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Foreword

The 3rd meeting of the Central European Tectonic Studies Group and the 10th meeting of the Czech Tectonic Studies Group took place in Felsõtárkány, Hungary, in April 2005. This was accompanied by the 1st meeting of the newly established Hungarian Tectonic Group (HUNTEK). It was decided at this meeting that the next one should be organized in Poland by the Galicia Tectonic Group, affiliated at the CeTeG since 2003. Reflecting to the announcement of the organizers, more than 120 participants registered at the 4th CeTeG meeting. Apart from the four nations of the Visegrád countries, contributions from Australia, Austria, France, Germany, Great Britain, Iran, Israel, Norway, Romania, Russia, Spain, Switzerland, and Vietnam will be presented.

The topics of the conference will concentrate on several aspects of tectonic, structural and petrological studies of sedimentary as well as of crystalline rocks in PANCARDI/Galicia area, Bohemian Massif and adjoining areas, Middle and Far East and high seas. It is also planned to organize sessions focusing on regional tectonics of the Carpathians, including recent results on neotectonics and the results of geophysical studies.

In Poland, the town Zakopane, located on the northern foothills of the Tatra Mountains, represents the conference location, whereas field excursions will mainly concern the medial portion of the Outer West Carpathians. One of these trips, "Structural development of the Magura Nappe (Outer Carpathians): From Palaeogene-Neogene subduction to Neogene-present-day collapse", will be devoted to the structural evolution of the Outer Carpathian accretionary wedge, including such topics as: early, soft-sediment deformation, succession of thrusting, large-scale Neogene rotations reconstructed from palaeomagnetic data, progressive vein mineralization, subduction-related(?) andesite intrusions and late Neogene to present-day deformation based on the analysis of fractured pebbles. The second trip, "Late Cretaceous–Neogene evolution of the Polish Carpathians", will focus on the Late Cretaceous – Neogene tectonosedimentary evolution of the Pieniny Klippen Belt and Outer (Flysch) Carpathian accretionary wedge, addressing: syntectonic evolution of the sedimentary basins, olistoliths, source areas, main tectonic units along the Zakopane-Kraków geotraverse, as well as timing and periodicity of thrusting.

Beautiful scenery of the Tatra Mts., Pieniny Klippen Belt, and Beskidy Mts., as well as local food, good beer and home-destilled specialties make Zakopane a perfect site for the conference. We do hope that these circumstances combined with high scientific level of participants' oral and poster presentations will contribute to a successful meeting.

The conference is sponsored by the International Visegrad Fund.

International
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Contents

4TH MEETING OF THE CENTRAL EUROPEAN TECTONIC STUDIES GROUP / 11TH MEETING OF THE CZECH TECTONIC STUDIES GROUP / 7TH CARPATHIAN TECTONIC WORKSHOP, ZAKOPANE, POLAND, APRIL 19–22, 2006: EXCURSION GUIDE

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Trace Element and Nd-Sr Isotopic Composition of Lamprophyres from the Ditrău Alkaline Massif, Romania

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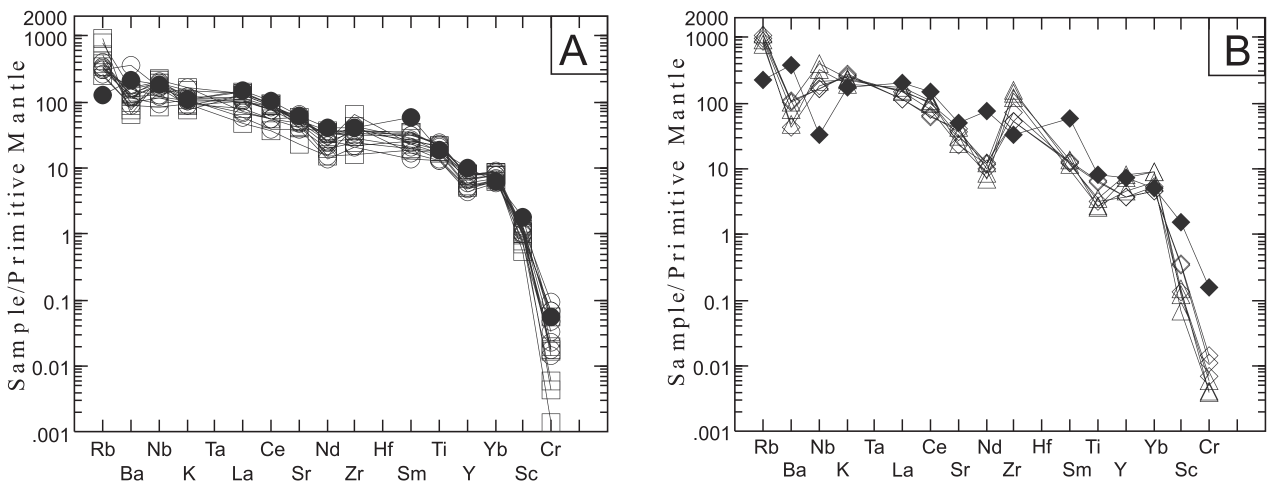
Lamprophyres are a group of alkali-rich igneous rocks containing essential amphibole and/or biotite-phlogophite and form subvolcanic dykes, sills, plugs, stocks, vents or margins to larger intrusions (Rock 1991). Alkaline lamprophyres (AL) have 'basanitic to nephelinitic' compositions, low-Si (41–43 % SiO_2), high-Na (3–4 % Na_2O) content and usually Na » K. Calc-alkaline lamprophyres (CAL) are 'shoshonitic' in composition and have high-Si (49–52 % SiO_2) and high-K (2–5 % K_2O) among lamprophyres. Both of them are rich in Sr, Rb, Ba, Th, Zr, LREE and volatiles (Rock 1991). This paper presents trace element, rare-earth element and radiogenic isotope data, and discusses variation in their chemical and isotopic composition.

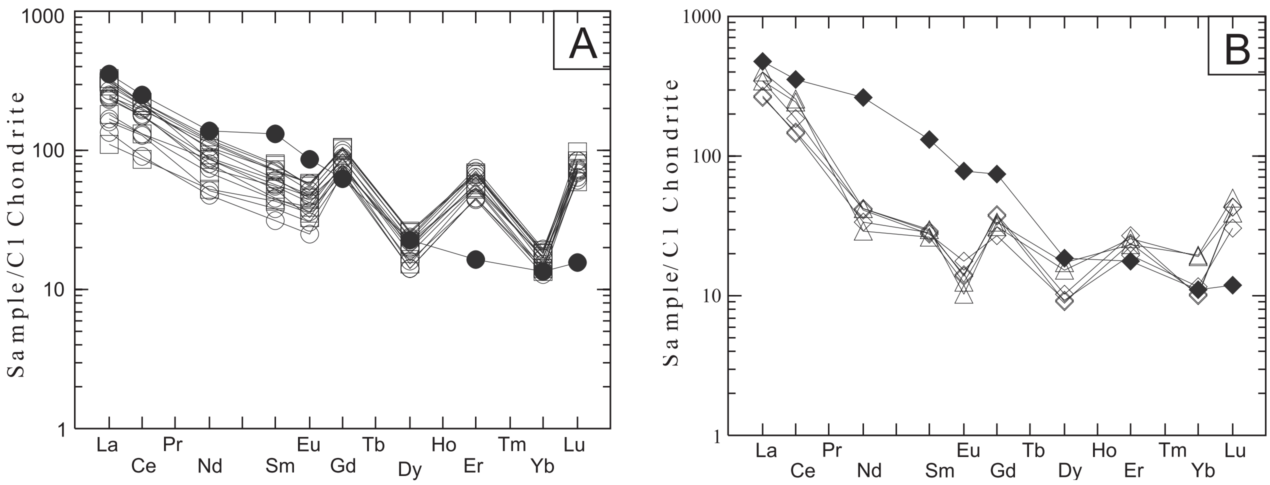
The Ditrău Alkaline Massif is a Mesozoic alkaline igneous complex and situated in the S-SW part of the Giurgeu Alps belonging to the Eastern Carpathians (Romania). This body intruded into the pre-Alpine metamorphic basement complexes of the Bucovinian Nappe Complex located on the east side of the Culimani-Gurghiu-Harghita Neogene-Quarternary calc-alkaline volcanic arc, and took part in the Alpine tectonic events together with these metamorphic rocks (Pál-Molnár 2000). The center of the DAM was formed by nepheline syenite, which is surrounded by syenite and monzonite. The northwestern and northeastern parts are composed of hornblendite, diorite (called Tarnica Complex, Pál-Molnár 2000), monzonite and alkali granite. The whole complex is cut by late-stage lamprophyre, alkali feldspar syenite and tinguaite dykes.

The studied area is the northern part of the DAM. The alkaline lamprophyres are camptonites (amphibole rich, plagioclase-bearing) and were collected from Tarnica Complex (Orotva, Tászok, Fülöp, Gudu Creeks), Török and Nagyág Creek. They are dark-grey, greenish-grey mafic rocks showing typical panidiomorphic texture and felsic globular structures. Camptonites from Tarnica Complex carry clinopyroxene phenocrysts, reddish-brown kaersutite, subordinate biotite microphenocrysts and interstitial plagioclase ($An_{4–34}$). Clinopyroxenes are aluminian subsilicic ferroan diopsides ($Ca_{0.9}Mg_{0.7}Fe_{0.2}Al_{0.26}Si_{1.73}O_6$). Kaersutites occur as euhedral groundmass minerals with up to 6.6 w% TiO_2. Subhedral biotites are strongly magnesian up to phlogopite (mg# = 0.73). Camptonites from Török and Nagyág Creek have only groundmass minerals without any phenocrysts. The groundmass consist of elongated magnesian hastingsite crystals (mg# = 0.49–0.55) displaying preferred orientation due to magma flow, biotite represented by small crystals with moderate mg# (0.47–0.50) and interstitial plagioclase ($An_{5–16}$). Apatite, titanite and magnetite are common minor phases in the studied camptonites. Secondary phases are tremolite-actinolite, chlorite, sericite and Ce-La allanite. Carbonate can occur as magmatic euhedral crystals and as secondary mineral as well.

The studied calc-alkaline lamprophyres occur only in alkali granites (northeastern part of the DAM) and are intermediate, less mafic dyke rocks. Samples collected from Török Creek are transitional types between minettes (biotite – K-feldspar rich) and kersantites (biotite – plagioclase rich). Rocks from Nagyág Creek are also transitional varieties amongst vogesite (amphibole – K-feldspar rich) and spessartite (amphibole – plagioclase rich). The minette – kersantite group contain andradite phenocrysts, phlogopite (mg# = 0.71–0.75), albite (An_1), K-feldspar (Or_{97}) and feldspathoids in the groundmass. The Ti-bearing an-



■ **Fig. 1.** Trace element patterns for the DAM lamprophyres normalized to primitive mantle (normalizing values from Taylor and McLennan 1985). ○ camptonites (Tarnica Complex), □ camptonites (Nagyág- Török Creek), ◇ minettes-kersantites (Török Creek), △ vogesites-spessartites (Nagyág Creek), ● AL average (Rock 1991), ◆ CAL average (Rock 1991).

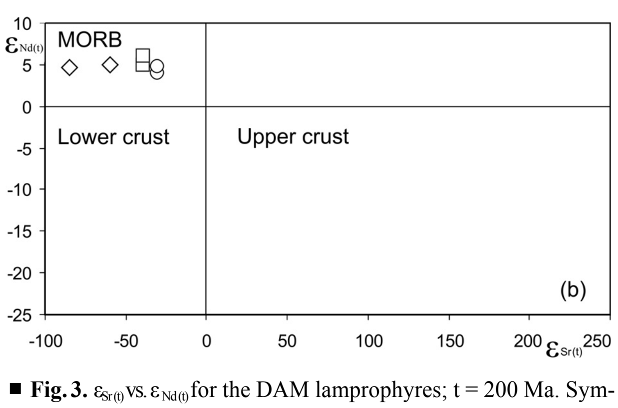


■ **Fig. 2.** Rare earth element patterns for the DAM lamprophyres normalized to C1 chondrite (normalizing values from Sun and McDonough 1989). Symbols are the same as in Fig. 1.

dradites are surrounded by secondary phlogopitic biotite, chlorite and magnetite. Feldspars are strongly sericitized. Other secondary phases are calcite and epidote. The vogesite – spessartite group carry clinopyroxene phenocrysts, ferro-eckermanite (CaO 3–10 w%, Na_2O 11–8 w%), ferro-richterite (CaO 12 w%, Na_2O 6 w%), small amount of magnesian biotite (mg# = 0.5–0.67), albite (An_0), K-feldspar (Or_{96}) and feldspathoids in the groundmass. Accessory minerals in all CAL are apatite, titanite, magnetite and zircon.

Major oxide composition were analysed on a Finnigan MAT Element spectrometer by HR-ICP-MS, the trace and the REE elements were determined by ICP-AES using a Varian Vista AX spectrometer at the Department of Geology and Geochemistry, University of Stockholm, Sweden. The concentration of major elements for the analysed samples falls within the range of alkaline and calc-alkaline lamprophyre as characterised by Rock (1991). Alkaline lamprophyres have low SiO_2content (42–50 %), high TiO_2 (1.7–3.7 %) content, high alkalies and incompatible trace elements such as LREE, Zr (629 ppm), Nb (124 ppm), Ba (1815 ppm), Rb (843 ppm), Sr (1142 ppm) and chondrite-normalized $(La/Yb)_N$ (>10) ratios. In terms of the mantle-normalized trace element abundances, the DAM alkaline lamlprophyres are similar to the average AL (Fig. 1A) described by Rock (1991). The chondrite-normalized REE patterns of the alkaline lamprophyres (Fig. 2A) show a decrease from La to Yb with strong positive Gd, Er and Lu anomalies compared to the AL average (Rock 1991). Calc-alkaline lamprophyres from the northern part of the DAM have higher SiO_2 content (54–57 %), higher alkalies (Na_2O 5–9.6 w%, K_2O w4–6 %), and much higher Zr (1301 ppm) than the previous AL group. The chondrite-normalized $(La/Yb)_N$ ratios are 17–29. The mantle- and chondrite normalized patterns slightly deviate in Rb, Ba, Nb, Cr (Fig. 1B) and strongly differ at Nd, Sm, Eu and Gd (Fig. 2B) from the CAL average (Rock 1991). The distinctive negative Eu anomalies (Eu/Eu^* = 0.33–0.65), the low Ti and Ba contents of the investigated CAL suggest that minettes – kersantites and vogesites – spessartites are more differentiated rocks, in contrast to the camptonites.

Sr and Nd isotopic data were obtained by a Finnigan MAT261 thermal ionization mass spectrometer at the Laboratory for Isotope Geology, Swedish Museum of Natural History, Sweden. Initial ratios were calculated using an age of 200 Ma. The DAM lamprophyres are characterized by variable $^{87}Sr/^{86}Sr$ (.7033–.7063) and composition of $^{143}Nd/^{144}Nd$ = .512729–.512832. Figure 3 displays the isotope results in a $\varepsilon_{Nd\,(t)}$ and $\varepsilon_{Sr(t)}$ diagram showing that the studied lamprophyres fall on the left side of the mantle array with $\varepsilon_{Nd\,(200\,Ma)}$ = +4.0 to +6.1. Based on the Sr-Nd isotopic composition the lamprphyres from the northern part of the DAM are mantle-derived rocks.



■ **Fig. 3.** $\varepsilon_{Sr(t)}$ vs. $\varepsilon_{Nd(t)}$ for the DAM lamprophyres; t = 200 Ma. Symbols are the same as in Fig. 1.

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Microstructural Analyse of Orthogneisses from the North-Eastern Part of the Bystrzyca Mts., West Sudetes

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Orthogneisses are one of the main lithological varieties in the Bystrzyca Mts., which is part of Orlica-Śnieżnik Dome, the easternmost unit of West Sudetes. Specimens were taken from the outcrops in Równia Łomnicka and Mostowice-Jagodna area. The main purpose of the analyse was determining temperature, kinematics and geometry of defomation, to which the gneisses were subjected. The method used in this study was measuring of the quartz *c*-axes orientation.

Diagrams for the eastern part of the Mostowice – Jagodna area are very similar. All indicate constrictional deformation (type II crossed girdles) under amphibolite facies temperature conditions. The microfabric asymmetry is "top-to-the-N". These gneisses could be classified as L<S tectonites (augen to augen-laminated) and S-tectonites (thin laminated).

Diagrams for the western part of this area are similar, but not so clear, and could indicate simple extension. C-axes microfabric shows asymmetry consistent with field observations of kinematics indicators (σ- or δ-clasts), but changing through the area. Deformation kinematics is here "top-to-the-S-SW". Gneisses have rodding texture (L-tectonites).

Diagrams for the Równia Łomnicka gneisses represent not fully-developed cross girdles. Sometimes it is hard to find, if it is type I or II. Positions of the maxima indicate also deformation in amphibolite facies conditions. Fabric asymmetry, consistent with field observations, is "top-to-the-N". These rocks have features characteristic for L-S-tectonites.

Spatial differentiation of deformation kinematics (top-to-the-N, top-to-the-S) with almost identically oriented foliation (gently dipping to the SW) is probably caused by folding of mylonitic foliation during the later tectonic event. It could be confirmed by existence of mineral lineation superimposed at low angle to earlier elongation lineation.

Summing up, texturally diversified Bystrzyca gneisses are metagranitoids. Deformation transforming granites into orthogneisses was multiphase. The stages were:

1. extension, that caused elongation of quartz-feldspar aggregates – lineation L_1,
2. flattening and shearing "top-to-the-N",
3. folding with eastern vergence, that caused mika flakes orientation – mineral lineation L_2.

Two first stages consisted in mylonitization connected with recrystallization of the gneisses mineral components.

Results of the quartz microfabric analysis indicate, that during the recrystallization:

- pseudohexagonal prism planes system in <*a*> direction dominated, what connected with maxima situated in the middle of the projection, indicates temperature conditions specific for amphibolite facies;
- strain was non-coaxial.

To a high degree microstructural and mesostructural features in the investigated orthogneisses are consistent.

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Composition of Biotites from Čierna Hora Granitoids (Western Carpathians) as an Indicator of the Granite Tectonic Setting

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Biotite – the dominant ferromagnesian mineral – has been analysed from Variscan granitoid rocks of the Čierna Hora Mts. (Western Carpathians) by an electron microprobe for completing of the existing data of their major elements and by Mössbauer spectroscopy for estimation of their Fe^{3+} vs. Fe^{2+} ratio. The direct relation between chemical composition of biotite and parental rock along with the presence of numerous minute inclusions of primary accessories trapped during biotite growth suggests the magmatic origin of biotite.

The analysed biotites exhibit a fairly wide range of X_{Fe} values and total Al content atoms per formula unit (apfu). Biotite from Ťahanovce area is characterised by high mean Al contents ~ 3.2 apfu and Fe/(Fe+Mg) values in the range 0.5 to 0.63. The positive correlation between Fe/(Fe+Mg) and Al_{tot} indicates the participation of sedimentary material on the granitoid petrogenesis. The trend of assimilation of aluminous crustal material to the magma is more significant in granites from Miklušovce complex because of higher mean Fe/(Fe+Mg) value (0.77) with more pronounced trend of increasing total Al (~ 0.8 apfu). On the other hand, biotites from Sokoľ and Sopotnica area show lower mean values of Al (approximately 2.97 apfu) and Fe/(Fe+Mg) ratio varies within 0.49 to 0.53 what indicate the I-type character of host rock. Concerning to the Fe valency, higher content Fe^{3+} (up to 20 wt. %) is characteristic for biotites from Sokoľ and Sopotnica granitoid bodies, whereas biotites from Ťahanovce granitoid massif show decrease Fe^{3+} content (around 5 wt. %). Such relation indicates the typical I-type oxidizing conditions due to the presence of higher water content during Sokoľ and Sopotnica granitoid evolution. Biotites from Ťahanovce area imply more reducing conditions with lower water content, and this is characteristic for the S-type granites.

According to the biotite chemistry we assume the affinity of granitoids from Sokoľ and Sopotnica massifs to the I-type granitoid suite which has been formed from slightly differentiated magma with mantle contribution. Contrary to it, Ťahanovce granitoid body and granitoids from Miklušovce complex show affinity to the S-type granitoid suite due to the precipitation of Fe-biotites from multiply contaminated melt by crustal material.

Granitic Rocks from Branisko Mts. (Western Carpathians): Geochemistry, Mineralogy and Tectonic Implications

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On the basis of textures, mineral composition and geochemical characteristics, the granitoid rocks of Branisko crystalline basement form two separate main groups: 1) syn-collision peraluminous leucocratic granites and granodiorites widely distributed in the S and W part of Branisko crystalline basement; 2) post-collision granodiorites inhabited mainly in the NE part of mention crystalline basement. Available mineralogical and geochemical data reveal that these two groups can be characterised by different magmatic evolution or protolith history.

The first group shows rather evolved geochemical characteristics. Major and trace element geochemistry of (leuco)granites clearly indicates their crustal origin. The main rock-forming minerals are K-feldspar + quartz + albite (An_{0-5}) + muscovite; essential accessory mineral phases are apatite, zircon (S_{1-3}; L_{1-3} types), monazite, xenotime, garnet ± rutile whereas REE contents (La_N/Yb_N ~ 19) are particularly controlled mainly by monazite. EMPA dating of monazite yielded age 342 ± 15 Ma for leucocratic granites (Bónová et al. 2005). Granodiorites which are occurred in western part of Branisko crystalline basement show slightly different features in comparison with granodiorites–tonalites from NE side. Higher volume of K-feldspars and significantly lower content of biotite or other mafic minerals is their dominant feature. Biotite exhibits a high total Al contents, reaching up to ~ 3.25 apfu. The biotite samples from investigated granodiorites define a relatively narrow range of Fe/(Fe+Mg) values, from 0.51 to 0.54 apfu and higher contents of TiO_2 (around 3.8 wt.%). Accessory minerals are zircon, apatite, rutile and monazite.

The granites of S and W parts of Branisko crystalline complex generally display affinity to S-type granites. In particular, we suggest that leucocratic granites have been formed by crystal fractionation within main meso-Hercynian (350–330 Ma) period as a part of S-type granite suite of the Western Carpathian basement complexes or as a result of continental collision during Missisipian (Lower Carboniferous).

The second group show less evolved geochemical character, biotite tonalite with hornblende and granodiorite prevail. Typical is enrichment in Ti, Sr, Zr and incompatible elements; normalized REEs are plotted with steep pattern ($La_N/Yb_N \sim 29$; Kohút et al. 2003). The main rock-forming minerals of these granitoids are represented by plagioclase (An_{34}) + quartz + K-feldspar + biotite. Biotite is defined by low Fe/(Fe+Mg) ratio (0.47 apfu) and total Al content reaches maximal 2.79 apfu. Significant abundance of titanite in an investigated rocks is compensated by lower TiO_2 content in biotite (1.89 to 2.73 wt.%). Typical accessories are zircon ($S_{12,16,17}$ subtypes), allanite, titanite, apatite and magnetite. The accessory mineral assemblage indicates an oxidation conditions in magma and geochemical aspects of investigated granodiorites suggest their competence to I-type suite of granitic rocks. This magmatism, similarly as in the other core mountains of the Western Carpathians, is rooted in the low crust and emplaced rather high to the middle crustal level. I-type plutonism in Western Carpathians is interpreted as an independent pulse during extension or transtensional regime within meso-Variscan orogenesis (Broska and Gregor 1992, Broska and Uher 1991, 2001, Haunschmid et al. 1997, Petrík et al. 1994, Petrík and Kohút 1997, Petrík et al. 2001 and other). Consequently, we assume that granitoids in Branisko crystalline complex of S and I-type affinities are genetically joined with two separate tectonomagmatic events.

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Attempt to Dating of Accretion in the West Carpathian Flysch Belt: Apatite Fission Track Thermochronology of Tuff Layers

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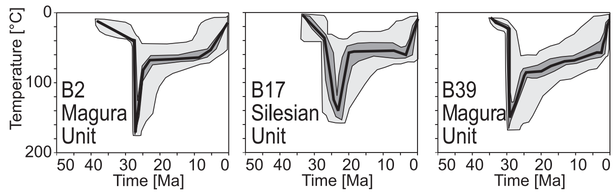
The Carpathians are a part of the European Alpine chain created by convergence and collision of the European and African plates (Golonka et al. 2000). The Outer Western Carpathians are a north-verging fold-and-thrust belt composed largely of Lower Cretaceous to Lower Miocene flysch sediments arranged into stacked complex of several nappes (from top to bottom: Magura, Dukla and Fore-Magura, Silesian, Sub-Silesian and Skole nappes. The tectonic evolution of the Outer Carpathians is subdivided into two successive shortening events: 1) NNW- (N) directed and 2) NE- (NNE) directed (Aleksandrowski 1989, Decker et al. 1997). During the first event the folding and thrusting started in the most inner, southern nappe (Magura nappe) and were propagated to the north. During the next event the previous thrust faults and folds were overprinted and refolded. According to Decker et al. (1999a), the first shortening event was taking place from the Eocene to Early Miocene time. The second shortening event started in the Early/Middle Miocene and lasted probably during the early Late Miocene time (Decker et al. 1999b, Wójcik et al. 2001). Organic maturation, clay diagenesis and fluid inclusion (Swierczewska et al. 1999, Hurai et al. 2002, Jarmolowicz-Szulc and Dudok 2005) of the different flysch units of Western Carpathians indicates that the nappes underwent some thermal overprint during their burial and/or the tectonic stacking.

We have performed apatite fission track (FT) thermochronology on ash layers and vitrinite reflectance measurements on organic rich pelites of Western Carpathian flysch in order to reconstruct the thermal and structural evolution of the flysch belt. The apatite FT dating were completed by the investigation of kinetic parameters in each dated crystals by Dpar measurements (Donelick 1995) and in some samples also by the determination of chlorine content using electron microprobe. The fission track results and Dpar measurements show rather homogeneous mineralogical composition of the dated grains and also indicate that the apatites of the samples derived form single sources. The only exception is sample B17, where we could identify two components, but by their separation it was possible to perform thermal modelling on a homogeneous sub-population. All FT ages are younger than the age of sedimentation or the dated age of the volcanic event. Geographically/structurally the age pattern is complex: the apparent FT ages are between 20 and 30 Ma in the Podhale Flysch (Inner Carpathians), along the PKB and the innermost part of the Magura nappe. In the Silesian nappe the ages cover a wider ranger between 9.7 and 29.5 Ma. The track length distributions are rather variable; the mean track lengths are between 13.6 and 15.2 µm. Mean track lengths do not show evident correlation with the apparent ages. Geochemically, these ash-layers apatites are usually chlorine-rich. The mean Dpar are between 2.6 and 3 µm and except sample B17 the chlorine content is always above the typical composition of Durango apatite (~0.4 wt%), thus both kinetic parameters indicate rather high closure temperature for the dated apatites.

The reset of fission track ages and the temperature indicators are evidently indicating that all investigated tectonic units have experienced significant burial. The post-sedimentary maximum temperature was around or above 100 °C. The apparent ages are different and the character of track length distributions is very variable, thus we can conclude that the investigated tectonic slices had rather different thermal histories. Considering the new data and Tmax determinations by clay mineralogical constraints further the results of fluid inclusion studies we have modelled the thermal evolution by HeFTy software (Ketcham, 2005). The input parameters of the thermal modelling are fission track apparent age, track length distribution, kinetic parameters (Dpar and chlorine content), formation age of the ash layer, time of the termination of sedimentation (by the overriding accretional wedge), temperature of sedimentary burial and maximum temperature. A very characteristic feature of the input data is the closeness of the end of sedimentation and the apparent FT age in case of every sample. The modelling resulted in slightly different time-temerature paths for the different structural units (Fig. 1). What is common in the thermal evolutions of the samples is the presence of the rather rapid increase of the burial temperature. After the end of sedimentation the temperature has reached the total reset conditions of apatite FT chronometer usually within 2 to 3 million years. The effective heating time is usually short, and soon after the thermal climax the flysch nappes were already cooling. The cooling rates are variable, but sometimes rather rapid (up to 50 °/Ma). We interpret these common characters of the thermal histories as the manifestation of the accretion process in the thermal history: (i) the rapid warming was related to the accretion of the flysch related with subduction process; (ii) the abrupt turn in the thermal histories were related to the collision process; (iii) the cessation of the increasing of temperature indicate the termination of the vertical displacement; (iv) the rapid cooling is probably related to normal faulting that exhumed some slabs of the accretionary complex faster than usual surface erosion.



■ **Fig. 1.** The results of thermal modelling of some charakteristic samples performed by HeFTy software of Ketcham (2005). Light gray fields represent the envelope of acceptable time-temperature paths, while dark gray belts are the enveloped by the runs of modelling with very good fitting to the measured data.

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REE Accessory Minerals in the Gneiss and Granulite Clasts from the Silesian Unit (Western Outer Carpathians, SE Poland) as Indicators of Metamorphic Processes

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Introduction

REE accessory minerals play the important role in metamorphic petrology. Their significance is related to their hosting of trace elements in metamorphic rocks and, therefore, providing important information about evolution that these rocks experienced. Phosphates (especially apatite, monazite and xenotime) are probably the most important group of metamorphic minerals, because they might be used in wide range of investigations, including geochronology and geothermobarometry (see e.g. Spear and Pyle 2002 and references therein). Moreover, reactions in which these minerals took part might provide significant information about history a rock experienced (e.g. Finger et al. 1998, Wing et al. 2002).

Gravel size extrabasinal clasts (so-called "exotics") of gneisses and granulites, which are present in the Silesian Unit (Western Outer Carpathians), preserve mineral assemblages, as well as structural relationships between minerals, that might provide important information about evolution of their source area – the Silesian Ridge. Monazite-(Ce), xenotime, apatite, zircon, uraninite and unidentified Th-phases as the main host of REE in studied rocks are roughly described in this report.

Sample selection and methods of investigation

Twenty one samples of gneisses and four samples of granulites collected in six localities (Bukowiec, Gorlice, Izdebnik, Krzesławice, Siekierczyna and Skrzydlna regions) in the Silesian Unit were studied. Transmitted light microscopy and SEM-EDS method were used during investigations. Samples are considered to be derived from the Silesian Ridge.

Chemical compositions of minerals were determined using cold field emission scanning electron microscope (FESEM) Hitachi S-4700 coupled with energy dispersive spectrometer (EDS) NORAN Vantage. Analyses were performed in the Laboratory of Field Emission Scanning Electron Microscopy and Microanalysis at the Institute of Geological Sciences of the Jagiellonian University, Kraków.

Results

Gneisses are mainly composed of plagioclase (ranging from oligoclase to andesine), quartz, biotite, muscovite and K-feldspar.

Apatite, zircon, monazite-(Ce), xenotime, uraninite, rutile, chlorite, Fe-Mn-garnet, epidote, allanite, barite, calcite, ankerite, Fe and Ti oxides, Fe, Zn, Cu and Pb sulphides are present as accessory minerals. Main minerals of granulites are as follows: plagioclase (oligoclase or andesine), Fe-Mg-garnet, alkali feldspar (microperthite), quartz and minor biotite, with accessory kyanite, apatite, monazite-(Ce), zircon, rutile, iron and titanium oxides, iron and zinc sulphides. Various degree of sericitization of plagioclase, as well as chloritization of biotite can be noticed in analysed samples. More detailed information about descriptions of samples are given by e.g. Budzyń et al. (2005, 2006, in print).

Monazite-(Ce) is present in most of studied samples and occur as primary and secondary mineral. Primary monazite-(Ce) forms subhedral to anhedral grains, up to ca. 70 × 210 µm. Monazite-(Ce) breakdown and its replacement by apatite is relatively common. Unusual formation of secondary monazite-(Ce) was stated in gneiss clast (sample B-2) from Bukowiec (see also Budzyń et al. 2006, in print). Monazite-(Ce) forms lamellae and irregular aggregates within and/or around calcite. These are usually enclosed in biotite and, rarely, are present at the borders of biotite crystals. Monazite-(Ce) from gneisses contains 3.24 wt.% of ThO2 and 0.66 wt.% of UO2 on average, and from granulites – up to 23.16 wt.% of ThO2 and up to 1.08 wt.% of UO2. Average Nd2O3/Ce2O3 ratio in gneisses and granulites is equal to ca. 0.35 and 0.38, respectively.

Xenotime forms anhedral grains, up to ca. 150 × 55 µm in size, and is present in samples from Bukowiec, Izdebnik and Skrzydlna regions. Monazite-xenotime pair (both grains are <15 µm in size), where xenotime partially mantles monazite was noticed in gneiss (sample B-1) from Bukowiec.

Zircon usually occurs as tiny, zoned grains (ca. 10 µm in diameter). However, grains up to ca. 220 × 90 µm in size were also noticed. Complex zoning pattern with xenocrystic cores mantled by newly grown magmatic zircon is present in large grains.

Allanite is present only in one gneiss sample (sample B-5) from Bukowiec and forms anhedral grains, up to ca. 10 × 30 µm in size. This mineral occurs as inclusion of monazite or contains inclusions of monazite.

Inclusion of uraninite (ca. 4 × 25 µm in size) in muscovite was noticed in gneiss (sample G-6) from the Gorlice region. Uraninite contains 92.67 wt.% of UO2 and 3.00 wt.% of PbO2. Moreover, unidentified Th-phases that accompany apatite were noticed in gneiss (sample K-2) from Krzesławice region.

Conclusions

Origin of zircon, as well as primary monazite is probably related to Precambrian and/or early Caledonian magmatic events that were stated in other exotics (see Michalik et al., this volume). Further investigations will provide more data concerning gneiss protoliths, as well as P-T conditions and timing of metamorphism which affected studied rocks.

Allanite-monazite reactions, formation of uraninite and unidentified Th-phases are probably related to relatively low temperature hydrothermal alterations. Chloritization and sericitization, as well as presence of barite, iron oxides, and iron, zinc, copper and lead sulphides also suggest relatively low temperature alterations. REE mobilization, and formation of Th-phases and uraninite might occur during the late Carboniferous-early Permian metamorphic episode dated in other clasts (e.g. Michalik et al. 2004, Poprawa et al. 2004). Younger age (ca. 225 Ma) was roughly determined using chemical method based on uraninite composition (content of U and Pb).

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Geology and Tectonics of the Vršatec Klippen Area (Pieniny Klippen Belt, Western Slovakia)

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The Pieniny Klippen Belt (PKB) represents a tectonic zone separating the External (Outer) Western Carpathians from the Central (Inner) Western Carpathians. It is a narrow, independent structural belt with extraordinarily complicated structure, which stretches from the vicinity of Vienna, through the Považie (Váh River valley) to the Orava region and extends via the Polish Pieniny Mts. to eastern Slovakia, Ukraine and Romania. Primary absence of pre-Mesosoic rocks, scanty representation of Triassic rocks, variable developments of Jurassic and Cretaceous successions and klippen-type (block-in-matrix) tectonic style are the principal characteristics of the Klippen Belt. Most of the klippen are tectonic by origin, though olistolites are present in places. According to the present views, the main phase of tectonisation and separation of the klippen took place after the Eocene–Oligocene and before the Sarmatian (Hók and Kováč 1996, Potfaj 1998). Our study concentrates on the structure and tectonic evolution of one of the most conspicuous parts of the PKB – the so-called Vršatské bradlá group of klippen in the Middle Považie region of western Slovakia. Taking as a whole, the Vršatec area represents the largest Czorsztyn-type klippe or a group of klippen in the Slovak part of the PKB.

The studied terrain is formed by two partially independent segments – the Vršatec-Javorník and Chmeľová regions. The southern, SW-NE trending, Vršatec-Javorník row of picturesque blocky klippen is composed of massive Jurassic – Lower Cretaceous limestones forming a steeply NW-dipping monoclinal slab with overturned stratigraphic sequence belonging to the Czorsztyn Unit. The sequence starts with the Upper Liassic – Aalenian dark hemipelagic marlstones (Krempachy Fm) followed by massive, light bioherm limestones (Vršatec Fm.). The latter were considered to be of Oxfordian age before (Mišík 1979). However, the recent study by Schlögl and Tomašových (in press) indicates their significantly older, most probably Lower Bajocian age. This age would make the structure of the Vršatec-Javorník klippen much simpler – only one overturned sequence is present, on contrary to two slices (one with normal, the other with overturned sequence) as proposed by Mišík (1979). Bioherm limestones are followed by crinoidal limestones (Smolegowa and Krupianka Fms), red nodular limestones (Czorsztyn Fm) and whitish to pink biodetrital limestones (Dursztyn Fm). From the SE side, this about 50–150 m thick slab of competent limestones is in contact with Upper Cretaceous red pelagic marlstones of the couches-rouges type (Púchov Fm). Both Lower Jurassic and Upper Cretaceous marlstones form the so-called "klippen mantle", i.e. a soft matrix in which the stiff klippen are embedded. The lithological contacts, though generally in stratigraphic sequence, are tectonically reactivated in most cases. To the SE, the Púchov marlstones are juxtaposed to various sediments of distinct units participating on the PKB structure (Orava and Drietoma Units – Schlögl et al. 2000).

The slab of competent Middle Jurassic – Lower Cretaceous limestones of the Vršatec-Javorník zone is truncated by numerous fractures and slickensides. The field study revealed the presence of oblique reverse faults and oblique dextral strike-slips, which are concentrated on the lithological boundary between the rigid klippen and the klippen mantle. These steeply dipping faults cut obliquely bedding and obviously control the lozenge shape of individual, variably large (metres to hectometres) blocks (klippen) into which the Vršatec-Javorník slab is divided. Unfortunately, due to weathering of fault surfaces in natural outcrops, only a limited number of faults suitable for kinematic and paleostress analyses were found and measured. Another group of spaced subvertical faults oriented at high angles to bedding may represent the transfer (tear) faults. In addition, marlstones of the Púchov Fm bear traces of an older deformation event recorded by planparallel solution cleavage that is slightly oblique to the bedding surfaces. Part of stylolites found in massive limestones might have recorded this event as well.

In the Chmeľová area, the Czorsztyn and/or a sort of "transitional" (akin to Niedzica-Pruské) succession crops out. It slightly differs from the above-described Czorsztyn succession by beds of allodapic crinoidal calciturbidites in the upper part of the Krempachy marl Fm, lack of the reef Vršatec limestones, much thinner Middle Jurassic sandy-crinoidal limestones, presence of radiolarite lenses (Czajakowa Fm) within the nodular limestones (Niedzica and Czorsztyn Fms) and by a newly discovered body of volcanic rocks. These occur in a core of a macroscopic syncline above the Dursztyn limestones and probably below the Púchov marls. The volcanic body is about 50 m wide and over 300 m long and consists of hyaloclastites of basanitic composition, which are partly mixed with surrounding sediments. Consequently, the volcanics are most probably of Early Cretaceous age and are a normal component of the stratigraphic succession as a submarine lava flow. It is one of the very scarce occurrences of Mesozoic volcanics within the PKB successions in Slovakia, but surely the largest one, for the first time discovered in this area (cf. Hovorka and Spišiak 1990, Mišík 1992).

This unit is internally tectonically more complicated, with alternating sectors of normal and reversed stratigraphic sequences. Unlike in the Vršatec-Javorník area, this region is dominated by macroscopic fold structures. Folding was enabled by a much thinner competent limestone layer (some 20–30 m only) sandwiched between incompetent strata. Folds are upright, slightly asymmetric, with SW-NE trending axes and with locally penetrative subvertical axial plane cleavage. The map view and presence of numerous slickensides reveal that folding was followed by reverse and strike-slip faulting that finally shaped the klippen tectonic style of the area.

In the most external position in the Chmeľová region, a narrow strip of the Kysuca Unit was discerned. It involves deep-water pelagic Jurassic strata (predominantly marlstones and radiolarites), but the Lower Cretaceous sediments are of special type with allodapic biodetrital limestones (Horná Lysá Fm – Mišík et al. 1994). The position of the Kysuca Unit is not clear; it is likely in an original tectonic superposition above the Czorsztyn-related units. To the NW, the Kysuca Unit is juxtaposed to the Maastrichtian – Paleocene flysch sediments of the Biele Karpaty Unit of the Carpathian Flysch Belt. This contact is most probably followed by a large wrench fault that forms the northern boundary of the PKB.

The PKB was formed during multistage ductile-brittle and brittle tectonic evolution that occurred in several deformation stages producing variable fold and fault structures. Probably the oldest event was thrusting of the presently most external Kysuca Unit over the Czorsztyn and Niedzica-Pruské Units. Then macroscopic folding due to orthogonal layer-parallel shortening affected especially the Chmeľová region with well-bedded Niedzica-Pruské succession, while the thick competent Vršatec-Javorník slab was steepened and partly overturned to the SE. Folding occurred at comparatively greater depths (but still beyond intracrystalline ductile deformation mechanisms) with development of cleavage formed by pressure solution. This strong compressional stage was followed by transpression, which is recorded by several generations of striated, mostly steeply dipping faults. Numerous reverse and strike-slip faults truncated the stiff limestones sandwiched between incompetent strata and produced klippen of two distinct morphostructural types: 1. the Vršatec type formed by vertical strata obliquely cut by faults into variously large, lozenge-shaped blocks arranged in one straight row; 2. the Chmeľová type with a more random arrangement of variously shaped klippen, dependent on which parts of pre-existing macrofolds (cores, limbs) were separated into klippen that slightly moved with respect to each other afterwards. Finally, the klippen style was affected also by slope movements and some independent blocky klippen are obviously loose blocks transported downslope by landslides.

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Tectonic Deformations in the Orava Basin Margins in the Western Carpathians, Based on the DEM Analysis and Geological Research

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The tectonic structure is the main issue of the Orava Basin geology but the Quaternary deposits make difficult presentation of the depression architecture. For the recognition of the linear morphological features which could depict the basement dislocations of the Orava Basin the digital elevation model (DEM) was made. Morpholineaments were drawn along straight segments of rivers and morphological edges. It allows distinguishing two groups of morpholineament – directions: the morpholineaments which form the margins of the Orava Basin and the morpholineaments which are perpendicular to the first group of lineaments and disturb them. Morpholineaments in the contour map, shaded relief map, and the DEM refer to fault lines buried beneath deposit cover which fill the Orava Basin (Chrustek and Golonka 2005). Moreover, the DEM allows determining the present image of the Orava Basin, relationship between ambient elevations and the Orava depression and effects of southern fringes uplift and northern fringes subsidence.

The geological research is in preliminary phase and hasn't been done yet in the whole investigated area. The tectonic deformations are represented by ductile and brittle faults which appear with breccias and also mixed faults in the north and south margins of the Orava Basin. The mentioned dislocations are longitiudinal and obligue faults. Moreover, most of them have strike-slip character. The field research allows determining places where the largest deformations are located as well. They are situated along south-east in the neighbourhood of the Pieniny Klippen Belt and north margin of the Orava Basin. The south-east fringe forms a steep scarp made up of Neogene gravels containing fractured clasts. The orientation of fracture sets observed in several outcrops from Domański Wierch to Stare Bystre in this area determines NE and NNE stress direction. According to Tokarski and Zuchiewicz (1998) these outcrops reflect dislocations in the basement. Probably these dislocations are active nowadays because along south-east fringe of the Orava Basin runs active zone called Myjava lineament (Bac-Moszaszwili 1993) or the Orava transforming fault (Baumgart-Kotarba *et al*. 2004). Besides, there are known earthquakes which had compressional character according to Baumgart-Kotarba (2001). In the northern part of the Orava Basin occur local overthrusts. In the border of thrust are reverse faults with overturned strata of the Magura series. These faults must have formed during compressional regime.

Directions of morpholineaments and faults founded in the investigated area are similar. It means that the faults situated on the surface are reflected by the morpholineaments distinguished on the basis of the DEM analysis.

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Olistostroms as Indicator of the Geodynamic Process (Northern Carpathians)

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Existence and transformation of the accretionary wedge in the southern in the Northern Carpathians are documented by occurrence of olistostroms. During the detail mapping and lithofacies investigations of deposits in a few main tectonic units of the Outer (Flysch) Carpathians the olistostroms has been identified. In Poland, Czech and Slovak Republic the olistostroms are known from the Cretaceous, Paleocene, Eocene, Oligocene and Miocene flysch deposits of a few main tectonic units. Those units are the Skole, Subsilesian, Silesian, Dukla and Magura nappes as well as Pieniny Klippen Belt.

The term olistostrome is derived from the ancient Greek and means «slide-layer» (Cieszkowski and Golonka 2005). An olistostostrome is a sedimentary deposit consisting of blocks of diverse origin that are immersed in a matrix. The Northern Carpathians this matrix consist of clay, mud, sand or their mixture forming turbidity package. The blocks in olistostrome are named olistolites. The size of olistolites varies, from centimeters to kilometers. Very large blocks could slide independently into the basin with no easily distinguishable matrix. The matrix in this case is the flysch sequence or even entire sedimentary-tectonic unit. The olistostroms formed in Northern Carpathians as debris flows during the different stages of the development of flysch basins, from rift trough post-rift to the orogenic stage.

In the southern Part of the Polish Northern Carpathians as well as in the adjacent part of Slovakia the olistostroms are known from the Cretaceous-Paleocene flysch deposits of the Pieniny Klippen Belt Złatne Unit and in Magura Nappe marking an early stage of the development of the accretionary prism. The most spectacular olistostroms have been found in the vicinity Haligovce village in Pieniny Klippen Belt and in Jaworki village in the border zone between the Magura Nappe and Pieniny Klippen Belt. The olistolites and large clasts are represented by igneous rocks including possible ophiolite basalts as well as a variety of carbonate rocks of Triassic - Paleogene age. This material represents the former PKB basinal and ridge sequences as well as Inner Carpathian terrane sequences. The Haligovce Klippen and Homole block represent largest Pieniny Klippen Belt olistolites (Golonka et al. 2005).

The olistostroms of different age are especially frequent in the Silesian Nappe. They have been created in some different stages of evolution of Silesian basin. Oldest are known from western-most part of Polish Outer Carpathians. In the Cieszyn beds olistostroms arrived during the Late Jurassic – Early Cretaceous extensional stage forming the Silesian basin. In the basal part of the Godula beds (Turonian – Campanian) represented by very thick-bedded sandstone turbidites a large flat blocks of the shales derived from the Lgota beds (Albian–Cenomanian) lies on slumped beds (Ślączka and al. 2005).

In area surrounding artificial Rożnów Lake a few olistostrome horizons are known from the Istebna beds (Maastrichtian–Paleocene) and from the Hieroglyphic beds (Middle-Late Eocene). There the large flat or plastically folded blocks of flysch deposits. Blocks of marls and occasionally limestones inhere in the debris-flow sandy-gravel matrix with pebbles. The pebbles represent different sedimentary, metamorphic and magmatic rocks.

During Late Paleogene final stage of the Silesian accretionary prism numerous olistostroms were deposited within the Oligocene-Early Miocene Menilite and Krosno Beds. In the abandoned quarry in Skrzydlna in the basal part of the Cergowa Sandstone in the Menilite beds (Oligocene) occur large olistostroms (Cieszkowski and Polak 2001) composed mainly of the Lower Cretaceous flysch deposits which represent Cieszyn beds, Hradiste beds and Verovice beds and minor addition of the Eocene gray or red marls and shales. The debris-flow deposits with pebbles of different sedimentary and crystalline rocks are frequent. The Silesian Ridge, which framed the Silesian Basin from south, was overridden by accretionary prism. Then, the ridge basement rocks, Paleogene deposits of the slope as well as older Cretaceous flysch deposits partly folded and thrust within the prism were slided northward toward the basin, forming the olistostrome.

The Sub-Silesian ridge deposits were partially included into the Subsilesian nappe, the ridge's basement rocks and part of its depositional form olistostroms and exotic pebbles within Menilite--Krosno flysch. The largest olistostroms were found in the vicinity of Andrychów and are known as Andrychów Klippen (Golonka et al 2005). The Fore-Magura and Silesian ridges were destroyed totally and are known only from olistolites and exotic pebbles in the Outer Carpathian flysch. They destruction is related to the advance of the accretionary prism. This prism obliquely overridden the ridges leading to the origin of the Menilite-Krosno basin. The Malcov Formation was deposited in the smaller piggy-back subbasin. During overthrusting the outer, marginal part of the advanced nappes was uplifted whereas in the inner part sedimentation persisted in the remnant basin. From that, uplifted part of the nappes big olistolites glided down into the adjacent, more distal basins. The nappes became detached from the basement and were thrust northward in the west and eastward onto the North European platform with its Miocene cover. Overthrusting movements migrated along the Carpathians from the west towards the east. The Outer Carpathian allochtonous rocks, as result of Miocene tectonic movements, have been overthrust onto the platform for a distance of 50 to more than 100 km.

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The Fault Tectonics of the Middle Skawa River Valley (Polish Flysch Carpathians)

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Upper Cretaceous and Paleogene rocks of the Silesian and Magura Nappes build up the complex structure of the investigated area, which is stretching along the Skawa River valley between Sucha Beskidzka and Świnna Poręba (Książkiewicz 1972, 1974a,b, Golonka et al. 2005). Middle and Upper Godula Beds, Istebna Beds, Eocene red shales and Ciężkowice Sandstone as well as Hieroglyphic, Menilite and Krosno Beds form the Silesian Succession. The sedimentary succession typical for the so-called Siary unit forms the Magura Nappe, which includes the following lithostratigraphic units: Jaworzynka Formation (Ropianka Beds) with Gołynia Shale member and Mutne Sandstone Member within the Paleocene part of profile, Łabowa Formation with the Skawce Sandstone Member, Beloveza Formation, and Beskid Makowski Formation with Zembrzyce Shale Member, Wątkowa Sandstone Member and Budzów Shale Member.

The Silesian Nappe stretches from Moravia (Czech Republic) to Ukraine where it looses its individuality (Golonka et al 2005). In the western segment of the Polish Carpathians, the Silesian Nappe is flatly overthrust onto the substratum. The southern part of the Silesian Nappe is hidden beneath the Magura Nappe. Between the Soła and Skawa Rivers, the Silesian Nappe is built up of several gently folded structures. According to Książkiewicz (1977), the imbricated folds gradually become more and more marked eastwards. The dislocation system of faults of located along the Skawa River known as so-called Skawa Fault divides the area into two different parts. East of the river the main structures of the Outer Carpathian fold-and-thrust belt have orientation East-West, while west of Skawa this orientation changes to WSW–ENE. Also the main overthrust of Silesian Nappe displaced 10 km northward west of the dislocation system (Książkiewicz 1972, 1974a,b). The western part f the nappe is included into the large Beskid Mały block. Along the fault Klecza Dolna–Łękawica–Dąbrówka (Książkiewicz 1974a,b), which belongs to the Skawa dislocation system , this block contacts the Pogórze Lanckorońskie Zone. The Beskid Mały block is uplifted, so the upper Cretaceous rocks contact along the Fault the Oligocene deposits of the Pogórze Lanckorońskie. The Merkowa, Jaszczurowa, Berszcz, Mucharz, Mucharza II, Świnna Poręba, Zagórz and Zagórza II, NNW-SSE faults cut the Beskid Mały block into smaller segments. Between Berszcz and Mucharza faults the pull-apart depression was formed during the strike-slip left-lateral activity of these dislocations (see Zuchiewicz 1998, Zuchiewicz et al. 2002, and Golonka et al. 2004). Similar depression exists in Skawce-Zembrzyce area within the Magura Nappe.

The Magura Nappe margin is displaced 5 km southward east of Skawa River. This displacement display gradual step-like character along several smaller NNW-SSE oriented faults. The so-called Krzeszów tectonic half-window is located here. The Magura Nappe has the independent fault systems, some of its dislocations have been acquired, however, from the Silesian Nappe.

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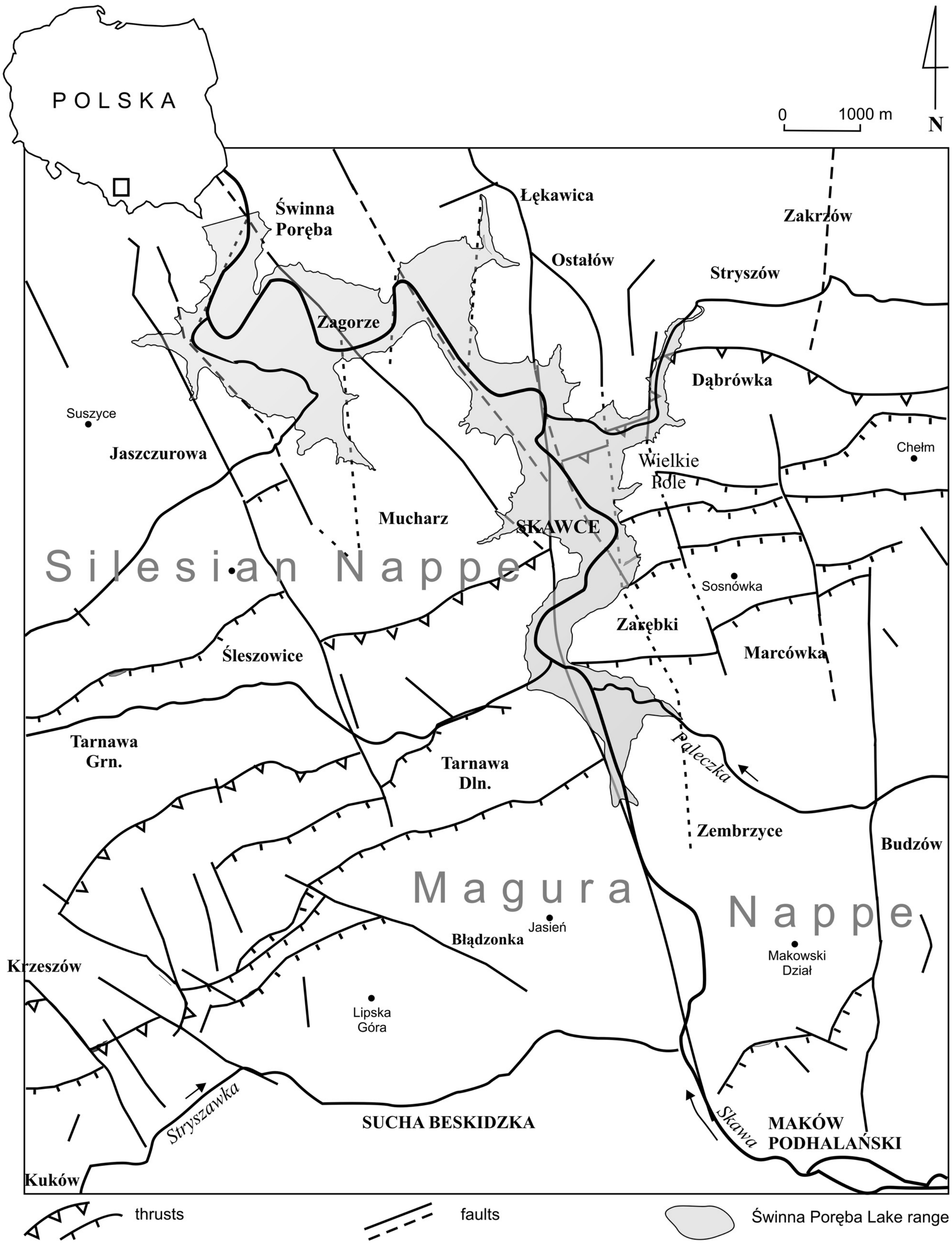

■ Fig.1 Tectonic map of the middle Skawa river valley (Polish Flysch Carpathians)

Are There Olistoliths on the Eperkés Hill? The Paleomagnetic Answer

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Epekés Hill is a classic exposure of the Jurassic-Lower Cretaceous successions in the Transdanubian Range, Hungary. Although the exposures of the Hill have been thoroughly studied during the last 40 years, facies interpretation is still subject of debate. According to some (Fülöp 1964, Konda 1970, Császár 1988a, 1988b) Upper Jurassic beds overlay the eroded surface of the Upper Triassic–lowermost Jurassic carbonates. In contrast, Galácz (1989) suggested that the Upper Triassic–lowermost Jurassic carbonates are large blocks (megabreccias or olistoliths) embedded into the Kimmeridgian through Berriasian limestones.

Recently, the results of new resistivity measurements (Palotai et al., in press) were interpreted as geophysical support for the megabreccia concept, since high resistivity patches (probably platform carbonates) seemed to swim in low resistivity material (probably pelagic limestones).

As the existence or absence of megabreccia is crucial for the interpretation of the structural evolution and paleogeography of the Transdanubian Range at the Jurassic-Cretaceous boundary, we decided to apply an other, independent geophysical method, paleomagnetism, to the same problem.

We collected paleomagnetic samples from two artificial exposures, from a 107 m long trench and from a large unearthed rock surface, both consisting of regular beds of Late Jurassic–Early Cretaceous limesones and of the suspected megabreccia horizon. We drilled several beds below and above the “megabreccia” horizon, respectively, and several points in suspected olistoliths. From every sampled bed and from every point three or more cores were taken so that the consistency of the paleomagnetic signal on site level could be checked. The samples were subjected to standard paleomagnetic laboratory processing (demagnetization was mostly carried out with the thermal method) and evaluation.

We found that magnetic parameters, like NRM intensities and susceptibilities are distinctly different for Upper Jurassic–Aptian beds of “normal” stratigraphic setting, on one hand, and for suspected olistoliths, on the other hand. The first group is characterized by NRM intensities of 2.1–12.5 mA/m and positive susceptibilities, while the same parameters for the second group are 0.04 to 0.43 mA/m and negative values, respectively. The paleomagnetic signal is highly consistent within nearly every site, no matter if they are regular beds or suspected olistoliths. However, there is a great difference in between-site consistency, which is extremely high for regular beds, but is non-existent for the sites of the “megabreccia” horizon. Thus, our results confirm that large blocks of late Triassic-Early Jurassic limestones were moved and re-deposited during the Late Jurassic-Early Cretaceous in the Transdanubian Range, therefore the geodynamic conditions must have been the same as in the Northern Calcareous Alps.

Acknowledgement

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Upper Mantle Mylonites: Evidence for Hydrated Mantle Wedge Beneath the Eastern Transylvanian Basin

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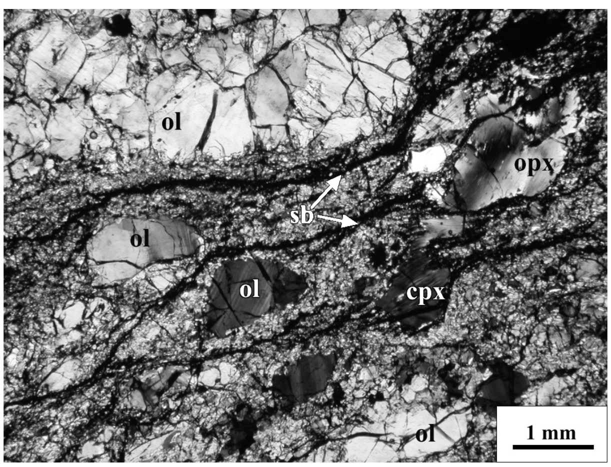
The severe effect of water on the deformation of the upper mantle (i.e. olvines) has only been recently recognized (e.g, Jung and Karato 2001, Katayama et al. 2004). The activation of slip systems (010)[001] ('b' slip) and (100)[001] ('c' slip) rather than the (010)[100] slip system ('a' slip) in H_2O-rich environments implies that hydrated upper mantle will have strongly distinctive texture from that deformed at H_2O deficient conditions (e.g., Jung and Karato 2001). Given that the development of the 'b' and 'c' type textures is extremely limited and at shallow mantle depth this texture only occurs within the mantle wedge in the presence of sufficient H_2O (>500 ppm) and stress (>100 MPa) (Kneller et al. 2005). Their recognition is a very strong evidence for identifying hydrated upper mantle developed in relation to subduction. Moreover, the occurrence of 'b' type textures in the hydrated mantle wedge explains arc-parallel fast shear waves observed in many supra subduction zone settings (e.g., Nakajima and Hasegawa 2004).

One of the best tools to directly study the deformation related textures in the upper mantle is from mantle xenoliths, hosted in deep-originated volcanic material, mostly alkali basalts. They are generally fresh and, in contrast to peridotite massifs and ophiolites, they have not experienced deformation during their emplacement to the crust. Moreover, they are relatively abundant in continental, back arc settings. The major disadvantage is that no direct tectonic implication can be derived from their textural analysis, because they cannot be directly fixed to any external reference frame. Instead, each xenolith must be fixed to its own reference frame (mostly foliation and lineation, if present), which is one of the major challenges, where texture analysis can go wrong.



■ **Fig. 1.** Anastomozing shear bands around strongly deformed porphyroclasts in a spinel lherzolite mylonite from the Eastern Transylvanian Basin (LGR03-01). Note extreme internal strain features in the porphyroclasts. ol – olivine; opx – orthopyroxene; cpx – clinopyroxene; sb – shear band. Cross-polarized light image.

In this study we represent 3 mantle xenoliths with mylonitic microstructures (Figure 1) from the Eastern Transylvanian Basin, which show the activation and dominance of (010)[001] and (100)[001] slip systems, indicating that they have been derived from the upper mantle which was deformed in an H_2O rich environment. This suggests that the subcontinental lithospheric mantle beneath the Eastern Transylvanian Basin represents a hydrated mantle wedge above the subducting European slab.

The mantle xenoliths are fertile, clinopyroxene rich peridotites. Clinopyroxenes show extremely primitive composition with high Al_2O_3 (4.5–5.3 wt%) and Na_2O (0.7–0.8 wt%) and low MgO (15.5–16.4 wt%) contents. Incompatible trace elements including Rb, Pb and U are depleted with respect to primitive mantle, whereas LREE is depleted with respect to MREE and HREE.

Our results strongly contradict recent models of delaminating continental crust (Knapp et al. 2005) and imply that the subduction beneath the Carpathian arc orogen, at least in part was 'normal' and involved the subduction of hydrated, most probably oceanic crust. We also confirm that the recent position of the slab is the consequence of rollback. Based on the presence of b-type olivine textures, we predicted an arc-parallel fast shear wave direction beneath the Eastern Transylvanian Basin. The observed chemical compositions disagrees with earlier results, which showed the enrichment of incompatible trace elements in the mantle wedge.

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Multicomponent Diffusion Modeling of Garnet: a Tool To Estimate Burial and Exhumation Rate of Metamorphic Complexes

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The occurrence of high-pressure rocks, mainly preservation and formation of concentration gradient within garnet or at its contact with other minerals provide evidence for the geodynamic processes of subduction and exhumation of crustal material within orogenic belts. With combination of the results from experimental data on diffusion coefficients and thermobarometric analyses from natural material, the measured concentration profiles in mineral can be used to estimate burial and exhumation rates of rock underwent high-pressure metamorphism. Diffusion modeling is based on definition of initial and boundary conditions, diffusion coefficients and the P-T-t history over which the diffusion takes place. During geological processes, diffusion in multicomponent minerals such as garnet results from simultaneous flow of more than two components. In garnet the diffusion of Mg, Fe, Mn and Ca are coupled to each other. The diffusion profiles measured in the microprobe can be modeled using the approach, proposed by Chakraborty and Ganguly (1991).

Starting with the initial profile shapes and calculated diffusion coefficients, a finite difference modeling scheme can be employed to calculate diffusion profiles. In this forward modeling approach, calculations are carried out until a good match is obtained with the observed, high resolution (points measured at 1 micron spacing) profile shapes near the interface of two minerals or two compositionally different garnets. It is here that a full multicomponent simulation is more useful than using effective binary diffusion coefficient – only for a rather limited range of time scales is possible to simultaneously reproduce the shapes of profiles for all elements. The time scale corresponding to the simulation that yields the best fit with the observed profiles is taken to be the one taken to traverse the P-T path being modeled. The method was used for burial and exhumation rate for two garnets of different composition, where the older garnet forms core and younger mantle. The simulation was don for a time scale estimated according to the P-T path. Diffusion coefficient was calculated for Mn, Mg and Fe, where Ca was treated as dependent component. The advantage of such simultaneous calculation is that it allows subtle details of variations in compositional profiles to be interpreted and considerably reduces the uncertainty in retrieved time scales that may be obtained from using only one profile. The modeling suggests that a minimum subduction / exhumation rate of ~4cm/a and heating / cooling rates on the order of 100–260 °C/Ma for a 60 °C subduction angel are required to preserve the observed compositional zoning overall while modifying the zoning at the interface between two garnets to the extent observed.

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Tertiary Tectonic Evolution of the Pannonian–Carpathian–Eastern Alpine Domain: a Personal View of from Pannonia in the Light of the Terminological Question of Tectonic Units

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Tectonic units are three-dimensional rock bodies with distinct physical boundaries and unique structural characters including temporal evolution. One of the main tasks of structural research is to define boundary surfaces or zones (structural elements), describe their geometry, characterize kinematic nature and determine temporal evolution. On the other hand, the presence of certain rock formations, their special facies or paleogeographic similarities are not distinctive features, although they can occasionally be useful to establish or better characterise some tectonic units.Tectonic units are evolving in time. Their boundaries can be shifted, their size can be increased or reduced. Frontal accretion, underplating, low-angle normal detachment, formation of strike-slip duplexes etc. may contribute to volume increase/decrease of a unit.

The definition of tectonic units (e.g. its boundaries and structural characters), and proper use of its name in temporal context is not merely a terminological problem. Although a perfect agreement on certain terms can hardly be achieved but on the other extremity, completely different usage of a tectonic unit reflect misunderstanding of the structural, and in consequence, the whole geological evolution. Tectonic units can be figured on geological maps, thus they are useful and necessary "tool" to disseminate results of "purely" structural geological research. On the other hand, results of modern structural research modify considerably geological maps and general knowledge. It is particularly true in Central Europe, where despite long research, and extensive geological knowledge, structural interpretation was not always integral part of the research and/or mapping. These are the reasons that I feel important to clearly define the units, discuss their nomenclature and structural characteristics.

I would like to embed the problem of tectonic units in a brief, simplified structural evolution of the area ain question, the western Carpathians, Pannonian area, and somewhat the Alps. As a sort of review, the presentation would be a selection of data what I feel important. Thus, it will be far from complete. The basic lines of the model were established by the clever reconstruction of Balla (1984), are coming from the first modern and straightforward structural synthesis of Tari (1994), from the genuine works of Frank Horváth, and from a great number of other colleagues, not listed below.

Looking from Pannonia, the Paleocene and early Eocene is a period of tectonic quiescence and terrestrial denudation, although geologists working in the Flysch Belt would argue for initial deformation of those areas. Basin subsidence started gradually from west to east from early Eocene to latest Eocene, a fact known from longtime (e.g. Báldi 1986) and re-summarized by Kázmér et al. (2003). Most authors agree on the compressional origin of the 'Central Carpathian' and Slovenian–Hungarian–South Slovakian Paleogene basins, although the obliquity of convergence, and the suggested forearc and retroarc position (Jablonsky et al. 1994, Tari et al. 1993) would not modify considerably the local structural geometry. Despite local problems, there is no doubt on the integrity of the Alpine-Carpathian orogen and the lack of a special 'Pannonian domain'.

One of the major structural elements of the entire Alpine–Carpathian–Pannonian–Dinaridic orogen is the Periadriatic Fault (PAF) which goes subsurface in north–eastern Slovenia. Following Kovács and Kázmér (1985), Balla (1984), Csontos et al. (1992, 1998), Kováč et al. (1994) the Hungarian (and hopefully all other "Carpathian") point of view is clear, that the continuation of this fault is within the "Mid-Hungarian Shear Zone", although the importance of any particular fault of this zone can be debated. Because the Periadriatic Fault and Mid-Hungarian Shear Zone is highly curved, it is improbable (although not completely excluded) that the entire fault system could still slip with a coherent kinematics. This shows that a unified, kinematically coherent PAF–MHZ system is a structural element of a certain time period, and was dismembered later (into PAF and MHZ) and then evolved separately. It is thus illogical to speak about *continuation* of the PAF into the MHZ in neotectonics; we can only speak occasional *connection* of the two fault zones.

Major issue is kinematics of the two fault zones, and the timing of the kinematics. Although the Western Alpine PAF seems to be better constrained in both respects (Schmid et al. 1996), the Pannonian area has still something to add. Intrusion of most of the tonalitic bodies along the PAF around ~30 Ma may indicate an important tectonic reorganisation, probably the establishment of dextral slip. This magmatism can be traced up to the Darnó Zone in NE Hungary (Benedek et al. 2004). On the other hand, this date may coincide with the major and dramatic subsidence in the whole Slovenian–Hungarian–South Slovakian Paleogene basin. Up to this date, the Paleogene basin was unique, but later was separated by the PAF–MHZ fault system.

Extrusion/escape tectonics is considered as a major event in the structural evolution. The displacing Alcapa unit is suggested to incorporate the eastern part of the Eastern Alps, the Western Carpathians and the northern and western Pannonian basin (Csontos et al. 1992). The process resulted in eastward motion of substratum of future Pannonian basin toward the stable European platform and ultimately resulted in shortening within and subduction below the Carpathian orogen.

Despite considerable research, physical boundaries of the Alcapa and the time span of its existence still merit a debate. The birth of Alcapa coincides with the onset displacement along its boundaries. To the west, extension of Penninic units of the Eastern Alps and boundary strike-slips are generally considered to be active from Early Miocene (Ratschbacher et al. 1989), from around 25 Ma. During the eastward motion of the extrusion, new nappes and slices of the former flysch basin(s) were accreted to the relatively rigid

Alcapa in its north-eastern periphery. In a strict sense, the consolidated flysch units became part of the Alcapa unit, because the major boundary structural element(s) were shifted from the front of one to the other (flysch) units.

Integration of paleomagnetic data may show differences between the Alpine and Carpatho-Pannonian segments of the Alcapa during the late Early Miocene (Márton 2001). While crustal extension ("orogenic collapse") and boundary strike-slip faults seem to persist in the early to mid-Miocene in the Eastern Alps, no notable extension existed before 19–18 Ma in Pannonia. The onset of upper crustal faulting coincides with the first rotation event, 30–50 ° counterclockwise rotation of the Western Carpathians – northern Pannonia between ~18–17 Ma. Because this rotation does not occur in the Eastern Alps, the rigid connection of western ('Alpine') and eastern Alcapa terminated (Márton and Fodor 2003). On the other hand, this rotation changed completely the southern boundary of the extruding Alcapa. While the Periadriatic Fault does not seem to be rotated, its continuation to Hungary, the Mid-Hungarian Zone sensu lato suffered the rotation. In consequence, the dextral slip along the Periadriatic Fault was transferred from the Mid-Hungarian to other fault zones in southern Pannonia or in the northernmost Dinarides (Fodor et al. 1998).

The Alcapa unit suffered considerable rearrangement at its south–eastern boundary. The Alcapa and the southern Tisza–Dacia units juxtaposed prior to or during the first major rifting phase (~18–14 Ma). From that moment, the Pannonian part of the Alcapa and the Tisza–Dacia units were moving eastward in a coordinated manner and their distinction as separate units is largely weakened. In my view, the only reason, which could still validate the usage of Alcapa and Tisza–Dacia units would be the verification of considerable strike-slip displacement between the two units, along the southern parts of the MHZ and its Eastern Carpathian continuation. In the lack of large displacement, I would say "Pannonian basin" and "Carpathians" or simply "Carpathian–Pannonian unit".

These considerations suggest that in the west the Alcapa was disintegrated around ~18 Ma into coherently moving, but distinct sub-units while increased by accreted new (flysch) units in the east. To solve the "terminological problem" we may have two solutions: (1) we can keep the term Alcapa from 25 Ma to 17 or 14 Ma, keeping in mind its continuous volume changes and accept at the same time that the rotations and rifting of the Pannonian basin (~18–14 Ma) is still part of the extrusion process, having affected a disintegrated unit; (2) we restrict the usage of Alcapa to post-25 to pre-18 Ma extrusion and speak about "rifting of the Pannonian–Western Carpathian–Eastern Alpine domain" after 18 Ma.

The disadvantage of the usage of the term Alcapa is more visible, when considering units during the late Miocene, the classical "post-rift phase", ca 11–6 Ma (Horváth 1993). Data available for me suggest that during the late Miocene subduction and frontal accretion ceased all along the Carpathians (expect probably the SE corner, Maţenco and Bertotti 2000), meaning a solid and fixed connection of the Carpathians and the European foreland. In this scenario, the definition of any Carpathian–Pannonian units would need a much better resolution of displacement rates along possible boundary structures than we have actually – there is only one tectonic unit merged with Europe(?).

The maintenance of the name Alcapa is more frustrating for the neotectonic phase (ca. 6–0 Ma). GPS-derived velocities, and structural data would indicate dextral motion along the Slovenian PAL, (Weber et al. 2005) a continuous (or reactivated?) eastward motion of the easternmost Alps, westernmost Pannonia and westernmost Carpathians, while the "north–eastern corner" (formerly part of the Miocene Alcapa) seems to be fixed to the European plate (Grenerczy and Kenyeres 2005). The projection back in time of the GPS data would result an "intra-Alcapa" accommodation zone with ca. 1 mm/y shortening. Despite similarities of certain structural elements, the boundaries of the Miocene Alcapa unit and neotectonic 'alter ego' are not the same, the southern PAF–MHZ was disintegrated and are moving with markedly different kinematics. This is the reason we used temporal names for three major neotectonic blocks of the Carpathian–Pannonian domain (Fodor et al. 2005) – hoping good suggestions and also better understanding of neotectonic movements and units.

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Kinematic and Rheological Model of Exhumation of High Pressure Granulites in the Variscan Orogenic Root: Example of the Blanský Les Granulite, Bohemian Massif, Czech Republic

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The structural pattern of the south Bohemian Moldanubian domain in the broad surroundings of Blanský les, Prachatice and Křišťanov granulite massifs is dominated by pervasive moderately NW dipping amphibolite facies foliation. This fabric parallels the trend of the Brunian and Saxothuringian margins and its attitude can be correlated to the flat lying amphibolite facies foliation dominating the eastern Moldanubian, ascribed by Schulmann et al. (2005) to a flow of Moldanubian rocks over the Brunia margin. In the vicinity of the granulite massifs this fabric is being disturbed to form irregular patterns passively adjusting a fold-like shape of rheologically stronger granulite massifs. Inside these rigid bodies, older Variscan fabrics have been well preserved, documenting two-stage exhumation history of the felsic granulites. Based on the kinematic model of granulite deformation history we use these fabrics to unravel the far-field stress changes in space and time during the Variscan collision.

The relict granulite facies fabrics allow for a reconstruction of the early exhumation mechanism in form of a vertical ascent channel because the subsequent cooling history froze these fabrics enabling us to observe them continuously on a km-scale. Analysis of the corresponding microstructure reveals very high plastic strain of quartz while the prevailing fine-grained feldspar dominated matrix shows only slight plastic deformation. Together with the presence of syndeformational intergranular partial melt this implies highly ductile behavior attaining characteristics of viscous flow. This offers an efficient way to transport the relatively small portions of lower crust rapidly upwards through the orogenic root.

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Structure and Petrology of the Western Part of the Meliata Unit, West Carpathians

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The Meliata Unit with bluschist facies metamorphic rocks, exposed in the western part of the Gemericum, occurs in several localities around the Štítnik and Nižná Slaná area. It forms tectonic slices overlying the early Paleozoic greenschist facies sedimentary and volcanic rocks with late Paleozoic metaconglomerates and also occurs beneath the unmetamorphosed or very low-grade metamorphosed Silica (Turna) Nappe formed mostly by limestones and dolomites. The Meliata and related rocks can be subdivided into two even three (upper, middle and lower) tectonic sheets. The most common rocks of the upper sheet are marbles with lenses of metabasite, different varieties of phyllites and rarely micaschists. Marble is pure calcitic, but at the contact with metabasites or phyllites it may contain amphiboles (glaucophane or actinolite), epidote and micas. The blueschist facies minerals in metabasites are glaucophane, epidote, albite, titanite and locally also garnet. Fine-grained, black phyllites, exposed on the basis of marbles are characterized by the presence of long (up to 1 cm) glaucophane crystals that cross cut the foliation and they are mostly replaced by chlorite and Fe-oxide (limonite). In some cases, brown colored mixed layered silicate of chlorite with mica is also present. Some of these glucophane-bearing phyllites may contain also chloritoid. The micaschists of the upper sheet occur between Nižná Slaná and Hankova and they are characterized by the presence of large (up to 5 cm) columnar crystals that have random orientation. The spectacular large crystals are formed either by pseudomorphs of chlorite with relicts of hornblende or by glaucophane. The rock with hornblende has relatively high amount of epidote (10 vol%), which mostly form inclusions in white mica. The studied micaschists occur adjacent to mafic blueschists, but their contact is not exposed. Minerals present in the blueschists are glaucophane, epidote, albite and chlorite. Some phyllites additionally contain quartz, phengite and rarely also paragonite and garnet. The middle sheet rocks are phyllites with chloritoid, although without glaucophane. They are exposed only locally, but lithologically correspond well to those in the eastern part of the Meliata Unit and are called the lower complex underlying marbles and blueschists. Chloritoid forms porphyroblasts in the fine-grained matrix and crosses cut the foliation. The lower sheet rocks are represented by the Permian conglomerates which are strongly deformed and show similar microstructural features as that in the middle and upper sheets. Compared to the upper and middle sheet rocks these rocks indicate very low-grade metamorphic conditions. Four deformation events were recognized in the studied area. The oldest deformation event (D1) was identified only locally in metabasites and it is characterized by the development of moderately east dipping metamorphic foliation bearing HP mineral assemblage. This foliation is transposed into a new metamorphic foliation showing retrogression features during the second deformation event (D2). The second deformation fabric is sub-parallel to the S1 and dips to the SE at medium angles bearing generally WNW-ESE trending lineation defined by shape preferred orientation of micas. The third deformation event (D3) is characterized by the development of large scale folds as well as small scale kink bands with steep N-S trending axial planes and subhorizontal NNE-SSW trending axes. This event shows westward shear senses and it is probably related to the late stage buttressing following the exhumation of the Meliata Unit. The forth deformation event (D4) related to Cretaceous deformation described also in adjacent Units is represented by the development of folds and kink bands with steep WSW-ENE trending axial planes and axes dipping at moderate angles to the NW. The folds and kink bands of the last two deformation events affect highly anisotropic phyllites and metaconglomerates preferentially.

The Zázrivá Fault – Paleostress History and Kinematics (Pieniny Klippen Belt, North Slovakia)

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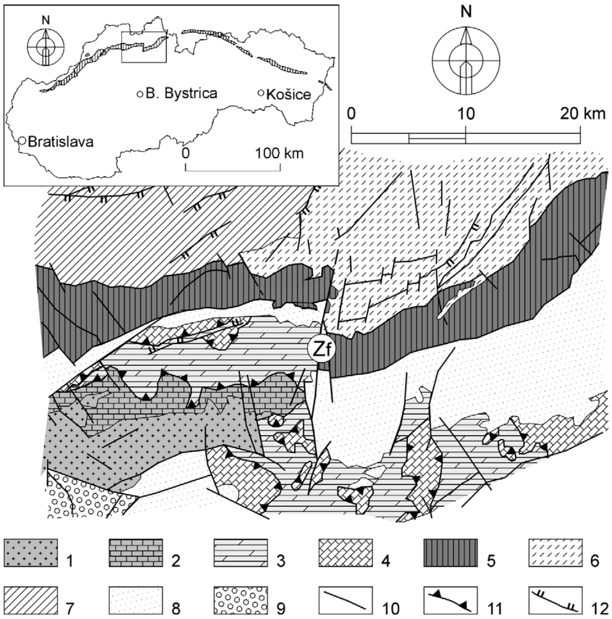
The Zázrivá fault is N-S trending discontinuity distinctively affecting the northern part of the Pieniny Klippen Belt in the Orava area of the Western Carpathians (Fig. 1). It is a part of the first order fault zone cutting whole structure of the Western Carpathians, even Panonian region traditionally called Zázrivá-Budapest fault. The Zazrivá fault represents the northernmost structural expression of the Central Slovak fault system (Kováč and Hók 1993, Nemčok 1994, Nemčok and Nemčok 1998).

The structural records of the Zázrivá fault activity were observed and analyzed at available outcrops in the vicinity of the Zázrivá and Istebné village. The structures were observed along the damage zone of the Zázrivá fault in the Outer Carpathians,



■ **Fig. 1.** Location of the Zázrivá fault and idealised tectonic sketch (according to Biely et al. 1996, modified). 1 – Crystalline basement of the Tatric Unit; 2 – Tatric sedimentary cover; 3 – Fatric Unit (Krížna nappe); 4 – Hronic Unit (Choč nappe); 5 – Pieniny Klippen Belt; 6 – Krynica Unit; 7 – Bystrica Unit; 8 – Central Carpathian Paleogene Basin sediments; 9 – Neogene sedimentary rocks; 10 – Faults; 11 – First order overthrust lines; 12 – Second order overthrust lines; Zf – Zázrivá fault.

Pieniny Klippen Belt and Central Carpathians units. Attention was focussed to the brittle deformations – slickensides, as well as tensional structures (veins, tension gashes) to reconstruct paleostress evolution within the area of interest.

Four paleostress fault-related events have been reconstructed. Thanks to the various age of fault-bearing rocks (from the Triassic up to the Middle Eocene) the approximate age and superposition of computed paleostress events were possible to establish there.

As the oldest one seems to be NW-SE compressional event followed by N-S, NE-SW compressional events and WNW-ESE, NNW-SSE tensional event. WNW-ESE tensional event was probably complementary one to the NE-SW compressional event. All detected tectonic events characterized by different paleostress fields and structural records are regarded to be the post-Eocene. They are very probably products of the young Tertiary tectonic activity. In recognized stress fields, the kinematic activity of the Zázrivá fault has been verified and model of Zázrivá fault tectonic evolution has been submitted.

The oldest period of tectonic evolution recorded by small-scale structures was NW-SE compression (Oligocene-Early Miocene). The dextral shearing along the E-W trending segment of the Pieniny Klippen Belt in the area of interest is related to this tectonic event.

N-S compression (Early-Middle Miocene), with frequent structural records was responsible for backthrusting tectonics affecting the whole structure of the contact zone in between Outer and Central Western Carpathians (Marko et al. 2005). Nevertheless, N-S tensional discontinuity – embryonic Zázrivá fault was founded as well.

In the third period (NE-SW compression, NW-SE tension respectively, Middle/Late Miocene), the Zázrivá fault was reactivated as dextral strike-slip, but with important tensional component of stress causing downthrow of the Eastern block.

This scenario continued also in the period of NNW-SSE tension (Late Miocene-?), when the dextral separation along the Zázrivá fault was enhanced and the subsidence of the eastern block continued. It led to the erosion of relatively uplifted units in the western block, and preservation of these units in downthrown eastern block. It resulted in great geological differences in the western and eastern block of the Zázrivá fault.

Acknowledgements

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Geodynamic Evolution of the Subsilesian Realm

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Subsilesian nappe in Poland and Czech Republic

Present-day Subsilesian Nappe extends between Moravia in Czech Republic and Eastern Part of the Polish Outer Carpathians. Westward it extends into Ždanice zone in Moravia, across Lanckorona-Żegocina zone in Beskid Mts., the eastward extension is unknown. It underlies tectonically the Silesian Nappe. In the western sector of the West Carpathians both nappes are thrust over the Miocene molasse of Carpathian Foredeep and in the eastern sector they are thrust over the Skole Nappe.

The Subsilesian unit has also been drilled in many boreholes between Bielsko, Cieszyn and Ustroń, and in the adjacent apart of Moravia beneath the Silesian Nappe (Picha et al. 2005, Ślączka et al. 2005 and references therein). This unit also appears in the Żywiec window (Geroch and Gradziński 1955). The intensely folded, and arranged in scales mostly lying in an N-S direction and steeply dipping to the west Subsilesian Unit rocks form the diapiric anticlinal uplift (Książkiewicz 1977). Eastwards, several tectonic windows under Silesian and Skole nappes occur. The Subsilesian Unit rocks are exposed in these windows. In the frontal part of the Silesian Nappe, north of the town Krosno the Subsilesian Nappe is exposed in the Węglówka tectonic half window. Deep wells connected with the Węglówka oil field show that the tectonic window is built of refolded thrust-faulted anticline. The Subsilesian Nappe is steeply overthrust onto the Skole Nappe. Further to the east the Subsilesian Nappe forms once more a narrow zone in front of the Silesian Nappe. Near the town of Ustrzyki Dolne the Subsilesian Nappe disappears from the surface, and the frontal part of the Silesian Nappe becomes a thrust-faulted fold and eventually joins with the Skole Nappe. There is also a possibility that tectonic prolongation of Subsilesian Nappe is the Rosluch scale in Ukraine.

In the western part of the Outer Carpathians near of Andrychów, along the Silesian Nappe there are several huge blocks built mainly by Jurassic limestones. They were regarded as tectonic klippen that were sheared off during the movements of the Silesian nappe (Książkiewicz 1977), however a new data suggest that they are olistolites in uppermost part of the Krosno beds of the Subsilesian nappe (Ślączka et al. 2005). It is possible that Andrychów and Subsilesian Upper Cretaceous and Paleogene rocks were deposited within the same ridge area. The Andrychów facies represent the central, partially emerged part of the ridge, while the Subsilesian much broader slope area.

Geodynamic evolution and sedimentation

The Silesian basin and Subsilesian sedimentary area have been connected during their early sedimentation period. In the Ždanice Unit shallow-water carbonate facies represent the oldest Jurassic-lower Cretaceous deposits (Picha et al. 2005 and references therein). They equivalent in Northern Moravia are known as Baška facies, which now is regarded as belonging to the Silesian Nappe. Similar rocks are also known from the North European Platform. They were drilled under the Carpathian Overthrust in the area south of Rzeszów. The more basinal slope facies are represented by Maiolica-type Upper Jurassic-Lower Cretaceous cherty limestones known from Targanice and Roczyny in the Andrychów area. It looks like during the Late Jurassic large carbonate platform existed uplifted part of European plate and Tethyan Penninic realm. The Maiolica deposits represent the deepest part of this platform. They were rimmed by shallower carbonate facies with carbonate buildups. As result of the fragmentation of European platform the Outer Carpathian rift had developed with the beginning of the Uppermost Jurassic-Lower Cretaceous calcareous flysch sedimentation. This Proto-Silesian basin was formed during the synrift process with a strong strike-slip component (Golonka et al. 2005). It included the oldest deposits of the future Subsilesian realm as well as the future Skole basin.

