SYMPOSIUM III: GEODYNAMIC OF THE ALPINE OROGEN

INTEGRATED MODELLING AND RHEOLOGICAL STUDY OF THE WESTERN CARPATHIANS

M. BIELIK[1], H. ZEYEN[2] and A. LANKREIJER[3]

[1]Geophysical Institute of the Slovak Academy of Sciences, Dúbravská cesta 9, 842 28 Bratislava, Slovak Republic; geofmiro@savba.sk
[2]Department des Sciences de la Terre, Universite de Paris-Sud, Bat. 504, F-91405 Orsay Cedex, France
[3]Institute of Earth Sciences, Vrije Universiteit, 1081 HV Amsterdam, The Netherlands

Key words: Integrated modelling, rheology, finite element algorithm, Western Carpathians.

Introduction

For the purpose of solving geological and geophysical problems in framework of the PANCARDI project we concentrated on the intergrated modelling of the geophysical fields. Alhough the Western Carpathians belong to the regions which are geophysically well surveyed, the study of its lithospheric structure has usually been performed by independent interpretation of the single geophysical fields. Lateral and temporal changes in the lithosphere rheology of the Carpathian-Pannonian region have been shown to have pronounced effects on lithosphere dynamics (Lankreijer et al. 1996). Therefore rheological constraints on geodynamic models in the studied area are very important. Integrated modelling and rheological study of the Western Carpathian lithosphere was carried out along deep seismic reflection transect 2T.

Methods

The applied integrated modelling is capable of determining two-dimensional thermal lithospheric structure. On the basis of a finite element algorithm (Zeyen & Fernandez 1994), the temperature distribution in steady state regime given an assumed lithospheric thickness (here defined as the 1300 °C isotherm) and heat production profile is calculated. Densities are then calculated, based on this temperature distribution. The lithospheric thickness is adjusted by trial and error until the model fits all the data. The input data are as follows: topography, gravity, heat flow density, lithology, thermal conductivity, heat production and density.

Due to the importance of mantle density variations, it does not make much sense to do isostatic modelling using the crustal density variations alone. For example the typical modelling that utilizes crustal seismics, gravity and topography does not make sense, if we do not also take the lithosphere-asthenosphere boundary into consideration, especially, in areas like the Carpathians and the Pannonian Basin with important variations in lithospheric thickness.

In the uppermost level of the mechanically strong part of the lithosphere, rheology is generally goverened by brittle failure (Bayerlee's Law). At the temperatures exceeding roughly half the melting temperature of rock, ductile creep processes become the dominant deformation mechanism. The rheology of the lithosphere is predicted on the basis of extrapolation of failure criteria, lithology and temperature models. The approach also allows us to predict effective elastic thickness (EET) variations in the lithosphere (Lankreijer et al. 1999).

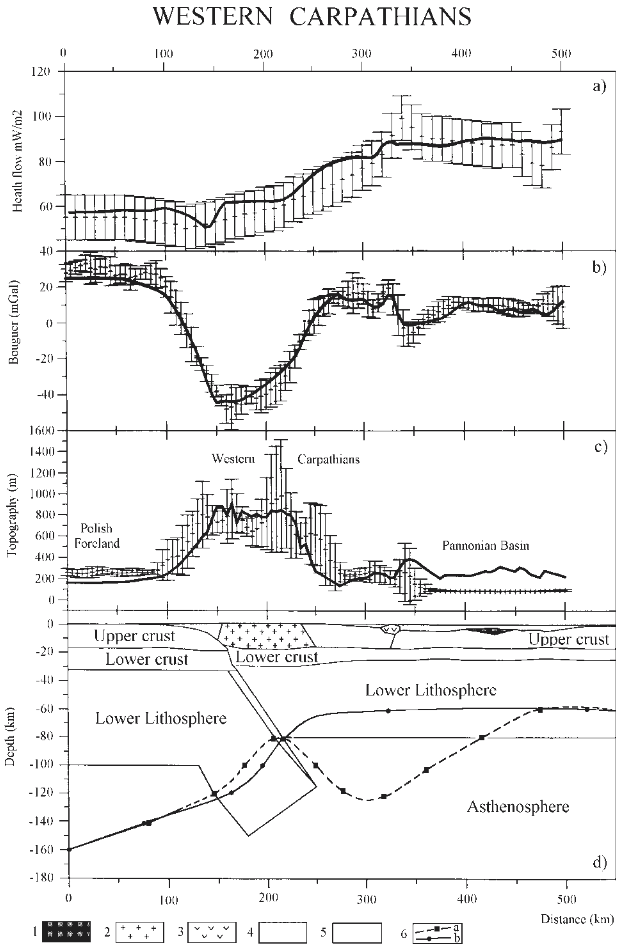

Fig. 1. Results of the integrated modelling along NNW-SSE Slovak Geotransect 2T. Explanations: (a) Surface heat flow. Commas represent measured values with corresponding error bars. The solid line represents calculated data. (b) Bouguer anomaly and (c) Elevation. Commas correspond to measured data and solid lines to calculated values. (d) Model geometry. Bodies: (1) Szolnok-Maramures flysch, (2) crystalline complex of the Tatricum, (3) Neogene volcanics, (4) Pannonian Basin Fill, (5) Mollasse and Flysch sediments of the external Carpathians, (6a) Lithosphere-asthenosphere boundary after Babuška et al. 1988, (6b) Lithosphere-asthenosphere boundary after Horváth (1993).

Results

The lithosphere thickness underneath the Polish foreland is about 100 km in our model (Fig. 1). It is much thinner than previously modelled (Babuška et al. 1988; Horváth 1993). One could decrease the heat flow there in order to thicken the European lithosphere. However, if we do this, the topography underneath the Polish foreland would become much too low. We also imply a decrease in upper-crustal heat production in order to be compactible with the measured heat flow data. Taking into account that the calculated mean value of the lithosphere thickness in Europe is about 100 km (Panza 1985), then the calculated thickness of the lithosphere by the integrated modelling in this area is similar (Zeyen & Bielik 1999). A presence of the lithospheric root can be observed underneath the highest topography of the Western Carpathians with a maximum beneath the northern slope of the Low Tatras Mts. A strong thickening is needed in order to avoid an excessively high topography of the northern part of the central Western Carpathians whereas a density decrease along the upper part of the thicknened area helps to keep the southern part of the central Western Carpathians at the observed high topography. Beneath the Pannonian Basin the calculated mean thickness of the lithosphere is 80 km.

Rheological study clearly indicates a general decrease in strength from the Polish foreland, via the Western Carpathians to the Pannonian Basin. In the Polish foreland area a horizontal rheological stratification of the lithosphere can be observed. Mechanically strong behaviour is predicted for the upper part of the crust, the uppermost part of the lower crust and the uppermost part of the mantle. On the other hand mechanically weak behaviour was calculated for the lower part of the lower crust and the lower part of the upper crust. These zones could act as a detachment levels. On the basis of strength predictions EET of 12 km was predicted. In the Western Carpathians, lower crustal strength completely disappears. The lithospheric strength gradually decreases towards the Pannonian Basin. This results from the increasing temperatures in this direction and the corresponding decrease of the thermally defined lithospheric thickness. The EET values of 15–23 km were calculated for the Western Carpathians. The Pannonian Basin is characterized by the extreme weakness of the lithosphere. Only one thin strong layer in the uppermost ten km of the crust can be observed. The EET is predicted at 5–10 km (Lankreijer et al. 1999).

Discussion

The overall agreements between the model that was calculated by integrated modelling and the data are quite good. But some differences in topography still exist. The calculated elevation is too low by about 200 m and in the Pannonian Basin it is too high by a similar amount. Based on analysis of the results obtained we suggest that the unexplained topography is due to the following reasons. In the Polish foreland it is, probably, due to the flexural bulge of the lithosphere. We have enough evidence to suggest (Bielik 1985; Zoetemeijer et al. 1997; Krzywiec & Jochym 1997; Lankreijer et al. 1999) that the foreland lithosphere behaves as an elastic plate that is governed by the flexural bulge behaviour. In the Pannonian Basin the problem could be associated with denser lithospheric mantle underneath this region. The lithospheric mantle has to be 5–10 kgm^{-3} denser than usual subcrustal mantle which may be explained by enrichment due to a plume.

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OLISTOLITHS — A PROOF FOR THE MIDDLE EOCENE MOBILITY OF THE MAGURA BASIN

J. BROMOWICZ

Faculty of Geology, Geophysics and Environment Protection,
Chair of Raw Rock Material Deposits University of Mining and Metallurgy,
al. Mickiewicza 30, 30-059 Kraków, Poland; bromow@@uci.agh.edu.pl

Key words: flysch, olistoliths, lithofacies, synsedimentary tectonics, Eocene, Polish Carpathians.

Olistoliths known from the Beloveza Formation (fm) are exposed in the Kamienica River valley, in the vicinity of Łabowa, by the road Nowy Sącz–Krynica (Fig. 1A, B). The rocks are located in the northern, marginal part of the Bystrica Subunit which is a fragment of the Magura Unit (Bromowicz 1998). To the north the Bystrica Subunit is thrusted over the Rača Subunit whereas to the south it plunges under the Krynica Subunit. The facial diversity of the structural units is illustrated in Fig. 1C, D and E.

Both mono- and polylithic olistoliths were distinguished. The former are represented by coarse- and fine-grained sandstones, and marls whereas the latter are bedded sets of sandstones, marls and shales in variable proportions.

Most of the olistoliths are composed of coarse-grained sandstones. The polylithic olistoliths consist of coarse-grained sandstones and marls interlayered with thin shale beds. These are accompanied by olistoliths showing distinctly higher percentages of shales, thin-bedded sandstones and marls and/or olistoliths composed exclusively of thick-bedded sandstones.

The olistoliths are enclosed within dark-bluish-grey marly shales strongly disturbed by submarine slides. The marly shales contain rare, mostly thin-bedded, fine-grained sandstones accompanied by irregular lumps of red shales.

The volumes of studied olistoliths vary from 0.24 cubic metres to, presumably, several millions of cubic metres. The largest olistoliths, although only partly exposed, are readily recognizable in morphology as they form distinct hills rising from the surrounding outcrops of shales of the Beloveza Formation (Fig. 2).

The olistoliths are mostly podiform. Spheroidal shapes are also common whereas discoidal and ellipsoidal forms occur less frequently. Olistoliths form the following structures: undisturbed slabs, bent slabs and anticlinally or synclinally deformed layers (folded slabs), occasionally tightly folded or cylindrical (Figs. 3, 4, 5). The outer surfaces of olistoliths show the presence of flute moulds, trace fossils as well as fine steps and chevrons in the form of wrinkles superimposed on hieroglyph moulds of various origin.

The origin of olistoliths was explained by the comparison of their lithology with those of sediments coexisting in the Magura Basin and belonging to the Beloveza, Żeleźnikowa and Magura Formations (fm). Sequences and their mineralogical compositions were analysed and compared with the use of the rank classification

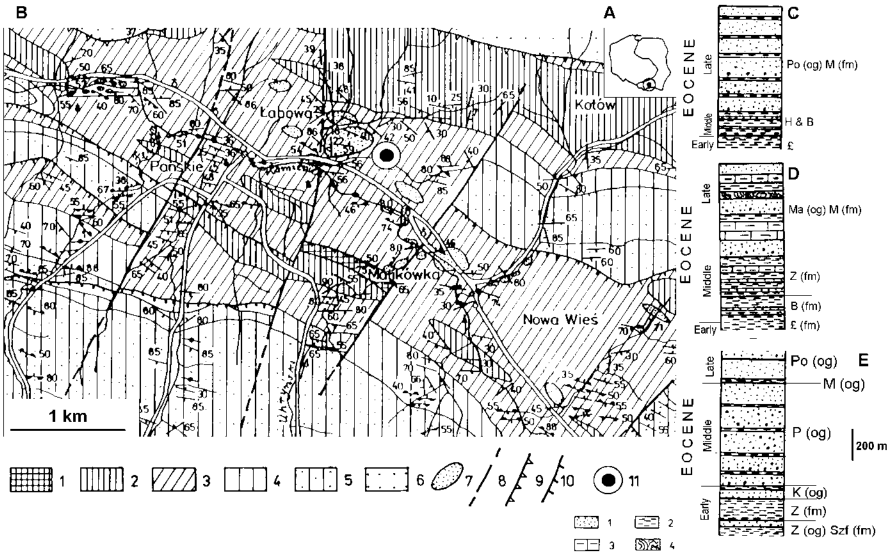

Fig. 1. Location of olistoliths and general geology of the studied area: **A** — location, **B** — geological map: 1 — "Inoceramian Beds"; 2 — Łabowa Shale Formation (fm); 3 — Beloveza Formation (fm); 4 — Żeleźnikowa Formation (fm); 5–6 — Magura Formation (fm): 5 — Maszkowice Member (og); 6 —Poprad Sandstone Member (og); 7 — olistoliths; 8 — faults; 9 — thrust of Bystrica Subunit over Rača Subunit; 10 — other overthrusts; 11 — olistoliths described in the text. **C–E** — generalized stratigraphy: 1 — sandstones; 2 — shales; 3 — marls; 4 — submarine slides, C — Rača Subunit: Łp — "Variegated Shales", H and B — "Hieroglyphic and Beloveza Beds", Po(og) — Poprad Sandstone Member; D — Bystrica Subunit: Ł(fm) — Łabowa Shale Formation, B(fm) — Beloveza Formation, Ż(fm) — Żeleźnikowa Formation, M(fm) — Magura Formation, Ma(og) — Maszkowice Member; E — Krynica Subunit: Sz(fm) — Szczawnica Formation, Ż(og) — Życzanów Member, Z(fm) — Zarzecze Formation, K(og) — Krynica Member, P(og) — Piwniczna Sandstone Member, M(og) — Mniszek Shale Member, Po(og) — Poprad Sandstone Member.