During Cretaceous time several ridges have been uplifted as an effect of the orogenic process (Golonka et al. 2005). This process started in Albian and was concluded in Paleocene. The Subsilesian Ridge originated between Silesian and Skole basin. Westward it extends into the shelf and slope of the European Platform. During the orogenic, mainly transpressional process the inversion of the proto-Silesian basin happened. The deepest part of the old basin became part of the newly formed ridge. New carbonate platform developed within the ridge and its slope area. The shallow-water Paleocene organogenic limestones are known from the Andrychów area (Książkiewicz 1951, Olszewska and Wieczorek 2001, Gasiński 1998). The Andrychów facies represent the central, partially emerged part of the ridge, while the Subsilesian facies much broader slope areas. These facies were deposited also in the deeper part of Silesian and Skole basins. The Subsilesian Late Cretaceous-Paleogene realm in Poland includes different deposits located between the central axes of the surrounding basins.

Variegated shales of the Cenomanian-Turonian age which pass upwards into a thick complex (about 700 meters) of red and green marls (Węglówka-type marls) which are Senonian to Mid Eocene (Ślączka et al 2005). During the Late Senonian grey marls (Frydek-type marls) often with exotic rocks (Książkiewicz 1977) developed at the same time with Węglówka-type marls in this area. Frydek-type marls represented submarine slumps from boundary of the shelf and bathial zones (Morgiel and Olszewska 1981). In the western part of the Subsilesian Unit sandy and conglomeratic complexes of the Upper Senonian and/or Paleocene were deposited. At the end of the Cretaceous an intensive activity of density currents started in the Subsilesian sedimentation zone, as a result of that a fine-grained sedimentation was interrupted by a sandy-shaly deposition. The Rybie sandstones, the Szydłowiec sandstones, the Gorzeń beds and the Czerwin sandstones are the effect of this sedimentation. At the end of Paleocene sedimentary conditions changes and are deposited muddy sediments called as green or variegated shales, which pass in to in marly shales with the Middle Eocene. The marly complex passes upwards into Globigerina Marls representing uppermost part of the Eocene.

The movement of Inner Carpathian terranes during Eocene-Oligocene led to the development of Outer Carpathian accretionary prism. This prism overrode the ridges, including the Sub-Silesian ridge. The ridge basement rocks and part of its depositional cover from olistostroms and exotic pebbles within Menilitic-Krosno flysch. The Oligocene begins in the Sub-Silesian realm with brown, bituminous shales (Menilite Beds) which grades upward into a complex of thick and medium bedded, calcareous sandstones and marly shales (Krosno Beds).

Finally, during the Miocene time the Outer Carpathian nappes were detached from the basement and thrust northward onto North European platform with its Miocene cover. The Subsilesian realm forms the present-day Subsilesian Nappe. The Outer Carpathian allochtonous rocks have been Overthrust onto the platform for a distance of 50 to more than 100 km.

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Phanerozoic Palaeogeography of Southeast Asia

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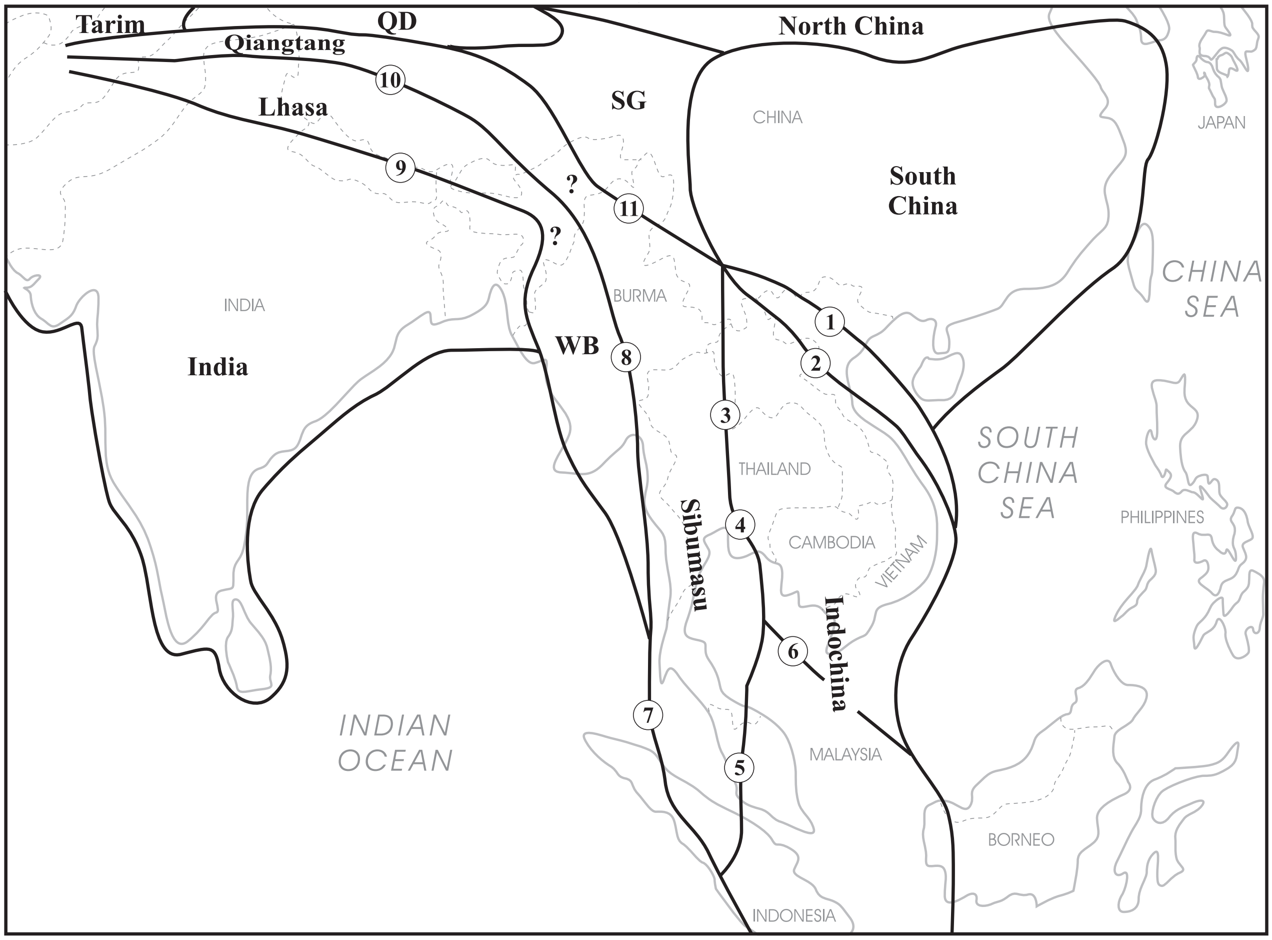
Methodology

Thirty two time interval maps have been presented, which depict the global plate tectonic configuration as well as palaeogeography and lithofacies for South-East Asia region (Fig. 1) from Cambrian to Neogene. The presented maps were primarily generated as Intergraph™ design files and CorelDraw™ files using computer software and databases. The plate tectonic model used to create palaeocontinental base maps is based on Plates and PALEOMAP tectonic reconstruction programs. These programs take tectonic features in the form of digitised data files and assemble those features in accordance with user specified rotation criteria. The detail information about the database, including the palaeopoles used can be found in the Plates homepage:

- http://www.ig.utexas.edu/research/projects/plates/plates.htm.

Plates maintains an up-to-date oceanic magnetic and tectonic database, continuously adding new palaeomagnetic, hot spot, geological, and geophysical data to extend the span and accuracy of global plate reconstructions. Plates' reconstructions are built around a comprehensive database of finite-difference poles of rotation, derived both from extensive plate motion research at UTIG, using the Plates interactive plate modeling software, and from published studies. Updated plate motion models are in turn



■ **Fig. 1.** Main plates, terranes and of Southeast Asia. Partially from Mecalfe, (1998). WB – Weast Burma, SG – Songpan Ganzi accretionary complex. QD – Quidam terrane. Sutures and major strike-slip faults: 1 – Red River zone, 2 – Song Ma, 3 – Nan-Uttaradit, 4 – Sra Kaeo, 5 – Raub Bentong, Three Pagodas, 7 – Woyla, 8 - Shan boundary, 9 – Indus Yarlung Zangbo, 10 – Banggong, 11 – Ailaoshan.

margin connected to South China Sea by extended continental crust (Fan 2000). The Indochina proper is separated from the plate southern part by the Three Pagodas Fault of NE-SW orientation.

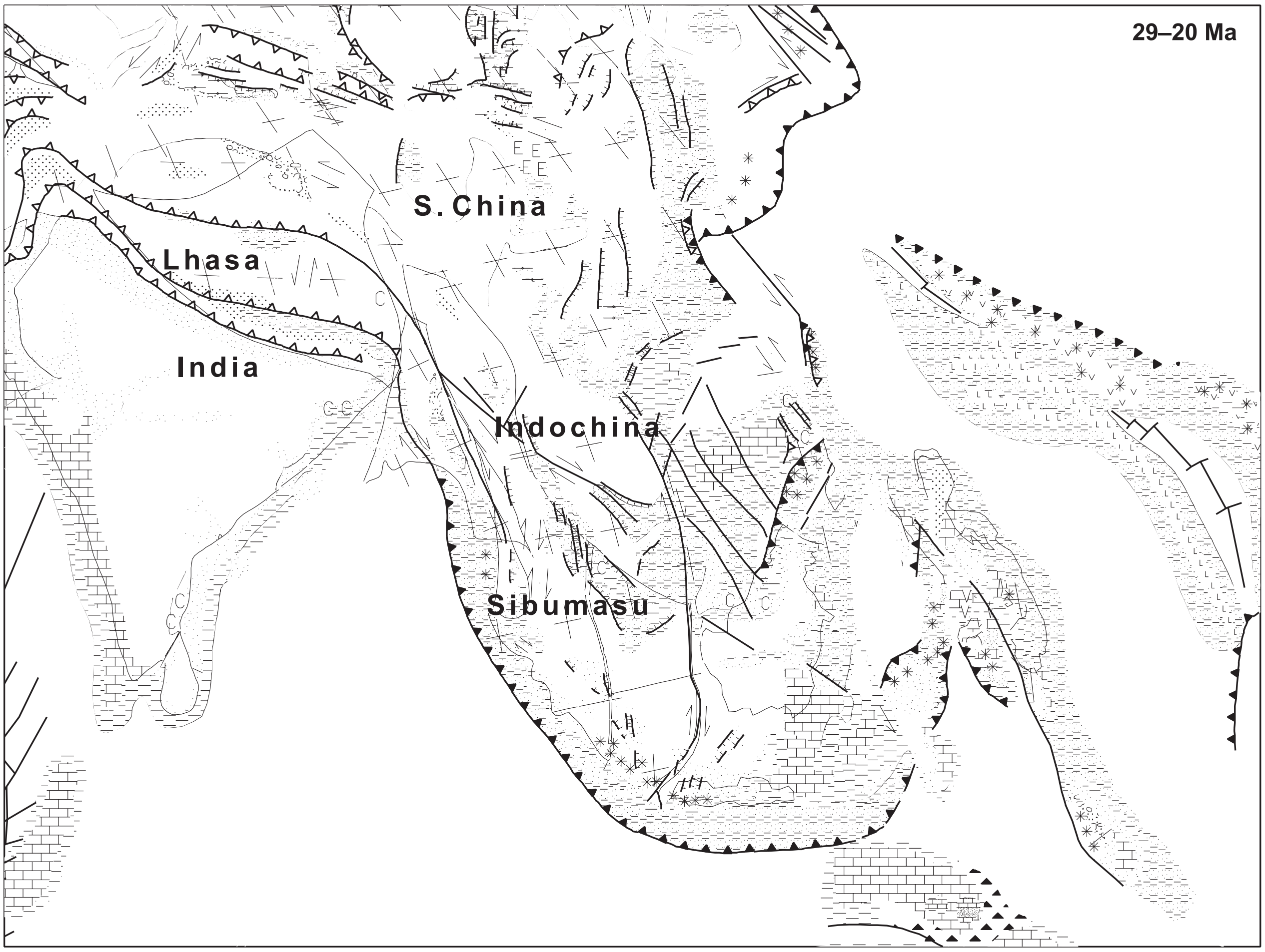
The South China plate includes southern part of China and northeastern fragment of Vietnam. It is separated from North China by Quingling-Dabie suture, from Indochina by Song Ma suture, from Sibumasu terrane by Ailaoshan suture, from Songpan-Ganzi accretionary complex by Longmenshan suture (Fig. 1) (Nie *et al.* 1990, Metcalfe 1998). The southeastern margin of South China is a passive margin connected to South China Sea by extended continental crust. To the east, the South China plate is bordered by the Taiwan foldbelt and the Okinawa trough passive margin. The Sibumasu terrane is bordered to the east by Raub-Bentong, Sra Kaeo and Nan-Uttaradit sutures, to the northeast by the Ailaoshan suture, (Fig. 1) (Nie *et al.* 1990, Metcalfe 1998, 2000), to the west its northern part is separated from the West Burma by the Shan suture. The Qiantang terrane is bounded on the south by the Bangong-Nuijang suture and to the north by the Lungmu-Yushu Zone. The Lhasa plate is bounded on the North by Lungmu-Yushu Zone and on the south by Indus Yarlung-Zangbo suture (Nie *et al.* 1990, Metcalfe 2002).

Outline of Geodynamic Evolution

The major South-East Asia plates originated during the Proterozoic as parts of Gondwana. They were detached during Palaeozoic time and drifted northward. The carbonate platforms were developed during the Devonian – Late Palaeozoic. The carbonate deposits were karstified later giving beautiful landscapes. The Palaeozoic history of detachment and collision is quite speculative. The equivalent of Caledonian orogeny followed by the formation of the Palaeotethys Ocean is quite possible. Climate record indicates major differences between Sibumasu, Indochina and South China during the Late Palaeozoic.

During Triassic time, as a result of the Indosinian orogeny and closure of the Palaeotethys Ocean, the South-East Asian plates joined the Asian continent. Strong tectonic deformations, metamorphism and magmatic intrusion and extrusion events were associated with the orogeny. The territory of South China was uplifted with mountains and intermountain basins with red beds, coals and volcanics. In the Indochina plate, during Jurassic and Cretaceous, terrestrial clastic sedimentation prevailed with red beds.

The onset of the collision of India with Asia occurred near the Palaeocene-Eocene boundary (e.g. Gaetani and Garzanti 1991). Pull-apart basins and strike-slip faulting occurred in China. Indo-



■ **Fig. 2.** Plate tectonic and lithofacies map of Southeast Asia during Late Tejas I.–Chattian–Aquitanian – 29–20 Ma.

china perhaps initiated the movement southeastwards, with respect to South China along the left-lateral Red River Fault (Lee and Lawver 1994, Golonka 2002), the main stage of this movement occurred, however, at a later time (fig. 2). The Red River Fault Zone in Yunnan, China and North Vietnam, up to 20 km wide, is one of the main strike-slip fault zones in SE Asia that separates the South China and Indochina blocks. The fault zone activity occurred in two phases: sinistral ductile shear active in 27–16 Ma, followed by exhumation and uplift from a depth of 20–25 km, and dextral, predominantly brittle shear active in Plio-Quaternary times. The Late Miocene change of the sense of motion is commonly related to the history of collision between India and Eurasia (e.g. Tapponnier et al. 1990). The opening of the South –East Asia basinal zones occurred as a result of complex tectonics during Palaeogene-Neogene time.

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Lithospheric Structure of the Carpathian Mountains, Pannonian basin and Eastern Alps Based on Seismic Data

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The network of seismic refraction profiles in the Central Europe covered now the area from the East European craton (EEC), along and across the Trans-European suture zone (TESZ) region in Poland to the Bohemian massif, and through the Carpathians and Eastern Alps to the Pannonian basin. The resulting seismic velocity models show strong variations in crustal and lower lithospheric structure (*Brueckl et al.,* submitted; *Grad et al.* 2006; *Środa et al.* 2006). In the Pannonian basin crustal structure is relatively simple. Beneath the sedimentary layer, two almost homogeneous crustal layers are observed with velocities 6.1–6.2 km/s in the depth interval 5–18 km, and 6.3–6.6 km/s in the lower crust. In this area, the Moho lies at depths of only 24–25 km.

In the Eastern Alps crustal thickness varies between 40 and 50 km. The most complicated structure is observed in the transition from the Pannonian basin to the EEC, which includes the Carpathians and the TESZ. In this area, the sedimentary cover with low velocities (Vp<5.5 km/s) reaches a depth of ~20 km, and the Moho deepens to ~50 km. Further to the northeast, the crustal structure of the EEC is typical for cratonic areas, with a thin sedimentary cover and a three-layer crystalline crust with velocities of 6.0–6.4 km/s, 6.5–6.7 km/s and 6.7–7.0 km/s, respectively. The depth of the Moho for the EEC varies between 42 and 48 km. Beneath the Moho lower lithospheric reflectors were found at depths of ~15 km beneath the Moho and at several deeper intervals.

The longest profile CEL05 (1420 km) shows clear crustal thickening from the Pannonian basin to the TESZ region, together with the configuration of the lower lithospheric reflectors. This result suggests northward subduction of mantle underlying Carpathian-Pannonian plate toward the north under the European plate. *Książkiewicz* (1977) postulated that subduction of the Pannonian lithosphere under the East European craton occurred during the Jurassic–Early (Lower) Cretaceous. In their paleogeographic reconstruction of the circum-Carpathian area *Golonka et al.* (2003) also proposed that north-northwestward subduction of the Meliata-Halstatt Ocean crust was completed by the end of the Jurassic, ~140 Ma ago and that the location of this closure corresponds to the Mid-Hungarian line. The northward subduction however conflicts with strong geological evidence for southward subduction, and we present three tectonic models for the CEL05 area, that are to not totaly mutually exclusive, to explain the lithospheric structure of the area: (1) northward "old" subduction of the Pannonian lithosphere under the East European craton in the Jurassic - Lower Cretaceous, (2) a collisional zone containing a "crocodile" structure where Carpatho-Pannonian upper crust is obducting over the crystalline crust of the EEC and the Carpathian-Pannonian mantle lithosphere is underthrusting cratonic lower crust, and (3) lithosphere thinning due to the effects of Neogene extension and heating with the slab associated with "young" subduction southward in the Miocene having been either detached and/or rolled-back to the east. In the last case, the northwestward dipping in the lithosphere can be interpreted as being due to isotherms that could represent the lithosphere/asthenosphere boundary in the Pannonian region.

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Tectonic Setting of Sokolov Basin in Relation to Prediction of Thermal Water Discharge Zones

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The study area – the western part of the Eger Rift (the Sokolov basin) – belongs to the European Cainozoic Rift System (Kopecký 1978, Sengör 1995, Prodehl et al. 1995, Adamovič and Coubal 1999, Dèzes et al. 2004). This system of graben structures and intraplate volcanic fields spreads over a distance of some 1000 km, including the French Massif Central, the Upper Rhine Graben, the Eifel, the North Hessian Depression, the Vogelsberg, the Eger Rift and the Elbe Zone. Graben structures evolved on top of uplifted basement blocks (Variscan massifs); Tertiary and Quaternary volcanism is mainly concentrated on the flanks of these graben structures along boundary faults or on the adjacent uplifted blocks. Dominantly (ultra-) alkaline, but also more evolved, magmas were erupted. The main rifting phase with incipient graben formation and voluminous intraplate alkaline volcanism lasted from about 42 Ma to 9 Ma. A detailed overview of the Cainozoic volcanic activity in the western part of the Bohemian Massif is given by Ulrych et al. (2003). The most recent expressions of magmatic activities within the European Cainozoic Rift System are the CO_2 degassing fields. The isotope (He, C, and N) composition of CO_2-rich gas emanations of mineral springs and mofettes from the western Eger Rift (Weinlich et al. 1999, 2003) gives evidence for the ascent of gases from fluid reservoirs in the European subcontinental mantle.

The Sokolov Basin is also a place of collision between long-term coal mining and spa Karlovy Vary protection. Both the technology and the method of coal mining in Sokolov Basin are strongly limited due to the existence of Karlovy Vary thermal springs resources that have priority importance. Considering that from the structural and geological point of view, the geohydrodynamic systems of these resources form one single structure, extending as far as the Sokolov basin brown coal deposits, the possibility of natural barrier layers being negatively impacted by human activity (i.e. mining technology in existing protection zones – especially in areas of hydrogeologically active faults and joint systems) is extremely strong (Trčková et al. 2000). This has also become evident recently in the case of uncontrolled opening of some old exit paths (old drills, old mining works, etc.) that had to be solved as emergency or warning states in relation to Karlovy Vary thermal springs. This problem may only be solved by conducting complex structural-tectonic analysis, based on parallel interpretation of geophysical methods followed by regional hydrogeological prospecting.

The Sokolov Basin proper is a bilaterally tectonically limited, transversally asymmetric depression, extending in WSW-ENE direction. In NW it is limited by the Krušné Hory Fault and also characterised by a system of minor parallel faults (especially the Lipnice, Grasset, Sokolov and Nové Sedlo Faults), forming a significant tectonic zone of lithospheric range (Ziegler 1990). According to Adamovič and Coubal (1999), most of this system's accompanying faults are younger than the main stage of the Ohře Rift volcanic and sedimentary development.

Another significant fault system of the Ohře Rift are the faults running in NNW-SSE to NW-SE direction (in the Sokolov Basin these are faults following the Svatava, Chodov and Karlovy Vary faults). This system is especially intensively developed in the neighbouring Cheb Basin, forming part of Mariánské Lázně tectonic zone (e.g. Špičáková et al. 2000). The analysis of the Ohárecký Rift filling has shown that some of these faults had already been active synsedimentary. In the area of Sokolov basin the Chodov fault zone striking NW-SE belong to this tectonic system. It interfered with SW limit (contact zone) of Variscan Karlovy Vary granite pluton and was reactivated later in post-rift stage.

Emphasised in the most recent studies of the Ohře Rift tectono-sedimentary development has been the significance of W-E faults that had already been active in the course of sedimentation as extension faults (Rajchl and Uličný 2000, Špičáková et al. 2000).

From the above it follows that the structural development and the current tectonic architecture of the Sokolov Basin, similarly as to the entire Ohárecký Rift, have been affected by several basic systems of normal faults, some of which show a less significant strike-slip component. Typical is above all the en-echelon arrangement of faults, horsetail-like virgation of faults, curvature in directional course, but also their normal fault listric geometry (se Fig. 1). Specific deformation conditions occur above

■ **Fig. 1.** Low-angle listric normal fault with rollover fold striking W-E in the coal seam Antonín (Jiří Mine, Sokolov Basin).

all in places of their mutual interference. To be expected in these anomalous zones is the substitution of classic dislocation zones, accompanied by mylonitisation, by systems of brittle fracture, above all by development of tensile joints. Thus, places of potential outflow of underground or possibly thermal waters due to joint secondary porosity have formed.

In spite of safety measures aimed at stabilising the hydrogeologic situation of gas-cut thermal springs of the Karvoly Vary type in the coal basin basement, situations arise in the Družba and Jiří open-pit mines (Sokolov Basin) during mining which represent uncontrolled interference with the natural regime of these groundwater bases. Especially dangerous are the old mine workings of abandoned deep mines. An example of such situation is the inrush of thermal waters in the Jiří open-pit mine in January 2003, occurring in the wall of abandoned working base, in close vicinity of the Grasset Fault. The inrush was monitored as to its yield and hydrogeochemical composition, before the place of inrush was sealed by inner dump.

In spite of the relatively high resistance of thermal structures to outside impacts, the disturbance of natural steady regime with resulting impact on Karlovy Vary thermal springs may not be ruled out. In such event the restoration of conditions would be most difficult and lengthy, with far-reaching consequences for the spa town of Karlovy Vary.

From the viewpoint of geotechnical stability of the seam basement, the most critical situation is in the vicinity of the Grasset and Nové Sedlo faults in Jiří open pit mine (thermal waters uncovered overpressure reaches up to 0.6 MPa).

When studying the hydrogeologic relations between the groundwater basement of the Sokolov coal basin and the Karlovy Vary termal spring used for balneological purposes, a systematic and complex approach was adopted. Both hydrogeochemical methods and methods of classic drill explorations, analyses of geological and tectonic data, analyses of diverse geophysical methods, and hydrogeological methods were applied. The clarification of hydrogeological and hydraulic relations will help to more effectively protect the natural mineral resources of the spa town Karlovy Vary from potential impacts of brown coal mining.

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Two Types of “Augen Gneisses” in the Śnieżnik Metamorphic Unit, West Sudetes, Poland

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Śnieżnik Metamorphic Unit is the easternmost part of the Orlica-Śnieżnik Dome. It comprises metasedimentary-metavolcanogenic rocks of the Stronie-Młynowiec formation and widespread gneissic formation, traditionally subdivided into Śnieżnik and Gierałtów gneisses (see Don et al. 1990). The first ones are defined as coarse- to even-grained, rodding to flattened, augen (ortho)gneisses, while the second ones are diversified assemblage of rocks, within which one can identify fine-grained and compositionally banded (para)gneisses to embrechnites, biotite-rich and biotite-poor aplite-like homogenous gneisses, and coarse-grained gneisses with porphyroblasts. All the gneisses of the Orlica-Śnieżnik Dome underwent Variscan constriction and mylonitization, which resulted in imparted similar outlook of many types of gneissic rocks, especially in local shearing zones. Thus the grain-size and/or briefly noticed similar attitude as main criteria of classification of different types of gneisses are not sufficient. Detailed structural analysis, carried out in the Międzygórze area, revealed that some gneisses with augen structure, traditionally accounted to the augen Śnieżnik gneisses, are in fact migmatitic rocks of the Gierałtów type, different in orientation and complication of their fabric and, most of all, position and genesis of title “augens”.

In the Śnieżnik augen gneisses (metagranites), the augen structure results from the constriction and flattening of the original porphyritic granite. They posses distinct pure constriction fabric, with no foliation, but ductilely stretched monomineral rods of quartz, K-feldspar, and mica streaks ($L3^{*}$=160–210/15–30). Subsequent strain converted into flattening lead to dominant planar fabric (S3 = 90–160, 200–260/15–30), formed by separate layers of HT dynamically recrystallized quartz; K-feldspar; plagioclase and micas, accompanied by augen porphyroclasts, around which the foliation anastomoses. Porphyroclasts are composed of [1] white and/or pinkish (hip)automorphic megacrysts (up to 10 cm) of K-feldspars (microcline) with tails of pressure shadows and minor [2] quartz, characteristically elongated and flattened. Such augens are more or less flattened porphyroclasts derived from porphyrocrystals of the original granite, which produce typical pinch-and-swell structures in the XZ sections of the local strain ellipsoid. Singly/simply twinned K-feldspar megacrysts are often affected with recrystallization. Subgrains occurring in the core of megacryst, being slightly bigger and definitely more lobate (dynamically recrystallized) towards megacryst edges are getting smaller and polygonal in shape (static recrystallization). In more flattened varieties, K-feldspars are deformed into the long, foliation-parallel ribbons. If persisted, the parent grain occurs in the central part of the lamella, and smaller, dynamically recrystallized grains project out of it.

This simple fabric is only locally overprinted by S-C’ bands and small-scale E-W trending folds. Their reorientation to NW-SE and NE-SW due to syn- and post-mylonitic distortions was accompanied by weak biotite lineation and local crenulation. The character of all these phenomena reflects an amphibolite facies conditions. Further deformation brought about large-scale, brittle, E-vergent kinking folds (F4), which have developed during compressional regime, with “top-to-the-E” kinematics. These contrasting with mylonitic conditions, much more brittle deformation must have developed in the retrograde greenschist facies conditions.

The rocks so far being considered as the Śnieżnik metagranite due to the presence of big feldspar augen-like blast are in fact coarse-grained, biotite-rich, often pinkish (ortho)gneisses, with two HT mylonitic foliations and characteristic felsic porphyroblasts (“augens”) and leucosome nests. They are characterised by mylonitic, monomineral layering expressed by alternation of disrupted layers of quartz, K-feldspar, plagioclase and micas, with microstructural evidence of HT shear deformation and strong recrystallization overprint (presently subhorizontal foliation S1). Small-scale isoclinal to disharmonic W-vergent folds (F1/2) with “z” asymmetry and “top-to-the-W” kinematics was accompanied by migmatization, as leucocratic aggregates commonly occupy the triangle dilatant sites of the small folds, and readily nucleate parallel to the axial plane foliation (S2, emphasized additionally by shearing zones and S-C’ structures), overprinting the mylonitic fabric in a shape of augen-like blasts. These “augens” are polymineral (K-feldspar + quartz + plagioclase ± biotite) and in contrast with the porphyroclasts in the Śnieżnik metagranite, these are porhyroblasts with clearly metamorphic/migmatitic provenance. These fabric is overprinted by indistinct “s” asymmetrical (“top-to-the-E” shearing), the same scale folds (F1/2), which are also convoyed by migmatization, as the leucosome big augens/nests grow completely disorderly over the existing fabric. All the blastic augens (up to 10 cm) are similar in composition. Bigger, elongated K-feldspars are gathering in the core of the augen, surrounded successively by a thin plagioclase mantle, then by a quartz mantle tapering gradually off from the augen and contributing to the mylonitic banding, and closed finally by a mica layer. Polymineral and zonal composition of such augens are in evidence of their secondary/migmatitic origin. Pure quartz augens are surrounded only by mica assemblages.

Over the complex fabric the second mylonitic flattening (S3 = 100–160/20–45) and elongation lineation (L3=160–190/30) so characteristic for the whole region is printed. These structures do not deflect/anastomose around migmatitic blasts, evidencing that none significant (directional) deformation after the growth of blasts has occurred. More strongly flattened varieties, with clearly distinguished S3 mylonitic layering, have occasional relicts ("ghost structures") of the hinges of F1/2 folds in the form of bent mica and recrystallized aggregates of quartz.

The late and asymmetrical kinking folds F4 are the same as observed in the Śnieżnik gneisses. All described features are characteristic for the migmatitic suite of the Gierałtów type. These rocks do poses the same style, sequence and amount of deformational events as the Gierałtów gneisses (Dumicz 1989). The only Variscan obliteration and mylonitization of the earlier, complicated fabric (twice folded and migmatized), make some of the Gierałtów gneisses locally very similar to the original augen Śnieżnik metagranites.

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* Chronology of deformational events after Dumicz (1989).

Variscan Hydrothermal Veins in the Prague Synform (Barrandien Area)

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A few papers dealt with hydrothermal veins of various relative ages in Lower Palaeozoic sediments of the Prague basin (e.g. Suchy et al. 2002). Not many of them were concentrated on study of character of the fluid systems related to deformation stages, which affected the basin. In our present research stage we are focused on a definition of the Variscan (synorogenic) fluids and conditions of their origin and migration. First of all it is important to specify P-T conditions of Variscan deformations and their genetic connection with relevant fluid systems and thermal histories (Glasmacher 2002).

Field work took place at several localities of SW part of the basin (e.g. Homolák quarry, Srbsko). Preliminary research on Variscan veins revealed that calcites are dominant mineral phase in veins, which are mostly deformed and recrystalised, and calcite is fine and medium grained (in Devonian limestones). Veins are not very long, frequently have irregular or lenticular shape and they are arranged into en echelon arrays. Veins in Ordovician quartzite-sandstone are filled with quartz showing a fibrous structure. In drusy cavities a black organic matter occurred. Older hydrothermal veins are deformed and may be penetrated by younger veins.

So far two fluid systems have been found in fluid inclusions (FIS) of calcites, aqueous and liquid hydrocarbons. Sizes of FIS are around 5 micrometers. Due to the small size of the fluid inclusions there were difficult to observe eutectic temperatures (Te).

Homogenisation temperature (Th) primary and/or pseudosecondary aqueous FIS have values between 77–120 °C and generally have lower salinities (0,2–7,9 wt.% NaCl equiv.). Primary inclusions rich in hydrocarbons show Th between 41–85 °C.

Parent aqueous solutions have $\delta^{18}O$ values between +0,4 ‰ and +2,2 ‰ SMOW. When fluids were isotopically buffered by wall rocks than isotopic composition of fluids is more positive.

Preliminary results suggest accord with other authors (e.g. Suchy et al. 2002) that tectonically deformed veins were generated in condition of the oil window. Relationships between tectonic evolution and hydrothermal veins of the Prague synform will be subject of the further study.

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Origin of Felsic Migmatites by Ductile Shearing and Melt Infiltration

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The Gföhl migmatite-gneiss complex forms the largest anatectic unit of the Variscan orogenic root domain. The origin of this migmatitic unit was classically attributed to the anatexis and the different degree of migmatitization explained by the variable degree of partial melting.

A new petrogenetic model of an origin of this felsic migmatites is proposed on a basis of the microstructural and petrological study. The detailed observation reveals that the migmatites originated by melt infiltration and contemporaneous shearing of the banded orthogneiss in a crustal scale shear zone. They are marked by gradual transition from the high-grade solid state banded orthogneiss with distinctly separated monomineralic layers via the migmatitic gneiss, the gneissic migmatite characteristic by disappearance of monomineralic layering to sheeted foliation parallel bodies of the granitic gneiss with no relicts of gneissosity. The disintegration sequence is characterized by: (i) progressive destruction of well equilibrated banded microstructure of the high-grade orthogneiss by a crystallization of new interstitial phases (Kfs, Plg and Qtz) along the feldspar boundaries and by a resorption of relict feldspars and biotite, (ii) variations of modal proportion of felsic phases reflecting the increasing amount of melt in the originally mono-mineralic aggregates, (iii) systematic grain size decrease of all felsic phases together and crystal size distribution curves (CSD) indicating increase of the nucleation rate coupled with preferential removal of large grains for all felsic phases with the increasing melt proportion. This evolutionary trend is connected with a decrease in grain shape preferred orientation (SPO) of all felsic phases, an increase of regular grain boundary distribution (dominance of unlike boundaries) and a decrease of grain boundary preferred orientation (GBPO) of unlike boundaries.

Melt topology reveals well oriented melt seems and pools at low melt fraction consistent with dislocation to diffusion creep regimes. At high melt fractions the absence of preferred orientation of melt patches corresponds to the distributed granular flow associated with a breakdown of rigid skeleton close to rheological critical melt percentage (RCMP).

SEM images show plagioclase zoning displaying non-diffusive 2–10 μm more sodic rims (An0-10) around oligoclase cores (An10-30). The whole textural sequence displays continuous increase of Na content in plagioclase cores and rims, increase of XFe in biotite and garnet coupled with decreasing Ti content in biotite towards the granitic gneiss. The increasing amount of discrete albite rims and complete disintegration of original banded texture are compatible with melt infiltration into progressively deformed rock. Additionally, the petrological observations indicate that the melt infiltration is connected with crustal exhumation along retrograde pressure and temperature path.

The Role of Melt Infiltration in the Formation of Migmatitic Orthogneiss

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The Gföhl orthogneiss is a widespread lithology in the Moldanubian orogenic root domain of the Bohemian Massif. Its apparent textural variations were classically attributed to the variable degree of anatexis, however, a recent textural study interprets some of the variations to be due to different degrees of melt infiltration. In this contribution, we describe mineral and bulk rock chemical changes from the original banded orthogneiss (textural type I) to granite-looking gneiss (type IV) and determine equilibration P-T conditions. We characterize what sort of fluid is involved, calculate its composition and deduce how it interacts with the original rock.

The mineral assemblage in all the rock types is garnet-biotite-sillimanite-K-feldspar-plagioclase-quartz. As muscovite is absent, the infiltrating fluid must be a melt and not an aqueous fluid. Garnet in the studied sequence displays the following changes: $alm_{75=>94}py_{17=>0.8}grs_{2.5=>1.2}sps_{2=>11}$; X_{Fe} 0.80=>1, and biotite X_{Fe} increases (0.45 => 0.99). Plagioclase in the original aggregates has higher anorthite content ($An_{25=>5}$) than interstitial grains or films tracing the K-feldspar boundaries and plagioclase rims ($An_{18=>0}$). In an AFM diagram, the assemblage garnet-biotite-sillimanite is divariant, in the presence of quartz, K-feldspar and melt, a systematic increase in X_{Fe} of the phases indicating a decrease in equilibration temperature. The compositional isopleths in pseudosections also point to temperature decrease, corroborated by average PT calculations (800 => 650 °C/6kbar).

There is no direct evidence of the composition of involved melt, apart from the mineral compositions with which it equilibrated. A melt composition that is in equilibrium with plagioclase-K-feldspar-quartz-sillimanite-garnet-biotite in the NCKFMASH system can be calculated if X_{An} and the P-T conditions are fixed. With a melt composition derived in this way we calculated T-x pseudosections for a bulk composition line between K-feldspar (or plagioclase) and melt in order to understand bulk composition changes. When in a K-feldspar layer, plagioclase starts to crystallize above the temperature of muscovite stability only if melt: rock is 9:1. Similarly, in the plagioclase layer, the K-feldspar crystallizes if more then melt: rock is 8:1. Such a high melt proportion is reasonable only if the edges of grains of solid phases are considered to be in equilibrium with melt covering the grain boundaries. Thus, a small proportion of melt is present in the whole rock at one time. In order to change the whole rock composition in such a way, a large, but currently unidentified quantity of melt must have passed through the rocks along the grain boundaries.

The observed compositional changes in individual layers as well as bulk rock chemistry changes are driven by equilibration with large quantities of infiltrating haplogranitic melt of unknown source. Such a process of a large quantity of melt passing through rocks at grain scale without any important signs of segregation might be an important mechanism for melt transport in a migmatitic crust. As the Gföhl gneiss appears as hundreds of km^2 bodies, the process of penetrative melt flow through the orogenic crust would be a crucial process for crustal differentiation and also for crustal rheology during orogeny.

New Seismo-Tectonic Activity near Zakopane (Poland) – Events Recorded by Broad-Band Stations Operated by IPE

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During the end of the year 2004, the seismo-tectonic activity in the Polish part of the Vysoké Tatry region was newly detected. This activity continued also during the year 2005. The broad-band stations operated by the IPE (Institute of Physics of the Earth, Masaryk University, Brno – stations JAVC, KRUC, MORC and VRAC) in the eastern part of the Czech Republic registered 25 events with local magnitude ML from 1.1 to 4.6.

The new exhibitions of the seismo-tectonic activity have started by the strongest event (local magnitude ML = 4.6) on 30. 11. 2004, which was macroseismically observed. The historical macroseismic observations are known in this region. But during about ten years long continuous registration of broad-band stations operated by the IPE, before 30. 11. 2004, these stations had not recorded any tectonic event with epicentre situated in the Polish part of the Vysoké Tatry region. In contrast to situation before the strongest event, the significant seismo-tectonic activity was observed during first three days of the December 2004 (13 recorded events with local magnitude ML from 1.1 to 3.5). Less intensive activity continued up to the August 2005 (11 recorded events with local magnitude ML from 1.5 to 3.4).

Using other stations operated by Polish, Slovak, Czech and Hungarian seismological institutes, 13 events were reliably located by program LocSAT. In the case of other 12 events, only approximate locations were possible due to small number of reliable records by accessible broad-band stations. Epicentres are situated near Zakopane, on the northern margin of the Central Western Carpathians. This region represents the NE prolongation of the significant seismoactive zone passing from the Mur-Mürz fault system in the Eastern Alps through the southeastern part of the Vienna basin into Western Carpathians and continuing along the Pieniny Klippen Belt to NE.

Seismo-Tectonic Activity in the NE Part of the Bohemian Massif – New Records in the Period 2004–2005

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In the NE part of the Bohemian Massif, the weak recent seismo-tectonic activity occurs. Micro-earthquakes are concentrated in numerous epicentral areas. During the period 2004–2005, more than 250 tectonic events were detected (more than 60 events were located) in the NE part of the Bohemian Massif by the seismological stations operated by IPE (Institute of Physics of the Earth, Masaryk University Brno).

The most frequent occurrence of the micro-earthquakes is observed in the area northwards of Šternberk. In the period 15. 11. to 19. 11. 2005, new swarm-like sequence of the tectonic events was observed in the area NW of Šterberk. The stations of IPE detected 33 events (the local magnitude ML of strongest events was 1.4) belonging to this sequence. Seismo-tectonic activity occurred also in some other areas near Šternberk, the sequence of micro-earthquakes including the event with local magnitude ML = 2.2 was detected in December 2005 in the area eastwards of Šternberk.

Significant seismo-tectonic activity was observed in the Hronov region. In this region, the strongest micro-earthquake (local magnitude ML = 2.8) occurred on 25. 10. 2005. The swarm-like sequence of weak events (five registered events with local magnitudes ML varying from 1.0 to 2.2) observed on 10. 8. 2005 represents another occurrence of the relatively significant seismo-tectonic activity detected in the Hronov region during the year 2005.

Other relatively significant exhibitions of the seismo-tectonic activity occurred in the areas near Bruntál, Budišov n. Budišovkou, Opava, and Hranice na Moravě in the period 2004–2005. Epicentres of 15 micro-earthquakes recorded in the period 20. 8. to 29. 8. 2004 and located into area near Hranice na Moravě (the local magnitude ML of strongest events was 0.9) are situated in the Western Carpathian flysh nappes, close to the front of these nappes. But, in respect of the depth of hypocenters which exceeds 10 km, the seismic activity occurs in the units of Bohemian Massif forming the basement of the Western Carpathian nappes, which have thickness of only 1–2 km in this area (for instance Menčík et al. 1979).

Also seismo-tectonic activity newly observed in the Vizovice region (18 events detected in the period 14. 3.–6. 7. 2004, the local magnitude ML of strongest events was 1.3) is probably connected with the faulting in the units of the Bohemian Massif under the Western Carpathian flysh nappes. The thickness of the Western Carpathian nappes reaches about 6 km in this region (for instance Menčík et al. 1979). The calculated depths of the located hypocentres vary from 12 to 16. These depths correspond to the hypothesis, that the hypocentres are situated in the basement formed by the Bohemian Massif, close to the base of the Western Carpathian nappes. But the determination of the depth is less accurate in comparison with determination of other coordinates.

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Geodynamic Implications of Flattened Equigranular Textured Peridotites from the Central Part of the Carpathian–Pannonian Region

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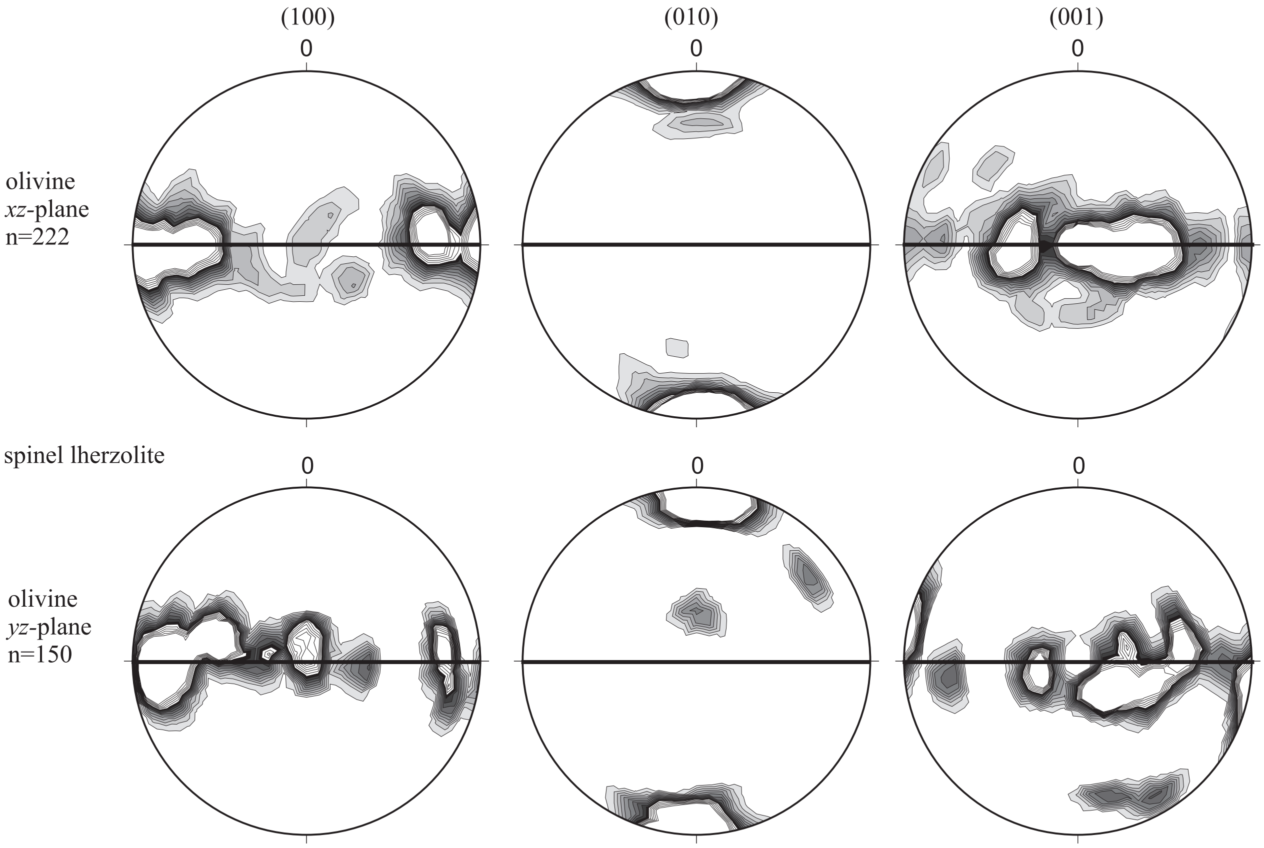
Peridotite xenoliths showing unusual tabular equigranular texture (addressed as flattened equigranular) were found in Neogene alkali basalts from the Bakony-Balaton Highland Volcanic Field (BBHVF), in the central part of the Carpathian-Pannonian Region. In this study we present a basic (major and trace element) geochemical, detailed fabric (polarized light microscope and computer tomography /CT/) and EBSD analysis of CPO of both olivine and orthopyroxene in three flattened equigranular textured peridotite xenoliths selected for this study.

Macroscopic foliation and mineral lineation in the studied upper mantle rocks are visible in hand specimens being defined by flattening and stretching of all mineral phases, respectively. On the CT images, foliation is also shown in 3D by the olivines. Regarding their textural type, the studied xenoliths are not common in the BBHVF and were reported extremely rarely among the worldwide-studied upper mantle peridotites. The petrographic features and uniqueness of the observed texture inspired us to address it flattened equigranular texture.

As a geochemical summary of the studied peridotites, based on their major element composition, they went through high degree partial melting (20–25 %), which is higher than the usual observed in common upper mantle peridotites of the BBHVF



■ **Fig. 1.** Crystallographic preferred orientation (CPO) patterns of olivine in a studied spinel lherzolite xenolith. Horizontal black lines denote the foliation, the lineation at 90°/0°. The thin sections had been cut oriented in xz- and yz-planes (i.e. perpendicular to the foliation and parallel to the lineation). Sectioning inaccuracies were corrected by rotating the data. Pole figures are lower hemisphere, equal area projections. n: number of grains measured by EBSD.

(~20%) (e.g., Downes et al. 1992, Embey-Isztin et al. 2003, Szabó et al. 2004). Furthermore, the mg#s of the pyroxenes and olivines in the common rock types of the BBHVF are usually lower (89–90) than those of the studied xenoliths (91–92), which correspond to the above mentioned higher degree of partial melting observed in the studied peridotites. Therefore, this depletion is also confirmed by the high orthopyroxene/clinopyroxene ratio (3–4) in the studied xenoliths, which means significant clinopyroxene loss during partial melting. The trace element composition of the xenoliths studied suggests similar geochemical evolution (early depletion followed by different degree enrichment in light rare earth elements).

The olivines have a characteristic crystallographic preferred orientation (CPO) with [010]-axes perpendicular to the foliation and the [100] and [001]-axes forming a continuous girdle in the foliation plane (Fig. 1). Contrarily, the CPO pattern of orthopyroxene is much more scattered, although a double maximum can be observed in [001] planes parallel and perpendicular to the plane of lineation (Fig. 2). In case of olivine, the activation of multiple slip systems: (010)[100] and (010)[001] is suggested. The deformation micro-mechanisms of orthopyroxenes are suggested to be a combination of intracrystalline glide on the (100)[001] system and some kind of other mechanism resulting in quite scattered patterns. We suggest that the unusual orientation patterns of olivines and orthopyroxenes are the result of the complex tectonic evolution of the region. The flattened equigranular xenoliths could represent a structural domain within the subcontinental lithospheric mantle beneath the volcanic field with particular seismic characteristics.

The occurrence of flattened domains in the upper mantle may considerably influence the percolation and residence time of mantle melts and fluids, which could promote or prevent melt/wall-rock interaction. Furthermore, the studied upper mantle xenoliths may provide insight into anisotropic nature of the lithospheric mantle beneath the central part of the Carpathian-Pannonian Region. We suggest anisotropic nature is very likely associated with the tectonic evolution of the Carpathian-Pannonian Region.

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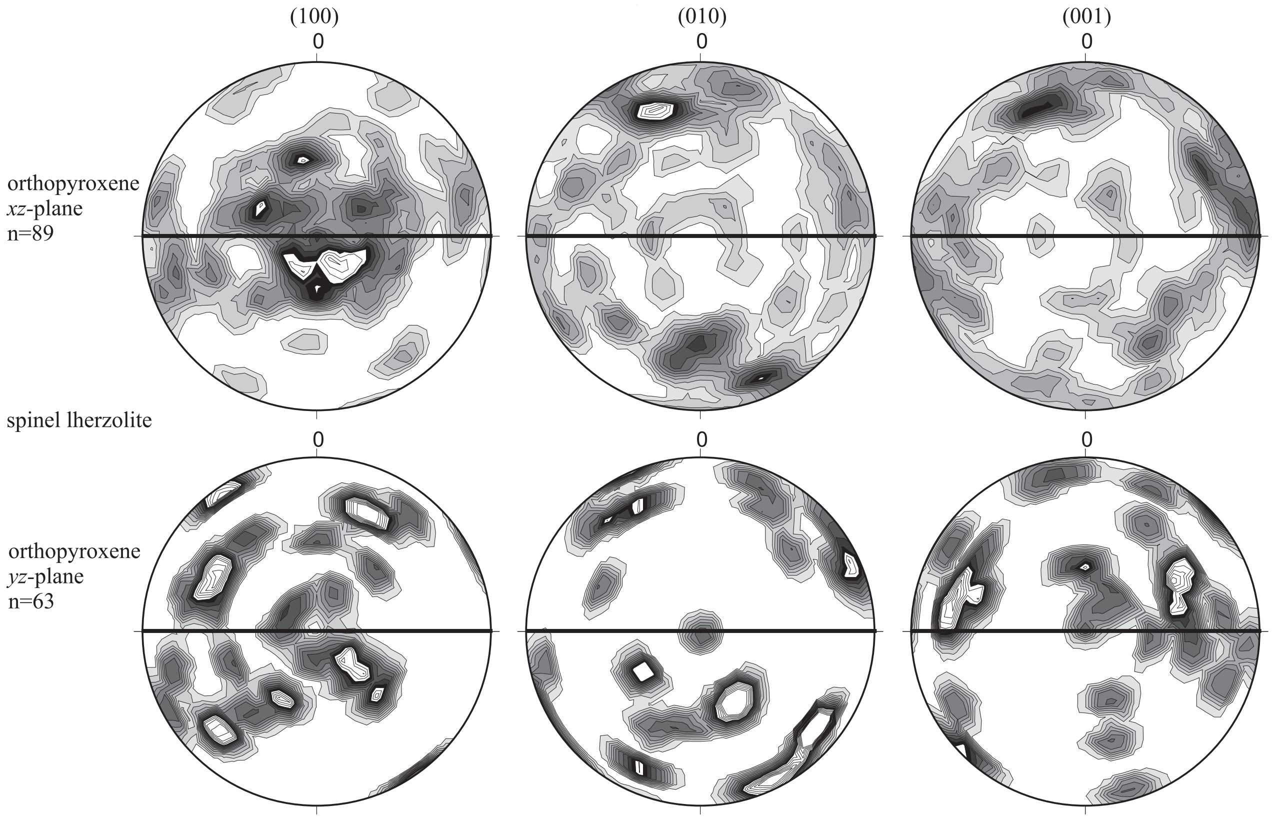

■ **Fig. 2.** Crystallographic preferred orientation (CPO) patterns of orthopyroxene in a studied spinel lherzolite xenolith. Horizontal black lines denotes the foliation, the lineation at 90°/0°. The thin sections had been cut oriented in xz- and yz-planes (i.e. perpendicular to the foliation and parallel to the lineation). Sectioning inaccuracies were corrected by rotating the data. Pole figures are lower hemisphere, equal area projections. n: number of grains measured by EBSD.

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Magnetic Fabric and Ductile Deformation Differences between the Magura and Krosno Groups of Thrust Sheets of the Flysch Belt of the West Carpathians and their Tectonic Implications

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The magnetic fabric in sandstones of the thrust sheets of the Western Sector of the Flysch Belt of the West Carpathians ranges from essentially sedimentary to mostly deformational in origin. In the thrust sheets at the margins of the Flysch Belt (Outer Krosno–Menilite Flysch in the west and Bílé Karpaty unit and Oravská Magura unit in the east), the magnetic fabric is mostly sedimentary in origin, the ductile deformation being very weak, hardly detectable by magnetic anisotropy.

In the central thrust sheets of the Inner Magura Flysch, the magnetic fabric is relatively strongly affected by ductile deformation represented by a combination of simple shear and lateral shortening, probably associated with creation and motion of the thrust sheets driven by a push from the rear side. The ductile deformation is generally stronger in the frontal areas of the individual thrust sheets than in their central areas.

The Krosno lithofacies, mostly occurring in the Outer Krosno-Menilite Flysch, represents the youngest synorogenic flysch sediments largely terminating the last depositional history in the Flysch Belt, embracing the interval from the Late Oligocene to Early Miocene. The deposition of this flysch lithofacies replaced the euxinic sedimentation of underlying Menilite Formation. This change in deposition was connected with Neoalpine orogenetic movements during the Oligocene, which evoked the fundamental re-arrangement of the orogenic belt. In terms of plate tectonics, this re-arrangement represents the stage of closing subduction and starting collision.

Fluid Inclusion, Stable Isotope and Geochronologic Evidence of Cretaceous Collision-Related Formation of Hydrothermal Veins in the Gemeric Basement (Western Carpathians)

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Palaeozoic basement of the Gemeric tectonic unit contains around 1300 siderite-sulphide and quartz-stibnite veins oriented parallel with regional cleavage structure. Origin of the veins has been widely discussed since introduction of modern geochronology methods. Granitic source of ore elements and Cretaceous age of the Gemeric hydrothermal deposits was proposed by Varček (1957). Magmatogenic models invoked Variscan granitoids (Ilavský et al. 1977) or deep mafic intrusions of Cretaceous-Eocene age (Rozložník 1989) as the main sources of ore elements. Metamorphogenic models favoured mobilization of the ore elements during Variscan (Grecula 1982) or Alpine (Varček 1985) tectono-metamorphic processes.

Žák et al. (1991) and Grecula et al. (1995) proposed a metamorphic-hydrothermal model, according to which precipitation of the Gemeric hydrothermal veins was induced by mixing of Variscan metamorphic fluids with evaporite-leaching meteoric waters within Permian rifts. High bromine concentrations in the ore-forming fluids (Hurai et al. 2002) ruled out the presence of evaporite-leaching meteoric water, which was replaced by residual, halite-fractionated seawater infiltrating the Palaeozoic basement from the periodically swamped and evaporated Permian rift/graben in the last versions of the metamorphic-hydrothermal model (e.g. Radvanec et al. 2004, Grecula and Radvanec 2005).

Available stable isotope and fluid inclusion data from the Gemeric hydrothermal veins (Hurai et al. 2002, Urban et al. 2006) are controversial with the concept of rift-related metamorphic-hydrothermal origin. Recalculations based on new fluid inclusion and stable isotope data define formation temperature of 177 to 217 °C, paleodepth of 6.0 ± 0.3 km, and thermal gradient of 33.5 ± 5.5 °C/km for the siderite stage of the Droždiak vein in the northern part of the Gemeric unit. The temperatures of 227–263 °C, paleodepth of 11.2 ± 0.6 km, and thermal gradients of 22 ± 3°C/km have been obtained from the siderite veins in the Rožňava ore field of the southern Gemeric unit. Uniform character of primary fluid inclusions in siderite, i.e. $NaCl-CaCl_2-H_2O$ brines with salinities between 18–25 wt.% NaCl equivalents, and oxygen isotope composition of the parental fluid positively correlated with the metamorphic grade of country rocks (from 5 ‰ in low-grade Permian to 11 ‰ in medium-grade Lower Palaeozoic rocks) are reminiscent of a closed, rock-buffered fluid system. The normal-to-low thermal gradients and paleodepths substantially exceeding available thicknesses of overburden during Permian-Triassic times rule out opening of the vein structures during the extensional tectonic regime incidental with rifting.

Sulphide stage of the Gemeric hydrothermal veins exhibits highly variable fluid compositions, ranging from high salinity (max. 35 wt.%) $NaCl-CaCl_2-H_2O$, CO_2-poor brines to CO_2-dominated aqueous fluids with signs of heterogeneous trapping. The contrasting fluid compositions indicate an open-system fluid behaviour. In the Cucma stibnite deposit of the southern Gemeric unit located near Roznava town, the carbonic fluid is extremely dense (up to 1.197 g/cm3) and admixture of minor CH_4 and N_2 is typical. Fluid inclusion trapping *PT* parameters in the Klement vein of the Cucma deposit correspond to 183–237°C, and 1.6–3.5 kbars, possibly up to 4.5 kbars. The *PT* conditions point to a 15–18 km thick overburden and low thermal gradients, corresponding to only 12–13 °C/km (Urban et al. 2006). These parameters are controversial with the partially molten hot continental crust, and up to 7 km thick overburden at the base of the south-Gemeric basement during the Permian-Triassic rifting. Composition of the gaseous mixture is typical of an externally derived metamorphic fluid, and high-salinity aqueous component probably represents basinal brine modified by cationic exchange reactions with crustal rocks.

U-Pb-Th age of monazite from quartz-tourmaline-white mica assemblage from the Cucma deposit revealed Early and Upper Cretaceous mineral-forming events culminating at 120±9 and 76±12 Ma. The first event is coincidental with thrusting of the Gemeric unit over the adjacent Veporic basement and the formation of the Alpine cleavage structure of the Gemeric basement. The second event corresponds to transpressive shearing and the formation of major trans-Gemeric shear zone.

Geometry of the Gemeric hydrothermal veins together with fluid inclusions, stable isotopes, K-Ar and U-Pb-Th dating support the model of vein opening at gradually increasing thickness of the overriding nappe piles, attaining ~4–5 and 6–7 km during crystallization of early siderite in the northern and the southern Gemeric basements, respectively. A 8–10 km thickness of the overthrusted nappe units must be expected in the south-Gemeric basement to explain high fluid pressures during precipitation of quartz-tourmaline-(white mica-phosphate) assemblage of the quartz-stibnite veins near Roznava (Urban et al. 2006).

K-Ar and U-Pb-Th dating shows short veining interval compared to the age span of the Gemeric cleavage fan formation. The fact that most veins are subparallel to the cleavage indicates Pfluid > T + σn (Pfluid – fluid pressure, T – tensional strength of rock, σn - plane-perpendicular stress; Cosgrove 1997). The opening of tensile fracture parallel to the main anistropy (i.e. Cretaceous cleavage) can be explained in terms of low differential stress (σ1 – σ3), corresponding to small difference between horizontal tectonic stress and vertical overburden pressure, and large difference between tensional strengths (Tp – Tn, where Tp is parallel to main anisotropy, and Tn perpendicular to it). Cosgrove (1997) showed that the tensional failure occurs parallel to main anisotropy in direction perpendicular to the main compressive stress, if (Tp – Tn) > (σ1 – σ3). Therefore, the veining event in the Gemeric unit might have occurred within a narrow time interval at specific stress conditions marked by building of high overburden pressure (vertical load due to thrusting) and strong horizontal stress (horizontal push related to the formation of the Gemeric cleavage fan).

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Metamorphism and Exhumation Processes of the Shotur Kuh Metamorphic Complex, Semnan Province (Central Iran Zone)

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The Shotur Kuh Metamorphic Complex (SKMC) represents an E-W trending elliptical tectonic window (area 20 × 11 km), exposed within Jurassic–Eocen sedimentary sequences 260 km SE of Semnan city (Semnan province, Eastern Iran). The SKMC consists of banded sequence of orthogneisses (metagranite-metatonalite) and amphibolites. U-Pb data obtained by laser ablation ICP-MS analysis of zircon yielded a Cambrian age of ca. 520 Ma, which can be interpreted as corresponding to the protolith crystallization. The absolute age of pervasive amphibolite facies metamorphism is not established yet, but an indication of pre-Jurassic age of metamorphism is indicated by the presence of pebbles of gneisses and amphibolite in Jurassic basal conglomerates. The gneisses are formed mostly by plagioclase ($An_{19\text{-}30}$), biotite, (X_{Mg} = 0.47 or 0.67, regarding to the whole rock composition), in some cases also garnet ($Alm_{58–70}$, $Grs_{24–36}$, $Py_{2–7}$, $Sps_{1–5}$) and accessory amounts of allanite with epidote rim. Tonalitic varieties may additionally contain amphibole. Some quartz-rich mica schists have variable amounts of muscovite. Amphibolites are mostly formed by plagioclase ($An_{18–20}$) and ferropargasite (X_{Mg} = 0.34, Na^{M4} = 0.3), and locally also garnet ($Alm_{56–61}$,$Grs_{30–34}$,$Py_{3–5}$,$Sps_{1–5}$). PT conditions of 520–560 °C and 6–8 kbar have been estimated using Grt-Bt, Grt-Amph thermometry and Grt-Amph-Pl barometry, in combination with thermodynamic calculations using the PTGIBS software. The amphibolite facies metamorphism was accompanied by strong deformation (D1) that resulted in formation of isoclinal syn-schistose folding of compositional layering and strong mineral foliation bearing stretching and mineral lineation trending in the E-W direction. All rocks are affected by variable degree of retrogression, mylonitization and brittle-ductile reactivation of the early high grade fabric. This late event was synchronous with very low-grade metamorphism and deformation (D2) developed in the adjacent Jurassic meta-sediments. Kinematic indicators, mainly asymmetrical buckling and asymmetrical boudinage of cherts in weakly metamorphosed limestones and dolomites suggest a N-S compression and top to the north shearing. Foliation in the Jurassic rocks which is parallel to mylonitic shear zones affecting the gneiss and amphibolite complex dips 60° to the north along the northern margin of the SKMC. However, in the southern and central parts of the complex the greenschist facies fabrics dip under 20–40° to the south. Cretaceous age of this tectonometamorphic process is inferred from the presence of pebbles of Jurassic slates and weakly metamorphosed carbonate rocks in the Upper Cretaceous basal conglomerates.
In addition, late thin skinned deformation (D3) and N-S verging folding in the Eocene sediments date the yougest tectonometamorphic process in this area. This analysis clearly demonstrate that the exhumation of the Shotur Kuh Metamorphic Complex resulted from active buckling of the basement during a N-S directed Cretaceous to Tertiary compression and de-coupled thin-skinned tectonics of the Mesozoic cover. However, the age of main metamorphic event affecting the SKMC remains yet not constrained.

Emplacement Mechanisms of the Thrust Sheets in the Barrandian (Bohemian Massif)

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The Prague Synform (Teplá–Barrandian region, Bohemian Massif), referred in old papers as a Prague Basin, represents remains of Ordovician to Middle Devonian sedimentary units folded into a large synclinorium. The term Prague Basin was a consequence of geological interpretations of the Prague Synform as a relatively small and isolated sag of synsedimentary origin. The issue of older models is broad and complicated, further details are discussed in Melichar (2004).

First of three major faults under our study was the Tachlovice Fault. This fault is at least forty kilometres in length, strikes in WSW–ENE direction, dips in 45 ° SE and could be traced from the surroundings of Beroun through Prague to Běchovice. Our research focused on fault accompanying structures which show a kinematic pattern in strong disagreement with the old models. The opposite sense of tectonic transport along the Tachlovice Fault, top-to-the SE ("normal" faulting) was documented by small-scale structures (fault-detachment folds, small ramps, S-C structures and others). If we examine the Tachlovice Fault on the map there are visible duplications of the rocks mainly of the Silurian and upper Ordovician age. A nice example is located e.g. in the surroundings of the Tachlovice village.

The Očkov Thrust is next important fault. It is nicely exposed in the old Na Kobyle Quarry near Koněprusy where the Lochkovian rocks were thrust over the Pragian ones, with an increased tectonostratigraphically distinct gap in close neighbourhoods (Silurian rocks of the Požáry Formation above the relicts of mid-Devonian Srbsko Formation). The Očkov Thrust is about forty-five kilometres long but of varying strike due to its curved shape. A longer part of the Očkov Thrust strikes similarly as the Tachlovice Fault in WSW–ENE direction but it dips to the NW and could be traced from the surroundings of Běchovice through Prague to the surroundings of Zadní Třebaň. There the fault changes its strike into W–E, going to Koněprusy where its further continuation is unknown. Looking on the Očkov Thrust in the map there is a visible three times repeating of the Silurian and upper Ordovician formations in the vicinity of Zadní Třebaň. The Koda Fault is last of the three major faults. It is located almost in the middle of the Prague Synform striking WSW–ENE and dipping to the NW. This fault is poorly exposed in the mouth of the Loděnice Creek (Kačák) into the Berounka River. It divides facially quite different Devonian and upper Silurian rocks and three times duplicates the Silurian Kopanina and Požáry formations and the Devonian Lochkov Formation near Radotín.

The folding style is characterized by brachyal folds (with wavelength greater than hundred metres) that are all connected in major faults. Almost everywhere they have steeper south oriented limbs against to the north verging limbs. There are relatively shallow and wide synclines and steep but tight anticlines. Average axis of the folds is oriented 243 °, but the orientation of folds is slightly changing to be in similar direction as a nearby lying fault. Small folds are abundant in less competent, mostly layered rocks and are very variable in shape and orientation of axis even in the scale of one outcrop. There is a wide variety of fold shapes. We can mention e.g. tight to isoclinals folds from the Srbsko village located on the left side of the Berounka River or isoclinals folds from the Jedlička Rock near Radotín. Disharmonic folding is known from the Budňany Rock in Karlštejn, Barrande Rock in Prague or Homolka Hill near Velká Chuchle. Kink style folds are less abundant except of the thin laminated siliciclastic Devonian Srbsko Formation. The vergence of folds is oriented mainly to the SE, with the exception of rarely occurring NW verging folds situated in the frontal parts of antiforms. Folds axes calculated from 177 folds show two very close maxima – 250/9 and 238/8 respectively.

Very interesting data were obtained from small ramp-and-flat faults which are often connected with intensive folding in accompanying rocks. A nice example can be traced in the Budňany Rock, just next to the parastratotype of the Silurian/Devonian boundary. Next examples are located in the Devonian Zlíchov Formation in the Mramorka Quarry near Chýnice and on the Tachlovice Fault near Beroun in the Silurian Liteň Formation. Small fold propagation folds are found rarely. A nice example is located on the Tachlovice Fault near Beroun in the Silurian Liteň Formation. All these structures are showing everywhere top-to-the SE sense of movement.

Our arguments are supporting top-to-the SE sense of the movement in the central parts of the Prague Synform. Large folds are related to major faults (e.g. the Amerika Anticline is related to the Koda Fault, Hradinov Hill Anticline is related to the Očkov Thrust). We suppose that this is caused by changes in geometry of faults (ramps) beneath the surface. There are great stratigraphic inversions and duplications along major faults (Silurian over Devonian, upper Ordovician over Silurian).We used a stratigraphic separation diagram (SSD) to prove that the Tachlovice Fault and the Očkov Thrust are faults with a ramp and flat geometry (Melichar 2004). We have arguments in support of thrusting of shallower Silurian facies (tuffs+limestones) over the Silurian graptolitic black shales as was proposed by Bouček (1941) and confirmed by Melichar (2004). Taking in mind top-to-the SE sense of tectonics movement and accepting allochthoneity of the shallower tuffitic Silurian facies of the Liteň and Kopanina formations we can assume at least 5 km transport of these rocks.

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Contrasting Petrogenesis of two Volcanic Suites in the Devonian Vrbno Group (Hrubý Jeseník Mts., Czech Republic)

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The problem of determining palaeotectonic setting of old volcanic suites becomes quite challenging in metamorphosed terrains out of original structural context. In Silesicum (NE Bohemian Massif, Czech Republic), near the eastern termination of the Rhenohercynian Zone of the Variscan chain (Franke 2000), petrologically extremely variable metavolcanites occur as a part of the palaeontologically dated Devonian sedimentary sequence of the Vrbno Group (VG). Current controversies in interpretation of the petrogenesis and geotectonic setting of the VG goes partly on account of the separate use of a relatively narrow range of geological, petrographic, geochemical and/or petrophysical methods in previous studies. In addition it reflects a polyphase tectonometamorphic overprint; the rocks of the VG were deformed, imbricated and metamorphosed jointly with their mainly metagranitic Cadomian basement (Schulmann and Gayer 2000 and references therein).

Regardless the presence of greenschist-facies metamorphic assemblages, volcanic structures are locally well preserved. Thus the primary character of the volcanic products can be determined: pillow lavas, ignimbrites, banded tuffs, agglomerate tuffs and subvolcanic dykes. In the studied southern part of the VG, volcanosedimentary and bimodal volcanic rocks occur in two approximately N–S trending belts, separated by little deformed Cadomian metagranitic parautochton (the Oskava Block) (see also Aichler et al. 2004): (1) The geochemically relatively primitive Western Volcanic Belt (WVB), restricted to a narrow rim of the Cadomian basement, is characterised by an abundance of metasediments accompanied by mostly basic–intermediate metavolcanites; acid volcanites are subordinate. (2) The more evolved Eastern Volcanic Belt (EVB), covering a significantly larger area between Malá Morávka and Uničov E of the Oskava Block, is predominantly metavolcanic. The relative proportion of acid volcanic products is much larger. In addition, there are rare felsic dykes (rhyolites and comendites/pantellerites) cutting the Oskava Block itself. Finally, numerous dolerite dykes penetrated both Cadomian and Devonian sequences.

The metavolcanites of the **Western Volcanic Belt** are exclusively calc-alkaline in chemistry. Basalts–andesites are of submarine origin as shown by locally preserved pillow lavas. The NMORB-normalized spiderplots (Sun and McDonough 1989) are characterized by marked depletions in Nb, Ti and Sr. The LILE contents are extremely variable, reaching up to *c.* 450 × NMORB. Such remarkable LILE/HFSE enrichments point to a continental arc geotectonic setting (e.g. Pearce and Parkinson 1993, Tatsumi and Eggins 1995). The chondrite-normalized REE patterns (Boynton 1984) are rather flat ($La_N/Yb_N = 3.60–7.45$; $La_N/Sm_N = 2.33–3.12$). Both ratios increase with SiO_2 as does the magnitude of the Eu anomaly (Eu/Eu* = 0.91–0.66). The Nd isotopic data are compatible with derivation from a moderately depleted mantle source ($\varepsilon^{390}_{Nd} \sim +3.3$, $T^{DM}_{Nd} = 0.83$ Ga – a two-stage Nd model age of Liew and Hofmann 1988).

The felsic (rhyolitic, $SiO_2 = 71.8–81.7$ wt. %) samples from the WVB show higher degree of LREE/HREE fractionation ($La_N/Yb_N = 4.39–8.04$; $La_N/Sm_N = 3.26–4.91$). The Eu anomaly is significantly deeper (Eu/Eu* = 0.75–0.14) and its magnitude generally increases with rising silica. The LREE and HREE drop in the same direction. The chemistry of rhyolites also resembles a volcanic-arc geochemical signature (Pearce et al. 1984) and their Nd isotopic composition is in line with their possible derivation from immature crustal source or by nearly-closed system fractional crystallization of the parental basaltic melts ($\varepsilon^{390}_{Nd} \sim +2.9$, $T^{DM}_{Nd} = 0.86$ Ga). The importance of feldspar(s) and apatite fractionation is supported by a marked drop in Sr, P, Eu and Ti with increasing SiO_2. Role for contamination by geochemically immature and isotopically undistinguishable Cadomian basement is difficult to assess, even though some upper crustal contribution is unequivocal based on $\delta^{18}O$ values (10.3–13.0 ‰ SMOW) elevated for all samples (Davidson et al. 2005).

In the **Eastern Volcanic Belt** abundant alkaline volcanics span the whole compositional range from alkaline basalt to comendite, with acid rocks prevailing in outcrops. At least partly, their structures indicate subaeric origin (agglomerate tuffs, ignimbrites). The NMORB-normalized spiderplots differ strikingly from the western belt by the absence of Nb trough. For the samples with $SiO_2 < 69$ wt. % is characteristic depletion in Ti, Sr, P and Eu. While the LILE exceed 1250 × NMORB, HREE are enriched only *c.* 1.5–5.5 times. The REE patterns are variable; the least fractionated samples are characterized by low total REE contents and practically lack any Eu anomaly (Eu/Eu* = 0.9, ΣREE ~ 320 ppm), whereas the most fractionated samples have high total REE contents and deep Eu anomaly (Eu/Eu* = 0.2, ΣREE ~ 890 ppm). The mafic alkaline rocks of the EVB are represented by a volcanic bomb in agglomerate tuffs, whose radiogenic Nd documents its independent position. It points to Devonian partial melting of a time-integrated, strongly LREE-depleted mantle source with little scope for crustal contamination (ε^{390}_{Nd} = +6.9, T^{DM}_{Nd} = 0.55 Ga).

The felsic rocks of the EVB have Al_2O_3 concentrations higher than 1.33 × (FeO^T + 4.4) and thus can be classified as comendite and comenditic trachyte (MacDonald 1974). They show generally highly fractionated REE patterns (La_N/Yb_N = 1.84 to 10.20; La_N/Sm_N =1.73–5.32,) with negative Eu anomalies deepening with increasing degrees of fractionation (Eu/Eu* =0.18–0.11). The total REE contents decrease from 943 to 187 ppm with increasing silica (i.e. increasing degrees of fractionation), reflecting a concomitant drop in LREE and HREE. The zircon typology (Wilimský et al. 2005) and whole-rock geochemistry of the acid volcanics resemble Within Plate Granites (WPG, Pearce et al. 1984). Additionally, these rocks show high contents of HFSE (Nb, Ta, Y, Zr) as well as high Ga/Al and Fe/Mg ratios, typical for within-plate, A-type igneous activity (Eby 1990, Collins et al. 1982). Their radiogenic Nd (ε^{390}_{Nd} ~ +2.8 to +3.8) and primitive $^{87}Sr/^{86}Sr_{390}$ (~ 0.704) rule out derivation from mature crustal sources; the rather heavy oxygen (13.7–15.7 ‰ SMOW), however, precludes a closed-system fractionation from the Earth's mantle (Hoefs 2004). Viable hypotheses thus involve intracrustal derivation, probably of the mainly granitic Cadomian basement of the Oskava Block (Hanžl et al. in review).

Most of the dykes penetrating the more westerly **Oskava Block** are alkaline, closely resembling the chemistry of the volcanic rocks from the EVB (ε^{390}_{Nd} = +2.8; oxygen slightly lighter, $\delta^{18}O$ = 12.0 ‰ SMOW). Rarer seem to be dykes with an overall calc-alkaline, WVB-like chemical signature.

Finally, the tholeiitic **dolerite dykes** and sills have remarkably primitive isotopic chemistry. The Nd isotopic signature is compatible with direct derivation from a Depleted Mantle source in Devonian times (with ε^{390}_{Nd} = +7.8 to +8.0, T^{DM}_{Nd} = 0.46–0.48 Ga) and this is also in line with the oxygen isotopic data ($\delta^{18}O$ = 5.5 to 6.6 ‰ SMOW). The elevated Sr isotopic ratios ($^{87}Sr/^{86}Sr_{390}$ = 0.705 to 0.706) and less radiogenic Nd compositions some of the samples (down to ε^{390}_{Nd} = +5.3) can be explained by crustal contamination. Such scenario is confirmed in many NMORB-normalized spider plots by positive anomalies of Rb, K, Sr and Pb as well as Nb troughs.

Patočka and Valenta (1996) with Patočka and Hladil (1997) outlined a model in which the volcanites of the VG originated in a volcanic arc geotectonic setting with a transition to a back-arc spreading. According to these authors, the apparent scarcity of volcanites with a destructive margin geochemical signature could be due to a deep erosion of the former arc, documented by accumulation of large masses of quartzites. The current study has indeed confirmed such a view. The metavolcanic rocks in the VG apparently form two distinct volcanic provinces: (1) western with a most likely convergent geotectonic setting and (prevailing) submarine origin, and (2) eastern, at least partly subaeric, back-arc rift-related alkaline suite. The original configuration of both volcanic sequences, preserved only as fragments, is still largely open to debate. Based on palaeomagnetic data, the original orientation of the Devonian basins in Moravia was E–W (Hladil et al. 1999). The subduction was most likely south-dipping (Franke and Żelaźniewicz 2000). The Devonian basins seem to have rotated *c.* 90 degrees clockwise in the Late Devonian–Early Carboniferous (Hladil et al. 1999). Following this rotation, the EVB could have been thrusted eastward (cf. Schulmann and Gayer 2000) over the Cadomian basement to which the WVB stuck as a relative parautochton. This scenario is in line with the conspicuously zoned distribution of the Devonian volcanic rocks as well as our observation of the tectonic contact between pillow lavas and overlaying lowermost members of the Devonian VG sequence in the WVB.

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On the Genesis of Two Meridionally Trending Lineations in Rocks of the Orlica-Śnieżnik Dome: Evidence from Marbles of the Stronie Formation

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Character and kinematics of the meridionally trending lineations in the Orlica-Śnieżnik Dome (OSD) have been widely discussed and diversely interpreted. Because this lineation is composite tectonic feature (neglection of that fact can lead to erroneous, simplified conclusions) its interpretation has to be carried out with respect to the superimposed deformational events distinguished in rocks of the OSD. The very important aspect of this investigation is the correlation of N-S trending tight recumbent folds preserved mostly in metapelites of the Stronie formation and similarly, N(NE)-S(SW) trending stretching lineation observable mostly in orthogneisses. The N-S trending lineation in the Stronie formation is considered to be associated with the N-S trending tight folds (e.g. Teisseyre 1975, Don 1982). In gneisses, the regional elongation along N-S trending rodding lineation could be the result of either coincidental strain due to N-S tectonic escape induced by the E-W shortening (Żelaźniewicz 1988) or the NE-SW strike slip in transpressional regime (Cymerman 1997). Żelaźniewicz (1988) connects development of N-S stretching lineation with the early tectonic stage of the OSD gneisses evolution, whereas Cymerman (1997) assumes that all tectonic features of the gneisses developed during one deformational event.

On the basis on structural reconstruction and geothermometric calculations carried out for marbles of the Stronie formation it can be stated that the N-S trending linear structures observed in the rocks of the Stronie formation result from two separate events characterised by different metamorphic and kinematic conditions. This explains the ascertained occurrence of two lineations: (i) intersection and (ii) stretching, where each of them becomes locally dominant. Marbles were chosen because of their rheological properties allowing for a good distinction between tectonic features developed during consecutive tectonometamorphic stages. The earliest distinguished N-S trending lineation in marbles is defined by the intersection of a tightly folded (F2) early metamorphic fabric S1 and the axial planar foliation S2 marked by the parallel alignment of flattened carbonates and plate- and needle-shaped silicates (Phl-Ms±Tr±Czo). On the folded planes, the intersection lineation manifests as frequently arranged thin trails of this new metamorphic lamination. On the exposed axial planes of tight folds F2, the lineation occurs as differently thickened and shaded bands, depending on the thickness and colour of the folded beds. The folding is interpreted to be induced by the E-W shortening leading to the lithospheric thickening (in accordance with Dumicz 1979). Tight folds formed during orogenic uplift related to relaxation of the lithosphere (also documented by the flattening of the inclusion trails in garnets in mica schists). At the onset of this uplift temperature peak of metamorphism in the upper amphibolite facies conditions took place.

The axial planes S2 are overprinted by locally observed new meridionally trending stretching lineation. This lineation is defined by elongated carbonate clasts and linearly accumulated carbonates in pressure shadows of less ductile domains e.g. boudinaged Czo porphyroclasts. Products of this tectonic stage have been recorded by locally developed dynamically recrystallized shear zones, which are characterised by strong grain size reduction and elongation of carbonate grains. Angular relations between developed S- and C'-type planes as well as geometry of σ-clasts point to top-to-the-N shearing along the reactivated former axial grain shape fabric. Due to progressive deformation rocks were locally transformed into the L-tectonite. These processes took place during retrograde conditions, at temperatures lower by ~100 °C than those accompanied with the folding stage. The top-to-the-N shearing recognised within the OSD can be correlated with the sinistral movements in the Złoty Stok – Skrzynka Zone, as stretching lineations in these areas have the same position in the sequence of deformation. This deformational stage could be possibly linked to NNE-directed thrusting of the OSD, when OSD interacted with adjacent domains. Within the OSD, the deformation was heterogeneous and partitioned into laminae, hence parallel to the former axial planes.

In conclusion, two generally N-S trending lineations could be distinguished in rocks of the Stronie formation. The orientation of the foremost intersection lineation delineates the Y-axis of the strain ellipsoid representing the tectonic stage related to the temperature peak of metamorphism. Formation of the subsequent stretching lineation was related to uplift and retrogression. Contrary to the intersection lineation, its orientation shows the direction of the maximum strain component. These observations could partially explain controversy regarding the presence of the mineral lineation nearly parallel to the fold axis. In orthogneisses, high temperature conditions, at which the rodding lineation was formed, point to its connection with the N-S directed tectonic escape induced by E-W shortening (according to Żelaźniewicz 1988). Transition from prolate to oblate shapes of the rodding lineation (Żelaźniewicz 1991) could be related to the flattening strain responsible for folding in the Stronie formation. Later top-to-the–NE reactivation at greenschist facies conditions (Żelaźniewicz 1991) concurs with the late shearing that gave the stretching lineation in the Stronie formation.

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Application of Newly Developed ArcGIS Software Extensions for Localization of Faults and Natural Zones of Methane's Escape by Morphotectonic Analysis (Moravosilesian Region)

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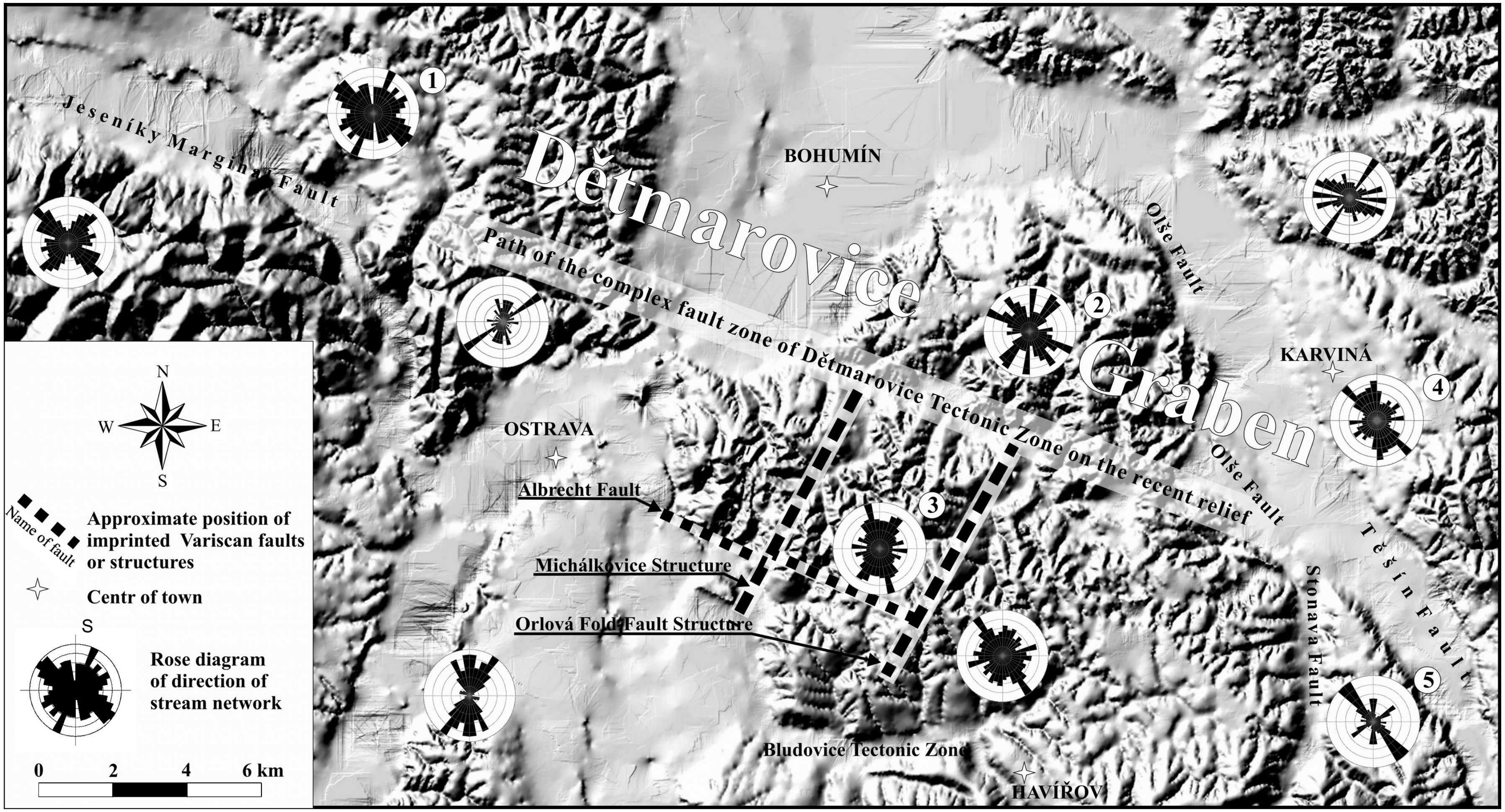
The framework of Variscan coal-bearing molasses represented in the Moravosilesian region by the Czech part of the Upper Silesian Coal Basin (USCB) was reactivated and modified by a sedimentary loading of the Inner Carpathian molasses and tectonic movement of accretion wedge of the Outer Carpathian nappes during the period of Alpine orogeny. The sedimentary and tectonic loading initiated significant rejuvenation of older Variscan structures. A lot of reactivated Variscan faults of the USCB were imprinted



■ **Fig. 1.** DTM (shaded relief) of the Ostrava-Karviná coalfield – the recent relief. A resolution of the DMT is 10 meter / grid cell. In figure imprinted main Variscan structures of the Czech part of the Upper Silesian Coal Basin into recent relief are marked. Rose diagrams are put to centre of analysed polygons of stream network. Minimum length of each stream segment is 100 meters.

through the Outer Carpathian nappes and sediments of Inner Carpathian molasses to recent relief (Jelínek and Grygar 2002). There is assumption about escape of methane up to relief by Alpine reactivated Variscan tectonic zones. In the fifties Petránek (1954) pointed up this hypothesis about methane's escape by reactivated tectonic zones as the first. His early ideas about the influence of young Tertiary tectonic on structural-tectonic conditions of Variscan accretion wedge of the USCB have been overlooked.