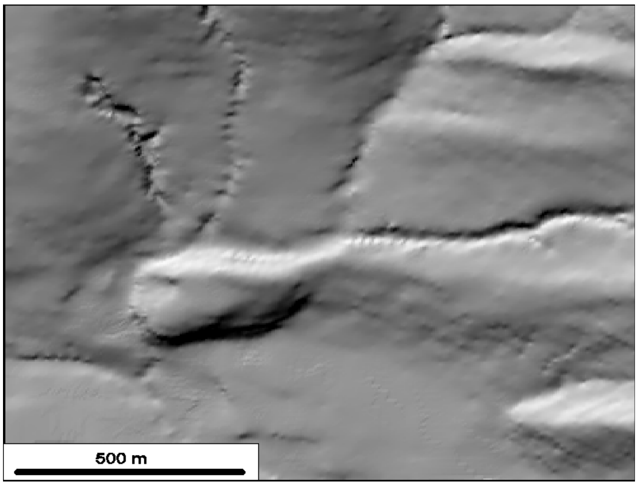

Fig. 2. Shaded DTM with the Kiczera Niżna mountain in central part. Olistolith marked in Fig. 1B.

Fig. 4. Marl olistolith of cylindrical shape.

Fig. 3. Polylithic olistolith as a recumbent anticline.

Fig. 5. Sandstone cylindrical olistolith.

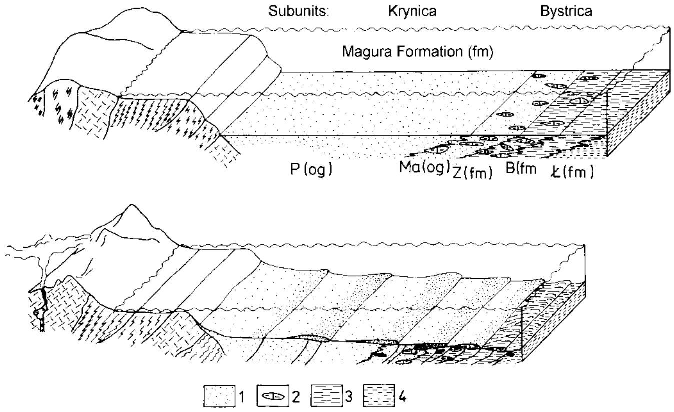

Fig. 6. Formation model of olistoliths as a result of changes in configuration of the Magura Basin. 1 — sandstones, 2 — marls, 3 — blue shales, 4 — variegated shales. P(og) — Piwniczna Sandstone Member, Ma(og) — Maszkowice Member, Ż(fm) — Żeleźnikowa Formation, B(fm) — Beloveza Formation, Ł(fm) — Łabowa Shale Formation.

based on Pearson's correlation coefficient. It was found that the olistoliths were derived from the Żeleźnikowa (fm) and the Magura (fm) Formations. The latter provenance is represented by the olistoliths of the Maszkowice Member (og) from the Bystrica zone and of the Piwniczna Member (og) from the Krynica zone. The shapes, outer surfaces and diagenetic features observed along the boundaries of the olistoliths and enclosing rocks point to their transport in semi-consolidated state.

The appearance of semi-consolidated olistoliths representing both the Żeleźnikowa (fm) and the Magura (fm) Formations within the Beloveza Formation proves the synsedimentary movements of material in the Magura Basin. The lithostratigraphic scheme of the basin (Oszczypko 1992) points to the Middle Eocene age of olistoliths displacement to the axial part of the basin. The movements were triggered by shortening of the basin presumably due to the subduction of the Magura Basin floor (Fig. 6).

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VEIN STRUCTURES — A FIRST OCCURRENCE WITHIN THE OUTER CARPATHIANS

M. ĆWIŻEWICZ

Institute of Geological Sciences, Research Centre in Kraków,
Polish Academy of Sciences, Senacka 1, 31-002 Kraków, Poland;
ndcwizew@cyf-kr.edu.pl

Key words: vein structures, synsedimentary tectonics, Skole Nappe, Outer Carpathians.

Introduction

The term *vein structures* was introduced by Cowan (1982) to describe subparallel planar structural discontinuities, which appear as dark linear, curved or braided traces. These structures were found in cores from the inner slope of the Middle-America Trench, off Guatemala.

Since then vein structures have been described from numerous localities, mainly located at the active plate margins (Lundberg & Legget 1986; Lundberg & Karig 1986; Karig & Lundberg 1990; Brown & Behrmann 1990). It is thought that they are the earliest brittle features formed in poorly indurated fine-grained sediments (Cowan 1982; Kemp 1990; Kimura et al. 1989; Lindsley-Griffin et al. 1990).

The origin of vein structures has been not fully explained. A series of hypotheses have been put forward (Carson et al. 1982; Cowan 1982; Ritger 1985; Hanamura & Ogawa 1993; Brothers et al. 1996). The results of the latest research (Brothers et al. 1996) suggest that vein structures are formed by combined action of downslope creep and seismic quakes, and that their orientation is controlled by regional stress field. It follows that vein structures can be used as a new stress indicator for paleostress analyses.

To my knowledge, the vein structures has not been described yet from the Outer Carpathians. However, features resembling vein structures were recognized by Antek Tokarski in 1997 in an exposure of the Oligocene strata of the Skole Nappe (Fig. 1). The object of the present paper is to provide the evidence that these features are indeed vein structures.

Methods

The discussed features have been studied in the Oligocene Dynow Marls exposed in the Straszydle quarry within the Skole Nappe (Fig. 1). Oriented samples were taken from a six-metre long section. Microscopic observations and structural analysis of microstructures were complemented by the cathodoluminescence (CL), back-scattered electron image (BEI), scanning electron microscope (SEM) and chemical analyses.

Description

The Dynow Marls (Kotlarczyk 1966) are thin-bedded, calcareous-siliceous rocks occurring in the lower part of the Menilitic Formation (Kotlarczyk & Leśniak 1990). The chemical composition of these rocks changes vertically within individual beds. The increase of calcium carbonate contents from 30 % to 74 %, is accompanied by the decrease of silica contents from 64 % to 19 %. These variations result in an inter-bedding lamination visible in exposure. Clay minerals contents are about 4 %. In thin sections, detrital components: quartz, feldspars and organic fragments as well as micritic carbonate and microcrystaline quartz patches are arranged parallel to the bedding.

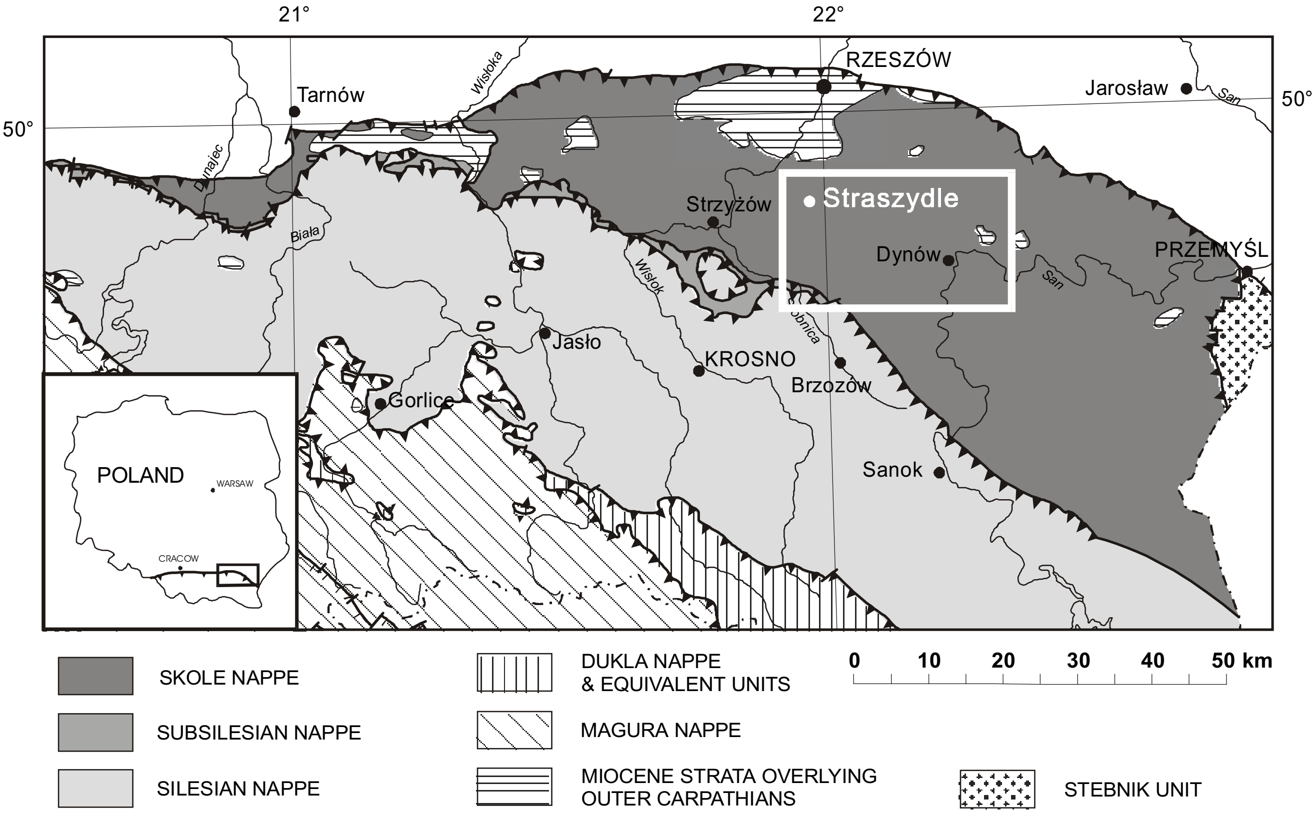

Fig. 1. Tectonic map of NE part of the Polish segment of the Outer Carpathians (after Poprawa & Nemčok 1988–1989; modified) with location of the study area (boxed).

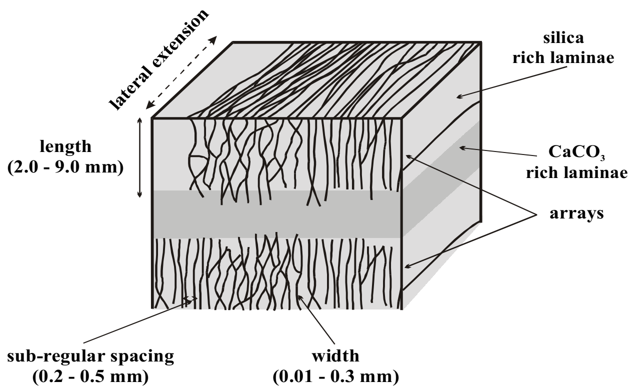

Fig. 2. Morphology and geometry of the studied structures.

In the plane perpendicular to the bedding the described planar features form dark linear or curved traces (Fig. 2), which are either braided or sub-parallel to each other. The individual structures are either sub-perpendicular or oriented at high angle to the bedding. They are arranged with sub-regular spacing to form arrays, which are oriented sub-parallel to the bedding. The spacing between the particular structures within an array is 0.2–0.5 mm. In thin-sections, discussed structures are 2.0 to 9.0 mm long, 0.01 to 0.3 mm wide and their lateral extension is up to 40.0 mm. The discussed features have been observed in silica-rich laminae. Most of these structures are confined to these laminae, and only a few of them cross the laminae's boundaries, terminating 1.0–1.5 mm further out. They form 2–3 sets striking sub-parallel to the strike of beds (Fig. 3). The boundaries between these structures and the host rock are sharp and are usually outlined by a brownish colouring. Based on BEI and CL analyses, the chemical composition of struc-

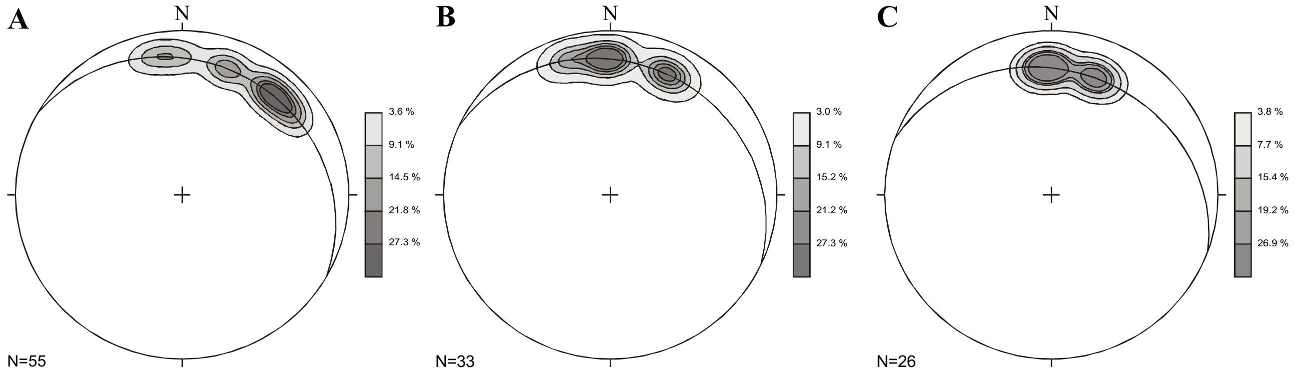

Fig. 3. Lower hemisphere plots of the studied structures at the top (A), in the middle part (B) and at the bottom (C) of the studied section. Great circles denote orientation of bedding.

ture fillings is similar to the host rock but in places they are enriched in ferrous pigment, probably formed by Fe-hydrooxides, giving the structure a rusty colour.

Discussions and conclusions

The discussed structures show three features diagnostic for vein structures.

1/ Similar to vein structures described by Cowan (1982) and Ogawa & Miyata (1985), the discussed planar structures occuring in the Dynow Marls form dark linear, curviplanar to sigmoidal traces oriented sub-perpendicular to the bedding. They show characteristically a sub-regular spacing and form arrays, which are sub-parallel to the bedding.

2/ Like vein structures (Ogawa & Miyata 1985; Kimura et al. 1989; Lindsley-Griffin 1990), the occurrence of the discussed structures is clearly controlled by the lithology of host rock. These structures occur mainly in siliceous fine-grained rocks.