The object of resolved project is aimed at detection of features extent of Variscan structures to recent relief in Moravosilesian area. Preliminary results of the study of character manifestation of tectonic pattern in recent relief by morphotectonic analysis are helpful for localization of potential natural zone of methane's escape. Modern methods of morphotectonic analysis make use of results of morphometric analysis of digital terrain models (DTM) confronted with results of structural-tectonical analysis (Jordan et al. 2003). Application of digital terrain analysis in GIS environment enables fast mutual confrontation of results of geomorphologic analysis with geological data. Unfortunately there is no accessible software at GIS, which would be able to perform combined morphotectonic and structural analysis on DTM data. It was necessary to create user friendly software for simultaneously statistic analysis of field structural data and morphostructural DTM data at ArcGIS 9.0 by ESRI company. The ArcGIS 9.0 software was chosen as common GIS software. An asset of ArcGIS is facile linkage with Visual Basic, which was used as programming language for created extensions. The standard software from ESRI does not include appropriate extensions, which would be able to resolve specific problems of structural and morphotectonic analysis. A disadvantage of ArcGIS is also a limited set of interpolation methods for calculation of DTM. For that reason DTM were calculated by Surfer 8 software. This program is more suitable for interpolation of DTM. Unfortunately at the present-day no software module for transforming grid data from GRD file format of Surfer to grid data ESRI format of ArcGIS exists. The first developed extension is conversion of grid data from Surfer 8 software into ESRI ArcMap. The function transfers feature data into ESRI shapefile format and grids into ESRI grid file format. The second extension is used for interactive fast saving of structural data of map to existent structural database. Another extension uses statistic methods for analysis of structural data or analysis of azimuths of polylines in different coverages. Typical examples are stream network (Fig. 1), contour lines, morpholineaments, photolineaments etc. Application finds out azimuth of each segment of polyline (e.g. river stream) located in defined polygon. Azimuths are shown in rose diagrams for each separate polygon. The orientation of each segment of polyline is picked out by different colour for easier confrontation with results of other directional analysis.

Until recently studies of geomorphological and structural-geological character of relief were resolved without mutual confrontation of results of morphological and geological analyses. Modern morphological methods of study of relief comprise structural-geological analysis of structure framework (Pánek 2004). For easier mutual confrontation of results of morphostructural methods with structural–tectonical methods, it was necessary to create a new modulus. The selection of analysed data in the created modulus proceeds simultaneously at all coverages of polylines or structural data. Till this time the selection has been proceeded individually for each polygon and for each coverage. Correctness of function and suitability of using created extensions at a morphotectonic analysis was tested in the Ostrava-Karviná coalfield.

Suitability of using the first module was tested not only at generation grid of DTM, but also at transmission, a grid data of second directional derivation of DTM. This analysis is employed at study of curvature relief and study of morpholineaments (Jordan et al. 2003). The impossibility to implement morphotectonic methods for analyses of DTM from ArcGIS to Surfer (and on the contrary) extends field of usage of created extensions. The possibilities of transforming various grids offer their uses in all series of other field of study not only in morphotectonic analysis. The GRD grid format with a square cell is the requirement for correct function of the first extension at transmission data grid from Surfer to the ArcGIS. Format ESRI is able to save only regular grid with square cells in contradistinction to GRD format, which allows storage of an irregular grid formed from rectangular cells. The user must already take into account this fact at generation of the grid in Surfer. Surfer normally sets maximum and minimum X and Y on the base of border input points at automatically loaded input data. Exactly square matrix cell grid will originate only quite rarely at partition data spaced on integer number of lines and columns. Therefore it is more suitable to order the same size of the grid cell along both X and Y axes and to adjust subsequently maxima and minima upon this axis.

The morphotectonic analysis of DTM, which is oriented to localisation of natural methane's escape by brittle deformation, together with results of structural-tectonical analysis detect a complicated structure relation between recent relief and tectonic deformation of research area. The methodical procedure of morphotectonic analysis included representative morphometric and special terrain analyses of DTM (e.g. the second directional derivation, the slope aspect, the aspect of slope orientation, the aspect of drainage orientation, the digital determination of topolineaments, the aspect of morpholineament orientation, the aspect of brittle deformation orientation, etc.) and paleostress analysis of terrain data. The resolution of analysed digital model is 10 meter / grid cell.

The recent relief of the Ostrava-Karviná coalfield is much more markedly formed by exogenous factors than by endogenous factors. However results of executed morphotectonic analysis in many aspects proved neotectonic rejuvenation of the study area. In relief, where exogenous factors are dominant, a significant influence of structural framework to character of relief occurs (Ahnert 1998, Bloom 1998, Ritter et al. 2002). The most markedly imprinted Variscan structure in recent relief of the Ostrava-Karviná coalfield is the Dětmarovice Tectonic Zone. This wide graben is put together from many partial faults. There is interpretation of imprinted complex faults zone of the Dětmarovice Tectonic Zone in Figure 1. The azimuth analysis of stream network supports link between river system and imprinted Variscan structures. The main direction of rose diagrams No. 1 and No. 2 is identical with complex fault zone of the Dětmarovice Tectonic Zone WNW-ESE (Fig. 1). The Dětmarovice Tectonic Zone connects the west of Doubrava and Eleonora Faults with the fault systems of the same direction in the western part of the Czech part of the USCB and continues as a Jeseníky Marginal Fault as far as to Opava town. The intrusions of neovolcanites and the occurrence of mineral water rich in CO_2 (Dopita et al. 1997) along this tectonic zone are an evidence of the Neoidic geodynamic activities.

Results of comparative morphotectonic analysis confirm also Alpine reactivation of many other Variscan structures (Michálkovice Structure, Orlová Fold-Fault Structure, Albrecht Fault, Olše Fault, Těšín Fault, Stonava Fault, Bludovice Tectonic Zone, etc.). The second main direction of rose diagram No. 3 (NNE-SSW) corresponds with direction of imprinted Michálkovice Structur and Orlová Fold-Fault Structure (Fig. 1). Rose diagrams No. 4 and No. 5 show main direction of drainage system (NW-SE) identical with Těšín Fault and Olše Fault. The Těšín Fault is part of the Jablunkov Tectonic Zone which is noticeable also in DTM of paleorelief of the Brunovistulicum with its Paleozoic cover (Jelínek and Grygar 2002).

Described Variscan structures reactivated by Alpine orogeny could be potential natural zones of methane's escape. The project has significant asset not only for localisation of natural zones of methane's escape but also for morphotectonic methodology. Developed ArcGIS extensions supplemented methods of morphotectonic and structure-tectonic study in GIS environment.

Acknowledgements

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Deformation Pattern Related to an Orogen Parallel Extension Event Recorded in the Vepor Unit, West Carpathians

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The Vepor unit composed of pre-Alpine basement and Late Palaeozoic to Mesozoic cover sequences is one of the major crustal segments incorporated into the Alpine structure of the Central West Carpathians. In this contribution we discuss deformation pattern related to Createaceous orogen parallel extension event recorded in the Vepor basement. The studied deformation heterogeneously affects late Variscan granitoids as well as Variscan high grade orthogneiss, migmatites and paragneiss. Cretaceous reworking of steeply inclined E-W trending Variscan fabric is characterized by the development of sub-horizontal mylonitic fabric in area of about 800 km^2 large. The mylonite foliation bears E-W trending stretching lineation, which is parallel to hinges of isoclinal folds preserved in low strain domains. The development of mylonitic fabric is associated with a prograde metamorphic mineral assemblage, which by using thermodynamic modelling in Perple_X indicates metamorphic P-T conditions 430–590 °C and 5–8 kbar. The distribution of P-T data in the central part of the Vepor Unit indicates an E-W metamorphic field gradient showing higher grade metamorphic conditions towards structural footwall in the west. This observation is in a good agreement with micro-structural analyses in this area showing higher temperature micro-structural features and bigger recrystallized quartz grain size in the west. Following our P-T profile, the mean quartz grain size shows an increase from 63 to 95 microns towards the west. Such grain size variations were also recognised in other parts of the Vepor Unit proposing an existence of similar metamorphic field gradients across the whole studied domain. As a common feature the grain size generally decreases towards the meta-sedimentary cover indicating lower metamorphic conditions in the cover. In selected samples the quartz lattice preferred orientation (LPO) was determined by using electron back-scattered diffraction and computer integration polarization microscopy methods. Both the methods indicate activity of basal a, rhomb and prism a slip systems during the recrystallization, which is in a good agreement with calculated P-T range. The LPO determination revealed single and crossed girdle rotation patterns showing conflicting westward or eastward shear senses detected even in the same thin section. The absence of uniform shear sense during the deformation suggests a pure shear dominated process of ductile thinning operated in the Vepor basement. To explain structural and metamorphic evolution of above described horizontal crustal scale shear zone, we propose that the studied deformation is related to the overthrusting of the southern Gemer Unit upon the Vepor Unit resulting in an orogen parallel extension within the Vepor basement.

Tectonic Control and Basin Evolution of the Northern Transdanubian Eocene Basins (Vértes Hills, Central Hungary)

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The Transdanubian Palaeogene Basin was situated behind the active Carpathian thrust front (Tari et al. 1993). Due to the overprinting Neogene tectonic phases and the poor outcrop conditions, the origin of this basin was long-time debated; suggestions include extensional, compressional, and strike-slip settings (Báldi-Beke and Báldi 1985, Fodor et al. 1992, Tari et al. 1993). We used surface mapping, structural measurements, tectono-sedimentary observations and borehole analyses to describe Eocene sediment pattern around the Vértes Hills (northern Transdanubian Range) and to better understand basin evolution.

After Mesozoic carbonate sedimentation, a long period of late Cretaceous to early Eocene terrestrial denudation resulted in a sub-horizontal peneplain and the occurrence of bauxite lenses. The Eocene (late Lutetian–Bartonian) sequence started with a lagunal-marine coal-bearing clastic unit. It is covered with shallow marine marl, than open marine claystone. Sedimentation on basin margins were characterized by the Szőc Limestone Fm. deposited on low-angle, relatively narrow carbonate ramps. The inner ramp is represented by 4 microfacies types, extraclast rudstone to extraclast-bioclast floatstone (basal beds of Szőc Limestone), bioturbated Foraminiferal-Molluscan-Echinoiderm packstone/ grainstone (interpreted as sea-grass meadows), skeletal grainstone (bioclastic sand shoals), and Nummulites perforatus rudstone/ packstone (Nummulites banks). Mid-ramp is characterized by the predominance of larger Foraminifera under the influence of occasional storms. On the outer ramp glauconitic bioclastic grainstone composed of mainly larger foraminifera, red algae, and bryozoa deposited in current agitated high-energy conditions. The main influencing paleoecological factors were depth, light intensity, hydrodynamic energy, substrate, nutrient content, and sedimentation rates. The inner ramp was characterized by high energy well-lit conditions with the highest nutrient content and highest sedimentation rates. The mid ramp records oligotrophic environment with moderate/low energy and light conditions. The outer ramp is characterized by high hydrodynamic energy with low light intensity and low sedimentation rate (Pálfalvi 2004).

Sediment pattern was determined by two NE striking elevated ridges, (the Dad and Vértes ridges) and two parallel depressions. The development of the Eocene sequence is different on the NW versus SE side of the southerly located Vértes ridge. Thickness is smaller in the NW (Oroszlány depression) than in the SE (Kincses-Magyaralmás depression). The latter was characterized by the deposition of alternating molluscan, Miliolina or Nummulites marl and limestone (Kopek 1980) in a restricted/open lagunal to open marine environments in permanently shallow water conditions.

The Vértes ridge was dissected by NW to W trending synsedimentary monoclines, which are frequently breached by syn-sedimentary faults. Major cross-structures include the north–eastern and southern boundary fault of the Tatabánya depression, the Gesztes fault, the Zámoly-Bükk fault, which all have a strike-slip character. The north–eastern margin of the Csákberény trough, and the Nagyegyháza depression seem to be bounded by normal faults. The surface-rupturing faults were mantled with fault-bounded breccia or conglomerate bodies (Bada et al. 1996). Abrasion frequently rounded clasts derived from these scarps. The fault planes themselves or the abrasional gravels on the fault scarps are frequently bioperforated (Kercsmár 2005). The scarp-related limestones were frequently deformed during the diagenesis, due movement of underlying faults. The syn-diagenetic structures include boudinage, intraformational breccias and sedimentary dikes. Sedimentary dikes also occur along major structures. Seismic activity related to faults could induce redeposition of shallow water sediments toward basin centres in form of different cohesive gravity flows and was generated distally

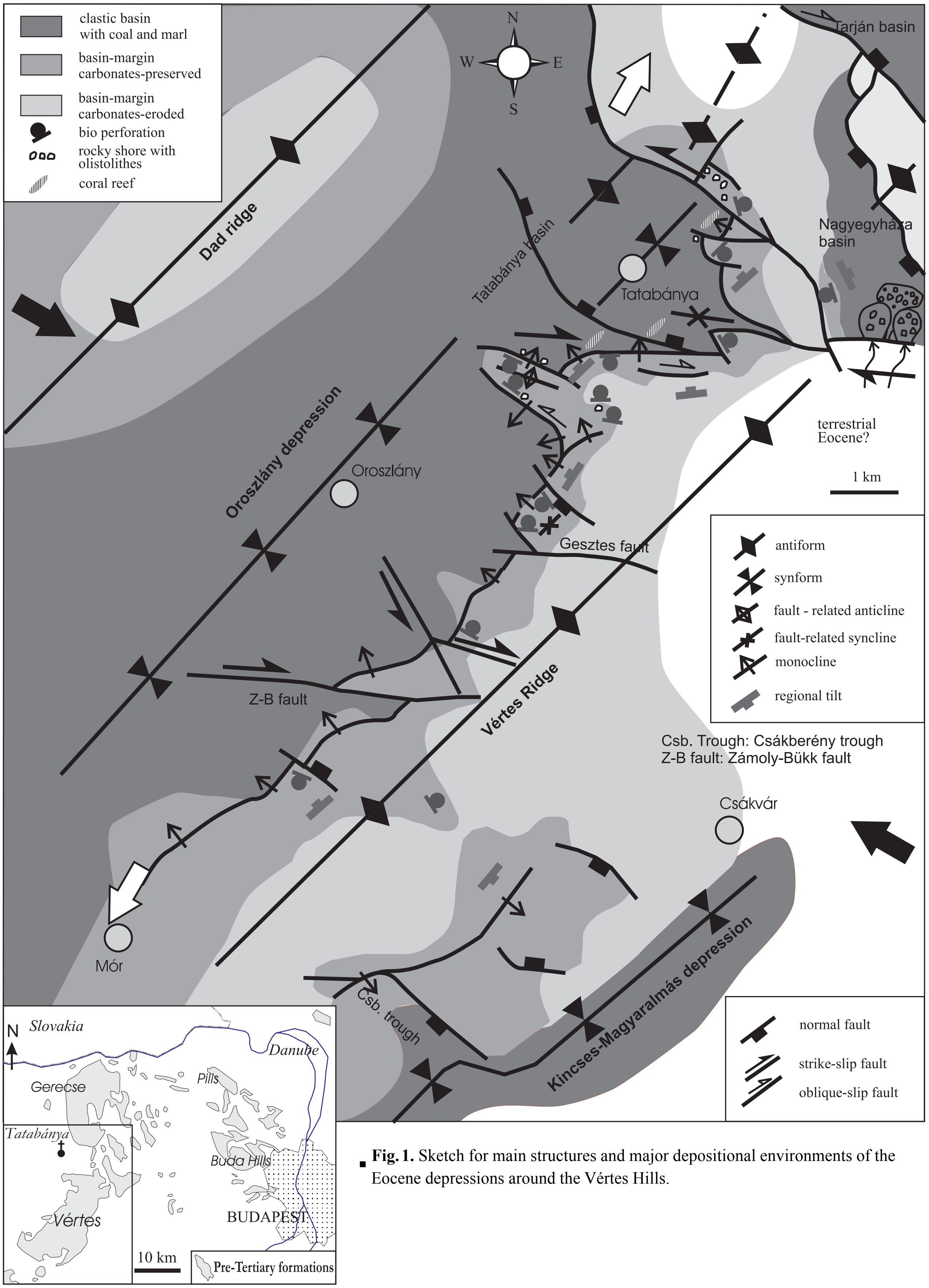

Fig. 1. Sketch for main structures and major depositional environments of the Eocene depressions around the Vértes Hills.

steepened carbonate ramp which developed rimmed carbonate platform on the northern part of the Vértes ridge. Coral reefs were grown on the platform margin dividing the fore and back reef facies.

Syn-sedimentary structures, and the bioperforated fault planes with striae permitted the approximation of middle Eocene stress field. The compression was oriented (W)NW-(E)SE, while the tension was perpendicular. The compression is perpendicular to the general trend of the local paleo-topographic features and might have induced gentle folding of the pre-Tertiary basement. Elevated ridges (antiforms) were colonised by carbonate-producing organism, and carbonate ramps formed on their fringes, along NE-striking monoclines. Depressions (synforms) were covered by slightly deeper water and trapped fine-grained siliciclastic detritus. The orientation some of the normal faults and sedimentary dykes were perpendicular to the compressional direction in the early stage of the middle Eocene tectonic processes. These structures are due to local upwarping and bending of the pre-Tertiary basement during the early stage of folding.

The observations are in agreement with the model of Tari et al (1993) about the compressional (retroarc) origin of the basin. Thickness difference may suggest that the Vértes antiform was slightly asymmetric and had a very minor SE vergency. Such suspected asymmetry is part of the model of Tari et al. (1993) and was documented in the neighbouring Buda Hills (Fodor et al. 1992). On the other hand, the local structural geometry is more complex than a single reverse fault or monocline. The E-W to NW-SE trending strike-slip and normal faults cross-cut the antiforms and seem to be more important in the localisation of the sediment traps. Alternatively, they represent structures post-dating an early phase of folding.

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Do You Separate Sets of Reactivated Faults Manually?

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Faults are brittle structures which are formed as the response to stress in the upper crust. Due to variation of the tectonic stress field in time, faults are frequently reactivated in subsequent stress phases. The fault-slip data set is heterogeneous if the slips recorded on the fault planes occured in multiple tectonic phases with different stress fields. The principle of inverse method in paleostress analysis is to find an optimal stress tensor by using fault-slip data measured in deformed rocks. But to use the inverse technique a homogeneous fault-slip data set is required, which means a group of faults activated only in one specific tectonic phase. The stress calculated by applying the inverse method on heterogeneous data does not characterize the real stress situation.

A new computer program was made to identify individual paleostress phases and to determine paleostress tensors from heterogeneous fault-slip data. The possible stress tensor solutions are calculated from the orientation of the fault planes and from the striations and sense of slip. The fault-slip data are combined into four-element groups and the reduced tensor (e.g., shape and orientation of the stress ellipsoid) is calculated for each group. Group with four homogeneous fault-slip data provides the true results wich characterize the real paleostress conditions. The stress tensor calculated for heterogeneous four-fault group is not reliable. These results were visualized using lower hemisphere equal-area projection in which these true and false results can be easily distinguished. Projections of directions calculated from heterogeneous data sets – false results – are dispersed whereas the true results obtained from homogeneous data are grouped in clusters. In case of large number of fault-slip data analysed, the computer program is needed to identify the density of solutions. Density maximum indicates some of possible directions of considered principal stress. The number of such clusters represents a number of paleostress phases. The introduced software has useful application in the study of striated faults. Development of the program was supported by the grant project GA AVČR IAA3013406.

Do We Have a Remnant of the Hanseatic Terrane and/or Rheno-Hercynian Ocean in the Western Carpathians? – A Case Study from the Devonian of the Považský Inovec Mts.

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Geodynamic evolution understanding in orogenic belts like the Alps, Carpathians or Himalayas that contain multistage metamorphic and magmatic episodes is often problematic. The polyorogenetic history of such orogenic belts, marked by incorporation of pre-Mesozoic polycrystalline basement blocks into young Alpine structures, resulted in formation of complicate rocks puzzle often characterised by juxtaposition of various terranes and/or blocks due to multistage tectonic evolution by large-scale nappe and strike-slip tectonics. However, unravelling orogenic episodes in the modern polyorogenetic belts is practically impossible without precise stratification and dating, as lithological and structural relations within various fragments are frequently ambiguous. The European Hercynian (Variscan) and Alpine mountain chains are typical collisional orogens, and are built up of pre-Hercynian basement blocks that, in most cases, originated at the north Gondwana margin. Such pre-Hercynian elements were part of a pre-Ordovician continental ribbon – the Hun superterrane in the former eastern prolongation of Avalonia, and their present-day distribution resulted from juxtaposition through Hercynian and/or Alpine tectonic evolution (Stampfli and Borel 2002, von Raumer et al. 2003). The Devonian was a period of relative silence in the Earth history between vanishing Caledonian movement in the Lower Devonian and beginning of Hercynian (Variscan) orogenesis in the Upper Devonian. Thick terrigeneous accumulations of so-called Old red sandstone, huge marine carbonatic and flysch sediments, as well as extensive products of submarine basic and/or bimodal volcanism represent the rocks record of this period. Sedimentary record has in European realm general feature of changing facies from terrigeneous clastic material at the north (Old red continent evolution) through a mixture of psammitic-pelitic depositions and/or neritic-pelagic interchange (Rheno-Hercynian evolution) to calcareous sedimentation with pelitic intercalation (Bohemian – Barrandian's evolution) at the southern margin.

Devonian in the Western Carpathians (WC) has not large area extension in general. There are known only the Gelnica and Rakovec Groups in the Gemeric unit that consist of metagreywackes, phyllites, lydites, carbonates and basic volcanics, the Harmonia Group in the Malé Karpaty Mts. (Tatric unit) with similar metamorphosed rocks (phyllites, greywackes, limestones and basic volcanics), and the Predná hola sedimentary-volcanogenic complex (Veporic Unit). Devonian limestones were sporadically described from deep boreholes at the southern Slovakia. Recently in the frame of construction new geological map of the Považský Inovec Mts., there was documented an unusual volcano-sedimentary complex for the Tatric unit of the WC. This complex was displayed in the official General map of Slovakian territory as amphibolites (Kamenický 1956, Kamenický in Buday et al. 1962). However, our field and petrological study proved that the dominant part of this complex consists of dark grey fine-grained laminar to weakly banded pelitic-psammitic metamorphosed rocks – metagreywackes and phyllites. There were identified locally sills of submarine basic volcanics – amphibolites and/or its pyroclastic analogues, layers of black schists respectively graphitic metaquartzites and lydites, as well as calk-silicate hornfels – erlans and whole complex was called as Hlavinka volcano-sedimentary metamorphic complex (Kohút et al. 2005). The most common rocks of this complex – metagreywackes and phyllites are composed by quartz, plagioclase, K-feldspar, biotite and organic matter (graphite), in accessory content are present garnet, zircon and monazite. Metamorphic overprint of original volcano-sedimentary sequence reach to upper part of greenschist facies respectively lower part of amphibolite facies with T = 500 to 550 °C and P = 300–350 MPa (Kohút and Siman 2005). Due to apparent dominance of amphibolites there was omitted stratigraphic determination indicated Devonian age of palynomorphs, tracheids and phyto detritus (Čorná and Kamenický 1976) separated from black schists, and this part of the Tatric crystalline was regarded as "deep Lower Paleozoic basement" till present. However, crucial for the Devonian classification was recent discovery of hematite metaquartzites – a typical analogue of the Lahn-Dill volcano-sedimentary iron ores (Kohút and Havrila 2006) within the Hlavinka Group. Geochemistry confirmed greywacke protolith character of prominent metamorphic rocks (metagreywackes and phyllites) from the Hlavinka Group, whereas these rocks were sedimented at continental slope in the back-arc basin. These greywacke were derivate from an acid and/or intermediate magmatic rocks source that originated in an active continental arc. Rather unusual MORB geochemical character of metabasic rocks – amphibolites was shown as standard for Rheno-Hercynian evolution of the Devonian (Floyd 1995). Relative lack of modern stratigraphic data from Hlavinka Group partially supply dating of uraninite and monazite with the electron microprobe (CAMECA SX-100) in an attempt to broadly constrain formation ages of greywackes and hematite metaquartzites. The uraninite origin (390–380 Ma) was the most probably synchronous to formation of submarine-exhalation iron ores, whereas monazite data (350–330 Ma) from identical samples indicate rather final Meso-Hercynian metamorphic overprint of volcano-sedimentary pile (Kohút et al. in preparation).

The Carpathians form part of an extensive, equatorial, orogenic belt extending from Morocco in the Atlas Mountains, through the Alps, Dinarides, Pontides, Zagros, Hindukush to the Himalayas and to China. The Western Carpathians are the northernmost, E–W trending branch of this Alpine belt, linked to the Eastern Alps in the west and to the Eastern Carpathians in the east. Nowadays verification of Devonian in the Považský Inovec Mts., call for some geodynamic questions and/or analogy not only within Alpine – Carpathians realm, but in the frame of whole Central European basement areas. Sedimentological or litho-facial evolution of Hlavinka Group – dominance of clastic terrigeneous psammitic-pelitic material (greywackes and schist), small contribution of organic matter (black schists) with minimal limestone intercalations and typical basic volcanism causing Lahn-Dill iron ores (hematite metaquartzites) clearly affined with Rheno-Hercynian evolution of the European Variscan mobile zone. One can observe an analogous evolution today from the Lizard complex in Cornwall, through the Ardennes, the Rhenish Massif, the Harz Mts., northern part of Bohemian Massif (BM) to the easternmost part of BM – Moravo-Silesian zone (Jeseník Mts., and Drahany Upland). The Devonian rock sequences in the frame of the Alps are well preserved in the Austro-Alpine realm, mainly in the Grauwacken Zone and Graz Paleozoic. Both occurrences form a part of the Ordovician – Carboniferous extensive volcano-sedimentary complexes with dominative calcareous and pelitic sedimentation during Devonian period, showing thus an affinity to Bohemian – Barrandian's evolution, although some faunal indications display strong Rheno-Hercynian similarity (Schönlaub 1995). Since Kossmat (1927) it is known that an elemental part of the Devonian basin remnants form Rhenohercynian zone of the European Variscides. The Rheno-Hercynian ocean was opened as a Devonian oceanic domain within the southern Laurussia margin, due to Gondwana-directed slab pull, and was situated between the southern margin of Laurussia (Avalonia) and Hanseatic terrane (Stampfli and Borel 2002, von Raumer et al. 2003). Due to Middle Devonian collision become weak Hanseatic block part of Hunic superterrane which collided during main Meso-Hercynian period (Visean) with Laurussia caused widespread granitization in whole Central European realm. It is evident now that Rheno-Hercynian ocean must have a continuation from the Moravo-Silesian zone to the Austro-Alpine domain and through the Western Carpathians join the Dobrogea suture in Romania and/or Moesia (Stampfli, personal communication). It is generally accepted that Hercynian basement of the Western Carpathians formed part of Hunic superterrane, however our study proved that before the docking of the Hunic terrane against Laurussia, part of the Carpathians pre-Hercynian basement recorded history of Rheno-Hercynian ocean and/or Hanseatic terrane, showing that affinity to Avalonia origin.

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Paleogene-Early Miocene Igneous Rocks and Geodynamics of the Alpine-Carpathian-Pannonian-Dinaric Region: an Integrated Approach

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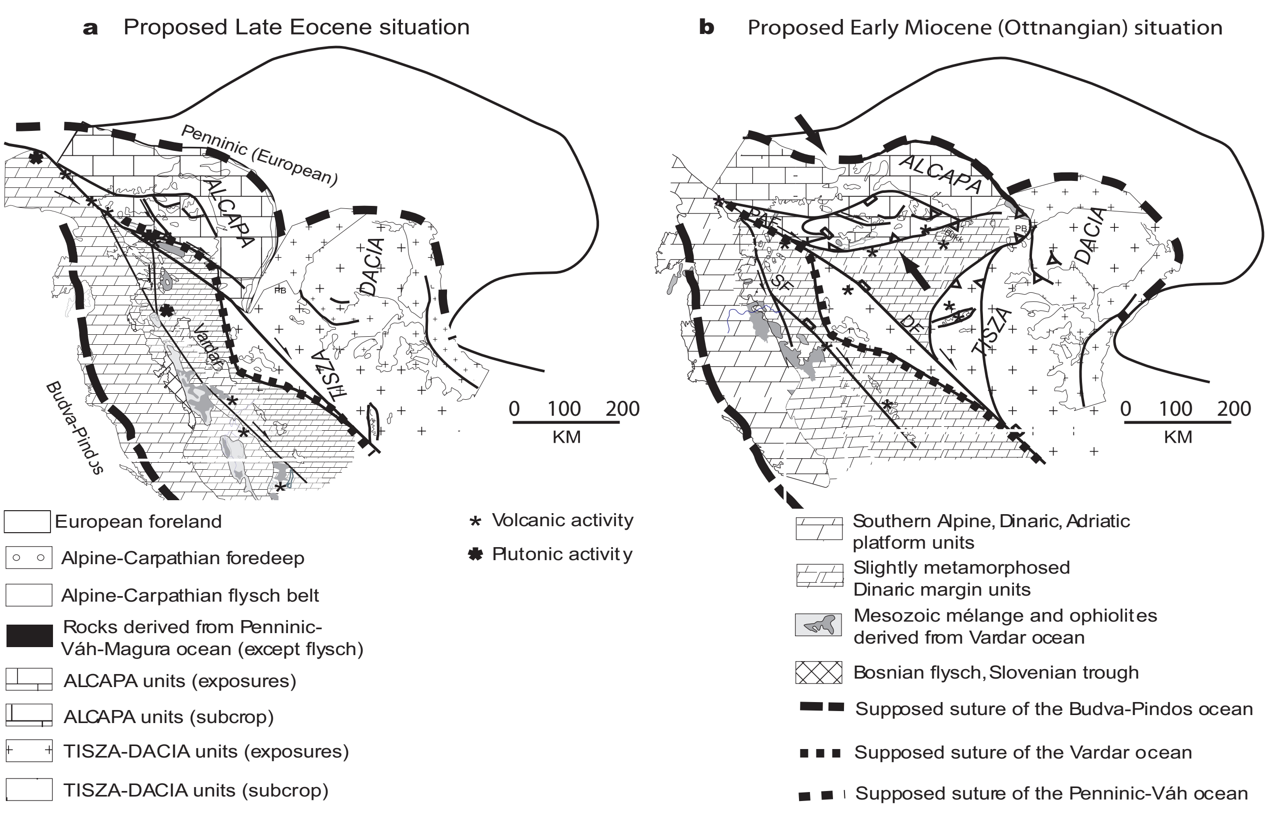
We attempt to reveal the geodynamic link between Paleogene-Early Miocene igneous rocks of the Mid-Hungarian zone and those of the Alps and Dinarides. Our summary suggests that the Paleogene-Early Miocene igneous suite in all studied igneous provinces of the Alps, Carpathians, Pannonian Basin and Dinarides was formed in the same time interval, with three peak episodes of magmatic activity in the Eocene, Late-Eocene-Early Oligocene and Late Oligocene-Early Miocene. The studied igneous rocks have similar geochemistry and petrology, which shows a subduction-related character. The magmatic belt along the Periadriatic zone can be followed along the Balaton fault, on the northern margin of the Mid-Hungarian zone to the Bükk Mts. (N Hungary). This continuity is supported by well-correlated Mesozoic and Paleogene-Early Miocene sedimentary sequence assemblages, forming the basement of these magmatic bodies (Fig. 1). On the other hand, these North Hungarian country rocks have their counterparts in the internal zone of the Dinarides, along the Sava-Vardar Zone, that also host similar magmatic rocks. We, therefore, suggest that all the Paleogene-Early Miocene magmatic rocks of the studied region are closely related and have a common, subduction-related origin (Fig. 1).

The study also highlights orthopyroxene-rich websterite mantle xenoliths from Western Hungary (Bakony-Balaton Highland) and East Serbia that was formed in the vicinity of a subducted slab. These orthopyroxene-rich websterite mantle xenoliths have common petrographic and geochemical signatures.



■ **Fig. 1.** Tectonic reconstruction of the Alpine-Carpathian-Pannonian-Dinaric region.

These mantle rocks must have formed close to a subduction in a fore-arc setting. The closest subduction scar to both occurrences is that of the Vardar Ocean.

We discuss the location and polarity of all potential subduction zones of the area that may account for the igneous rocks and orthopyroxene-rich mantle rocks. Results of seismic tomography on subducted slabs beneath the studied area combined with geological data demonstrate that igneous rocks and mantle rocks may not be explained by the same subduction process. Instead, we propose that the western portion of igneous rocks in the Periadriatic zone is related to the Penninic subduction, whereas most of their Paleogene-Early Miocene counterparts in the Mid-Hungarian and Sava-Vardar zone could have originated from the Budva Pindos subduction. The most likely solution is that these oppositely dipping and synchronous subductions relayed each other and accommodated together the Europe-Africa convergence during the Paleogene.

The present diverging shape of the proposed arc has been achieved by considerable shear and rotations of major continental blocks. The Paleogene magmatic belt was strongly affected by lithospheric scale, arc-parallel, right lateral strike slip shear. This shear was initiated in the Late Eocene but maximum motion was achieved during Oligocene. The fault-induced pervasive fracturing could have localized magmatic activity, and vice versa, the heat impulse of the magmatic activity could have rheologically softened the country rocks and rendered them more easily deformable. This second case is strongly suggested for the Mid-Hungarian zone, which must have experienced highly intense deformations during the Early Miocene rotations (Fig. 1).

The occurrence of Late Oligocene-Early Miocene and also a part of the Middle Miocene-subrecent igneous rocks along the Mid-Hungarian zone in the Pannonian Basin match with those of the Late Eocene-Early Oligocene igneous rocks. We speculate that the genesis and ascent of Late Oligocene-Early Miocene magmas was initiated when the rotations took place. Thus, the rotated blocks could have also brought their earlier metasomatized lower lithosphere into the Carpathian embayment. The large volume of Early Miocene magmatic rocks associated with the Mid-Hungarian zone also suggests that the ascent of these magmas was controlled or facilitated by the most deformed part of this structural zone. Furthermore, a part of Middle Miocene igneous rocks close to the Paleogene-Early Miocene igneous rocks might have had the same subduction-related mantle source that had been transported to its present position from the former Paleogene arc. Consequently, it is proposed here that the same enriched mantle source could be reactivated several times during its geodynamic evolution providing different styles of volcanics depending on the process responsible for magma generation.

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Depositional Systems and Lithofacies of the Zlín Formation near the Contact between the Bystrica and Rača Units (Magura Nappe, Outer Carpathians, Eastern Slovakia)

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Magura basin has a complex tectono-sedimentary history with several phases of its evolution. Tectonic activity and sea-level changes controlled the sedimentary supply to the basin and induced formation of individual depositional elements and their migration in time and space.

Geological mapping (Žec et al. 2005) and detail sedimentological analysis on selected, relatively well exposed profiles revealed the significant differences between sedimentary character of Zlín Formation in Bystrica and Rača Units near their contact. Studied area is situated between Oľka and Laborec river valleys in the eastern Slovakia. Stratigraphic range of the Zlín Formation in the area is from the upper part of the Middle Eocene up to the Late Eocene. The formation is developed in overlayer of the Early to Middle Eocene Beloveža Formation which sedimentary conditions were affected by the subsidence and sea-level rise. During the uppermost Middle Eocene and Late Eocene, a significant uplift in the Magura basin was recorded (e.g. Oszczypko et al. 2003). This event influenced the sedimentary conditions in the area and triggered origin of depositional elements of the Zlín Formation.

The Zlín Formation of Bystrica Unit is characterized by higher sand/mud ratio (usually > 2) and the coarse-grained lithofacies resemble to those from Krynica unit (Strihovce Formation sandstones). Medium- to coarse-grained sandstones (lithofacies B1.1 sensu Pickering et al. 1986) are laterally continuous and often amalgamated. Grading is absent or poorly developed (coarse-tail grading), in some cases fluid-escape structures are preserved. Lithofacies B1.1 originated by the deposition from quasi-steady concentrated density flows (Mulder and Alexander 2001) and is alternating with the thin- to thick-bedded, normaly graded fine-grained sandstones representing turbidites sensu stricto with well preserved Bouma´s intervals (lithofacies C2.1, C2.2, C2.3 sensu Pickering et al. 1986). Tabular geometry, great lateral extent and lack of channelization suggest deposition in lobe and interlobe environments commonly interpreted to be diagnostic of an outer submarine fan (Mutti and Normark 1987).

Sedimentary fill of the Zlín Formation in Rača Unit (near the contact with Bytrica Unit) has a different character. Thick (up to 50–100 m) mudstone dominated horizons are alternating with several metres or tens of metres thick sandstone packages. Mudstone horizons are characterized by low sand/mud ratio (<0.5). Thin to very thick fine-grained sandstone beds are overlain by very thick mudstone drapes (up to 10–15 m, lithofacies C2.4 sensu Pickering et al., 1986). These sandstone/mudstone couplets probably originated by ponding of huge turbidite flows, in which the mud component of the flow was retained within a tectonically restricted depocenter. Paleoflow direction inferred from sole structures are usually oriented from SE to NW. However, some ripple and dune orientations indicate flow direction at a high angle to that deduced from associated sole structures. These different directions are caused by reflection off containing slopes (e.g. Haughton 1994). We suppose the slopes were paralel with basin axis (NW-SE trend) and the gravity flows dispersed from the basin margins were forced by basin topography to flow along the axis (longitudal filling).

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Depositional Environments and Biostratigraphy of the Lower Part of Rača Unit Paleogene (Magura Nappe, Outer Carpathians, Eastern Slovakia)

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We have studied sedimentology and biostratigraphy of the lower part of Rača Unit sedimentary infill at several localities in the eastern Slovakia – near Mrázovce village in the southern part, and in Vyšný Komárnik village and Dolhonec valley in the northern part of the the unit. Near Mrázovce village, the upward-fining and thinning bed succession was interpreted in more then 220 m long, well exposed profile from the basal sandstone-conglomerate horizon to overlying thin-bedded and fine-grained lithofacies of the Beloveža Formation (Kováčik and Bóna 2005).

In the lower part of profile the coarse-grained lithofacies, deposited by concentrated density (gravity) flows (sensu Mulder and Alexander 2001) in submarine channels, are presented. Cobble to pebble conglomerates, coarse-grained to granule sandstones are thick- to very-thick-bedded, massive (lithofacies A1.1, A1.4, B1.1 sensu Pickering et al., 1986), graded (lithofacies A2.2, A2.7) or partially stratified (lithofacies A2.8). The thick-bedded sandstones are rich in intraclasts of grey calcareous mudstones containing mixed foraminifera fauna (plankton>>benthos), ostracods, and inoceramid prisms. Benthos is mostly calcareous. Plankton with *Globigerinelloides subcarinatus*, *Gansserina wiedenmayeri*, and *Globotruncanella petaloidea* evidences the Maastrichtian age. Fauna is of "Frydek-type" biofacies. The mudstones were originally deposited under the well-oxygenated outer-shelf settings and later eroded and transported by gravity flows to the site of deposition (base of slope?).

Towards the top (in the middle part of the profile) the finer-grained, thin- to medium bedded turbidites (lithofacies C2.3, C2.2.) gradually prevail above the coarse-grained lithofacies having a more organized character with well preserved Bouma's intervals, sole casts and positive grading. Fine-grained sandstones are often horizontally, ripple-cross or convolute laminated and alternate with grey mudstones. Lamination is usually emphasized by the plant and mica detritus. Well preserved flute and groove casts prove the flows directions prevailingly from W or NW to E or SE. The lithofacies of this horizon were deposited in the transitional zones between channels and overbanks or directly in overbanks.

The uppermost part of profile is represented by the Beloveža Formation built by very thin- to thin-bedded, fine-grained turbidites and hemipelagites. The beds of variegated (red brown, green grey) noncalcareous mudstones represent the condensed horizon. These mudstones contain agglutinated foraminifera fauna dominated by the tubular astrorhizids (*Nothia* sp., *"Rhizammina"* sp.) accompanied by abundant *Glomospira charoides*, *Hyperammina nuda* and *Ammodiscus planus*. No stratigraphically significant taxa were observed among agglutinated species. Single specimen of *Subbotina crociapertura* (?reworked) indicates the early Middle Eocene age. Paleoenvironment can be characterized as oligotrophic, well-oxygenated lower slope or basin plain below the CCD. Oligotrophic conditions are also proved by abundant findings of trace fossils (*Chondrites*, *Paleodictyon*, *Scolicia*, *Helminthopsis*) which is a widespread phenomenon related to global warming in the late Paleocene to middle Eocene (e.g. Uchman 2004)

In variegated shales from both Vyšný Komárnik and Dolhonec localities the solely agglutinated foraminifera fauna dominated by tubular astrorhizids (*Nothia* sp., *"Rhizammina"* sp.) was found. Rare *Saccaminoides carpathicus* evidences the Early Eocene age for both localities. Agglutinated foraminifer fauna indicates the lower slope depths below the CCD. Dominant *"Rhizammina"* sp. and abundant radiolarians from Vyšný Komárnik may indicate eutrophic conditions. Moreover, the finely pyritized radiolarians indicate oxygen-minimum zone in the water column according to the taphonomic interpretation of Bąk (2000). However, the deposition of variegated mudstones lasted until the Middle Eocene, as was recently proved from another localities near Vyšný Komárnik (Kender et al. 2005).

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Temporal Investigations and Retrograde Metamorphism of the North-Eastern Part of the Bohemian Massif

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Indentantion of a lithospheric Brunia continent with the Moldanubian orogenic root produced a crustal wedge, which shows following metamorphic zonality from the east to the west: chlorite-biotite, garnet, chloritoide, and staurolite in the easterly par-autochthon (the Desná dome), staurolite-sillimanite-andalusite in the deeper part and staurolite and garnet at the upper part of the westerly lower allochthon (the Keprník nappe) and kyanite zone in metapelites and eclogite boudins in the westernmost upper allochthon (the Velké Vrbno unit). The structural mapping distinguished fabrics related to burial, reworked by transpressional deformation and folding and finaly by heterogeneous extensional deformation associated with voluminous magmatism. Th-U-Pb dating on monazites provides information on the prograde and retorgrade parts of PT evolution while the closure of the K-Ar isotopic system in muscovite and biotite allows determining the time when the rock passes through the isotherms of about 360 and 320 °C, respectively. Four micaschist samples collected from chlorite-biotite zone (300–400 °C) at the eastern border of the Desná dome yield the K-Ar ages of muscovite that provides the age of metamorphic peak (from 320 ± 4.7 to 343

±5.1 Ma). Samples from the western part of the Desná dome (the staurolite zone), shows cooling ages on muscovites from 260 to 302 ±4.4 Ma. In the Keprník nappe (staurolite-sillimanite zone) K-Ar method yields ages on muscovites from 285 to 300 ±4.5 Ma while in the western part affected by syn-extensional magmatism the youngest age was depicted (300 ±4.5 Ma, muscovite). Micaschists from the Velké Vrbno unit provide information about the earliest increments of synconvergent exhumation (331 ± 4.5 Ma, on biotite). Samples for monazite dating cover all units of wedge and inclusions of monazite in garnet and monazites in matrix were measured separately. In the Desná dome and in the Keprník nappe, both types of monazites yield similar ages ranging 250 Ma to 297 ±30 Ma. However, the samples from the Velké Vrbno unit exhibit two generations of monazites: in the garnet porphyroblasts the average ages correspond to 340 ±30 Ma, while in the matrix average age of 302 ±30 Ma was depicted. In order to correlate the K-Ar ages with the exhumational P-T path, the conditions of trapping of fluid inclusions in quartz and quartz –andalusite bearing tensional gashes were studied. In andalusite are identified ambigeous primary inclusions, while in quartz are observed primary and secondary inclusions. Primary incluisons of quartz were trapped from a homogeneous fluid phase. Interval of trapping is 280 to 380 °C at 2–2.8 kbar and consequently the ages of exhumation 282–293 Ma determined from K-Ar dating on micas are related to this event. In conclusion, the burial of continental crust occurred around 340 Ma, followed by synconvergent exhumation at 330 Ma and the up-doming of the crustal wedge accompanied with intrusion of large scale granite and final extensional collapse at about 300 Ma.

Tectonic Position of the Latest Triassic–Jurassic Sequences of Rudabánya Hills, NE Hungary – The First Steps in a Puzzle

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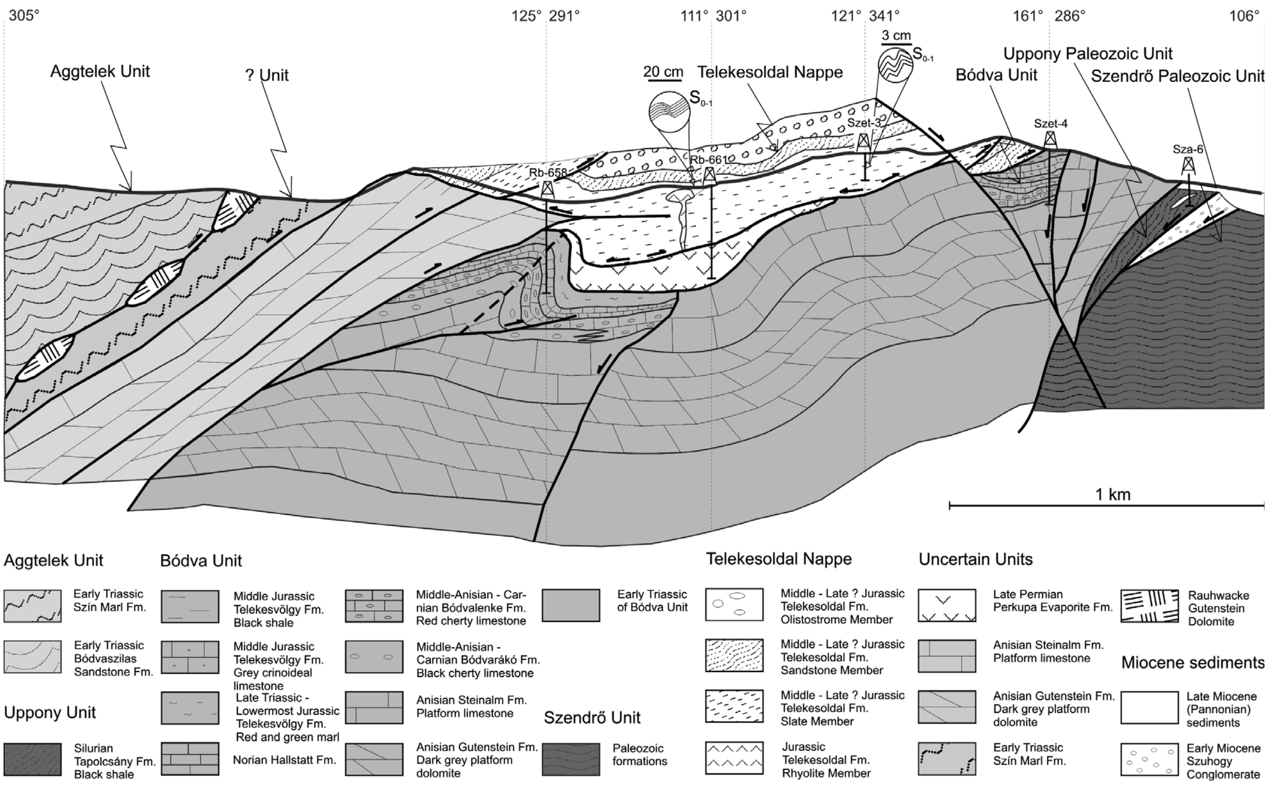
The Jurassic rocks of the Rudabánya Hills have been studied since the middle of the 19th century. However, our knowledge on their tectonic position, the number of the tectonic units, the history of their deformation, their exact age, their Triassic basement, and anyway, the correct order of the formations are not satisfying, yet. On the other hand, the Jurassic age of these formations was generally accepted only about 15 years ago, based on the works of Grill, Kozur and Dosztály.

The previous researchers divided the Jurassic rocks into two sequences by right of lithological and paleontological data (Grill and Kozur 1986, Dosztály 1994). The Telekesvölgy Formation (TV Fm.) starts with red and green clay marl, containing limestone olistholites and redeposited beds . This sub-unit likely has latest Triassic age, according to the Conodont fauna of the olistholites and redeposited limestone beds. Although the contact or transition is not clearly outcropping, the green and red marl is followed by grey crinoideal limestone and marl with no exact age, and finally, by black shale, with Bajocian – Bathonian radiolarians (Grill 1988, Dosztály 1994).

The Telekesoldal Formation (TO Fm.) contains silicified slate and marl with subvolcanic rhyolite bodies, black shale with sandstone olistholites (reinterpreted here as sandstone turbiditic layers) and a sub-unit of olistostromes with varying clast composition (limestones, rhyolite). From the sedimentary rocks, the only age we have, is Bajocian by right of the radiolarian fauna of the lowermost slate–marl member (Grill and Kozur 1986, Dosztály 1994). Radiometric age of the rhyolite was estimated with Rb/Sr (158 ±34 Ma) and K/Ar (120 ±6 Ma) methods (Grill 1988). The first method gives a wide age interval for magma intrusion; the other one may reflect the Early Cretaceous metamorphic event thus the exact formation age of the rhyolite remains poorly constrained.

In our work we made field works and measurements, thin sections, illite Kübler index measurements and radiolarian investigations, and re-examined several borehole material. The rocks of TO Fm. have a bedding-parallel foliation (S_1); in the olisthostroms foliation is connected to strong layer-perpendicular shortening (flattening of clasts). Rarely the bedding and the foliation intersect each other at an oblique angle. In this case, the foliation is an axial plane cleavage (S_2), connected to closed folds. The bedding-parallel foliation was overprinted by a folding phase, resulting in small-scale kink folds (F_3), at the transition of brittle-ductile deformation field. The effects of such ductile phases cannot be determined neither on the members of the TV Fm., nor on the Triassic rocks of Bódva series (Middle Triassic platform carbonate, red, basin facies limestones and chert from Middle Anisian to Middle Norian).

According to the radiolarian investigations, and the illite Kübler index data, measured on the rocks of TV Formation correspond to the diagenetic zone (Árkai 1982, Árkai and Kovács 1986), and contains poor Middle - Jurassic (Bajocian – Upper Bathonian) radiolarian fauna with the dominance of Nassellarians (deeper water) (Do-



■ **Fig. 1.** Geological cross section of the central part of Rudabánya Hills, NE Hungary.

sztály 1994, Ozsvárt pers. comm.). In contrast with it, the rocks of TO Formation suffered ductile deformation, in most cases show anchimetamorphic illite Kübler index values (Árkai 1982, Árkai and Kovács 1986), contains Bajocian radiolarian with the dominance of Spumellarians (shallower water) (Dosztály 1994, Ozsvárt).

We specially investigated the contact of the carbonate-dominated and clastic-dominated latest Triassic to Jurassic formations, and the relationship of the two (latest Triassic–) Jurassic formations. In some borehole (P-74, Szet-4, Rb-658) we found, that the transition between the Hallstatt Limestone of the Bódva series and the red and green marl of the TV Formation is continuous, so we have indication that the Bódva series being the original Triassic basement of the TV sequence. The illite Kübler index (IK) data, measured so far (Árkai 1982, Árkai and Kovács 1986, and this study) confirm it, because both Triassic and Jurassic sequences have only diagenetic IK values. The relation between the Bódva-type Triassic and TO Jurassic is always tectonic – both in field and boreholes – and there is a considerable difference not only in the metamorphic grade, but also in the style of deformation. So the Jurassic TO Fm. cannot be related to the Bódva-type Triassic sequence, in contrast to previous interpretation (Grill 1988, Less et al. 1988). Superposition of the two latest Triassic–Jurassic formations was in fact penetrated by the borehole Rb-658, in which the TO Fm. is in upper position. The reinterpretation of Szet-4 borehole and its surroundings suggests their superposition, too (Fig. 1.).

We suggest that the TV Formation is the original Jurassic cover of the Triassic carbonate rocks of the Bódva series. The TO Formation is supposed to be a nappe, overlying the Bódva nappe (including TV Fm.). Because the upper "Telekesoldal nappe" is very probably slightly metamorphic, this juxtaposition would only be possible after metamorphism. This overthrust supposed to be a relatively early event, and was followed by several smaller-scale thrusts. We have not direct field data on the vergence of the main nappe emplacement so far. The most frequent dip directions of the formations are W or NW, so the transport direction can be from E to W (hinterland-dipping) or from W to E (foreland-dipping). The vergence of the younger reverse faults and fault propagation fold is to E or SE. This event can be connected to the deformation of the Alsótelekes gypsum-anhydrite body, located somewhat to the S (Zelenka et al. 2005).

There are normal faults among the observed brittle structures, too. Part of them supposed to be active even in the Middle and Late Triassic, causing considerable thickness variations of basinal formations. They can be more than 100 m thick, but in some outcrop it appears only as neptunian dykes in the underlying platform limestone (Steinalm Fm.) and as a few condensed beds. Some of them are reactivated during the Miocene.

According to our investigations and geological sections constructed (Fig. 1.) so far, the following deformation phases can be detected. The first S_{0-1} foliation of the TO Fm. was most probably formed due to deep tectonic burial (e.g. thrusting) under anchimetamorphic conditions, representing the first D_1 event. The S_2 axial plane cleavage of the D_2 folding phase can be observed only in few outcrops. The last phase of ductile deformation (D_3) is represented by kink folds (F_3), can be seen in the outcrops and borehole materials of the silicified slate of TO Fm. This kind of deformation could occur at shallower crust depth, at the transition of brittle-ductile deformation field. The influence of these ductile phases cannot be observed on the rocks of Bódva series (including TV Fm.). During the D_4 deformation event the anchimetamorphic „Telekesoldal nappe” thrust over the Bódva nappe. In the D_5 phase, characterized by NW-SE compression, SE verging reverse faults and fault-propagation folds were formed (Fig. 1.). The alternating dip values measured on the rocks of Bódva series and those of TO Fm. are probably the reason of a late folding phase (F_4), characterized by open folds with long wavelength. The semi-vertical dipping of the Late Triassic basinal limestones can be formed by movements along SE verging fault-propagation folds (F_5). The ongoing NW-SE compression resulted in SE verging thrusts. Among them, an uncertain unit of Gutenstein Dolomite, Steinalm Limestone and Early Triassic marl thrust upon the Bódva Unit. During this thrust the ramp fault might have not reach the surface, but connected to roof thrust of duplexes. The juxtaposition of Aggtelek Unit and Bódva Unit can be related to this phase, but it is more likely to be an older structure. Younger transpressive strike-slip and normal fault movements (D_6–D_7), connected to Darnó Zone, juxtaposed the Mezosoic formations of Rudabánya Hills and the Paleozoic rocks of Uppony and Szendrő Hills. Parts of these movements are Tertiary in age, indicated by the involved Szuhogy Conglomerate and Pannonian sediments (Szentpétery 1997). This model can be extended to the major part of Rudabánya Hills, because it has a great similarity to the previous investigations made in other part of the Rudabánya Hills (Fodor and Koroknai 2000, Kövér et al 2005)

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Multiple Magmatic Fabrics in Episodically Emplaced Granites in Transtensional Setting: Tectonic Model Based on AMS Study and Numerical Modeling

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The three successive sigmoidal sheet-like granite intrusions (Thannenkirch, Brézouard, Bilstein granites – BBT Complex) in the Central Vosges Mts. (France) separates medium to high grade (~700 to 800 °C, >9 kbar) gneiss and granulites to the north from low-pressure (~700 °C, ~4 kbar) anatectic migmatites to the south. The entirely compressional fabrics in the northern gneiss contrast with the pervasive extensional deformation in the south. This different structural record reflects the latest deformation event in the south while the relics of first compressional steep foliations are preserved in both regions. In addition, the entirely discordant northern contacts of BBTC with respect to the host rock foliations contrast with the southern intrusion margins that show fabrics perfectly coherent with the host rock. The BBTC intrusions show systematic transition from the dominant magmatic fabric in their northern and central parts to the MT subsolidus deformation terminating in LT mylonites in the south. The latter fabrics are developed in conjunction with E-W trending foliation and subhorizontal stretching lineation. The progressive decrease of deformation temperature is confirmed by quartz c-axis fabric patterns that suggest transition from the activity of prism <a> slip system towards the rhomb <a+c> and the basal <a> slip systems. AMS study reveals bimodal fabric pattern with the central-northern margins showing NW-SE trending foliations and lineations, low intensity (P parameter) and the southern parts with steep E-W trending magnetic foliation, horizontal lineation and high intensity (P parameter). The AMS within the central and northern parts is consistent with the AMS fabrics in southern migmatites. Telescoped ^{40}Ar-^{39}Ar cooling and U-Pb crystallization ages (~328–325 Ma) of BBTC and migmatites in the south proved that the exhumation occurred during a short period of time and that the intrusions of granitoids were coeval with the ductile thinning of southern domain. In contrast, the granulites to the north show cooling path related to compressional exhumation (~335–330 Ma) followed by reheating (~325 Ma) during intrusion of northern granite sheet (Thannenkirch pluton) of the BBTC. Based on our structural study, we suggest that the preexisting E-W trending compressive fabrics structurally controlled the distribution and the emplacement of the granitic magmas. The SSW-NNE extensional traction operated along steep mechanical anisotropy at high angles which generated oblique transtensional regime. The internal fabric within individual plutons is therefore interpreted in terms of partitioning of transtensional deformation, with pure shear dominated area in their northern and central parts and wrench dominated domains along their southern margins. To asses realistically the obtained AMS pattern we compare two numerical models of AMS fabrics in transtension with respect to originally isotropic and pre-deformational intrusion-related fabrics. It is the latter model which returns more realistical fabric data. The asymmetrical microstructural and geochronology patterns are further discussed using thermal 1D modeling which indicates the sequences of individual intrusion from north to south and their mutual thermal interferences leading to successive reheating of southern margins of northerly intrusions that can accommodate prolongated viscous deformation compared to granite northern regions that are cooled down almost instantaneously.

Caledonian Orogeny in Southeast Asia: Questions and Problems

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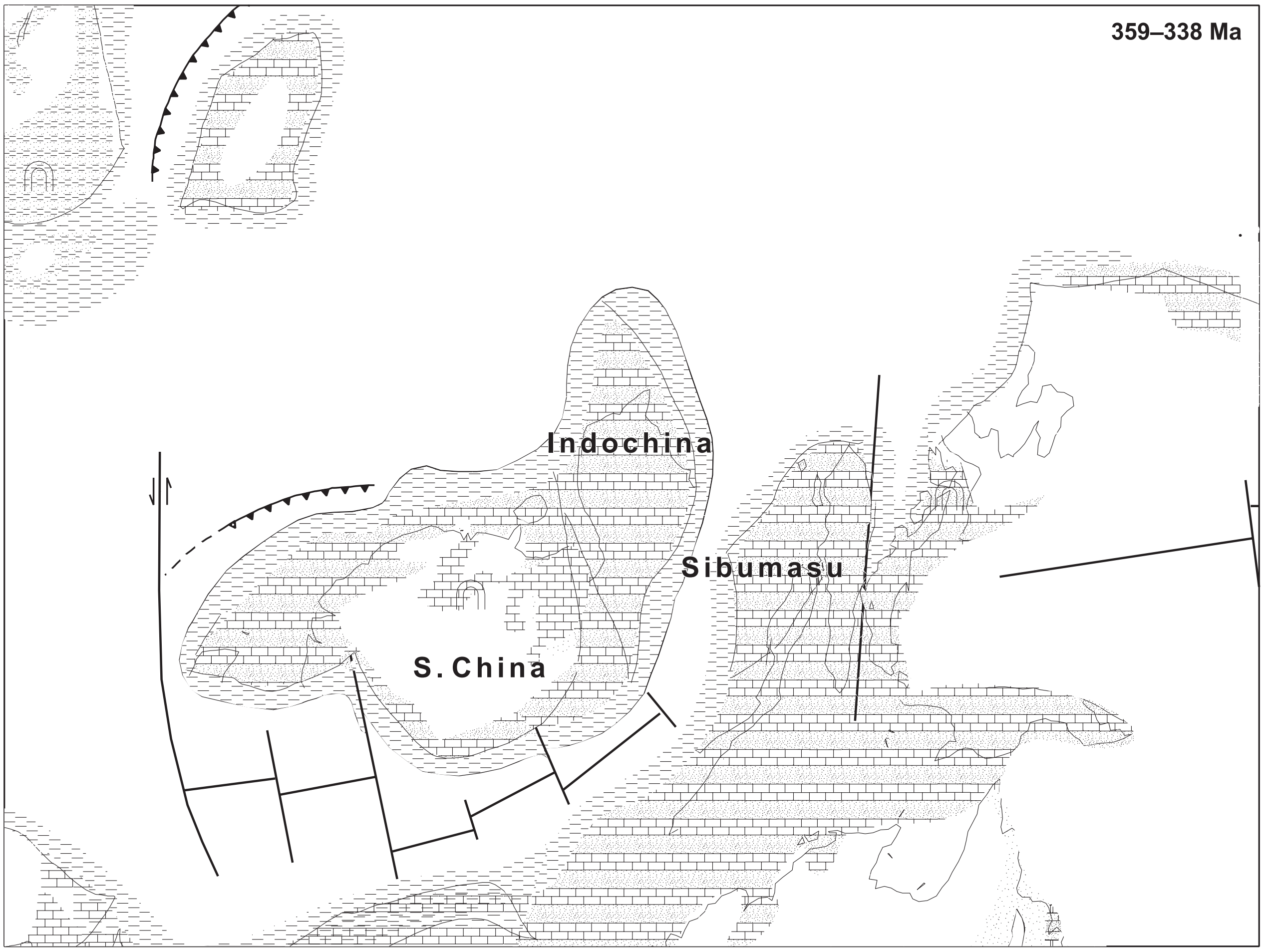
Avalonia probably started to drift from Gondwana and move towards Baltica in the late Tremadocian and was in a drift stage by the Llanvirnian (McKerrow *et al*. 1991, Torsvik *et al*. 1996, Golonka 2002). Between Gondwana, Baltica, Avalonia and Laurentia, a large longitudinal oceanic unit, known as the Rheic Ocean (McKerrow *et al*. 1991, Golonka 2002) was formed. Traditionally the continent of Avalonia consists of northwestern and possibly southern Poland, and their foredeep, terranes in northern Germany, the Ardennes of Belgium and northern France, England, Wales, southeastern Ireland, the Avalon Peninsula of eastern Newfoundland, much of Nova Scotia, southern New Brunswick and some coastal parts of New England. The Brunovistulicum terrane, some accreted terranes in the basement of East Carpathians parts of the Scythian platform, parts of Kazakhstan and Southern Mongolia terrane could constitute the eastern extension of the Avalonia (Paul *et al*. 2003a, b). The Turkmen (Zonenshain *et al*. 1990) and Solonker (Sengör and Natalin, 1996) oceans in Asia could constitute the eastern parts of this Rheic Ocean. Relationship of eastern peri-Gondwana terranes and Avalonia plates remain unknown and speculative. On presented maps the South China and Southeast Asia plates remain attached to Gondwana according to the previously published global paleoreconstructions (Golonka 2002). The alternative reconstructions (Paul et al. 2003a, b) suggest the possibility of extension of Rheic toward the easternmost part of Gondwana. It is not impossible that South China and Indochina plates were rifted from Gondwana in Ordovician. The uplift and volcanic rocks (Fig. 9) support such a possibility. According to Shouxin and Yongyi (1991) the Ordovician conformably overlies the Cambrian over most of the South China plate. The northern part of the plate (Yangzi Platform was covered with carbonates and mixed carbonate/clastic facies. The southern part of the plate is partially uplifted and partially covered by deep water synorogenic clastic deposits – more than 4000 m of weakly metamorphosed flysch, sandstones and graptolitic shales. Similar rocks formed on the margins of Indochina plate. They are known as Pa Ham formation (Ordovician-Silurian).

Late Siluruian was the time of the major development of the Caledonian orogeny and final closure of the Iapetus. The collision between Baltica and Greenland continued, marked by nappes in Norway and Greenland. After the complete closure of the Iapetus Ocean, the continents of Baltic, Avalonia, and Laurentia formed the continent of Laurussia (P. Ziegler 1989). It is quite possible, that at that time several microplates rifted away from the Gondwana margin to arrive at Laurussia and Kazakhstan at the Devonian-Permian time (Golonka 2002). The exact time and the nature of rifting of these terranes and their relationship to Southeast Asia and Chinese plates remain speculative. Accor-



■ **Fig. 1.** Plate tectonic and lithofacies map of Southeast Asia during Kaskaskia III time – latest Devonian–Early Carboniferous – 359–338 Ma

ding to Shouxin and Yongyi (1991) following orogenic movements (Guanxi orogenic episodes), the Late Silurian was a time of regression within South China plate.

This was the time of the final phase of the Caledonian orogeny, transpressional collision of Gondwana and Laurentia and formation of the Oldredia supercontinent, which included all major plates. Most of Oldredia was located between the South Pole and the Equator. According to Golonka (2002) collision of South and North America occurred during Early Devonian time. The Caledonian orogeny was concluded in Europe and North America. A late stage of thrust related deformation occurred in northern Scandinavia, eclogites formed about 410 Ma in Norway, in an over-deepened root of Baltica, which had developed in the ductile lower crust, as a response to extreme crustal shortening (Golonka 2000, 2002). The peak of orogenic process occurred during this time within Southeast Asia and South China. In Northern Vietnam deep water Ordovician and Silurian synorogenic deposits were replaced by continental Early Devonian red beds (Tran Van Tri 1979, Tran Due Luong and Nguyen Xuan Bao (Eds.) 1988, Phan Cu Tien 1989). This red beds can be observed in Mai Chu area. The important unconformity is visible between Early Paleozoic rocks and Middle-Late Devonian carbonate deposits in North Vietnam. The similar unconformity exists in the adjacent part of China. According to Leloup et al. (1995), within Yangtzi paraplatform south of Kunming, the lowermost sediments are folded (schistosed Proterozoic shales, carbonates and volcanoclastics). These sediments are covered by Lower Devonian conglomerates and sandstones followed by Upper Devonian, Carboniferous and Permian shallow-water carbonates. Perhaps the Late Silurian – Early Devonian was time of accretion of terranes to South China plate in the collisional process. The paleogeography of this event is unknown and will be a subject of future research. Perhaps these events were related to the global Caledonian orogenic process and closing of extension of Iapetus Ocean (see Paul et al. 2003a, b). This orogenic process is, according to Leloup et al. (1995) well established in Guangzi province where metamorphic Siluro-Ordovician schists are covered by Lower Devonian Old Red sandstones. The orogenic process was perhaps also related to the onset of rifting of South China, Tarim, and Indochina from Gondwana that happened, according to Metcalfe (1998), during this time leading to the future development of new generation of oceanic realms (Fig. 1).

In the western part of South China plate (Shouxin and Yongyi 1991) and in Indochina (Tran Van Tri 1979, Tran Due Luong and Nguyen Xuan Bao (Eds.) 1988, Phan Cu Tien 1989, Brookfield 1996) the previous synorogenic and postorogenic facies were replaced by shallow water carbonates. The mixed character of this carbonates is changing upward leading to the deposition of pure limestone and dolomites.

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Comparative, Velocity-Dependent Gravity Modeling of the Density Section Along Three Carpathian DSS Profiles

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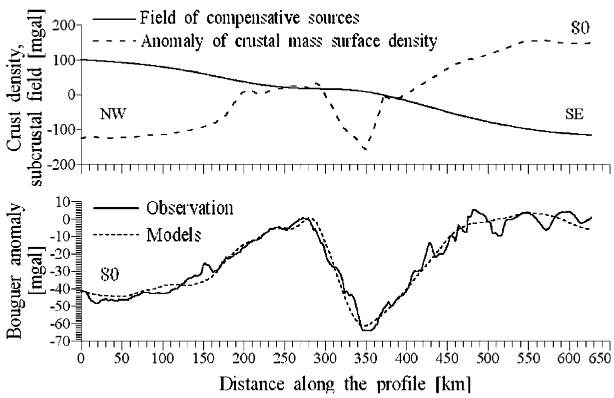
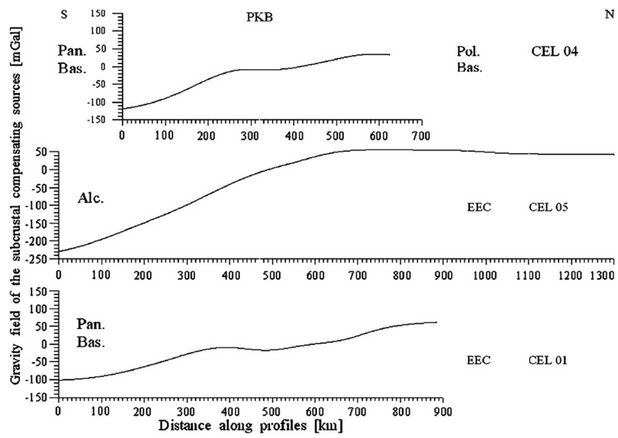
Difficulties arisen during gravity modeling along the three Carpathian DSS profiles (CEL 01, CEL 04, CEL 05) crossing boundary of the orogen, were a reason of searching for a modifications of the simplest method of constant layer densities applied at the beginning. The problem was resolved by taking more extensive advantage of the structural information contained in the velocity distribution in the cross section. A stable and convincing results were obtained by using a modified concept of the density field, where density is a function of p-wave velocity in each layer and the density satisfies limitations for



■ **Fig. 1.** Example of successful gravity modeling for profile CEL 04.

its value. Now, the modeling process can be regarded as successful in all three cases (Fig. 1).

The most interesting general results of tectonic character concern the presence of subcrustal isostatic compensation and its characteristic depth. Pronouncing regularity in the spatial distribution of the field of compensating sources (Fig. 2) seems to be a clear documentation of the state of the lower lithosphere



■ **Fig. 2.** Comparison of the spatial distribution of the field of compensating sources for the three Carpathian DSS profiles (CEL 01, CEL 04, CEL 05) crossing boundary of the orogen.

showing a large anomaly below Panonian Basin, anomaly being a record of the rift process in the basinal area during formation of Carpathians. Another results concern the significance of the crucial Carpathian tectonic boundary and its present dynamical state.

Timing and Structural Style of Final Thrusting Movements of the Carpathian Orogenic Wedge, S Poland

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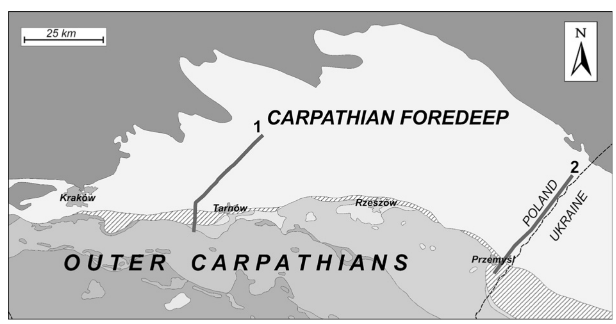
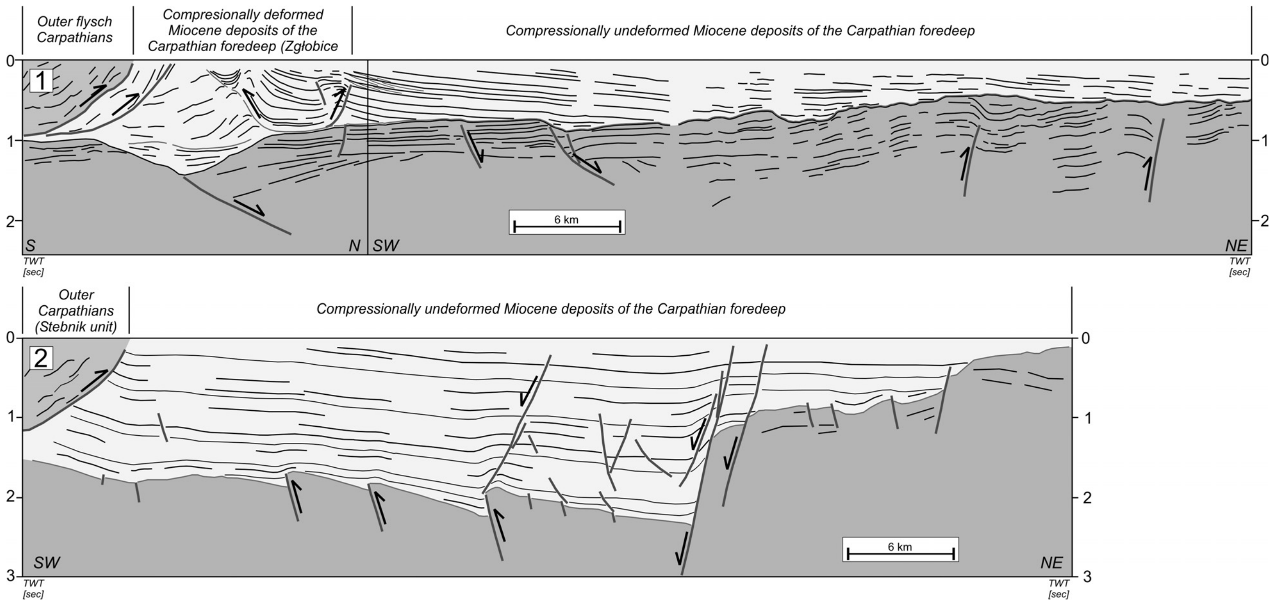
During progressive evolution of the thrust-and-fold belt deposits of the foredeep basin become progressively incorporated into the orogenic wedge. Such process is often syn-depositional, and consequently syn-kinematic (growth) strata from the foredeep infill could be used to decipher modes and timing of the thrusting movements.

Carpathian foredeep basin developed in front of the advancing Carpathian orogenic wedge (cf. Oszczypko et al. 2006 for further details and references). Its outermost part, presently located in front of the flysch (pre-Miocene) Carpathian units (nappes) is filled by the Badenian – Sarmatian deposits. In this paper two regional seismic lines are presented that illustrate very different gross structure of the orogenic front, foredeep basin and the foreland plate in central and E segments of the Polish Carpathians (fig. 1), and are used to constrain modes of final thrusting movements in this segment of this orogenic belt.

First profile (profile 1 – Fig. 2) is located in the central part of the Polish Carpathians and their foredeep basin, in vicinity of Tarnów. In this area frontal part of the orogenic wedge is built of relatively wide zone of deformed Miocene (Upper Badenian – Sarmatian) foredeep deposits that form the so-called Zgłobice Unit. These unit has been interpreted as a triangle zone cored by passive-roof duplex (Krzywiec et al. 2004). Formation of the triangle zone was controlled by morphology of the Mesozoic basement as well as by distribution of Upper Badenian evaporites. Within this zone numerous evidences of the syn-kinematic deposition have been identified, attesting to the latest Badenian – Sarmatian age of the final thrusting movements. They include progressive unconformities, localized thickness reductions within the crestal parts of the fault-related folds and small-scale fan deltas developed in front



■ **Fig. 1.** Location of regional seismic profiles from the central and eastern segments of the Polish Carpathian foredeep basin. Deformed foredeep deposits (older – Stebnik unit, and younger – Zgłobice unit) are shown by obliquely patterned area.



■ **Fig. 2.** Regional profiles from the central (1) and the eastern (2) segments of the frontal Polish Carpathians and their foredeep basin. Note very different structure of the orogenic front and the foreland plate observed along both profiles.

of the growth folds (cf. Krzywiec 2001). Foreland plate is characterized by rather gentle flexural profile without signs of any major flexural extension.

Very different picture is shown on profile 2 from the E part of the Polish Carpathians and their foredeep basin, located above the Teisseyre – Tornquist Zone (fig. 2). In this area Carpathian front is defined by rather sharp frontal thrust fault along which Miocene deposits of the Stebnik unit together with the Carpathian flysch nappes are overthrust above the compressionaly undeformed Badenian- Sarmatian deposits of the outer foredeep basin. Foreland plate shows large amount of flexural extension related to the late Badenian – Sarmatian reactivation of the fault zones belonging to the T-T Zone. Traditionally, using geometrical relationship between the hangingwall and the footwall of the thrust front, it was assumed that the thrusting movements were post-depositional in respect to the foredeep infill presently located in front of the orogenic wedge. Such a model would however require formation of large-scale fault-bend fold above the presently preserved foredeep infill. Considering structure of the thrust front and the foreland plate as well as the geometry of the Miocene infill in front of the orogenic wedge it is proposed that final thrusting occurred during flexural extension of the foreland plate and related sedimentation within the foredeep basin. Consequently, foredeep infill presently located in front of the Carpathian frontal thrust could be regarded as a syn-kinematic, not pre-kinematic, similarly to the central part of the Polish Carpathians.

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Salt Tectonics in Compressional Settings: Comparison of the S Pyrenees and the N Carpathians

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Evaporites in general, and rock salt in particular, are of key importance for evolution of fold-and-thrust belts, as evaporitic layers often form preferred levels of detachments within the orogenic wedge. The combined effects of the foredeep basin morphology during deposition of evaporites and distribution of the surrounding non-evaporitic depositional systems influence the position, extent and thickness of the foredeep evaporitic successions. The continuous forward propagation of the thrust front often result in forward and upward migration of the evaporitic units.

The Carpathians and the Pyrenees belong to the Alpine – Himalayan orogenic belt formed by the closure of the Tethys Ocean. At present, the frontal part of the S Pyrenees is well exposed, whereas front of the Polish Carpathians is mostly buried, especially in their central segment described below. In both the S Pyrenees and N Carpathians, the foredeep evaporitic layers constitute the principal detachment levels for the late development of both fold-and-thrust systems, with a strong coupling between tectonics and sedimentation and vice versa.

The external folded domain of the Southern Pyrenees is detached above several middle Eocene to middle Oligocene foreland syntectonic evaporitic layers (e.g. Vergés et al. 1992). The position, extent and thickness of these evaporites as well as the shape of the southwards transported Pyrenean thrust front constrain the position, geometry and trend of the series of detached anticlines in the Ebro Basin. The most important foreland detachment is located above the middle Eocene, 300-m thick Cardona salts on top of which a trend of continuous NE-SW trending anticlines developed (e.g. Sans and Vergés 1995). The tip line of the Pyrenean shortening corresponds in this region to a backthrust with NW vergence. Towards the SW boundary of the Cardona salt basin, the detachment climbs to the about 1,000 m thick middle Oligocene Barbastro evaporites. Above the ramp, the Barbastro and Sanaüja anticlines developed with their forelimb detached as a backthrust along the overburden – evaporites contact (Sans et al. 1996, Sans and Vergés 1995).

The Outer Carpathians are genetically linked to the Carpathian foredeep basin that developed in front of the advancing orogenic wedge (for recent summary see Oszczypko et al. 2005). Presently, in front of the Outer Carpathian flysch (pre-Miocene) units, a zone of deformed foredeep deposits exists of variable width, reaching max. 10 km in the area located between Kraków and Tarnów. Undeformed foredeep infill preserved in front of the Carpathians consists of the Upper Badenian – Sarmatian siliciclastic succession, with important evaporitic level at its bottom. Major detachments of this Miocene thrust system are related to the upper Badenian evaporites. Within the frontal part of the triangle zone of the Wojnicz slice two evaporitic horizons overlap laterally, one being uplifted by passive backthrusting of the Biadoliny slice, and another being preserved in its autochthonous position beneath the triangle zone (cf Krzywiec et al. 2004). Such configuration suggests that in this area two overlapping evaporitic horizons were deposited, contrary to previous models assuming single evaporitic level developed within the entire basin.

Using Cardona model proposed for the S Pyrenean front as well as triangle zone model proposed for the Tarnów sector of the Polish Carpathians, a new conceptual model for the Wieliczka sector (S from Kraków) was developed. In this part of the frontal Carpathians famous Wieliczka salt mine is located within the strongly deformed Miocene evaporitic zone in front of the flysch Carpathian nappes. Previously published cross-sections of the Wieliczka salt mine (Tołwiński 1956) suggest that also in this area triangle zone might have developed, and that the entire salt mine might be located within the core of this zone. Formation of the Wieliczka triangle zone and associated backthrust might have been controlled by lateral facies changes of the evaporitic unit – transition from thick salt-dominated domain to thin anhydrite-dominated domain. Additional control on both evaporitic facies distribution as well as tectonic style and location of backthrusting could have been exerted by morphology of the pre-evaporitic basement.

Acknowledgements

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Role of Basement Tectonics in Evolution of Salt Diapirs: the Mid-Polish trough Versus the Dead Sea Basin

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During the extensional basin subsidence, salt layers underlying a thick sedimentary overburden can start to flow, giving rise to the development of a variety of halokinetic structures, such as salt diapirs, salt pillows and salt walls. Salt flow can be triggered by extensional faulting of the sub-salt "basement" (Koyi et al. 1993), as well as by thin-skinned extension of the post-salt sedimentary cover (Vendeville and Jackson 1992a, b). In intracontinental settings salt structures are particularly often related to sub-salt fault zones as in such basins localised extension and subsidence is associated with significant faulting within the pre-salt basement.

The Mid-Polish Trough (MPT) formed the axis of the Polish Basin which belonged to the Permian-Mesozoic system of West- and Central-European epicontinental basins (Ziegler 1990). During the Permian, the MPT formed the easternmost part of the Southern Permian Basin. Prior to its Late Cretaceous – Paleocene inversion, the MPT was filled with several kilometres of Permian and Mesozoic sediments, including thick Zechstein salts. The presence of these Zechstein salts gave rise to the development of a complex system of salt structures in the central and northwest segments of the MPT.

Recently completed regional analysis of seismic reflection data from the entire territory of the Mid-Polish Trough allowed to formulate some rules concerning relative roles of the basement, cover and salt tectonics. Using results of interpretation of seismic data basin-scale sub-Zechstein basement fault pattern responsible for the Mid-Polish Trough subsidence and inversion was proposed together with its role for development of salt structures (Krzywiec 2004a, b, 2006a, b). Basement extension has resulted – since the Triassic – in initiation of salt pillows. In some areas, peripheral (i.e. located outside the zone of maximum subsidence) fault zones formed within the Mesozoic sedimentary cover. Within the central (Kuiavian) segment of the basin very intense extension and faulting led to development in Late Triassic of the salt diapir that extruded onto the basin floor and was covered by uppermost Triassic and Jurassic deposits. During Late Cretaceous inversion salt structures present within the Mid-Polish Trough have been reactivated, both due to basement mobility (uplift of the basement block along reverse faults) as well as compressional stress field acting within the sedimentary cover. Compressional reactivation is best observed for the Drawno – Człopa – Szamotuły salt structure system (Krzywiec 2006b). In this area compression resulted in active diapirism, and growing diapir caused development of numerous unconformities in its vicinity that document consecutive stage of its development. During inversion within peripheral parts of the Mid-Polish Trough salt pillows were formed entirely related to the Mid-Polish Trough inversion and related lateral salt flow. Growth of such salt structures is documented by local thickness variations of the Upper Cretaceous deposits. Analysis of seismic data provided also information on Cenozoic reactivation of selected salt structures. Within the Drawno – Człopa salt structure system extensional reactivation of their topmost parts is observed. Similar activity connected with significant localised subsidence and deposition of brown coal seams has been described above the Damasławek salt diapir (Krzywiec et al. 2000).

A similar interaction between basement faulting, a thick salt layer and its supra-salt sedimentary cover was documented in many other basins, with good example provided by the Dead Sea Basin. This basin is a continental depression located within the rift valley that accompanies the Dead Sea Transform (DST). It is widely agreed that the basin is a rhomb-shaped pull-apart graben that was formed due to the left-lateral displacement along the segmented DST. The basin is bounded on the east and west by a series of oblique-normal (basinwards) faults, which suggest that the basin underwent active transtensional rifting.

Within the Dead Sea basin, the presence of thick Pliocene salt and active Quaternary normal faulting resulted in the development of numerous salt structures (e.g. Sedom and Lisan diapirs) and in different degrees of decoupling between the thick-skinned basement tectonics and the thin-skinned cover tectonics (cf. Al-Zoubi and Ten Brink 2001, Al-Zoubi et al. 2001, Larsen et al. 2002). The location of the Sedom salt diapir was dictated by the existence of oblique-normal faults in the margins of the basin. Presently, salt entirely pierced its overburden and extruded on the surface where it presently forms Mount Sedom with surface expression up to 200 m (Weinberger et al. 1997, Weinberger et al. 2006a). The present uplift rate of Mount Sedom, calculated from precise leveling and Interferometric Synthetic Aperture Radar (InSAR) is 6–9 mm/y, and is similar to the average Holocene rate (Weinberger et al. 2006b). Present-day crustal configuration, including intensly faulted basement resambles the Late Triassic structure of the Kłodawa salt diapir and its basement, located within the central part of the Mid-Polish. Such similarity is also related to possible influence of strike-slip movements on the Triassic evolution of the Mid-Polish Trough.

The comparative study of the Mid-Polish Trough and the Dead Sea Basin is being completed within the International Lithospheric Programme Task Force "Origin of Sedimentary Basins".

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Internal Flow Fabric Study of Viscous Lava Domes in Central Slovakia by Means of AMS and Quantitative Microstructural Study

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Two examples of acid and intermediate volcanic bodies in the central Slovakia (middle miocene in age, Konečný et al. 1995) were studied in order to understand the mode of magma flow and emplacement mechanism of highly viscous volcanics. We present preliminary results of integrated AMS and microstructural study. The garnet-bearing andezite dome Breziny (Neresnica Formation., Badenian) was investigated in detail. The quarry located cca 1,5 km NE from the Breziny village exposes southern margin of an extrusive andezite dome. Ten samples were collected from three quarry levels for detailed AMS and textural analysis. It was suggested that the internal magmatic fabric in the quarry forms a fan-like pattern (Konečný et al. 2004), typical for andezite extrusive domes in the region. The strikes of magnetic fabric in the quarry show a sinusoidal trend with one limb subparallel to the dome margin in the map and rather steep dips (80–90°) and do not show any trend of dips incident to fan-like pattern. The magnetic lineations are subhorizontal (0–35°) and trend NW-SE on the southern margin of the quarry. In the centre (2nd and 3rd quarry level) lineations are vertical and the AMS fabric there is also characterised by slightly lower values of T parameter than in the rest of the samples. Susceptibility-temperature curves in HT and LT document a presence of paramagnetic minerals (amphibole, biotite) and titanohematites. For the purpose of textural analysis, slabs parallel with AMS K1K3 and K2K3 planes were prepared and the fabric intensity and geometry defined by alignment of dark (mainly amphiboles) and white (mainly plagioclase and pyroxenes) minerals was statistically quantified from the slab photographs. The digitization of mineral objects and statistical evaluation of the textures was carried out using ArcView and extenstion PolyLX in Matlab environment (e.g., Lexa 2005). We compare the rock textures using the eigenvalue ratios of the bulk orientation tensor and aggregate distribution throughout the quarry. Furthermore we try to discriminate signatures of magma evolution from the grain size distribution (CSD) curves (Cashman and Ferry 1998). In addition, preliminary results from the rhyolite Jastraba Skala dome (sarmatian-panonian) comprising the relationship between magmatic textures and AMS are briefly discussed.

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Vertically Decoupled Thickening and Exhumation Processes in Orogenic Supra- and Infra-Structure During Building of Gemer-Vepor Continental Wedge

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Thickening of Gemer supracrustal unit occurred through development of wide positive cleavage fan (GCF) structure recently dated at 130–120 Ma using K/Ar and monazite U/Th method. This crustal scale structure is characterised by development of steep fabric in the core of the GCF associated with vertical extrusion of deeper portions of the Gemer Unit. In contrast, the Vepor infrastructural unit shows development of flat mylonitic fabric in deeper part of the basement associated with homogeneous burial. The internal deformation of the Vepor basement is poorly dated ,but it is bracketed by onset of inversion of the Zliechov basin to the north (~110 Ma) and 40Ar/39Ar micas and hornblende cooling ages in range 80–90 Ma. These two contrasting tectonic regimes were separated by greenschist facies mylonitic basement rocks and large portions of weakly deformed basement material. The plausible tectonic model explaining structural and metamorphic evolution of both crustal levels suggests existence of neutral level that is most likely located between Gemer and Vepor interface (Gemer-Vepor Contact Zone – GVCZ). This zone served as a decoupling horizon separating vertically elevated rocks from those, which were simultaneously buried. The hanging-wall Gemer Unit thickened by convergent flow while the Vepor Unit burial occurred by divergent flow or “syn-burial ductile thinning”. These competitive processes are registered by development of the GCF in the Gemer Unit and by PT gradients of different structural levels in the Vepor Unit. The lower crustal flow in the Vepor infrastructure progressively generated strong horizontally oriented mechanical anisotropy leading to continuous decrease of buckling resistance of the pile followed by large scale folding of the Vepor-Gemer multilayer system at ~80 Ma. The weakly deformed upper part of the Vepor basement surrounded by weaker Lower Paleozoic Gemer rocks and mylonitized lower crust dominated by amphibolite facies micaschists and gneisses represented a rigid layer controlling wavelength of crustal scale buckles. During folding the orogenic lower crust was exhumed by viscous extrusion along narrow belts when the folding mechanisms passed from active to passive amplification. We propose, that during this process the GVCZ was reactivated by fold hinge parallel slip (Trans-Gemer Shear Zone) of the suprastructure, commonly termed as “unroofing” of the Vepor basement. This process likely results from non-cylindrical growth of crustal buckle as well as from possible changes in far filed forces responsible for development of large-scale Upper Cretaceous sinistral shear zones.

Young Tectonics of the Orava Basin and Southern Magura Nappe, Polish Western Carpathians, in the Light of Gravity Studies: A New Research Proposal

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The Department of Geophysics, AGH University of Science and Technology in Kraków initiated in 2004 a new research proposal, the aim of which is to compare the results of studies throughout the Polish segment of the Western Outer Carpathians showing differentiated young tectonic movements. It is a multidisciplinary project; analysis of young crustal movements will be conducted basing on the results of three years long gravimetric, geodetic, geological and morphostructural studies. The problem consists in quantitative interpretation of the obtained gravimetric results, i.e. in combining temporal gravity changes with those of geodynamic crustal processes.

From the geodynamic point of view, the Orava-Nowy Targ Basin is an extremely interesting structure. This is a bi-partite basin, formed in Miocene time and superimposed on structural units that build the contact between the Inner and Outer Carpathians, namely: the Central Carpathian Palaeogene Basin, Pieniny Klippen Belt, and Magura Nappe. The maximum drilled thickness of sedimentary infill of the Orava Basin is 950 m, including 922 m of fresh-water Neogene molasses (Watycha 1976). The thickest Quaternary sediments (117 m) are confined to the Wróblówka Trough, in the northern part of this basin.

The Orava Basin is a tectonic trough which is bounded to the north and south by a system of longitudinal normal faults of throws up to a few hundred metres. These are cut by several transverse, mostly strike-slip faults that are oriented NNW-SSE and NE-SW (Pomianowski 1995, 2003). The basin-bounding faults became reactivated in Quaternary times (cf. Baumgart-Kotarba 1996), and their recent activity is indicated by earthquakes of magnitudes up to 4.3 (Guterch et al. 2005).

Our studies concentrate along a ca. 40-km-long, N-S trending, transect: Dzianisz–Czarny Dunajec–Wróblówka–Spytkowice–Wysoka, which cuts the contact between the Inner and Outer Carpathians, showing contrasting tendencies of young (Pliocene-Quaternary) tectonic movements. The Wróblówka Trough, situated in the medial segment of the transect, reveals Late Pleistocene and Holocene subsidence, while the southern portion of the Magura Nappe, in the northern portion of the transect, displays minor uplift. The location of stationary points was selected in such a way that each of them represents a different structural unit (Central Carpathian Palaeogene, Pieniny Klippen Belt, Orava Basin, Krynica and Bystrica subunits of the Magura Nappe). The construction of individual benchmarks enables for both gravity and geodetic measurements.

Gravity surveys across the profile will be carried out at yearly intervals. The choice of such methodology results from the fact that we want to study changes of the gravity field statistically, taking into account the expected small values of temporal anomalies of the gravity field and, first of all, their changes with time.

In July 2004 and July 2005, the first and second series of gravity measurements were made at fixed benchmarks of the profile, using three gravimeters (two CG-3 SCINTREX, and one La Coste & Romberg). Errors were calculated after each series. The calculation error was determined for each gravity value between the stations and for the average gravity value. The measurement precision was ca. 0.01 mGal, or 0.005 mGal during cycling measurement. An analysis of errors showed that in several cases only, i.e. for the measuring date, the error was larger than 0.005 mGal, and the overall error credibility limit attained a value of 0.01 mGal.

The average gravity error has a little bigger value, although not exceeding 0.01 mGal. This maybe a result of too small a number of gravimeters used. Therefore, the temporal gravity analysis will be done basing on gravity measurement values for each gravimeter, and not for the average gravity value.

The first measurement series made in July 2004 is a base series, to which measurements of successive series will be referred. Additional measurements were conducted at points situated 1 km apart along the profile steps. These results, combined with geological data, will be used in gravity modelling aiming at determining mutual connections between gravity anomalies and geological structures. This modelling will also be helpful in describing the source of changes of the gravity field. The second measurement series, conducted in July 2005, enables for the first comparison between gravity changes measured in 2004 and 2005. The determined trend will be verified by successive measuring campaigns.

A series of earthquakes occurred in the Orava Basin in November 2004, pointing to recent tectonic mobility of this area. Hence, changes of the gravity field observed between July 2004 and July 2005 appear to be particularly interesting.

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Structural Position and Metamorphism of Peridotite and Eclogite Bodies within Granulite in the Bestvina Unit, Bohemian Massif

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The Bestvina unit represents the westernmost part of the Kutná Hora crystalline complex and occurs in the central part of the Bohemian Massif. It tectonically overlies the Varied group of the Moladnubian domain in the SW and consists mainly of retrogressed felsic granulites, biotite gneisses, migmatites and several small isolated bodies of peridotite and eclogite (Pouba et.al. 1987, Synek and Oliveriová 1993). Two different deformation planar fabrics have been recognized during field works in the Bestvina unit. The older F_1 fabric has been observed in coarse-grained and weakly retrogressed granulites and in eclogites. It is parallel to layers of olivine and garnet-bearing pyroxenite occuring in peridotites. This planar fabric strikes NE-SW and dips under steep angles to NW or SE. Isoclinal fold of the F_1 fabric has been observed inside the eclogite body having steep axial plane striking SW-NE and subhorizontal fold axis. A younger F_2 fabric is extensively developed in the majority of the unit, where F_1 fabric is preserved in the low-strain domains only. The F_2 fabric moderately dips towards ENE to NE and it is developed in retrogressed and mylonitized granulites and migmatites. The mylonitic foliation F_2 is parallel to the foliation observed in paragneisses of the Varied group below the Bĕstvina unit.

Detailed petrological study has been carried out on ultramafic and eclogite bodies near Spačice and on peridotite at Doubrava and Úhrov in order to constraint PT conditions of the studied rocks. The Spačice eclogite forms ca 60 m long sigmoidal lens that occurs within retrogressed granulite. The original metamorphic assemblage of eclogite consists of omphacite, garnet, kyanite and rutile. Two textural and compositional varieties of garnet and of clinopyroxene are present in this eclogite. The eclogite facies garnet – Gr I (Py_{36}, Grs_{34}, Alm_{28}) contains rutile inclusions and associates with omphacite Cpx- I (Jd_{29}). Omphacite in some samples is characterized by the presence of quartz rods that usually occurs in ultrahigh-pressure metamorphic rocks. Garnet I is partly replaced by Al-rich clinopyroxene (Cpx II) and anorthite. A new Ca-rich garnet Gr-II (Py_{10}, Grs_{65}, Alm_{23}) that forms either individual grains or rims of the coarse-grained eclogite facies garnet, indicate textural equilibrium with Al-rich clinopyroxene and plagioclase. There is a sharp compositional jump with a very weak diffusion profile between these two garnet varieties. Mn content is low in both garnet, but the Ca-rich garnet has relatively higher Mn, suggesting decomposition of older garnet. Small amount of tschermakitic amphibole replacing Ca-rich garnet is also present. The Doubrava peridotite forms ca 40 m wide lens-shaped body surrounded by coarse-grained and retrogressed granulites. Within the peridotite body, there is small body of garnet rich eclogite, and olivine and garnet-bearing pyroxenites. Pyroxenites form up to 30 cm wide parallel oriented and steeply dipping layers providing primary anisotropy of peridotite body. The garnet peridotite has relicts of olivine, orthopyroxene, clinopyroxene, spinel and rarely of amphibole. Chromium-rich spinel forms inclusions in garnet and in clinopyroxene. Compositional maps indicate progressive formation of garnet after spinel. Garnet is rich in Mg (Py_{69},$Grs_{11}Alm_{18}$) and forsterite content in olivine is about 93 mol%. Clinopyroxene is diopside with $X_{Mg} = 0.9$. Orthopyroxene with $X_{Mg} = 0.8$ has Al_2O_3 about 1.7 wt.%. Spinel corresponds to Al-chromite with composition of $Mg_{0.54}Fe_{0.47}Al_{0.73-1.0}Cr_{0.8-1.19}O_8$. The eclogite within garnet peridotite has relatively high-Mg garnet (Py_{42}, Grs_{34}, Alm_{22}) and omphacite with Jd_{30}. Similar to eclogite from granulite, garnet is replaced by Al-rich clinopyroxene and anorthite ± amphibole and kyanite by anorthite, spinel and locally clinopyroxene. Garnet contains rutile needles that mostly have parallel orientation. The Úhrov peridotite is poorly outcropped and does not provide any strain features in the field our analyses showed its textural and mineralogical similarity to the Doubrava petridotite. Maximum PT conditions of ~4 GPa at 700 °C were calculated for Spačice eclogite as well as for eclogites inside the Doubrava and the Úhrov peridotites. Garnet peridotites reveal pressure conditions similar to eclogite but at high temperature of about 1000 °C. Textural relations and chemical composition in all rock types, but mainly the presence of Ca-rich garnet in eclogite, suggest that decompression was followed by rapid cooling.

Field structural data suggest that peridotites penetrate granulites during large scale folding of coarse-grained granulites along steep axial plane striking SW-NE. Pyroxenite layers provide very likely compositional as well as mechanical anisotropy in peridotite that helped to dismember mantle rocks to small bodies during variscan exhumation. However, PT estimated form peridotites, eclogites and granulites (Medaris et. al. 1995, Medaris et.al. 1998) show large disagreement between each other and it is difficult to link individual tectonic events with metamorphism. Later fabric F_2 corresponds to mid crustal deformation event that affect whole Běstvina unit together with Varied group in the Moldanubian domain.

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Neotectonic Investigations of the Érmellék Region (NE Pannonian Basin, NW Transylvania)

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Neotectonic investigation has been carried out along the Ér-river valley, and between the Ér- and Berettyó-river valleys (Érmellék region). This ENE-WSW striking hilly region is situated on the northeastern part of the Pannonian Basin and NW of the Transylvanian (Apuseni) Mountains. The aim of the study was to find evidences for the hypothetised neotectonic control on river network development of the Körös Basin. The Émellék region represents a natural link between the uplifting Apuseni Mountains and Körös Basin which is the deepest sub-basin of the subsiding Great Plain. The Érmellék region is famous for its presumed neotectonic activity is shown by two larger historical earthquakes occurred in 1829 (M = 4,9; I_{max} = VII) and 1834. (M = 6,3; I_{max} = IX) (Réthly 1952).

The hilly part of the region is mainly covered by loess and "red clays" (Sümeghy 1944). The latter is a brown forest type paleosoil complex of the loess sequence which is resistant to erosion and dominantly covers the top of the ridges. The age of the loess sequence was not dated till this time, but was preliminary correlated to paleosoil horizons and may represent loess up to Middle Pleistocene (Upper sequence of the Old Loess series of the Paks Loess Formation, Marsi et al. 2004). In the Ér-river valley Late Pleistocene – Holocene alluvial sand and aleurolite can be found at different topographic height which are probably the remnants of terraces of the palaeo-Tisza, which was flowing along the northeast-southwest striking Érmellék depression (Ér-river valley) during the Late Pleniglacial (Gábris and Nádor in press).

We investigated the outcrops of the above mentioned Quaternary sediments of the region by structural, tectono-morphological and sedimentological methods to quantify the main fault directions in the field, and analysed the morphology and river network to determine the style of neotectonic deformation. We found two phases of deformations, based on microtectonic investigation of the area. The older is reflected by NE-SW trending normal faults, joints and dykes in the loess, filled with reddish, brown aleuritic clay. This is a redeposited material of the brow forest paleosoil complex. The younger/second phase is mainly reflected by rejuvenated shear faults of the first phase and Riedel-faults. These are usually filled by greyish-brown aleuritic clay wich are probably originated from chernozem-brown paleosoil of the eroded Upper Pleistocene paleosoil complex or recent zonal soil. Apart from small scale faulting, the most characteristic neotectonic feature is surface undulation. This phenomenon is probably related to folding, based on the en-echelon arrangement of the ridges of elongated undulations.

Combination of microtectonical datas with the morphotectonical observations and river network analysis, we concluded that the Érmellék region was a left lateral ENE-WSW striking fault zone with NE-SW compression and perpendicular extension up to the Middle Pleistoce. The second phase was a reactivation of the „first" phase, generated by WNW-ESE compression, and caused right lateral transpressions. This seems to be active till this time. Active deformation is also supported by the presence of historical earthquakes, too. This zone is in the norteastern continuitation of those tectonical lines which were analysed from seismic sections of the Körös Basin and caused main tectonic control on river network developement during the Late Pleistoce (Nádor et al., in press).

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REE Accessory Minerals as Regional Metamorphic Processes Indicators: An Example from Wedel Jarlsberg Land, Svalbard

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Accessory minerals are commonly used in reconstructions of metamorphic evolutions and geotectonic interpretations – from thin section scale to large regions. Significant role in that play REE- and REE-bearing minerals, such as monazite, xenotime, allanite and apatite. These minerals were previously used as indicators of metamorphic processes and their P-T conditions by numerous authors (e.g. Finger et al. 1998, Spear and Pyle 2003, Wing et al. 2003). In this paper we present use of interpretations based on reactions involving monazite, xenotime, apatite and allanite compiled with geochronological data in reconstructions of metamorphic evolution of the Isbjørnhamna Group rocks (see also Majka and Budzyń 2006).

Polimetamorphic tectonic block, composed of the metasedimentary Isbjørnhamna Group, conformably covered by metavolcanosedimentary Eimfjellet Group, is distinguished in SW part of Wedel Jarlsberg Land in Svalbard (Czerny et al. 1993). This complex was affected by metamorphism two times: firstly under amphibolite facies conditions (Barrovian type), and secondly under greenschist facies conditions (Majka et al. 2004).

Fine-grained mica schists from the Isbjørnhamna Group were studied. Quartz, biotite, muscovite, garnet, chlorite (progressive) and plagioclase are present as main minerals. Kyanite, staurolite, chloritoid occur in some samples. Moreover accessory tourmaline, zircon, sphene, apatite, monazite, xenotime, allanite, unidentified Th-phases, ilmenite, hematite and magnetite are common. Partial or complete replacement of garnet and biotite by chlorite, disintegration of muscovite, sericitization of plagioclase and kyanite indicate changes related to the low temperature metamorphism.

Euhedral monazite grains generally without zonation occur in the Isbjørnhamna Group metapelites. Chemical U-Th-total Pb method performed on monazite grains (some of them enclosed in garnets) provided uniform Cadomian (643 Ma) ages. Basing on the fact that monazites enclosed in garnets yield the same age, it is unquestionable that first metamorphic event took place during that time and also indicate, that this event was a result of orogenic movements in large scale. Previous geochronological results of Ar-Ar dating performed on micas and hornblende, reported by Manecki et al. (1997) indicate similar Cadomian ages (616 Ma for Hbl, 584–575 Ma for micas).

Investigated region was affected by later changes of P-T conditions resulting in breakdown of primary monazite and formation of apatite and/or allanite coronas. It is important to notice, that these secondary minerals are stable in lower P-T conditions than monazite, characteristic for low and middle greenshist facies (lower than Bt-in isograd). These changes are probably connected with younger Calledonian metamorphic event indicated by Ar/Ar dating (459 Ma; Manecki et al. 1998), or could be connected with cooling during the exhumation of orogen after Cadomian metamorphic event.

Compilation of the geochronological data and closing temperatures of investigated monazites – as well as hornblende and micas analyzed by Manecki et al. (1998) – provides the cooling ratio of the whole orogen equal to ca. 100 °C / 20 Ma. These data indicate maximum of metamorphism at ca. 643 Ma followed by slow exhumation of the orogen and erosion without significant uplift what took place till early Cambrian (ca. 575 Ma). Connecting of such geological history of this part of Wedel Jarlsberg Land with tectonic and litostratigraphical knowledge provides description of evolution of unique exotic terrain in Svalbard Archiepalgo.

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The Tectonic History of the Mýto – Tisovec Fault (Western Carpathians)

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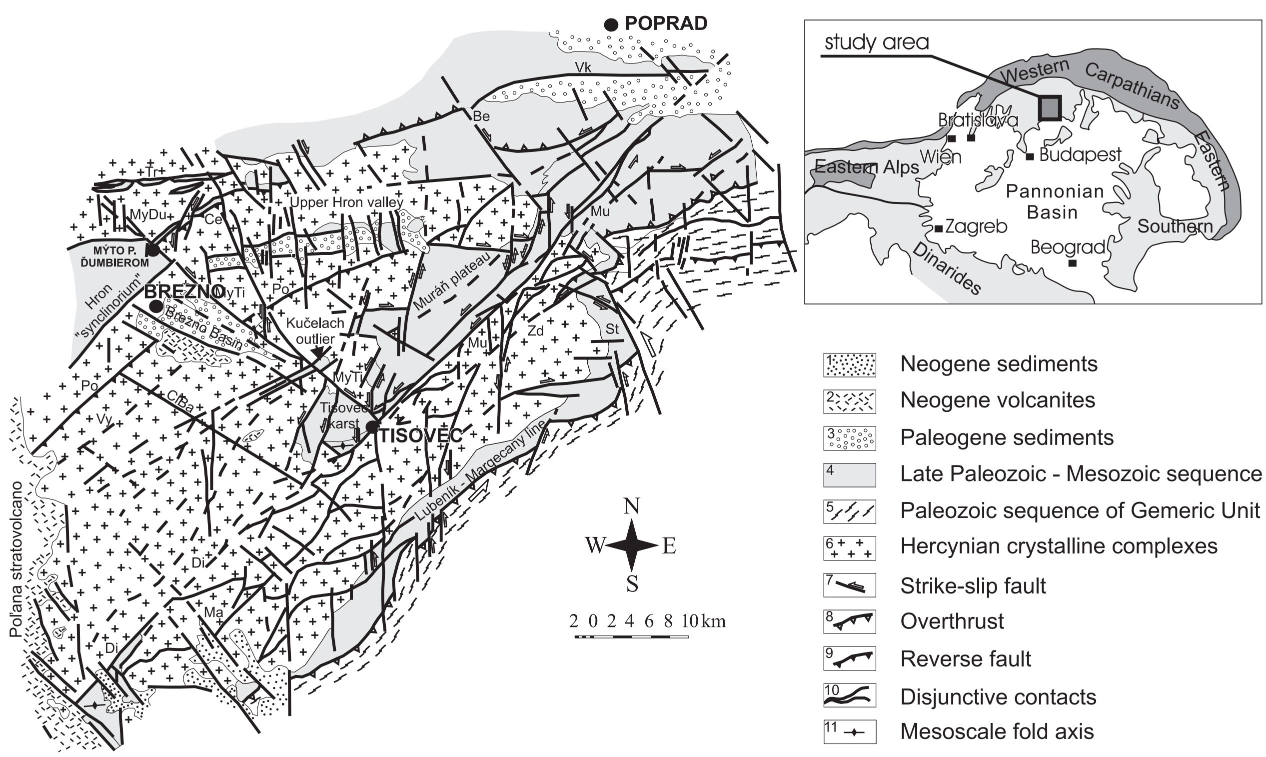
The Mýto-Tisovec fault has been first described as the Mýto fault by Zoubek (1935), according the village Mýto pod Ďumbierom in his vicinity. The name Mýto-Tisovec fault (Marko 1993a) used herein unambiguously defines the southern studied segment (in between Mýto pod Ďumbierom and Tisovec) of this important NW-SE striking disslocation (Fig. 1).

Methods

As the research approach to solve the topic, combination of field mesostructural observations and map-scale structures analysis has been applied. Structural research has been focussed to investigation of brittle deformations related to paleostress field studied along the map trace of the Mýto-Tisovec fault. Area of



■ **Fig. 1.** Structural-tectonic map of the northwestern Veporic area (Marko 1993b). Coded names of faults and shear zones: Be – Benkovo fault, Ce – Čertovica shear zone, CiBa – Čierny Balog f., Di – Divín f., Ma – Málinec f., Mu – Muráň f., MyDu – Mýto-Dúbrava f., MyTi – Mýto-Tisovec f., Po – Pohorelá s. z., St – Štítnik f., Tr – Trangoška f., Vk – Vikartovce f., Vy – Vydrovo f., Zd – Zdychava s. z.

the Mýto-Tisovec fault zone has been regarded as a one structural domain, allowing combination of faults from different localities for paleostress analysis (direct inversion method, Angelier 1984).

Results

NW-SE striking Mýto-Tisovec map-scale brittle fault distinctively affects internal zones of the Western Carpathians. It cuts and evidently offsets Meso-Alpine tectonic units and structures and represents the zone of important geophysical anomalies as well. Using methods of structural analysis, complex tectonic evolution of this long living fault has been restored. Six successive fault-slip related regional paleostress events, controlling the activity of the Mýto-Tisovec fault has been distinguished. The oldest recognized paleostress event, with NNE-SSW maximum principal stress axis operated after the Late Cretaceous and before the Middle Eocene. Orientation of the Miocene maximum principal stress axis clockwisely rotated from NW-SE in the Early Miocene to the NE-SW direction in the Middle Miocene and E-W direction in the Late Miocene - Pliocene. NNW-SSE trending compression has been estimated for the Quaternary stress field. Correspondingly three periods of Miocene tensional paleostress events with NE-SW, NW-SE and N-S orientation of minimum principal stress axis has been restored as well. The Mýto-Tisovec fault kinematically fluctuated in the changing paleostress field. However, the most evident structural records are related to the dominant dextral strike-slip regime. Dextral transtensional tectonic regime was responsible also for opening of narrow and deep depositional depression – Brezno Basin, related to the Mýto-Tisovec fault, where the Late Eocene – Early Miocene sediments of the Central Carpathian Paleogene Basin (CCPB) fill have been deposited and later preserved.

Restored evolution of the Neogene – Quaternary paleostress field submitted herein fits well with paleostress evolution of the ALCAPA junction area (Nemčok et al. 1989, Csontos et al. 1991, Marko et al. 1995, Fodor 1995). In spite of this similarity, no blok rotations (well known from ALCAPA junction area) has been taken into account in geodynamic model of the Mýto-Tisovec fault area evolution. It has been decided due to the lack of paleomagnetic data from the northwestern Veporic, even a few zero paleomeridian rotations were measured in Jurassic rocks (Kruczyk et al. 1992) in the similar terrane – the area of the Vysoké Tatry Mts.

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A New Method in the Geologic-Tectonic-Hydrogeologic Documentation of Shafts and Tunnels

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The base of the documentation is a photo series with known and programmable camera models made by the self-developed ImaGeo® Photorobot equipment. This equipment is combined with a laser distance measurer which measures wall distances in a programmable density. A self-developed software displays the composite image in 3D and evaluates the images from a geologic-tectnic-hydrogeologic point of view. The composite image and the evaluation are oriented by means of geodesic measurements.

The development of this new method was triggered by the project aimed to find an underground final disposal place for the low and intermediate level radioactive waste produced in Hungary. Within the frame of this project, two shafts are being driven near Bátaapáti (Geresd Hills, SW Hungary) in the Mórágy Granite Formation.

The base of the documentation is the photo series of each advance taken of the front wall and the shaft wall. The photos are taken by the self-developed ImaGeo® Photorobot, which takes each photo with a known and adjustable camera model. The camera model includes the focus and measured distance values, the spatial angle of the camera, and the exposition values as parameters. All these parameters can be programmed in the photorobot. Since the photorobot can be rotated around two axes, it is able to take photos in any spatial angles. Besides the rotating mechanics and the control electronics, it also contains a laser distance measurer, which measures distances within a photo in programmable numbers and places. The places of the measurements can be identified in the photos, and the distance data can be assigned to the matching the pixels. These distance data provide a mass of points in space, onto which the surface of the front wall and the shaft wall can be fitted. This is our shaft wall model. With the help of the camera models, the photoseries are fitted onto the shaft wall model, which becomes 3D this way. The geologic-tectonic-hydrogeologic documentation can be fixed in this 3D model. The model is placed into a georeferred space by doing geodesic measurements in three points on the shaft wall and one point on the front wall. Therefore, the photos of the advances can be fitted together and the documentation can be connected through several advances.

The ImaGeo Photorobot is served by also self-developed softwares. Besides the controlling software, CoreDump has been developed, in which the composite images are made from the individual photos and fit into the shaft wall model. The software makes it possible to draw objects on the composite images. Thus, a spatial system of lines results, where the objects belonging to different pnenomena can be placed on separate layers. The objects may be appended by a database. A line can be appended by a plane, a geologic-tectonic classification, an azimuth, a dip degree, surface parameters, a geometry, infillings, hydrogeological parameters, etc. The database can be shaped according to the actual research project, quieries can be made, and the results can be visualized: stereograms, pole distribution diagrams, density histograms can be made.

The database is exported from CoreDump to AutoCad, where the final geologic-tectonic evaluation and the plotting is done.

Reconstructing Post-Carboniferous History of the Krkonoše Piedmont Basin Using Detrital Apatite Fission-Track Data

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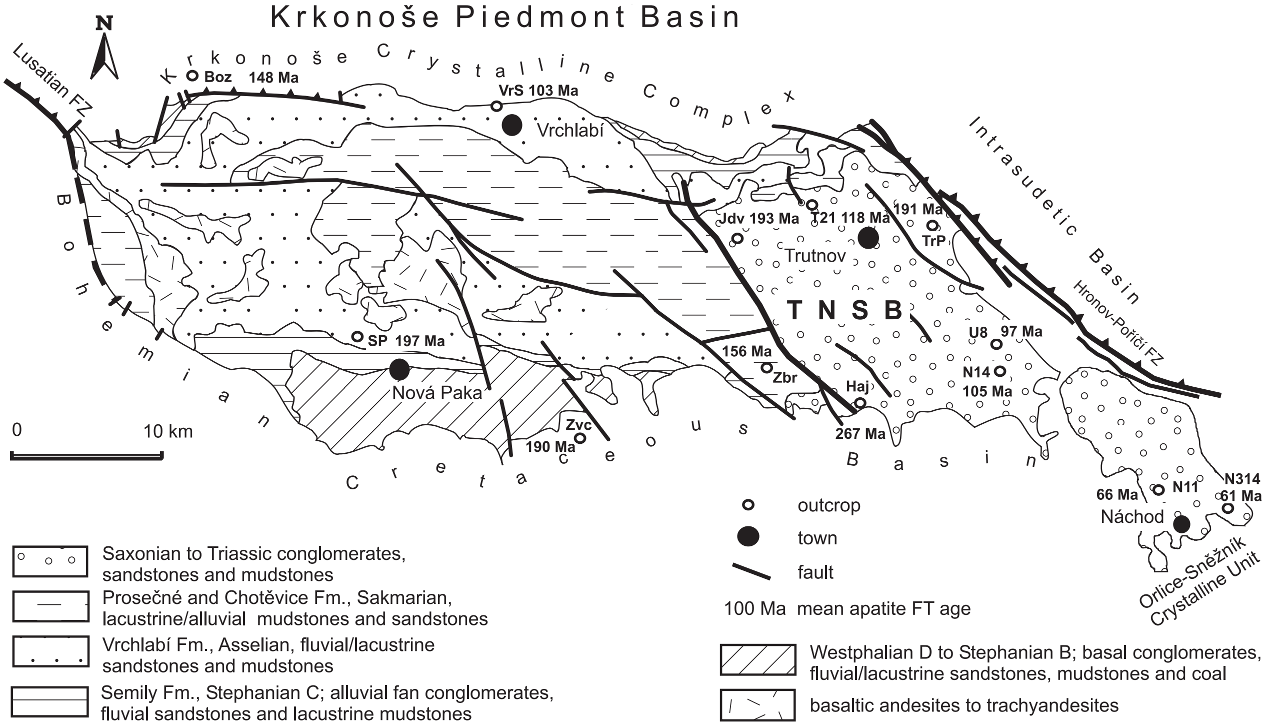
The Krkonoše Piedmont Basin (KPB) belongs to a system of post-orogenic extensional / transtensional basins of the Bohemian Massif. The age of the KPB sedimentary basin fill (non-marine red beds) spans between Westphalian D to Lower (or ?Middle) Triassic (ca. 305–240 Ma). The older parts of the KPB fill underwent partial deformation during the formation of the Trutnov-Náchod sub-basin (TNSB, Saxonian-Triassic age). Post-Variscan history of the KPB was documented only by indirect evidence: 1) post-Variscan left-lateral brittle to semi-brittle kinematics on the West Sudetic fault zones (Aleksandrowski et al. 1997), 2) Palaeogene deformation of the Lusatian fault zone (Coubal 1990) and 3) Upper Cretaceous kinematics based on sedimentary depocenter migration (Uličný 2001). This study brings a new low-temperature geochronological evidence of the post-Carboniferous history of



■ **Fig. 1.** Geological sketch map with sample location.

the KPB even for intervals where stratigraphical and structural record is missing.

Fission-track (FT) apatite data (12 samples) yield ages indicating Upper Permian annealing below 120 °C (PAZ, partial annealing zone) of the Palaeozoic rocks (both crystalline rocks and Permo-Carboniferous sediments). Assuming present-day thermal gradient of 25 °C/km, an average burial is suggested deeper than 4 km. Time-temperature modelling reveals four significant cooling/uplift periods (from 120 °C to present-day conditions, see Fig. 1):

I. Early Jurassic (190–197 Ma), documented by activity on the Hronov-Poříčí Fault Zone, Pilníkov Fault, Nová Paka Anticline, and Zvičina Crystalline Unit
II. Late Jurassic (148–156 Ma), uplift of the Chotěvice Block, and Železný Brod Crystalline Unit
III. Early/Late Cretaceous (Aptian/Cenomanian, 97–118 Ma), fault activity in the central and northern part of the TNSB, and northern part of the KPB
IV. Late Cretaceous/Paleocene (61–66 Ma), fault activity in the south-eastern part of the TNSB

Lower (?Middle) Triassic sandstone sample from the southern part of TNSB differs from all other samples. Mean detrital apatite FT age is older (ca. 267 ± 23 Ma, 1 sigma) than sedimentary age (ca. 240 Ma), which is interpreted as apatite annealing (below 120 °C) in sedimentary source areas. FT data modelling reveals another fast subsidence event (300 m/Ma, which would produce burial up to 2–3 km assuming thermal gradient 30 °C/km) in Early/Middle Triassic (ca. 250–240 Ma).

Oligocene/Miocene fast uplift (5–30 Ma, 100–150 m/Ma) is a common feature for all studied samples.

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The Contact Zone between the ALCAPA and Tisza-Dacia Mega-Tectonic Units of Northern Romania in the Light of New Paleomagnetic Data

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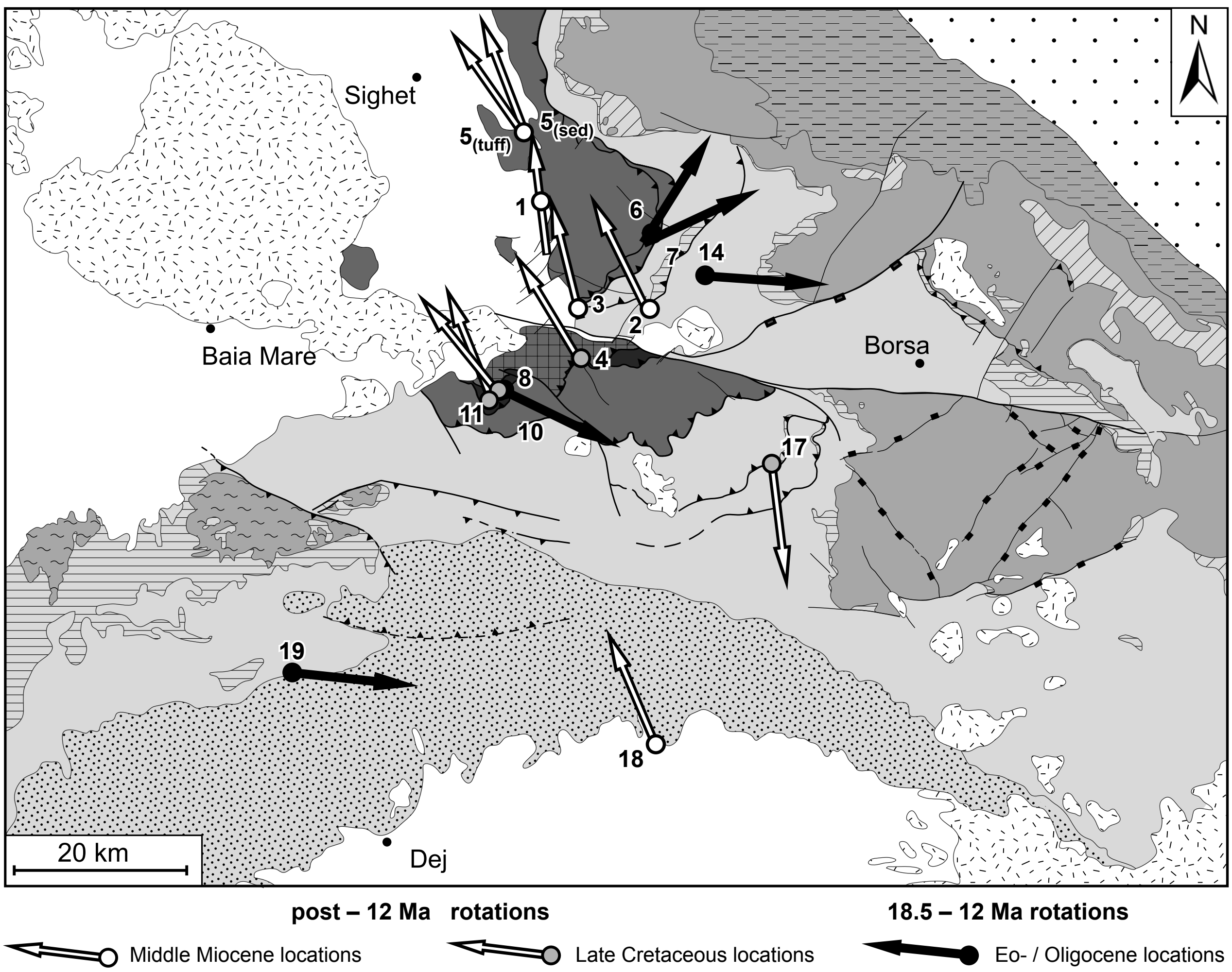
Paleomagnetic analyses were carried out on samples from 19 localities within two different mega-tectonic units in Northern Romania: Tisza-Dacia (11 localities) and ALCAPA (8 localities). The samples cover a range of different lithologies: (1) Late Cretaceous red-coloured marl to marly limestone, (2) Eo- Oligocene flysch sediments, and (3) Mid-Miocene (Langhian) tuffite (Dej tuff and related sediments). The Late Cretaceous and mid-Miocene specimens carry secondary paleomagnetic signals exhibiting a counterclockwise deflection of the paleo-declinations by some 30°, while the Eo-Oligocene localities indicate an overall clockwise deflected (between some 45° and >90°) paleo-declination with respect to present-day north. Clockwise rotation postdates the age of sedimentation (Lower Oligocene), as well as (at least partially) thrusting of the Pienides onto the Tisza-Dacia mega-tectonic unit, which occurred between 20.5 and 18.5 Ma. Clockwise rotation predates post-12 Ma counterclockwise rotations inferred for the mid-Miocene localities.

Surprisingly the clockwise rotations of the first rotational stage affected not only the (par-) autochthonous sedimentary cover of



■ **Fig. 1.** Paleo-declinations, plotted relative to present-day north on a geological map. The secondary magnetizations of Mid-Miocene and Late Cretaceous localities indicate a post 12 Ma counter-clockwise rotation of about 30° (white arrows). The Eo-/Oligocene locations show consistent clockwise rotations that pre-date the counter-clockwise rotations (black arrows).

the Tisza-Dacia mega-tectonic unit, but also the allochthonous flysch nappes of the Pienides, i.e. the eastern tip of the ALCAPA mega-tectonic unit. Well-documented opposed rotation of the remainder of ALCAPA necessitates a detachment of this eastern tip of ALCAPA after 18.5 Ma. The most likely location for this detachment zone is along the margins of the Transcarpathian depression. During a second (post-12 Ma) stage, counterclockwise rotations of up to 30° affected the entire working area. Regarding timing and magnitude, these second stage rotations are similar to rotations documented for the East Slovak basin, but different from those reported from the South Apuseni Mountains and the Central and Inner West Carpathians located west of the East Slovak basin.

First Paleomagnetic Results from the Oligocene Sediments of the Silesian Nappe, Western Outer Carpathians

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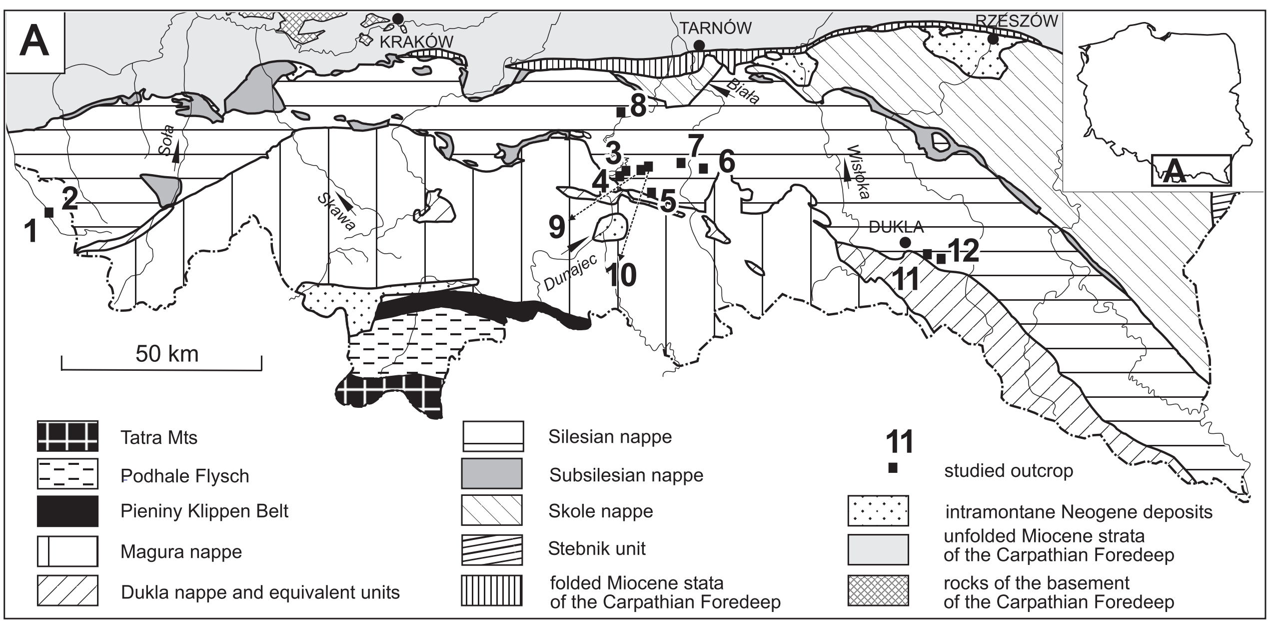
Western Outer Carpathians consist of several north-verging Nappes. The Silesian Nappe, the subject of the present study, is situated between tectonic units from which there are a number of Tertiary paleomagnetic results, the Magura Nappe and the molasse zone of the foredeep. It consists of Late Jurassic – Early Miocene rocks, mostly flysch sediments. The rocks of the Silesian Nappe form an arc, which is gently convex to the north. The regional fold axes are almost parallel to the trace of the frontal thrust of the nappe, thus they are WSW–ENE oriented in the west, E–W striking in the central segment and WNW–ESE oriented in the east. Bending followed regional folding and thrusting, after the mid-Miocene.

Krosno Formation consisting of shales and sandstones in different proportion represents the youngest strata of the Silesian Nappe. We sampled for paleomagnetic study the shaly members at 12 localities (and also a limestone bed at locality 2), distributed along the bend from the Czech–Polish boundary to Krosno. The sampled localities are of Oligocene age.

As a result of standard paleomagnetic measurements and evaluation, 10 localities yielded statistically good paleomagnetic results; for one locality (loc. 7) the statistical parameters are poor but the direction is still in line with the others. Tilt test on regional scale (including 11 localities) is positive, best statistics is obtained at 105 % untilting; the overall mean paleomagnetic direction is D = 310° I = 65° (k = 30 $\alpha 95$ = 8°). Although the remanence is of pre-folding (tilting) age, there is no correlation between individual declinations and the regional fold axes.

Magnetic fabric is dominantly foliated and is basically of sedimentary character. Yet lineation directions are E–W oriented (exceptions are locality 1, in the western and locality 9, in the central



■ **Fig.1.** Paleomagnetic sampling localities (1–12) in the Silesian Nappe.

segment of the Silesian unit). Variation in the lineation directions is not dependent on the orientation of regional fold axes.

The paleomagnetic and magnetic anisotropy results of the present study suggest that the Silesian Nappe rotated en block in a stress field of uniform direction, after Oligocene. The rotation angle is about 50 °, the sense is counterclockwise. As the paleomagnetic signals are of pre-folding/tilting age, our results do not support the bending origin of the systematic trend in the orientation of the regional fold axes.

Compared to the paleomagnetic results for the Magura Nappe, the present results are superior, since the paleomagnetic signal is better defined and is of pre-folding/tilting age. Yet, the overall rotations are the same. Between the two Nappe systems, there is the Dukla Nappe, which have to be studied in the future before we can explicitly state that Magura and Dukla Nappes rotated as one tectonic body.

An other important aspect of the present results is the originally stable European connection of the study area, since the results characterize the hinterland of the molasse basin, where rotation measured was only 30 °counterclockwise. In this case, there are also missing links, i.e. the Miocene rotations of the Subsilesian and Skole Nappes are not constrained paleomagnetically. Future studies will show if the rotation angle gradually changes from the Silesian Nappe system towards the molasse zone or the change is abrupt at the northern front of the Western Outer Carpathians.

Dating of the Gneiss Clasts from Gródek at the Jezioro Rożnowskie Lake (the Silesian Unit, SE Poland) Based on U-Pb Method

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Zircons from the gravel size gneiss clasts collected in Gródek at the Jezioro Rożnowskie Lake (the Silesian Unit, Western Outer Carpathians) were dated using U-Pb method with the aid of CL images. Previously other geochronological investigations of these samples were reported by Jacher-Śliwczyńska (2004) and Michalik et al. (2004).

Analyses were performed on separated zircon grains from four gneiss samples with use of LA-ICP-MS at the University of Arizona (Tucson, AZ, USA). Cores and rims or tips of zircons were analyzed when possible. 139 analyses were conducted, of which 63 are from cores and 76 are from rims and tips.

Gneisses are composed of quartz, K-feldspar, plagioclase, muscovite and biotite. Fe-Ca garnet, zircon, epidote, apatite, monazite, uraninite, Fe oxides, rutile, pyrite and zinc sulphides are present as accessory minerals. Detailed information about microstructure, alterations and chemical composition of selected minerals as well as whole rocks are given by Michalik et al. (2004).

Zircon forms euhedral to anhedral, doubly-terminated prismatic, <150 μm in size crystals. All grains exhibit complex zoning pattern. Xenocrystic cores (rarely subrounded) mantled by newly grown magmatic zircon commonly occur. Presence of fractures or inclusions is common. U content and U/Th ratio are 75–10092 ppm and 0.4–31.4, respectively, for cores, and 64–4597 ppm and 0.9 to 14.3 for rims. Cores as well as rims have experienced Pb loss.

Cores yield wide spectrum of Precambrian ages – ca. 1250 up to ca. 2747 Ma. Younger ages were obtained from rims. Tight clusters of eight analyses from sample JR-6 and seven analyses from sample JR-12 indicate ages 533 ± 19 Ma and ca. 572 Ma, respectively.

Zircon shapes and internal structures indicate their magmatic origin. Geochronological results are related to at least several magmatic events: older – Precambrian, and younger – Cadomian – early Caledonian. These results are particularly consistent with monazite CHIME ages of the same samples reported by Michalik et al. (2004) and indicate age of the protolith of gneisses. That there are no Variscan zircon U-Pb ages, that might be correlated with previously provided micas K-Ar ages (e.g. Poprawa et al. 2004) or monazite CHIME ages (e.g. Michalik et al. 2004, Poprawa et al. 2005).

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Traces of the Pre-Variscan Tectono-Thermal Event in Rocks of the Orlica-Śnieżnik Dome

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The sequence of deformations, established for the Stronie formation and referred to the history the Orlica-Śnieżnik Dome (OSD) gneisses could be interpreted as one continuous process starting from subhorizontal shortening caused by collision between two different crustal units forming East and West Sudetes, and subsequent orogenic uplift with subvertical shortening and flattening strain (Murtezi and Jastrzębski 2004, Murtezi 2005). Shortening and associated constriction of the first stage (D1) led to the development of upright folds with N-S trending axes in rocks of the Stronie formation and to constrictional stretching along the N-S direction of the Śnieżnik gneisses. Intrusion of the Śnieżnik metagranites took place in the extensional environment at c. 500 Ma (Turniak et al. 2000), Sharp intrusive contacts of metagranites with angularly unconformable relations to the schistosity of the enveloping metasedimentary rocks indicates that the Stronie formation was already deformed prior to the intrusion of the Śnieżnik gneiss protolith. At this point it is necessary to consider the possibility of existence of any deformational episodes (D0) that affected rocks of the Stronie formation prior to D1. Structural record connected with such an early deformational event (or sequence of events) within rocks of the Stronie formation is strongly obliterated by the later developed sequence of structures and in result very difficult to distinguish. As the Sudetes constitute a mosaic of terranes it is possible that early deformation and metamorphism of the studied rocks resulted from the interaction between crustal units later amalgamated to form the Sudetic part of the Bohemian Massif. At the opposite sides of the Sudetes are situated two crustal blocks generally formed of the Cadomian basement intruded by the lower-Paleozoic plutons, these units are: the Lausitz-Izera block on the west and the Bruno-Vistulicum on the east. Units situated in between form a mosaic in which different authors distinguish different number of terranes marking out their frontiers in different places. According to one of the newest interpretation of terrane distribution in the Sudetes proposed by Aleksandrowski and Mazur (2002), the OSD as a part of the Moldanubian (Gföhl) terrane took part in the bilateral collision with the Tépla-Barrandian terrane on the west and with the Moravian and Brunovistulian (Silesian) terranes on the east. Postulated by these authors, intra-Devonian subduction of the Central Sudetic oceanic basin and finally collision of the OSD as a part of the Moldanubian domain with the units adjacent to the west (Cadomian basement of the Tepla-Barrandian) can be taken into account as a process responsible for buttressing activity of the OSD and piling-up of crustal slices forming now the Silesian domain (SD). The OSD had to collide with units situated to the W and E as a previously consolidated crustal block. In this view, the western margin of the OSD constitutes a major discontinuity (a terrane boundary) and so, the OSD would have occupied the more internal part of the collision zone than the SD. It is in disagreement with the observed in rocks of the OSD decrease of metamorphic grade (from maximum at the boundary with the SD) towards the contact with the Nové Město Belt in the SW. Under this circumstances it may be the case that a suture zone between the crustal fragment represented by the OSD and units situated to the W should be located further westward, beyond the Nové Město Belt.

Basing on the comparison of the P-T-d record of rocks from the OSD and the SD, the earliest recognisable (having still distinguishable sequence of structures in rocks of the OSD) stage of deformation – D1, can be classify as the event that proceeds the Variscan collision of those two units (can result from the western collision of the OSD or even be a pre-Variscan). Nevertheless, this event starts the Variscan P-T-d path proposed for rocks of the OSD, considered as a record of the collision between the OSD and the SD. Fact that this event has no clear structural record in rocks of the SD can be explained by the lateral movements along the N-S direction on the late stage of the collision, shifting primarily colliding parts of the West Sudetes and the Bruno-Vistulicum. Despite these facts, lithotectonic architecture of the OSD provides some evidence on the existence of deformation that precedes the emplacement of the 500 Ma granite having intrusive discordant contacts. Inclusions of the gneissic enclaves in the 500 Ma metagranite, described by Grześkowiak and Żelaźniewicz (2002) in the Międzygórze anticlinorium can indicate on an older deformation of the surrounding rocks (probably Cadomian basement and Stronie formation volcano-sedimentary cover). This early tectono-metamorphic imprint can be a trace of a Cambro-Ordovician tectono-thermal event connected with the separation of a Variscan terrane assemblage due to progressive subduction and rifting at the northern margin of Gondwana continent. An extensional environment created during this event could be the location for the postulated early deformation and metamorphism of the Stronie formation. The Śnieżnik metagranites would have been emplaced into such, already deformed, supracrustal cover. Similarly in rocks of the Silesian domain Schulmann and Gayer (2000) distinguish at least one pre-Variscan tectonic event.

Significant improvement in the isotopic age evidence is required for the further progress in revealing the pre-Variscan history of the OSD and other Sudetic units. Despite the fact that the today observed tectonic architecture of the OSD is a result of a Variscan collision (or series of collisions) between Sudetic terranes, it seems reasonable to conclude that development of the earliest structures observed in rocks of the Stronie formation at in part of the OSD gneisses took place before Variscan orogeny.

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Thrusts and Folds in the Neyriz Ophiolite and Associated Rocks, Iran

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As the youngest collisional orogenic belt, the Zagros orogenic belt, has caused widespread folding and thrusting relative to crust thickening and uplifting, and was recognized generally by the international geoscientific community (e.g., Ricou; Alavi 1980, 1994; Pamic, et al.). The southeastern part of Zagros orogenic belt is located in around of Neyriz region and comprises a number of NW--striking thrust faults, ductile-brittle shear zones, folds, ophiolite, ophiolite mélange and tectonic slices. Ophiolites are major features of an orogenic belt and they are dominantly oceanic crust and mantle emplaced by collision of mantle-rooted thrust fault with a continental margin or island arc. Ophiolite nappes thus represent remnants of lithospheric plates; their basal thrusts are remnants of subduction zones. The Neyriz ophiolite is part of the upper mantle and Tethyan oceanic crust that stretched along the Zagros suture. The ophiolite consists of several small and large thrust sheets each having its own internal layering. Rocks in this complex are include compositionally layered, serpentinized peridotites (mainly harzburgites, lherzolites, dunites and pyroxenites), both massive and layered gabbros, sheeted dikes, pillow lavas, and thinly and uniformly bedded Jurassic to Upper Cretaceous radiolarian cherts interbedded with red lutites and thin beds of pelagic limestones (Nadimi 1999, 2002). Compositionally layered amphibolites and metamorphosed sedimentary rocks as schists have observed locally. In the central part of the Neyriz area, the ophiolite thrust sheets transported over the Upper Cretaceous shallow-water shelf/platform carbonates and overlain unconformably by the uppermost Cretaceous limestones (Tarbour Formation). The initial emplacement as slivers of Neo-Tethyan oceanic crust over the Afro-Arabian continental shelf must have been a Cenomanian-Maastrichtian event (Alavi 1980, 1994). Sheets of recrystallized limestones that strongly sheared and/or brecciated at their contacts are also presents.

In cross section, major thrust sheets show an imbricate pattern. Within each thrust sheet, rocks intensely folded and sheared. Folds are of angular parallel type and disharmonic; shear zones are brittle with discrete, anastomosing shear planes and associated cataclasites. Structurally, the northeastern high mountains, the ophiolite, and southwestern units are characterized by NW-SE trending folds and thrusts, which exhibit shortening in a NE-SW direction and show evident southwestward thrusting in general. The northeastern limit of the ophiolite thrust sheets distinguished by a system of breaching thrusts, which has resulted in transportation of the Mesozoic continental shelf sedimentary rocks over the ophiolites and severe crushing, and intermingling of various rock units. And also the southwestern limit of the ophiolite thrust sheets distinguished by a system of nappes and interesting folds in beds of pelagic limestone and radiolarites, which has resulted in obduction of the ophiolite and associated sedimentary rocks over the Mesozoic continental shelf sedimentary rocks of the Arabian platform. The radiolarites and pelagic limestones have shortened about 35–40 % during folding.

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Contrasting Metamorphic Evolution of HP Rocks in the Gföhl Unit of the Kutná Hora Crystalline Complex and the Moldanubian Zone in Austria

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The uppermost part (Běstvina and Malín formations) of the Kutná Hora Crystalline Complex is due to presence of HP/HT granulites, garnet peridotites and eclogites are generally correlated with the Gföhl complex in the Moldanubian zone. We have studied felsic granulites, kyanite-bearing migmatites and migmatitic gneisses of the Gföhl-relted unit in the Kutná Hora Crystalline Complex and in the Moldanubian zone of Waldviertel in Austria. In order to analyse PT evolution of individual lithologies we have constructed pseudosections in the NCKFMASH system using the THERMOCALC software (Powell et al. 1998). The calculated isopleths in the pseudosections were compared with composition of garnet, plagioclase and K-feldspar in different stages of the rock evolution. The overall PT trajectory was further improved by calculation of "Average pT" method (Powell et al. 1998).

Granulites and migmatites occuring in the uppermost parts of the Kutná Hora Crystalline Complex, are characterized by the presence of kyanite, garnet and feldspars and gneisses contain also white mica. The migmatite contains clusters of muscovite and biotite with small grains of garnet and kyanite. Textural relations indicate that biotite was formed during the late stage of metamorphism. It replaces or rims muscovite being in textural equilibrium with garnet and kyanite. Garnet in both migmatite and in gneiss is homogeneous and only weak retrograde zoning was observed in the rims of garnet from granulite. Garnet in migmatitic gneiss is rich in Fe (Alm_{77-83}, Py_{10-13},Grs_{2-6}). Relatively high-Mg garnet is present in granulite (Alm_{56-60}, Py_{28-32},Grs_{5-11}). Analysed muscovite has relatively high phengite component with Si = 3.2 a.p.f.u. Plagioclase is usually rich in Na and the anorthite content ranges between 6–11 mol% in migmatitic gneiss. Granulite has nearly pure albite with $An_{0.07-0.09}$.

The HP/HT metamorphic conditions of 875 ± 95 °C and 15.6 ± 1.4 kbar were calculated, using the average PT (Powell et al. 1998) for the assemblage Ky-Grt-Plg-Kfs-Ms-Qtz ($X_{H2O\ in\ melt}$ = 0.5) in the Malín and Běstvina migmatitic gneisses. The Běstvina granulite gave temperature and pressure of 831 ± 53 °C and 16.5 ± 1.8 kbar ($X_{H2O\ in\ melt}$ = 0.6) for kyanite, garnet core, perthitic feldspars and biotite in garnet. Metamorphic PT conditions for the MP/LT stage were calculated from the matrix biotite, garnet rims and the recrystallized grains of plagioclase, K-feldspar and quartz. The results yielded PT conditions of 712 ± 39 °C and 10.6 ± 1.8 kbar for migmatites and 705 ± 97 °C and 14.4 ± 2.1 kbar for granulites. The calculated PT conditions, consistent with the lack of sillimanite, suggest that the retrogression occurred still at high pressure in the kyanite stability field.

PT conditions obtained for the HP/HT metamorphic stage in felsic granulites from the Moldanubian zone in Austria correspond to 912 °C ± 54 °C and 13,7 kbar ± 1,5 kbar. Their exhumation to the middle custal levels was accompanied by formation of LP/HT mineral assemblages that yield pressure and temperature of 890 °C ± 72 °C and 8,5 kbar ± 2,0 kbar. The Gföhl migmatitic gneisses contain relics of kyanite armoured within feldspars indicating the earlier HP metamorphic stage. The dominant assemblage within the late foliation is Sill-Grt-Plg-Kfs-Bt determining conditions of the metamorphic overprint at 877 °C ± 69 °C and 6,7 kbar ± 1,7 kbar. The development of spinel in the ky-kfs granulites indicates, that the metamorphism continued to shallow depths at still high temperature.

According to the observed mineral assemblages and calculated PT conditions, exhumation of the Kutná Hora Crystalline Complex rocks occurred at different PT path comparing to that of the Moldanubian zone in Austria. In the Moldanubian zone, decompressional anatexis resulted in stabilization of LP mineral assemblages with sillimanite and spinel in migmatites and LP overprint of eclogites and HP granulites. In contrast, the Malín and Běstvina migmatites and felsic granulites are characterized by a PT path implying almost isobaric cooling. The presence of kyanite in leucosomes of migmatites suggests that anatexis occurred already during the high-pressure stage. The kyanite-bearing migmatites and HP felsic granulites of the Malín and Běstvina formations may represent well-preserved segment of extensively granulitized lower crust of the Variscan orogenic root, which was not affected by late LP overprint probably due to very high exhumation and cooling rates.

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Geodynamic Significance of Late Cretaceous Lamprophyres from the Carpathian-Pannonian Region

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Late Cretaceous lamprophyres as dyke swarms have been found in two major locations within the Carpathian-Pannonian Region: 1) Villány Mts situated in Tisza megaunit – S Hungary and 2) NE Transdanubia situated in Alcapa megaunit. Previous studies (Szabó et al. 1993, Nédli 2004, Nédli and M. Tóth, submitted) showed that these dyke swarms petrographically and geochemically are similar and their melts probably derived from the same or very similar asthenospheric mantle sources. The lamprophyric melts originated from significant depth in the upper mantle and show no signs of crustal contamination. In this way they carry specific geochemical, petrologic information about the Late Cretaceous lithosphere-asthenosphere beneath the region. Their xenolith content reveals also the physical-chemical characteristics and processes of the pre-Paleogene crust and mantle. Considering that these dykes are situated on two different microplates (Tisza and Alcapa), their study can contribute to the geodynamic evolution of the microplates composing the Carpathian-Pannonian Region.

Dykes studied petrographically are alkali lamprophyres with porphyritic texture containing olivine and clinopyroxene phenocrysts and a fine-grained, pyroxene-rich groundmass. Whole rock K/Ar data for the dykes indicate Late Cretaceous age (Dunkl 1991, Molnár and Szederkényi 1996). Based on the geochemical characteristics, these dykes are thought to have been originated from an enriched (EM II-type) garnet lherzolite mantle source by low degree partial melting. Significant negative Nb-Ta anomaly, extreme LILE and LREE enrichment, moderate enrichment in HFS elements (Fig. 1) and geochemical similarity to other EM-type mantle sources worldwide (Weaver et al. 1986) suggest the subduction-related origin of the enriched component of the source).

Both dykes contain abundant xenocrysts and xenoliths of lower crust or upper mantle origin (Szabó 1985, Kubovics et al. 1989, Molnár and Szederkényi 1996, Nédli and M.Tóth 1999). Xenoliths from Villány Mts. are mainly four phase spinel lherzolites, whereas in Transdanubian lamprophyres occur a large variety of xenoliths (e.g. pyroxenites, lherzolites, websterites, wehrlites). Some samples from both localities contain OH-bearing minerals suggesting that hydrous metasomatism effected the source mantle region of the xenoliths. Both xenolith series reveal a deformed and re-equilibrated subcontinental mantle.

The significant chemical, petrographic, petrogenetic and age similarities of lamprophyres from the Villány Mts. and NE Transdanubia suggest similar or the same mantle source for these igneous activities. Their mantle source region must be of asthenospheric origin, due to the unequivocally different lithospheric composition of the Alcapa and Tisza units, separated by the Mid-Hungarian Zone (e.g. Csontos and Nagymarosy 1998, Kovács et al. 2006). These lamprophyres carry the evidence of the presence of similar or the same subduction-related, enriched asthenospheric subcontinental mantle component beneath the Tisza and Alcapa microplates in the Late-Cretaceous, which were located about 400 km from each other in the Late Cretaceous (Csontos and Vörös 2004).

On reconstructions of Late Cretaceous structural elements in the Alpine-Carpathian-Pannonian area (Csontos and Vörös 2004), it is clear that the localities containing lamprophyre dykes are aligned along the Peri-Adriatic Fault, with the Villány Mts near Beograd, whereas the Transdanubian Region is situated between the Southern and Central Alps. In time and space the closest suspected subductions in the surroundings can be related to the Budva-Pindos or Meliata-Vardar oceans. Extension stresses related to the Budva-Pindos subduction, started in the Mid-Late Cretaceous (Csontos and Vörös 2004), could affected the formation and ascending of the lamprophyric melts in the Tisza sector. However, this subduction seems too late to cause the upper mantle deformation and metasomatism detected in the xenoliths; at least 24 Ma years is necessary for an extensive metasomatism of the mantle wedge (Greya et al. 2002). This process probably could be due to the former closure of the Meliata-Vardar ocean, ended by the Middle Cretaceous (Csontos and Vörös 2004). Even if these subductions could be reasonable candidates for mantle enrichment and deformation in the case of Villány Mts, they can hardly explain the mantle enrichment beneath the Transdanubian region, not being extended west-bound until the Alcapa microplate (Csontos and Vörös 2004). This suggests that, in this case, at least two separated subduction-related enriched mantle domains existed beneath the Alpine-Carpathian-Pannonian area in the Late Mesozoic. Explain the subduction-related deformation and enrichment beneath the Alcapa microplate is even more difficult, the Penninic-ocean seems to be a reliable candidate. However, dating of its opening and closure is uncertain and there have been no direct subduction-related magmatic traces of its subduction detected previously.

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Tertiary Stress Field Evolution in the Eastern Part of the Bükk Mountains, NE Hungary

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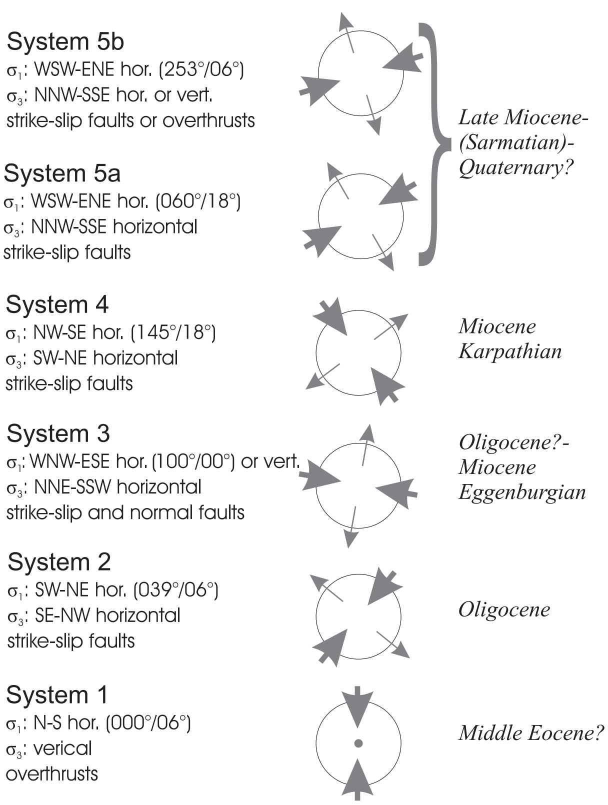
The Bükk Mts. consists of Paleozoic and Mezozoic rocks surrounded by Tertiary sediments and volcanics of the Pannonian basin, so it can be regarded as an uplifted and exhumated part of the basement. According to the 43±8 MY cooling ages of Dunkl et al. (1994) (apatite fission track, cooling under 70°C) and the Late Eocene age of the oldest known Tertiary sedimentary rocks (limestone with Nummulites) of the Bükk Mts. the first Tertiary exhumation occurred in the Middle Eocene. After some periods of sedimentation and volcanism the present state is the almost perfect exhumation of the basement rocks with well-defined edges around and with few remnants of the cover inside the mountains. These basement rocks on the steep slopes of the mountains offer several outcrops with well-preserved striation on fault planes of variable orientation. Some fault planes were active in two or more phases of the deformation history, so based on the observed overprinting relations the phases can be separated and the relative ages of the movements also can be stated. These features are very favourable for fault slip analysis and stress inversion. On the other hand, lack of the coeval sediments prevents the observer from direct dating of the deformation phases.

In my analysis I used the fault slip data of about 70 outcrops (22 of those with more than 10 unambigous data of different plane-striae pairs, 655 such pairs altogether) from the eastern part of the Bükk Mts, on a cca. 80 km^2 area collected during the last two years. I measured the orientation of the fault planes and the slickenlines (striae or slickolites) on these planes. I observed the shear sense indicators and overprinting relations where it was possible. Based on these, I defined sets of striae which could be formed during a single deformation phase with a nearly constant stress field. For stress inversion I used the p-T dihedra method (Angelier 1984) because the necessary inputs of this method were suitable for the collected data set and, because of the limited validity of the assumptions needed for the inversion, there is no possibility to reach a level of accuracy higher than provided by this method. Construction was carried out by the StereoNett 2.46 software of J. Duyster.



■ **Fig. 1.** Movement systems with constructed principal stress directions, characteristic fault types and assumed ages.

The sets of striae were grouped into "movement systems" based on the kinematic similarity of the movements on similarly oriented planes of each outcrop. I strived to minimize the number of these movement systems, so the ones with a difference less than 10° between the estimated principal stress directions were merged if overprinting relations did not contradict it. In this way there remained five movement systems which were active in the whole area and possibly beyond it; some other systems occur only on some adjoining sites and seem to have a local role only. As direct dating of the movement systems was impossible, in the last step I tried to fit in the data with the regional stress field analyses (Márton – Fodor 1995) and with the sedimentary and volcanic records (Pelikán 2002), in this way associating an assumed age to each of them. This step was made from the youngest system towards the previous ones, so I introduce them in an inverse chronological order (fig 1).

System 5 characterized by E-W shortening and WNW-ESE sinistral strike-slip faulting is connected with the faults of the last uplift and the edges of the present Bükk Mts (Németh 2005), so its assumed age is Late Miocene-Pliocene; it may have been active even in the Quaternary. System 4 characterized here by dextral faulting on E-W strike correlates well with the Karpathian state of the regional stress field according to Márton and Fodor (1995). System 3 is characterized by NNE-SSW extension and correlates with the Eggenburgian system of Márton and Fodor (1995). System 2 characterized by NE-SW directed highest principal stress and corresponding strike-slip faulting cannot be correlated with known regional systems, but it is definitely older than the overprinting System 3 so its assumed age is Oligocene. System 1 is characterized by N-S shortening and large-scale overthrusts with considerable amount of uplift. It seems to be connected with zigzag folds and a deformation style change, so I assume it to be coeval with the Middle Eocene cooling age determined by Dunkl et al. (1994). Brittle deformation elements older than these always are associated with folding.

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Geochemistry of Metabasites in the Stronie Group and Nové Město Group, the Orlica-Śnieżnik Dome, West Sudetes

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The Stronie Group (SG) supracrustal rocks, together with a variety of gneisses enclosing small bodies of granulites and eclogites, form the middle and eastern parts of the Orlica-Śnieżnik dome (OSD). The westernmost flank of the OSD is formed by the Nové Město Group (NMG). Metabasites within the Stronie Group appear as small lensoid or irregular bodies, rarely exceeding 0.5 km^2, associated with mid-grade metamorphosed metapelitic schists, subordinate marbles, quarzites and acid metavolcanic rocks. They form two groups: alkali basalts of WPB type and subalkali tholeiites similar to MORB, with typical ratios of characteristic elements. No low-Ti metabasites has been identified. Alkali basalts are characterized by Nb/Y > 1.5, Ti/V > 50, Zr/Y > 4, Zr/Nb < 5. In metatholeiites the ratios are: Nb/Y < 0.7, Ti/V < 50, Zr/Y < 4, Zr/Nb > 20. Metabasites of WPB type occur as laminated biotite amphibolites which pass laterally to mica schists or to calcareous schists next to marble bodies. At least part of these amphibolites likely originate from tuffitic protolith merged with clastic rocks and represent pyroclastic products of early stage of a continental rift volcanism. MORB-like metatholeiites are more frequent, though their occurences are widely separated and differ in size. They have fine-grained gabbroic or diabasic protoliths interpreted as hypabysal lava bodies or dykes feeding individual volcanoes during more advanced rifting. Marbles contain problematic fossils of early Cambrian age (e. g. Gunia 1984), which suggests similar age for the interlayered metatuffs. U-Pb SHRIMP analyses of zircons from acid metavolcanics throughout the SG yielded similar ages of ~500 Ma (Murtezi and Fanning 2005). Accordingly the same age is assumed for the SG metabasites.

In the NMG, Domečka and Opletal (1980) and Opletal et al. (1990) identified subalkalic tholeiites interpreted as ocean floor basalts and co-magmatic calc-alkaline felsic volcanites, apparently different from (meta)basites of the Stronie Group at the OSD core. Our studies show that the NMG metabasites, generally similar to the main series tholeiites of Floyd et al. (1996, 2000), can be subdivided into 3 different types: within-plate tholeiites (WPT) which dominate in the region, less common MORB-like tholeiites, and scarce Ti-tholeiites, based on Zr/Y, Ti/Y, Ti/V, Zr/Nb, $(La/Yb)_N$ and $(La/Sm)_N$ ratios and abundances of HFSE and REE. In the field, the WP-tholeiites (Zr/Y ca. 3.57 to 5.70; Ti/Y ca. 290–450) and MORB-like tholeiites (Zr/Y < 3.5; Ti/Y < 320) form roughly parallel belts concordant with the regional meridional strike of the main lithological boundaries and that type of their distribution continues E-ward to the Stronie Group. It is observed in the field that the WP tholeiites and Ti-tholeiites are accompanied by felsic volcanites which injected the former and the MORB-like tholeiites intruded other mafites. Both field evidence and geochemistry confirm earlier conclusion of Opletal et al. (1990) who concluded that the bimodal magmatic rocks were derived from the same magma source. U-Pb SHRIMP datings of acid metavolcanics in the NMG reveal an age of ~500 Ma, identical with that determined for the SG rocks.

In the NMG, the WPT and MORB-like metabasites have rather uniform isotopic signature εNd(500) ~5–7 suggestive of similar mantle source and possibly weak contamination with crustal material. A MORB source transitional between N-MORB and E-MORB and an enriched mantle source are suggested. A N-MORB-like source might possibly be mixed with an enriched OIB-like source (plume). For the WPT, both the isotopic and characteristic elemental ratios may also point to mild contamination due to some crustal admixtures. The SG MORB-like metabasites have similar isotopic signature to the NMG metatholeiites εNd(500) ~5–6. They come from a common depleted mantle source. Isotopic signature εNd(500) ~2–4 of the SG alkali basalts suggest derivation of the alkali magmas from a different more enriched mantle source than the source of the metatholeiites.

The mafic magmatism recorded in the NMG and SG likely developed in a relatively short period of Mid-Late Cambrian magmatic episode in a continental rift-related (probably back-arc) setting which reached the stage of new oceanic crust production represented by MORB-like metabasites. The scarce alkali basalts from the eastern part of the SG area are assigned to earlier stages of the rifting. Palaeogeographic assignment of the rift basin needs dating of its sedimentary-volcanogenic infilling. Further U-Pb datings are under way.

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Granitoids from the Ditrău Alkaline Massif, Transylvania, Romania

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Re-examination of a part of the Ditrău Alkaline Massif [DAM], identified earlier as granite, has revealed that it has a complex petrography, containing monzonites, syenites and granites. These rocks are peraluminous and peralkaline, moderately to highly fractionated. The most fractionated are the oversaturated rocks (granites) representing the subalkaline branch of the magmatic evolutionary trend, while alkaline branch contains quartz-monzonites, quartz-syenites, syenites and probably nepheline syenites.

According to the discrimination diagrams, Ditrău granites are A1-type and within-plate granites. K/Ar ages obtained from amphibole and biotite vary between 217.6 ± 8.3 and 196.3 ± 7.4 Ma, which fall close to the age of nepheline syenites (216.0±8.1 Ma) and hornblendites (226.0 ± 9.6 Ma) from the rocks of the massif which also support the coeval relations between them.

Occurrence of characteristic accessory minerals, zircon morphology and data from geochemistry and microthermometry suggest a mantle derived parental magma from which the series of derivates were formed by fractionation, differentiation and contamination processes in the upper crust in an extensional, within-plate tectonic setting.

Introduction

The mineral composition, structure and magmatic evolution of the DAM (Transylvania, Romania) have been discussed for more than 150 years. During this time numerous researchers attempted to explain the genesis of the massif. Summarising extensive previous research, two possible hypotheses have emerged concerning the origin of the granite: (1) it is either a differentiation product of mantle-derived ultramafic melts or (2) it was formed by crustal contamination process. The purpose of this paper is to augment understanding of the formation of the granitoid rocks in the DAM by using zircon morphology, geochemistry, fluid inclusion analysis and geochronology.

Ditrău Alkaline Massif has a complex petrography: a wide variety of igneous rocks have been described from the DAM. Within a short distance from a peridotite to granite, nepheline syenite type rocks crop out. The granitoid rocks occur in many places in the DAM, the largest body occurring in the north-eastern part of the massif.

Petrography of the granitoid rocks

The examined rocks are dominantly leucocratic, pale-grey, pale-pink or reddish in colour. Modal analyses, mineral composition and textural studies enabled a clear differentiation of the following rock types: syeno- and monzogranites; syenites, quartz-syenites, alkali-syenites; monzonites, quartz-monzonites.

On the basis of the mafic minerals, two well-separated groups can be distinguished within the studied rocks. The first contains biotite±hornblende, whereas in the second displays alkali pyroxene (aegirine) and alkali amphibole (arfvedsonite) are the main mafic components. These two groups mean different chemical character (Shand 1947, Clarke 1992), the first one is metaluminious and the second one is peralkaline. The most frequent accessory minerals are apatite, zircon, sphene, allanite and opaque minerals.

The zircon crystals of the studied populations are dominantly transparent, colourless, pale-yellow and pale-brown or rarely reddish-brown in colour and all grains are euhedral. Their zonality shows more than one crystallisation phases. The most frequent subtypes (Pupin 1980) are: P_4, P_5, P_3, S_{19}, S_{20}, S_{24} és D in the examined rocks, plotting at the boundary of the subalkaline and alkaline fields which suggest that the zircon crystals were formed in a hyperalkaline or hypoaluminous geochemical environment.

Geochemistry

The examined rocks are moderately to highly fractionated. The SiO_2 contents ranges between 63.5–77.1 wt%. The calcium and magnesium contents are low: 0.1–0.9 and 0.1–0.6 wt% but sample AGK-6831 is less fractionated and contains 2.1 wt% CaO and 1.0 wt% MgO. The FeO*/MgO is relatively high: 4.6 to 10.7 wt%. The alkali contents are also high; K_2O varies between 4.7–6.5 and Na_2O ranges between 4.4 to 6.1 wt%. According to the Q-P diagram (Debon Le Fort 1983) the examined rocks are classified as granites, monzonites, quartz-syenites and syenites which separate two distinct groups regarding SiO_2 vs. (Na_2O+K_2O) diagram where plotted on the fields of alkaline and subalkaline series. Their chemical character is peraluminous and peralkaline but metaluminous also occurs in the A/CNK vs. A/NK diagram after Maniar and Piccoli (1989). Two separated groups of samples can be seen in the Harker variation diagrams which are correspond to the alkaline and subalkaline series.

The REE patterns of the examined granites can be charactерised by a moderately falling LREE part with marked negative Eu anomaly and a slightly lifting HREE part. The ratio of Eu/Eu* ranging from 0.02 to 0.48. On the multi-element variation diagram where element contents are normalized to chondrite shows negative Ba, Sr, Eu anomalies.

In the discrimination diagrams of Whalen et al. (1987), the examined rocks plot within the anorogenic or A-type field. Eby (1992) subdivided into two groups of A-type granites. According to Y-Ce-Nb and Y-3Ga-Nb diagrams, the examined rocks fall into A1 field which represents mantle differentiates from the same types of sources that produce ocean-island, intraplate and rift-zone magmas.

Fluid inclusion petrography

The studied primary fluid inclusions (Roedder 1984) can be found in the xenomorphic quartz crystals that are characteristic of DAM granites. All measured fluid inclusions are two-phase (liquid and vapour) inclusions with ice melting temperature (T_m ice) from –8.4 to –4.1 °C and homogenisation temperature (T_h) from 176 to 228 °C. Microthermometry results indicate that the fluid inclusions are saline in character and therefore can be modelled in a $NaCl$-H_2O system. Using T_m ice and T_h data the water salinity was calculated according to Bodnar (1993) equation. The diagrams of salinity frequencies and salinity as function of homogenisation temperature show a very large salinity distribution (2.07 to 17.52 wt%) with average value of 8.6 wt% NaCl equivalent.

Discussion and Conclusions

There is a magmatic evolutionary differentiation and fractionation relationship between the examined rocks: the oversaturated rocks (granites) represent the subalkaline branch and Qtz-monzonites, Qtz-syenites, syenites fit on alkaline branch of magmatic trend (Upton 1974).

The value of $(Eu/Eu^*)_{ch}$ shows the different degree of fractionation among the examined samples. The lowest value (0.10) represents the most fractionated sample which is monzogranite while the highest value (0.48) indicates slightly differentiated Qtz-monzonite sample. Nb/Ta ratio are varies between 13.2 and 32.3 which is a wide range, indicates heterogenites and can refer to fractionation and differentiation trend as well.

On the basis of the comparison of major and trace elements of the examined rocks with the typical granite types, the granitoid rocks of DAM have higher Al_2O_3, Na_2O, K_2O, Rb, Sr, Nb, Zr, Ga and lower MgO, CaO, Ba, Pb, Y, Ni, contents. These data suggest that the examined rock belong to A-type granite which is forced by discrimination diagrams. The examined rocks can be classified as A1 subgroup of the A-type granitoids. In the light of this result it is possible that the source of these rocks can be mantle derivatives.

Zircon crystals were formed in a high temperature and Pupin's classification scheme indicates that the host rock is a mantle-derived granitoid rocks.

Projecting fluid inclusion homogenisation temperature (Th) and salinity data on $NaCl$-H_2O P-T diagram and to the water-saturated alkaline granite solidus (Johannes 1984), proving that the development of the quartz crystals in granitoid melts took place between 640–680 °C temperature and 6.7–8.5 kbar pressures in crust conditions after magmatic differentiation.

New K/Ar ages from amphibole and biotite fractions from granites yielded variations between 217.6 ± 8.3 and 196.3 ± 7.4 Ma. These K/Ar ages of amphibole and biotite fractions correspond to that of the amphiboles of hornblendite, 237 ± 9.1 (Pál-Molnár 2000) which suggests the mantle derivated hornbelndite and granites are coeval.

We conclude, therefore, that the examined rocks represent mantle-derived material which was subsequently modified by differentiation and crustal contamination processes in a rifting continental regime.

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Tertiary Tectonics and Paleostress Reconstruction in the Central Carpathian Paleogene Basin (Orava Region)

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The studied area is situated in the Orava River valley and its vicinity. In term of orographic classification, the studied area appertain to the Chočské vrchy Mts., Podtatranská brázda depression, Skorušinské vrchy Mts., Oravská vrchovina Mts., and the south part of the Orava Basin. The most expressive elevation structure are Chočské vrchy Mts. (Veľký Choč hill 1661 m. n. m) and Skorušinské vrchy Mts. (Skorušiná hill 1314 m. n. m.). In the north, the studied area is limited by the strong morphological mountain range of the Oravská Magura Mts. The river belong to the Váh drainage area, from that predominantly to the local Orava drainage area.

The aims of this work was fault and fold deformation of the Paleogene sedimentary sequence of the Subtatras Group and the Neogene sedimentary sequence of the Orava – Novy Targ Basin. The main works were oriented to the fault – slip analysis and paleostress reconstruction during the Oligocene to Quaternary period.

The area of the southern and eastern Orava has complicated geological structure because is situated near contact zone between the Inner and Outer Western Carpathians. The area is affected by strong strike – slip deformation along the Peripieniny Lineament. Five basic tectonic superunits are present in the area of interest. In the direction from the south to the north, they are. (a) The area of the Chočské vrchy Mts., that composed of the Mesozoic complexes of the Tatric, Fatric and Hronic Nappe Units. The mountains is strictly limited by the Choč-Subtatric fault in the south. (b) The Subtatric Group formed by Paleogene sediments, situated between the Chočské vrchy Mts. and the Pieniny Klippen Belt. (c) The Pieniny Klippen Belt, a tectonically composite structure with the development of the Kysuce, Czorsztyn, Orava and Klape successions, formed by the Jurassic–Cretaceous rocks. In the north, the Pieniny Klippen Belt is limited by the Cretaceous and Paleogene rocks of the Outer Carpathians. (d) Neogene sediments, that belong to the Orava–Nový Targ Basin. (e) Quarternary sediments, belonging to several genetic types.

The fault-slip data has been computed by the TENSOR software package (Delvaux 1993, Delvaux and Sperner 2001) using right dihedra and inversion methods for the paleostress reconstructions. In this work, deformational fault and fold structures occurring in the Paleogene (the Borové, Huty, Zuberec, and Biely Potok Formations) and Neogene sediments in the Orava region were analysed. The Orava region was affected by a paleostress field which changed several times. The studied area was also strongly influenced by a NW-SE compressive tectonic regime which formed a NE-SW syncline structure. The chronology of the Neogene paleostress stages in this area is very complex because younger sediments than the Paleogene age are not preserved in the Orava region, except the Pliocene sediments of the Orava – Nowy Targ Basin. The deformational stages were separated on the basis of the relationship between sedimentary sequences and deformational structures, and the successive relationship between faults at the outcrop, as well. It was also detected by the paleostress reconstruction. The chronology of the determined tectonic regimes is as follows:

- The youngest deformational stage is characterized by the pure W-E extensive tectonic regime. This regime was identified in the Pliocene sediments of the Orava–Nowy Targ Depression and it was dated at the Upper Pliocene to Quaternary Period.
- The second deformational stage was developed during the pure NW-SE compressive tectonic regime and has been tenuously dated at the Upper Badenian to Sarmatian Period. This deformational stage is dominant over all studied localities.
- The third deformational stage was activated during the Upper Karpathian–Lower Badenian on the basis of its relationship with the previous stage. The pure NW-SE extensional tectonic regime activated this deformational stage and the evolution of normal faulting in this area.
- The fourth generation faults were formed during the pure NE-SW extensive tectonic regime. This deformational stage has been tenuously dated at the Otnangian–Lower Karpatian Period.
- The fifth deformational stage is characterised by the strike-slip tectonic regime generated during NNW-SSE compression and WNW-ESE extension. This stage was activated from the Egenburgian till to the Otnangian Period.
- The fault and fold structures of the sixth deformational stage were activated during the Upper Eocene to the Oligocene Period. This stage is characterised by the E-W compressive strike-slip tectonic regime.