3/ Resembling vein structures (Lindsley-Griffin et al. 1990; Pickering et al. 1990), the discussed structures show no composition contrast between structure fill and host rock. The fills in places are enriched in ferrous pigment.

It follows that the structures occurring in the Dynow Marls are vein structures.

Based on their morphology these vein structures are anastomosing and parallel type of vein structures distinguished by Ogawa & Miyata (1985) (Fig. 2).

The orientation of the discussed vein structures is similar in the whole studied section. This could suggest a tectonic control on their origin (cf. Brothers et al. 1996).

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OLISTOSTROME/MÉLANGES IN YUGOSLAVIA AND HUNGARY: AN OVERVIEW OF THE PROBLEMS AND PRELIMINARY COMPARISON

N.M. DIMITRIJEVIĆ,[1] D. M. DIMITRIJEVIĆ[1], S. KARAMATA[2], M. SUDAR[3], S. KOVÁCS[4], L. DOSZTÁLY[5], Z. GULÁCSI[5] and P. PELIKÁN[5]

[1]Hadži Melentijeva 82, 11118 Beograd, Yugoslavia; quax@EUnet.yu
[2]Faculty of Mining and Geology, Djusina 7, 11000 Beograd, Yugoslavia; cvladica@EUnet.yu
[3]Faculty of Mining and Geology, Kamenicka 6, 11000 Beograd, Yugoslavia; sudar@Eunet.yu
[4]Academic Research Group, Department of Geology, Múzeum krt. 4/a, 1088 Budapest, Hungary; skovacs@iris.geobio.elte.hu
[5]Hungarian Geological Institute, Stefánia út 14, 1143 Budapest, Hungary; geo@vax.mafi.hu

Key words: olistostrome-mélanges, ophiolite belts.

An overview of the problems

Chaotic rock complexes recognized at the end of the 19th and at the beginning of the 20th century in the Himalaya Mts. and containing blocks of "flysch-type" sediments, exotic, basic to ultrabasic rocks, were interpreted in two different ways, which have influenced the interpretation of such complexes in different parts of the world for many decades. These were: the genesis as a "volcano-sedimentary formation", or the mixture due overthrusting of regional nappes.

The term "mélange" did not exist at that time. In 1919, Greenly, describing the fabric of the of the Gwna Group from the Mona Complex of Anglesey Island at Scotland, coined the term "autoclastic mélange", writing: "The essential characters of an autoclastic mélange may be said to be the general destruction of original junctions, whether igneous or sedimentary, especially of bedding, and the shearing-down of the more tractable material until it functions as a schistose matrix in which the fragments of the more obdurate rocks float as isolated lenticles or phacoids." It is clear, that such a fabric in modern language of structural geology will be defined as the product of transposition of S-surfaces, representing a purely structural phenomenon.

The term "mélange" was further on used, without distinction, for all chaotic mixtures of various rocks not obeing the classical rules of sedimentary sequences (the law of superposition, the law of original continuity, and so on; cf. Gilluly et al. 1959).

During the discussion of Beneo's lecture at the 4th World Petroleum Congress (Rome 1955), Flores introduced the term and notion of olistostrome, and Gansser (1959) used this mechanism to explain partly the origin of the ophiolitic "colored mélange" of Iran and Beluchistan (olistostrome + tectonics). Thus two main branches of thinking have been opened for the origin of ophiolitic mélange — tectonic and sedimentary.

The advent of plate tectonic concept made a specific kind of mélange one of the very important features in the global architecture. This was the ophiolitic mélange — a mixture of matrix, different in composition, and exotic blocks of various size, where ophiolites, cherts and greywackes played an important rule.

General rules of mélange organization were given by Hsü 1968 and in 1974 he proposed criteria to make distinction between "mélanges" (advocating strictly their tectonic origin, without using a genetic adjective) and "olistostromes" (of sedimentary origin). However, if a "sedimentary mélange" endured subsequent pervasive shearing, a distinction becomes very difficult or impossible, as emphasized by Dimitrijević & Dimitrijević 1973a (see also the "Addendum" in Hsü 1974, p. 331).

At the same time the classification of mass gravity transport processes was developed (Middleton & Hampton 1973), allowing the recognition that olistostromes (= pure "sedimentary mélanges") are deposits of debris flows (Rupke 1978). Based on the examples recognized in NE Hungary, Kovács (1988) proposed a correlation between types of subaqueous mass gravity transport and their resulting deposits.

Besides the mentioned difficulties to distinguish "sedimentary" and "tectonic" mélanges even in well exposed ophiolite zones, in poorly exposed, hilly regions the interpretation of small outcrops of hard rocks, outcropping seemingly at random from below shale or marl debris, is further complicated, if the relationship to the "matrix " is not visible. In such cases even rocks of normal successions (for example, a sequence of deep water limestone – radiolarite – shale), disturbed only by folding or imbrication, but with undestructed stratigraphic connection can be misinterpreted as "sedimentary" or "tectonic" mélange. Therefore, a careful study of "block" composition, relationship to the matrix and structural relationships is needed for a real justification of the complex.

Summarizing this chapter, the "mélange" means simply a mixture and has itself no genetic significance without a descriptive adjective. The use of the term without such an adjective should thus be abandoned. The "ophiolitic mélange" should be used for regional chaotic mixtures connected with ophiolites and inclusions of exotic rocks. The genetic explanation (olistostromal: Dimitrijević & Dimitrijević 1973a; tectonic: Hsü 1974) depends largely on the paradigm adopted, the degree of terrain exposure, the quantitative relations between land- and ocean-derived ingredients, and the (subsequent) shearing of the complex.

Ophiolite mélanges in Yugoslavia

Ophiolite mélanges (the "Diabase–Chert Formation" in the former literature) occur in two broad geotectonic zones in former Yugoslavia: in the Vardar Zone in the East and in the Ophiolite Belt (called also "Dinaridic Ophiolite Belt) in the West. They are separated in the middle sector, situated in present Yugoslavia and dealt with in detail in the present contribution, by the Drina–Ivanjica Element/Terrane, from the NW end of which ("Tuzla–Zagreb sector") the separation becomes obscure. Nevertheless, the ophiolite mélange continues towards NW and from the Zagreb–Zemplin Zone (interpreted as a large Tertiary transcurrent zone) it turns towards NE (Dimitrijević 1974). In the territory of Hungary, borehole evidence of Bükk-type developments prove the connection to the Bükk Mts. and its surrounding in NE Hungary ("Igal–Bükk eugeosyncline" of Wein 1969). To the south the Ophiolite Belt is bound to the Mirdita Zone in Albania, as a probable issue of the Vardar (Tethys) Ocean.

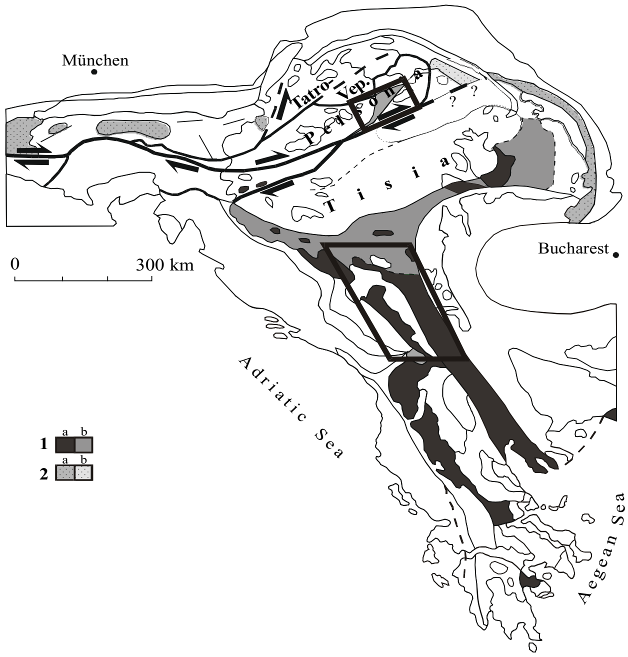

Fig. 1. Position of the compared areas (framed) on the simplified geotectonic sketch map of the SE part of Alpine Europe (simplified from Kovács et al. 1996–1997, Fig. 1), showing only the ophiolite zones. Legend: 1—(Neo)Tethyan ophiolite zones (a—outcropping, b—covered) (with Vardar Zone and Dinaridic Ophiolite Belt in Yugoslavia and Bükk area in NE Hungary); 2—Penninic and Outer Dacidic (or Civcin–Severin) ophiolite zones (a—outcropping, b—covered).

The Ophiolite Belt (OB): The matrix of the ophiolite mélange (OM further on) is composed of siliciclastic detritus deriving from the Drina-Ivanjica Element (DIE), accumulating in the apron along the border of the terrane. In places blocks of turbidites of the same composition are recognizable. Inclusions in the matrix are clasts (cm to dm), olistoliths (m to decam) and 100 m to km sized olistoplakae/olistonappes. They belong to two groups: 1) All formations of the Triassic DIE Carbonate Platform (Dimitrijević & Dimitrijević 1991), ending with olistonappes/gravitational nappes covering large areas of the OM; exotic blocks include Permian limestone and conglomerates with Permian clasts; 2) Blocks from the ocean floor: ophiolites (frequent pillow lavas, as well as blocks of cumulates and sheeted dyke complex, ultramafics) and cherts yielding both Triassic and Jurassic radiolarians (Gorican, in Obradović & Gorican 1989 and in Dimitrijević et al. 1996; Dosztály unpubl.). Characteristic of the OB are large ultramafic masses (Zlatibor being the largest), in places with preserved metamorphic sole. These are interpreted by Karamata (1988) as hot masses coming up in the accretionary prism. The matrix is usually weakly sheared. The oldest overstep sequence of the mélange can be found in Bosnia on the NW and is represented by the Pogari Series containing Latest Jurassic to Earliest Cretaceous clasts. The central segment of the OB is interpreted as an oceanic tract opened in the Middle Triassic and subducted in the Late Jurassic beneath the DIE, with gravity transport of huge olistoplakae into it and onto (=olistonappes) it.

The Vardar Zone (VZ): The Ophiolite mélange appears mostly in the External (Western) Subzone, with some occurrences in the Central Subzone. Its major part is of Jurassic age, with a thin veneer of Upper Cretaceous mélange at the western border of the VZ. In contrast to the OB, the mélange here shows the following specific features: 1) inclusions of limestone clasts are rare: 2) blocks of turbiditic sandstones are frequent; 3) matrix is intensely tectonized and tectonic mixing with adjacent rocks (even Cretaceous) is conspicuous in zones; 4) ocean-derived magmatics are profuse, but pillow lavas are less frequent, than in the OB; 5) large peridotite masses (like the Ibar mass) lack metamorphic sole and were thrust as cold masses towards the W; 6) large olistoplakae of sedimentary rocks are absent; 7) some mélange masses exerted regional (suprasubduction?) metamorphism of greenschists facies, others not. The Upper Cretaceous mélange occurs along the Zvornik suture (the present boundary towards the Dinarides). From the Majevica Mt. toward NW this suture is covered by Tertiary strata, but its continuation seems to pass there between the mélange of the Ophiolite Belt and the "tectonized mélange" north of it, regarded as part of the Vardar Zone (Mojicević et al. 1977; Jovanović & Magas 1986). This mélange east of Majevica contains fragments of Senonian Globotruncana limestone, and the "tectonized mélange" in the west bears olistoliths of basalts interlayered with the ?Upper Turonian limestone. According to Karamata et al. (1994) this area up to the Upper Cretaceous represented a backarc oceanic basin, with slightly different closing times of different segments.

Olistostrome-mélanges in NE Hungary

These mélanges occur in the Darnó and Szarvaskő ophiolite complexes of the Bükk Composite Terrane. The Mónosbél Unit of the Szarvaskő Complex consists of olistostromes with shaly matrix, interfingering with carbonate turbidites. The latter are of two types: ooidal limestone (Bükkzsérc Lmst., rather similar to the Grivska Formation of the DIE), and homogeneous microsparite. Olistoliths are of the same limestones, and also sandstones, Jurassic radiolarites and radiolarite breccia, with rare basalt, Triassic basinal limestone, and quartz conglomerate. The mafic rocks of the Szarvaskő Unit correspond to the back-arc setting (Harangi et al. 1996). Specific are Bódva-type Triassic limestone slide-blocks with basalt, identical to those in the Darnó Complex.

The Darnó Complex, as seen in drill-hole cores, is built up of an upper, magmatic unit with subordinate abyssal radiolarites and mudstones of both Triassic and Jurassic age, representing an accretionary prism and/or large slide blocks. Basalts and gabbros are of MOR-type (Dosztály & Józsa 1992). The lower, sedimentary unit comprises sedimentary rocks showing typical features of slumping, debris flow, and distal turbidites. They were derived from a pelitic and a marly-calcareous source area, respectively, less frequently from a sandy one. Shales are often silicified. Exotic slide-blocks are of the Triassic Bódvalenke type (reddish cherty limestone with basalts). A block of Upper Permian Nagyvisnyó Lmst. was also found. This lower unit, similar to the Mónosbél Unit, represents the proximal, toe-of-slope part of the trench complex.

A specific, volcanic-arc related olistostrome complex with rhyolite and limestone clasts occurs in the Telekesoldal Unit of the Rudabánya Mt. (Kovács 1988). The Ophiolite mélange was found also in a few boreholes in the SW part of the "Igal-Bükk zone", near to the national border, representing the continuation of the Repno Complex of the Kalnik Mt. (Haas et al. 1995).

A few summarizing remarks

In the Hungarian-Yugoslavian area the belts of ophiolitic mélange point to the presence of several Mesozoic oceanic realms: the Vardar Ocean as the main and highly complex Tethys part between the southern Tethys and the ALCAPA region (?Paleozoic to the Upper Jurassic, with a backarc basin closed in the Upper Cretaceous), the Dinaridic Ophiolite Belt ocean (Middle Triassic to the uppermost Jurassic, with a possible branch between present Dalmatian-Herzegovinan and East Bosnian-Durmitor Terranes, and displaced oceanic remnants along the large transcurrent fault in the "Igal-Bükk zone". Mélanges of these oceanic realms differ significantly in composition, depending on the geotectonic setting and composition of the surrounding continental blocks and oceanic areas. Nevertheless, some types of blocks common in the mélange of the Darnó–Mónosbél units and of the Dinaridic OB (Bódva-type limestones with basalt, basalts with red cherts, both Triassic and Jurassic, or the Grivska–Bükkzsérc Lmst. analogy) are worth mentioning.

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LATE PHASE OF THE CARPATHIAN THRUSTING IN RESPECT TO FLUID MIGRATION

J. FRANCŮ[1], O. KREJČÍ[1], J. MILIČKA[2], M. PERESZLÉNYI[3] and Z. STRÁNÍK [1]

[1]Czech Geological Survey, Leitnerova 22, 658 69 Brno, Czech Republic; francu@cgu.cz
[2]Department of Geochemistry, Comenius University, Mlynská dolina, 842 15 Bratislava, Slovak Republic
[3]VVNP Bratislava, Slovak Republic

Key words: Flysh Belt, thermal maturity.

The outer part of the Western Carpathians comprise allochthonous units which were transported over tens to hundreds of km from their original basin locations. A detailed study of the thermal history recorded in the rocks offers new insights into the burial, erosional, and thrusting evolution of this area.

The Flysch Belt consists of the Magura and Krosno-Menilite groups of nappes. They were gradually detached from their depostional environment and incorporated within the thrust-and-fold belt.

The geological units underwent three types of thermal alteration. The first one occurred due to burial below younger sedimentary formations in the basins. The second type of heating was induced by tectonic burial under the frontal thrust sheets of the moving orogenic belt. The third setting occurred when the unit was imbricated, stacked and buried its own lower scales to depth and increased temperature.

The analyses of organic matter in the deep boreholes in the Flysch Belt and its foreland show highest thermal maturity in the Magura (Rača) unit, lower in the Menilite-Krosno unit and the lowest in the Carpathian Foredeep. The saw-tooth profile of increasing maturity and reversals is typical of mainly pre-tectonic thermal alteration of the upper thrust sheets.

From the analysis of maturity distribution in profiles it is concluded that the Magura unit experienced all three types of heating and its significant part was eroded during stacking and prior to colision with the Krosno-Menilite units.

The foreland (the North European Platform) was the least and last thermally affected. Its margin was bent down, buried under the Western Carpathians and exposed to temperatures high enough to generate fluids only in pre-Neogene source rocks. The major part of the fluids generated within the overthrust sheets probably escaped before the nappe complex arrived at the margin of the North European Platform.

CONTINUATION OF THE PERIADRIATIC LINEAMENT, ALPINE AND NW DINARIDIC UNITS IN THE PANNONIAN BASIN

J. HAAS[1], P. MIOĆ[2], J. PAMIĆ[3], B. TOMLJENOVIĆ[4], P. ÁRKAI[5], A. BÉRCZI-MAKK[6], S. KOVÁCS[1], E. RÁLISCH-FELGENHAUER[7] and B. KOROKNAI[1]

[1]Geological Research Group, Hungarian Academy of Sciences–Eötvös Loránd University, Budapest, Hungary; haas@ludens.elte.hu
[2]Geological Survey of Slovenia, Ljubljana, Slovenia
[3]Croatian Academy of Sciences and Arts, Zagreb, Croatia
[4]University of Zagreb, Croatia; bruntom@rudar.rgn.hr
[5]Laboratory for Geochemical Research, Hungarian Academy of Sciences, Budaörsi út 45, H-1112 Budapest, Hungary
[6]MOL Hungarian Oil and Gas CO, Hungary
[7]Hungarian Geological Institute, Hungary

Key words: structural units, lineament, Pannonian Basin, Hungary, Slovenia, Croatia.

Introduction

Intense explorations in the last half century revealed that a strongly tectonized narrow belt traverses the Pannonian Basin in NE–SW direction where structural units (terranes) of Austroalpine, South Alpine and Dinaridic origin and a large dismembered block of the Eurasian plate margin (Tisza Unit or Tisia Terrane) got into juxtaposition. These units are bounded by major fault systems, namely the Periadriatic-Balaton and the Zagreb-Zemplin (or Mid-Hungarian) Lineament systems having crucial importance in the geology of the Pannonian Basin and surrounding Alpine-Carpathian-Dinaridic mountain ranges.

The aim of the present paper is to define the structural units of the area studied on the basis of comparative analyses and draw conclusions for the continuation of the Alpine and Dinaridic units in the basement of the SW part of the Pannonian Basin (Fig. 1).

Results

The extension and relationships of the structural units of the pre-Tertiary basement and the major discontinuity belts (lineament systems) are shown in a simplified way in Fig. 2. In addition to the data collected from the exploratory bore-holes, several previously published maps were used for the compilation of the map (Fülöp et al. 1987; Flügel 1988; Placer in Turnšek 1997).

Conclusions

Summarizing and re-evaluating the relevant surface and subsurface data for the studied area, we came to the following conclusions:

1. According to the diagnostic characteristics of the Periadriatic Lineament system: 1/ it is a first order structural element which separates the South Alpine and the Austroalpine and Penninic units and 2/ late Variscan and Paleogene magmatites of similar geochemical features occur along it. The Balaton Lineament system may be rightly considered the continuation of the Periadriatic Lineament, as it has been assumed by a number of previous authors (Bögel 1975; Wein 1978).

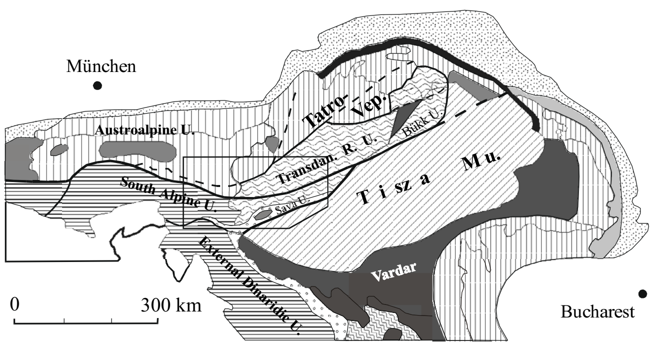

Fig. 1. Location map of the study area.

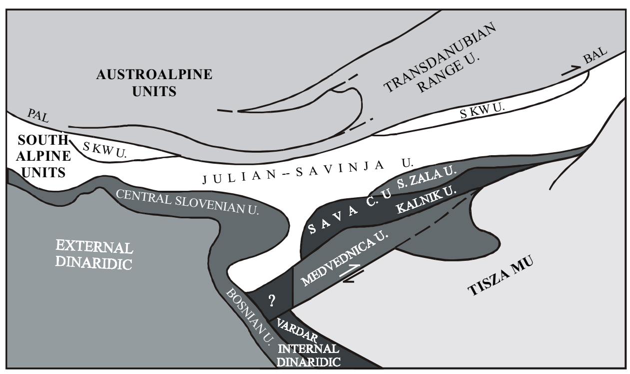

Fig. 2. Sketch map of the structural units and the major lineaments.

2. Windows exposing the Penninic Unit, metamorphic complexes of the Lower- and Middle Austoalpine units, Upper Austroalpine nappes and the Transdanubian Range Unit are located north of the Periadriatic-Balaton Lineament system. Due to its pre-Tertiary paleogeographical setting, the Transdanubian Range Unit shows mainly South Alpine facies relations but as it is thought to overlie the Lower–Middle Austroalpine units (Tari 1996), its structural position is identical with that of the Upper Austoalpine nappes.

3. Between the Periadriatic-Balaton and Zagreb-Zemplin Lineament systems heterogeneous structural units are juxtaposed forming the Sava Composite Unit. In the northern part of this composite unit non-metamorphosed nappes (South Karawanken and Julian-Savinja units) occur which can be regarded as the eastern continuation of the South Alpine units. These nappes are overthrust onto the Internal Dinaridic units, that is onto the Central Slovenian Unit, the previously metamorphosed Medvednica and South Zala units and the ophiolite mélange of the Kalnik Unit which is the prolongation of the Vardar Zone. This means that in the Sava Composite Unit both South Alpine and Internal Dinaridic elements occur, and along Tertiary (Oligocene–Miocene) thrust faults the South Alpine elements are overthrust onto the Dinaridic ones.

4. The Bükk Composite Unit is located northeast of the Sava Composite Unit, on the northern side of the Zagreb-Zemplin Lineament. It is made up of structural units having crucial features akin to those in the Sava Unit: Paleozoic–Triassic successions of South Alpine-Dinaridic affinity, affected by Alpine (Cretaceous) metamorphism (Árkai et al. 1995), and ophiolite mélange of Internal Dinaridic affinity.

5. The Zagreb-Zemplin (Mid-Hungarian) Lineament separates the Sava Unit from the Tisza Megaunit (Tisia Terrane) which is characterized by a Variscan and Alpine evolution fundamentally different from that of both the South Alpine and the Inner Dinaridic units and shows affinity to the evolution of the Tethyan margin of the Eurasian plate during the early stage of the Alpine evolution.

6. The present-day setting of the structural units forming the basement of the Pannonian Basin is the result of a series of different structural processes. Subsequent to the Variscan cycle, the main stages of structural evolution were as follows: multiple Neotethys opening from the Middle Triassic to the Middle Jurassic, ocean closure, obduction, nappe formation, regional metamorphism in the Late Jurassic–Cretaceous interval, sizeable overthusting of the South Alpine units onto the Dinaridic ones and strike-slip displacements along the major lineaments and within the Sava Composite Unit in the Tertiary.

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OCEANIC REALMS IN THE CENTRAL PART OF THE BALKAN PENINSULA DURING THE MESOZOIC

O.S. KARAMATA[1], N.M. DIMITRIJEVIĆ[2] and D.M. DIMITRIJEVIĆ[2]

[1]Serbian Academy of Sciences and Arts, Knez Mihajlova 35, YU-11000 Belgrade, Yugoslavia; cvladica@eunet.yu
[2]Hadzi Melentijeva 82, YU-11000 Belgrade, Yugoslavia; quax@eunet.yu

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

At the beginning of this century two ophiolite zones were already noted in the central part of the Balkan peninsula. These are very well expressed at the Geological map of SFRYu 1:500,000 (1970), and numerous papers deal with some parts or some members of these zones, included in this text are only those which offered new data. Correlative synthesis of the main belts, based on modern geological approaches has been first given by Dimitrijević & Dimitrijević (1973). More recent comparative presentations appeared in 1998, in papers by Karamata, Dimitrijević & Dimitrijević (a and b). The present paper is a further elaboration of thoughts developed in these papers.

Existing data point to the existence of three different ophiolitic belts (Fig. 1), resulting from separate oceanic realms and basins, and differing in geotectonic framework, age of birth and duration of oceanic expanses, being reflected in presence/absence of specific sedimentary and magmatic members and their sedimentologic, petrologic and geochemic characteristics. These data are summarized in the Table. The ophiolitic belts are:

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

1a. The main ophiolite belt of the Vardar Zone (MVZ) as scar of the main Vardar oceanic realm — the Tethys. This oceanic realm had a long continous existence as continuation of the Lower Pale-

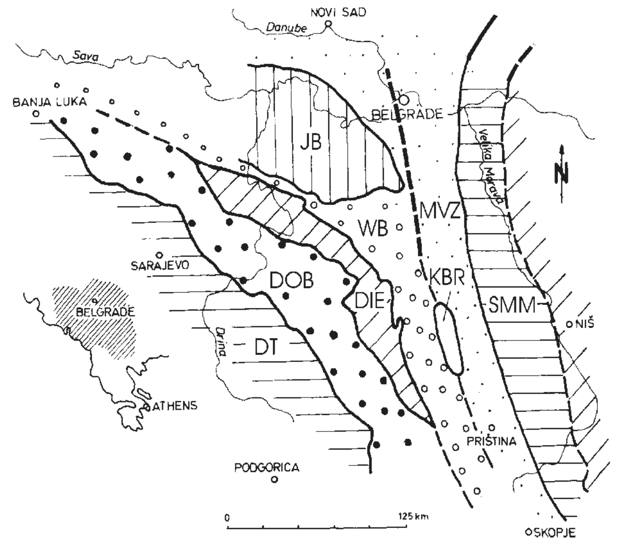

Fig. 1. Present position of ophiolite belts, relics of Mesozoic oceanic realms in the central part of the Balkan peninsula. DT — main Dinaridic trunc; DOB — Dinaridic Ophiolite Belt; DIE — Drina-Ivanjica Element; WB — western belt of the Vardar Zone; KBR — Kopaonik block and ridge, MVZ — main Vardar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

Table 1: Correlation of oceanic realms in the present Balkan peninsula.

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

Time-scale is not linear
* Northern part of the belt
Capitals – characteristic members

← V → opening; → ← closing

ozoic (or older ?) oceanic realm; it closed before the end of Jurassic; and is characterized by the presence of the Paleozoic island arc relics (“Veles Series”); prevalence of material from the higher parts of the oceanic crust in the olistostrome, and the absence of limestone olistoplakae.

1b. The western ophiolitic belt of the Vardar Zone (WB) as scar of the western marginal basin of the Vardar ocean separated from the main Vardar ocean by the Kopaonik block and ridge unit. This basin existed from the (middle part ? of) Upper Triassic to the uppermost Senonian. Its most expressive features are the prevalence of graywackes in the olistostrome; the absence of Paleozoic metamorphics and granitoids; the presence of Upper Jurassic ophiolitic sole metamorphics and Upper Cretaceous limestones fragments.

2. The Dinaridic Ophiolite Belt (DOB), with the southern continuation into the Mirdita Zone, as a relic of the oceanic tract which ran through the Dinarides and was along its southeastern border connected locally with the Vardar ocean. It existed from the Middle Triassic to the end of the Jurassic. Its characteristics are the prevalence of greywacke and Triassic limestone as olistoliths in the olistostrome; bodies of Carboniferous granite and kilometric lenses of deep-sea silicious rocks; olistoliths of Middle Triassic to Upper Jurassic chert; large olistoplakae of Triassic limestone; ultramafic plates and diapirs with contact-metamorphism in the floor or around respectively.