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Tectonic Evolution of the Pieniny Klippen Belt and its Structural Relationships to the External and Central Western Carpathians

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The Pieniny Klippen Belt (PKB) is a narrow zone dominated by wrench tectonics (Nemčok and Nemčok 1994, Ratschbacher et al. 1993, Kováč and Hók 1996) that separates the External Western Carpathians (ECW – Flysch Belt, Tertiary accretionary wedge) from the Central Western Carpathians (CWC – Cretaceous nappe stack). The PKB involves predominantly Jurassic, Cretaceous and Paleogene sediments with enormously variable lithology and intricate internal structure (e.g. Andrusov 1931, 1938, 1953, Birkenmajer 1977, 1986, Mišík 1997). During more than a century of intense research these have been subdivided into numerous lithostratigraphic and tectonic units. However, the mutual structural relationships of various PKB units and the neighbouring tectonic zones remains a matter of controversy and no general agreement has been achieved even in some fundamental questions.

We present a series of generalized cross-sections through various segments of the PKB in Slovakia and Poland aimed at illustration of the internal structure of the PKB and its tectonic interactions with adjacent ECW and CWC zones (see Fig. 1 as an example). Sections are constructed based on regional geology using published maps and profiles, lithostratigraphic contents of tectonic units and our own structural data (e.g. Jurewicz 1994, 1997, Plašienka 1995a, Plašienka et al. 1998). In general, we distinguish units of the Pennine tectonic system structuralized during the Tertiary and those of the Slovakocarpathian (Austroalpine) tectonic system structuralized and emplaced predominantly during the Late Cretaceous, but partially reactivated during the Tertiary (Plašienka 1995b).

Four superposed superunits compose the Pennine system: 1) the Magura superunit (North Penninic) consisting of the Upper Cretaceous – Oligocene sedimentary formations composed mostly of flysch formations; 2) the Biele Karpaty superunit (North Penninic) as the innermost element of the ECW involving Cretaceous – Eocene strata (incl. synorogenic flysch of Jarmuta and Proč Fms); 3) the Oravic (Middle Penninic) superunit comprising Lower Jurassic – Upper Cretaceous sediments that is subdivided into several well-known PKB units of their own (e.g. Czorsztyn, Niedzica-Pruské, Kysuca-Branisko, Pieniny s.s., Orava etc); 4) the Vahic (South Penninic) superunit composed of Upper Jurassic – Upper Cretaceous oceanic sediments (Belice and Iňačovce-Krichevo unit).

However, the PKB and zones adjacent to its inner side involve also units derived from the CWC Cretaceous nappe systems, namely (from bottom to top): 5) the Tatric superunit consisting of the pre-Alpine crystalline basement and its Scythian–Turonian sedimentary cover; the structurally lowest and most external Infratatric units (Inovec, Kozol) include also Pennsylvanian–Permian rocks; 6) the Fatric superunit of detached Middle Triassic – Cenomanian sedimentary formations (Krížna nappe s.l.) that was emplaced in the Late Turonian; 7) some external partial Fatric and/or Tatric units that were strongly reactivated during the Early Tertiary and were largely incorporated into the PKB structure (Drietoma, Manín, Klape, Haligovce, Humenné subunits); 8) the Hronic superunit of detached Pennsylvanian – Hauterivian complexes (Choč nappe s.l.).

Except of the nappe superunits, four sets of overstep sedimentary complexes have been discerned. Still partially synorogenic Gosau Supergroup consists of: a) the Senonian Brezová Group; b) Paleocene – Lower Eocene Myjava-Hričov Group; c) Eocene – Egerian Podhale-Podtatra Group (incl. Eggenburgian sediments in places); c) Neogene sediments and volcanics.

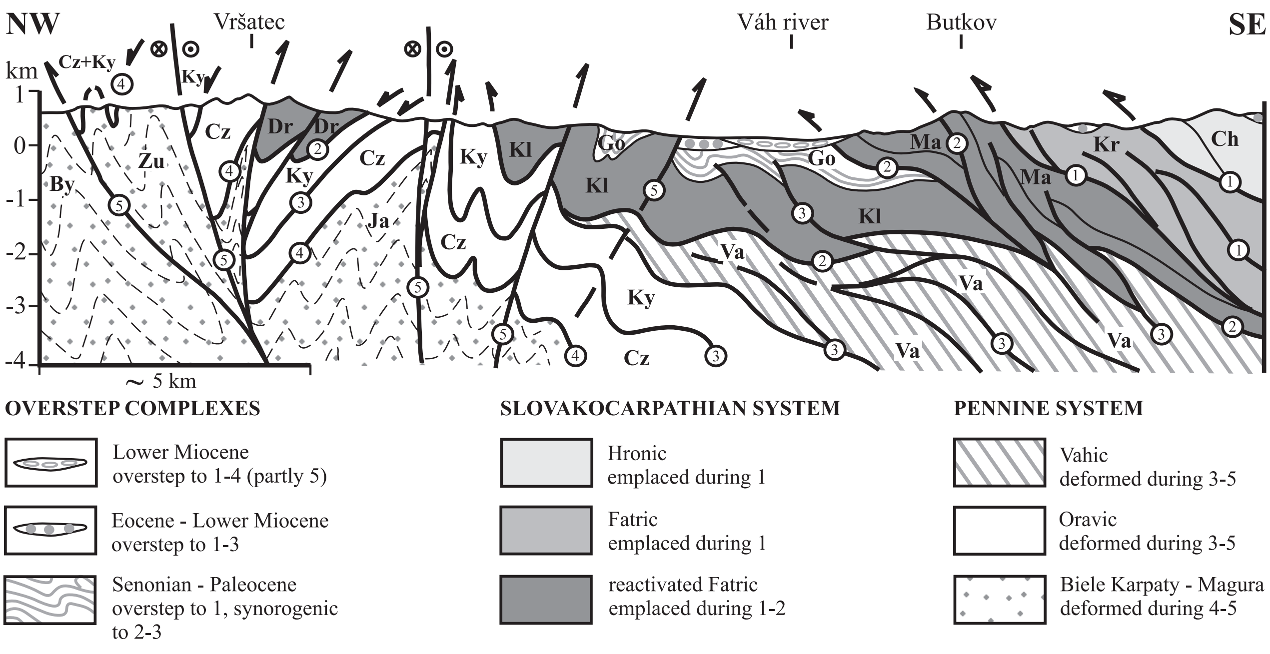
Tectonic evolution of listed units was very complex and proceeded in the following principal steps:

1. Late Turonian or slightly younger emplacement of the Fatric and Hronic cover nappe systems that overrode the Tatric substratum; transgressive Coniacian sediments of the piggy-back Gosau Brezová Group are syn- to posttectonic with respect to this event;
2. Senonian subduction of the South Penninic oceanic substratum and subcretion of the Vahic Belice unit beneath the frontal Infratatric units of the Tatric basement sheet, with piggy-back Fatric-Hronic nappes; Late Senonian (post-Campanian) partial reactivation of the most external Fatric units (Drietoma, Manín, Klape), somewhere with piggy-back Hronic, which overrode the Vahic and southern Oravic elements; Senonian sediments are inversion-related in the Magura, preorogenic (pelagic swell) in the Czorsztyn, synorogenic (lower plate) in the Kysuca and synorogenic piggy-back (upper plate) atop the Fatric-Hronic stack; thrust-related structures (ductile/brittle shear zones, low-grade metamorphism) in the Vahic and Infratatric units;
3. Latest Cretaceous – earliest Paleogene closure of the Vahic ocean, collision-related detachment of the Jurassic – Cretaceous successions of the Kysuca and allied “transitional” Oravic units (partly with piggy-back reactivated Fatric and synorogenic Gosau) to overthrust the ridge-related Oravic Czorsztyn succession; incipient fold-and-thrust belt of detached and piggy-back units and synorogenic Gosau sediments;
4. Paleocene to Early Eocene detachment of the Czorsztyn unit from its basement that was underthrust beneath the CWC and its thrusting (along with piggy-back higher Oravic, reactivated Fatric and synorogenic Gosau Myjava-Hričov Group) over the most internal Biele Karpaty elements with synoro-

genic Jarmuta and Proč Formations; duplexing and recumbent folding accompanied by shearing-related cleavage within the Czorsztyn unit;

5. After the Eocene – Oligocene extensional event associated with deposition of the Podhale-Podtatra Group overstepping the CWC and partly also the inner PKB zones, the PKB experienced strong dextral transpression during the Late Oligocene and Early Miocene, with development of a positive flower structure – the outer forward-thrust limb is located within the rear parts of the frontally accreted Biele Karpaty – Magura units, back-thrusts of the inner limb intensely affected the especially boundary PKB/CWC zones; the flower is usually centred by a narrow, generally vertical zone of the PKB s.s., in which strike-slipping prevailed that lead to the formation of the typical "klippen" (block-in-matrix) tectonic style caused by pervasive brittle faulting; Early Miocene dextral transpression is presumably related to counter-clockwise rotation of the CWC block with respect to the EWC, which was most probably accompanied also by narrowing and oroclinal bending of the PKB; sets of post-folding cleavages and shear zones record transpression on a mesoscopic scale;
6. Neogene evolution is complex as well and its kinematics strongly varies in individual PKB sectors – mid-Miocene to Pliocene sinistral transtension occurred along the inner side of the SW-NE trending PKB in western and northern Slovakia (Mur–Mürz–Leitha–Dobrá Voda–Považie–Žilina wrench corridor related to the tectonic escape of the ALCAPA block from the Alpine collision); mid-Miocene orthogonal shortening followed by the Late Miocene–Pliocene sinistral transtension (Orava Basin) affected the northernmost W-E striking PKB sector in Poland; the NW-SE trending eastern Slovakian sector underwent only dextral wrenching – first transpression during the Early–Middle Miocene, then transtension pursued by general extension during the Late Miocene and Pliocene (opening of the Transcarpathian depression).

Summing up, the overall tectonic scenario includes piggy-back mode of forward thrusting, formation of a fold-and-thrust belt capped by synorogenic sedimentary basins and some out-of-sequence thrusting as the principal tectonic regime during the Late Cretaceous and earliest Paleogene, followed by Oligocene–Lower Miocene dextral transpression responsible for the steepening and narrowing of the PKB that acquired its final tectonic style. The overall N-S convergence during the Middle–Late Miocene affected the already bended PKB, consequently its western SW-NE part underwent inversion to sinistral wrenching, whilst the eastern NW-SE trending part remained in a dextral regime. The klippen style resulted form the rheological contrast between a relatively thin, later disintegrated, stiff layer of Middle Jurassic to Lower Cretaceous limestones sandwiched within Lower Jurassic (décollement horizon) and much thicker Upper Cretaceous incompetent



■ **Fig. 1.** Schematic cross-section through the PKB and adjacent zones in western Slovakia (middle Váh Valley). Note that the Vršatec area represents the "PKB s.s." with tight imbricated flower structure, whereas the Váh Valley represents the "Periklippen Belt" as a mixture of Penninic and Slovakocarpathian units and several overstepping formations. Abbreviations: Go – Gosau Supergroup (Senonian – Paleocene); Ch – Choč nappe system (Middle Triassic – Hauterivian); Kr – Krížna nappe system (Middle Triassic – Cenomanian); Ma – Manín Unit (Lower Jurassic – lower Turonian); Kl – Klape Unit (Aptian – lower Turonian); Dr – Drietoma unit (Upper Triassic – Cenomanian); Va – Vahic units (Upper Jurassic – Senonian); Ky – Kysuca-Pieniny unit (Lower Jurassic – Campanian); Cz – Czorsztyn unit (Lower Jurassic – Maastrichtian); Ja – Javorina and Zu – Zubák nappe (Campanian – Lower Eocene); By – Bystrica unit (Senonian – Oligocene). Numbers refer to the age of principal fault and fold structures: 1 – late Turonian; 2 – late Senonian; 3 – Senonian/Paleocene; 4 – early Eocene; 5 – Oligocene to early Miocene.

marly and flysch sediments, as well as due to long-termed deformation progressing in several stages, including nappe-thrusting, extension and transpression.

The constructed cross-sections show that various PKB sectors record all these complex tectonic movements very irregularly, especially as far as the presence and distribution of tectonic units and overstep complexes are concerned. On the other hand, all sectors are dominated by the stage 5 dextral wrenching that finally shaped the PKB units. In many cases the stage 5 wrench faults juxtapose segments with fairly different collections of units, their thickness, pre-5 structure and erosional level. The fan-wise transpressional belt attains its maximum breadth of about 20 km in the middle Váh Valley near Považská Bystrica due to accumulation of at least 5 subvertical strike-slip duplexes developed within the Klape and underlying units. Relics of older thrusting stages are variably recorded, sometimes totally obliterated by wrenching. Nevertheless, if present, the superposition and lithostratigraphic content of units with different affiliation helps much in deciphering complicated tectonic history of this worldwide unique tectonic phenomenon.

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Late Jurassic to Early Miocene Tectonic Evolution and Palaeogeography of the Western Outer Carpathians

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The recent contribution aims to discuss tectonic processes, controlling the late Jurassic to early Miocene evolution of the Western Outer Carpathian (WOC) sedimentary basins and their source areas. In particular it attempts to compile in a coherent model constraints obtained from analysis of tectonic subsidence history of sedimentary basins (Poprawa et al. 2002), uplift history of source area based on analysis of deposition rate (Poprawa et al. 2006), as well as geochronology of basement of the sediment source area based on isotopic and chemical dating of crystalline pebbles (Ślączka 1998, Poprawa et al. 2004, 2005, Kusiak et al. 2004, Michalik et al. 2004, Malata et al. 2005). The model was confronted also with other data, like general facies evolution of the WOC basins (e.g. Książkiewicz 1963, Koszarski 1985), structural data (e.g. Świerczewska and Tokarski 1998, Nemčok et al. 2001) as well as geochemistry and petrology of igneous rocks (e.g. Narębski 1990, Dostal and Owen 1998).

At the end of Jurassic and beginning of Cretaceous in the WOC a rift-related extension led to development of the deep marine gabens with flysch and pelagic sedimentation, the elevated horsts, supplying the basins with sediments, and the zones of shallow marine carbonate sedimentation. Syn-rift extension tectonics is concluded mainly from characteristic pattern of tectonic subsidence (Poprawa et al. 2002) and sedimentological model for the Cieszyn beds in the Silesian subbasin (Słomka et al. 2002). Transition to the Early Cretaceous and Cenomanian post-rift thermal sag stage was responsible for general ceasing of subsidence of basins and tectonic activity in the source areas. Extension related to the late Early Cretaceous development of the basic volcanics (Narębski 1990, Dostal and Owen 1998, Nemčok et al. 2001) is regarded here as a minor reactivation of the previous rift structures. In the Barremian-Albian northern, external sources for sediments were uplifted due to compression (Poprawa et al. 2006), which was caused by the orogenic collision to the west and north of the Central Carpathians, related to subduction of the Penninic ocean (Vahicum), and/or by the orogenic collision in the Middle and Outer Dacides. This was coeval in time, and possibly genetically related to uplift and erosion of the SW and NE parts of the Polish Basin (particularly area of the Fore-Sudetic Monocline).

The Silesian ridge, rapidly elevated and eroded during the Late Cretaceous and Paleocene and supplying vast amount of detritus to the Silesian subbain, is interpreted here as a thick-skinned collision zone, composed of individual thrust sheets. Convergence, thrusting and stress transmission towards foreland caused flexural subsidence of the proximal zone of foreland (the inner Silesian basin) and uplift in the distal zone (including: the outer Silesian subbasin, the Subsilesian facies zone, the Skole subbasin and the northern sediment source areas). Concept of nappe structure of the Silesian ridge is derived from presence of contrasting types of detritus eroded from this isolated, intra-basinal source area. Sedimentary rocks, including cannibalised Lower Cretaceous flysch of the Silesian basin, are eroded from the Silesian ridge together with metamorphic rocks of low grade and high grade (e.g. granulites) metamorphism. The Late Paleocene–Early Eocene intensification of supply of the upper-most Jurassic to lower Cretaceous carbonates from the Silesian ridge is explained by thrusting of an another nappe onto the ridge. Coexistence of the Carboniferous to Permian metamorphic rocks together with unmetamorphosed sedimentary rock of the Devonian-Carboniferous age within the Silesian ridge also stands for presence of individual nappes with contrasting geological setting (Poprawa et al. 2004, 2005). Collisional uplift of the Silesian ridge allows to explain very high rates of deposition for the sediments derived from the source area (Poprawa et al. 2006). At the same time sediment source area north of the WOC also was compressionally uplifted. The late stage of Late Cretaceous-Paleocene compression, inversion and uplift of the WOC is coeval in time with compression, inversion and uplift of the Polish Trough. This confirms a concept of strong tectonic coupling between basement of the northern Tethian domain and European plate (peri-Tethian domain).

The Eocene alternating shallow marine deposition and exposition for erosion of the Silesian ridge is interpreted as controlled by both episodic tectonic activity and eustatic sea level changes. At this time a new collision zoned developed south of the Magura subbasin, which supplied vast amount of detritus to the Magura beds. The Eocene collision and development of the accretionary prism caused flexural bending of its broad foreland, subsidence, relative facies unification and decrease of activity of the source areas located north of the Magura subbasin. The Oligocene progress of plates/microplates convergence, and relocation of the zone of tectonic shortening towards the north, led to compressional uplift of the source area located both to the north of the WOC basins and source area located to the south of the Silesian facies zone, the later composed of crystalline basement and sediments of the partly deformed Magura unit. This sources supplied with detritus the upper Oligocene-lower Miocene Krosno Beds, being diachronic continuation of synorogenic deposition of the Magura Beds.

According to results of 'exotic' pebbles analysis the WOC developed on the basement, being the south-western prolongation of the Trans-European Suture Zone, composed predominantly of the Variscan and 'Cadomian' terrains (Poprawa et al. 2004). The contact zone of the Silesian basin and Silesian ridge is coinciding with palaeoboundary between basement of Variscan and 'Cadomian' orogenic consolidation to the south and north respectively. Therefore Silesian ridge developed on the suture of two tectono-stratigraphic terranes. The other suture of terrains, i.e. border between southern prolongation of the Brunovistulicum and Małopolska blocks/terrains, is expressed in detritus derived to Skole subbasin from the north.

During the Late Cretaceous-Paleogene-early Miocene an important tectonic shortening across the WOC took place, accommodated mainly in the source areas (Poprawa et al. 2004, 2006, Malata et al. 2005). This indicates that palaeogeographic relations between the Silesian subbasin, the Magura basin and the Central Carpathian Paleogene Basin were changing during the Cretaceous and Cainozoic. In the time span of the Albian to Oligocene in the zone palaeogeographically located between the Magura subbasin and the Central Carpathians three separate source area were active, each characterized by different geological setting. These sources were replacing each other in time, suggesting significant collisional and/or strike slip reorganisation of the zone during that period. The collision in the WOC evolved with time from thick-skinned mode during the late Cretaceous-Paleogene to thin-skinned one during the Miocene.

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Geochronology of the Crystalline Basement of the Western Outer Carpathians' Source Areas – Constraints from K/Ar Dating of Mica and Th-U-Pb Chemical Dating of Monazite from the Crystalline 'Exotic' Pebbles

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The Western Outer Carpathians (WOC) are composed of several major tectonic units, partly characterised by individual facies development. The units are composed of sedimentary fill of individual basins/subbasins. The basin-fill was detached from the original basement, deformed and thrusted over the European plate. The WOC sedimentary basins were supplied with detritus from several source areas, both external with relation to the WOC, like its northern rim, and internal, like e.g. Silesian ridge and Southern Magura ridge (e.g. Książkiewicz 1965). Judging from composition of pebbles deposited in the WOC flysch basins, often referred to as 'exotics', the source areas were build of broad variety of sedimentary and crystalline rocks (e.g. Książkiewicz 1965, Wieser 1985). Contribution of the later one to understanding of the WOC evolution is still rather limited, mainly due to lack of geochronological constraints.

Only during the last years geochronology of crystalline 'exotic' pebbles become a subject for systematic studies. First single K/Ar datings of mica from crystalline rocks derived to the WOC from the northern source area were reported by Lis (1980) and Ślączka (1999). More extended analysis, presented by Poprawa et al. (2004) and Malata et al. (2005), allowed to differentiate characteristics of the northern rim of the WOC, the Silesian ridge, the Southern Magura ridge and Marmarosh ridge respectively. CHIME analysis of monazite were conducted so far by Hanžl et al. (2000) for granite pebbles eroded from the western Silesian ridge, Michalik et al. (2004) for gneisses supplied form the central part of the Silesian ridge, Kusiak et al. (2004) for detrital monazites from the Dukla unit, as well as by Poprawa et al. (2005) for mostly metamorphic rocks derived from the northern rim of the WOC, the Silesian ridge, and the Southern Magura ridge.

The recent contribution is based on results of K/Ar dating of mica for 32 samples of metamorphic 'exotic' pebbles and for another 8 samples of igneous 'exotic' pebbles, representing every of the main source areas for the WOC basins. The results of K/Ar dating of different metamorphic rocks typically correspond to time of post-metamorphic cooling. In a case of biotite obtained age could reflect last tectono-thermal event. For the igneous rock the obtained ages most probably represents time of magmatic/volcanic activity. In a case of 3 sandstone samples detrital muscovite could be both of metamorphic and plutonic origin (Wieser 1985), with the first option more probable. Moreover for 6 'exotic' pebbles of metamorphic rock and one of granite, a single spot ages of monazite were analysed with chemical Th-U-total Pb isochron method (comp. Suzuki and Adachi 1991). In most cases for each individual monazite grain the CHIME age was determined for a few single spots. The main objective of the research is to obtain new constrains on geological setting of the source areas as well as on tectonic evolution and palaeogeography of the WOC.

Pebbles/blocks deposited to the WOC basins from the external, northern source record mainly the Neoproterozoic to Cambrian cooling after metamorphism (Poprawa et al. 2004). This, together with characteristic assemblage of crystalline and sedimentary 'exotics', suggest that the source of sediments was related to the Brunovistulicum and/or Małopolska massifs/terrains. The only exception is a sample of the upper Variscan granulite, eroded from the Variscan orogen, transported to the east and deposited in the Variscan foredeep, and afterwards recycled during the late Early Cretaceous from the Brunovistulicum to the Silesian subbasin. CHIME dating of monazite from metamorphic pebbles, representing the northern source, give the late Neoproterozoic to Caledonian single spot ages. For the northern source presence of the late Neoproterozoic-Cambrian granitoids was also revealed, both by K/Ar dating of mica and chemical Th-U-total Pb dating of monazite (Poprawa et al. 2005). K/Ar dating shown also, that the northern source supplied the WOC basins with detritus of the late-most Carboniferous to Permian andesites and porphyrites, typical for the NE part of the Brunovistulicum. Particularly interesting was a sample of granite pebble from the Skole unit, which revealed the late-most Cambrian to early-most Permian K/Ar age of mica. Directly to the north of Carpathians such granites are known only from the contact zone of the Brunovistulicum and Małopolska massifs/terrains.

The upper Cretaceous to Paleocene detritus of crystalline and sedimentary rocks from the Obidowa-Słupnice unit have characteristics resembling the northern source (Poprawa et al. 2004, 2005). K/Ar ages of biotite and CHIME single spots ages of monazite for gneiss pebble reveals the late Neoproterozoic to Caledonian metamorphism. It coincides with detritus of Carboniferous coal specific for Brunovistulicum. This together indicates that the Obidowa-Słupnice unit could be associated with Skole one and both represent common sedimentary basin, as proposed by Żytko and Malata (2001).

The Silesian ridge and the southern Magura ridge supplied the WOC basins with pebbles/blocks recording predominantly the Late Carboniferous to Permian metamorphism (Poprawa et al. 2004, 2005). This is documented by both K/Ar dating of mica and CHIME dating of monazite. Comparison of this data with the CHIME ages presented by Hanžl et al. (2000), shows that the late Variscan metamorphism was roughly coeval with granite emplacement in the western part of the Silesian ridge. Metamorphism and magmatic activity were related to Variscan orogeny (e.g. Nejbert et al. 2005).

The late Silurian to middle Devonian CHIME single spot ages obtained for the gneiss sample from Blizne (eastern part of Silesian subbasin, supplied by Silesian ridge) could be tentatively interpreted as related to the gneiss protolith. The Mesozoic CHIME ages might reflect tectonically induced thermal and diagenetic events, in particular the late Jurassic rifting in the WOC system (Poprawa et al. 2002, 2005). Other gneiss 'exotics' from Gródek at Rożnów Lake (central par of the WOC), derived also from the Silesian ridge, were studied also by Michalik et al. (2004). In that case in a spectrum of obtained ages the Variscan and older ones were rather rare and incoherent, while majority of spot gave ages falling into a time span of the Permian to Jurassic, concentrating within the Triassic. For a gneiss pebble collected from the same conglomerate level directly to the east of Gródek (location: Siekierczyna) Poprawa et al. (2004) obtained the early Permian K/Ar mica cooling age, which corresponds to the upper limit of the main cluster of CHIME ages from Michalik et al. (2004). However development of majority of monazites from Gródek gneisses postdates the main cooling phase after metamorphism, and could be related to tectono-thermal events, e.g. rifting (op. cit.). CHIME analysis for one sample of gneiss, eroded from the Southern Magura ridge shows that the main cluster of single spot ages on monazite document the Variscan metamorphism (Poprawa et al. 2005). This is coherent with K/Ar cooling age on muscovite for the same sample. Characteristic for this sample are clusters of CHIME ages being an equivalent of the Permian to Middle Triassic, and the Jurassic. Taking into account a palaeogeographic position of the source area south of the Magura basin, it is possible to speculate that the Triassic ages might reflect tectono-thermal event related to the Triassic rifting in the Pieniny basin, which was reconstructed by Birkenmajer (1988). For the other pebble from the location a biotite was dated with K/Ar method, giving the late-most Jurassic to early-most Cretaceous age, which might be in some relation with the Jurassic cluster of CHIME ages in the discussed sample. The late Jurassic K/Ar and CHIME age are tentatively related here with the rifting phase, causing development of the WOC flysch basins (Poprawa et al., 2002, 2005).

For a pebble of gneiss, which was supplied to the Dukla basin during the late Senonian-Paleocene from the north by unconstrained source area, the results of K/Ar dating document the late Variscan cooling after metamorphism. According to analysis of

Bąk and Wolska (2004) the protolith of the gneiss was a granite. A rather limited number of monazite CHIME single spot ages indicate, that the granite originated during the late-most Neoproterozoic to middle Cambrian time (Poprawa et al. 2005). It is however difficult to exclude an alternative interpretation, in which the cluster of older CHIME ages in this sample would be related to a protolith for the granite.

During the late Oligocene poorly rounded blocks of anchimetamorphic rocks (chlorite-rich fyllite and the chlorite-muscovite schist) were deposited in the eastern part of the Krosno Beds basin. Two analysed samples reveal the Albian to Cenomanian? cooling after metamorphism (Poprawa et al. 2004 Malata et al. 2005). This blocks were derived from a very proximate source, referred to as Marmarosh "cordillera". This clearly contrasts with facies of the Albian-Cenomanian sediments in the section of this part of the basin, which are represented by fine-grained, deep marine flysch and shale, characterized by a limited thickness. Such sediments can hardy develop in direct proximity of metamorphic domain. This stands for a significant Late Cretaceous-Paleogene convergence between Silesian basin and Marmarosh Massif, driven by collisional tectonics.

Comparison of characteristics of detritus derived from the individual source areas allows to conclude, that the WOC developed on the basement, being the southern prolongation of the Trans-European Suture Zone, composed of the Variscan and 'Cadomian' terrains. The contact zone of the Silesian basin and Silesian ridge coincided with palaeoboundary between basement of Variscan and 'Cadomian' orogenic consolidation to the south and north respectively. Development of Silesian ridge on the suture of Variscan and 'Cadomian' terrains could explain also a specific composition of detritus documented in one outcrop near Skrzydlna in central part of the WOC. In this location olistostrome in Menilite beds contains blocks of the late Variscan gneiss together with other type of Cadomian gneiss and unmetamorphosed Palaeozoic sedimentary cover. The suture of terrains, and therefore also location of Silesian ridge, represents a significant rheological contrast between two types of crust, strongly influencing development of the WOC. As it comes from above, conventional subdivision of the Western Carpathians into the Inner and Outer ones, documented by differences in orogenic phases and the Mesozoic facies development, does not corresponds to the basement domains.

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A Ramp-and-Flat Geometry of Thrust Faults in the Pavlov Hills, Western Carpathians, Czech Republic

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The Pavlov Hills are situated in the westernmost part of the West Carpathians at their contact with the Eastern Alps. The Jurassic to Cretaceous pre-flysh sediments form slices incorporated in the Paleogene flysh nappe (Ždánice unit). The Upper Jurassic part consists of dark gray deep marine claystones to limestones (Klentnice Formation), which prograde into light shallow-marine limestones (Ernstbrunn Limestone). The Cretaceous mostly siliciclastic sediments overlie these limestones. The structure of Pavlov Hills was produced by thrusting in the Carpathian acretionary wedge during the younger phase of the Alpine orogenesis (lower Miocene).

During our tectonic research in the Pavlov Hills the orientation of bedding in sedimentary rocks was studied. Two main maxima of bedding orientation were recognized. The bedding planes were buckled into several upright anticlines with fold axis very gently plunging to NE. Very well documented large brachyanticline is situated in the eastern part of Mikulov (Svatý kopeček). This structure was recognized during the new reinterpretation of seismic sections under the surface.

New geological mapping (with more detailed stratigraphic division) and compass data together with the data in old boreholes (Nové Mlýny-3) show several thrusts with stratigraphic inversions and tectonic duplications of the Jurassic formations. Thrusts are marked by high-strain zones with large amount of small tectonic slices of different age (Jurassic, Cretaceous, Paleogene).

The anticline structure is accompanied by duplexes. Thrusts are usually subparalell to bedding and mostly striking in NE-SW direction and dipping to the SE. Detachments are distinguished in the Klentnice Formation, in the “nodular limestones” (middle Tithonian) and at the top of the Ernstbrunn Limestone. Some more steep parts situated in the Ernstbrunn Limestone are interpreted as ramps. The angle Φ between the flats and the ramps is 20°. This value was obtained by the weighted average of friction angles of the failure-tested Jurassic rock using Mohr-Coulomb failure model?

The anticlines were formed in a ramp-and-flat geometry regime, so the balanced cross-section could be constructed based on seismic data across the Pavlov Hills.

Late Neogene Tectonic Activity of the Central Part of the Carpathian Foredeep, South Poland

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Normal faults of different orientations appear to be the youngest manifestations of faulting in the Polish Outer Carpathians, composed of the Lower Cretaceous trough Lower Miocene strata, and the related Carpathian Foredeep which is filled with the Lower to Middle Miocene sediments. In the Outer Carpathians, the folds and thrusts produced by accretion-related shortening were formed between the Palaeocene and early Late Miocene. The origin of normal faults is still debatable, since it is not known whether these faults were a result of multidirectional extension produced in a single collapse event, or differently oriented extension proceeding in a series of successive events.

Structural studies of the Late Miocene-Pliocene(?) fresh-water molasses of the Witów Series and the overlying Pleistocene loessial complex provide a possibility to reconstruct the Late Neogene – Pleistocene (to Recent?) stress field in the central part of the Polish Carpathian Foredeep and, indirectly, in the central part of the Polish Outer Carpathians. The strata of such an age are unique features in the Polish Carpathian Foredeep, providing thereby a key record of structural deformation during the latest stages of orogenic evolution of the Carpathian orogen. The molasses are cut by joints, and normal and strike-slip faults which were formed in two successive events: (1) a syn-depositional one for the molasses (Late Miocene-Pliocene?), proceeding under NNW-SSE to N-S oriented horizontal compression, possibly coeval with reactivation of a NE-striking sinistral fault of the Kurdwanów-Zawichost Fault Zone in the basement; (2) a post-depositional one for the molasses (Pliocene to Middle Pleistocene) during N-S to NE-SW- oriented extension, and (3) both syn- and post-depositional ones for the loessial complex (Late Pleistocene).

In the first event, reactivation of the NE-striking sinistral fault led to formation of N-S- oriented joints, as well as NW-striking dextral, and NNW-trending normal faults. This event was probably contemporaneous with sinistral reactivation of some thrusts in the Western Outer Carpathians, induced by eastward-directed extrusion of crustal blocks in the Carpathian internides. In the second event, both W-E and NW-SE- oriented joints and WNW-striking normal faults were formed. The latter most probably originated due to reactivation of the Early Palaeocene WNW- and NW- striking normal faults in the basement. In the third event, both NE-SW and NW-SE- oriented joints and NE-striking normal faults were formed as a result of reactivation of the SW- and WSW-striking faults in the basement. Therefore, normal faults detected in the Outer Carpathians and Carpathian Foredeep appear to be a result of not a single collapse event but of different successive events. This extensional episode lasted at least to the late Pleistocene.

We also provide evidence for the recent, N- to NNE-directed, tectonic compressive stress, typical for that segment of the Carpathian arc. This stress resulted in the formation of an orthogonal system of joints striking N-S and W-E, produced during the (4) event.

Pattern of the Mesoscopic Thrust Faults in the Eastern Part of the Silesian Nappe (Polish Western Outer Carpathians)

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Polish Western Outer Carpathians are a stack of nappes composed of Lower Cretaceous to Lower Miocene flysch sediments. One of the biggest nappe is the Silesian nappe. The outcrop of this nappe forms the bend convex to the North. The eastern part of this nappe, located to the southeast of the Wisłoka River, is the object of the research. In the Polish Outer Carpathians the folding and thrusting were connected with the two shortening events, which were characterized by different orientation of horizontal compression: 1) NNW- (N), and 2) NE- (NNE) directed ones. During the first event the thrusting has in-sequence character and the deformational front migrated approximately from the south to the north. The regional folds were formed during this event. During the next event, the previous thrust faults and related folds were overprinted and refolded. The regional fold axes in the Silesian nappe are approximately parallel to the frontal thrust of this nappe. In the eastern part of the Silesian nappe the regional fold axes and the strikes of the regional thrust faults are directed WNW-ESE and NW-SE.

The tectonic structures were investigated in 34 outcrops. The studied mesoscopic thrust faults were divided into two groups formed in: 1) horizontal strata and 2) tilted strata. Numerous thrust faults of (1) group were tilted, together with the host strata during folding. Such faults were backtilted at the beginning of the structural analysis. There were measured 572 thrust faults, 255 faults of (1) group and 317 faults of (2) group.

The thrust fault planes dip mostly about 25°. The dip angles of the (1) group faults are clustering around the value 25°. The dip of the (2) group fault planes is characterized by bigger changes and ranging between 15 and 70°. The strike of thrust faults usually varies between N 60°E and N140°E, showing single prevailing orientations N 125°E in the both groups of thrust faults. Other fault planes of the (2) group are striking WNW-ESE, W-E and more rarely ENE-WSW. The orientation of the strikes of the (1) group fault planes are mostly clustering around the direction N 125°E. The direction of thrusting varies mostly from N-S to NE-SW. The dominant direction of thrusting along the (1) group faults is NNE-SSW and along the (2) group faults is NE-SW. The differences of the thrust fault strikes and the thrusting direction between the (1) and (2) group of thrust faults are visible but not significant. According to crosscutting relationship older thrust faults are commonly these faults striking WNW-ESE and younger are these faults striking ESE-WNW. The reconstructed orientation of the horizontal compression is N 18°E for the older thrust faults of the (1) group and N208°E for the younger thrust faults of the (2) group.

The dominant age relationship of the thrust faults could be caused by clockwise rotation of the horizontal compression. Numerous thrust faults of (1) group vary in strike orientation with the host strata, this fact may suggest that the rotation of the horizontal compression took place locally before folding. There is also second explanation of such thrust faulting: the NE-oriented, stable horizontal compression and counterclockwise rotation of host strata.

In the eastern part of the Silesian nappe the differences between the orientations of the older thrust faults of the (1) group and the younger ones of the (2) group are smaller comparing such differences, which are observed in the central part of this nappe. Therefore in the eastern part of the Silesian nappe the total amount of the rotation was smaller. When we are taking into consideration that the host strata rotated counterclockwise, then the bending of the whole Silesian nappe could cause the clockwise rotation of the eastern part of the Silesian nappe. Such clockwise rotation of the eastern part of this nappe could decrease the effect of couterclockwise rotation of the whole nappe.

The Black Box of the Stress Analysis Based on Calcite Twinning

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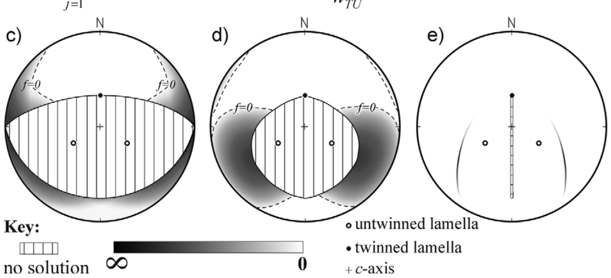
Since the deformation origin of calcite twin lamellae (e-twins) and their crystallographic laws have been determined in the end of the 19th century, it was recognized as the main deformation mechanism of calcite polycrystalline aggregates at low temperatures, low confining pressures and low finite strains (8 %; e.g. Turner 1963, DeBresser, Spiers 1996).

The e-plane will twin if, and only if the shear stress τ_i exceeds the critical value $\tau_c \approx 10$ MPa, which is believed to be independent of normal stress, temperature and strain rate (e.g. Laurent et al. 1981). This is one of basic assumptions of the Etchecopar inverse method modified by French authors Laurent, Lacombe et al. (e.g. Lacombe, Laurent 1996). It is based on applying numerous (500 to 1000) randomly generated reduced stress tensors and then selecting the best-fit tensor using a penalisation function. When a stress tensor [*T*] is applied on a set of twin planes, there are four possibilities for any twin plane: 1) the plane is twinned and [*T*] should twin it (compatible twinned plane); 2) the plane is untwined and [*T*] should not twin it (compatible untwined plane); 3) the plane is twinned and [*T*] should not twin it (incompatible twinned plane); 4) the plane is untwinned and [*T*] should twin it. In an ideal case only compatible untwined and twinned planes occur. In most cases all four cases are present. Whereas incompatible twinned planes can be caused by polyphase deformation or stress perturbations during deformation, the incompatible untwined planes are caused by wrong orientation or shape ratio of [*T*]. This means, that incompatible untwined lamellae can be most effectively used as a criterion for estimating the best-fit tensor.

a) $f = \sum_{j=1}^{n} \left(\tau_j - \tau_a \right)$ b) $f = \frac{n_{CT}}{w_{TU}} + n_{CU} w_{TU} - n_{IU} w_{TU}$



■ **Fig. 1.** a) penalization function *f* (Laurent et al. 1981), τ_j – shear stress for incompatible untwined planes, τ_a – the lowest shear stress of all compatible twinned planes (i.e. τ_c); b) penalization function *f* preferred by authors, n_{CT} – number of compatible twinned planes, n_{CU} – number of compatible untwined planes, n_{IU} – number of incompatible untwined planes, w_{TU} – twinned/untwined planes ratio; the lowest values of *f* for all possible stress tensor orientations represented by σ_1 orientation at three different stress tensor shape ratios: c) $\Phi = 0$, d) $\Phi = 0{,}5$, e) $\Phi = 1$, for one grain with one twinned lamella.

Laurent et al. (1981) proposed a criterion (so called penalization function *f*) for selecting the best-fit tensor (fig. 1a). It is a sum of differences of shear stresses for incompatible untwined lamellae and the least shear stress for compatible twinned lamellae, considered as the critical value for twinning τ_c. It is clear, that the value of this function is strongly dependent on the amount of incompatible untwined lamellae and compatible twinned lamellae. As shown in Fig. 1, there exist stress tensors with $f = 0$, which represent stress tensors with no solution, the spatial distribution of which depends on the shape ratio Φ. The number of such "*solution–less tensors*" decreases with decreasing number of stress tensors examined and with increasing number of grains. Using this penalization provides results that may not always be reliable.

A penalization function with solutions for all stress tensors and a systematic search of all possible stress tensors within engaged range should be preferred. Such procedure commonly provides more maxima (minima) of the penalization function but eliminates all wrong solutions. At this time it seems that the most useful penalization function is a weighted sum of compatible twinned and untwined planes minus incompatible untwined planes (Fig. 1b). It provides less scattered maxima clusters and no solution–less stress tensors.

Even thought calcite stress inversion method by Laurent et al. (e.g. 1981) provides credible solutions and has been proven by experiments, a detailed revision of the penalization function used revealed some uncertainties due to its discreteness. The presented penalization function is less discrete then *f* by Laurent and Lacmbe (e.g. 1981), but in some cases it requires additional parameters to provide a unique solution (e.g. the lowest sum of stresses along incompatible untwined lamellae.

Acknowledgements

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Neotectonic and Landscape Evolution of the Gödöllő Hills, Central Pannonian Basin, Hungary

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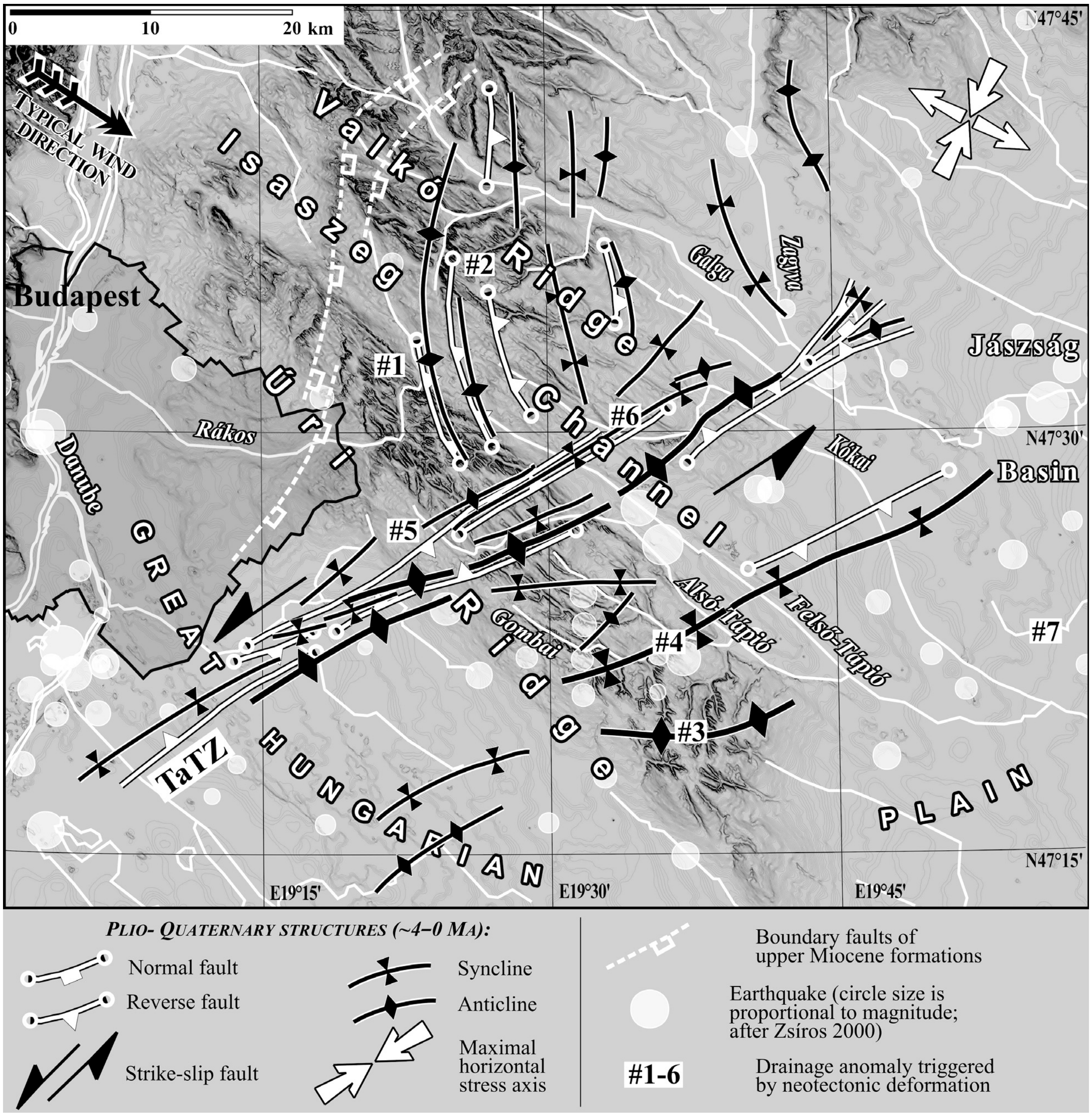
Neotectonic inversion in the SW part of the Pannonian Basin started during the latest Miocene (~6 Ma) and gradually extended into the basin interior during Plio-Quaternary times (e.g. Fodor et al. 1998, 1999, Bada et al. 2001). During this phase overall extension and subsidence (syn- and post-rift phases) of the basin system stopped and compressional stresses triggered differential vertical motions – i.e. simultaneous uplift and subsidence – in the basin interior (e.g. Fodor et al. 1999, 2005, Bada et al. 2001). The Gödöllő Hills is a rolling hilly area of 105 to 344 m asl. height in the central Pannonian Basin, east of Budapest, capital of Hungary (Fig. 1). It is part of the transitional zone between uplifting and subsiding regions and is composed of late Miocene–Pliocene delta and fluvial sequences covered by up to 40 m Quaternary loess and/or eolian sand. Morphologically it consists of two relatively elevated and dissected ridges (Valkó and Úri Ridge) separated by a wide valley with smooth surface (Isaszeg Channel; Fig. 1). River deflections and drainage pattern anomalies suggest that neotectonic activity had considerable influence on landscape evolution. Joint geologic-geomorphologic study, seismic reflection profile analysis and structural mapping were carried out to constrain neotectonic warping of this region. Main goals of this study are (1) to define how neotectonic motions influenced Quaternary landscape evolution; and (2) to recognize the relative role of neotectonic deformation versus climate controlled surface sculpturing.

The uppermost imaged reflectors on seismic profiles belong to the late Miocene (post-rift, Pannonian) phase, occasionally up to the Pliocene fluvial sequences. The upper Miocene delta succession is a well reflecting formation below the Gödöllő Hills. Its characteristic layered-cake structure usually appears as continuous, gently dipping reflectors. The inversion in the study area is mainly expressed by the gentle folding of the uppermost imaged horizons with a maximum amplitude of a few hundred m, which verifies Pliocene–Quaternary deformation (~4–0 Ma). The structural map presented in Fig. 1. is depicting the deformation affecting the uppermost imaged reflectors, description of older structures is beyond the scope of this study.

In the northern part of the Gödöllő Hills reverse reactivation of N-S to NNW-SSE trending earlier – mainly syn-rift – normal faults, and connected folding of the uppermost imaged horizons is typical. The central area is crossed by a WSW-ENE trending fault zone, which was described by Tari et al. (1992) as a syn-rift transfer fault. Sudden changes in thickness and syn-tectonic wedges of lower upper Miocene layers refer to rejuvenation of transtensional flower structures during the beginning of the post-rift phase, recognised already by Csontos and Nagymarosy (1998). It was named by Fodor et al. (1999) Tóalmás Zone and dated as middle to late Miocene (~14–6 Ma). In this morphotectonic study this strike-slip zone is called Tápió–Tóalmás Zone (TaTZ, Fig. 1.) because the characteristic surface expression of its neotectonic transpressive reactivation in the Alsó-Tápó valley. Inversed negative flower structure and en echelon arrangement of the faults and related folds suggest sinistral motion and transpressional character of the TaTZ. This is indicative of a NE-SW compressional-transpressional stress field for the neotectonic phase (Fodor et al. 2005). Surface ruptures have not been documented in the northern area nor in the TaTZ. Reverse slip along reactivated syn-rift faults has been distributed into several small-scale fault branches within the upper Miocene layers and/or was accommodated by drag folds above the fault tips. South of the TaTZ the upper Miocene reflectors are folded into wide ~WSW-ENE trending anticlines and synclines with 100–300 m amplitude, and 10–20 km wavelength (below Gomba Depression and Pánd Antiform on Fig. 1). Earthquakes (Zsíros 2000, Fig. 1) in the southern area, in the Jászság Basin and in the W elongation of the TaTZ (S of Budapest) suggest that deformation may last until the Holocene.

Joint investigation of mapped structures and geomorphology revealed that peculiar drainage pattern indicates considerable influence of neotectonic deformation on landscape evolution however, some typical landforms lack structural control.

(1) The gentle SE slope of the smoothed envelope surface of the Gödöllő Hills resembles the common SE dip of the upper Miocene reflectors. The transitional position of the Gödöllő Hills between the uplifting Hungarian Mountain Range and the subsiding Great Hungarian Plain is allowed for this tilting. No structural lines are reconcilable with these valleys consequently, the characteristic NW-SE valley strike in the Gödöllő Hills is surface expression of the tilt of the entire area. Similarly, the sharp rectilinear margins of the Valkó and Úri Ridges and the Isaszeg Channel lack structural control.
(2) Some anomalies of the drainage pattern correspond to locations of structural deformation. Anticlines of the uppermost reflectors can frequently be connected to topographic highs and similarly synclines to topographic depressions. The uplift of anticlinal hinges led to enhanced surface erosion, river piracy and development of radial drainage networks. In the central part of the Úri Ridge dissected landscape suggests river incision triggered by young uplift of the underlying anticline. The same structure could have induced the



■ **Fig. 1.** Neotectonic deformation of the Gödöllő Hills superposed on the slope distribution map (steeper slopes are darker). See explanation in text. (modified after Fodor et al. 2005).

capture of the Rákos Creek (#1 on Fig. 1), which has been deflected from its consequent SE flow direction within the Isaszeg Channel and its lower reach is cutting through the Úri Ridge towards the Danube River. Further towards the N radial drainage pattern has developed most probably because of a growing anticline observed on seismic profiles (#2 on Fig. 1). In the SE part of the Úri Ridge the Pánd Antiform (#3 on Fig. 1) is also characterised by radial drainage pattern, which also has developed above an anticline of the upper Miocene layers. In subsiding depressions, like e.g. the Gomba Depression (#4 on Fig. 1) N of the Pánd Antiform, centripetal drainage pattern and alluvial sedimentation occurred.

(3) The upper and lower reaches of the Alsó-Tápió and Kókai Creeks follow the overall SE tilt of the area however, their intermediate section flows in a WSW-ENE direction. These

river reaches have been deflected by anticlines formed above en echelon segments of the TaTZ (#5 and #6 on Fig. 1). Eastwards the morphologic expression of the TaTZ decreases because of the smaller amplitude of the structures and the proximity of the subsiding Jászság Basin.

Typically minor amplitude of surface undulations respective to the amplitude of folding of the uppermost Miocene layers indicates that several episodes of Plio-Quaternary erosion smoothed the deforming topography. Characteristic NW winds and local variation of wind power led to the development of the typical NW-SE striking landforms. Accordingly the Valkó and Úri Ridges of the Gödöllő Hills form large scale yardangs separated by a wind channel (Isaszeg Channel). In the Isaszeg Channel the surface expression of the structures is further weakened by the strong areal denudation of the wind. The central and SE part of the ridges evolved in a wind-shielded position where climate oscillations led to various phases of loess deposition and fluvial erosion. Here surface dissection is significant yet overall lowering was smaller (see slope distribution on Fig. 1).

According to the seismic reflection profiles structural inversion is younger than ~4 Ma. Chronostratigraphy of the outcropping loess-paleosoil sequences suggest that surface expression of the neotectonic deformation – i.e. valley sections developed in consequence of river deflections in front of growing anticlines – is at least 400–600 ky old. The significant lag between the onset of the structural deformation and the appearance of its surface expression may be explained by two reasons. Firstly, the denudation processes were stronger during Pliocene – early Pleistocene times, thus they obliterated the surface deformations. Secondly, the deformation in the first period of the neotectonic phase was slower and the vertical motions have been accelerating towards present.

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Lower Crustal Channel Flow in Hot Orogens in Space and Time Exemplified by the Variscan Eastern Margin

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Recent considerations of detailed petrological, geochronological, geophysical and structural data allow us to make progress in understanding mechanisms of crustal-scale exhumation of orogenic lower crust associated with lithospheric indentation. Current numerical models (e.g. Beaumont group) suggest an emplacement of “hot-nappes” in subsurface channel-flow powered either by gravity potential or by an indentation of a weak hot root with a lower crustal rigid promontory attached to the subducting plate. Geological examples of channel-flow are based on localized occurrence of high-grade rocks along the S. Himalayan front resulting from ductile extrusion driven by gravitational collapse and focused erosion.

We present an example of several thousand square kilometres of flat-lying orogenic lower crust underlain by a basement promontory located at the retroside of the Variscan orogen along a 300 km long collisional front (Poland, Czechia and Austria).

Gravity surveys show that the limit of basement promontory extend about 100 kilometres towards the internal part of the orogenic root from todays exposure of the orogenic front. Combined structural and petrological studies revealed that the orogenic lower crust (high-pressure granulites and mafic eclogites) was vertically extruded from depths of about 70 kilometres parallel to the western steep margin (ramp) of the basement promontory. The observed transition from steep to flat fabrics occurs in different depths from 35 to 15 kilometres and is marked by different P-T-t paths of exhumed lower crustal blocks. The vertically extruded rocks are reworked by flat fabrics reflecting the flow of hot material into some horizontal channel developed between the upper boundary (flat) of the basement promontory and the overlying orogenic lid. The flow kinematics in this horizontal channel are controlled by plate movements as documented by structural and paleomagnetic investigations. A simple 2D thermokinematic model is used to show that the differences in P-T-t paths are controlled by three major parameters: thickness of the indenter, plate velocity and thermal structure of the orogenic root. We suggest that the exhumation of orogenic lower crust in large hot orogens is an extremely heterogeneous process controlled by local parameters, essentially driven by indentation. Orogenic flat fabrics commonly reported in hot orogens result neither from lower crustal flow nor gravity driven collapse of an orogenic system but rather reflect the deformation fronts and geometries of crustal indentors.

Extraction of Morphotectonic Features from High-Resolution Photogrammetric DEM (Mecsek Mts., Hungary)

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Photogrammetric digital elevation models (DEMs) belong to the 2nd generation of DEMs. Compared to the preceding generation usually achieved by interpolation between contour lines, these are produced from aerial orthophotos. Under optimal conditions, they provide higher resolution and are devoid of some interpolation errors typical of the first generation.

A methodology developed earlier for tectonic feature extraction from traditional DEMs (Jordan et al., 2003, 2005; Sebe 2005) is applied on the photogrammetric DEM of Mecsek Mts. (SW Hungary). Mecsek Mts. and their foreland are characterized by the dominance of strike-slip fault systems and the presence of neotectonic (latest Miocene – Pliocene – Quaternary(?)) activity including young vertical movements. The area has already been studied from the aspect of tectonic geomorphology using DEM, traditional geomorphology and geology, and new concepts about young evolution history have been outlined (Sebe et al. 2006). The objective of the present study is to further improve our understanding on the tectonics of the area and to compare the two DEM types (contour-based and photogrammetric) in terms of morphotectonic interpretation.

Anthropogenic features such as roads and bridges were first removed from the photogrammetric DEM by means of mathematical morphology image processing methods. Detailed digital terrain analysis applied smoothing filters to the DEMs using a sequence of kernel sizes in order to detect morphotectonic features on various scales ranging from local to regional. For each re-scaled DEM several morphometric parameters of tectonic significance, such as aspect, slope, curvatures, directional derivatives and local relief were calculated and displayed as maps. These maps were analysed visually and statisctically to locate geomorphic features of tectonic origin. Tectonic study was enhanced by the examination of drainage network extracted from the DEM (for methods see Jordan et al. 2003).

Results show that the new photogrammetric DEM with a resolution higher than that of the traditional contour-based type provides important additional information, in particular along major morphotectonic features such as fault scarps and linear valleys, although it also has its characteristic error types that can hinder tectonic interpretation to a certain extent. In the photogrammetric DEM more fault lines and other tectonic features could be located, many of which are not indicated in geological maps. The new DEM seems to be especially useful in areas of low relief.

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Gravity Modelling of the Krkonoše-Jizera Pluton

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A detail gravimetric survey (in the scale of 1:25 000, i.e. with the density of 4–5 stations per 1 sq. km) was realized in the northern Bohemia during 2001–2003. The total extent of the surveyed area was 800 sq. km. It included the eastern marginal part of the Lužice (Lusatian) Granodiorite Massif, the Jizera Metamorphic (mostly orthogneiss) Complex, the Ještěd-Kozákov Belt of the South Krkonoše Metamorphic Complex and the substantial Jizera part of the Krkonoše – Jizera Granite Pluton.

The data processing and interpretation stages which followed the field works during 2004–2006 embraced an enlarged rectangular area sized 90 km (W-E) and 65 km (N-S), i.e. the area of 5850 sq. km situated on the German–Polish–Czech borderland. The corners demarcating this enlarged area are situated near the towns of Reichenbach (NW)–Zlotoryja (NE)–Trutnov (SE)–Mnichovo Hradiště (SW). The evaluated area includes the whole Krkonoše – Jizera Pluton, the whole Jizera Metamorphic Complex, substantial part of the Kaczawa Metamorphic Complex, the whole South Krkonoše Metamorphic Complex, the South Krkonoše Piedmont Late Paleozoic Basin, the eastern part of the North Sudetic Depression and the marginal part of the Czech Cretaceous Basin. The unified gravimetric maps of this extended area were compiled using gravimetric data advanced by the Polish Ministry of Environment in Warszaw, by the Saxonian State Department of Environment and Geology in Dresden, and by the CGS-Geofond in Prague.

The regional (low pass) map is depicted in the Fig. 1. The most remarkable gravity anomaly is a large gravity zone L1-L2-L3 situated in the central part of the area studied. This zone is almost 100 km long with the axis drawn-out in the direction of WSW-ENE. Three partial gravity lows are developed along this axis. The westernmost low (L1) reaching −51 mGal is situated in the northern marginal part of the Czech Cretaceous Basin, the central low (L2) with the extreme of −40.5 mGal occurs in the Jizera part of the Krkonoše – Jizera Pluton (in the surroundings of the town of Liberec) and the largest eastern partial low (L3) of −48 mGal is on the northern Polish slope of the Krkonoše Mts. in the southern vicinity of the town of Jelenia Góra. All the three partial minima represent the effect of the low-density variscan granite rocks of the Krkonoše – Jizera Pluton (in case of the L1 low there is also a substantial influence of the "light" Cretaceous sediments covering the buried granites). The large Jizera Metamorphic Complex situated to the N of the Krkonoše – Jizera Pluton which is mostly built by various kinds of orthogneisses and migmatites also partly contributes to the central gravity low.

The partial lows L1 and L2 are separated one from another by the local gravity high H1 caused by the Proterozoic to Early Paleozoic sedimentary-volcanic sequences building the Ještěd-Kozákov Mountain Belt.

The steepest horizontal gradients rim the main gravity low especially on its southern and eastern margins. In the direction to the South, it reflects the density contrast toward the South Krkonoše Metamorphic Complex built by metamorphosed Proterozoic and Early Paleozoic sedimentary-volcanic sequences covered by the South Krkonoše Piedmont Late Paleozoic Basin. The Late and Early Paleozoic Complexes cause the gravity high H2. In the direction to the E the steep horizontal gradient manifests the contact with the Kaczawa Metamorphic Complex covered partially also with the Late Paleozoic Formations. The Kaczawa Complex and its Late Paleozoic cover create the gravity highs H3 and H4.

The gravity modelling focused to the shape and deep position of the Pluton body was solved using the software GM-SYS along the two almost 150 km long profiles. The first one of the S-N direction started in the Czech Cretaceous Basin, crossed over the western Jizera part of the Krkonoše – Jizera Pluton (partial gravity low L2) and finished in the Odra Lineament Zone. The second one of the SSW – NNE direction began also in the Czech Cretaceous Basin, crossed over the gravity high H2, the western partial gravity low (L3), then the whole Fore-Sudetic Block and finished also near the Odra Lineament Zone. The resulting models show very steep (almost vertical) southern wall of the Pluton. The bottom boundary of the Pluton is expected to be in depth of about eight kilometers in the Jizera Mts. and to ten kilometers in the Krkonoše eastern part of the Pluton.

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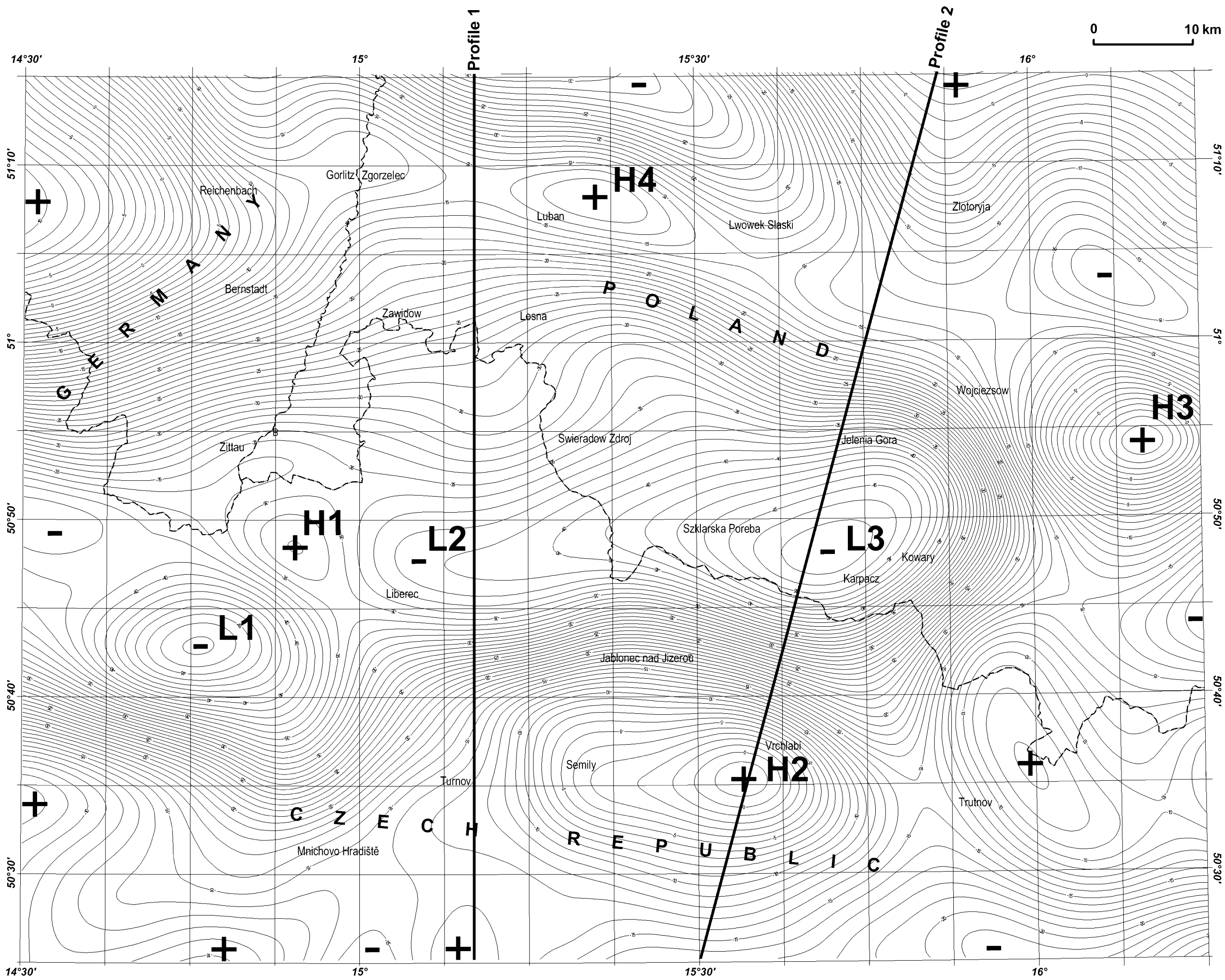

■ **Fig. 1.** Regional gravity anomalies with location of the two interpretation profiles.

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The Problem of Garnet Composition in Eclogite-Bearing Gneisses from the Śnieżnik Metamorphic Complex (Western Sudetes)

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In the Śnieżnik Metamorphic Complex (ŚMC) – the eastern part of the Orlica-Śnieżnik Dome, Western Sudetes, there are numerous bodies of eclogites. They outcrop as small, usually several tens of meters long, lensoidal bodies inside gneisses. Mostly they are surrounded by two-feldspar orthogneisses, rarely they also contact with plagioclase paragneisses. Discovery of possible pseudomorphs after coesite in omphacites resulted in conclusion, that the eclogites experienced ultra-high pressure metamorphism (UHPM) (Bakun-Czubarow 1992). During retrogression the (U)HP metabasites were extensively amphibolitised, especially the outer parts of the bodies.

The orthogneisses surrounding the eclogites frequently contain accessory grains of garnets. Some of the garnets display unusual composition. They are almandines with a high content of grossular mole fraction, reaching up to 50 %. Simultaneously they are poor in pyrope component (1–7 %). These Ca-Fe garnets have been interpreted as indicators of ultra-high pressure metamorphism of the migmatic orthogneisses directly contacting with the eclogites (Bröcker and Klemd 1996). They were recognized together with rutile as remnants after UHP mineral assemblage. The clinozoisite and sphene, also present in the gneisses, were interpreted as the products of retrogressive reaction *Grt(Grs30−50)+Coe+Rt → Zo+Spn*. Garnets with high grossular content were also observed in the ultra-high pressure gneisses from the classical UHP terrane: Dabie-Sulu in China (e.g. Carswell et al. 2000), what strengthens the above-mentioned interpretation. The recognition of the Grs-Alm garnets as UHP mineral relics in gneisses is a good support for generalized interpretation of the whole Śnieżnik Metamorphic Complex as an ultra-high pressure terrane (Gordon et al. 2005).

The high-grossular garnets have been reported by Bröcker and Klemd (1996) in only one part of the ŚMC, so called the Międzygórze unit (MU). The garnets occurring in eclogite-bearing gneisses outside the MU normally do not have a high-Ca composition. They are typical almandines (*Alm 60–87.5, Sps 2.5–20.5, Grs 1.5–18 (mostly<12), Prp 0.2–10.5, And 0–1.5*). The diversity of mole fraction ranges in the almandines is strongly connected with a whole rock chemistry of the gneisses, in which they occur. Regardless of Grs-Alm or typical Alm composition, the garnets usually have anhedral shape, sometimes they form skeletal or atoll grains. All the garnets often display strong effects of resorption, what gives them a relic look. Mostly the blasts have small size: from tens to hundreds of micrometers. Occasionally they are being replaced by biotite and chlorite. No inclusions of high pressure minerals (Jd, Ky, Zo, Rt) or quartz pseudomorphs after coesite have been observed inside the garnets.

The studies of the present author have revealed rare cases of Ca-Fe garnets in the localities outside the Międzygórze Unit, in the Gierałtów (GU) and Śnieżnik Units (ŚU) (Stawikowski 2005). Borkowska et al. (2003) also mentioned about the find of high-Ca garnets in the orthogneisses from the Gierałtów Unit. The granite gneisses from the GU and ŚU containing the Ca-rich garnets, like in case of the MU, either show the direct evidence of migmatisation or, if not, gradually pass into migmatites. Usually the Grs-Alm garnets were found in them in the nearest neighbourhood of eclogites, more precisely, close to their amphibolitised outer shells. They were sampled at distances of milimeters or centimeters from post-eclogitic amphibolites. Also in the MU, the biggest number of garnets was observed in the rocks situated very close to the metabasites. Such a regularity led the author to the hypothesis about the growth of high-Ca garnets in upper-amphibolite facies conditions, due to migration of Ca ions from metabasites to gneisses during eclogite retrogression (Stawikowski 2005).

Recently, the high-grossular garnets have been discovered near Strachocin in the Gierałtów Unit, about 50 meters from the large body of post-eclogitic amphibolites. The garnets occur in the migmatic orthogneisses, displaying partially pegmatitic appearance. The migmatites are composed of an assemblage *Qtz+Pl+Kfs+Bt+Grt+Spn+Ap+*Zr. The rocks comprise numerous newly-grown porphyroblasts of antiperthitic plagioclases, reaching up to 30 mm. The size of the garnet blasts is also bigger than in majority of eclogite-bearing gneisses from the ŚMC. They form crystals up to 5 mm in diameter. Their composition is *Alm 59.5–71.5; Grs 21–34.5; Prp 2–6; Sps 1–3; And 0–2*. The diversity of garnet chemistry is caused by retrogression. The crystals are heavily fractured and show bigger amount of Fe and Mg, whereas smaller content of Ca, in the areas situated close to chloritised biotites. They do not contain the inclusions of HP mineral relics. Frequently, garnet grains have subhedral to anhedral shapes and are in equilibrium with the big neoblasts of feldspars. The Ca-Fe garnets are accompanied here by sphene, usually absent in the granite gneisses outside the Międzygórze Unit. The textural relationships between the minerals building the migmatised metagranites indicate common growth of the feldspar porphyroblasts and garnets, as well as the sphene. One can deduce, that they build together the mineral assemblage connected with a migmatic event. As the migmatisation must have been induced either by temperature rise or drop of pressure (decompression), both these explanations do not fit well to the conception of blastesis of the Ca-rich garnets within the investigated gneisses in UHPM conditions.

The ambiguity of the UHPM genesis of Grs-Alm garnets in the orthogneisses is emphasized by different composition of garnets appearing in ultra-high pressure granulites from the ŚMC. The bimodal granulites build here several km long belt called the Stary Gierałtów Granulitic Complex. The garnets in the felsic granulites display significantly higher amount of magnesium, typical for HP metamorphism (*Alm 48–53, Prp 19–23, Grs 25.5–29.5, Sps 0.5–1, And 0–0.5*). On the other hand, the whole rock chemical composition of the orthogneisses and felsic granulites is analogous. Also the garnets from thin intercalations of felsic high-pressure rocks in the ŚMC eclogites contain more Mg than the ones from eclogite-bearing gneisses (*Alm 37–62.5, Prp 8–35, Grs 4.3–37.5, Sps 0.5–13.5, And 0–2*).

In conclusion, the high-Ca composition of garnets occurring in part of the eclogite-bearing gneisses may not be a sufficient evidence for common, in situ (ultra-)high pressure metamorphism of the eclogites and their host orthogneisses in the Śnieżnik Metamorphic Complex. Origin of these Grs-Alm garnets can be connected not with the burial to the mantle depths, but with migmatisation, possibly in the upper amphibolite-facies conditions.

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The Basement of Eastern Part of the Polish Carpathians in the Light of Geophysical Data Interpretation

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The study area is located in the eastern part of the Polish Carpathians, east of the Wisłoka River valley, and includes a transition zone between the Western and Eastern Carpathians. The structural rebuilding of the Carpathian orogen and its basement characteristic of that zone reflects in changes of geophysical fields, e.g. in the distribution of gravity anomalies (Bojdys and Lemberger 1986). Specific 3-D deformations of the flysch cover observed in the area suggest strike-slip faults in the basement. In the western part of the area, there probably occur a major tectonic zone in the basement that separates zones of different tectonics (Żytko 1997).

The recognition of the basement in that zone is not complete, generally because of the complex structure of the Carpathian overthrust and lack of deep boreholes, making interpretation of geophysical data difficult. As only a few boreholes located in the marginal zone of the overthrust reached the sub-Paleogene basement, its recognition is based on surface geophysical investigations. As a result of the complex structure of the orogen, the efficiency of the reflection seismic method is lesser. Hence, magnetotellurics, gravity method, geomagnetic soundings and refraction seismics are of greatest importance to investigations in that area.

Deep geomagnetic and magnetotelluric suondings have been made in Polish Carpathians since 1960s (Jankowski et al. 1991). Since then, wide regional surveys applying equipment of two different technological generations were made and different geological interpretations were presented (Woźnicki 1985, Ryłko, Tomaś 1995, Żytko 1997, Stefaniuk 2001). During the period of 1997–2002, a regional survey with the use of high-frequency MT system was made in the framework of "The project of magnetotelluric survey in Carpathians" (Stefaniuk 2003). Seven profiles crossing transversally the orogen and two profiles paralell to the general strike of Carpathian outocrops were located in the eastern part of the Polish Carpathians. Results of MT data interpretation enabled the structural map of the top of high-resistivity basement and maps of horizontal resistivity distribution for selected depths to be constructed. Resistivity cross-sections including elements of geological interpretation were made along measurement lines.

The origin of gravity anomalies in zones of folded orogens is connected mainly with thick under-compacted sedimentary series that fill deep basement depressions. The other source of gravity anomalies in the study area is a low-density zone in the upper mantle (Bojdys and Lemberger 1986). The qualitative analysis of residual gravity anomalies computed for selected depth intervals enabled extreme zones of horizontal gradients connected with vertical or steep density boundaries to be evaluated. Such gradient zones are probably related with tectonic contacts.

Based on the above mentioned results of magnetotelluric and gravity data interpretation, a spatial model of the basement was constructed. The model includes a structural map of the top of the Precambrian basement related to main regional resistivity boundary, a structural map of the top of Mesozoic and Paleozoic basement and major tectonic zones. The map of refraction horizon related to the top of sub-Paleogene basement was also used in model construction.

Acknowledgments

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Application of Paleomagnetic Methods for the Tectonic Study of Northern Variscan Thrust Front

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Paleomagnetic study has been carried out within a 150 km long segment of the Variscan external fold–and–thrust belt of N France and S Belgium. Main target was about its tectonic development and particularly the origin of its curved shape. The carbonate rocks have been sampled in numerous locations in the Ardennes Massif. The sites have been localized along the fold-belt at the similar distance from its front in order to compare the paleomagnetic records from the tectonic structures characterized by different orientations of the fold axes but with a same age of deformations. Some others carbonate sites has been spotted inside specific tectonic structures in single outcrop (Bettrechies) in order to compare relative age of deformation and remanence acquisition. The sandstones have been collected in the Ardennes and in the Artois Massif. In the Ardennes the sites containing sandstones have been located both in the middle part of the thrust-belt and in its marginal part.

Within magnetite and pyrrhotite bearing Devonian and Carboniferous carbonates, two secondary components were evidenced. Inclination-only tests indicate the synfolding origin of both components: the high temperature component (HT) was acquired during the early stage of deformation while the low temperature component (LT) appears during the late stage. Results from Bettrechies enable to correlate diagenesis events with remagnetizations episodes and progressive folding. Outcomes obtained for the Lower Devonian reddish sandstones indicate presence of a hematite carrier and syn- or post-folding magnetization, depending on the sampling site location.

Paleomagnetic directions from the carbonates display dependence on the local tectonic trend. Declinations of the HT component are similar to the directions known for Laurussia in regions of NE-SW orientation of the fold axis. Conversely, areas characterized by E-W to WNW-ESE structural trends show declinations rotated clockwise. Declinations show a correlation with the structural trend for both HT and LT components, but in the case of the LT component the magnitude of the declinations deviation is smaller. Results from the sandstones confirm the presented outcomes and additionally prove the heterochronic age of the deformations that differ between the marginal and the internal zone of the fold–and–thrust belt.

The presented declination data support only local oroclinal bending which give rise to the strike deviations in the thrust-belt. In the Ardennes clockwise rotations of the thrust occurred only within narrow transpressive zones, active during the propagation of the thrusts. It is also suggested that the long segment of WNW-ESE trending thrust-belt, that includes the Massive Artois, represents the oblique transfer zone between the Ardeno-Rhenish and SW England frontal belts.

Mechanics of Large-Scale Sand Injection – Understanding the Hamsun Giant Sand Injectite Complex

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Sandstone intrusions (injectites) are intriguing features as, despite their widespread occurrence, their origin is poorly constrained. The lack of process understanding poses a challenge to anyone dealing with post-depositional sediment remobilization. The formation of large-scale sand injectites has been attributed to various factors and processes such as: overpressure build-up, fracture propagation, fluidization, etc. Overpressure build-up can be caused by a variety of mechanisms such as disequilibrium compaction, loading by mass transport deposits, earthquakes, bolide impacts, or injection of fluids external to the sand body, such as, for example, hydrocarbons. Fractures start to propagate when pore-fluid pressure in a sand body exceeds the vertical or horizontal stress and the tensile strength of the host rock. Pressure-differential forces sediments to flow and fill fractures in the host rock. Depending on pressure conditions in the source bed and the seal and on the rheological properties of the host rock, sand injectites may form a range of geometries.

Clastic injectites occurring in the form of sills or dykes have been described for many decades. The size of clastic intrusions varies on a scale from sub centimetre to hundreds of metres. Recently, they have been recognized not only in outcrops but also on seismic data. A spectacular example is the Hamsun giant sand injectite complex that is located in the Paleogene of the North Sea. This complex is believed to be world's first sand injectite that was deliberately (and successfully) drilled by Marathon Oil UK as a hydrocarbon prospect, adding several tens of millions of barrels of oil to their Alvheim development. The Hamsun complex is sourced from the Hermod sand which occurs in Sele Formation and is believed to be of early Eocene age. The injectite complex was investigated by means of multi-volume-based 3D seismic interpretation and visualization in order to gain detailed characterization of the complex body. Overall shape of the body was analyzed, including its thicknesses, angles, depths, heights and relation to faults. Borehole core from two locations along the injectite were examined and constitute the ground truthing of the 'remote sensing' 3D seismic datasets. The investigations enabled drawing some conclusions about the Hamsun complex, like for example multi – phase injection.

Sand injectites are currently the subject of a concerted research effort at the University of Aberdeen, drawing on data from key outcrop analogues and selected sand injectite oil fields to catalogue the range of injectite styles, grain size variations, geometries and sizes, in order to establish genetic models and assist in reservoir modelling of sand injectite oil fields.

Record of Motion Along the Red River Fault Zone in Provenence Studies, Northern Vietnam

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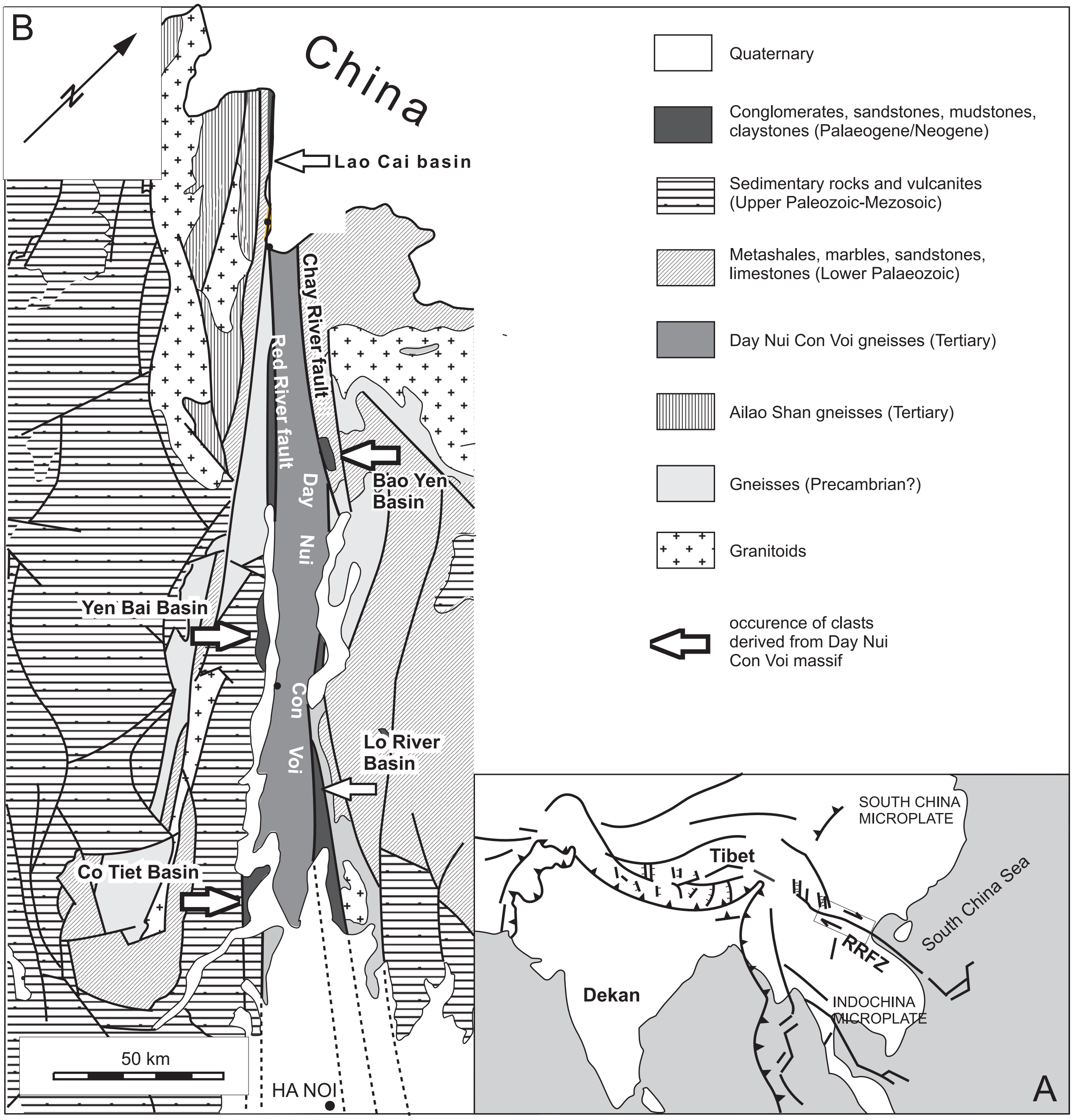
Provenance studies and, clast analysis in particular, are a valuable source of information on timing of uplift and denudation in source area. These studies may also document motion of a source area for basins related to strike-slip faulting. In this paper we present first results of clasts analysis from sedimentary basins adjoining the Red River Fault Zone (RRFZ) in Northern Vietnam.



■ **Fig. 1.** A. Location of studied region in SE Asia (rectangular). B. Geological sketch map of NW Vietnam showing location of the Paleogene/Neogene basins along the RRFZ.

The Red River Fault Zone (RRFZ) is a large strike-slip zone that separates Indochina and South China microplates. In the northern Vietnam (Fig. 1), the NW-SE trending RRFZ comprises two main faults: Red River Fault and Chay River Fault separated by up to about 25 km wide metamorphic Day Nui Con Voi massif (DNCV). The Red River Fault bounds the DNCV to the SW whereas Chay River Fault to the NE. Origin and uplift of this massif, formed by amphibolite facies paragneisses with minor contributions of mica schists, marbles and amphibolites, were related to left lateral movement along the RRFZ (Leloup et al., 2001 and references therein). However, dating of motion and estimation of amount of shifting along the RRFZ as well as time of exhumation are subjects of debate.

In the northern Vietnam, gneisses with bodies of amfibolite and metasedimentary rocks adjoin the RRFZ to the SW. Metasedimentary rocks dominate to the NE of the RRFZ. Close to the northern part of the RRFZ granitoid massifs are also present.

A few small sedimentary basins occur along the discussed faults. Their origin was probably connected to the tectonic activity of the RRFZ. (Leloup et al., 2001, Wysocka, Świerczewska 2002). The basins are filled with clastic Paleogene/Neogene strata representing fluvial and lacustrine depositional environments. The strata are characterised by numerous local facies changes.

Our studies were focused on two basins associated with the Chay River Fault and three basins associated with the Red River fault. The composition of the clast assemblages shows strong variability both in particular basins and between the basins. In all studied basins, the local source areas located outside of the DNCV are clearly marked. Clasts derived from the DNCV are recognized only in a few sites. The occurrence of these clasts in the basins associated with the Chay River fault show distinct differentiation. The clasts of the DNCV gneisses are observed in two exposures of deformed conglomerates in the Bao Yen Basin. The gneiss clasts were not observed in the second basin.

Along the Red River fault, in the Lao Cai Basin, only single clasts of gneisses were observed. The vast majority of the clasts are formed of granitoid clasts derived from Ailao Shan massif. It is not clear if the gneisess are derived from the DNCV. Further SE, in the Yen Bai Basin, the clasts of gneisses occur in undeformed conglomerates what suggest post-motion age of the conglomerates. In the Co Tiet Basin, clasts of DNCV gneisses occur in deformed, probably Miocene, strata. Like as the gneiss clasts, detrital garnets are common only in some samples of heavy minerals separated from Paleogen/Neogene sandstones of studied basins. Composition of these detrital garnets points to differentiation of metamorphic source area: from amphibolite to greenschists facies.

Presented results show that only for small portion of the fill of the basins, the high metamorphic DNCV was a source area. Poor dating of the sediments filling basins (palyonological data only) does not allow to precise stratigraphic position of strata containing clasts derived from the DNCV. Basing on degree of deformation of sampled strata it seams that the relationship between sedimentation of gneisses-bearing strata and deformation related to RRFZ activity is different for particular basins. Only in vicinity of Bao Yen and Cot Tiet basins uplift and exhumation of the DNCV was coeval or pre-dated deformation recorded in the sedimentary rocks. For these basins the offset along the RRFZ up to 200 km cannot be excluded.

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Structural Analysis and Paleostress Reconstruction of the Spišská Magura and Podhale Region

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The area of Spišská Magura and Podhale region is composed mostly of Mesozoic and Paleogene sequences. The most measurements were done in the Paleogene sediments.