3. The Civcin-Severin ophiolite belt in the east. Insufficiently known, due to very complex cover of young nappes, but existing data point to some similarities with western belts.

To conclude, in the central part of the Balkan peninsula relics of several Mesozoic oceanic basins are visible, differing in the opening and closing times. The main oceanic realm was the Vardar ocean, which existed from the lower Paleozoic (or even earlier), whilst other basins represented marginal basins or oceanic crust generated between the dispersing continental fragments. The composition of olistostromes, deposited in subduction troughs of these oceanic basins, differs significantly, depending on the width of the oceanic area, geotectonic position and lithology of the trough margins.

These data show that ophiolitic belts of the central Balkan peninsula represent relics of different Mesozoic basins with oceanic crust, and could not be regarded as products of the one sole oceanic realm.

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THE JASŁO NAPPE *VERSUS* GORLICE BEDS. A CONTROVERSY AT THE FRONT OF THE MAGURA NAPPE IN THE JASŁO-GORLICE REGION (POLISH MIDDLE CARPATHIANS)

A. KOSZARSKI

Institute of Geological Sciences, Jagiellonian University, ul. Oleandry 2a, 30-063 Kraków, Poland

Key words: Carpathians, Jasło Nappe, Fore-Magura zone, Krosno Beds, Gorlice Beds, olistostromes.

Introduction

The problem of the front of the Magura Nappe and its basement in Jasło-Gorlice region has a complicated history. Tołwiński (1921) described the Jasło Nappe as a large tectonic nappe lying on the Silesian Unit (Middle Group) and below the covering Magura Nappe. The facial development of the Jasło Nappe was different from both units. This nappe has been preserved as tectonic outliers in Jasło vicinity.

The Tołwiński's concept of the Jasło Nappe has been abandoned without publishing any criticising paper. That happened probably because of lack of more modern and more detailed Tołwiński's writing explanations of his idea.

Later this zone was incorporated by many authors (for example Świdziński 1953) to the Magura Nappe as its tectonic "peninsulas" and outliers. The eastern "peninsula" was called as "the Harklowa peninsula". However all authors described this marginal zone of the Magura Nappe as different compared more southern zones of the large Magura Nappe.

Following older suggestions (Tołwiński l.c.; Burtan & Sokołowski 1956; Książkiewicz 1972) a new presentation of the Jasło Nappe was given by Koszarski & Koszarski (1985). According to the authors this nappe has a continous flysch succesion of the Paleogene to Early Miocenian deposits. The Late Eocene-Early Oligocene deposits are formed by certain facies (mostly shaly), called "the Duląbka Beds". Younger deposits are equivalents of Menilite type beds (containing the Menilite Hornstones) and Cergowa Shales (containing the Tylawa Limestones). The Krosno Beds are the youngest member of the Jasło Nappe. Sediments are strongly influenced by the presence of olistostromes and debris-flows especially well developed in the Duląbka Beds. Some reworked carbonates are common. Geometrically, the Jasło Nappe is an independent tectonic unit lying on various tectonical units (Silesian, Dukla, Grybów nappes). The Magura Nappe is thrust over the Jasło Nappe. The position of the Jasło Nappe is considered (Koszarski & Koszarski l.c.; Koszarski 1985; Ślączka 1998) as resembling the position of the S Fore-Magura Scale (sensu Burtan 1968).

Recently, Jankowski (1997, 1998, 1998a) modified older Szymakowska's (1976) ideas and described the Gorlice Beds as the youngest deposits of the sedimentary succession of the Silesian Unit in the Gorlice area. These beds (after Jankowski l.c.) are chaotic sediments of Early Miocene age containing numerous olistostromes and debris-flows deposits. According to the new founding of Early Miocene age shaly deposits Jankowski (l.c.) included all the succession of the Jasło Nappe to these chaotic Gorlice Beds.

Methods

The relatively well outcropped region of "Harklowa peninsula" and its neighbourhood was investigated SE of Jasło Town. Classical field mapping was a background. Archival drill cores and other oil industry materials yielded additional data.

Results

Investigations show that the Jasło Nappe in the Jasło-Gorlice region is a reality. The stratigraphical sequence of the Jasło Nappe described generally earlier by Koszarski & Koszarski (l.c.) can be applied, however the internal facial variability exists. The olistostromes are grouped in some horizons. Locally is possible to observe them on surface on beds extention traces. Locally is also possible to correlate olistostrome horizons in different limbs of fold structures. The olistostromes sometimes show features of internal olistostromes.

The tectonic style of the Jasło Nappe stated earlier by Koszarski & Koszarski (l.c.) is true. In the northern part of the "Harklowa peninsula" slightly deformed folds and scales occur. Chiefly they are possible to be indicated on maps. In the southern part of the "peninsula" at the front of the Magura Nappe complications are stronger. In that zone small imbricated scales are dominant. The Jasło Nappe is strongly cut by numerous mostly cross faults.

The Jasło Nappe is structurally independent from its base. The thrust surface of this nappe is rather flat. The Magura Nappe is thrust over the Jasło Nappe.

Discussion

The presented results show that the Jasło Nappe in the Jasło-Gorlice region is a fact. This statement excludes out the concept of the Gorlice Beds (sensu Jankowski l.c) which are artificial and have origin different from their definition. These Beds Jankowski (l.c) combined on the base of two separate elements. The first element was the whole Jasło Nappe. The second one was probably some olistostromes on the Krosno Beds of the Silesian Nappe. Such horizons in the Krosno Beds occur and were notified also in the Jasło region (Jasionowicz & Szymakowska 1963). The fact of the presence of the Early Miocene age deposits in the Jasło Nappe succession explains only the time of the final stage of sedimentation in this zone. And/or: some shaly deposits were tectonicaly transported to the surface from below the Jasło Nappe during later stages of tectonisation (i.e. thrusting, folding, scales forming, duplexing, faulting). In such a case it estimates a time of ending of sedimentation at the front of the Jasło Nappe. In every case it is obvious that the Gorlice Beds do not exist.

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GEOLOGICAL EVOLUTION OF MIDDLE PONTIDE AREA (NORTHERN TURKEY) AND EASTWARD CONTINUATION OF THE MELIATA-HALLSTATT OCEAN

H.W. KOZUR[1], M. AYDIN[2], O. DEMIR[2] and H. YAKAR[3]

[1]Rézsü u. 83, H-1029 Budapest, Hungary; kozur@helka.iif.hu
[2]Türkiye Petrol. A.O., Exploration Group, Mustafa Kemal Mahallesi, 2. Cadde No. 86, 06502 Esentepe-Ankara, Türkiye
[3]Türkiye Petrol. A.O., Research Center, Mustafa Kemal Mahallesi, 2. Cadde No. 86, 06502 Esentepe-Ankara, Türkiye

Key words: Paphlagonian Ocean, Karakaya Ocean, Meliata-Hallstatt Ocean, *Torlessia mackayi.*

The Paleozoic and Mesozoic geological development of Turkey is of decisive importance for the understanding of western Tethyan paleogeography. The continuation of the Izmir-Ankara Ophiolite Belt in the Axios-Vardar Zone of Greece, Macedonia and Serbia is undisputed among Turkish and Western European geologists. No equivalents of North Alpine and Western Carpathian units have been found within and south of the Izmir-Ankara Belt.

The Karakaya Ocean was situated north of the Izmir-Ankara Belt, within the Sakarya Composite Terrane, in the latest Permian–Triassic. The view of Okay & Mostler (1994), Pickett et al. (1995) and Ustaömer & Robertson (1995) that the Karakaya Ocean was a large Late Paleozoic-Triassic Paleotethyan ocean cannot be confirmed. The assumed Lower Permian red radiolarites of the Çal Unit belong to the latest Permian late Dorashamian Stage and the assumed Upper Permian basic volcanics of the Çal Unit were dated as Middle Triassic (Kozur 1997, 1999). The Middle Permian to Dzhulfian shallow-water platform carbonates (present over a distance of more than 1000 km) are not remnants of sea mounts as assumed by Pickett et al. (1995) and Ustaömer & Robertson (1995). Neither the facies and fossil content of these platform carbonates nor their very wide distribution and absence of slope and basinal sediments of this age support the sea mount hypothesis. The Nilüfer Formation (basic volcanics and pelagic limestones, partly in glaucophane schist facies) of the Karakaya Ocean, with an assumed Permian–Triassic age was dated in its type area to the Early Triassic (Kozur et al. in press b).

Both Ladinian red radiolarites and pillow lavas (as in the Meliaticum) and Middle Triassic Hallstatt Limestones and cherty limestones (as at the northern slope of the Meliaticum) occur in the Karakaya Ocean, but Upper Triassic siliciclastic flysch and a post-middle Norian/pre Rhaetian closure exclude that the Karakaya back-arc basin was the eastern continuation of the Meliata-Hallstatt Ocean.

The Küre Complex in the Middle Pontides, often regarded as a remnant of the Paleotethys (e.g. Sengör 1985) was proven to be a Variscan metamorphic oceanic complex (Bekirli Group s.s. of Aydin et al. 1995) which is discordantly overlain (relations often tectonically obscured) by the typical North Alpine shallow-water Lower-Middle Triassic Sirçalik Formation (Kozur et al. in press a). The latter begins with Alpine Buntsandstein (with a basal conglomerate at its base). This is overlain by typical Werfen Beds, a hypersaline horizon at the Scythian-Anisian boundary, Gutenstein Beds and Steinalm Dolomite. These beds are (?unconformably) overlain by marginal parts of the Akgöl Group (dark siliciclastic beds). Towards the central part of the basin (originally in the north), the Akgöl Group consists of dark deep-water shales, siltstones, turbidites and olistostromes, well exposed, e.g. along the road Küre-Inebolu, somewhat south of the Black Sea coast. The matrix contains the upper Carnian *Torlessia* n. sp. and the lower to middle Norian *Torlessia mackayi* (Kozur et al. in press a), both known from New Zealand, *T. mackayi* also from the Antalya nappes in southern Turkey and Burma. Rich deep-water (but not abyssal) trace fossils are also present. The olistoliths consist both of shallow-water, slope and basinal limestones of Middle Triassic age, with oldest pelagic faunas of the Chiosella timorensis Zone (basal Anisian), black Anisian and very rarely red Ladinian (?) radiolarites, basic volcanics and clasts of ultrabasites. The shallow water, slope and basinal rocks are very similar to the deposits of the southern shelf and slope of the Meliaticum, whereas the Middle Triassic ophiolites and pelagic sediments are similar to that of the Meliaticum. The Middle Triassic opening of the ocean is also similar to the Meliaticum, but the siliciclastic rocks begin earlier, within the Carnian, whereas the closure was close to the end of the Middle Jurassic, as in the Meliaticum.

Around Çalça in the Middle Pontides, Jurassic siliciclastic rocks and olistostromes contain large slices of Pelsonian to Norian Hallstatt Limestones (Kozur et al. in press c). All lithofacies types of the North Alpine and Western Carpathian Hallstatt Limestones are present. The Jurassic siliciclastic, but partly marly rocks with the Hallstatt Limestone slices are named as the Çalça Formation. Ophiolite slices are present within (?) Jurassic turbidites adjacent to the Hallstatt Limestone slices. The turbidites with ophiolite slices and the Çalça Formation are united in the Çalça Unit. The Jurassic siliciclastic rocks of the Çalça Unit with pre-Jurassic ophiolites and large slices of typical North Alpine Hallstatt Limestones are analogous to those of the Eastern Alps and Western Carpathians. As towards the south equivalents of the Meliaticum and North Alpine Triassic are unknown, the continuation of the Meliaticum (with its typical northern slope and outer shelf Hallstatt Limestones) through the Kotel Zone (Middle Jurassic turbidites with Hallstatt Limestone blocks) was in the northern Middle Pontides as assumed by Kozur & Mock (e.g. 1997) and Kozur (1991).

After the closure of a Caledonian Ordovician-Silurian deep water trough (connected with strong thermal alteration and very low-grade metamorphism), the northernmost unit of the middle Pontides, that is the Zonguldak Terrane was a marginal part of the east European Platform. Immediately south of this unit, siliciclastic turbidites and olistostromes of the Jurassic-Lower Cretaceous Beykoz Group occur, and contains block from the Zonguldak Terrane. Beside these blocks, olistoliths of pelagic Carboniferous to Middle Permian rocks (pelagic limestones, cherts) are present. The same rocks can be found as olistoliths in the Tauridian flysch of SE-Crimea. They were separated from the Beykoz Group by the later opening of the Black Sea. The Late Paleozoic deep-water rocks belong to a Late Paleozoic ocean (remnant basin of the Variscan ocean ?) at the margin of the East European Platform, which is named the Paphlagonian Ocean. It has no westward continuation.

The following mutual relationships between the three units with oceanic sequences of the northern part of the Middle Pontides (original south-north arrangement: Akgöl Unit, Çalça Unit and Beykoz Unit) can be assumed. The Akgöl Unit and the Çalça Unit were part of the Cimmerian Ocean. Its southwards directed subduction began during the Carnian, as indicated by thick siliciclastic deep-water turbidites and olistostromes (Akgöl Group). During that time, the typical North Alpine Hallstatt Limestones were deposited at the northern, passive margin of the ocean or on an intra-oceanic ridge. The Paphlagonian Ocean was either subducted during the Late Permian-Early or Middle Triassic and by this the Cimmerian Ocean opened by supra-subduction zone sea-floor spreading, or it was the oldest part of an ocean that closed latest by the southwards-directed subduction of the Cimmerian-Paphlagonian Ocean. The fact that the siliciclastic turbidites and olistostromes become continuously younger from south to north (mainly Late Triassic in the Akgöl Unit, Jurassic in the Çalça Unit and Late Jurassic to Early Cretaceous in the Beykoz Unit) speaks in favour of the latter model. The fact that no pelagic Triassic limestones and cherts have been found as olistoliths in the Beykoz Unit supports the first model. However, this may reflect insufficient investigation of the olistoliths in the Beykoz Unit because olistoliths of pelagic Triassic rocks are known from the Tauridian flysch of Crimea. However, it is not clear, whether they occur in the same unit as the olistoliths of pelagic Permian rocks (which are present in equivalents of the Beykoz Unit) or in a different unit corresponding to the Çalça Unit.