In the study area we distinguished ten deformation stages connected with 1) E-W compression generated in strike-slip stress regime (?Paleocene), 2) NW-SE compression and perpendicular (NE-SW) tension generated in compressive strike – slip regime (Egerian–Eggenburgian), 3) NNW-SSE extension generated in pure extensive tectonic regime (Eggenburgian – Ottnangian), 4) NE-SW extension generated in pure extensive stress regime (Ottnangian–Karpatian), 5) NW-SE extension generated in pure extensive stress regime (Karpatian), 6) NW-SE compression activated in pure compressive stress regime (Badenian), 7) NNW-SSE compression generated in pure compressive stress regime (Sarmatian–Pannonian), 8) NE-SW compression activated in compressive strike-slip stress regime (?Pannonian), 9) NW-SE extension activated in pure extensive stress regime (?Pontian–Pliocene), 10) ENE-WSW extension generated in pure extensive stress regime (?Pliocene–Quaternary).

These ten deformation stages we divided in two groups.

The first group contains structures that were rotated to their recent position depending on uplift of the crystalline core of the High Tatras Mts. that started in Upper Miocene, according to FT dates from apatites (Kráľ 1977, Kováč et al. 1994, Struzik et al. 2003). This group contains the first six deformation stages originated from ?Paleocene up to the Badenian period.

The second group contains last four deformation stages that are the youngest structures originated after tilting of the High Tatras Mts. from ?Sarmatian up to the Quaternary period.

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Garnet Pyroxenites from Eastern Transylvanian Basin: an Integrated Textural and Geochemical Study

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Since the lower crust and the upper mantle cannot be sampled and studied directly, deep seated xenoliths from basaltic, kimberlitic and lamproitic extrusions provide important information on the petrologic and geochemical composition, rheological state, thermal evolution of the lithosphere. These xenoliths, fragments of wall rocks entrained by magmas at upper mantle and lower crustal levels, have been carried to the surface by alkaline basalts extreme rapidly, probably in less then 60 hrs (Kushiro et al. 1976, Mercier 1979).

Petrologically, the mantle xenoliths are mainly peridotites (lherzolite or harzburgites) with lower amount of spinel and garnet pyroxenites which represent less than 10 % of the total volume of shallow mantle in the Carpathian–Pannonian Region based on our experience. Pyroxenite layers (veins? lenses?) can be seen as small-scale heterogeneities in the geophysical studies, however these methods cannot offer a detailed picture of the lower crust and the upper mantle (Chen et al. 2001). Garnet pyroxenite xenoliths are rare in alkaline basalts; some examples are: Israel (Esperanca and Garfunkel 1986, Mittlefehldt 1986), SE Australia (Irving 1974, Wilkinson 1974, Griffin et al 1984, O'Reilly and Griffin 1995), SW USA (Shervais et al. 1973), Hawaii Islands (Wilkinson 1976, Frey 1980) and Eastern Transylvanian Basin, Romania (this study).

The Persani Mts. in the Eastern Transylvanian Basin is the easternmost Plio-Pleistocene alkaline basaltic volcanic field in the Carpathian–Pannonian Region. The products of the volcanic activity are lava flows and pyroclastic rocks, in which peridotites as xenoliths from the upper mantle can often be found. Besides peridotite xenoliths, spinel and garnet pyroxenites are also common. Garnet-bearing pyroxenites composed mainly of primary garnet, spinel, ortho- and clinopyroxene. The secondary mineral phases in the studied xenoliths are plagioclase, amphibole, spinel and ortho- and clinopyroxene. Textural observations suggest deformation events and mineral reactions, as the results of changes in stress, P-T conditions and melt/rock interaction during the evolution of the upper mantle beneath the region. Primary clino- and orthopyroxene frequently contain exsolution lamella of the other pyroxene (sometimes they are curved). Garnet often contains, amphibole, ortho- and clinopyroxene inclusions, exsolved needles of rutile and is always surrounded by symplectitic intergrowth of secondary ortho- and clinopyroxene, spinel and plagioclase.

Thermobarometric calculation was carried out based on electron probe microanalysis data of the primary rock forming minerals. Equilibrium pressure was estimated using garnet-orthopyroxene barometry (Harley and Green 1982), yielded between 1.4 and 1.7 GPa, whereas equilibrium temperatures are in the range of 1030–1140 °C (based on the garnet-clinopyroxene thermometers of Ellis and Green, 1979). The majority of the primary clinopyroxenes shows the usual chondrite normalized REE pattern of upper mantle clinopyroxenes coexisting with garnet (i.e. enriched in LREE and depleted in HREE). However, some of them are enriched in HREE, which is a simple enrichment in HREE of "normal" clinopyroxenes without changing their LREE concentration. The REE pattern of primary garnets shows depletion in LREE and enrichment in HREE, whereas that of the symplectite coronae around primary garnets is slightly enriched in LREE, showing flat REE pattern, sometimes with negative Ce anomaly. The bulk trace element composition of the garnet pyroxenites was calculated based on the garnet and clinopyroxene compositions and their modal abundance. The calculated trace element patterns are quite similar to each other and very similar to MORB composition, too.

The wide petrologic variability of the studied mantle xenoliths shows that the upper mantle beneath the Eastern Transylvanian basin is more heterogeneous than it was described previously (e.g., Vaselli et al. 1995, Chalot-Prat and Boulier 1997). Based on the textural relationships (e.g. the appearance of symplectites, plagioclase, curved exsolution lamellae) and the thermobarometric results, the evolution of the xenoliths can be outlined, indicating deformation and pressure decrease (upwelling) in the lithospheric mantle before alkaline basaltic volcanism. The inferred P-T-path of the Persani Mts. garnet pyroxenites agrees well with the previously studied former garnet peridotites (Falus et al. 2000).

The estimated paleogeotherm (older than the Plio-Pleistocene) beneath the region, shows slightly higher temperature than the present day heat flow calculations and, therefore, indicates significant cooling of the upper mantle after the cessation of the alkaline basaltic volcanism in the Persani Mts.

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Neotectonic Character of the Horná Nitra Depression

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The Horná Nitra Depression is situated in the western part of the Central Western Carpathians, and it is the elongate Upper Miocene to Quaternary structure in the N-S direction. This depression is bound by fault structures which were observed and measured during the neotectonic research. The aim of this work is to identify and define the main fault structures on the basis of the relevant tectonic geomorphology and structural geology methods used. The next step was the dating of fault activity during the Plio-Quaternary Period, and testing its ability to generate seismic events. The faults observed in the Horná Nitra Depression have been divided into three categories.

The first category consists of neotectonic active faults. In the Horná Nitra Depression, these consist of the Malá Magura fault and the north-west segment of the Pravno fault. These are faults whose activity during the Plio-Quaternary Period was able to be independently determined using several methods. The Malá Magura fault is the tectonic structure which divides the Tatric crystalline basement of the Malá Magura Mts. from the sedimentary fill of the Horná Nitra Depression. It is a typical mountain-front fault with a N-S striking and a dipping to the east. The dominant component of the movement on the fault plane is a normal slip, and the length of the fault is 16.71 km. The neotectonic activity is shown by the superposition of the Quaternary alluvial fans, by the value of the mountain-front sinuosity, by the mountain-front faceting, by the valley floor-to-height ratio, by the valley cross-section ratio, by the interpretation of aerial photographs and satellite images, and also by the geophysical measurements. The north-west segment of the Pravno normal fault is also neotectonically active, and it divides the sedimentary fill of the Horná Nitra Depression from the Pre-Tertiary rocks. This segment of the fault measures 4.71 km and it is in a NW-SE direction with an inclination towards the SW. The neotectonic activity is shown by the relationship with the Quaternary alluvial fans which are cut by this fault, by the considerable change of morphotectonic parameters (e.g. relief slope and segmentation, etc.), and the aerial photograph and satellite image interpretation.

The second category consists of faults which may possibly still have been active during the neotectonic period. In the studied area, these consist of the Nedožery, the Brezany, and the Hájske faults. These are faults whose activity during the Plio-Quaternary Period was not able to be unambiguously determined. The Nedožery fault is a N-S intra-depressional normal fault structure with a westward dipping and a visible length of 13.77 km. It separates the extent of the Pliocene Lelovce Formation on the east from the Quaternary alluvial fans on the west. This fault influences the Nitra river pattern, and it is identified in aerial and satellite images. The other methods of tectonic geomorphology do not reflect its activity during the Plio-Quaternary Period. The Brezany fault is also a N-S intra-depressional fault structure dipping towards the west. The dominant component of the movement on this fault plane is a normal slip, and the fault has a visible length of 10.86 km. This fault system divides the Lelovce Formation from the Biely Potok Formation (Oligocene), and it breaks the south-eastern segment of the Pravno fault. The relationships between the fault and the Quaternary sediments have not been definitely determined. Other methods of tectonic geomorphology do not reflect its activity during the Plio-Quaternary Period. The Necpaly fault is a NE-SW intra-depressional fault structure dipping towards the SE, with a visible length of 6.42 km. This fault structure breaks the volcanic sedimentary formations of the Upper Miocene (Sarmatian), and it cuts the south-eastern segment of the Pravno fault. This fault also limits the extent of the Lelovce Formation towards the Quaternary alluvial sediments, and it probably influences the size of the Holocene alluvial fans. The Hájske fault is a NE-SW normal intra-depressional fault dipping towards the NW with a length of 8.75 km. This fault limits the extent of the Lelovce Formation towards the south-east and it breaks the volcanic sedimentary formation of the Upper Miocene (Sarmatian), and it evidently cuts the south-eastern segment of the Pravno fault. The fault influences the Handlovka river and its Quaternary sediments, and it is clearly visible in aerial and satellite images. Other methods of tectonic geomorphology do not dipict its activity during the Plio-Quaternary Period.

The last category consists of neotectonically inactive faults in the Horná Nitra Depression. These are the Šútovce fault and the south-eastern segment of the Pravno fault, whose activity during the Plio-Quaternary Period, has been unambiguously eliminated. The Šútovce fault is a NW-SE polygenetical strike-slip structure with a subvertical dip. This fault is 9.51 km and it divides the Mesozoic and Paleogene sediments from the Tatric crystalline basement of the Malá Magura Mts. The last tectonic activity on the Šútovce fault was probably in the Middle Miocene age. Younger tectonic activity was not detected by any geological, tectonic or morphotectonic methods. The fault is not identified on the satellite images. The south-eastern segment of the Pravno fault separates the Tatric crystalline basement of the Žiar Mts. from the Mesozoic and Paleogene sediments. The fault length is 15.13 km with NW-SE striking and it dips towards the south-west. The dominant component of movement on the fault plane is a normal slip. The last tectonic activity on the fault probably occurred during the Upper Miocene age. Younger tectonic activity was not detected by any geological, tectonic or morphotectonic methods. Its NW end is covered by the Lelovce Formation, and the fault is clearly visible and identifiable on the satellite images.

The present-day stress in the Horná Nitra region Earth's crust was determined by paleostress analysis and tectonic geomorphological criteria. The principal maximum horizontal compressive stress SHmax was computed to be in a NNW-SSE direction, and the principal minimum horizontal compressive stress Shmin is perpendicular to this direction. This stress-field orientation may generate movement on the Malá Magura and the north-west segment of the Pravno faults. This data may be useful in compilation of a seismotectonic model of the area.

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Tectonics of Variscan Foreland Coalbearing Basin on Example of Karvina Subbasin – Upper Silesian Coal Basin

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The Upper Silesian Coal Basin (USCB), the part of the Moravosilesian area, could be classify as the foreland basin located in the toe domain of the European Variscan accretion wedge (Grygar and Vavro 1995, Dopita et al. 1997, Grygar et al. 2000). The Karviná sub-basin represents the most eastward transverse structural depression (Grygar et al. 1989) of the USCB. The coal-bearing Karviná formation (continental molasse – Namurian B) is cropping out on buried Pre-Alpine basement relief and is covered by sedimentary filling of West Carpathian Tertiary Foredeep and Outer Carpathian nappes.

The Variscan orogenic belt of central Europe represents a complex of crustal blocks accreted to the Laurussian foreland during closure of oceanic domains between Laurussia and Gondwana. The general kinematics of accretion in Saxothuringian and Moldanubian/Lugian zone of the Variscan belt was towards the northwest up to NNE, however presence of Brunovistulian Pan-African (Cadomian) terrane (microcontinent) on the eastward flank in the Moravosilesian area resulted in more complex and anomalous kinematics and transpressional character of thrusting inside of Variscan sedimentary accretion wedge of Moravosilesian zone (Grygar and Vavro 1995).

Transpressional thickening of the inner Lugo-Moldanubian domain of Variscan orogeny (Grygar and Vavro 1995, Štipská et al. 2001) was contemporaneous with grow and sedimentary filling of synorogenic Moravosilesian flysh foredeep and subsequently more outer (eastward) coal-bearing foreland molasse basin located on the Brunovistulian foreland. Final crust thickening of internal Variscan orogeny domains during Upper Carboniferous orogeny stages resulted in consequent top-to-SE up to E-ward thrusting of sedimentary accretion wedge. Underplated Brunovistulian foreland with its pre-Carboniferous (mostly Devonian limestone facies) sedimentary cover carried out essential role in character, kinematics development and space distribution of regional deformations structures. Most significant role in Variscan accretion wedge thrusting played oblique tectonic ramps. They correspond to subequatorial transverse (in relation to longitudinal main fold-thrust structure trend) tectonic zones and by them limited structure elevations and depressions. The Karviná subbasin corresponds to this type of structure depression. This structure pattern is well evident in the case of Karviná Central Thrust Zone (Grygar et al. 1989). A final stage of nappe thrusting is related to widely extend dextral transpressional along WNW–ESE and NW–SE striking mostly brittle shears zones, which are very common in Karvina subbasin as in the whole USCB.

Progressive development of deformation inside accretion wedge was conditioned by layer parallel slip (Fig. 1) and detachment thrusting promoted by high bedding anisotropy of cyclic coal-bearing lithology. Similar role belongs to lithological inhomogeneities (Devonian carbonate versus flysh facies, sandstones layers versus coal seams and/or shales etc.). Slickensides on the bedding planes and intrafolial fault indicate WNW–ESE up-to NW–SE compression. Recently known easternmost limit of thrust front reaches today post-erosional eastern limit of the Karviná subbasin (easternmost coal field of Czech part of USCB – ČSM Mine – Grygar et al. 1989, 1998, Koníček and Ptáček 1999 etc.). Next progressive deformation stage was represented by tectonic ramping (Fig. 2) and fault-bend folds structures. Main fold-thrust system (e.g. Michalkovice Antikline, Orlová fold-thrust structure etc.) striking NNE-SSW. Dominant kinematics asymmetry (vergency) of folding and direction of thrusting is E to SE-ward. In the whole accretion wedge so as inside Upper Silesian Coal Basin entire thrusts system display also statistically conjugated (fan-like) structure pattern and kinematics. However back-thrusting is primarily limited only along the western domain of flysh foredeep and also partially on the western zone of Upper Silesian coal basin (Ostrava subbasin).

■ **Fig. 1.** Brittle-ductile boudins and layer parallel mylonitisation in the tectonic zone of Central Thrust (roof of Seam No. 30 – Saddle Member – Namurian B, Coalface No. 300205, ČSM Mine).

■ **Fig. 2.** Duplex and small ramps tectonics in the roof of Seam No. 30 (Saddle Member – Namurian B, Coalface No. 300205, ČSM Mine) represents easternmost thrust tectonics (Central Thrust Zone) of the Variscan accretion wedge in the Karviná subbasin (USCB).

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Tectonic Features within the Cap Rock of the Mogilno Salt Structure, Central Poland

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Cap rocks mantelling upper parts of salt structures are attributed to rock salt dissolution occurring at the salt structure contact with overlying formations in response to circulation of unsaturated brine/ground water within the salt surroundings. They are commonly thought as uniform films protecting salt structures from outer factors. However, seismic and geological studies (e.g. Krzywiec *et al.* 2000, Wilkosz 2005) as well as salt mine catastrophic inundation (e.g. inundation of Wapno Salt Mine, Poland in 1977) have shown that cap rocks have complex structures, they are fractured and faulted, thus, they do not isolate salt series from the surrounding rocks that perfectly. The studies carried out over the Mogilno Salt Structure, central Poland, has proved that tectonic processes exert significant impact not only on a cap rock stability but also on its internal structure.

The Mogilno Salt Structure is one of 11 salt structures in Poland piercing through the Mesozoic cover up to the shallow subsurface (up to about 60 m below the surface). The structure has developed between the Triassic and present, therefore it remains in contact with various Mesozoic and Cenozoic formations. Its cap rock locally borders with Pleistocene deposits, indicating, thus, relatively recent episode of structure's uprise. The cap rock has differentiated thickness (77–190 m), morphology of the surface (± 100 m of relative height difference) and lithology. The latter was revealed by boreholes which evidenced three dominant rock constituents of the cap rock: gypsum, anhydrite and clays, forming altogether varying lithofacies. Additionally allochtonous sediments (gravels, sands, muds and lignite) occur within the cap rock.

This study aimed to analyse tectonic meso-scale structures occurring in the cap rock material in three drill cores. Alas the cores were not spatially oriented, thus only qualitative analysis was possible. The set of tectonic structures documented in the analysed rocks can be divided into two groups according to relative time of their development: (i) inherited tectonic structures and (ii) structures developed in the cap rock *sensu stricto*. The first group includes features developed in salt series during salt flow and they are observed in competent rocks (anhydrite/gypsum) incorporated as blocks into the cap rock. These are stylolites, slickolithes, joints/shear fractures and veins. All structures have varying orientation relative to the core axes dependent on overall block orientation and the stylolites, slickolithes, and veins depict variable geometry and petrographical characteristics throughout the cap rock. Phase changes between anhydrite and gypsum are also evidenced in their structure. The second group includes structures preserved in the gypsum-anhydrite-clay rocks originated due to salt series dissolution and these are represented by joints, shear fractures, shear zones and veins. Due to small area of observations the distinguishing between joints and shear fractures is arbitrary: joint system is attributed to rare fractures cutting the lamination almost vertically and the shear fracture system to those making almost constant angle of 30–50° throughout the cap rock. Joints are dominantly preserved in sulphate rocks and the shear fractures (as well as shear zones) are observed both in sulphate rocks and clays. Some shear fractures has transformed into microfaults as evidenced by slickensides and gypsum coatings with clear fibre lineation and older-vein offset. The latter features are also observed in shear zones which are demonstrated by 10 cm-wide zones of closely spaced fractures that make an angle of about 30° with the shear zone boundary. Generally the angular relationships between all types of fractures and the primary bedding in the cap rock indicate that the fractures developed due to vertical interaction of the salt structure occurring beneath the cap rock and the load overlying it. Timing of their origin can not be deciphered at present.

Progressive growth of gypsum crystals within the fractures (some crystals exceed 10 cm in length) has resulted in formation of continuous veins of varying thickness and locally to substitution of primary clay layers by gypsum ones. This observation indicates that tectonic factors both lead to disintegration of primary structure of the cap rock and to transformation of its lithological composition.

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Quaternary Tectonic Activity of the Central Part of the Polish Carpathian Foredeep, Evidences from Archaeological Open Site at Brzezie near Kraków

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The fossil graben and associated with it the normal faults and joints within the Vistulian and Holocene sediments are the object of considerations here. These structures were observed in the archaeological open site at Brzezie, in the central part of the Polish Carpathian Foredeep (Fig. 1A).

The normal faults cut the Pleistocene gleyed loess, laminated loess, Eoholocene buried soil and the lower part of the Mezoholocene deluvium that includes an archaeological artefacts from the Neolith and early Bronze Age (Fig. 1B). These structures die out within the middle and upper part of the Neoholocene deluvium including archaeological artefacts from the Lusatian culture. The normal faults strike mostly NNE-SSW and dip steeply about 65–850 (Fig. 1C). Some of them, the master normal faults, bound the fossil graben (Fig. 1B). The surfaces of the normal faults are slightly striated. The fault-slip analysis shows that the maximum principal stress axis (σ1) was in subvertical position, the minimum principal stress axis (σ3) was horizontal and WNW-ESE-directed (Fig. 1D). The joints occur within the graben and outside of it. They group into three sets: 1) the NNE-SSW-trending; 2) the WNW-ESE-trending and 3) the ENE-WSW-trending (Fig. 1E). The joints of the two first sets predominate. They form an orthogonal joint pattern, where the joints of the (1) set strike parallel to the normal faults and the joints of the (2) set strike perpendicular to them. Additionally, these joints are closely spaced close to the normal faults. Stewart and Hancock (1990) described the similar relationships between joints and faults and suggesting that the development of joints was connected with the normal faulting. Therefore we believe that jointing was simultaneous with faulting at Brzezie. The basement of the study area is cut by NE-SW-trending faults that represent fragment of the Kurdwanów-Zawichost Fault Zone (Fig. 1A). There are some evidences of sinistral reactivation of this fault zone during the Late Miocene and later (Rauch-Włodarska et al. 2005). The normal faults and joints observed at Brzezie could be caused by activity of the Kurdwanów-Zawichost fault zone during the Pleistocene and Holocene.

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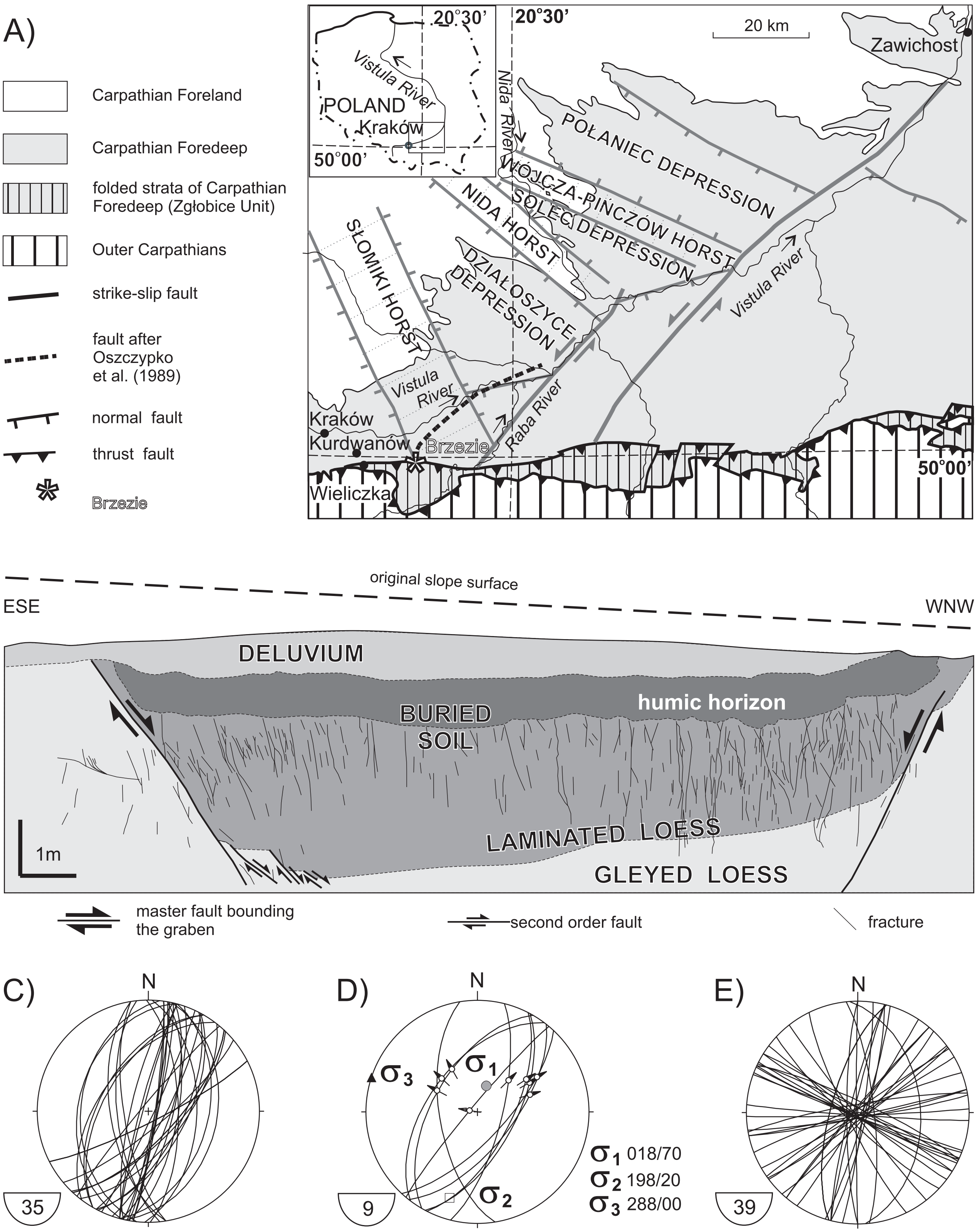

■ **Fig. 1.** A) Tectonic sketch of the central part of Polish Carpathian Foredeep (after Krysiak 2000) showing location of Brzezie, the Zgłobice Unit after Połtowicz (1991, simplified); B) Cross-section of graben; C) Plot of normal fault surfaces; D) Plot of normal faults with striations and orientation of reconstructed principal stress axes (using program TectonicsFP); E) Plot of joint surfaces. All plots on the lower hemisphere.

Kinematic Indicators of Slip Sense Along Faults in Loess Deposits: a Case Study from Fossil Graben at Brzezie, Polish Carpathian Foredeep

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The meso and microscopic indicators of slip sense along faults in a poorly indurated sediments are the object of consideration here. The mesoscopic structural data and sediment samples for microscopic studies were obtained from the archaeological open site at Brzezie in the central part of the Polish Carpathian Foredeep. Here, the Pleistocene loess deposits and the Holocene pedogenic and deluvium layers are cut by numerous fractures. The fractures are often enriched by dark grey fine grained material which is macroscopically similar to the Eoholocene humic horizon of the pedogenic layer covering the loess deposits. The pattern of fractures arrangement is complex. There are major fractures and minor fractures. The major fractures run regularly at a distance of 2 metres at least and dip approximately 65–85 °. The minor fractures have shorter length and wide scale range of dipping. They often display an anastomosing pattern. It is difficult to recognise directly offsets and slip senses along the fractures due to the absent of internal layering within loess deposits. Only the stratigraphic marker that is represented by lower and upper surfaces of the Eoholocene humic horizon of the pedogenic layer show decimetric-scale vertical offsets along some of these fractures. The mesoscopic slickenlines scarcely present on the surfaces of the fractures. Based on these rare indicators we recognise a few normal or oblique-slip faults (Rauch-Włodarska et al. in prep). Some of these faults are master faults which bound the small fossil graben.

There are other mesoscopic kinematic indicators for slip sense determination of both the faults and some fractures with negligible offsets on a mesoscopic scale. The minor fractures which display a geometry of Riedel shears or C-S shears predominate. They are represented by synthetic R fractures, antithetic R' fractures, synthetic hybrid fractures and synthetic C or S fractures. These fractures do not constitute a composite planar fabric but occur as single structural elements connected with major fractures and faults. Some of the minor fractures join a tip points of two parallel major fractures and faults, forming isolating lenses which are typical of linking damage zones (Kim et al. 2004). The R' fractures often make an angle 60 ° with major fractures or faults. The absence of R fractures observed here shows that the synthetic slip could be accommodated by slip along major fractures or faults. The R fractures play an important role in asymmetric boudinaging of dark grey layers occurring in core of the major fractures and faults.

The microscopic appearances of analysed fractures and faults were analysed using the images obtained from the optical microscope and SEM (back-scattered electron imagery). They are composed of distributed subsidiary shear zones and fissures. The shear zones are defined by both the elongated domains of alignment clay platelets and the clay interweaving bunches. These "clay particles" display a C-S or Riedel shears geometry. The sigmoidal fractures observed within the shear zones are represented by two groups. The sigmoidal tension gashes are arranged 'en echelon'. The extensional steps bordered these gashes dip in the opposite direction to the C-S shears. The offsets between these steps range from 22 μm to 0.23 mm. The other group of sigmoidal fractures contain extensional forms which are similar in terms of shape and orientation to the C-S shears. The offsets between steps observed here are similar (micrometric-scale). We believe, that this group of sigmoidal fractures uncharacteristic of shear zones were produced by shrinkage of "clay particles" during dewatering of sediments. In some places the shear zones are accompanied by the microscopic drag folds composed of "clay particles". Here, the sigmoidal tension gashes occurring within the normal limb show flexural-slip between folding layers of the "clay particles". The other microscopic kinematic indicators as delta-clasts, microthrusts and minor asymmetric folds around silt-sized particles were rarely recorded within the shears zones and thus play a minor role in determination of the slip sense along analysed major fractures and faults.

The slip sense defined on the basis of both meso and microscopic kinematic indicators generally agree with one another. Moreover, in the case of faults, it is confirmed by mesoscopic offsets. The described examples of kinematic indicators show that loess deposits are worthy of greater interest of structural geologists, exactly in their neotectonic studies.

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Analogue Modeling of AMS Development During Emplacement of Shallow Level Volcanic Bodies (Extrusive Domes and Laccoliths)

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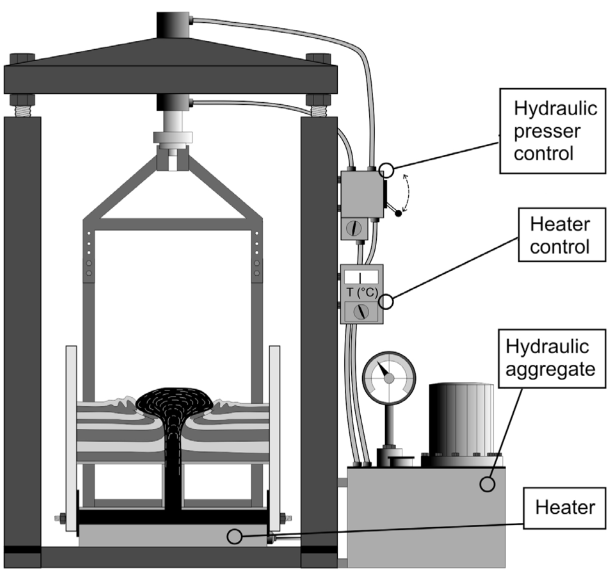
Volcanic domes and laccoliths are typical features of monogenic volcanic fields and are formed by highly viscous magmas either due to magma composition (andezites and rhyolites) and/or high crystal content (e.g. trachytes and phonolites). Although the evolution of shapes of extrusive domes and laccoliths was modelled for ideal newtonian fluids (Talbot and Jarvis, 1984, Koch et al. 1981) the evolution of internal fabric pattern during growth of such bodies is poorly understood. This is probably due to lack of good vertical cross-sections through such bodies and the microscopic nature of the fabric elements in volcanic rocks. Some aspects of the internal fabric development during viscous flow of magma within lava domes can be explained by the results of analogue and mathematical modelling of strain within lava flow and dome extrusions on flat surface (Buisson and Merle 2004). The models have shown that in the upper part of a lava flow (or flank of the dome), the maximum stretching axis is oriented perpendicular to the magma flow direction, while in the lower part lineations are parallel with the flow direction and diverge, where the flow extrudes radially. As the authors suggested, the results should be tested in field for various types of magma rheologies (e.g., shear-thickening dilatant rheology for crystal rich magmas, Smith 2002). Besides that natural magmatic domes often differ from the ideal flat based droplet geometry and their outer shape (and internal fabric) is controlled by complex interaction between the magma and host sediments.



■ **Fig. 1.** Scheme of hydraulic squeezing apparatus used for AMS modelling.

The aim of our study is to investigate the fabric generated by viscous flow within domes and laccoliths emplaced into weak sedimentary sequence by means of AMS analogue modeling. We follow a procedure of Kratinová et al. (2006) and use a hydraulic analogue apparatus (see Fig. 1) equiped with steel squeezing board and tube conduit and a perspex container filled with sand. The initial plaster and sand layers can be colored to visualize the deformation pattern within and around the model bodies. Sedimentary sequence is formed by pure sand or sand with clay layers, which induce zones of low tensile strength necessary for the emplacement of laccoliths.

Plaster still remains the most suitable analogue material for AMS modeling of viscous flow. It is cheap, easily colored and handled during the experiment; it is also easily homogenized with magnetic material. We have tested other analogue materials. Asphalt seemed promising, while it shows a range of temperature dependent viscosities, however it can not be colored and it is hardly handled in larger volumes. Silicones are expensive and can not be polymerized in order to bring them to solid state at the end of the experiment and carry out AMS sampling. Therefore we continue with rheological measurements of plaster of different mixing ratios of plaster and water with ambition to scale down our experiments for different magma rheologies, which are strongly dependent on the amount of solid particles (crystals). Flow of magma during filling of laccoliths and domes is for example typical with successive magma pulses (e.g. Mock and Jerram 2003). Such pulses penetrate discordantly the already present magma within inflating dome/laccolith and refold the surrounding magmatic fabric (Závada et al., this volume).

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The Mode of Flow and Emplacement of a Trachyte Body of the České Středohoří Mts. Studied by Means of AMS and EBSD Techniques

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The structural investigation of volcanic rocks is restricted by the small size of fabric elements, if macroscopic fluidality is not present. Therefore AMS is often employed, which is a powerful technique precisely investigating the orientation of the magnetic minerals (for review see Tarling and Hrouda 1993). The internal fabrics induced by flow of magma studied by AMS have focused mainly on the basaltic types, forming dykes or lava flows (e.g. Elwood 1978, Herrero-Bervera et al. 2001). Detailed structural analysis of more viscous volcanics due to magma composition and/or high crystal content (e.g. Smith et al. 1993) often forming domes and laccoliths was carried out much less frequently and AMS was rarely employed. The crystal-rich volcanic rocks often show conjugate textural domains (or microshear zones) interpreted to form due to extension or shear of the solidifying magma induced by "viscous drag" of still mobile magma and the bisector of conjugate shear sets indicates directions of maximum stretching and shortening (e.g., Smith et al. 1993).

In our study we have used an integrated AMS and EBSD approach to investigate the kinematics of magmatic flow within a trachyte body Hradiště u Habří and outline the style of its emplacement. We refer to the excursion guide of Šmíd et al. (2003) for the geological characteristics and petrology of the studied trachyte. The structural investigation was carried out using oriented thin-sections parallel with K_1K_3 and K_2K_3 planes of the AMS ellipsoid. The orientation of crystals within the textural domains was measured using the EBSD technique from total 12 thin-sections. The symmetry of the fabric was revealed on the basis of relative aerial representation of synthetic and antithetic microshear domains from image analysis of microphotograph sets of both perpendicular thin-sections. The correlation of the image analysis results and cluster patterns of the susceptibility directions of individual AMS specimens (8 cubes / locality) revealed three types of fabric, which form due to compression at high angle to the magmatic layering. Type I fabric shows equally developed conjugate sets of textural domains in both sections and is matched by girdle of K_2 and K_3 directions from 8 trachyte cubes measured. Type II fabric is typical with clusters rather than girdles of K_2 and K_3 directions and shows well and equally developed textural domains overprinting the primary crystal alignment exclusively in the K_1K_3 section. Type III fabric shows predominance of synthetic shear domains in the K_1K_3 section and equally developed conjugate domains in the K_2K_3 section and are characterized by very narrow clusters of the AMS directions. Type I and II fabrics are denoted as bearing orthogonal and Type III monoclinical symmetry. Since the fabric symmetry corresponds to the symmetry of deformation which caused it (Sander 1970, in Smith 2002) we can use the discrimination of fabric types to unriddle the symmetry of deformation and shear sense throughout the studied cupola using the AMS stereoplots. The interpretation of AMS clustering patterns using this classification assigns coaxial flattening and stretching parallel with the steep western margin of the body in the western rim, strong coaxial flattening in the central part resulting in intense subhorizontal fabric and non-coaxial flow on the northern margin and the eastern slope. In contrast to the rest of the body typical with strongly flattened fabric coupled with intense fabric-parallel fracturing, the southeastern part of the body exposes outcrops irregularly folded or trachytic fabric that is less clearly defined and less intense fracturing. The fabric in this area shows intense folding of the steep vertical magmatic fabric in the K_2K_3 section developing crenulation folds like in metamorphic rocks and the K_3 direction is perpendicular to the newly developed planar fabric element (c-planes). Textural domains showing regular kinks developed due to layer parallel compression of primary subhorizontal fabric are also present. The kinematic analysis revealed that compression axes inferred from several localities which show folding converge to one point and thus locate the feeding conduit. The magma flow and emplacement of the studied body is therefore also characterised by ascent of successive pulses, which cut discordantly through the already present magma and produce folds in the surrounding trachyte due to inflation of each magmatic pulse.

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Tectonometamorphic Evolution of the Svratka Crystalline Complex (NE Bohemian Massif): Evidence for Wrench-Dominated Tranpression along the NE Margin of the Variscan Orogenic Root

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Based on our structural and petrological data from the Svratka Crystalline Complex (SCC) in the northeastern part of the Variscan orogenic root (Bohemian Massif), we interpret tectonometamorphic processes during the later stages of the Variscan orogeny. The SCC is made up of high-grade migmatites, mica-schists, paragneisses and metagranites. The dominant regional fabric observed in these rocks is represented by ~NW-SE metamorphic foliation that dips at steep to moderate angles to the NE or SW. This foliation bears gently to moderately plunging NW or SE stretching lineation. The regional foliation is also roughly parallel to the contacts against the nearby geological units. Various stages of fabric development were recorded in microstructures of the coarse-grained and porphyric metagranites where two domains with different microstructures and finite strain patterns were recognized: (i) Low-strain domain (Vysoký kopec) is characterized by prolate finite strain ellipsoid, slightly fractured quartz aggregates retaining their magmatic shape, initial stages of K-feldspar recrystallization where the lattice preferred orientation (LPO) of new grains is homogenous and discordant to the regional fabric, and total recrystallization of biotite and muscovite. (ii) High-strain domain (Rabuňka) recorded oblate finite strain and is characterized by complete recrystallization and micro-scale deformation of all mineral phases with compositional banding, mechanical twinning and albite exsolution lamellae. LPO of the recrystalized aggregates is in this domain sub-parallel to the regional fabric. Furthermore, our petrological study from the micaschists indicates that that the SCC reached maximum PT conditions of 9 kbar and 670 °C. However, the regional fabric rather reflects the retrograde metamorphic conditions of 6 kbar at 640 °C.

Therefore, we argue that the regional fabric along the NE margin of the orogenic root recorded dextral wrench–dominated transpression at mid-crustal level. This study is supported by projects of Czech Geological Survey (CGS6328 and CGS6352) and by MSM 0021622412.

Textural Relations and Mineral Compositions of Retrogressed Low-Grade, High-Pressure Metabasites and Phyllites around Krkonoše-Jizera Complex and near Kraslice in Krušné Hory Mts.

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The rocks from both areas are strongly retrogressed into greenschist facies assemblages. Relatively fresh blueschists with strong foliation, defined by blue amphibole, phengite and epidote, are locally preserved in the eastern part of the Krkonoše–Jizera Complex. Some coarse-grained unfoliated varieties with primary igneous pyroxene and pseudomorphs of plagioclase are also present in this area. Beside blue amphibole, albite, epidote and phengite, other blueschist facies minerals in metabasites are garnet, chlorite, titanite and aegirine. Blue amphibole is usually replaced by chlorite and albite or it occurs in the core of actinolite grains. Boundaries between these two amphiboles are mostly sharp but continuous. Composition of blue amphibole ranges from glaucophane to riebeckite with $X_{Al}=Al/(Al+Fe^{3+})=0.37–0.75$ and $X_{Mg}=Mg/(Mg+Fe^{2+})=0.4–0.68$. Variation of X_{Al} contents in the blue amphibole is result from the whole rock composition, but also from zoning in individual crystal, where riebeckitic variety occurs in the core and glaucophane on the rim. Na-Ca amphibole was not found yet in these blueschists. Na-pyroxene occurs as thin rim around igneous diopsidic augite, which is rarely preserved in some coarse-grained rocks (gabbros?). It is aegirine (Di_{43-48}, Aeg_{40-45}) with low jadeite content (Jd_{8-12}). Epidote is rich in Fe ($X_{Al}=0.656–0.886$). Backscatter images show zoning of epidote represented by increase of X_{Al} from core to rim. Composition of phengite ranges in Si from 3.3 to 3.4 a.p.f.u.. Accessory biotite found in some retrogressed blueschists is rich in Fe ($X_{Mg}=0.535$ to 0.699). Chlorite composition ranges between $X_{Mg}=0.37–0.622$ and it is difficult to distinguish different chlorite generations. Garnet associating with blue amphibole occurs only in metabasites from the Kopina Hill, locality in the Poland side. Its presence was already described by Smulikovsky (1995). Garnet forms idioblastic grains with numerous inclusions of epidote, white mica, quartz and opaque minerals. It is rich in Fe (Alm_{55-70}, Grs_{25-35}, Py_{1-3}, Sps_{1-15}) and shows progressive zoning with decrease of Mn and Ca and increase of X_{Mg} towards rim.

Surrounding phyllitic rocks from the Krkonoško-Jizera Complex contain porphyroblasts of chloritoide in the fine-grained matrix composed of white mica, quartz and chlorite. Chloritoid forms small needles of different orientations and usually it crosses cut the foliation. In some very fine-grained rocks, chlorite forms porphyroblasts with interlayers of white mica. Chloritoid is rich in Fe with $X_{Mg}=0.078–0.083$, chlorite has $X_{Fe}=0.64–0.68$ and white mica is relatively rich in Si = 3.2 a.p.f.u.

The mafic rocks from the vicinity of Kraslice in the Krušné Hory Mts. are strongly retrogressed and the minerals relating to the HP/LT metamorphic stage are preserved only rarely. Amphibole composition ranges from Ca-Na- to Ca-amphibole. The Ca-Na amphibole corresponds to winchite with $X_{Al} = 0.16–0.22$ and $X_{Mg} = 0.7–0.72$. Calcic amphibole is actinolite in composition. Epidote is rich in Fe ($X_{Al} = 0.66-0.74$). Chlorite has $X_{Mg} = 0.32–0.62$ and accessory stilpnomelane was also observed. Plagioclase is pure albite with anorthite component of about 1 %. Surrounding phyllites contain, similar to that in the Krkonoško-Jizera Complex, Fe-chloritoid ($X_{Mg}= 0.09–0.12$) and chlorite ($X_{Fe} = 0.66–0.69$). In some fine grain varieties, chlorite forms porfyroblasts and it associates with phengitic white mica (Si= 3.2 a.p.f.u.).

Thermobarometric calculations with help of the PTGIBBS (Brandelik and Massone 2004) program used for phyllites with chloritoid indicate temperature of about 400 °C and pressure of 12 kbar. PT conditions, estimated based on mineral composition of blue amphibole, chloritoid and phengite in metabasites, are comparable with the epidote blueschist composition 6 of Evans (1990). Textural relations from the metabasite indicate breakdown of blueschist facies minerals, mainly of glaucophane and formation of chlorite and albite. The presence of biotite and actinolite rimming blue amphibole could be the result of nearly isothermal decompression after maximum pressures were reached.

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Cadomian Versus Variscan Fabrics in the Desná Dome Basement Rocks, East Sudetes

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In northern Moravia, the Czech Republic, the so-called Silesian units form two antiformal domes in which NE-SW elongate Neoproterozoic basement cores are flanked by medium and low grade metasediments of Devonian age. The eastern unit is referred to as the Desná dome (DD) and its cover on the east and north is known as the Vrbno Group (VG). In the DD core, medium to high grade metamorphosed schists, paragneisses, orthogneisses, migmatites, and amphibolites occur, dated isotopically on zircons between 644 Ma and 502 Ma (Pb-Pb evaporation, Kröner et al. 2000). The VG comprises Devonian quartzites, arenites and mudstones which underwent multiphase deformation and low grade metamorphism dying out easterly in Late Carboniferous times. Although an Alpine-type nappe stacking is the most commonly accepted explanation of the Variscan tectonics in the Silesian units, many problems like kinematics and sequence of deformations and metamorphism, direction of thrusting, extensional vs. compressional regime during dome formation, backthrusting, etc. are still unclear and debated. In the basement rocks, migmatitic fabric is usually interpreted as a Precambrian feature while mylonitic fabric is usually taken as a Variscan overprint, but no more detailed discrimnating criteria have been given.

In the Głuchołazy area, East Sudetes, SW Poland, there are isolated outcrops of Neoproterozoic basement and its Devonian cover considered a continuation of the DD i VG from the Jeseníky Mountains. In one of the outcrops, biotite paragneisses contain disrupted and folded layers and pods of quartz-amphibole-epidote schists. There are also subalkali tholeiite sills turned to amphibolites boudins. The latter resemble the Písečna-type massive amphibolites from the nearest part of the Jeseník Amphibolite Massif assigned to the Devonian. In the outcrop, the three types of rocks are crosscut by felsic injections of muscovite pegmatites to grt-bearing aplites which later underwent folding and shearing. Such relationships help to distinguish pre- and post injection episodes in the structural history of the gneisses and allow to constrain timing of the two deformational episodes. The new data help to clear some of the above problems and create some new ones.

The felsic injections range from up to 40 cm thick dykes pegmatite dykes down to < 1 cm quartzo-feldspathic variably discordant to concordant veinlets. Thinner veins are deformed in open asymmetric to ptygmatic folds. Larger folds have in their opposite limbs "z-type" and "s-type" parasitic folds respectively, which testifies to buckling and flexure. The relevant strain ellipsoid indicates an overall top-to-the W kinematics of folding. Besides, strain appears to be localized in narrow shear zones that displace the veins W-ward by up to 70 cm. Parts of the veins which got into the shear zones may acquire geometry of sigma–type clasts with top-to-the W sense of transport. Such high strain shear zones are heterogeneously distributed with spacing of 50 to 120 cm. The intervening areas are subject to flattening which accomodated the other part of the overall shortening.

The biotite paragneiss is composed of bi-pl-qtz-grt(chl) assemblage forming the excellent foliation dipping gently to the east. On the foliation, biotite flakes are arranged in one direction which defines mineral lineation plunging gently E-ward (parallel with weak corrugation lineation related to late kink folds). In sections perpendicular to the lineation, asymmetric pods and sigma clasts derived from the foliation-parallel quartz segregations show variable sense of movements, top-to the E and top-to the W. The latter is interpreted as concurrent with the deformation of the veins. The former is clearly older than the felsic veins as they intersect quartz-amphibole-epidote layers and pods which were earlier folded and/or dismembered by the top-to-east shearing. The same is true about asymmetric boudins of the metabasite sills.

Zircons retrieved from the foliated pegmatitic vein were analyzed by SHRIMP II machine. They have typical magmatic Th/U ratio of 0.2-0.9 and on the concordia diagram show 3 clusters of U-Pb ages at: 1420 Ma, 615 Ma, and 575 Ma. The youngest group is interpreted as a time of pegmatite injection in the post-tectonic period. The other Neoproterozoic age group likely reflects the main thermal event (anatexis, magma underplating) in the basement connected with the reworking of Mesoproterozoic crust (the oldest group). Accordingly the deformational structures produced with top-to-east kinematics (normal at the present-day orientation of the foliation) must be considered Precambrian and related to Cadomian tectonics, the termination of which is constrained by the pegmatite intrusion of late magmatic stages at ca. 575 Ma. It is suggested that a remarkable share of mylonitic features observed in basement rocks and attributed to Variscan orogeny is in fact Cadomian, although these may be hard to distinguish in the field.

Muscovite fractions from the same foliated pegmatite yielded an Rb-Sr isochron age of 289±2 Ma. The result shows that the Rb-Sr system in the analyzed micas was presumably totally reset in Palaeozoic times. The obtained data seems to imply post-tectonic cooling and uplift (influence of the Žulová granite is unlikely because of 10 km distance) after the deformation that involved an important phase of W-vergent tectonics in which the felsic veins became foliated and folded. Regionally, this corresponds well with Ar-Ar cooling ages of ~290 Ma obtained for amphiboles and micas from the Žulová granite pluton (Maluski et al. 1995) and with Ar-Ar cooling ages of 285–279 Ma determined in white micas from the Devonian Jegłowa Beds in the Fore-Sudetic Block (Szczepański 2002). They all point to uplift of the Moldanubian Fault Zone footwall at the Carboniferous/Permian turn when the the easterly tectonic transport was eventually stopped. Timing and significance of the W-vergent phase are not well constrained. Considering the E/NE-vergent piling of Devonian-Carboniferous flysch, it may represent the retrothrust/fold deformation in the Desná basement backstop. This explanation would correspond well with the presence of the W-vergent Andělská hora thrust identified by Cháb (1990).

In general, the Carboniferous deformation of the Silesian units has been defined as top-to-the NE thrust movements (D2) and dextral transpressional shearing (D3) in narrow, steeply dipping NE-trending shear zones (Schulmann and Gayer 2000). In northern part, in the continuation of the Vrbno Group to Poland, the WNW/NW-trending folds with predominant SSW vergence (D2), NE-trending folds with NW vergence (D3) and NW-trending folds with NE vergence (D3) have been described by Żaba et al. (2005). Our observations in the Vrbno quartzites and quartz-staurolite-garnet schists next to the Desná gneiss outcrop indicate early contractional deformation with ~S-vergent folding and thrusting followed by transpressional deformation which gave rise to the moderately to steeply plunging folds developed in the N/NW-dipping steep shear domains with dextral kinematics (or E/SE-vergent folds if foliation is gently dipping). However, none of those observation is consistent with the formation of the W-vergent structures (giver their Variscan origin) in the basement gneisses and veins. Therefore, we expect a significant detachment to exist between the Precambrian basement and the Devonian-Carboniferous cover units on the eastern side of the Desná dome (part of the Andělská hora thrust system?).

Some doubts probably can also be cast on the assumed Devonian age of the Jeseník amphibolites. If the observed massive metabasite sills in Głuchołazy are truly equivalent to the Písečna amphibolites, than the Jeseník massif amphibolites, or parts of it, must be assigned to the Precambrian units, and the metabasists are only folded and thrust together with basement paragneisses and Devonian metasediments.

Yet another constrain for the regional geology may come from the K-Ar analyses of the muscovite samples from the studied pegmatite (M. Banaś, Kraków K-Ar Lab). They yield an age of 233 Ma. Having accepted that the K-Ar system in muscovites (with the blocking temperature of ~250 °) was also entirely homogenized till post-Variscan times, this data may imply that the slow uplift and cooling of the area was punctuated by Triassic event, possibly related to the onset of Pangaea rifting and break-up. Similar Ar-Ar low temperature extraction ages around 220 Ma were determined in the Silesian units by Maluski et al. (1995).

A by-product of our data is a strong confirmation that the Silesian crust does not originate from West Africa Craton in contrary to the Saxothuringian and Moldanubian crust. This would mean that the Ordovician rift zone envisaged in the Staré Město belt (Schulmann and Gayer 2000) went beyond the stage of the thinned continental crust, reached the stage of full oceanic separation, and the pre-rift Cadomian counterparts of the Silesian units cannot be located in the so-called Lugian domain of the West Sudetes.

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Excursion Guide

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Conference Excursion 1: Structural Development of the Magura Nappe (Outer Carpathians): From Subduction to Collapse

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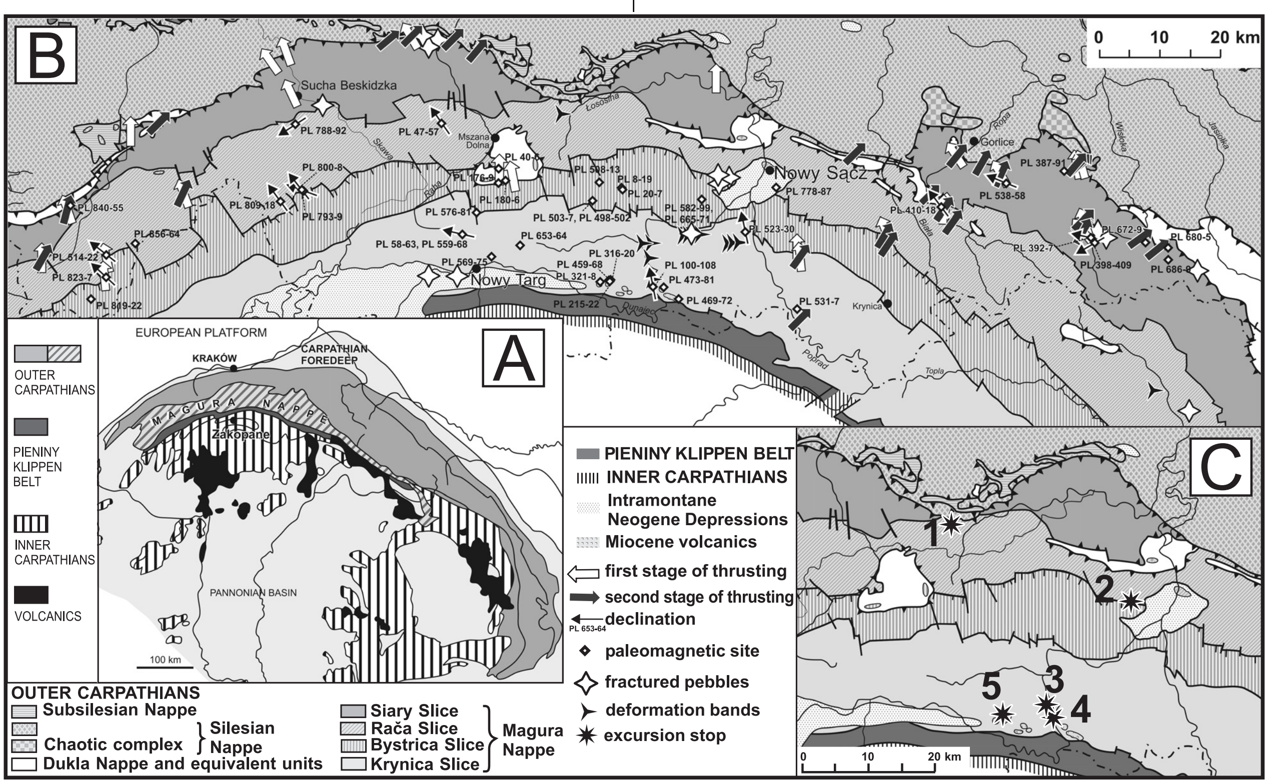
Introduction

The Outer Carpathians are a thrust-and-fold-belt, north-verging in the Polish segment (Fig. 1A). The belt, composed largely of Lower Cretaceous to Lower Miocene flysch strata, comprises several nappes. The innermost and largest of the nappes is the Magura Nappe. This nappe is subdivided by north-verging reverse faults into four slices which are named (from south to north) Krynica, Bystrica, Rača and Siary slices (Fig. 1B). To the north, the Outer Carpathian nappe pile is thrust over the Carpathian Foredeep, whereas to the south, the Magura Nappe contacts along steep faults with the Pieniny Klippen Belt (Fig. 1A). The Outer Carpathian nappe pile was formed due to Palaeogene-Neogene, southward-directed subduction of oceanic or sub-oceanic crust intervening between continental crust of the European plate and continental crust of the ALCAPA unit. The features produced during subduction-related shortening are overprinted by normal faults (Zuchiewicz et al. 2002 and references therein), which were formed during gravitational collapse. These faults bound intramontane basins filled with Neogene strata.

The map-scale structures of the Outer Carpathians have been recognised for a long time due to repeated mapping and numerous deep wells. On the other hand, still ten years ago, very little structural research have been done in the Outer Carpathian. During this excursion (Fig. 1C), we are going to show the results of our research of the past ten years.



■ **Fig. 1.** A – structural scheme of the Carpathian arc; B – map showing: (1) results of kinematic analysis of the Magura Nappe (after Decker et al. 1999 and Galicia T. Group unpublished), (2) results of palaeomagnetic study in the Magura Nappe (after Márton et al. 1999, 2004 and Márton and Tokarski unpublished), (3) location of exposures of gravels and paraconglomerates bearing fractured clasts and (4) location of exposures in deformation bands were studied, C – itinerary of excursion.

Kinematic analysis

We have analysed small-scale structures at over 120 exposures located within all nappes. The results (Fig. 1B) show that the Outer Carpathian nappe pile was formed due to two successive stages of thrusting, the first one directed towards the NW, the second one to the NE. The inferred succession of thrusting confirmed earlier results of Aleksandrowski (1985) from the western part of the Magura Nappe. On the other hand, our results raised important question: whether the succession of differently verging thrusts results from clockwise far-field stress rotation or from anticlockwise body rotation of the Outer Carpathians. To answer this question, we have undertaken extensive palaeomagnetic study.

Palaeomagnetic study

From the Magura Nappe, fine-grained members of the flysch sequence, preferably marls, were collected for palaeomagnetic study at 34 localities (300 samples) distributed along the arc of the nappe (Fig. 1B). From the Mszana and Szczawa tectonic windows, additional four localities (24 samples) and from the Upper Badenian marls of the Nowy Sącz intramontane basin, two localities (21 samples) were sampled. As a result of standard palaeomagnetic measurements and evaluation, 15 localities from the Magura Nappe and one locality from the Nowy Sącz Basin yielded statistically good or acceptable palaeomagnetic directions (Table 1). They all indicate counterclockwise rotation, which is, however, characterizing post-folding/tilting tectonic movements, since the remnant magnetizations were acquired after folding/tilting (exception may be locality Pl 840-855). There seems to be no systematic change in declinations as we proceed from the west to the east, and although inclinations cluster around 60°, there are a few localities with much shallower inclinations increasing the within-locality scatter.

Based on 14 localities (Pl 840-855 excluded) the palaeomagnetic mean direction of post folding/tilting age is Dec = 306°, Inc = 57° (k = 15, α_{95}° = 11), suggesting a general counterclockwise rotation of about 50° of the Magura Nappe, which could have taken place before the late Badenian, since declination for a single locality from the Nowy Sącz Basin does not show any deviation from the expected stable European declinations. For more details see **Stop 5**.

Timing of deformation

The termination of folding of the Magura Nappe is fairly well dated by: (1) the age of Miocene andesites which cut the already folded Magura sequence, and (2) the age of the unfolded Badenian strata filling Nowy Sącz intramontane basin. Conversely, the timing of the onset of folding has been not known; however, it had been traditionally accepted that it took place during the Early Miocene or Late Oligocene times (e.g. Roca et al. 1995). This opinion mostly comes from the apparent lack of angular unconformities within the flysch sequence, except for the local unconformities at the base of the Oligocene strata (Książkiewicz and Leško 1959), and within the Upper Eocene strata (Węcławik 1969); both unconformities occurring in the inner part of the Magura Nappe.

Locality	N (n)	D°	I°	k	α_{95}°	Dc°	Ic°	dip
Tenczyn Pl 47-57	7/11	321	+48	14	16	284	+70	168/30
Łąkcica Pl 100-108	5/9	149	−28	40	12	151	−5	352/25
Wołowiec Pl 392-397	6/6	282	+60	44	10	248	+23	220/50
Radocyna Pl 398-409	5/10	240	+64	26	15	224	+27	211/39
Łosie 410-418	8/9	323	+47	31	10	260	+42	205/56
Złatna Pl 514-522	5/9	128	−60	40	12	217	−86	120/30
Barcice Pl 523-530	6/9	168	−61	82	8	153	−59	266/9
Ropica Pl (538-558) 547-551	5/5 (21)	292	+57	37	13	264	+32	230/36
Klikuszowa Pl 559-568	4/11	283	+66	41	14	263	+37	245/32
Gołynia Pl 788-792	4/5	237	+59	25	19	183	+40	138/40
Zubrzyca I. Pl 793-799	5/7	290	+66	31	14	280	+69	160/5
Zubrzyca II. Pl 800-808	4/9	322	+31	28	18	302	+58	170/33
Zawoja Pl 809-818	7/10	324	+59	37	10	320	+36	315/23
Glinka Pl 823-827	4/5	313	+46	71	11	315	+47	345/5
Milówka Pl 840-855	7/16	84	+21	25	12	85	−20	150/85

■ **Tab. 1.** Palaeomagnetic results from the Magura Nappe.

The goal of our studies (Świerczewska and Tokarski 1998) was to demonstrate that folding in the Outer Carpathians started earlier than hitherto thought. We attempted this by dating the history of folding in relation to increasing induration of the strata involved. This was done by the analysis of deformation bands microstructures which indicate the degree of induration during a deformational event.

Deformation bands (DB) (Antonellini et al. 1994 and references therein) are roughly planar features which occur in porous granular materials. In sandstones, they form bands which are up to few millimeters thick and up to few hundred meters long. DB are accommodating offset like faults, but they do not contain planes of displacement discontinuity. DB are widespread in the strata of the Outer Carpathians. We have observed them within the Silesian and Magura nappes, in strata ranging from the latest Cretaceous through Late Oligocene in age.

We studied DB within strata of the Magura nappe (Fig. 1B) which is the innermost tectonic unit in the Polish segment of the Outer Carpathians. Assuming piggy-back style folding and thrusting in the Outer Carpathians (e.g. Decker and Peresson 1996 and references therein), it appears reasonable that folding first started in the Magura Nappe. Altogether, we have studied DB at 10 localities. For details and results see **Stop 1**.

PT conditions

Illite/smectite studies

X-ray diffraction studies of illite-smectite separated from about 270 claystone samples were use for a reconstruction of the burial and thermal history of the Magura Nappe (Świerczewska in press). Maximum palaeotemperatures were estimated based on the degree of smectite to illite transformation. The palaeotemperatures were calculated using the plot of Šucha et al. (1993) with a 10 °C correction, as suggested by Clauer et al. (1997).

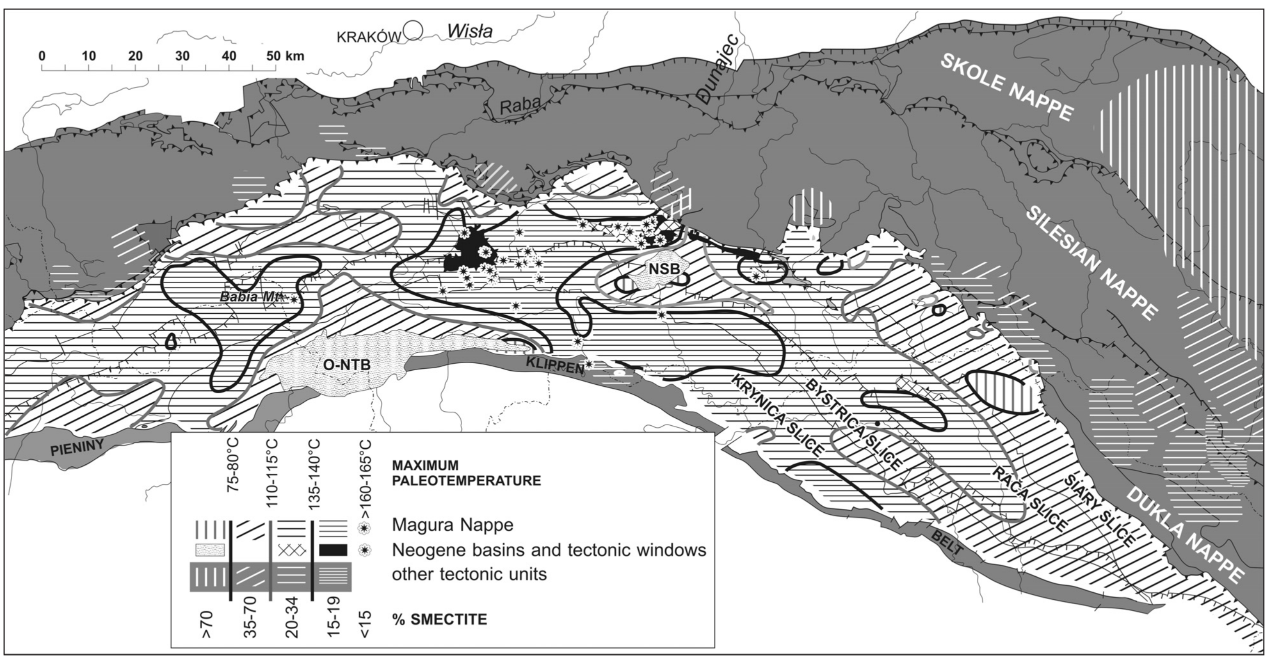
The Magura Nappe claystones contain from ≪5 to 90 % smectite in illite-smectite (I/S). These compositions indicate that the observed thermal alteration of the rocks on the present-day erosion surface of the nappe reflects temperatures that ranged from ≪75 °C to 200 °C (Fig. 2). These temperatures were related to tectonic burial at depths of between <4 and 11 km, assuming a mean value of 18 °C/km for the palaeo-geothermal gradient. For those exposures where this low palaeogradient was constrained by fluid inclusion data, burial depths calculated using I/S data agree with those based on fluid inclusions (cf. Hurai et al. 2004). This coincidence indicates that the influx of water-methane fluids and the maximum thermal alteration were coeval. The calculated values are a measure of the maximum thickness of eroded cover rocks.

Summary results of I/S studies for individual slices of the Magura Nappe and for the entire nappe show that for each slice, and for the entire nappe, there is a positive correlation between the age and degree of alteration of the sampled rocks. Only in small discrete portions of the nappe this correlation is not visible. Within the entire Magura Nappe, there is a clear general decrease in thermal alteration from the inner part outwards. A local and less distinct trend of decreasing thermal alteration inward, towards the Inner Carpathians, is apparent in the innermost part of the Magura Nappe. The maximum thermal overprint, corresponding to 160–165 °C and more, is registered in the central part of the nappe, especially in the region where tectonic windows occur. Rocks affected by the lowest palaeotemperatures (≪75 °C) occur only on the edge of the intramontane Nowy Sącz Basin, and in scattered localities in the eastern part of the Siary slice. In the central segment of the frontal overthrust, the Magura Nappe is thrust over significantly less altered rocks of the Silesian Nappe.

The observed thermal structure of the Magura Nappe was largely established during accretion, before emplacement of the Magura Nappe rocks in the present day tectonic setting. The outward-directed decrease in thermal alteration, positive correlation between rock age and thermal overprint in each of the slices and in the entire nappe, correlate with the growth of the accretionary wedge. The accretion-related thermal structure was greatly modified during the later thrusting of the Magura Nappe over its foreland. This event involved differential uplift and erosion. The uplift was significantly influenced by the morphology of Carpathian basement. The erosion was greatest above and inward from the regional basement slope distinguished by Ryłko and Tomaś (2001). This erosion resulted in exhumation of the most altered rocks. Two main stages of uplift and erosion can be distinguished. The first stage was related to the uplift of the accretionary wedge, when up to 4 km could have been removed by erosion. The second stage of uplift and erosion was related to thrusting of the Magura Nappe over its foreland and to coeval deformation within the latter.

Vein mineralization vs. structural development

Mineral fillings of fractures record conditions of structural evolution of a given tectonic object. Conversely, knowledge



■ **Fig. 2.** Results of illite/smectite studies, palaeoisotherms on the present day erosion surface, NSB – Nowy Sącz Basin, O-NTB – Orava-Nowy Targ Basin (after Świerczewska in press; geology from Żytko et al. 1989, Ryłko and Tomaś 2003).

of structural development enables for dating of mineralization phases. The results of our studies on the interrelationship between structural development and phases of calcite and calcite-quartz mineralization in the Magura Nappe (e.g. Świerczewska et al. 2000b) allowed us: (1) to date the phases of mineralization in relation to successive tectonic stages in a fragment of the Outer Carpathians and, (2) to reconstruct the PT conditions which took place during these stages.

In our reconstruction, the relationship between structural development and mineralization seems to be well proven for the last stage of the structural development (collapse). On the other hand, we were not quite satisfied of the relationship for the earlier stages of structural development. We believed that the suspected unfit could have resulted from the adopted model of jointing which comprised both shear and extensional joints (e.g. Dunne and Hancock 1994). Therefore, we decided to test our interpretation by adopting exclusively the extensional model of jointing (e.g. Dunne and Hancock 1994). For the object of this study we have chosen an exposure of Paleocene-Lower Eocene strata where the mineralization has been already intensively studied (Świerczewska et al. 2000a and b, 2001).

The exposure at Krościenko town (Fig. 1C, stop 4) is located in the innermost part of the Magura Nappe, the structural development of which is fairly well known (Świerczewska and Tokarski 1998). In that part of the Magura Nappe, mineral veins in Paleocene-Eocene sandstones are largely restricted to early cross-fault joints and faults. Vein textures show that mineralization occurred progressively with the evolving stress regime. The reconstructed process of mineralization fits well into the scheme of the Outer Carpathian structural development. Mineralization was most abundant during gravitational collapse, the last stage of structural development. The extensional model of jointing has been positively verified. For more details see **Stop 4**.

Fluid inclusions

Temperature of fluids penetrating rocks of the Magura Nappe was evaluated basing on fluid inclusion studies. These inclusions are hosted in calcite and calcite-quartz association filling fractures. Only quartz and calcite formed during collapse do contain immiscible aqueous_methane fluid inclusions. Trapping PT conditions for methane-bearing fluids are 160–210 °C and 0.75 to 2 kbar (Świerczewska et al. 2000a). Quartz –calcite veins were only found in the central segment of the Magura Nappe. They are more common in the basement of this nappe, namely in the Dukla Nappe and its equivalents exposed in tectonic windows. Fluid inclusions data collected from the basement show higher temperature (200–220 °C) and pressure (2.1–3.3 kbar) of trapping than those in the Magura Nappe.

Calcite formed in all stages of structural development contains mainly aqueous_inclusions. The latter homogenise between <50 and 145 °C (Świerczewska et. al. 2001). For most of the studied blocky calcites these temperatures are between 80 and 100 °C (A. Kozłowski unpublished), whereas for the columnar calcites they are more differentiated. Quartz overgrowths, which predate calcite mineralization, show homogenisation temperature below 50 °C.

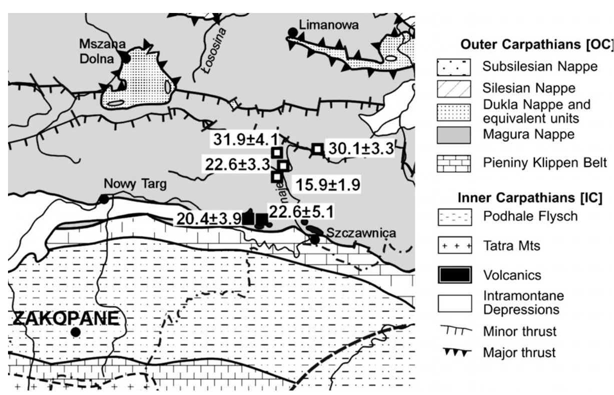
Volcanism

Andesites occur within the Pieniny Klippen Belt and the adjoining portion of the Magura Nappe (Fig. 1B). These rocks form numerous small intrusions (mostly dykes) exposed along the so-called Pieniny Andesite Line. There are two systems of intrusions representing two successive phases of volcanic activity (Birkenmajer and Pécskay 2000 and references therein).

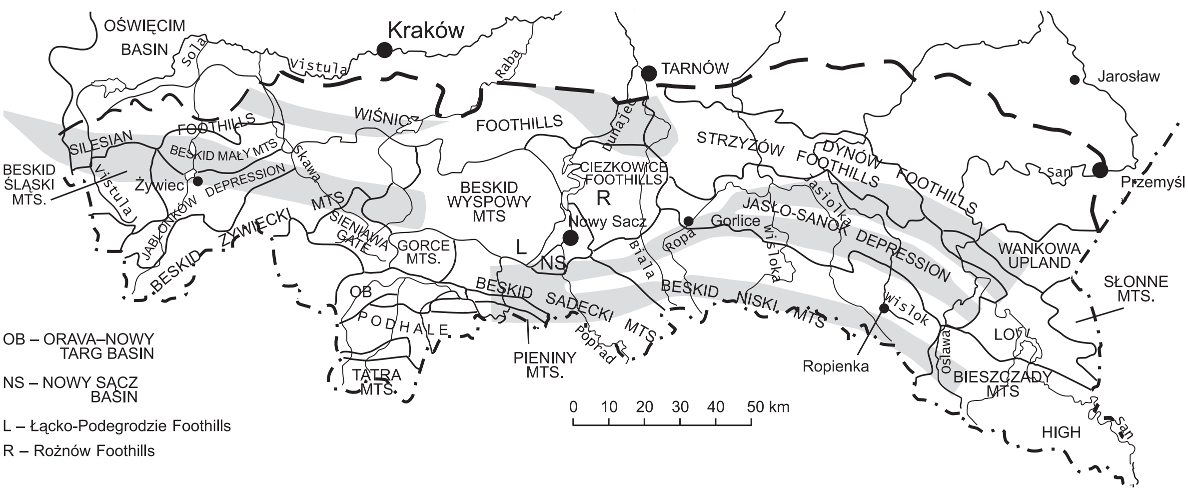
The Pieniny andesites represent basaltic andesites and andesites (according to TAS classification; LOI free basis). The SiO_2 content ranges between 51.5 and 61.5 wt%; LOI value varies from 1.3 to 3.8 wt% (Michalik et al. 2004). Numerous varieties of andesites were described according to phenocrysts assemblage composition (e.g. Małkowski 1958).

According to the results of K/Ar dating, the ages of andesites of both systems range from 13.5 to 11 Ma (Birkenmajer and Pécskay 2000). Both andesites and the host strata are hydrothermally altered (Szeliga and Michalik 2003 and references therein). Andesite samples bearing more altered mafic phenocrysts are rich in secondary chlorite, chlorite/smectite and Ba-enriched K-feldspars in matrix, whereas samples with less altered mafic phenocrysts containin illite/smectite (or vermiculite/smectite) (Michalik et al. 2006 submitted). The age of newly formed biotite in hydrothermally altered andesite is 11.35 Ma (Birkenmajer et al. 2004). This can suggest that either hydrothermal activity occurred directly after magma crystallization, or that dates of Birkenmajer and Pécskay (2000) correspond to the hydrothermal event.

New apatite fission track (AFT) analyses of two andesite samples from Mt. Wżar (Fig. 3) yielded ages of 20.4 ± 3.9 Ma and 22.6 ± 5.1 Ma. (Anczkiewicz unpublished). These ages are similar to the AFT ages obtained for the sandstones from the surrounding flysch of the Magura Nappe, which span an interval of 16–30 Ma. All AFT ages are younger than stratigraphic age of the strata involved and are clearly reset. Hence, the obtained ages both for andesites and for the surrounding flysch strata are interpreted as reflecting cooling of the Magura Nappe. Such interpretation implies that andesitic intrusions took place at least about 20 Ma ago. These results strongly contradict earlier K-Ar results. Clearly, further geo-



■ **Fig. 3.** AFT dating results from the Pieniny andesites (black squares) and Magura Nappe strata (white squares). For location see Fig. 1C (Stop 5).



■ **Fig. 4.** Physiographic units of the Polish segment of the Carpathians (based on Starkel 1991, modified), showing zones of Quaternary uplift marked by abnormally high river bed gradients (modified from Zuchiewicz 1998).

chronological studies, involving other geochronometers are needed in order to establish reliable timing of the Pieniny andesites.

Birkenmajer (2003) provided explanation of the origin of andesite magma. It can be a product of hybridisation of primary magma of mantle origin related to subduction of the European Plate. An alternative explanation (Birkenmajer 2003) suggests that the magma chamber was formed at the base of the accretionary wedge of the Outer Carpathians (possibly at a depth of 10 to 12 km). The continuation of the Pieniny Andesite Line matches the Odra Fault Zone (Lower Silesia), indicating a relationship between the Pieniny andesites and the Lower Silesian alkaline silica-undersaturated mafic rocks of the eastern termination of the Central European Volcanic Province (Birkenmajer and Pécskay 1999). According to Pin et al. (2004), the limited volumes of magmas were produced by partial melting of mafic rocks in the lower crust and/or enriched domains in the mantle, probably as a result of local decompression in a transpressive geodynamic setting. Jurewicz and Nejbert (2005) link the origin of andesitic magma with decompression during shear movements within a deep tectonic zone which can be the SE continuation of the Kraków-Lubliniec fault zone (tectonic contact of the Upper Silesia and Małopolska blocks).

Postscriptum (neotectonics)

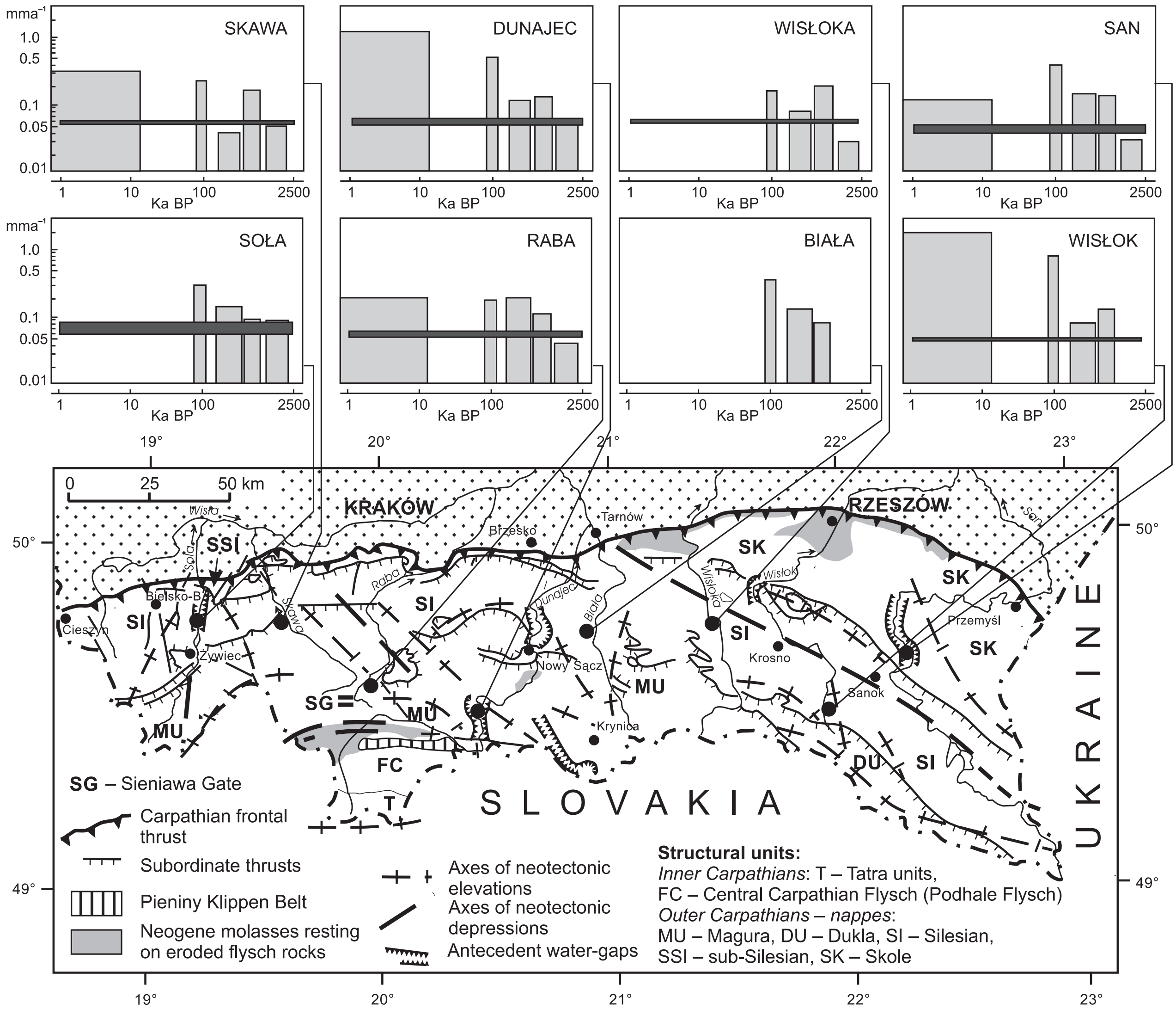
General features

The Outer Carpathian stack of nappes witnessed differential uplift in the Pliocene and Quaternary (Zuchiewicz et al. 2002 and references therein) (Figs 4, 5). The uplift in Pliocene times was probably of block-type in the western segment of the belt, whereas that in the medial and eastern segments was restricted to subparallel and relatively narrow upwarped zones. The total size of uplift, approximated by the amount of erosional dissection of the Plio-Pleistocene planation surfaces and straths of Quaternary terraces, varied from 150 m to ca. 900 m, averaging at around 300 m. The uplift could have been a result of both post-orogenic isostatic rebound related to erosional unroofing, particularly intensive in the western part of the area, and of the steepening of frontal parts of some nappes.

Episodes of intensified Quaternary uplift, restricted to relatively narrow zones oriented subparallel to the structural grain of the area and coinciding with frontal parts of overthrust nappes and larger slices, occurred during the Cromerian-Elsterian 1/2, Eemian-early Weichselian and Weichselian Late Glacial-Holocene times, at rates varying from 0.15 to 2.0 mm/yr (Fig. 5). These long (100–250 km) and narrow (15–25 km) zones of localized uplift appear to be a result of the Quaternary relaxation of horizontal stresses (cf. Zuchiewicz 1998 and references therein).

The available pieces of structural evidence imply that during the Late Neogene times structural development of the Polish segment of the Outer Carpathians was controlled by normal faulting. This interpretation is corroborated by geomorphic data indicative of *en block* uplift in the western part of the belt. However, there is no unequivocal evidence to decide whether the faulting was due to successive phases of alternating N-S and E-W extension or owing to one or more phases of multidirectional extension. Moreover, the geomorphic data from the medial and eastern parts of the belt suggest the occurrence of compressional stress regime during Pliocene times (cf. Zuchiewicz et al. 2002).

The data available for Quaternary times show an apparent contradiction. On one hand, different pieces of geomorphic evidence imply compressional stress arrangement, with the maximum compressive stress being oriented roughly perpendicular to the belt. This interpretation is compatible with the present-day orientation of the S_{Hmax} inferred from the breakout analysis and from focal solutions of the Krynica earthquakes (Fig. 6) (cf. Jarosiński 1998, Zuchiewicz et al. 2002). The zones showing tendencies to Recent uplift tend to be aligned subparallel to frontal thrusts of individual nappes and larger slices, suggesting the presence of Plio-Quaternary horizontal stresses in the flysch nappes. En echelon arrangement of theses zones, however, slightly different in the western and eastern parts of the study area, appears to indicate young sinistral motions along the Kraków-Lubliniec fault zone (Fig. 6) in the substratum of the overthrust nappes (Zuchiewicz 2001).



■ **Fig. 5.** Neotectonic sketch map of the Polish Carpathians (based on Zuchiewicz 1998, Zuchiewicz et al. 2002). Diagrams illustrate rates of Quaternary river dissection in those segments of river valleys which are located in neotectonically uplifted structures. Solid lines denote average rates of Quaternary dissection.

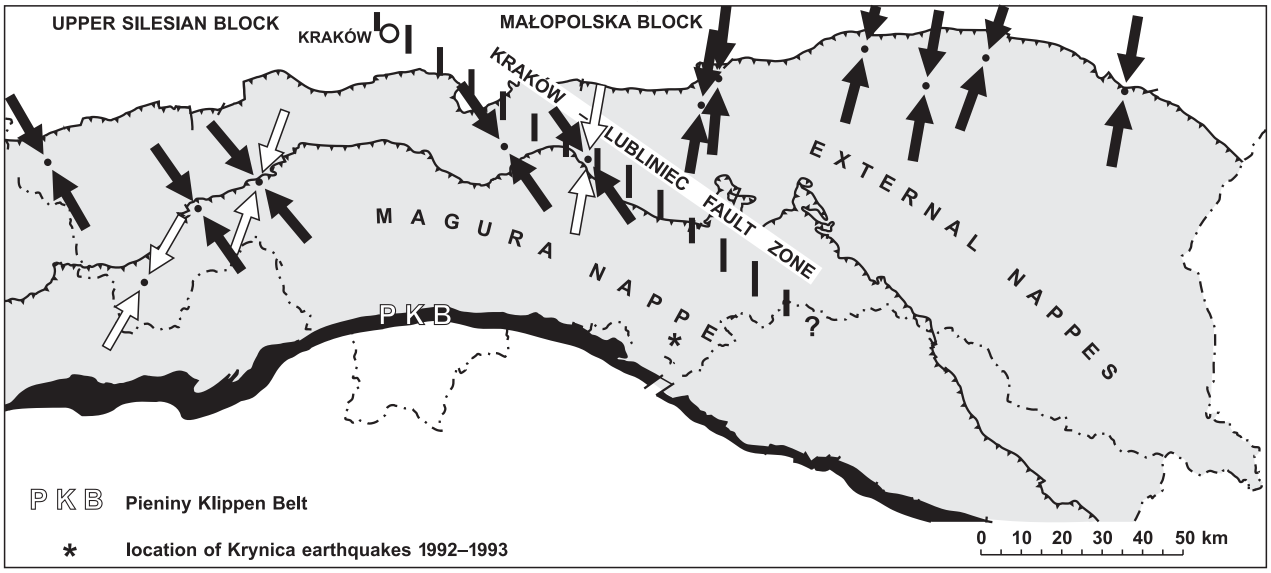
On the other hand, Quaternary normal faulting within the intramontane basins and in localized narrow zones of frontal parts of nappes and larger slices points to extensional stress arrangement. This contradiction can be explained by a concept of normal faulting restricted to certain zones affected by horizontal stresses. These processes were probably not uniform, as shown by differentiated rates of erosional dissection of Quaternary straths in individual geomorphic units within different Quaternary stages (Figs. 5, 7). Another, although not contradictory explanation, lies in the general isostatic post-orogenic uplift, being overprinted by coeval horizontal motions within the flysch cover.

The rates of recent vertical crustal motions in the Polish Outer Carpathians range between 0 mm/yr in the western and medial segment to ca. +1 mm/yr in the east (Wyrzykowski 1985), whereas those in the Pieniny Klippen Belt do not exceed 0.5 mm/yr (Ząbek et al. 1993, Czarnecki 2004). The results of recent GPS campaigns (Hefty 1998) point as well as borehole breakout analyses (Jarosiński 1998), in turn, to NNE-directed horizontal motions throughout the area (cf. Fig. 6).

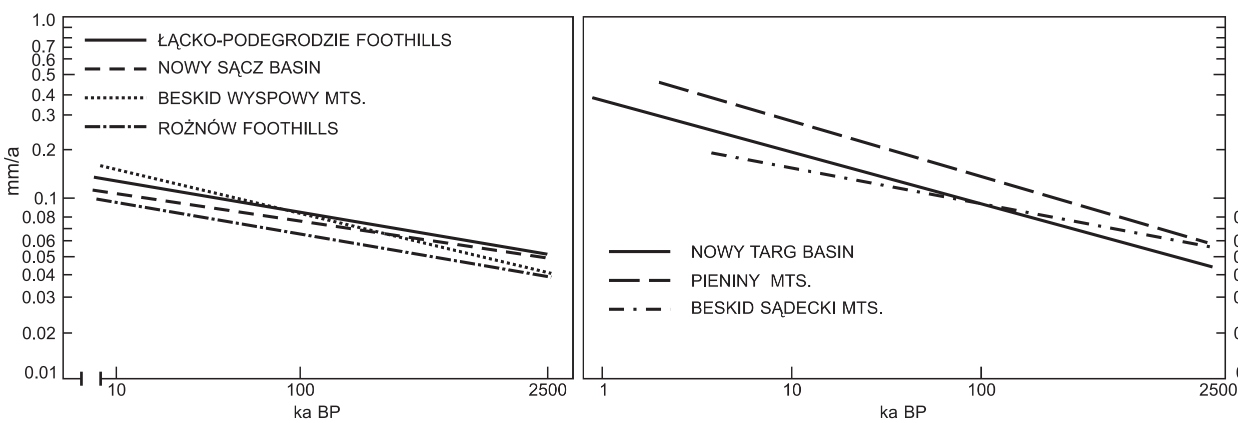
Magura Nappe

The apparent amount of Late Neogene-Quaternary uplift of the medial segment of the Magura Nappe has ranged from 150–430 m in the south and 360–900 m in the medial sector, to 180–310 m in the north. The size of purely Quaternary uplift of the southern part of the Magura Nappe has been estimated for some 150 m (cf. Zuchiewicz 1995).

A comparison between the pattern of elevated and subsided structures of the Magura floor thrust and the results of morphotectonic studies shows that in the western part of the Polish Outer Carpathians (Figs. 8, 9) the highest-elevated neotectonic structures (in the southern portion of that area) coincide with depressions of



■ **Fig. 6.** Mean SHmax directions obtained from breakouts in the autochthonous basement (black) and the Carpathian nappes (white) (based on Jarosiński's data in Zuchiewicz et al. 2002 simplified).



■ **Fig. 7.** Rates of erosional dissection of straths in different physiographic units of the medial portion of the Polish Carpathians (based on Zuchiewicz 1998).

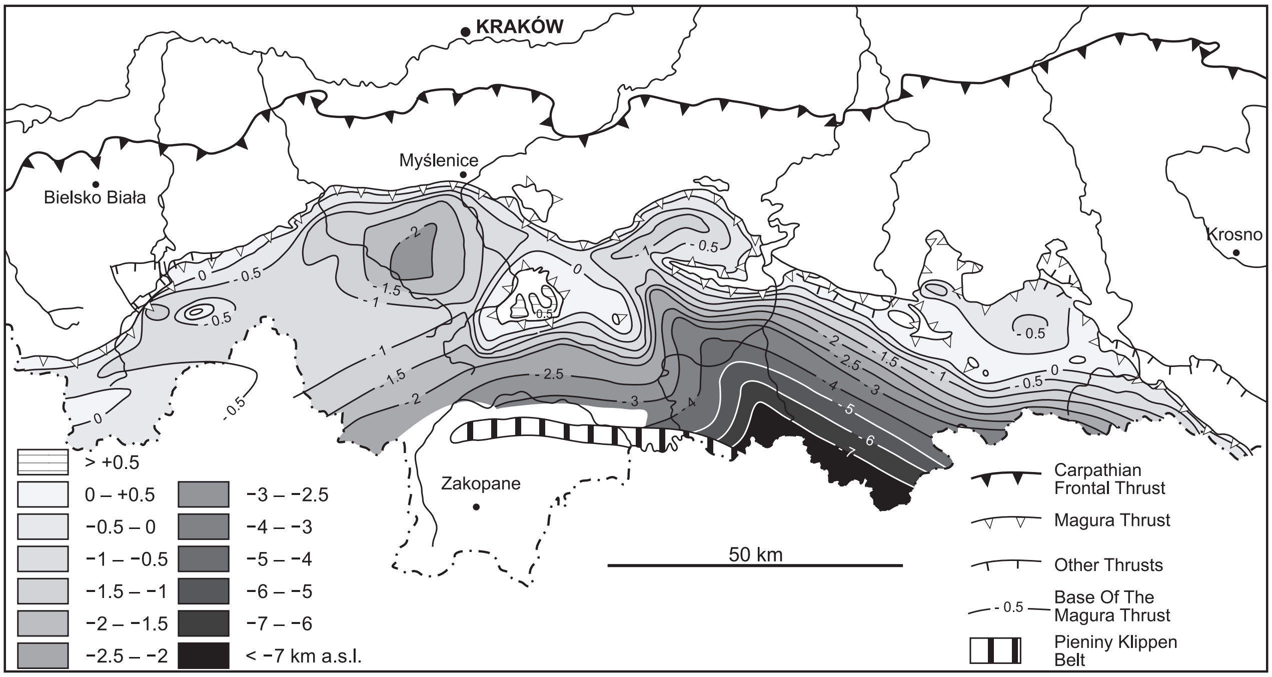
the Magura floor thrust, whereas farther north a reverse pattern becomes dominant: neotectonic elevations coincide either with exposures of the Magura frontal thrust or with elevations of its surface (Fig. 8). This is particularly true for an area comprised between 20° and 20° 30′E meridians. Moreover, the strongly uplifted region in this part of the Outer Carpathians is situated shortly south of the main elevation of the Magura floor thrust, represented by the Mszana Dolna tectonic window (Fig. 9).

The origin of such relationships is difficult to explain. We infer that one of possible factors could be Pliocene-Quaternary reactivation of faults cutting the Magura floor thrust, as well as the basal thrust of the Outer Carpathians, and particularly that which appears to separate the western-medial segment of the Outer Carpathians from the more eastern portion. A similar conclusion has been proposed by Zuchiewicz et al. (2002) when analysing morphometric parameters of uplifted neotectonic structures in the western and eastern portions of the Outer West Carpathians, whose en echelon arrangement is different on either side of the reactivated, probably sinistral, Kraków-Lubliniec deep fault zone (Fig. 6), located beneath the overthrust nappes. For more details see **Stop 3**.

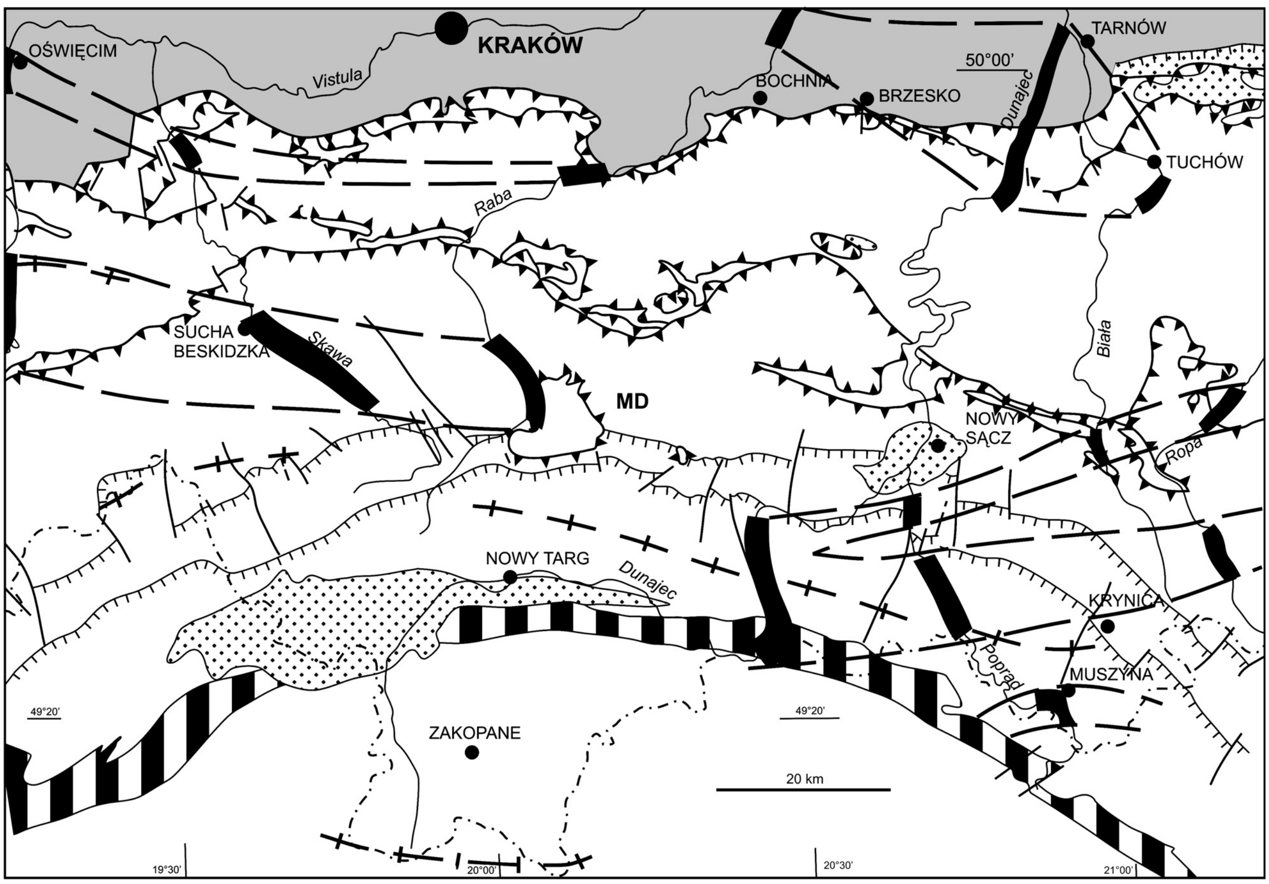
Fractured clasts

Analysis of fractured clasts in gravels and conglomerates has commonly been applied to palaeostress studies and dating of faulting during the past few tens of years (for review see: Tokarski and Świerczewska 2005). In the Polish segment of the Outer Carpathians, fractured clasts are fairly numerous in young Cenozoic (Miocene through Holocene) gravels and paraconglomerates which crop out close to regional overthrusts and faults (Fig. 1B), pointing to recent activity of these features. So far, the kinematics of this activity is poorly understood (e.g. Tokarski et al. submitted).

Fracture architecture in fractured clasts is either well-organized or chaotic. It is likely that only well-organized fracture architecture is suitable for kinematic analysis. The aim of our studies



■ **Fig. 8.** Topography of the Magura floor thrust (based on unpublished Oszczypko's data).



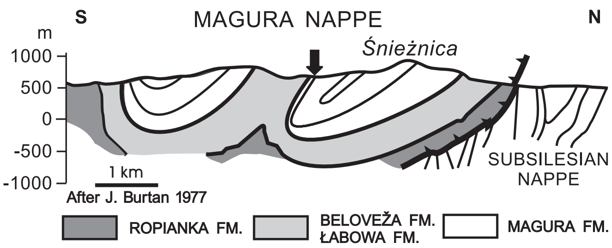
■ **Fig. 9.** Neotectonic sketch of the medial portion of the Polish Carpathians. Black vertical segments – Pieniny Klippen Belt, dots – Late Neogene molasses resting on eroded flysch units, shaded elongated segments – segments of river valleys showing abnormally high river-bed gradients, thick fence lines – neotectonically elevated areas.

has been to describe the influence of textural properties and lithology of gravels and paraconglomerates on fracture architecture, and to make an attempt at kinematic analysis of fractures. Up to now, we managed to document that: (a) the number of fractured clasts in a given clast population is positively correlated with the clast diameter and negatively correlated with the size of grains within clasts of detrital rocks; (b) the fractures include both those inherited after joints cutting host strata in the source areas, and those formed *in situ* in the studied rock series; (c) the inherited fractures, showing chaotic architecture, are mainly oriented at right angle or nearly perpendicular (80–90 °) to clast a-b surfaces; (d) neofractures formed *in situ*, showing well-organized architecture, are oriented both perpendicular (80–90 °) and at smaller angles (<80 °) versus a-b clast surfaces. For more details see **Stop 2**.

Stop 1. Deformation Bands in Eocene Sandstone, Gruszowiec

Description

Small, abandoned quarry in the Gruszowiec Village (Fig. 1C), located 300 m from the "Pod Cyckiem" (Under the Tit) bar. Thin- to thick-bedded sandstones intercalated with mudstone and claystone are exposed in the quarry. The strata belong to the Late Eocene Poprad Sandstone Member of the Magura Formation within the Rača slice (Fig. 1B). The exposure is situated in the overturned limb of the Śnieżnica syncline (Fig. 10). Four groups



■ **Fig. 10.** Cross-section (after Burtan 1977) of the Śnieżnica syncline showing tectonic position of the discussed exposure. For location see Fig. 1C (stop 1).

of planar minor structures have been observed (Fig. 11A): (1) water-escape sheets, (2) deformation bands (DB), (3) joints, and (4) minor faults. Water-escape sheets (1) and DB (2) were observed only within sandstones, whereas joints (3) and faults (4) also cut mudstones and claystones.

(1) Water-escape sheets extend upwards from the bases of sand-stone beds. Some of the sheets root in wedge-shaped injections penetrating from the underlying mudstone. The water-escape sheets are oriented subvertical and cross-fold.

(2) Eighty-four DB have been studied in thin sections. The majority of DB are oriented fold-parallel, a minor number of bands shows of cross-fold orientation, and very few DB are oriented obliquely to the fold (Fig. 11B). Following the classification by Antonellini et al. (1994), three types of DB have been distinguished: (1) deformation bands with no cataclasis (NDB), (2) deformation bands with traces of feldspar cataclasis (TDB) and, (3) deformation bands with strong feldspar cataclasis (SDB). All types of DB contain traces of calcite cement, but we have not observed any calcite cataclasis. In the host rocks distribution of calcite cement is not uniform in relation to DB. Usually, DB separate areas showing different degree of cementation.

Inside the NDB and the TDB, mica flakes and elongated grains are oriented parallel to the band boundaries. The grain size, compared to the host rock, is only slightly reduced in the NDB, whereas it is reduced by factors between 5 and 10 in the SDB. The boundaries with the host rock are transitional for the NDB and clear cut for the SDB. The latter contain more matrix than the host strata whereas only local slight enrichment in matrix occurs in the former. The microstructures of the TDB are transitional between the NDB and SDB. The cross-fold DB are exclusively NDB. Occasionally, these bands cut fold-parallel NDB. Some of the cross-fold DB pass laterally into water-escape sheets. Some of the NDB pass laterally into the SDB. Numerous SDB cut NDB.

In all cases where the sense of shear can be determined, the observed offset along the fold-parallel DB is reverse. The fold-parallel DB display different dip angles ranging from subhorizontal to subvertical (Fig. 11C). Steep dips prevail for the NDB whereas the majority of TDB and SDB display shallow dip. Numerous DB form conjugate sets which intersect one another at 30–60 ° (Fig. 11A). The orientations of the bisecting planes of conjugate bands change from bedding-parallel to bedding-perpendicular.

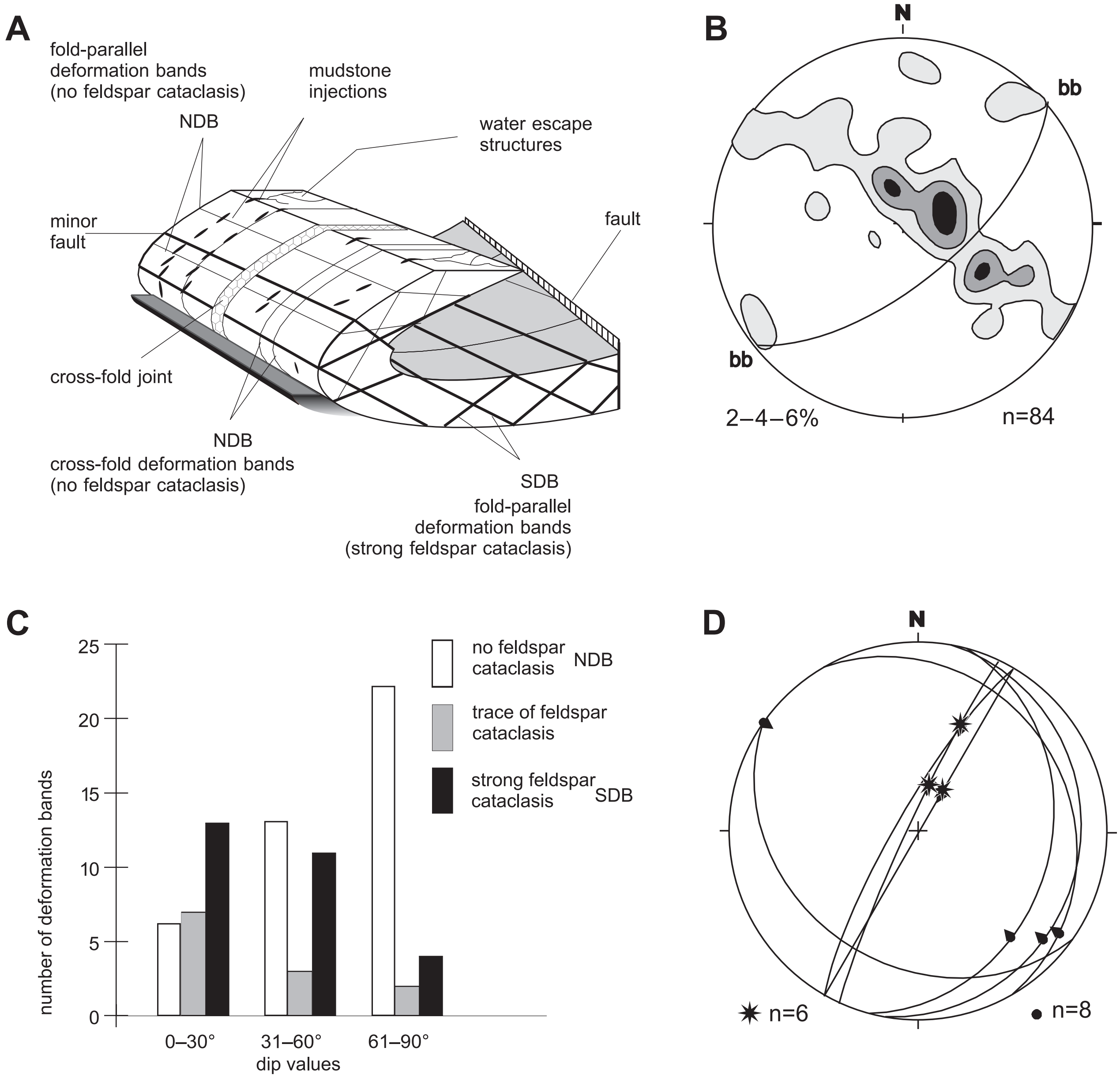
(3) Joints are sub-vertical and cross-fold (Fig. 11A). Some of the joints are filled by calcite veins up to 10 cm thick. In the host strata, close to the calcite-filled joints, the content of calcite cement diminishes abruptly away from joints.

(4) Minor faults are lined by breccia zones, up to 0.5 mm thick. The boundaries of these zones are clear-cut. In distinction to the DB, the breccia zones lining minor faults are devoid of mica flakes and display quartz and calcite cataclasis. The minor faults studied comprise shallow-dipping reverse faults, and subvertical, strike-subparallel, oblique faults (Fig. 11D).

Discussion

The orientation of water-escape sheets, joints and DB are either fold-parallel or cross-fold. This indicates that formation of these structures was related to regional folding.

Water-escape sheets are the features which form penecontemporaneously with sedimentation. This indicates that regional folding was already initiated during sedimentation of the strata involved (Fig. 11A). Some of the water-escape sheets pass laterally into cross-fold DB, whereas some of the latter cut some of



■ **Fig. 11.** Minor planar structures at the Gruszowiec exposure. A – model showing attitudes of structures in different structural positions; B – stereoplot of deformation bands; C – histogram of DB type versus DB dip; D – stereoplot of minor faults.

the fold-parallel DB. These relationships indicate that both cross-fold and fold-parallel DB also started to form during sedimentation. This conclusion is corroborated by the nature of the cross--fold DB which are exclusively NDB.

The fold-parallel DB present the whole spectrum of orientations, from bedding-parallel DB to bedding-perpendicular ones. They display reverse offset and at least some of them can be grouped into conjugate sets. We believe that all fold-parallel DB were formed as conjugate shear failures. In our interpretation (Fig. 12), these shear failures were occurring progressively during folding, becoming more and more steeply inclined to the bedding. Several fold-parallel NDB display shallow dips (Fig. 11C), indicating that they formed when the strata attained subvertical orientation close to the completing of folding. It follows that folding was completed when the host strata were still poorly indurated.

The number of fold-parallel TDB and SDB is negatively related to their dip angles. It follows that the majority of these bands formed when the strata were oriented subvertical, at least some of them as conjugate shear failures. The NDB are commonly cut by SDB, indicating that fold-parallel SDB which are steeply inclined with respect to bedding formed after completing of folding, when the host strata became more indurated than during the formation of the NDB.

We believe that those SDB which display steep dips also formed after completing of folding, resulting from reworking of the NDB. The observation that some NDB pass laterally into SDB corroborates this opinion. Not a single DB displays calcite cataclasis. This indicates that the calcite mineralization was introduced into host strata and into DB only after formation of SDB and, therefore, after completing of folding. Distribution of calcite cementation vs. DB indicates that DB, probably due to their low permeability, formed barriers to calcite-bearing fluids. It appears that this mineralization was introduced along the cross-fold joints (Fig. 11A). The studied minor faults display calcite cataclasis. It follows that the faulting occurred after introducing of calcite mineralization into the host rock and therefore after completing of folding.

Conclusions

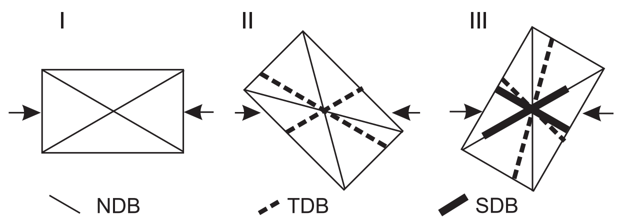
The above data provide good evidence for progressive pre-, syn- and postlithification of small-scale tectonic deformations. Microstructural studies indicate that deformation started in loose sediment as hydroplastic features and lasted until the host strata became fully indurated.

(1) The formation of DB was controlled by regional stress field in a compressional regime. Distinct types of DB formed progressively during and after regional foldind, during increasing induration of the host strata. Deformation bands with no cataclasis (NDB) started to form when the host strata were still in horizontal position and continued to evolve until the completion of regional folding. Deformation bands with traces of feldspar cataclasis (TDB) and bands with strong feldspar cataclasis (SDB) formed during and after regional folding.

(2) The folding occurred when the host strata were poorly indurated and possibly under low confining pressure. The folding took place during deposition of the host strata.

(3) At the completion of regional folding, the host strata were still poorly indurated. Introduction of calcite mineralization, which occurred along the joints, post-dated folding.

(4) Within the studied portion of the Magura Nappe, regional folding started not later than during the Eocene. Folding was completed when the host strata still were not fully indurated, most probably also during the Eocene.



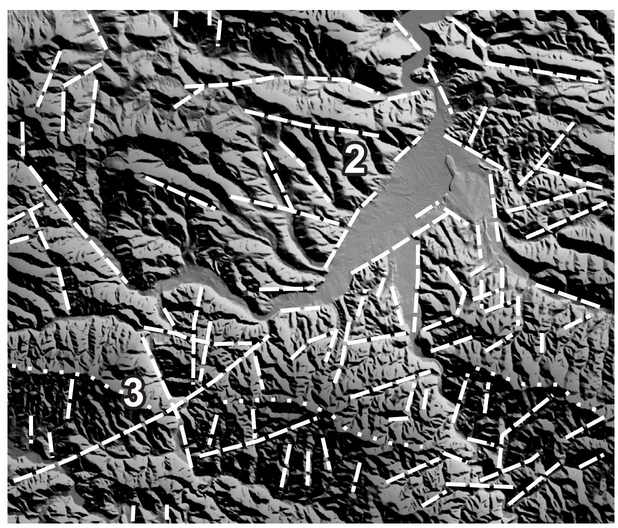
■ **Fig. 12.** Cartoon showing the succession of appearance of particular DB types, I – before folding, II – during folding, III – after folding.

Stop 2. Fracured Clasts in Miocene and Pleistocene Paraconglomerates, Brzezna

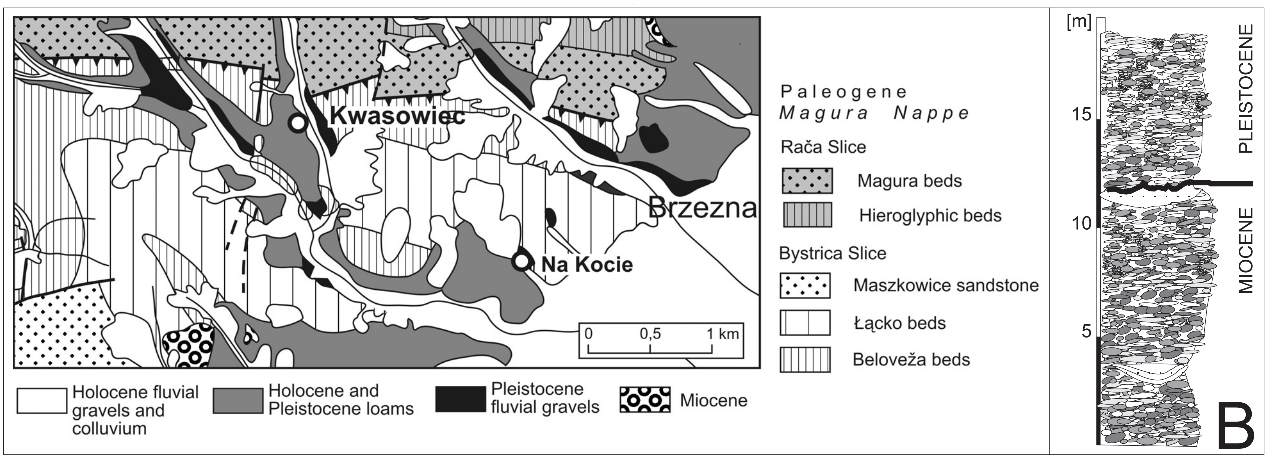
Natural exposure at the "Na Kocie" (Over the Cat) site is located in Brzezna village (Figs. 13, 14A). The choice of study object results from its interesting tectonic setting. This site is situated close to a regional thrust fault on the margin of the Nowy Sącz Basin, which is also tectonically controlled. In addition, the analysed strata bear a wealth of lithologically differentiated fractured clasts, and are disturbed by both small-scale (individual clasts) and larger-scale tectonic deformations. Finally, site "Na Kocie" is placed not far from an exposure of Pleistocene gravels at Kwasowiec, recently described by two of us (Tokarski and Świerczewska 2005). The latter gravels, showing poor roundness measures and monotonous lithology and bearing numerous fractured pebbles, provide good comparative material.

The studied exposure is situated in the hanging wall of the Bystrica slice thrust over the Rača slice, ca. 1,200 m away from the thrust surface. The present-day erosional margin of the Nowy Sącz Basin, coinciding with the extent of the Neogene infill, is placed 300 m from "Na Kocie" exposure.

The analysed rock series is exposed in a 19-m-high, undermined slope of a stream valley. The lower part, 11 m high (Fig. 14B), is composed of paraconglomerates bearing sandstone intercalations. This complex is of Late Badenian and/or Early Sarmatian age (Oszczypko et al. 1991 and references therein). The Miocene complex is unconformably overlain by 8-m-thick Pleistocene paraconglomerates dated to the Elsterian-2 (Butrym et al. 1989, Osz-



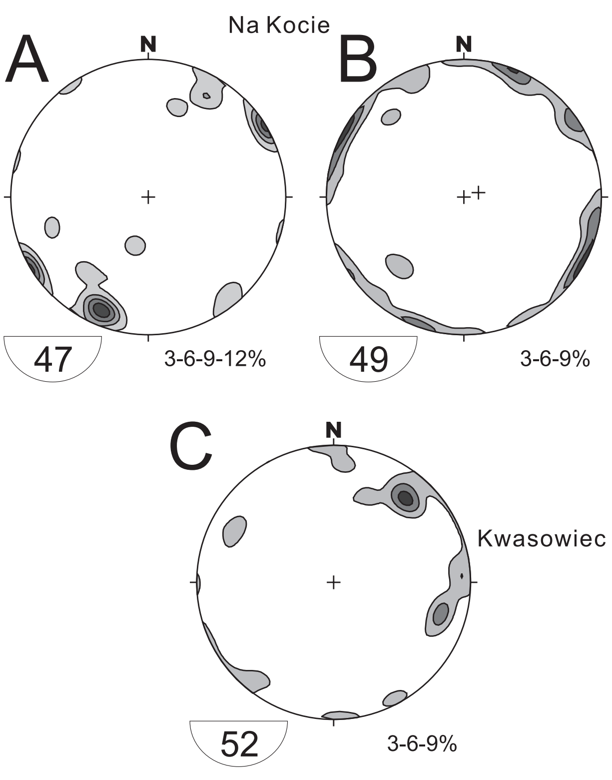
■ **Fig. 13.** Digital elevation model of the medial segment of the Polish Outer Carpathians, showing location of Brzezna (Na Kocie; NK) and Tylmanowa (TY) stops. Dashed lines refer to well-marked topolineaments, dotted line marks the axis of the most important neotectonic elevation of this region. Note a well-pronounced fault on the NW margin of the Nowy Sącz Basin and intersecting topolineaments close to Tylmanowa. For location see 1C (stops 2 and 3).



■ **Fig. 14.** A – study area showing the location of "Na Kocie" and "Kwasowiec" exposures (geology based on Oszczypko and Wójcik 1989); B – section of "Na Kocie" exposure; for location see Fig. 13.

czypko et al. 1992). Both complexes are separated by an erosional surface.

The clasts in the Miocene paracoglomerate are composed exclusively of rocks occurring in the Magura Nappe.The clasts are orderly arranged. Clast "a" axes plunge gently (<30°) to the NNW and SSE, less frequently to the NNE and SSW. The lower portion of the Miocene complex bears a lens-like intercalation of sandstones alternating with siltstones and claystones.

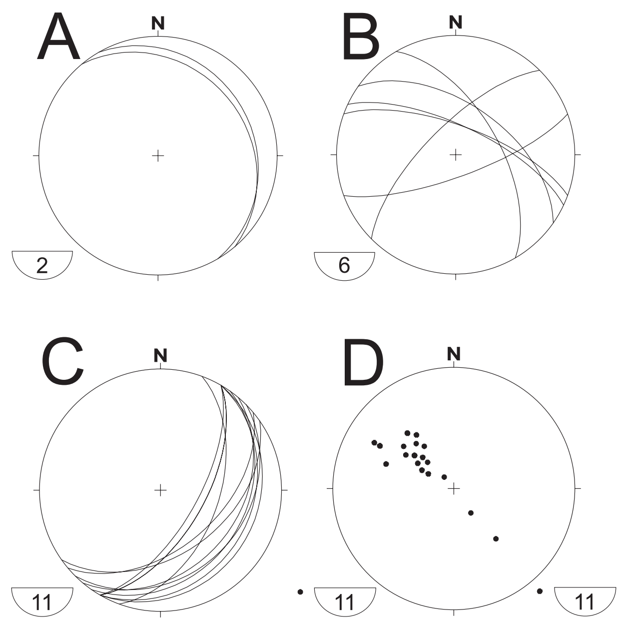


■ **Fig. 15.** Orientation of clast-cutting fractures: A – Miocene conglomerate at exposure "Na Kocie"; B – Pleistocene conglomerate at exposure "Na Kocie"; C – Pleistocene gravels at Kwasowiec.

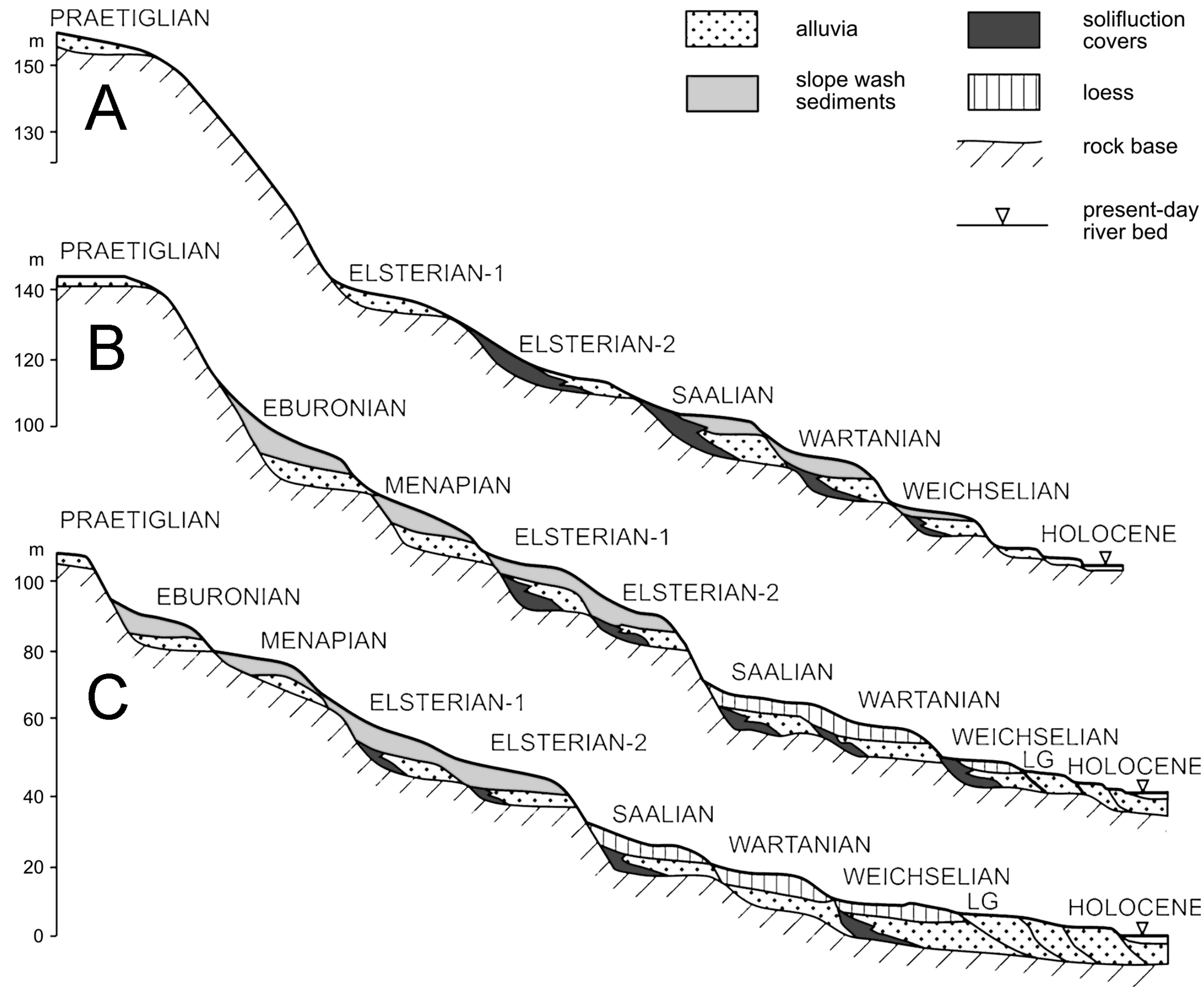
The clasts in the Pleistocene paraconglomerate are derived both from the Magura Nappe and from the Inner Carpathians (cf. Butrym et al. 1989, Oszczypko et al. 1992). The clasts are arranged orderly. The majority of clast "a" axes plunge gently (<30°) towards the west and south.

Fractured clasts

Paraconglomerates of both complexes bear numerous clasts cut by one fracture and some clasts cut by several fractures. These fractures are restricted to single clasts; the matrix is not fractured.



■ **Fig. 16.** A, B – normal faults cutting clasts within conglomerates of Miocene (A) and Pleistocene (B) age; C,D – structures comprised within sandstone intercalation in the lower part of the Miocene complex: A – reverse faults; B – poles to reverse faults (asterisks) and bedding surfaces (dots).



■ **Fig. 17.** Flights of Pleistocene and Holocene straths and complex-response terraces in three physiographic units dissected by the Dunajec River : A – Beskid Sądecki Mts. (site Tylmanowa), B – southern portion of the Beskid Wyspowy Mts. (Łącko – Podegrodzie Foothills), C – Nowy Sącz Basin (see Fig. 4 for location). LG – Weichselian Late Glacial.

In the Miocene conglomerate, a population of 100 large clasts (>2 cm) comprises 50 % of fractured clasts, and 50 % of unfractured clasts, while in the Pleistocene conglomerate, the respective share of these groups is 45 %, and 55 %. In the Miocene conglomerate, a population of 50 small clasts (<2 cm) contains 22 % of fractured clasts, and 78 % of unfractured clasts, and in the the Pleistocene conglomerate, these groups amount to 26 % and 74 %, respectively. Different lithologies represent different numbers of fractured clasts; clasts of detrital rocks tend to have more fractures compared to quartzites and magmatic rocks.

Clast architecture is well-organized and similar in both conglomerates (Fig. 15A, B). Most of the fractures are arranged subvertically, and tend to form two sets oriented NW and NE. The proportion of fractures situated at right angle and nearly perpendicular (80–90 °) to clast a-b surfaces amounts to 37 % and 50 % in the Miocene and Pleistocene conglomerates, respectively. In both conglomerates, infrequent clasts are cut by normal faults (Fig. 16A, B) which do not pass into the matrix.

Structures within sandstones

A sandstone intercalation exposed in the lower part of the Miocene complex is cut by numerous reverse faults of (Fig. 16C). Poles to both fault planes and bedding surfaces plot on a single great circle, whose axis is horizontal and oriented NE-SW (Fig. 16D). Joints cutting this sandstone body are oriented both perpendicular (NW) and parallel (NE) to the reconstructed axis of the great circle. Identically oriented fractures (NW and NE) cut a pebble situated in the lower portion of this body.

Interpretation

The architecture of clast-cutting fractures is well organized and comparable in both conglomerates (Fig. 15A, B), implying that fractures in the two complexes were formed *in situ* and are of the same origin.

We infer that the sandstone intercalation in the lower part of the Miocene complex originated during deposition of the latter as a channel fill which was later folded. This is testified to by the position of poles to bedding on a single great circle (Fig. 16D), the axis of which is horizontal and oriented NE-SW. The same great circle bears poles to reverse faults which cut the rock body in question. It means that both folding and reverse faulting took place in a compressional stress field of σ1 oriented NW-SE.

Most of clast-cutting fractures in the two conglomerates are subvertical and clustered into two sets oriented NW and NE. We conclude that these fractures represent joints formed in a compressional stress field of σ1 oriented NW. The NW-striking fractures are extensional features parallel to σ1/σ2 plane, whereas NE-striking fractures represent joint surfaces formed during stress relaxation (cf. Caputo 1995). In addition, fractures cutting clasts in the two conglomerates are parallel to joints within sandstone intercalation in the lower part of the Miocene complex. This implies that all these structures, except for normal faults, originated in the same stress field.

Discussion and conclusions

(1) In both complexes, the number of fractured clasts is positively correlated with the clast size, and negatively correlated with the diameter of grains in clasts of detrital rocks. The number of fractured clasts increases in clasts of detrital rocks, compared to those of quartzites and magmatic rocks. These data point to the role of clast size and lithology in controlling the number of fractured clasts, as already shown by Tokarski and Świerczewska (2005).

(2) Clast-cutting fractures are well organized (Fig. 15A, B), irrespectively of the orientation of clast "a" axes which are different for each complex. This means that such fractures were formed *in situ* and that clast orientation does not influence the bearing of fractures.

(3) The two conglomerates studied at "Na Kocie" exposure bear clasts cut by fractures oriented both at right angle or nearly perpendicular (80–90 °) and at smaller angles (< 80 °) to the clast a-b surfaces. This observation supports our previous conclusion (Tokarski and Świerczewska 2005) that such an orientation of fractures can be diagnostic for fractures formed *in situ* in the studied gravels and conglomerates.

(4) Fractures cutting clasts within the conglomerates represent extensional joint fractures that were formed in a compressional stress field, of σ1 oriented NW-SE. The sandstone intercalation in the lower part of the Miocene complex was folded and reverse-faulted in the same stress field. It is probable that the origin of these deformations was related to neotectonic activity of the Bystrica thrust. Similar architecture of joints cutting clasts in the Pleistocene fluvial gravels at Kwasowiec (Fig. 15C) appears to suggest that this activity must have been of a wider extent. In contrast, the origin of normal faults cutting clasts of both complexes was probably different and associated with neotectonic activity of the margin of the Nowy Sącz Basin.

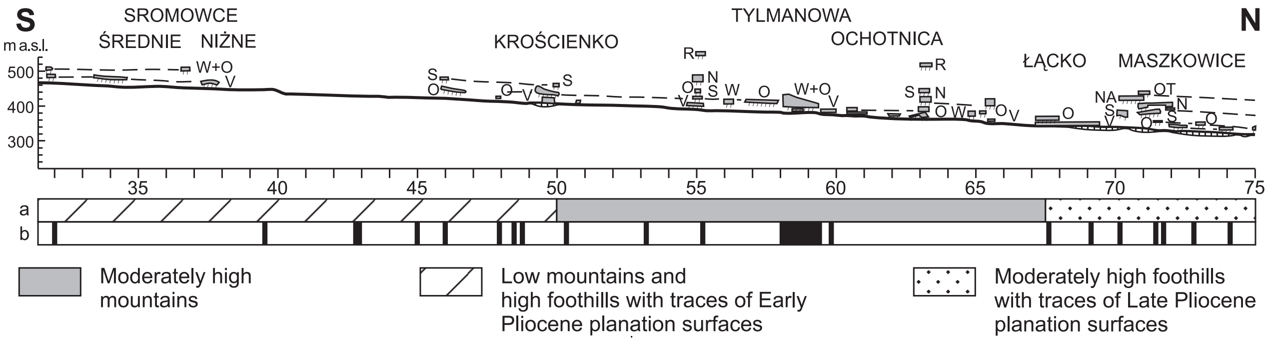
Stop 3. Uplifted Terraces in the Antecedent Dunajec River Gorge, Tylmanowa

Site Tylmanowa is situated in an antecedent gorge of the Dunajec River, dissecting the Beskid Sądecki Mts. Another gorge of this type cuts the Pieniny Mts., a few kilometres farther south. The "Beskid Sądecki" gorge segment includes two deeply cut meanders which are separated by a rectilinear valley, parallel to a fault line. This area is situated at a place of intersection of NNW, NE, and N-S striking topolineaments (Fig. 13).

The gorge is 15 km long and up to 700 m deep, its width changing from 75–100 m within the meanders to 450–500 m in other segments. The river-bed is cut into solid bedrock and its long profile is ungraded and of exceptionally high gradient, compared to the upstream and downstream valley reaches (cf. Fig. 9). The eastern valley sides are steep (50–66 %) and dissected by a network of short (up to 1.5 km) and high-gradient (200 %) minor tributary valleys and ravines. Outlets of tributary valleys are usually hanging over the present-day river bed, up to 10–15 m. Headwater parts of some of these tributaries represent hour-glass valleys. The surrounding ridges bear traces of four planated surfaces that rise 900 m, 770–830 m, 500–590 m, and 450–500 m a.s.l.

The eastern valley sides are mantled by weathering debris and loams, while the western ones are dominated by a flight of straths and complex-response terraces (Fig. 17), whose alluvial covers were deposited during the Pleistocene glacial stages: Praetiglian (150–155 m to 154–161 m), Elsterian-1 (75–84 m to 78–96 m), Elsterian-2 (51–55 m to 52–65 m), Saalian (26–41 m to 29–41 m), Wartanian (17–24 m to 20–31 m), and Weichselian (10–11 m to 16–18 m), as well as in the Holocene (6–10 m, 4–5 m, 2–3 m). The thickness of terrace alluvium is between 3–4 m and 10–14 m (cf. Zuchiewicz 1995 and references therein). These covers are composed of poorly rounded and poorly sorted, both Outer Carpathian flysch (sandstones, siltstones, rare conglomerates) and Tatra-derived (granites, quartzites) gravels and cobbles. Limestones shed from the Pieniny Klippen Belt can only be found within the youngest, i.e. Weichselian and Holocene alluvium; limestone clasts of older fluvial series became completely dissolved. The Early and Middle Pleistocene covers include a large proportion of angular clasts, pointing to the role of intensive solifluction within glacial stages. All Pleistocene fluvial covers interfinger with solifluction tongues, those dated to the last and penultimate glacial stages being also overlain by slopewash and/or solifluction-slopewash covers, which are 3–8 m thick. Such interfingering enables for relative dating of the preserved terrace covers to individual glacial ages.

Long-profiles of Pleistocene straths (Fig. 18) clearly show increased relative heights of the latter within meander loops. These



■ **Fig. 18.** Longitudinal profile of Pleistocene terraces of the Dunajec River valley between the Nowy Targ Basin and the southern portion of the Beskid Wyspowy Mts., showing location of site Tylmanowa. Symbols: a – landscape types, b – fault zones within bedrock. Grey patches on the profile refer to alluvial covers, barbed lines mark straths of individual terraces. Pleistocene glacial stages: R – Różce (Praetiglian), OT – Otwock (Eburonian?), NA – Narew (Menapian?), N – Nida (Elsterian-1), S – San (Elsterian-2), O – Odra (Saalian), W – Warta (Wartanian), V – Wisła (Vistulian, Weichselian). Numbers below the profile refer to distance (in kilometers) from Nowy Targ.

heights are greatest within the entire Polish segment of the Outer Carpathians; straths of equivalent age within the remaining Dunajec River valley reaches, and those of other Carpathian river valleys are lower, even by 30 m in case of the oldest Pleistocene straths (cf. Zuchiewicz 2001 and references therein). Deformations of Pleistocene straths combined with intense erosional downcutting appear to indicate Pleistocene uplift of the axial part of the Beskid Sądecki Mts., part of the most strongly elevated neotectonic structure in the Polish Outer Carpathians (cf. Starkel 1972, Zuchiewicz 1995, cf. also Figs. 4, 9, 13). The amount of fluvial accumulation throughout the Middle and Late Pleistocene in the Dunajec River gorge can be estimated at 555×10^6 cubic metres, that removed by erosion attaining ca. 652×10^6 cubic meters. The lack of fluvial covers from the Eburonian and Menapian stages points to intense, tectonically-controlled, erosion before the Elsterian; the equivalent-age terrace covers are to be found immediately north of the gorge (Fig. 18).

Stop 4. Joints and Mineral Veins in Paleocene Sandstone, Krościenko

Description

Natural exposure of the Paleocene-Lower Eocene Szczawnica Formation on the Dunajec River bank at Krościenko town (Fig. 1C) is the site where the model of jointing has been tested.

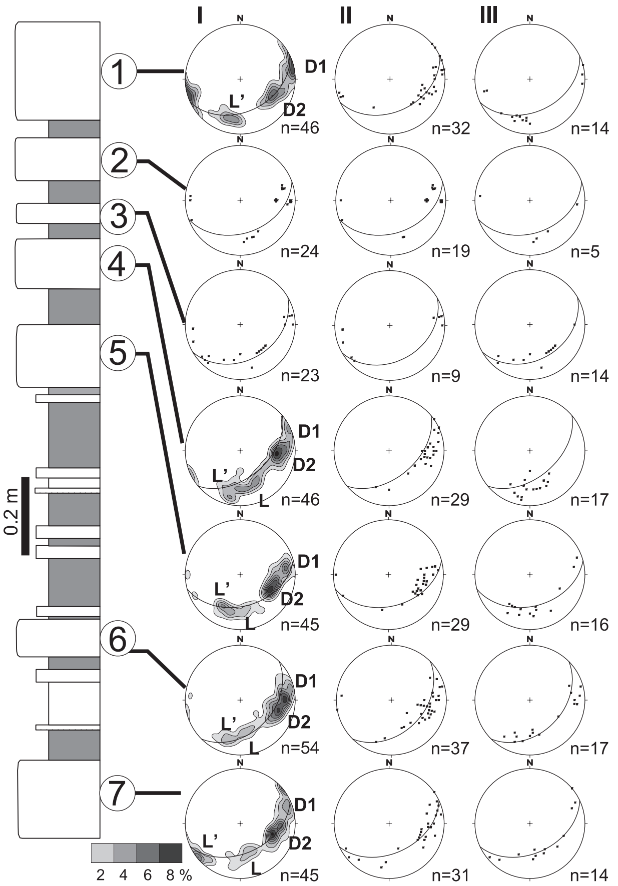
Our study focused on mineralized joints and small-scale faults that cut sandstone beds in a sandstone-mudstone-claystone sequence, 2 m thick (Fig. 19). Altogether, attitudes of 283 joints and 15 faults were measured, and all joints, mineral veins and small-scale faults cutting 7 sandstone beds were examined in detail.

The architecture of joints and faults was reconstructed. Macroscopic observations of mineral veins, especially their cross-cutting relationships, were carried out in the field. The results of our already published microscopic and geochemical studies were also used in the interpretation (Świerczewska et al. 2000b, 2001).

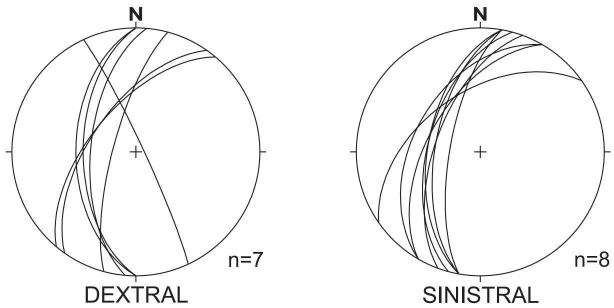
Joints. Attitudes of 283 joints were measured. Four sets of joints (D1, D2, L, and L') can be distinguished in most of the beds. The joints of sets D1 and L, and joints of sets D2 and L' are oriented sub-perpendicular (81–107°) to each other. The joints of sets D1 and D2, and those of sets L and L' are oriented under small angles (17–42°) to each other. The architecture of mineralized joints is distinctly different to that of the barren ones. Most joints of sets D1 and D2 are filled by mineral veins, whereas joints of sets L and L' are mostly barren.

Small-scale faults. Attitudes of 15 minor faults were measured (Fig. 20). Only strike-slip faults, both dextral and sinistral ones, were observed. The faults display strikes from N25W to N50E. The dips of the faults are similar to those of the joints which strike parallel to the faults. Whenever striae occur on fault surfaces, the striae are parralel to the intersection between fault plane and bedding plane. All studied faults are filled by mineral veins.

Textures of mineral veins - microscopic observations. Most of the D1 and D2 joints are healed by columnar (CC), blocky (BC) and drusy (DC) calcite, as well as by blocky calcite and blocky to drusy quartz (CQ). Simple veins filling these joints are composed of CC, BC or DC. In composite veins, the following successions are observed: (i) from CC to BC to DC, (ii) from CC to BC, (iii) from CC to QC to DC. The L and L' joints are filled exclusively by CC. Crystals oriented either perpendicular (CCP) or oblique (CCO) to the walls of veins were observed in



■ **Fig. 19.** Lithological log and attitudes of joints cutting particular beds (1–7): I – all joints, II – mineralized joints, III – barren joints (after Świerczewska et al. 2005).



■ **Fig. 20.** Attitudes of small-scale faults.

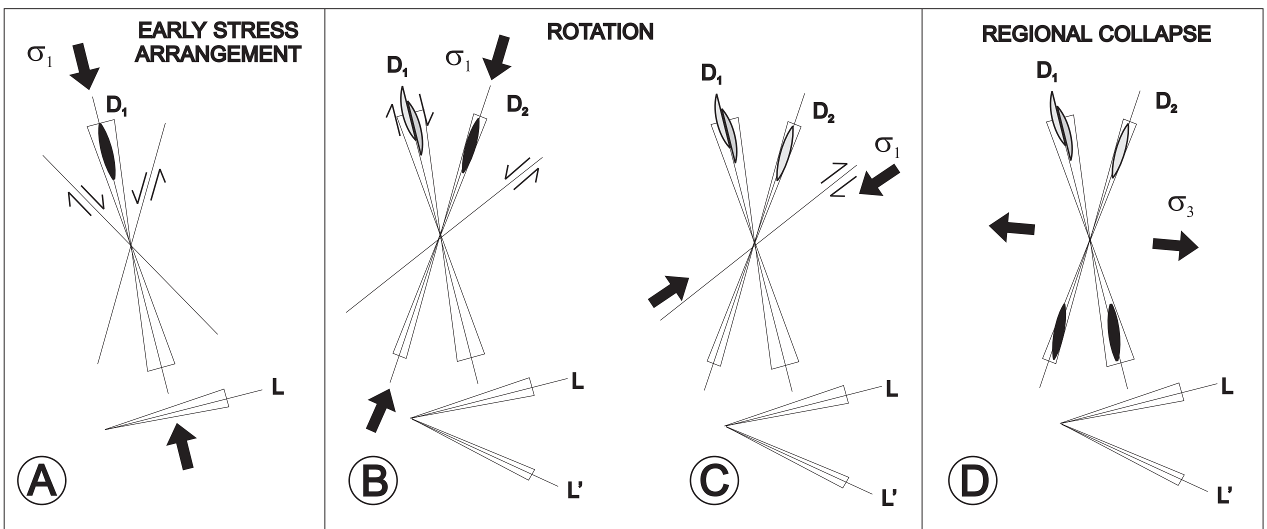
joints	I	II	III	IV
D_1	CCP	CCO	CCO	CCP CCO BC-CQ-DC
D_2		CCP	CCO	CCP CCO BC-CQ-DC
L	CCP	CCO	CCO	
L'		CCP	CCO	

time →

■ **Fig. 21.** Observed mineral successions in mineral veins filling D_1, D_2, L and L' joints.

joints of all sets. CCO indicate either sinistral or dextral sense of strike-slip movement along microfaults (cf. Fig. 20). Synkinematic fibrous calcite was found on surfaces of some small-scale strike-slip faults. In some faults CQ post-date this fibrous calcite.

Microscopic observations were supplemented by macroscopic examination of vein intersections. Results of studies showing succession in filling of the D1, D2, L and L' joints are summarized in Fig. 21. Cross-cutting relationships between composite veins yielded most information about mineral sequences. Basing



	A	B	C	D
D_1	CCP-*BC-DC*	CCO	CCO	CCP CCO BC-CQ-DC
D_2		CCP-*BC-DC*	CCO	CCP CCO BC-CQ-DC
L	CCP	CCO	CCO	
L'		CCP	CCO	
FAULTS	FC CCO			

■ **Fig. 22.** Cartoon (drawn in horizontal plane) illustrating progressive formation of extensional joints marked by ellipses (black when active, grey when inactive) and active small-scale faults (lines with half-arrows); hypothetic sequences of textures in mineral veins related to continuous mineralization are presented in the bottom part of the figure (parts of sequences which were not observed are marked in italics); see text for further explanation.

on these cross-cutting relationships, as well as on cross-cutting of composite and single veins it was possible to recognize CC formed during different stages of infilling.

Discussion and conclusions

The described architecture of joints fits very well the extensional model of jointing (e.g. Engelder and Geiser 1980). Results of kinematic analysis of small-scale faults show that the faults and all joints were formed progressively before and during anticlockwise rotation of the Carpathians (Fig. 22A-C). The amount of the rotation was at least 60°. Collapse-related deformation (Fig. 22D) was due to a WNW-ESE oriented extension.

Hypothetic sequences of textures in mineral veins related to progressive mineralization in a changing stress arrangement are presented in the bottom part of Fig. 22. Except for the *BC-DC* sequence (stages A and B), all sequences can be observed in veins (compare Fig. 21).

Cross-cutting relationships between mineral veins filling joints and small-scale faults appear to be complicated. However, these relationships fit in the hypothetic sequences of textures in mineral veins. Therefore, we believe that the adopted extensional model of jointing is valid in the discussed case. Moreover, the observed cross-cutting relationship of veins shows that the bulk of mineralization filling joints and microfaults was introduced during collapse. This confirms our earlier conclusion (Świerczewska et al. 2000b). Furthermore, there occur two temperature ranges of homogenisation (<50 °C and 100–145 °C) obtained from fluid inclusions trapped in CCP (Świerczewska et al. 2001). This appears to confirm our conclusion that CCP was formed during three stages of structural evolution (A, B, D – Fig. 22).

Stop 5. Pieniny Andesites at Wżar Hill – Mineralogy, Petrology and Palaeomagnetism

Two exposures will be presented: (1) old quarry close to the Snozka pass (road Krościenko-Nowy Targ) – two generations of andesite dykes; petrographic varieties, contact metamorphism, palaeomagnetic results, and (2) exposure near lower station of the ski-lift in Kluszkowce village - two petrographic varieties of andesite; high degree of hydrothermal alterations.

In the old quarry two generations of andesite dykes are well visible. The older generation is usually rusty and significantly altered. The younger one (which was quarried here) is "fresh-looking", dark grey in colour, with porphyritic texture. Abundant amphibole and pyroxenes phenocrysts are typical of this variety. Feldspar phenocrysts are less abundant (feldspar is common in rock matrix). Xenoliths of sedimentary rocks and dark enclaves (composed of agglomerations of mafic minerals), and mafic megacrysts are relatively common in both varieties (Kardymowicz 1957, Bakun-Czubarow and Białowolska 2004).

The older generation of dykes belongs to the first phase intrusions (e.g. Birkenmajer and Pécskay 1999). These are oriented subparallel to the northern boundary fault of the Pieniny Klippen belt. The dykes of younger generation (second phase intrusions) are related to transversal SSE-NNW faults (e.g. Birkenmajer and Pécskay 1999).

Andesites at Wżar hill are emplaced in Palaeogene flysch rocks of the Magura Nappe. Both andesite and sedimentary rocks are altered near contacts. Contact alterations are well visible in the old quarry (changes in colour, hardening of sedimentary rocks) (e.g. Birkenmajer 1958). Contact metamorphism is expressed in formation of garnet and pyroxene in sandstone with abundant carbonate cement (Pyrgies and Michalik 1998). Two chemical varieties of both garnet and pyroxene could be related to two phases of andesite emplacement. Andesite at contact is significantly enriched in calcite. Hydrothermal activity–related minerals overlap those related to contact metamorphism (Pyrgies and Michalik 1998).

In the exposure near lower station of the ski-lift also two varieties of andesite are exposed. Feldspar phenocryst-rich rock, characterized by numerous voids filled with secondary minerals, contacts with a variety dominated by mafic minerals. Hydrothermal alterations zones and hydrothermal veins exposed in this outcrop have not been studied in detail. (e.g. Gajda 1958 a, b, Szeliga and Michalik 2003).

Mineralogy. Plagioclase phenocryst composition varies from oligoclase to bytovnite. Most plagioclase phenocrysts are zoned. In some samples, cores of plagioclase phenocrysts rich in glass inclusions are rimmed by euhedral and inclusion-free zones (Michalik et al. 2004). Plagioclases are partly replaced by calcite and chlorite (especially in feldspar phenocryst- rich variety from the outcrop at Kluszkowce village). Amphibole phenocrysts represent magnesian hastingsitic hornblende and magnesian hastingsite and ferroan pargasite, ferroan pargasitic hornblende, edenitic hornblende, according to Leake's (1978) classification. Amfiboles are often replaced completely or partly by chlorite, calcite, Fe-Ti oxides, and titanite. Coronas of complex composition (fine-crystalline aggregates of plagioclases, pyroxene, Fe-Ti oxide minerals) are common around amphiboles. Pyroxene phenocrysts are represented by diopside (Morimoto's 1988 classification). Alterations of pyroxene result in formation of chlorite and calcite. Fe-Ti oxides phenocrysts belong to ulvöspinelo-magnestite series. Fibrous illite-like mineral is commonly observed using SEM. Mn-Fe oxides of corn-flake morphology aggregates are developed around mafic phenocryts (Michalik et al. 2004, 2005).

XRD study of < 2 μm fraction separated from andesite samples indicates that it is composed of illite/smectite (and vermiculite/smectite?) and cristobalite. Less than 2 μm fraction separated from andesite samples and studied using XRD method are composed of illite/smectite (and vermiculite/smectite?) and cristobalite (samples from Wżar) (Michalik et al. 2005). Abundance of secondary minerals related to hydrothermal activity and presence of newly formed K-containing minerals (illite) are the reason for difficulties in K-Ar dates interpretations.

Palaeomagnetism. On Wżar Hill both the older and younger Miocene andesites crop out in abandoned quarries. The contact zone between the two phases of intrusions is well exposed, but heavily weathered. Palaeomagnetic sampling of the andesites required special care, for the intrusions themselves are also quite altered. Thus, we drilled long cores and found that weathering indeed fully remagnetized the upper 10 cm or so of the andesites. Useful palaeomagnetic signal was only obtained from the deepest part of the about 20-cm-long cores. Both phases of intrusions possess composite natural remnant magnetization. The mostly dominant component was reversed polarity and exhibits no rotation with respect to stable European reference declination. The other, more stable, component shows moderate counterclockwise rotation (of about 30 ° in average). This component was not found during previous palaeomagnetic investigations, most probably because the samples were demagnetized by alternating field, a method which was not suitable to separate components.

The Wżar Mts. andesites are of key importance for the tectonic interpretation of palaeomagnetic data obtained also from other Pieniny andesite outcrops. They suggest that the area intruded by the andesites rotated counterclockwise after the second intrusion phase.

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Conference Excursion 2: Late Cretaceous–Neogene Evolution of the Polish Carpathians

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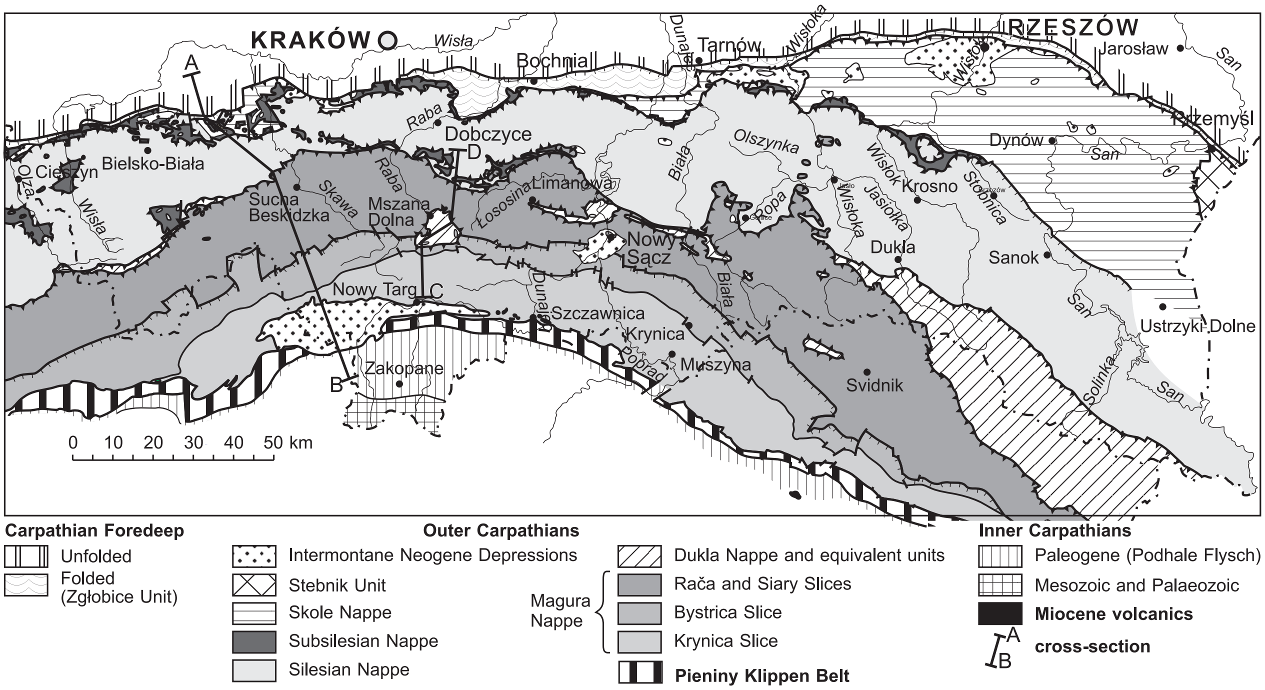
Itinerary: Zakopane – Maruszyna – Nowy Targ – Obidowa – Rabka – Mszana Dolna – Skrzydlna – Poręba Górna – Tylmanowa – Szczawnica Spa – Zabaniszcze – Czorsztyn – Snozka Pass-Wżar – Zakopane

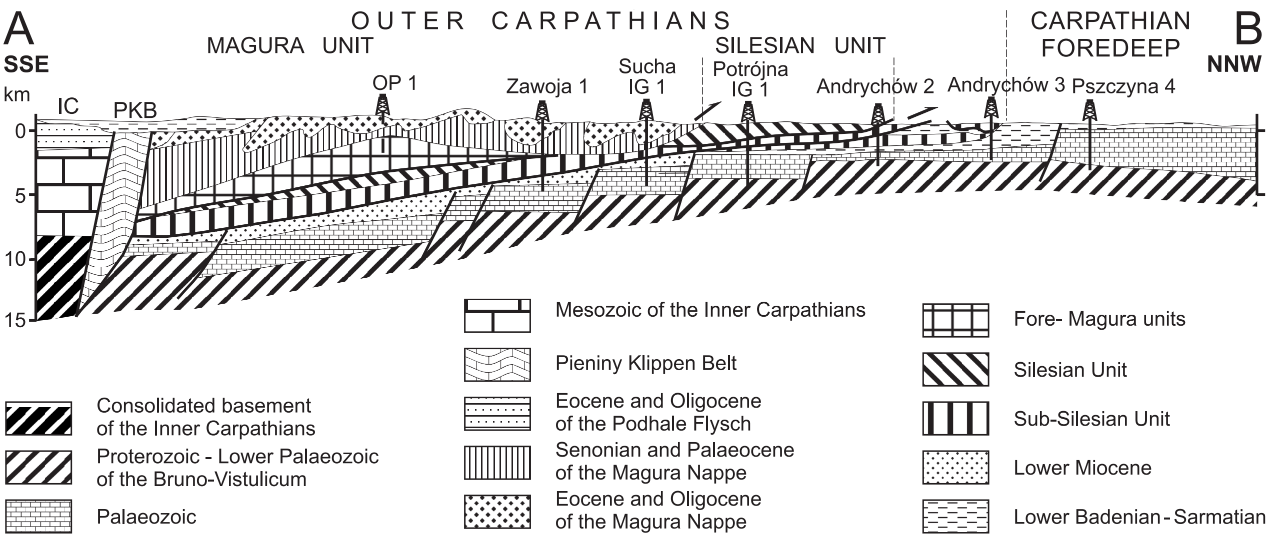
Regional setting

The Polish Carpathians are a part of the great arc of mountains, which stretches for more than 1300 km from the Vienna Forest to the Iron Gate on the Danube. In the west, the Carpathians are linked with the Eastern Alps and, in the east they pass into the Balkan chain. Traditionally, the Western Carpathians have been always subdivided into two distinct ranges. The Inner Carpathians are considered the older range and the Outer Carpathians the younger one (Fig. 1) (Książkiewicz 1977). The Inner Carpathians are regarded as a prolongation of the Northern Calcareous Alps, and formed part of the Apulia Plate in regional sense that is a promontory of the African Plate (Picha et al. 2005 and references therein). Sedimentation in the Inner Carpathians was mainly calcareous, and took place from the Early Triassic to the mid-Cretaceous. The Inner Carpathians were folded during the Late Cretaceous tectonic movements.

Between the Inner and Outer Carpathians the Pieniny Klippen Belt (PKB) is situated. It is Tertiary strike-slip boundary, which is a strongly tectonized terrain about 800 km long and 1–20 km wide (Birkenmajer 1986). The Outer Carpathians are built up of stacked nappes and thrust-sheets, which reveal different lithostratigraphy and structure (Figs. 1, 2). The Outer Carpathians are composed of the Late Jurassic to Early Miocene mainly turbidite (flysch) deposits, completely uprooted from their basement. The largest and innermost unit of the Outer Carpathians is the Magura nappe – a Late Oligocene/Early Miocene accretionary wedge. The Magura nappe is flatly overthrust onto the Moldavides (Ślączka et al. 2005 and references therein) - an Early/Middle Miocene accretionary wedge, which consists of several nappes: the Fore–Magura–Dukla group, Silesian, Sub-Silesian, Skole and Boryslav-Pokuty units. In the Outer Carpathians the main decollement surfaces are located at different stratigraphic levels. The Magura nappe was uprooted from its substratum at the base of the Turonian-Senonian variegated shales (Oszczypko 1992), whereas the main decollement surfaces of the Moldavides are located in the Lower Cretaceous black shales. All the Outer Carpathian nappes are flatly overthrust onto the Miocene deposits of the Carpathian Foredeep (see Oszczypko and Tomaś 1985, Ślączka et al. 2005 and references therein). However, along the



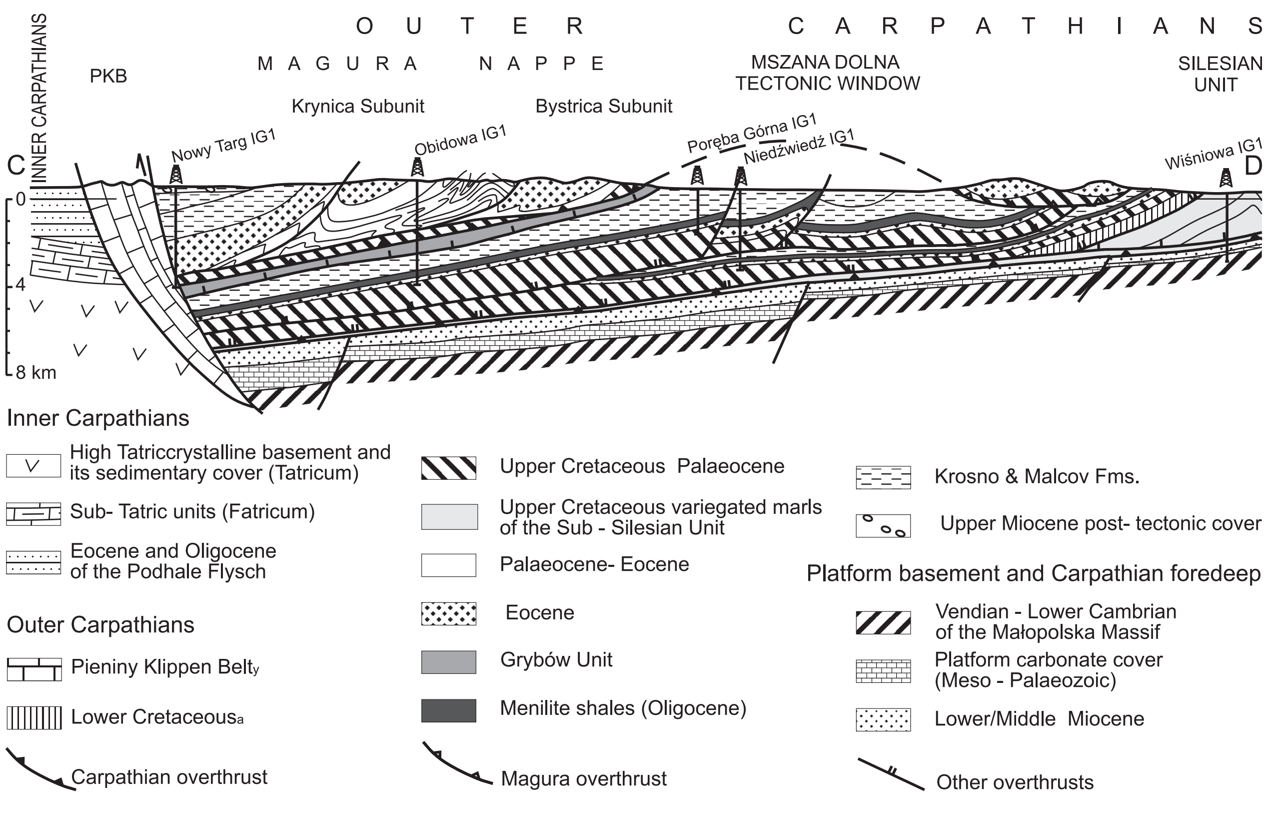
■ **Fig. 1.** Geological map of the Polish Carpathians (based on Żytko et al. 1989).



■ **Fig. 2.** Geological cross-section Orawa-Sosnowiec (after Oszczypko et al. 2005).

frontal Carpathian thrust a narrow zone of folded Miocene deposits developed [Stebnik (Sambir) and Zgłobice units (Figs.1, 2)]. The detachment levels of the folded Miocene units are connected with the Lower and Middle Miocene evaporites.

The basement of the Carpathian Foredeep represents the epi-Variscan platform and its cover (Oszczypko et al. 1989, 2005). The depth to the platform basement, recognized by boreholes, changes from a few hundred metres in the marginal part of the foredeep up to more than 7000 m beneath the Carpathians (Figs. 1, 2, 3). The magneto-telluric soundings in the Polish Carpathians have revealed a high resistivity horizon, which is connected with the top of the consolidated – crystalline basement (Ślączka et al. 2005 and references therein). The depth of the top of magneto-telluric basement reaches about 3–5 km in the northern part of the Carpathians, drops to approximately 15–20 km at its deepest point and then peaks at 8–10 km in the southern part. The axis of the magneto-telluric low coincides, more or less, with the axis of gravimetric minimum. South of the gravimetric minimum and, more or less parallel to the PKB, the zone of zero values related to of the Wiese vectors, was recognised by geomagnetic soundings (Ślączka



■ **Fig. 3.** Geological cross-section Nowy Targ IG 1 – Wiśniowa IG 1 (after Oszczypko 2004) et al. 2005 and references therein). This zone is connected with a high conductivity body occurring at the depth of 10–25 km and is located at the boundary between the North European Plate and the Central West Carpathian Block. In the Polish Carpathians, the depth to the crust-mantle boundary ranges from 37–40 km at the front of the Carpathians and increases to 54 km towards the south and then, peaks along the PKB at 36–38 km (Oszczypko 2004).

The deep seismic reflection profile 2T was located southwest from the Polish state boundary, (Golonka et al. 2005a). In the north of Pieniny Klippen Belt, this profile demonstrates two groups of south dipping reflectors, which are probably related to the Middle Miocene subduction of the Moldavides (Tomek and Hall 1993, Bielik et al. 2004). The upper reflection between 1–3 s (ca. 4.5 to 8 km) belongs to a plate boundary between the upper nappe (the Magura-PKB terrain), and the lower accretionary wedge complex (Dukla-Silesian-Subsilesian group of units). The lower reflectors represent the crystalline basement of the lower plate (North European Plate), and its sedimentary cover.

During the excursion the following Outer Carpathian units will be passed: The Magura Nappe, Dukla-Grybów units (Mszana Dolna tectonic window) and Silesian/Sub-Silesian units

Outer Carpathians

The Outer Carpathians are built up of a stack of nappes and thrust sheets\spreading along the Carpathians, consisting mainly of up to six kilometers thick continuous flysch sequences of Jurassic to Early Miocene time span. All the Outer Carpathian nappes are overthrusted onto the southern part of the North European Plate (Figs. 1, 2, 3) and covered by the autochthonous Miocene deposits of the Carpathian Foredeep at the distance of 70 km at least (Książkiewicz 1977, Oszczypko 2004, Ślączka et al. 2005 and references therein). Boreholes and seismic data indicate that the Carpathian overthrust extends at the distance at least of 60 km (Fig. 1, 2, 3). During overthrusting, the northern Carpathians nappes became uprooted from the basement and only their basinal parts were preserved. A narrow zone of folded Miocene deposits, developed along the frontal Carpathian thrust and known as the Zgłobice-Wieliczka Unit in the Northern Carpathians has its equivalent in the Subcarpathian (Borislav or Sambir-Rozniatov Unit of the Ukrainian part) and Romanian parts of the Eastern Carpathians (Książkiewicz 1977, Ślączka et al. 2005 and references therein). From the North to the South, the succession of nappes from the lowest to the highest one includes Skole (Skiba) Nappe (mainly easternmost part of Carpathians), Subsilesian Nappe, Silesian Nappe, Fore-Magura group of nappes and Magura Nappe (Figs. 1, 2, 3).

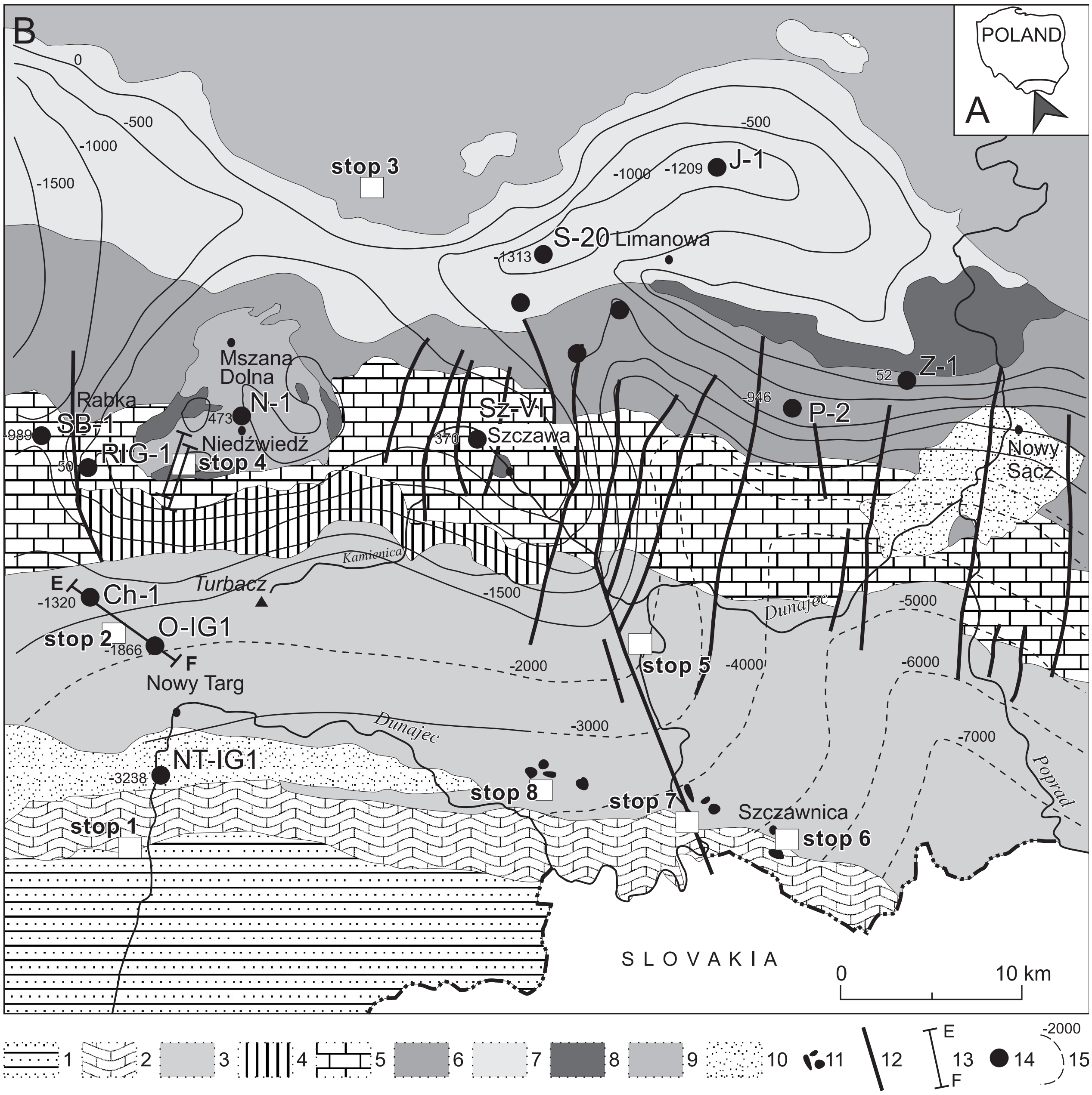
The **Skole Nappe** (Ślączka et al. 2005 and references therein) occupies a large area in the northeastern part of the Polish Outer Carpathians. Towards the east on the territory of Ukraine, it goes to be wider but towards the west, its width diminishes until it eventually disappears from surface, plunging beneath the Silesian and Subsilesian Nappes. The Skole Nappe consists of several narrow elongated thrust folds. There is predominance of the Oligocene Menilite and Krosno beds cropping out on the surface in the inner zone of this unit while outer is mainly built of Cretaceous strata. In the Skole Basin sedimentation started not later than in Hauterivian. Shales and marls (the Belwin Marls) and black shales (the Spas Beds) represent lower Cretaceous.

The **Subsilesian Nappe** underlies tectonically the Silesian Nappe. In western sector of the West Carpathians both nappes are thrust over the Miocene molasse of the Carpathian Foredeep and in eastern sector they are thrust over the Skole Nappe. This nappe consists of Upper Cretaceous–Palaeogene flysch deposits.

The **Andrychów Ridge** unit is represented by several huge blocks in the boundary between the Silesian and Subsilesian units, near Andrychów Town. Probably they are remnants of carbonate platform, which has been situated between Silesian and Subsilesian sedimentary areas, or represent a part of Subsilesian substratum. The composition of klippes differs from the adjacent units, although the Upper Cretaceous sediments show some similarity to the sequences of the Subsilesian Unit. The non-flysch, calcareous facies are very characteristic of the Andrychów Ridge sequences (Książkiewicz 1951, Gasiński 1998). The basement of the ridge consisted of granite-gneiss or mylonitised rocks of unknown age. Crinoidal and shallow water Middle-Upper Jurassic limestones represent the sedimentary sequences. Transgressive Lower Campanian conglomerates as well as marls, limestones and shaly marls of Campanian and Maastrichtian age cover these. Organogenic limestones and shales represent Palaeocene and Lower Eocene. The more basinal or slope facies are developed as the Maiolica-type Upper Jurassic-Lower Cretaceous cherty limestones (Olszewska and Wieczorek 2001).

The **Silesian Nappe** occupies central part of the Outer Carpathians, pinching out from below of the most internal nappes. Sedimentary facies of the Silesian Nappe represent continuous succession of deposits from the Upper Jurassic to the Lower Miocene. In the Western Carpathians, the oldest sediments of the Silesian Nappe are known in Moravia and Silesia areas only. They consist of Kimmeridgian-Lower Tithonian dark grey, calcareous mudstones (Lower Cieszyn Shales), which begin euxinic cycle that has lasted without notable gaps until Albian. Some of these mudstones represent slump deposits. The mudstones pass upwards into turbiditic limestones and marls namely 200 m thick Cieszyn Limestones of Tithonian-Berriasian age. The material of the detritic limestones was derived from the neighboring and northerly-situated shallow water, carbonate platforms. The younger, Valanginian-Hauterivian dark grey, black calcareous shales with intercalations of dark, thin and medium bedded, calcareous sandstones (Upper Cieszyn Shales, up to 300 meters thick) are known from the whole Silesian Unit. Barremian-Aptian black shales (Verovice Shales, up to 300 meters) overlie them. During the Hauterivian and Barremian time, several complexes (up to 200 meters thick) of sandstones and conglomerates have developed (Grodziszcze Sandstones). There are evidences that the Silesian and Subsilesian basins have been connected in the Cretaceous time.

The **Magura Nappe** is the largest and innermost tectonic unit of the Western Carpathians and is linked with the Rheno-Danubian flysch of the Eastern Alps. During the overthrust movements, the Magura Nappe has been completely uprooted from its substratum along the ductile Upper Cretaceous rocks. The oldest Jurrassic-Lower Cretaceous rocks are only found in this part of the Magura basin which was incorporated into the Pieniny Klippen Belt (i.e. the Grajcarek Unit, see Birkenmajer 1977, see also Oszczypko et al. 2004).



■ **Fig. 4. A** – Position of the investigated area. **B** – Sketch-map of the middle part of the Polish Carpathians (after Oszczypko et al. 1999a, Oszczypko-Clowes and Oszczypko 2004). 1 – Podhale Flysch, 2 – Pieniny Klippen Belt; Magura Nappe: 3 – Krynica Subunit, 4 – Tobołów-Turbaczyk thrust sheet, 5 – Bystrica Subunit, 6 – Rača Subunit, 7 – Siary Subunit, 8 – Grybów Unit, 9 – Dukla Unit, Silesian & Sub-Silesian units, 10 – Miocene onto the Carpathians, 11 – Miocene andesites, 12 – faults, 13 – cross section, 14 – boreholes, 15 – isobathe of the Magura nappe overthrust.

Stratigraphy and tectonics. The Albian /Cenomanian spotty shales remained in the southern margin of the Mszana Dolna tectonic window (Oszczypko et al. 2005b). More recently, Hauterivian-Albian deposits have been recognised in a few localities in Southern Moravia. On the basis of facial differentiations with regards to the Palaeogene deposits, the Magura Nappe has been subdivided into four facies-tectonic subunits: the Krynica, Bystrica (Nowy Sącz), Rača and Siary (Fig. 1). The Upper Cretaceous-Palaeogene deposits of the Magura Nappe may be subdivided into three turbidite complexes: the Campanian/ Maastrichtian-Palaeocene, Lower-Upper Eocene, and Upper Eocene-Lower Oligocene. Each of them begin with pelitic basinal deposits (variegated shales) which pass into thin-and medium-bedded turbidites with intercalations of allodapic limestones /marls, and then into thick-bedded ones. Finally, there are thin-bedded turbidites. In the Krynica, Bystrica and Rača subunits the youngest deposits of the Eocene complex belong to the Magura Fm., which is of Lower-to Upper Eocene age (Birkenmajer and Oszczypko 1989, Oszczypko 1991). This formation is reached in the Krynica Subunit, 1200 to 1400 m thickness and at least 500–2000 in the Bystrica and 1000 m in the Rača subunit.