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PRE-ALPINE STRUCTURE OF THE CARPATHIANS AND ADJACENT AREAS

A. MEDVEDEV and A. VARITCHEV

Institute of Geology and Geochemistry of Combustible Minerals NAS of Ukraine, Naukova Str. 3a, 290053 Lviv, Ukraine; igggk@ah.ipm.lviv.ua

Key words: Carpathians, paleotectonic reconstruction, pre-Alpine events, tectonic setting.

The reconstruction of the pre-Alpine tectonic situation on the territory of the Carpathians and adjacent areas, based on data, which were obtained during the investigation of pre-Mesozoic formations in outcrops (Inner Carpathians), boreholes (Transcarpathian depression, Carpathian Foredeep and adjacent areas of the platform) and rock fragments in flysch and molasses (Outer Carpathians), allowed us to determine the most characteristic features of the pre-Mesozoic formations, their type and tectonic setting (eugeosynclinal, miogeosynclinal, orogenic, platform), to make correlations between regions, to establish the stages of formation of the Earth´s crust of this region, and to reconstruct the paleotectonic setting. This showed, that before the Mesozoic the place of the Carpathians was occupied by a mountain and fold construction, which was a component part of the structure of the South-West frame of the East-European platform, in particular — part of the Middle-European Hercynides with homologues (or, perhaps, a direct extension) of their zones (Rhenohercynian and Saxoturingian), and also of the Hercynian foredeep. This foredeep is now overlapped by formations of the Alpine stage, but its existence is proved by fragments of rocks, which composed this foredeep (coal-bearing Carboniferous and Permian conglomerates verrucano), found in the flysch and molasses of the Carpathians. The pre-Alpine history of the Carpathians was finished by the formation here of the crust of a continental type, which, at the beginning of the Alpine stage, underwent extension and breaking, which caused, the complete destruction of the granite-metamorphic layer, in some places, as is proved by the presence in various parts of the Carpathians of magmatic rocks of oceanic type. Thus, in relation to the previous structure of the Carpathian region, the Alpine geosyncline is new formed, not inherited from earlier stages.

We present the results of the investigation in graphic form, as a table (Fig. 1) and scheme (Fig. 2), that, in our opinion, gives a clear idea of pre-Alpine events in the Carpathian territory and the tectonic situation, to which these events led.

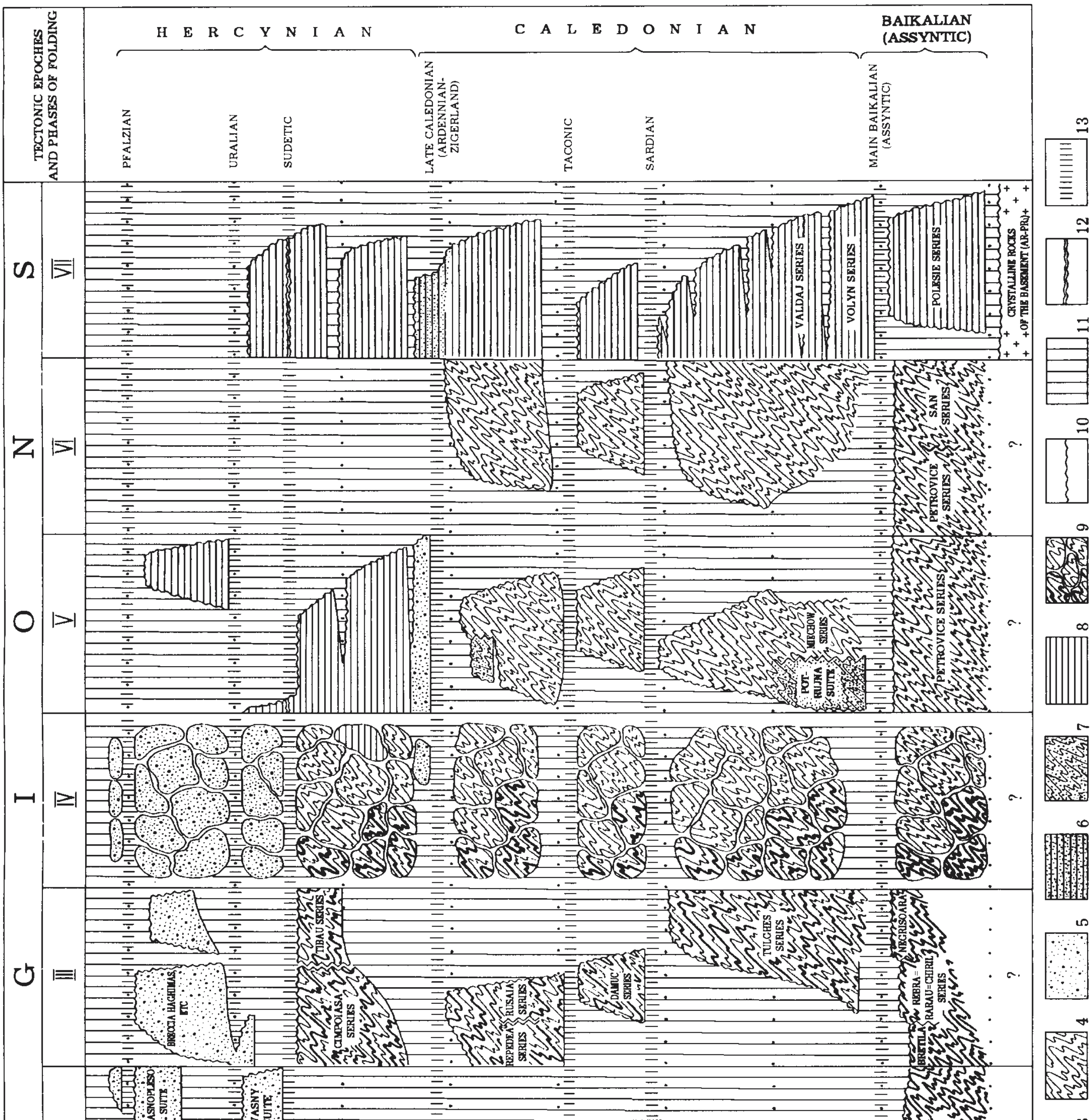

Fig. 1. Periodization of the pre-Alpine tectonic events on the territory of the Carpathians and adjacent regions. 1-2 — eugeosynclinal mesozonally (1) and epizonally (2) metamorphosed complexes; 3-4 — miogeosynclinal greenschist (3) and non-metamorphosed (4) complexes; 5 — orogenic complexes; 6 — teleorogenic complexes (Old Red of the Volyno-Podillia); 7 — geoanticlinal orogenic complexes; 8 — platform complexes; 9 — fragments of pre-Alpine rocks in flysch and molasses of the Carpathians; 10 — discordances; 11 — absence of deposits; 12 — boundary between paralic and liminal Carboniferous formations in the Lviv-Volyn basin; 13 — phases of folding. Regions: I — Western Inner Carpathians, II — Ukrainian part of the Eastern Inner Carpathians, III — Romanian part of the Eastern Inner Carpathians, IV — Outer Flysch Carpathians, V — Polish Carpathian Foredeep, VI — Ukrainian Carpathian Foredeep, VII — Volyno-Podillia.

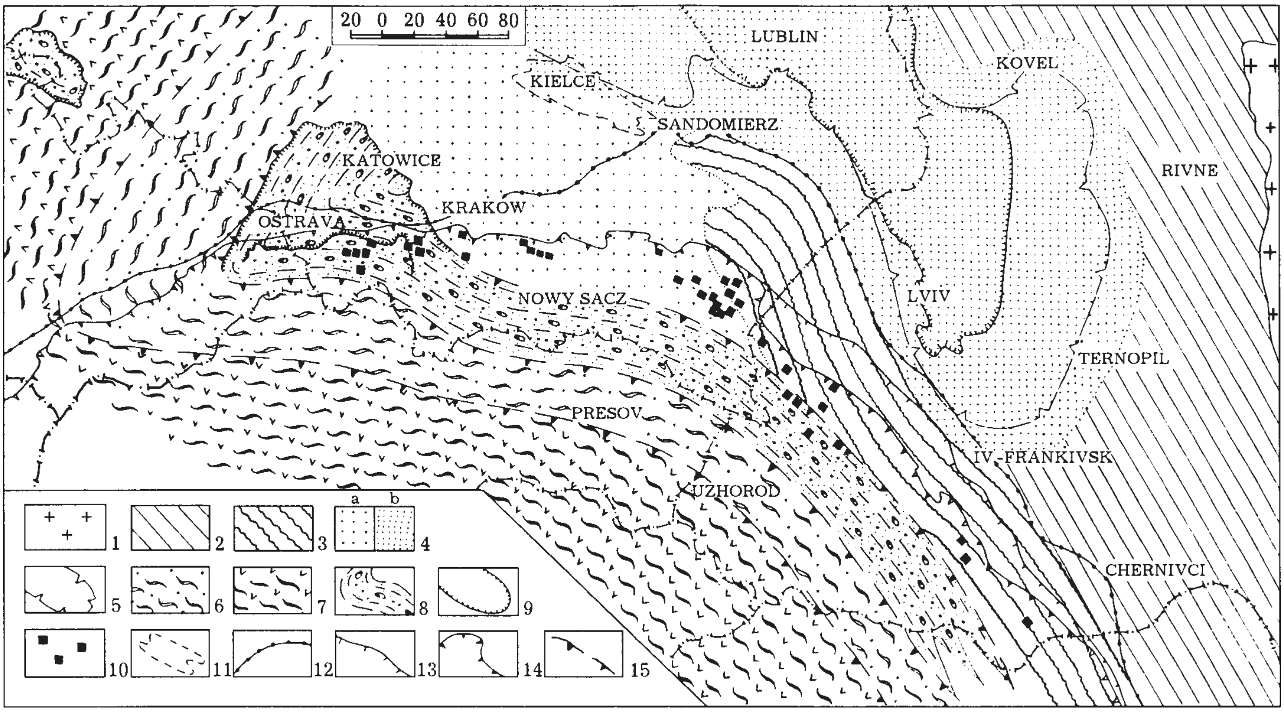

Fig. 2. Pre-Alpine tectonic situation in the territory of the Carpathians and adjacent regions. 1 — Ukrainian shield; 2 — pre-Hercynian complexes of the East-European platform's cover; 3 — pre-Hercynian fold complexes on the surface (marginal platform uplift); 4 — Hercynian platform complexes: a — on the epi-Caledonian platform (Miechow depression), b — in posthumous foredeep (Lviv-Lublin); 5 — present-day margins of posthumous foredeep; 6-8 — Hercynian mountain and fold construction: 6 — miogeosynclinal zone; 7 — eugeosynclinal zone; 8 — foredeep; 9 — present-day margins of Upper Carboniferous depressions; 10 — location of coal deposits in the Carpathian flysch deposits; 11-14 — present-day boundaries of: 11 — Gory Swietokrzyskie Mts.; 12 — Outer zone of the Carpathian Foredeep; 13 — Inner zone of the Carpathian Foredeep; 15 — Hercynian thrusts.

RAUHWACKES — A KEY TO UNDERSTANDING OF THE SUPERFICIAL THRUSTING MECHANISMS: CASE STUDY FROM THE MURÁŇ NAPPE

R. MILOVSKÝ[1*], D. PLAŠIENKA[1**] and J. SOTÁK[2]

[1]Geological Institute, Slovak Academy of Sciences, Dúbravská cesta 9, 842 26 Bratislava, Slovak Republic; [*]geolben@savba.savba.sk; [**]geolplas@savba.savba.sk
[2]Geological Institute, Slovak Academy of Sciences, Severná 5, 947 01 Banská Bystrica, Slovak Republic; sotak@gu.bb.sanet.sk

Key words: Muráň nappe, rauhwacke, thrusting, fluid pressure, hydrofracturation.

Introduction

Rauhwackes of various origins and tectonic positions were studied by several authors in the Alps and Appenines. Some of the authors (Masson 1972; Jaroszewski 1982; Plašienka & Soták 1996) recognized these rocks as basal cataclasites of overlying thin-skinned nappes, at the time of formation saturated with fluid, and thus facilitating the movement of the nappe body. In our research, rauhwackes occurring below the base of several West-Carpathian cover nappes — namely the Muráň, Drienok, Silica s.s., Choč and Križna nappe — were also considered to be the basal tectonic breccias, which accommodated the displacement of related nappes. The data presented in this short contribution have been obtained from the rauhwackes of the Muráň nappe and aim to show, how the investigation of the geometry, anatomy and alteration processes of rauhwackes may give deeper insight into the phenomena of thin-skinned tectonics. Moreover, it can provide important knowledge about the pre-thrusting substratum and timing of the thrusting event.

Methods

Detailed mapping was carried out in the Drienok nappe (district of Ponická Lehôtka), Muráň nappe (district of Tisovec) and Choč nappe (district of Malužiná), a number of samples — ca. 400 — was collected also from the other above mentioned nappes. Rock textures were described in the field and after detailed observation in optical microscope and stereomicroscope, cathodoluminiscence fabrics were investigated as well. Samples were dissolved in acetic acid and authigenic minerals from the insoluble residue were identified by means of powder X-ray diffraction and electron microprobe and further subjected to a fluid inclusions study.

Results

Rauhwackes (in some places only pronounced as "zone of rauhwackization") were observed to be a continuous layer underlying the base of the nappe, reaching up to 150 m in thickness. Texturally the rock is a tectonic breccia, typically with dominating portion of a calcitic matrix. Polymict clastic material is derived both from the substratum and, prevailingly, from whole pile of the overriding nappe — in some cases mostly from its upper levels. Therefore the parent mass of "protorauwacke" must have primarily gathered at the nappe front and was continually overridden by the advancing thrust, as shown on the Fig. 1.

In the space between the substratum and moving upper plate the debris was submitted to extensive cataclasis enhanced by a high hy-

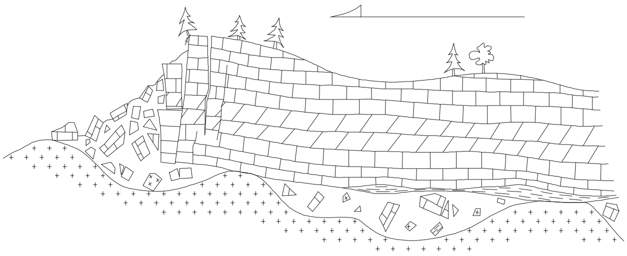

Fig. 1. Hypothetical sketch of the rauhwacke development. Features of both sedimentary and tectonic breccia are obvious.

draulic pressure, and severe chemical and mineral alterations. Dedolomitization, dissolution of quartz and feldspars, subsequent authigenesis of these minerals, as well as pyrite, apatite, phlogopite, illite, chlorite, tourmaline of dravitic composition all indicate that the system was highly fluid dominated and should be described in terms of hydrothermal, rather than metamorphic petrology. The physicochemical properties (pH, Eh...) of the fluid must have been changing in course of advancing "rauhwackization", as inferred from the presence of counteracting processes such as dissolution and precipitation of quartz and feldspar, crystallization and oxidation of pyrite. From the pattern of neoformed minerals it is also obvious that the fluid was saturated with respect to Na^+, K^+, Mg^{2+}, Fe^{2+}, Al^{3+} and SiO_2. Moreover, the presence of anhydrite, forming inclusions in quartz documents the activity of sulphatic anion. In the fluid inclusions chlorides NaCl, KCl and CaCl were determined cryoscopically with overall NaCl-equivalent concentrations between 43–50 wt. % (using equations in Sterner et al. 1988). Liquid CO_2 was observed in flat inclusions in authigenic feldspars, its mole fraction in the entrapped fluid we estimated to be ranging from 3.7 to 7.6 %. Besides these 3-component inclusions pure NaCl and pure liquid CO_2 inclusions were also found, pointing to complicated immiscibility phenomena, not interpreted yet.

Fluid pressure (p_{fluid}) might be assessed from the presence of chaotic hydrofracturation as supralithostatic, since the tension must overcome lithostatic pressure increased by rock strength to produce hydraulic fractures in all directions. More concrete assessment of p_{fluid} is offered by fluid inclusions, which, at gas-phase disappearance at 220–370 °C and total homogenization temperatures 380–420 °C exhibit pressures between 0.7 and 3.2 kbar (estimation after Bodnar 1994; Schmidt et al. 1995). Interpretation of pressure corresponding to the depth of 10 km is troublesome, because stratigraphically based estimations of overburden thickness infer ca. 2 km. Therefore we must assume that highly excessive pressures arose only locally, but relatively frequently in many places.

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DISTINCTIVE STRIKE-SLIP AND VERTICAL MOVEMENTS ALONG THE NE MARGINAL FAULTS OF THE EAST SLOVAK BASIN AND THEIR CONSEQUENCES

M. MOŘKOVSKÝ[1], J. MAGYAR[2], R. RUDINEC[2], F. JUNG[2] and J. SOTÁK[3]

[1]Geofyzika a.s., Brno, Czech Republic
[2]Nafta a.s. Gbely, PTS Michalovce, Slovak Republic
[3]Geological Institute, Slovak Academy of Sciences, Severná 5, 974 01 Banská Bystrica, Slovak Republic; sotak@gu.bb.sanet.sk

Key words: East Slovak Basin, strike-slips, thrust faulting, hydrocarbon accumulations.

Introduction

The East Slovak Basin and its pre-Neogene basement represents a noteworthy area of the West Carpathian–East Carpathian junction. Since of 1958, the East Slovak Basin has been investigated by shallow, medium and deep drillings and seismics, which have provided information on basin tectonics, depocenters, hydrocarbon source rocks and reservoirs. The latest seismic research, which was done in 1986–1995, allows to precise the structural-tectonic interpretation and origin of the East Slovak Basin. By this, the basin research yields reliable data for the recognition of distinctive normal and strike-slip faulting.

Basin history

The deep boreholes, which penetrated the Neogene sediments and basin basement, were investigated by gravimetric, magnetic and seismic methods. The results provide an evidence for a "pull-apart" origin of the East Slovak Basin followed by the polystage extension and different intensity of horizontal and vertical movements (Vass et al. 1988).

Early Miocene sediments were deposited along the NE margin on the Mesozoic and Paleogene substrates. After the Ottnangian emersion and hiatus the basin was downfaulted again along the NE margin being infilled by shallow marine and lagunar deposits belonging to the Karpatian (sandstone-siltstone sediments, dark and variegated claystones, evoporites, etc.).

In the Early Badenian, the East Slovak Basin occurred under large-scale extension with a distinctive deepening and marine sedimentation. Acid volcanism was also highly active in this time. The Middle Badenian was characterized by the halite deposition. In the Upper Badenian the volcanogenic sedimentation became dominant in the basin and adjacent areas as well. The Badenian evolution of the East Slovak Basin finished with a new phase of basin deepening iniciated partly by the rotational movement of the Western Carpathians.

In the Lower Sarmatian, the East Slovak Basin ceased to subside, and the basin depocentre was inversed and shifted to the S and SE, being succeeded by new appearances of volcanites (Hrádok, Lesné). The basin was filled by the thick formations of sandstone-siltstone deposits, which are sealed off by claystones and volcanoclastics.

Structural-tectonic interpretation

Since the Lower Miocene the eastern and central part of the East Slovak Basin has been actively subsided (Baráth et al. 1998). The

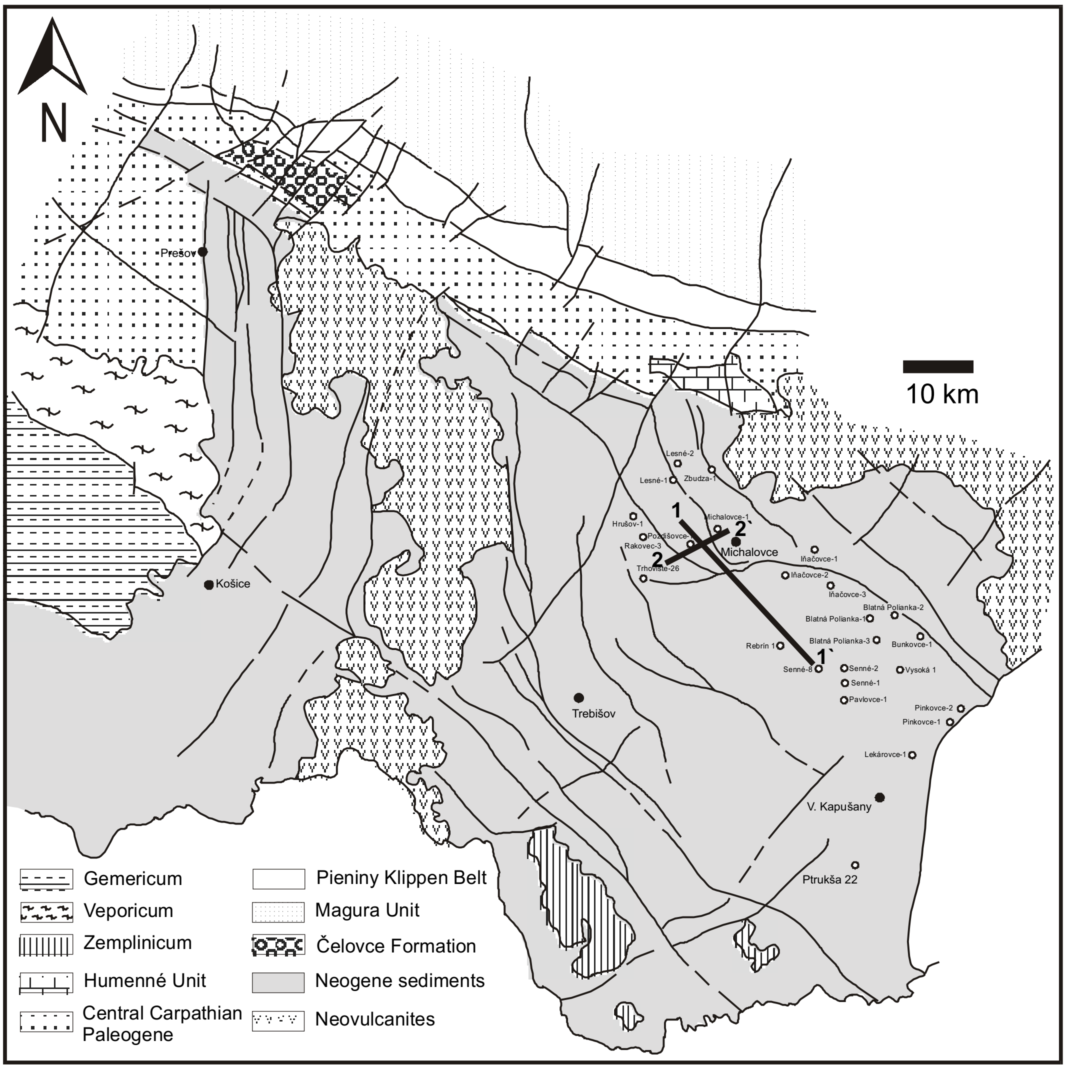

Fig. 1. Gravity survey in the area of the East Slovak Basin and the location of profiles 1, 2.

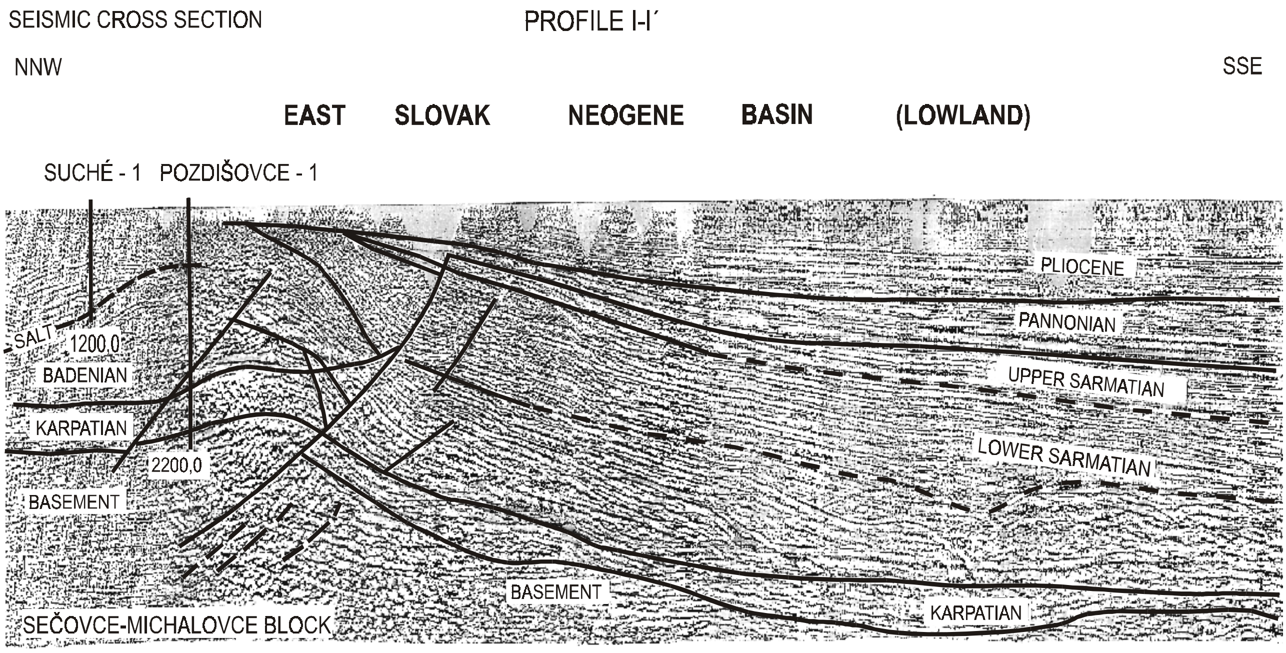

Fig. 2. Seismic cross-section (amplitudes envelopes + phases) showing the interpretation of structural styles of the East Slovak Basin along the Profile 1–1´.