The Magura Fm. is represented by the thick-bedded turbidites and fluxoturbidites. The Globigerina marls (Upper Eocene-Lower Oligocene), Menilite shales and the Malcov Fm (Late Oligocene), locally overlay the Magura Fm. In the northernmost part of the Magura Nappe (Siary subunit) the youngest deposits are composed as the thick-bedded glauconitic sandstones (Wątkowa Sandstones) of the Lower Oligocene age, and finally by marls with intercalations of glauconitic sandstones (Budzów Beds, Oligocene). Traditionally, the Oligocene Malcov Formation was regarded as the youngest strata of the Magura nappe. However, in the peri-Pieniny Klippen Belt area and the vicinity of Nowy Sacz the Early Miocene flysch deposits have recently been discovered (see Oszczypko et al. 1999a, Oszczypko and Oszczypko-Clowes 2002, Oszczypko et al. 2005c and bibliography therein).

The Magura Nappe is flatly thrust over its foreland, built up of the Fore-Magura group of units and partly by the Silesian Unit (Figs. 1, 2, 3). The amplitude of the overthrust is at least 50 km, and the post Middle Badenian thrust displacement is more than 12 km. The northern limit of the nappe has an erosional character, whereas the southern coincides with a more or less vertical strike-slip fault along the northern boundary of the PKB. The thrust developed mainly within the ductile Upper Cretaceous variegated shales. The sub-thrust morphology of the Magura foreland is very distinctive. The shape of the northern limit of the Magura Nappe and the distribution of tectonic windows inside the nappe are connected with denivelation of the Magura basement. As a rule, the "embyemets" of the marginal thrust are related to transversal bulges in the Magura basement, whereas the "peninsulas" are located in the depression of basement. At a distance of 10–15 km south from the northern limit of the unit the zone of the tectonic windows, connected with uplifted Fore-Magura basement, is located. The biggest is the Mszana Dolna tectonic window, situated in the middle part of the Polish Carpathians. This tectonic window developed as the duplex structure during the Middle Miocene thrusting of the Magura Nappe. South of the zone of tectonic windows the inclination of the Magura thrust surface increases, and at the northern boundary of PKB the thickness of the Magura Nappe is more than 5 km. The Magura Nappe has been subdivided into four structural subunits (thrust sheets) namely the Krynica, Bystrica (Nowy Sącz), Rača and Siary (Fig. 1). These subunits coincide, largely, with the corresponding facies zone. In the Magura accretionary prism, three structural complexes can be distinguished: the Late Cretaceous-Palaeocene, the Early to Late Eocene, and the Oligocene to Early Miocene. These complexes revealed a decreasing degree of tectonic deformation from the base to the top of the nappe. The basal part of the nappe, built up of Upper Cretaceous-Palaeocene flysch rocks is strongly deformed in the area surrounding the Mszana Dolna and Szczawa tectonic windows. The broad W-E trending synclines and narrow anticlines dominate in Lower to Upper Eocene flysch of the Rača and Krynica subunits. The southern limbs of synclines are often reduced and overturned. In the Bystrica (Nowy Sącz) Subunit, sub-vertical thrust-sheets are common. Both the northern limbs of the anticlines and southern limbs of the synclines are tectonically reduced and usually overturned. The youngest (Malcov Fm, Late Eocene–Early Oligocene), weekly deformed, deposits of the Magura Nappe uconformably overlaid the older Eocene flysch deposits in the Krynica and Rača subunits.

Tectonic evolution. The Magura Basin was probably located on the oceanic floor and/or the thinned continental crust, and began to develop during the Liassic-Dogger continental rifting (Oszczypko, 1992, 1999). That rift was relocated into the Silesian Basin during the Early Tithonian. This event was followed by a long-lasting Berriasian-Cenomanian (35 my) period of basin expansion, and deep-water pelagic deposition, probably connected with passive thermal subsidence. This period was characterised by very low rates of sedimentation (0.5–5 m/my). Before the Albian, the Magura Basin was separated from the Silesian Basin by the Protho-Silesian submerged ridge, but during the Cenomanian – Turonian time there was a unification of sedimentary conditions in the whole basin of the Outer Carpathians, and radiolarian shales were deposited. At the turn of Cenomanian, radiolarian shales followed by red clays with intercalations of basinal turbidites were deposited below the CCD. In the northern and middle part of the Magura Basin this type of sedimentation persisted up to the Campanian, whereas in the Krynica zone until the Maastrichtian. The rate of sedimentation of variegated shales oscillated between 15 to 25 m/my. During the Maastrichtian/Palaeocene time, a considerable reorganisation of the Magura Basin took place. This was connected both with compression at the southern margin of the basin and an inversion of the forbulge at the northern margin (Silesian uplifted ridge). It was accompanied by a deposition of the Upper Senonian-Palaeocene flysch (so called Inoceramian beds). The rate of sedimentation oscillated between 50 and 75 m/my. Since the Palaeocene/Early Eocene, the accretionary prism has begun to develop in the southern margin of the Magura Basin, close to folded and thrust Pieniny Klippen Belt. The moving load, in front of this accretionary prism, has caused subsidence and a progressive northward shift of depocentres. The Early Eocene axis of deposition was located in the Krynica zone, and then during the Middle and Late Eocene migrated northward, towards the Bystrica and Rača zones respectively. In this part of the basin narrow and very long submarine fan (few hundred km) was formed. The clastic immature material of the fan was supplied from a southeast direction, and was derived from an erosion of the exposed part of the accretionary prism. During the Early to Middle Eocene time, the deepest part of the basin, often beneath the CCD, was located in the northern part of the basin. The rate of sedimentation varied from 10–15 m/my on the abyssal plain, to 75–100 m/my in the outer fan and, between 200 to 300 m/my in the area affected by middle fan-lobe system. Simultaneously along the northern margin of the basin (Siary zone), small fans developed. During the Late Eocene, the southern part of the Magura basin was involved in submarine folding, caused by a southward subduction of the Magura Basin beneath the Pieniny Klippen Belt/Central Carpathian Block. Late Eocene to Early Oligocene subsidence and an intensive deposition in the Siary zone, supplied from the Silesian Ridge by mature glauconite sandstones and massive turbidite marls, followed this event. After the Late Oligocene folding, the Magura Nappe was thrust northwards and during Burdigalian its front reached the S part of the Silesian basin. These movements were almost contemporaneous with that of the Northern Calcareous Alps and Rhenodanubian Flysch. Simultaneously the southern part of the Magura Nappe was transformed into the piggy-back basin flooded during the Early Burdigalian high stand of sea-level. The Early Burdigalian rise in sea level enabled the connection between the Magura piggy-back basin and the Vienna basin via Orava. In the Magura piggy-back basin occurred the deposition of Kremna facies close to PKB and the Zawada Formation in the more northern part of the basin (see Oszczypko et al. 1999a, Oszczypko and Oszczypko-Clowes 2002, Oszczypko et al. 2005c). The terminal flysch deposition in the Krosno basin and the Magura piggy-back-basin was followed by the Intra-Burdigalian folding, uplift and overthrust of the Outer Carpathians onto the foreland platform (Oszczypko 1997, 1998, Kovač et al. 1998).

Pieniny Klippen Belt

Outline of geology

The Northern Carpathians are subdivided into an older range known as the Inner Carpathians and the younger one, known as the Outer or Flysch Carpathians (Figs. 1, 2). The Pieniny Klippen Belt (PKB) is situated at the boundary of these two ranges. The Inner Carpathians nappes contact along a Tertiary strike-slip boundary with Pieniny Klippen Belt.

The relationship between Pieniny Klippen Belt and the Magura Nappe changes along the PKB strike. In the Vah and Orava valleys these two units are divided by the Miocene sub-vertical strike-slip fault and both units are involved in the complex flower structure. Present day confines of the Pieniny Klippen Belt are strictly tectonic. They may be characterized as a (sub)vertical faults and shear zones, along which a strong reduction of space of the original sedimentary basins took place. The NE-SW striking faults accompanying the Klippen Belt have the character of lateral slips. It is indicated by the presence of flower structures on the contact zone of the Magura Unit and the Klippen Belt, or by the structural asymmetry of the Inner Carpathian Palaeogene Basin.

The tectonic character of the Polish section of PKB is mixed. Both the strike slip and thrust components occur here (e.g. Książkiewicz 1977, Birkenmajer 1986, Golonka et al. 2005b and references therein). In general the subvertically arranged Jurassic-Lower Cretaceous basinal facies display the tectonics of the diapir character originated in the strike-slip zone between two plates. The ridge facies are often uprooted and display thrust or even nappe character. The Niedzica Succession is thrust over the Czorsztyn Succession, while the Czorsztyn Succession is displaced and thrust over the Grajcarek Unit (e.g. Książkiewicz 1977, Birkenmajer 1986, Golonka et al. 2005a, b and references therein). The Grajcarek Unit is often thrust over the Krynica Sub-Unit of the Magura Nappe. The Upper Cretaceous-Palaeogene flysch sequences of the Złatne Furrow (Birkenmajer 1986, Golonka et al. 2005a, b and references therein) are often thrust over the various slope and ridge sequences. In the East Slovakian section of the PKB, the back-thrusts of the Magura Nappe onto PKB, as well as PKB onto the Central Carpathian Palaeogene, are commonly accepted (Golonka et al. 2005 a,b and references therein). The PKB tectonic components of different age, strike-slip, thrust as well as toe-thrusts and olistostromes mixed together, are giving the present-day melange character of the PKB, where individual tectonic units are hard to distinguish.

The Pieniny Klippen Belt is composed of several **successions** of mainly deep and shallower-water limestones, covering a time span from the Early Jurassic to Late Cretaceous (Birkenmajer 1977, 1986, Golonka and Krobicki 2004, Golonka et al. 2005 a, b and references therein). This strongly tectonized structure is about 600 km long and 1–20 km wide, stretching from Vienna to the West, to Romania to the East.

During the Jurassic and Cretaceous within the Pieniny Klippen Basin the submarine Czorsztyn Ridge (= "pelagic swell", mainly so-called Czorsztyn Succession) and surrounding zones formed an elongated structure with domination of pelagic type of sedimentation (Birkenmajer 1977, 1986, Golonka and Krobicki 2004, Wierzbowski et al. 1999, Golonka et al. 2005 a, b and references therein). The Pieniny Klippen Basin trends SW to NE (see discussion in Golonka and Krobicki 2001, 2004). Its deepest part shows the presence of deep water Jurassic-Early Cretaceous deposits (pelagic limestones and radiolarites) of Złatna Unit (Golonka and Sikora 1981, Golonka and Krobicki 2001, 2004). Somewhat shallower sedimentary zones known as the Pieniny, Branisko (Kysuca) successions have been located close to central furrow. Transitional slope sequences between basinal units and ridge units are known as Czertezik and Niedzica successions (Podbiel and Pruské successions in Slovakia) near the northern (Czorsztyn) Ridge, and Haligovce-Nižná successions near the southern so-called Exotic Andrusov Ridge (Birkenmajer 1977, 1986, Golonka et al. 2005 a, b and references therein). The strongly condensed Jurassic-Early Cretaceous pelagic cherty limestones (Maiolica-type facies) and radiolarites of the Grajcarek Unit were also deposited in north-western Magura Basin.

Geodynamic evolution

The **earliest stage** of the basin history is enigmatic and documented only by pebbles in the Cretaceous-Tertiary flysch. These pebbles indicate the possibility of an existence of an enigmatic embayment of the Vardar-Transilvanian Ocean which separated the Tisa (Bihor-Apuseni) block from the Moesian-Eastern European Platform (Golonka 2004). The pelagic Triassic limestones in the exotic pebbles in the Pieniny Klippen Belt 1990) and Magura Unit could have originated in this embayment (Golonka et al. 2005 b and references therein) The embayment position and its relation to the other parts of Tethys, Vardar Ocean, Meliata-Halstatt Ocean, Dobrogea rift remain quite speculative. The Pieniny rift opened during Pliensbachian – Aalenian time forming a part of the global system related to the opening of the Alpine Tethys. The Alpine Tethys, that is Ligurian, Penninic and Pieniny/Magura Oceans constitute the extension of the Central Atlantic system (Golonka 2004). The basins' opening is related to the closure of the Meliata Ocean. The restricted environment prevailed in this newly formed basin. The synrift stage lasted in the Pieniny/Magura Basin from late Early Jurassic to Tithonian.

Generally, the Pieniny Klippen Basin sedimentary history is tripartite (i-iii) – from the (i) oxygen-reduced dark/black terrigenous deposits of the Early-early Middle Jurassic age (Gresten-type and Fleckenkalk/Fleckenmergel facies) trough (ii) Middle Jurassic-earliest Cretaceous crinoidal, nodular (of the ammonitico rosso type) or cherty (of the Maiolica = Biancone type) limestones and radiolarites up to the (iii) Late Cretaceous pelagic marls (i.e. Scaglia Rossa = Couches Rouge = Capas Rojas) facies and/or flysch/flyschoidal series (i.a. Birkenmajer 1986, Golonka and Krobicki 2004, Golonka et al. 2005b and references therein).

The **oldest Jurassic** rocks of the Pieniny Klippen Belt (e.g., Birkenmajer 1977, Krobicki et al. 2003, Golonka and Krobicki 2004 and references therein) consist of different type of Gresten-like dark/black sediments (Hettangian-Sinemurian). Spotty limestones and marls of oxygen-depleted, widespread Tethyan Fleckenkalk/Fleckenmergel-type facies (Pliensbachian – Early Bajocian in age) and Bositra ("Posidonia") black shales with spherosiderites (Birkenmajer 1986) overlay them.

One of the most rapid changes of sedimentation/paleoenvironments within this basin took place during Early Bajocian (**Middle Jurassic**) when well-oxygenated multicoloured crinoidal limestones replaced dark and black deposits of Early-early Middle Jurassic age (Birkenmajer 1986, Wierzbowski et al. 1999, Golonka and Krobicki 2004). Sedimentation of younger (from latest Bajocian) red nodular ammonitico rosso-type limestones was an effect of Meso-Cimmerian vertical movements which resulted in subsidence of the Czorsztyn Ridge and produced tectonically differentiated blocks, and associated neptunian dykes and scarp-breccias (e.g. Birkenmajer 1986, Wierzbowski et al. 1999, Aubrecht and Tunyi 2001, Golonka et al. 2003).

The formation of the Czorsztyn Ridge within the Pieniny/Magura Basin took place during the spreading phase in the Bajocian. In the same time the radiolarite sedimentation appeared in the axial, basinal sequences of PKB (Birkenmajer 1977, Wierzbowski et al. 2004). This strongly contrasted facies pattern between basinal and ridge successions have existed throughout Jurassic and Early Cretaceous times.

Late Jurassic (Oxfordian-Kimmeridgian) history of the PKB reflects strong facial differentiation within sedimentary basin where contrasted siliceous to carbonate sedimentation took place. This may be at least partly attributed to radical and fast paleogeographic evolution of the Pieniny Klippen Basin, as indicated by recent palaeomagnetic results in its eastern section (Lewandowski et al. 2003a, 2005). Greatest deepening effect is indicated by widespread Oxfordian radiolarites which occur in the all the basinal and transitional successions, whereas the shallowest one (Czorsztyn Succession) is completely devoid of siliceous intercalations showing sedimentation of ammonitico rosso facies.

The Czorsztyn Succession during **latest Jurassic–earliest Cretaceous** time (Tithonian-Berriasian) consisted of hemipelagic to pelagic organogenic carbonate deposits of medium depth, for example ammonite coquinas and white and creamy *Calpionella*-bearing limestones. Several tectonic horsts and grabens were formed, rejuvenating some older, Eo- and Meso-Cimmerian faults (Birkenmajer 1986, Golonka and Krobicki 2004 and references therein). Such features resulted from the intensive Neo-Cimmerian tectonic movements and have been documented by rapid facies pattern changes, hardgrounds and condensed beds with ferromanganese-rich crusts and/or nodules, sedimentary-stratigraphic hiatuses, sedimentary breccias and/or neptunian dykes, and/or fauna redeposition (Birkenmajer 1986, Krobicki 1996, Golonka and Krobicki 2001, Golonka et al. 2003 and references therein). In the same time within deeper successions (mainly Branisko and Pieniny ones) cherty limestone of maiolica-type (= biancone) facies deposited (Birkenmajer 1977, Golonka et al. 2003 and references therein). These white-gray, micrite well bedded calpionellids-bearing limestones built now highest part of the Pieniny Mts (e.g., Trzy Korony Mt, Sokolica Mt etc).

During the **Tithonian** time, subduction jumped to the northern margin of the Inner Carpathian terranes and began to consume the Pieniny Klippen Ocean (Birkenmajer 1986). The Tethyan plate reorganization resulted in extensive gravitational fault movement.

Lower Cretaceous (Berriasian and Valanginian) rocks are represented by both pelagic, maiolica-type cherty limestones as basinal facies and more shallow-water different type of organodetritic rocks (brachiopod-crinoidal limestones and/or crinoidal limestones).

Deeper facies have continued sedimentation in younger time, whereas shallower one manifested a numerous breaks of sedimentation. Latest Early Cretaceous deposits are connected with first step of unification of sedimentation within Pieniny Klippen Basin (Birkenmajer 1977).

Upper Cretaceous pelagic deposits dominated by scaglia rossa pelagic, foraminiferal, multicoloured green/variegated/red marl deposits originated during the latest, third episode of evolution of the Pieniny Klippen Basin (Birkenmajer 1977, 1986). Flysch and/or flyschoidal facies are younger (Santonian-Campanian) with several episodes of debris flows with numerous exotic pebbles (Late Albian-Early Campanian) (Birkenmajer 1986, Golonka and Krobicki 2004, Golonka et al. 2005b and references therein).

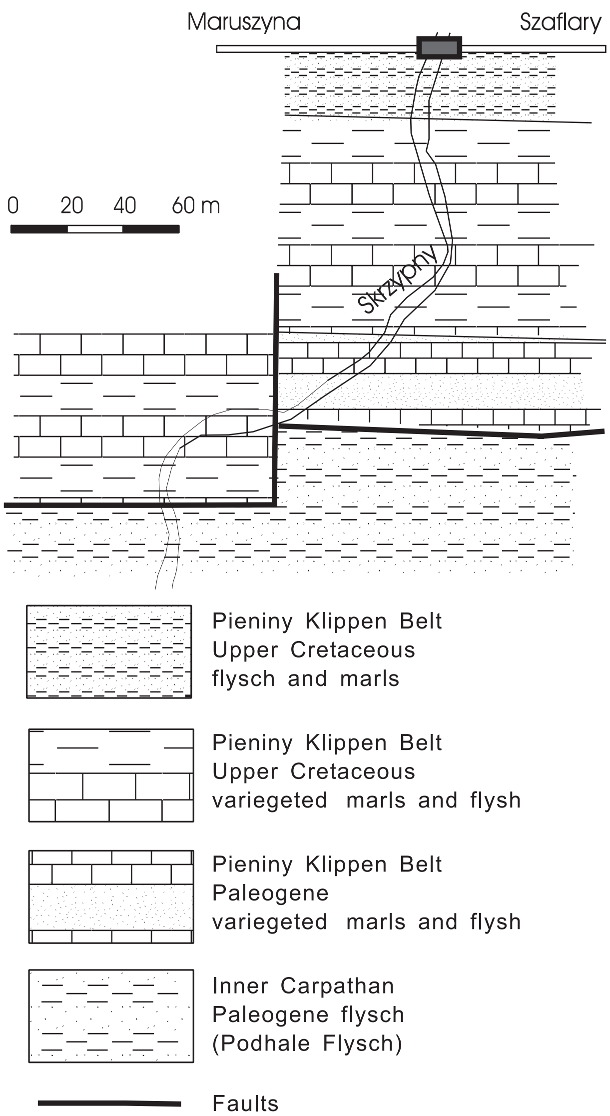
With the development of the Inner Carpathian nappes, forearc basin was formed between the uplifted part of the IC terrane (so-called Andrusov ridge) and the subduction zone. The flysch of the Klappe and Złatne successions was formed in this area. Behind the ridge, the Manin Succession was deposited within the back-arc basin. The Pieniny Klippen Belt Basin was closed at the Cretaceous/Tertiary transition as an effect of a collision between Czorsztyn ridge and European Platform. The accretionary prism had overridden the Czorsztyn Ridge. The subduction zone moved from the southern margin of the Pieniny Klippen Belt Ocean to the northern margin of the Czorsztyn Ridge (Golonka et al. 2005b, Oszczypko et al. 2005c and references therein).

The Pieniny Klippen Belt was finally formed as the melange in the suture zone between Inner Carpathian-Alpine (Alcapa) terrane and the North European plate. Part of allochthonous Outer Carpathian units and perhaps fragments of basement were also located in this suture zone. With the eastward movement of the Alcapa plate, system of strike-slip faults originated. Good visible effect of several tectonic phases of folding and deformations within Pieniny Klippen Belt is geomorphologic view of tectonically isolated klippes of Jurassic and Cretaceous hard rocks surrounded by softer shales, marls and flysch deposits. The last important event in the Pieniny Klippen Belt was **Middle Miocene (Sarmatian)** volcanism represented by calc-alkaline andesite dykes and sills which cut mainly Palaeogene flysch rocks of the Outer Carpathians (Magura nappe) and they formed so-called Pieniny Andesitic Line (PAL) (Birkenmajer 1986).

Stop 1. The Skrzypny Stream – Contact Between Pieniny Klippen Belt and Inner Carpathian Palaeogene

Jan GOLONKA

The Skrzypny Stream at the eastern border of Maruszyna village (fig. 5) displays limestones variegated shales, marls and flysch deposits of the Pieniny Klippen Belt succession (fig. 6) as well as sandstones and shales of the Inner Carpathian Palaeogene (fig. 7) (Podhale Flysch). The Maiolica-type cherty limestones of Pieniny Limestone Formation (Birkenmajer 1977) exposed below the bridge on Skrzypny Stream on the road between Szaflary and Maruszyna represent reduced latest Jurassic-Early Cretaceous sequence of Złatne Succession (see Golonka and Sikora 1981, Golonka et al. 2003, 2005b, Golonka and Krobicki 2004). Upstream in tectonic contact with the cherty limestones the series of Upper Cretaceous variegated shales, marls and flysch grey, grey-greenish and blue-greenish sandstones deposits is exposed (Fig. 6). These rocks belong to Złatne Succession sensu Golonka and Sikora (1981) or Maruszyna Succession (see Kostka 1993 and references therein). Southward the contact between Pieniny Klippen Belt and Inner Carpathian Palaeogene runs along the major strike slip fault. This fault was caused by rotation and movement of Inner Carpathian terrane (Golonka et al. 2003, 2005a, b) in relation with North European plate). The Pieniny Klippen Belt rocks were placed in the suture zone between two plates, displaying flower structure and tectonic mélange features. The Inner Carpathian Palaeogene



■ **Fig. 5.** Geological sketch of the Skrzypny Stream between Maruszyna and Szaflary. Podhale region, southern Poland.

■ **Fig. 6.** Deformed red marls of the Złatne Succession (Pieniny Klippen Belt). Skrzypny Stream, Podhale region, southern Poland.

■ **Fig. 7.** Deformed flysch rocks of the Szflary Beds (Inner Carpathian Palaeogene), near the strike-slip boundary with Pieniny Klippen Belt Skrzypny Stream, Podhale region, southern Poland.

flysch rock are only moderately deformed across in the central part of Podhale region, in the border zone, they are heavily deformed (fig. 7) due to the activity of the border strike-slip fault.

Passage: Skrzypne – Chabówka

Marek CIESZKOWSKI

From Skrzypne the route leads to Nowy Targ. It passes tectonic contact between the Pieniny Klippen Belt and the Magura Nappe, and goes northward to the Orava – Nowy Targ Depression. It is marked in morphology as a plane surrounded from the South by Podhale Upland and in Slovakia Oravska Magura range, and from the North by Babia Góra massif and Gorce range. The depression is situated at the border between the Inner and Outer Carpathians. Its origin has been caused by the collision of the Pannonian microplate (part of the Appulian plate) and North European platform, during its last stage after folding and thrusting of the Outer Carpathian nappes. The intermountain basin formed here is filled by the Late Miocene, Pliocene and Quaternary fresh water deposits. The highest summary thickness about one thousand meters deep been reached in the deep borehole Czarny Dunajec IG 1. Within the Miocene and Pliocene gravels, sands and clays are known occurrences of small seams of brown coal. On the plane, in the centre of the depression, several Quaternary peat bogs occur.

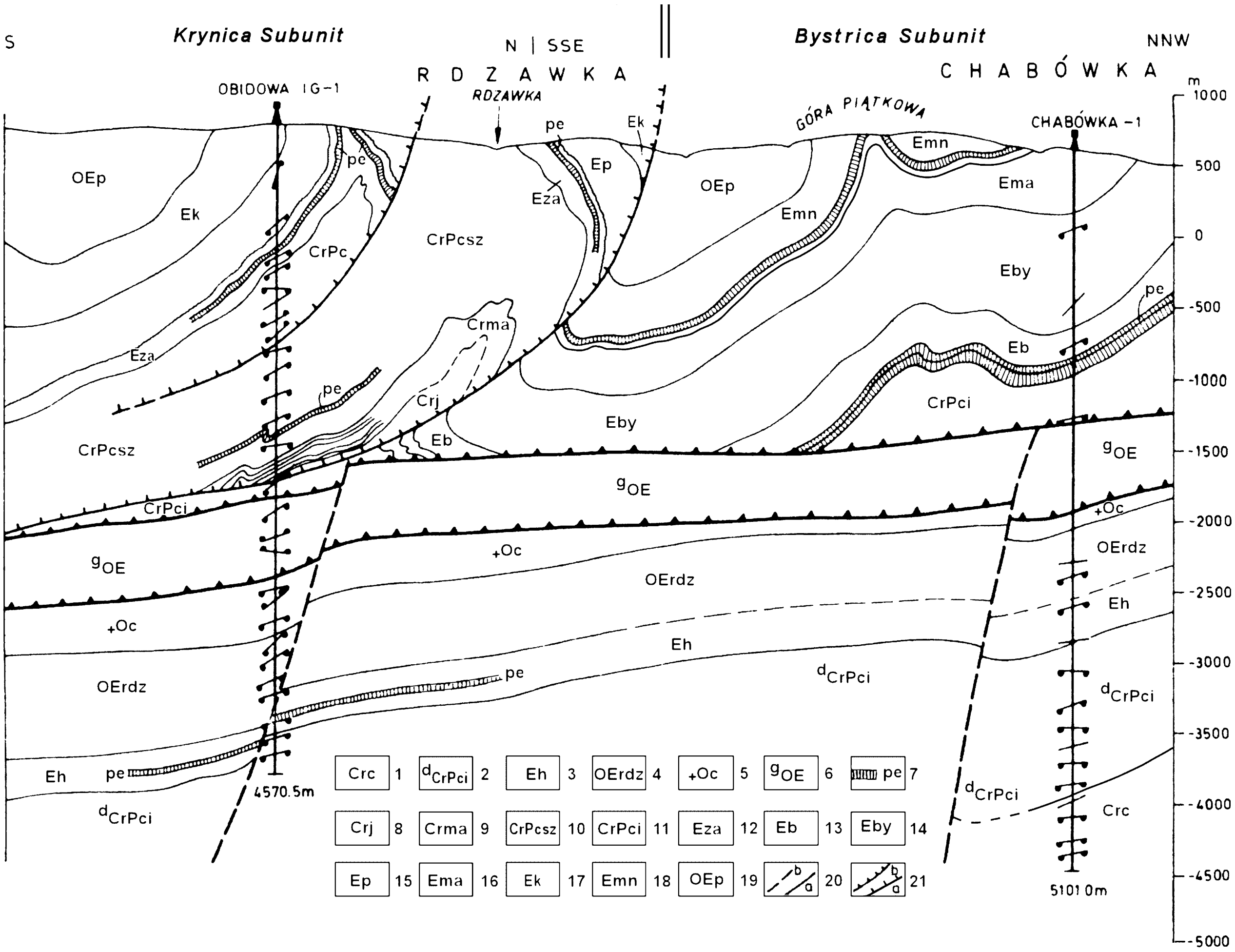
Between Nowy Targ and Rdzawka the route crosses the Krynica Subunit of the Magura Nappe. The Gorce mountains are well visible from the plane. There the soft morphology of their southern foothills at Nowy Targ and East of this town covered by fields and meadows, marks an occurrence of the thin-bedded shaly-sandstone flysch of the Malcov Formation (Cieszkowski and Olszewska 1986). Between Nowy Targ and Stare Bystre Miocene marine deposits, that overlain the Malcov Formatin has been discovered (Cieszkowski 1992, 1995). The higher parts of Gorce, covered by forests, are built of sandstones of the Magura Formation (Early Eocene–Early Oligocene). Following northwards, route is passing Klikuszowa village. There, on the right hand, is visible the quarry (Cieszkowski et al. 1998) with the outcrops of thick turbidites of the Poprad Sandstone Mb. (Late Eocene–Early Oligocene) which belongs to the Magura Formation. The rout follows uphill and at the top it passes in Rdzawka village where the Obidowa IG-1 deep borehole (see Fig. 3) is located. After about 3.5 km the route crosses the tectonic contact between the Krynica and Bystrica subunits. Here different lithostratigraphic divisions of the Krynica Subunit thrust over the Magura Formation of the Bystrica Subunit. Westwards of Rdzawka intersection of the thrust is well visible and easy for geological mapping, because there deposits of the Szczawnica and Zarzecze formations of the Krynica Subunit overthrust the Magura Fm. of the Bystrica Subunit. Westward is not so easy to fix the line of thrust, because the sandstones of the Magura Fm. of the Krynica Subunit contact directly with the same sandstones of the Bystrica Subunit. After about 1.5 km, on the area occupied by deposits of the Bystrica Subunit, near Chabówka village, is located stop no 2.

Stop 2. Chabówka: Structures of the Flysch Carpathians between Nowy Targ and Rabka

Marek CIESZKOWSKI

The stop 2 is located near the 17th century Holy Cross wooden church at Piątkowa Góra Mt. (715 m a.s.l.) close of Chabówka village (Cieszkowski 1985). The geological scenery of the surroundings includes three main tectonic units of the Magura Nappe: the Krynica Subunit on the South, Bystrica Subunit in the middle and Rača Subunit on the North. The formal lithostratigraphic inventory of the Magura Unit constitutes here mainly on divisions proposed by Oszczypko (Oszczypko and Birkenmajer 1989, Oszczypko 1991, Oszczypko et al. 2005b). The Krynica, Bystrica and Rača subunits have common older strata development as green spotted, partly radiolarian, Albian-Cenomanian shales (with manganiferous concretions) of the Jasień Formation, variegated Turonian-Early Senonian shales of the Malinowa Shale Formation, and the Late Senonian-Palaeocene deposits of the Inoceramian beds and similar to them the Szczawnica Formation. Differ lithostratigraphic development of Eocene and Oligocene deposits lets divide the subunits of the Magura Nappe. The Rača Subunit forms the highest mountains of Żywiec Beskid range – Babia Góra Mt. (1725 m asl) (the highest mount of the western sector of the West Outer Carpathians) and Polica Mt. (1369 m asl), as well as Beskid Wyspowy (Island Beskid) range with Zembalowa and Luboń Wielki (1113 m) in the east. It contains variegated shales of the Łabowa Formation (Early and Middle Eocene), the shaly-sandstone flysch of the Beloveža Formation – Hieroglyphic beds facies (Middle and Late Eocene) and thick-bedded "Magura sandstones" of the Magura Formation (Late Eocene-Early Oligocene). There inversion of morphology is well expressed by synclinal, weathering-resistant sandstones of the Magura Formation forming highest mountains, and by softer Upper Cretaceous-Middle Eocene sediments, included within anticlines, and forming valleys and passes.

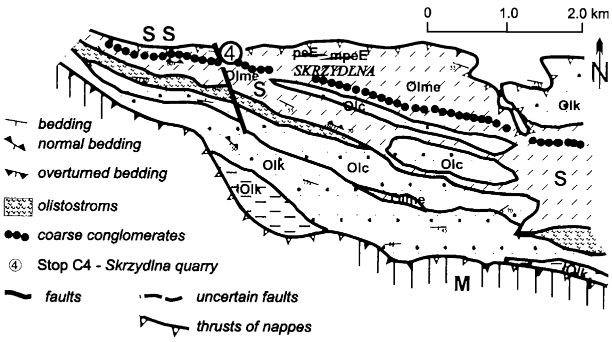
The Bystrica Subunit occupies a wide morphological Chabówka-Jordanów Depression and continues westward to the Orawa Depression. Eastwards it prolongs along the Raba River valley and northern slopes of the Gorce Range. Lithostratigraphic log of the Palaeogene deposits includes here: the Łabowa Shale Formation (Early Eocene), the Beloveza Formation (Early Eocene), the Żeleźnikowa and Bystrica formations (Middle or Middle-Early Eocene), both predominated by the Łącko marls lithotype, and Magura formation (Late Eocene-Early Oligocene). In the southern zone of the Bystrica Subunit the Magura sandstones occur earlier and the Magura Formation consists of three mem-



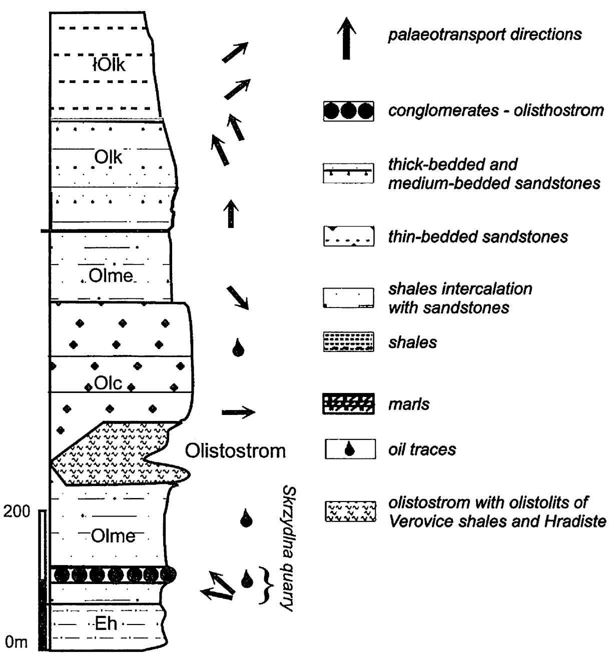
■ **Fig. 8.** Geological cross-section of the Outer Carpathian units between the deep drillings Obidowa IG-1 and Chabówka 1 (after Cieszkowski 1985, partly changed). Dukla Nappe, Obidowa-Słopnice Zone: 1 – Cisna andstones (Senonian), 2 – Inoceramian beds and Bukowiec Wielki sandstones, nondivided (Senonian-Palaeocene), 3 – Hieroglyphic beds (Late Palaeocene-Late Eocene), 4 – Rdzawaka beds (Late Eocene-Early Oligocene), 5 – Cergowa beds, shally facies (Early Oligocene); Grybów unit: 6 – nondivided Hieroglyphic, Menilitic (Grybów) and Cergowa beds (Middle Eocene-Early Oligocene); 7 – variegated shale intercalations in various Creaceous and Palaeogene strata; Magura Nappe, Krynica and Bystrica subunits: 8 – Jasień Formation (Cenomanian), 9 – Malinowa Shale Formation – variegated shales – (Turonian-Kampanian), 10 – Szczawnica Formation with Życzanów Sandstone Member (Kampanian-Palaeocene), 11 – Inoceramian beds (Late Senonian-Palaeocene), 12 – Zarzecze Formation, Krynica Sandstone Member (Early Eocene), 13 – Łabowa Shale Formation (pe) (Late Palaeocene-Early Eocene) and Beloveža Formation (Early Eocene), 14 – nondivided Bystrica and Żeleźnikowa formations (Middle Eocene), Magura Formation: 15 – Piwniczna Sandstone Member, 16 – Maszkowice Sadstone Member, 17 – Kowaniec Member, 18 – Mniszek Member, 19 – Poprad Member;Tectonic signatures: 20 – faults: a – proven, b – supposed; 21 – overthrusts: a – of nappes, b – of thrust-sheests.

bers; the Maszkowice Sandstone Mb., shaly-sandstone Beloveža-like flysch of Mniszek Member (Middle-Late Eocene) with level of variegated shales, and the Poprad Sandstone Member (Late Eocene–?Early Oligocene). Between Rabka and Spytkowice at the top of the Maszkowice Sandstone Mb. occurs olistostrome (Cieszkowski et al. 2003). There in the debris-flow consisting of coarse sandy and fine gravel material occur balls of fine-grain calcareous sandstone and olistholites of variegated shales or thin-bedded flysch.

The Krynica Subunit, forming the highest ranges of Gorce and Działy Orawskie ranges is characterized by extreme development of the Magura sandstone lithotypes. These sandstones appeared already in the Early Eocene and continued through the Middle and Late Eocene and Early Oligocene. North of Nowy Targ the Magura Fm. riches up to 2000 m in thickness. Above the Szczawnica Formation, which includes complex of thick bedded sandstones and conglomerates of Życzanów Mb., occurs the Zarzecze Formation (Early Eocene). Here, in the western section of the Gorce range, almost all thin- and medium-bedded flysch typical for this formation is replaced by thick bedded sandstones and conglomerates of the Krynica Sandstone Mb. The Magura Formation (Early Eocene–Early Oligocene) consists of three members: the Piwniczna Sandstone Mb., the Kowaniec beds (thick-bedded Magura sandstones with intercalations of the Hieroglyphic beds-like flysch and Łącko marls), and the Poprad Sandstone Mb., here more than 1000 m thick. The shale-sandstone Malcov Formation (Oligocene) and marly-sandstone Stare Bystre beds of Miocene age (Cieszkowski and Olszewska 1986; Cieszkowski 1992, 1995) overlie the Magura Formation. Both the Magura sandstone lithotypes occurs as intercalations.



■ **Fig. 9.** Geological scetch of the area of Skrzydlna (after Polak 2000). S – Silesian Nappe: *mpe*E – variegated marls (Eocene), *pe*E – variegated shales, Eh – Hieroglyphic beds – (Middle and Late Eocene); Menilite beds (Oligocene): Olme – Menilite shales, Olc – Cergowa sandstones; Krosno beds (Oligocene): Olk shales and sandstones (Early Oligocne), *ł*Olk – shales (Late Oligocene); SS – Subsilesian Nappe (Skrzydlna tectonic window): M – Magura Nappe; 4 – location of Skrzydlna Quarry.

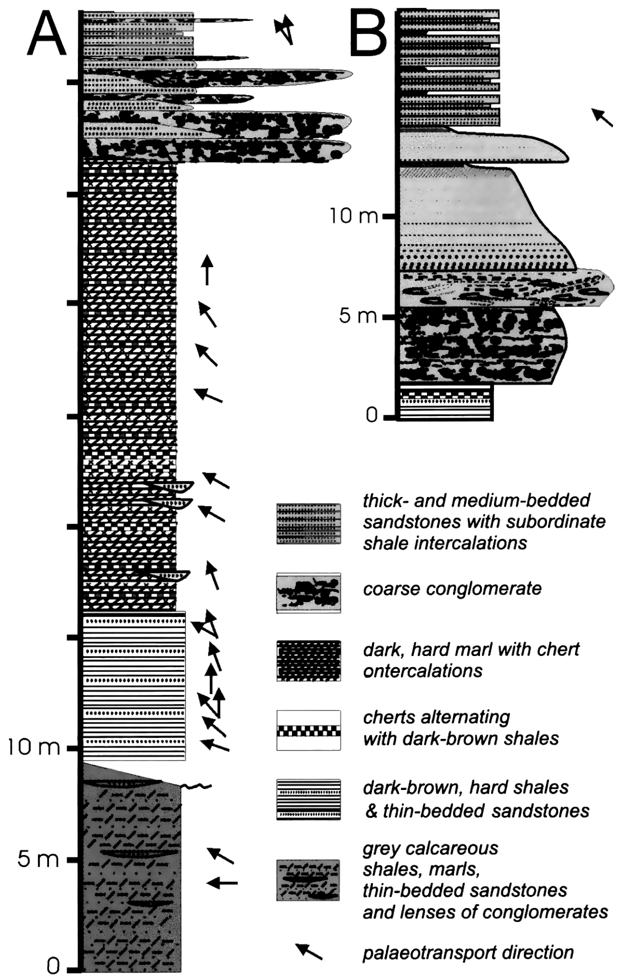


■ **Fig. 10.** Lithostratigraphic log of the sedimentary sequence of the Silesian Unit in the area of Skrzydlna (after Polak 2000). *pe*E – variegated shales, Eh – Hieroglyphic beds – (Middle and Late Eocene); Menilite beds (Oligocene): Olme – Menilite shales, Olc – Cergowa sandstones; Krosno beds (Oligocene): Olk – shales and sandstones (Early Oligocne), łOlk – shales (Late Oligocene)

Heavy mineral investigations, that have been done in the area of Nowy Targ, show the traces of chromian spinels within sandstones of the Magura Fm. (Cieszkowski at al. 1998), as well as traces of gold in the Krynica Mb. and in the Malcov Fm.

The deep structure of the Magura Nappe and its tectonic substratum have been recognized by deep boreholes Obidowa IG-1 and Chabówka 1 (Figs. 3, 4), as well as seismic and magnetotelluric data. In the boreholes, the base of the Magura Nappe has been found at depth from 1800 to 2500 m below the surface. Under the Magura there occur units of the Fore-Magura Zone (Cieszkowski 2001): strongly tectonically reduced Palaeogene deposits of the Grybów Unit and Dukla Nappe – Obidowa-Słopnice Subunit (Cieszkowski et al. 1981a, b; Cieszkowski 2001, see also Oszczypko-Clowes and Oszczypko 2004).

In cores of Hieroglyphic beds (Middle and Late Eocene), shally facies of Cergowa beds and Grybów shales (Oligocene) that represent the Grybów Unit has been identified. In the Obidowa-Słopnice Subunit the Inoceramian (Ropianka) beds (Late Senonian-Palaeocene) with medium to thick bedded sandstones resembling the thick bedded sandsones of the Cisna beds (Senonian) in the lower part and upwards of the section the Bukowiec Wielki sandstones (Palaeocene), the Hieroglyphic beds with horizon of va-



■ **Fig. 11.** Detailed lithologic log of the Menilite beds in: A – Skrzydlana Quarry, B – Stróża-Stróżkiewicze Creek (after Polak 2000).

riegated shales (Late Palaeocene-Late Eocene), the Rdzawka beds (Late Eocene-Early Oligocene) developed as "black Eocene" facies (thin bedded sandstones and black shales), conglomeratic sandstone and gravelly mudstones (Zboj sandstones) in the upper part and finally shally facies of the Cergowa beds (Early Oligocene) occur. Within the Grybow Nappe and Obidowa-Slopnice Subunit the traces of hydrocarbons, mainly of gas, has been recognized.

Passage: Chabówka-Skrzydlna

From Chabówka the route leads downhill to Rabka and at strong curve of road it passes location of the deep borehole Chabówka 1. Rabka is a famous spa for children, known of special microclimate and mineral salt water springs. Hydrothermal water (28 °C) has been also discovered here. In the center of town, there is the wooden church of the XVIII Century, recently changed to the museum of local, mountaineer folk, manufacture and art. Leaving town of Rabka the route crosses the Bystrica subunit of the Magura Nappe and after passing Zaryte hamlet and Raba Niżna village it arrives to a small town of Mszana Dolna. In the area of Mszana Dolna, gentle morphology of hills, covered with fields and meadows, marks the area of Mszana Dolna tectonic window. In this window predominantly Oligocenen strata of the Fore-Magura group of Nappes (Grybów and Dukla nappes) crop out. Higher morphology of the Magura Nappe that builds Beski Wyspowy and Gorce mountain ranges surrounding the window is covered by forests. North-East of Mszana Dolna the rout leaves the window and goes by Kasina village to Skrzydlna (cf. Polak 1999). On this distance, we cross the northern-most part of the Magura Nappe, called Siary Subunit. There above the Łabowa Shale Formations occur the Zembrzyce (Sub-Magura) and Budzów (Supra-Magura) beds (Late Eocene-Early Oligocene) developed as shaly- or marly-sandstone, thick- or very thick-bedded turbidites. A complex of thick-bedded, glauconitic Wątkowa sandstones (called previously glauconitic facies of the Magura sandstones) separates both divisions. Going downhill from Kasina the route passes frontal thrust-fault of the Magura Nappe. The Magura Nappe thrust here over the Silesian Nappe (Figs. 2, 3, 8). On the distance of one kilometer, we pass the Oligocene Krosno beds and Menilite beds (Cergowa sandstones) of the Skrzydlna thrust-sheet, that create here the southern limb of so-called Lanckorona-Żegocina structural zone, including numerous of tectonic windows in its axial part. There, in the windows, deposits of the Subsilesian Unit arrive on the surface from below of the Silesian Nappe. After about 1 km the route comes to a small quarry situated at the road in southern periphery of Skrzydlna village (Figs. 9, 10, 11). In this quarry the Oligocene deposits of the Menilite beds crop out.

Stop 3. Skrzydlna: Conglomerates and Olistholites in the Menilite Beds of the Silesian Nappe in the Skrzydlna Quarry

Marek CIESZKOWSKI

The lithological section of the Menilite beds in Skrzydlna Quarry (Polak 2000, Cieszkowski and Polak 2001) (Fig. 11) begin dark grey, marly shales intercalated by grey marls. Above are developed carbonateless, dark brown, silicified, bituminous shales, partly with black cherts, those present very characteristic deposits of the Menilite beds. The layers of fine grained, thin-bedded silicified sandstones interbed those marly and shaly deposits. Occasionally, the lenses of coarse sandstones or poorly cemented conglomerates occur there. The sandstones and conglomerates are composed of quartz, feldspars, muscovite, as well as clasts of green clayey shales, and the Carboniferous coal. A few sandstone clastic dykes cut vertically the layers of shales. The solemarks show that the clastic material of the sandstones and conglomerates have been derived from S and SE.

Above arrives 22 meters thick complex of thin and medium bedded, hard, silicified marls with cherts. The marls are dark grey, brownish, light when weathered, usually massive or with parallel lamination. Some marly layers have got small-scale solemarks showing paleotransport direction from S and SE. The lenses of fine conglomerates, composed of quartz, feldspar and, clasts of carbonate rocks and coal, occur in the lower part of the marly complex. A few thin clastic dykes cut vertically the layers of marls. Within the brown shales and marls of the Menilite beds, fossilised fish remnants, mainly scales, occasionally occur. Here, very rare are complete fish skeletons and once finding of reptile remnants was happened.

Marls and cherts pass rapidly to the thick bedded, coarse conglomerates and sandstones. The conglomerates consist of more or less rounded pebbles or angular blocks and clasts of various rocks. Their dimensions usually oscillate from 5 to 15 centimetres but larger pebbles and clasts have often appear, even, occasionally above one and half of a meter or even more. Up to the top of a layer conglomerates pass to the coarse-grained sandstones. The layer boundaries are erosive with a set of channels.

The conglomerates consist mainly of pebbles and clasts of sedimentary rocks: limestones (62 %), cherts (2.5 %), quartz granules and quartzites (7 %), sandstones (15.5 %), conglomerates (3.5 %) and others, mainly detrital rocks (7.5 %), and only small amount (2 %) of crystalline rocks, dominated by gneisses and granitogneisses. Various sedimentary rocks of different age have been distinguished there, e.g.: Devonian limestones with stromatoporoids e.g. *Stromatopora* sp., *Actinostroma* sp., *Stachyodes* sp. and *Amphipora ramosa*, Givetian – Franian limestones rich of foraminifera, black Carboniferous limestones (wackstone) with brachiopods *Productus* and *Leptena* and occasional trylobites; Carboniferous conglomerates; dark craned limestones; light, Stramberk-like Jurassic limestones with corals, molluscs, onkoides, peloides, sponge spicules, radiolarians, etc.; Eocene nummulitic limestones with occasional molluscs and crinoid remnants, as well as corals, briozoans and algae; the Palaeogene allodapic limestones with T*abcd* Bouma intervals containing molluscs shell remnants; problematic in age light micritic limestones, brownish marls, glauconitic T*abcd* sandstone turbidites, cherts, mudstones and shales. In the area surrounding Skrzydlna village vein quartz pebble with gold mineralization has been found.

The coarse conglomerates represent channel facies. They have been derived from the remnants of the Silesian Ridge to the Silesian Basin by high-density flows. The pebbles represent partly the rocks from the crystalline basement of the ridge core, and its Meso-Paleozoic sedimentary cover. Part of them represents also the Upper Cretaceous and Palaeogene sediments that have been deposited on the narrow shelf surrounding the ridge and on its northern slope.

South of the quarry the pebbles and sandstones pass up to the complex of brown shales with occasional cherts and next to the Cergowa beds. The Cergowa beds are typical for the Dukla unit, but occur also in the inner zone of the Silesian nappe. They replace part of the chert horizon within the Menilite beds. In the Skrzydlna thrust-sheet the Cergowa beds are developed as thick bedded, massive sandstones. Above the Cergowa beds thin complex of the brown Menilite shales arrives occasionally once more, and the Menilite beds pass up the section to the light grey, sandstone-shaly flysch of the Krosno beds.

Within the Skrzydlna thrust-sheet, at the basal part of the Cergowa sandstones and close of the lithostratigraphic boundary of the Menilite and Krosno Beds, the large olisthostroms occur (Figs. 10, 11). These olisthostroms consist mainly of the Lower Cretaceous flysch deposits of the Upper Cieszyn, Hradiste and Verovice beds. Minor frequency of olistholites of the Upper Cretaceous flysch deposits as well as Eocene non-turbiditic, grey, greenish or red marls and shales have been also noticed there. The occurrence of the olisthostroms within Oligocene sequence of deposits are known in Poland from the inner zone of the Silesian Nappe east of the Dunajec called Fore-Dukla zone, however especially well developed are in the Bitla zone in Ukraine.

During the Early Oligocene tectonic movements in the Outer Carpathian basinal area, the Silesian Ridge has been partly collapsed. It happened when the plate of the Silesian Basin's basement underthrusted southerly situated Silesian Ridge, and caused origin of a local accretionary prism. Then, the Palaeogene deposits of the slope, as well as the older Cretaceous flysch deposits, partly folded and thrusted within the prism, have been uplifted. In consequence, some of them sliced northward to the basin, forming the olisthostromes.

From the Skrzydlna quarry the route leads northward and after passing the tectonic border (thrust) between the Silesian Nappe and Subsilesian Unit occurring here in tectonic window (Fig. 10), it arrives to Skrzydlna village. The village is situated on the area of the Skrzydlna tectonic window. There, in the center, is an old beautiful wooden church of XVI century with the crucifix of XIV century. From Skrzydlna the route come back along the same way to Mszana Dolna.

Stop 4. Poręba Górna – Southern Margin of the Mszana Dolna Tectonic Window

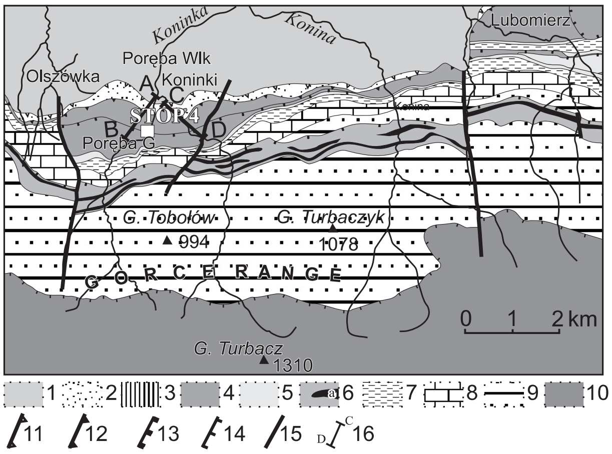
Nestor OSZCZYPKO, Marta OSZCZYPKO-CLOWES and Dorota SALATA

Description

The characteristic feature of the middle part of the Magura Nappe in the Polish Outer Carpathians is an occurrence of the Mszana Dolna tectonic window (MDW). The central and most uplifted part of this window is dominated by the Oligocene Krosno Formation of the Dukla Unit, whereas the narrow, marginal part of it is occupied by the Cretaceous-Oligocene deposits of the Grybów Unit (Fig. 4). The youngest deposits of the Mszana Dolna tectonic window belong to NP 24 and NP22 calcareous nannoplankton zones for the Dukla and Grybów units respectively (Oszczypko-Clowes and Oszczypko 2004).

The Mszana Dolna tectonic window is the big duplex structure, which developed during the Middle Miocene thrusting of the Magura Nappe against its foreland. The floor thrust developed along the frontal ramp formed by the Dukla Unit, whereas the roof thrust is related to Magura Nappe. Between the floor and roof thrusts the imbricated horses of the Grybów Unit developed. The southern margin of MDW is build up of the Cretaceous-Palaeogene deposits, which could be correlated with the Bystrica Subunit (Fig. 2*)*, however the occurrence of fragments of the Rača Subunit in basal Magura thrust (Poręba Wielka-Koninki thrust-sheet) could not be excluded.

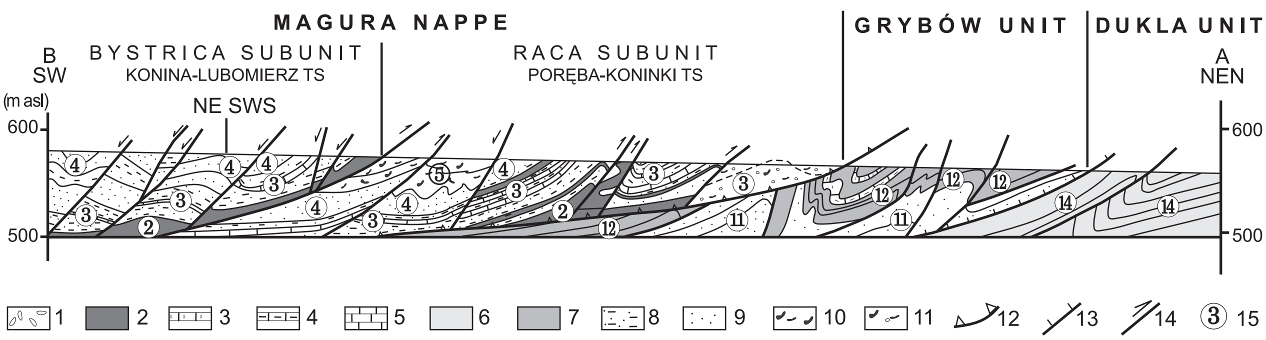
The section is exposed along the bedrock of Poręba Górna Stream on the southern margin of the Mszana Dolna tectonic window (Figs. 1, 4, 12 see also Oszczypko et al. 1999b). The section is nearly 400 m long and displays the contact zone between the Grybów and Magura units and structure of the basal part of the Magura Nappe (Figs. 12, 13, 14). The exposed succession can be subdivided in-to 4 minor units differing in structure. The incomplete unit 1st belongs to the Oligocene Grybów beds of the Grybów Unit. The section begins with the SW dipping, massive muscovite sandstone of the Cergowa type, passing upwards into black shales and mudstones with intercalations of thin-bedded sandstones and siderites nodules. These beds are incorporated into three strongly deformed, outcrop scale thrust-sheets. These thrust-sheets display numerous NWN-SES to N-S trending mesoscopic, subvertical thrust-fault propagating folds. The unit 2 begins with SW dipping sole trust of the Magura Nappe (Poręba-Koninki thrust-sheet). The inclination of the thrust surface is around 30°. The basal portion of the Magura nappe is represented by the 3 m packet of the medium-bedded sandstones and grey-greenish shales of the Senonian



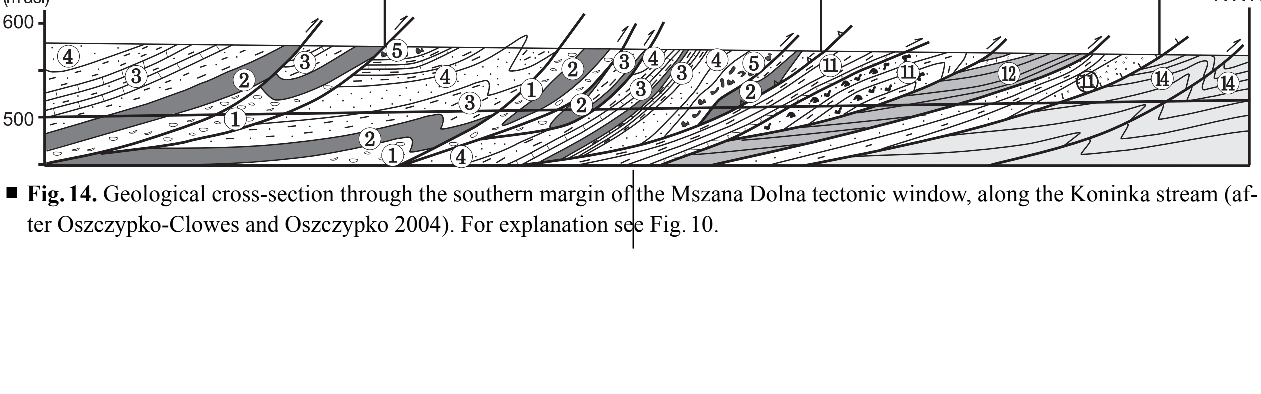
■ **Fig. 12.** Geological map of the Magura Nappe on the southern margin of Mszana Dolna tectonic window (after Oszczypko et al., 1999b). 1 – Oligocene Krosno beds of Dukla Unit, 2-Grybów Unit, (3-9) Magura Nappe – Bystrica Subunit: 3 – Albian-Cenomanian deposits, 4 – Cenomanian-Palaeocene; Eocene: 5 – Łabowa Fm., 6 – Zarzecze Fm., a-variegated shales, 7 – Beloveza Fm., 8 – Bystrica and Żeleźnikowa fms., 9 – Maszkowice Mb. of the Magura Fm., 10 – Krynica Subunit, 11 – Grybów overthrust, 12 – Magura overthrust, 13 – Bystrica Subunit internal overthrusts, 14 – Krynica overthrust, 15 – faults, 16 – cross-sections.

age. This stratified unit is covered by 40–50 metres complex of chaotic deposits, known as the Poręba Wielka beds (see Burtan et al. 1978). The stratified and chaotic units contact along the south dipping inverse fault. The chaotic deposits display fragments of blue-greyish, medium-grained non calcareous sandstones of different size (from 1 cm to 1.5 m boulders) and shape, dispersed in the green-greyish and dark-greyish, non-calcareous, clay-claystone material. Among these sandstone fragments, the small lumps of drag-folds have been observed. The sandstone fragments showing primary fractures often with calcite mineralization. The sandstone clasts and shales sometimes revealed the remnants of primary stratification. The chaotic deposits occur in the layers of few to 50 centimeters thick, with two-type of boundaries; fluidal-and shear fracture plane boundaries. Calcite veins often accompany the shear plane-type boundary. Both type of boundaries are gently dipping and are more or less parallel to each other, towards the NE and SW in the basal and top part of chaotic body respectively. These subhorizontal planes are occasionally cut by W-E trending subvertical, south dipping inverse faults with calcite mineralization. The lower strongly chaotic part is covered by the upper less chaotic part with more frequent primary type of stratification. Towards the top of this unit random dispersed sandstones fragments are progressively replaced by the boudine-like fragments.

The chaotic unit (Poręba Wielka beds) is tectonically followed upward by the 50 m thick unit 3, which is characterized by oc-



■ **Fig. 13.** Geological cross-section through the southern margin of the Mszana Dolna tectonic window, along the Poręba Górna stream (after Oszczypko-Clowes and Oszczypko 2004). 1 – spotty shales, 2 – variegated shales, 3 – spherosiderites, 4 – marls, 5 – turbidite limestones, 6 – calcareous shally flysch facies, 7 – black marly sheles, 8 – thin to medium-bedded turbidites, 9 – thick-bedded sandstones, 10 – submarine slumps, 11 – chaotic deposits, 12 – Magura overthrust, 13 – Grybow thrust, 14 – fault, 15 – lithostratigraphic units: 1 – Hulina Formation, 2 – Malinowa Shale Formation and Hałuszowa Formation, 3 – Kanina Beds, 4 – Szczawina Ss., 5 – Ropianka Beds, 6 – Łabowa Shale Formation, 7 – Beloveža Formation, 8 – Bystrica Formation, 9 – Żeleźnikowa Formation, 10 – Maszkowice Member of the Magura Formation, 11 – Jaworzynka Beds, 12 – Grybów Beds, 13 – Cergowa Beds, 14 – Krosno Beds.



■ **Fig. 14.** Geological cross-section through the southern margin of the Mszana Dolna tectonic window, along the Koninka stream (after Oszczypko-Clowes and Oszczypko 2004). For explanation see Fig. 10.

currence of outcrop scale NWN-SES trending recumbed and imbricated folds of the Malinowa Fm. (Turonian-Santonian) and Białe Formation (Campanian). This unit passed into steep south-west dipping thin-bedded flysch of the Białe Fm. and the thick-bedded Szczawina Sandstone Formation (Maastrichtian/ Palaeocene, see Oszczypko et al. 2005b). These sandstones showing brittle type of deformation with numerous small-scale W-E and WNW-ESE trending, and S and SWS dipping inverse faults.

In the Szczawina Formation occur heavy mineral assemblages composed mainly of very stable minerals such as zircon (12–34 %), tourmaline (29–59 %, represented by schorl-dravite series) and rutile (10–24 %). They also contain apatite (1–17 %), besides in some samples higher amounts of garnet (about 20 %, represented by almandine) and small amounts of brookite were found. The heavy fraction of Białe Fm in the southern part of the Bystrica subunit consisted mainly of zircon population (about 65 %) and smaller amounts of tourmaline (about 18 %), rutile (about 8 %) and single grains of garnet. In sandstones of the Białe formation grains of chromian spinel reaching amount of 8 % were also found (Oszczypko and Salata 2005).

Interpretation. The chaotic deposits from the Poręba Górna section were interpreted both in sedimentary (submarine slump, Książkiewicz 1958, see also Cieszkowski et al. 1986, Mastella 1988) as well as tectonic (metamorphic tectonites or "wildflysch", see Burtan and Łydka 1978, Burtan et al. 1978) terms. According to Burtan and Łydka (1978) and Mastella (1988) the temperature during overthrusting reached 300–350 °C. In our opinion, the chaotic deposits revealed the direct relation to the Magura Nappe sole thrust. Its is supports by observation of degree of the tectonic deformation, which drastically decreased outside of the Magura thrust, both in the Magura as well as the Grybów successions. The thrusting of the Magura Nappe onto the Fore-Magura (Grybów and Dukla) begun to develop during the Late Oligocene. This process was probably simultaneous with deposition of the Zawada Fm in the Magura piggy-back basin (Oszczypko et al. 1999a, Oszczypko and Oszczypko-Clowes 2002, 2003). The last act of the overthrusting probably took place during the post-Middle Badenian time (see Oszczypko 1998). It means that overthrusting lasted 9–10 Ma. This process was probably initiated in the submarine condition, when the front of the Magura Nappe reached the Dukla-Grybow subbasin. As a result the Magura Nappe at least 2.5–3 km thick, build up of compacted and impermeable rocks loaded and sealed under the compacted, clayey-sandy deposits of the Menilite-Krosno Fm. of Grybów succession. It build up overpressured zone on the contact of the Magura Nappe and Grybów succession (see also Mastella 1988), which was affected by the fracturing and frictional sliding. The process of deformation in the Magura/Grybów thrust zone was continued during the Middle Miocene thrusting and development of the Mszana Dolna duplex structure (Oszczypko-Clowes and Oszczypko 2004).

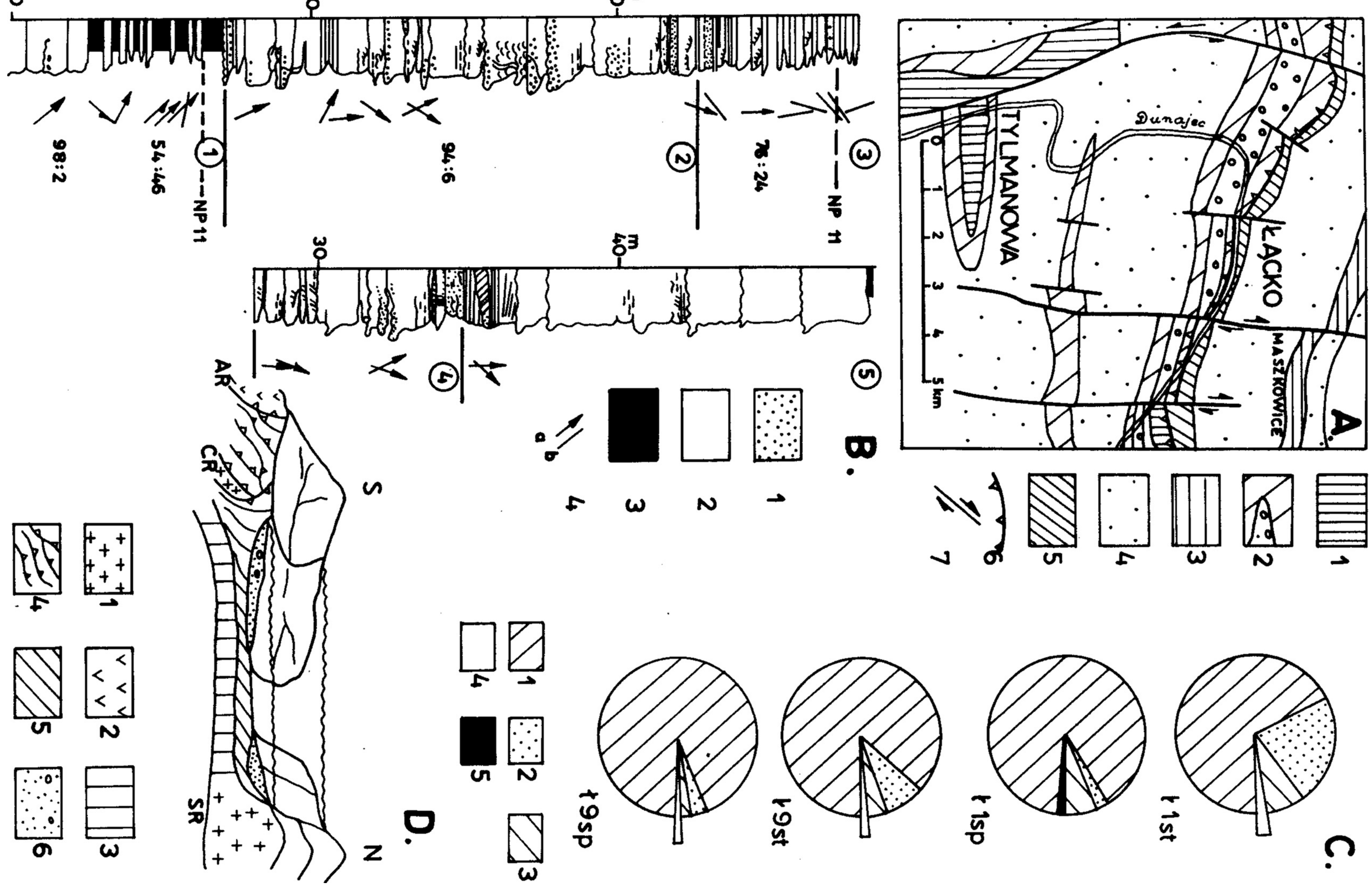
Stop 5. Tylmanowa–Piwniczna Member of the Magura Formation

Nestor OSZCZYPKO

The section is exposed along the highway linking town of Stary Sącz and Krościenko (Figs. 4, 15A, see Oszczypko and Porębski 1985, Oszczypko et al. 1992). The section is nearly 50 m thick and consists of medium to thick-bedded, dominantly medium-grained sandstones with two intercalations of thin-bedded turbidites. This sequence belongs to the Piwniczna Sandstone Member (Lower-Middle Eocene) of the Magura Formation. These sandstones are followed upward by shales with numerous, thin, medium to fine turbidites displaying Bouma's Tbc and Tc divisions. The exposed succession can be subdivided in 5 minor sequences differing in facies assemblages and vertical textural trends (Fig. 15B). The incomplete basal sequence l begins with a few metres of massive, medium-grained sandstones

Interpretation. The exposed succession can possibly be interpreted in terms of middle-fan distributary channel-depositional lobe systems, supplied from SSE. In this context, the sequence 2 may represent a channel fill. At the base of this sequence, there occurs a 1.5 m thick zone of disturbed bedding involving imbricated sandstone sheets and overturned folds. These structures suggest slumping of semi-consolidated sediment toward the E, i.e. perpendicularly to the NNW-directed palaeocurrents.

The overlying sequence 3 may record a gradual abandonment of the channel due to widening of its cross-section with successive filling episode, which increased overbank spilling of channelised turbidity currents. The coarsening-up sequence 4 may have originated through development of a depositional lobe which cou1d possibly follow a topographic low left after the filling of the underlying channel. The progradation of the lobe was succeeded by the incision and filling of a feeding channel, a process envisaged for the sequence 5. The large-scale traction structures preserved at the base of this sequence, may record an open channel (up to 5 m long and 0.7 m deep) flow conditions and sediment by-passing before the channel became plunged by a thick flow.

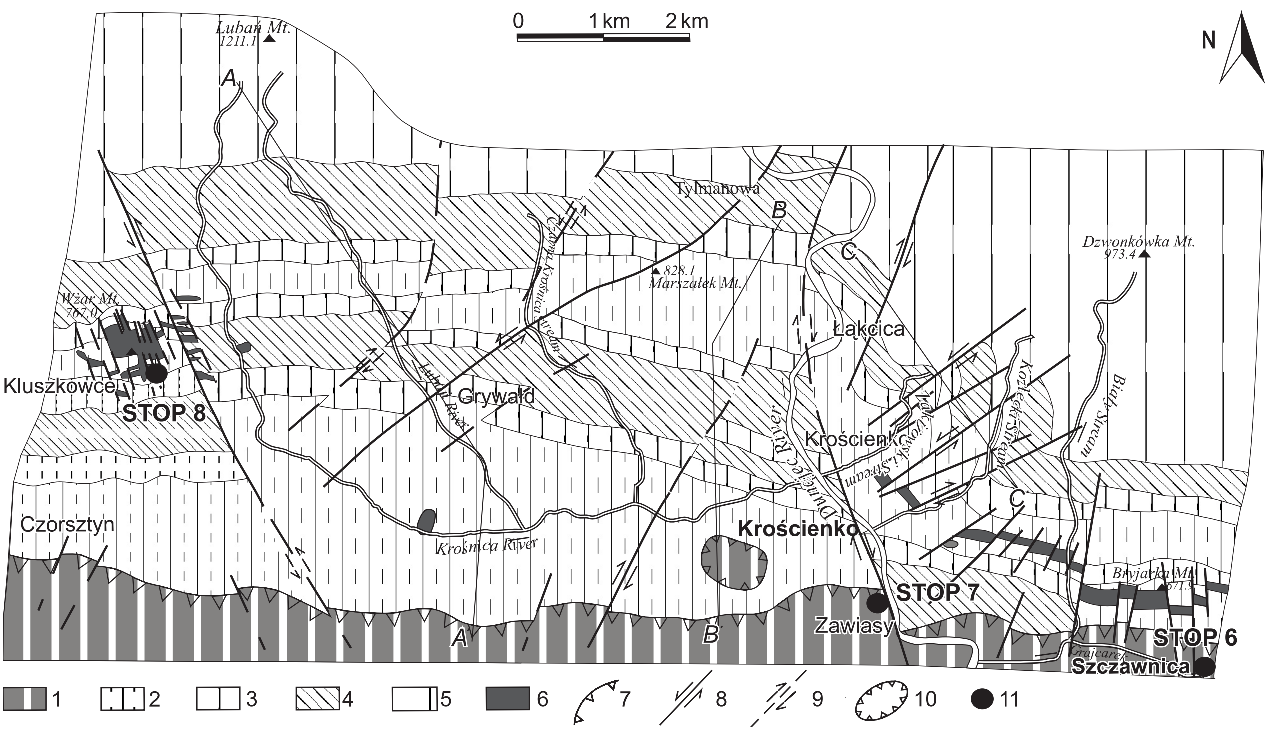


■ **Fig. 15.** Geology of the Dunajec river valley between Tylmanowa and Łącko (after Oszczypko et al. 1992). **A** – geological map: 1 – Szczawnica Formation, 2 – Zarzecze Formation, a – Krynica Sandstone Member, 3 – Żeleźnikowa Formation, 4 – Magura Formation, 5 – Mniszek Shale Member; 6 – thrust of Krynica zone upon Bystrica zone, 7 – faults; **B** – Sedimentological log of the Piwniczna Sandstone Member at Tylmanowa (based on Oszczypko and Porębski 1985): 1 – conglomeratic sandstones and conglomerates, 2 – sandstones, 3 – mudstones and clayey shales, 4 – attitude of sole marks: a – flute casts, b – drag groove casts; 94:6 – sandstone:shale ratio. Numbers refer to sequences described in the text by N. Oszczypko and S. Porębski; **C** – composition of heavy minerals: 1 – garnet, 2 – turmaline, 3 – rutile, bruckite, anatase, 4 – zircon, 5 – staurolith; ł1,ł9-bed nymbers;st- top, sp- base; **D** – depositional model of the Magura basin during early Eocene times:1 – cotinental crust of the Silesian cordillera (SR) and Czorsztyn Ridge (CR), 2 – exotic ridge of Andrusov (AR), 3 – oceanic or transitional crust, 4 – Upper Cretaceous units of the Pieniny Klippen Belt, 5 – pelagic and hemipelagic deposits, 6- thick-bedded turbidites. (Birkenmajer and Oszczypko 1989, Oszczypko and Porębski 1985, Oszczypko et al. 1992).

Stop 6. Szczawnica Wyżna (Zabaniszcze)–Grajcarek Unit

Jan GOLONKA, Michał KROBICKI, Nestor OSZCZYPKO and Dorota SALATA

On the left slope of Grajcarek Stream in Szczawnica-Zabaniszcze (Szczawnica Wyżnia) the Upper Jurassic – Lower Cretaceous sequence of the Magura Succession of the Grajcarek Unit occurs (Fig. 16). The deposits are represented by siliceous-carbonate, strongly sedimentary condensed sequence, which whole thickness does not exceed 14 m. The oldest deposits are grey, manganiferous thin-bedded radiolarites (3 m thick) of the Sokolica Radiolarite Formation (Birkenmajer 1977), dated by radiolarian assemblages as Oxfordian (Unitary Associations 7–8) (Widz 1991, 1992). These are overlain by green and red radiolarites of the Czajakowa Radiolarite Formation (Podmajerz Radiolarite Member and Buwałd Radiolarite Member – 8 m and 0.6 m in thickness, respectively) of which radiolarians indicative of the Upper Oxfordian and Kimmeridgian (Unitary Association 8–9) have been recognized (Widz 1991, 1992). These radiolarite formations are succeeded by hard, cherry-red marls of the Palenica Marl Member of the Czorsztyn Limestone Formation with a maximum thickness of 0.5 m. Aptychi may be locally abundant and suggest Kimmeridgian age of these deposits (lower part of aptychus subzone VI_1 of Gąsiorowski – see Birkenmajer 1977), but ammonites are absent. On the other hand, stomiosphaerids indicate Early Tithonian age (Malmica Zone after Nowak 1971, 1976). These marls are both facies and stratigraphic equivalents to the South Alpine Rosso ad Aptychi limestones/marlstones. In the Polish Carpathians they occur exclusively within Magura Succession of the Grajcarek Unit (Birkenmajer 1977), and are known only from the northern boundary of the Pieniny Klippen Belt between Szczawnica spa and Szlachtowa village. The sequence grades up to well-bedded light-grayish cherty limestones of the Pieniny Limestone Formation, which correspond to widespread Tethyan Maiolica (=Biancone) facies. A small thickness of these lime-



■ **Fig. 15.** Geology of the Dunajec river valley between Tylmanowa and Łącko (after Oszczypko et al. 1992). **A** – geological map: 1 – Szczawnica Formation, 2 – Zarzecze Formation, a – Krynica Sandstone Member, 3 – Żeleźnikowa Formation, 4 – Magura Formation, 5 – Mniszek Shale Member; 6 – thrust of Krynica zone upon Bystrica zone, 7 – faults; **B** – Sedimentological log of the Piwniczna Sandstone Member at Tylmanowa (based on Oszczypko and Porębski 1985): 1 – conglomeratic sandstones and conglomerates, 2 – sandstones, 3 – mudstones and clayey shales, 4 – attitude of sole marks: a – flute casts, b – drag groove casts; 94:6 – sandstone: shale ratio. Numbers refer to sequences described in the text by N. Oszczypko and S. Porębski; **C** – composition of heavy minerals: 1 – garnet, 2 – turmaline, 3 – rutile, bruckite, anatase, 4 – zircon, 5 – staurolith; ł1,ł9-bed nymbers;st- top, sp- base; **D** – depositional model of the Magura basin during early Eocene times:1 – cotinental crust of the Silesian cordillera (SR) and Czorsztyn Ridge (CR), 2 – exotic ridge of Andrusov (AR), 3 – oceanic or transitional crust, 4 – Upper Cretaceous units of the Pieniny Klippen Belt, 5 – pelagic and hemipelagic deposits, 6- thick-bedded turbidites. (Birkenmajer and Oszczypko 1989, Oszczypko and Porębski 1985, Oszczypko et al. 1992).

stones (only 1.5–2.0 m) is effect of strong condensation: several stomiosphaerid and calpionellide biozones have been recognized here (Obermajer, 1986), which cover stratigraphical interval from Upper Kimmeridgian to Upper Hauterivian (and may be even Barremian). The uppermost Jurassic/lowermost Cretaceous sequence shows highest condensation effect, which is known within Carpathian basins near Jurassic/Cretaceous boundary, and that's why the discussed succession could indicate the most axial, basinal zone of the Pieniny-Magura Ocean (see – Introduction) (Birkenmajer 1977, Golonka and Sikora, 1981, Golonka et al. 2003, Golonka and Krobicki 2004).

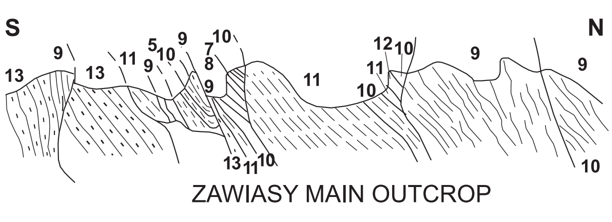
The Biancone type limestones pass upward (see Birkenmajer 1977) into the spotty marls and limestones of the Kapuśnica Formation (Aptian-Albian) and black-green shales of the Wronine Formation (Albian). In the western part of the section, the Jurassic-Lower Cretaceous klippen is thrust over the red shales of the Malinowa Formation (Cenomanian-Maastrichtian) and thick-bedded sandstones and conglomerates of the Jarmuta Formation (Maastrichtian-Palaeocene). Sandstones of the Jarmuta Formation outcropping in Szczawnica Zabaniszcze in heavy mineral set contain 20–48 % of zircon, 25–42 % of tourmaline (representing schörl-dravite series), 7–15 % of rutile, 1–23 % of garnet (with prevailing almandine molecule), 2–9 % of chromian spinel and 2–6 % of apatite. Chemical composition of tourmalines and garnets indicates their provenance from metamorphic rocks. The presence of chromian spinels and their composition suggests existence of ocean floor peridotites or rocks of an ophiolite sequence in the source area (see Oszczypko and Salata, 2005).

Stop 7. Krościenko-Zawiasy Klippe-Contact Zone between the Pieniny Klippen Belt and the Magura Nappe

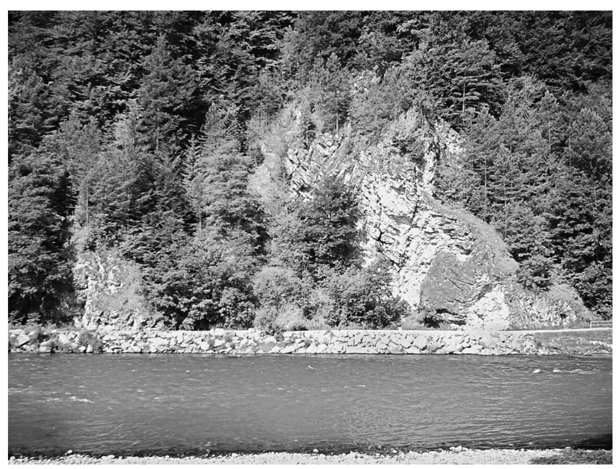
Jan GOLONKA, Michał KROBICKI, Nestor OSZCZYPKO and Dorota SALATA

On the western bank of Dunajec River across the in the Zawiasy Klippe is visible (Figs. 16, 17, 18). All rocks exposed here represent the fragment of the Magura Basin incorporated into the Pieniny Klippen Belt. They were deposited between the Czorsztyn Ridge and deeper part of the Magura Basin during Middle Jurassic through Palaeocene. The main ridge of the Pieniny Mountains built of Czertezik Succession is located South of Zawiasy Klippe. The Zawiasy profile is similar to that of the Branisko Succession (Birkenmajer 1977). The Branisko Succession is located however south of the Czorsztyn Ridge. Perhaps the separate Zawiasy Succession representing the slope deposits of the Magura Basin should be distinguished.

The main bulk of the Klippe is formed by the Biancone/Maiolica type cherty limestones (Pieniny Limestone Formation) – of latest Jurassic–Early Cretaceous age. The *Hedbergella* microfacies with *Hedbergella* sp., *Praeglobotruncana* sp and *Tahalmaninella ticenensis* Gandolfi) was found in the uppermost part of the limestones (Golonka and Sikora 1981). It suggests the Albian age of the youngest part of the Pieniny Limestone Formation in the slope succession of the Magura Basin. The well developed sedimentary breccias could be also observed in the cherty limestones. They were formed as a result of extensive gravitational faulting, which took especially during and Neo-Cimmerian (Upper Jurassic-Lower Cretaceous) movements. Similar breccias are known from the Czorsztyn Succession (Birkenmajer 1986, Golonka and Krobicki 2001). The exact age of the breccias in the Zawiasy area is currently under investigation. Presumed initial stages of subduction of the oceanic crust of the Pieniny Klippen Belt Basin riftogenesis volcanic activity and even paleooceanographical conditions (Birkenmajer 1986, Golonka and Krobicki 2001) are most probably connected with Neo-Cimmerian tectonic event. Alternatively, the formation of such allodapic rock beds are also interpreted as an effect of eu-



■ **xFig. 17.** Cross-section of Zawiasy main outcrop (modified from Golonka and Sikora, 1981). 5 – green Globotruncana marls – Jaworki Formation. 7 – spherosideritic shales-Skrzypny Shale Formation, 8 – super-Posidonia beds – Podzamcze Limestone Formation, 9 – cherty Biancone/Maiolica type limestones – Pieniny Formation, 10 – Globigerinal-Radiolaria beds – Kapuśnica Formatiion, 11 – Globotruncana marls–Jaworki Formation, 12 – Upper Cretaceous flysch Sromowce Formation 13 – Jarmuta Formation.



■ **Fig. 18.** Stop 7 at Krościenko. View on the Zawiasy Klippe.

static events (lithohorizone Be-7) and correspond very well with the Berriasian part of the Nozdrovice Breccia within Inner Carpathians 1996), which developed as scarp breccias along active submarine fault slopes. On the other side, the eustatic changes are perhaps connected with the global plate reorganization that took place during Tithonian-Berriasian time (Golonkaet al. 2003, 2005 b). This global plate reorganization is also related to the Tethyan Neo-Cimmerian tectonic activities.

The uppermost partion of the Zawiasy section (in overturned position) belongs to the Jaworki Formation (Cenomanian-? Campanian) composed of the variegated pelagic and marls of the Scaglia Rosa (= Couches Rouge = Capas Rojas) type. These pelagic, foraminiferal, multicoloured green/variegated/red marl deposits originated during the latest, third episode of evolution of the Pieniny Klippen Basin (Birkenmajer 1977, 1986, Bąk 2000), when unification of sedimentary facies took place within all successions (Albian-Coniacian). Flysch and/or flyschoidal facies are younger (Santonian-Campanian). During this syn-orogenic stage of the development of the Pieniny Klippen Basin the flyschoidal deposits developed as submarine turbiditic wedges, fans and canyon fills (Birkenmajer 1986) with several episodes of debris flows with numerous exotic pebbles (Late Albian-Early Campanian). The main "exotic source area" in the PKB was emerged, so-called Exotic Andrusov Ridge, as effect of Czorsztyn Ridge/Andrusov Exotic Ridge collision (Birkenmajer 1986, Golonka et al. 2003). During latest Cretaceous, flysch and/or flyschoidal deposits of the Jarmuta Formation (Maastrichtian-Palaeocene) were deposited. The contact between these formations is tectonic.

Composition of heavy mineral fraction in Jarmuta Formation in Krościenko Zawiasy is similar to the mineral set from Szczawnica-Zabaniszcze. The heavy fraction contains 26–88 % of zircon, 36–60 % of tourmaline (representing schörl-dravite series), 9–17 % of rutile, 1–4 % of garnet (with prevailing almandine molecule), 2–3 % of chromian spinel and 0–4 % of apatite (provenance of minerals see stop Szczawnica Wyżna-Zabaniszcze; Oszczypko and Salata 2005).

The major strike-slip right lateral fault is located along the Dunajec River. Mountains on the left side of Dunajec belong to Pieniny Mountains built of the Pieniny Klippen Belt successions. The successions exposed here are from south to north: Pieniny, Czertezik and Zawiasy. Mountains on the right side of Dunajec belong to Beskid Sądecki range built of flysch rocks of Krynica unit of the Magura nappe. Small fragment of this unit is exposed on the left bank of Dunajec north of the main Zawiasy outcrop. It is represented by Eocene Łącko type marls belonging to Zarzecze formation. The main tectonic boundary between Pieniny Klippen Belt and Magura nappe runs from west to east along the slopes of Pieniny Mountains south of Krościenko Town. Eastward this boundary runs from west to east along the slopes of Małe Pieniny range south of Szczawnica town. This boundary is developed as major strike-slip right-lateral fault. The Dunajec fault is younger than the main W-E boundary fault. It displaced Pieniny Klippen Belt 4 kilometers northward.

Stop 8. Pieniny Andesites at Wżar Hill – Mineralogy, Petrology and Palaeomagnetism

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For explanations see Conference excursion 1 – Stop 5.

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Notes