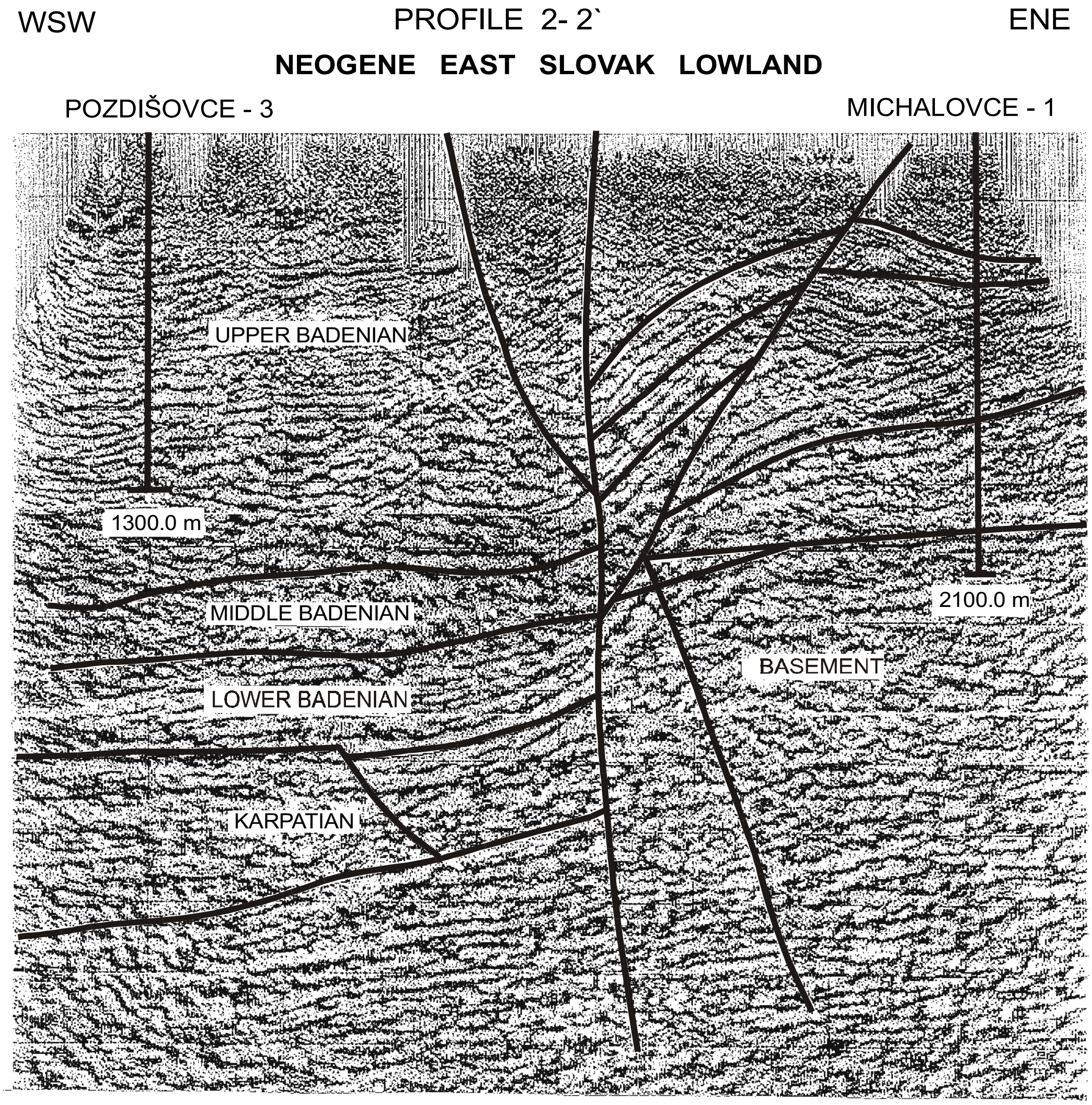

Fig. 3. Hilbert transformation of a time section (amplitudes envelopes + phases) showing the structural styles of the central part of the East Slovak Basin in the region of Michalovce (Profile 2–2´).

main mobile structure of subsidence activity can be identified with the master normal fault (décollement) dividing the Neogene sedimentary cover from the Pre-Neogene basement. New seismic data from the area of Michalovce proved and precised knowledge on listric nature of downfaulting (Fig. 1). The originally overthrust faults in structure of the Iňačovce-Krichevo Unit, were later reactivated as normal faults giving a force for the exhumation of this unit as well as for basin subsidence. The main detachment fault between Neogene sediments and the basin basement is apparent in seismic sections for its pronounced reflectivity. Because of the recognition of similar structures in the Pre-Neogene basement, the detachment faults grade up to higher structural level. So that, the youngest detachment fault of this system overprinted the sedimentary cover-basement boundary. Listric normal faults caused the cyclic subsidence of the basin during the Neogene extension reaching up to 7000 m thickness of sedimentary infillings.

During the Sarmatian–Pannonian, the strike-slip tectonics became dominant in the East Slovak Basin (Magyar et al. 1997). The main wrench zone is linked with the Močarany-Topľa Fault, revealing the dextral strike-slip movement from the NW to SE. This motion was compensated by oblique-stepping slips along the Trhovište Fault. In the Michalovce-Pozdišovce area, the compressed blocks between the strike-slip faults rose up to the intrabasinal elevations. Around this elevation, the depocentres oscillated in dependence on the magnitude and frequency of the block movements. In the seismic sections, the Michalovce-Pozdišovce structure appears as an asymmetrically elevated block bounded by dip-slip faults with distinctive vertical offset (up to 800 m, Fig. 3). The movement between the strike-slip faults is evidenced by eastward thrust faulting (Fig. 2). The most important fact is, that the Upper Badenian sediments are overthrust above the Lower Sarmatian sediments, which limited the age of strike-slip duplex formation to the Latest Sarmatian up to Pannonian. Accordingly, numerous positive flower structures originated during the Late Sarmatian. The seismic measurements indicate, that the Upper Sarmatian formations are refolded mainly above the flower structures (Fig. 3). The brachyanticlinal structures are related to the younger synsedimentary faults, which fade out below the Upper Sarmatian formations.

The coincidence of upper basement relief with the zones of thrust faulting was identified by Mořkovský & Lukášová (1986), alike that, Soták et al. (1993) reported the superposition of the basement metamorphic complexes above the sandstone sediments as a consequence of thrust faulting. It seems, that the Pozdišovce block was initially overthrusted and subsequently uplifted, which is obvious mainly from the deformation of the Badenian halite-bearing formation. The structural pattern of strike-slips, fault-bounded blocks and thrust faults reveals a transpressive regime of the East Slovak Basin during the Late Sarmatian–Pannonian.

Conclusions

The central part of the East Slovak Basin reveals an intensive deformation related to detachment faulting and strike-slip tectonics. As a result, an elevate megastructure with several apexes (Pozdišovce, Trhovište, Bánovce) originated. The elevated structures are responsible for trapping the natural gas occurred in the structural levels of the Upper Badenian and Lower Sarmatian sediments reaching the depth of 400–1600 m.

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FROM REMNANT OCEANIC BASIN TO COLLISION-RELATED FORELAND BASIN — A TENTATIVE HISTORY OF THE OUTER WESTERN CARPATHIANS

N. OSZCZYPKO

Institute of Geological Sciences, Jagiellonian University, Oleandry 2a, 30-063 Kraków, Poland; nestor@ing.uj.edu.pl

Key words: basin evolution, remnant ocean basin, foreland basin, Mesozoic, Cenozoic, Western Carpathians.

Introduction

The Western Carpathians are part of a great arc of mountains (Fig. 1). They are subdivided into the Inner and Outer Carpathians, separated by the Pieniny Klippen belt (PKB). The Outer Carpathians are composed of Late Jurassic to Early Miocene flysch deposits and are completely uprooted from their basement. The outer range consists of a Late Oligocene/Middle Miocene accretionary wedge, built up from several nappes and sub-horizontally overthrust onto the Miocene deposits of the Carpathian Foredeep. During its Mesozoic–Cenozoic history, the Outer Carpathian basin developed from a remnant ocean basin to a collision-related foreland basin. This work considers how that evolution was recorded in the burial history of the Outer Carpathian basin.

Methods

The burial history of the Outer Carpathians has been constructed on the basis of 7 selected sections from the Polish Outer Carpathians. Four of the sections are located in the Magura Nappe and the other three are in the Silesian, Sub-Silesian and the Skole units (Fig. 2). For computation of the subsidence curves, the procedure of Angevine et al. (1990) was used. This procedure makes it possible to plot the decompacted depths versus time for the stratigraphic units and "tectonic" subsidence. The corrections for paleobathymetry and sea level fluctuations have not been considered.

Results

In the Magura basin the first period of subsidence was probably initiated during the Berriasian and was followed by an Early/Late Cretaceous passive, thermal subsidence (Fig. 2). This period was characterized by deep-water deposition (below CCD) with very low rates of sedimentation (0.5–5 m/Ma) and followed by the Maastrichtian-Paleocene inversion and deposition. A new episode of subsidence began at the turn of the Paleocene in the Krynica sub-basin, and was accelerated during Lutetian and Priabonian in the Bystrica and Rača sub-basins with the deepest part (beneath the CCD) located in the north (Fig. 2). The rate of sedimentation varied from 6–18 m/Ma on the abyssal plain to 103–160 m/Ma in the outer fan, and between 180 and 400 m/Ma in the area affected by the middle fan-

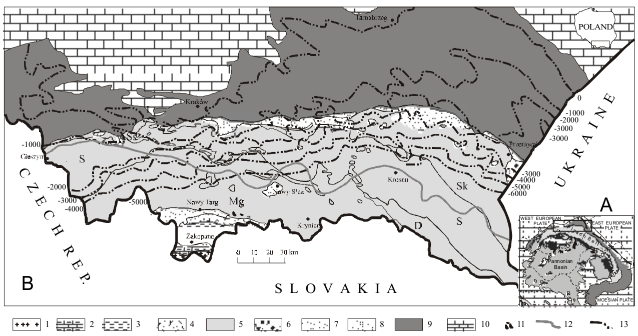

Fig. l. A — position of the Polish Carpathians. **B** — Sketch-map of the Polish Carpathians and their foredeep (after Oszczypko 1998 suplemented); 1 — crystalline core of Tatra Mts., 2 — high and sub-Tatra units, 3 — Podhale Flysch, 4 — Pieniny Klippen Belt, 5 — Outer Carpathians, 6 — Stebnik Unit, 7 — Miocene deposits upon Carpathians, 8 — Zglobice Unit, 9 — Miocene of the foredeep, 10 — Mesozoic and Paleozoic foreland deposits, 11 — andesites, 12 — northern extent of Lower Miocene, 13 — isobath of Miocene substratum.

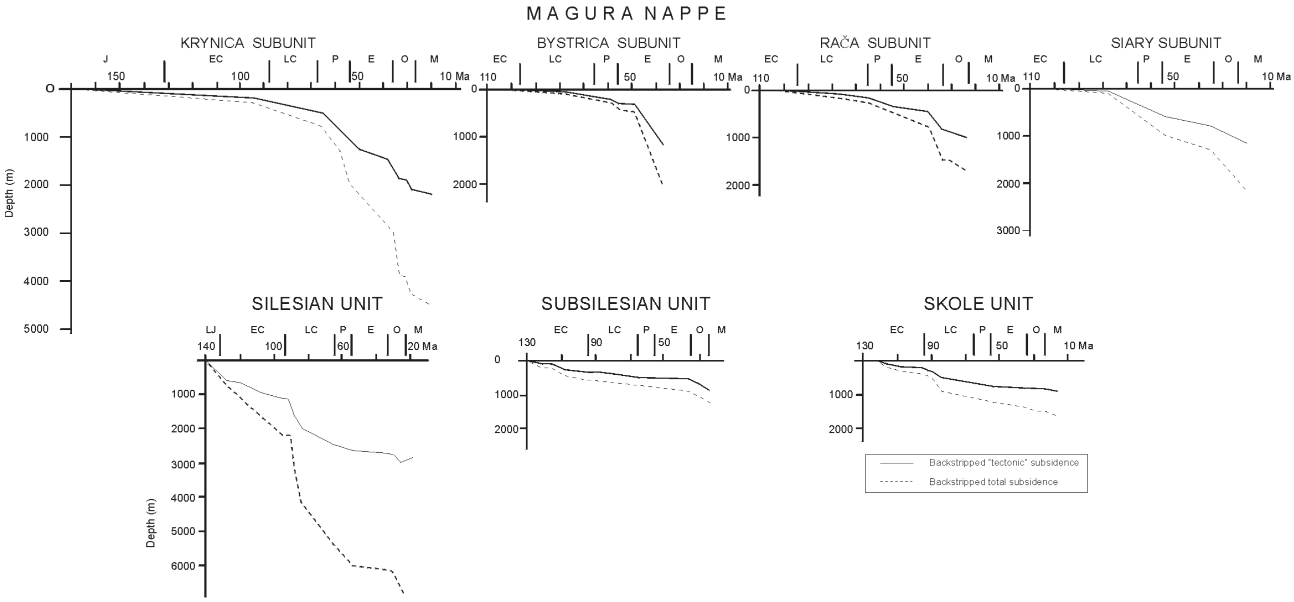

Fig. 2. Backstripped burial diagrams of the selected units of the Polish Outer Carpathians.

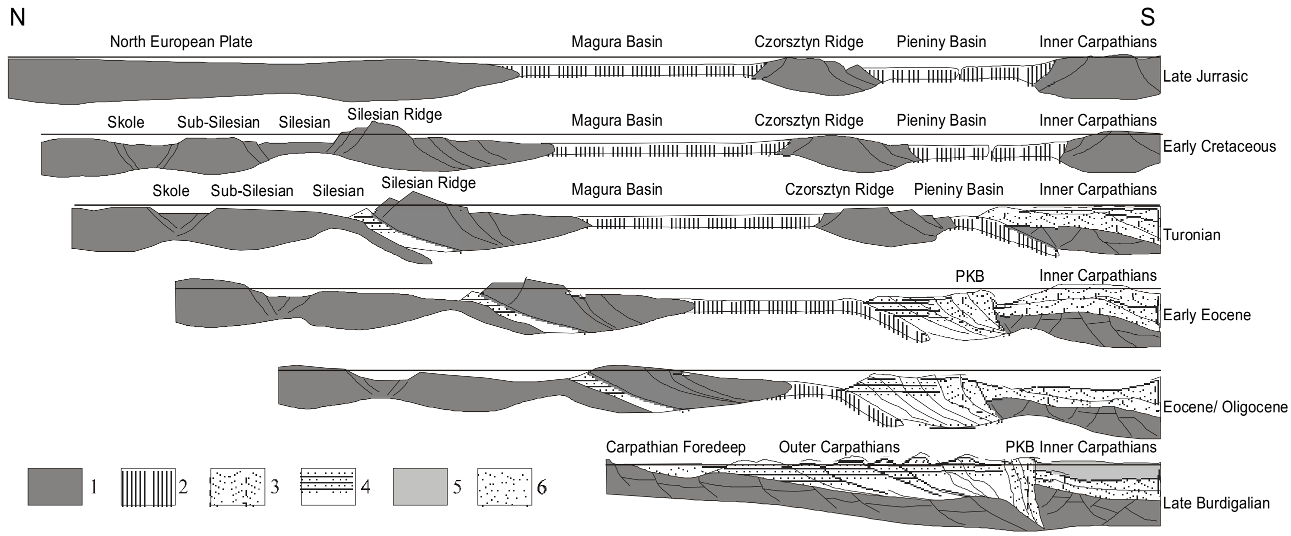

Fig. 3. Palinspastic evolution model of the Outer Carpathians (partly after Birkenmajer 1986). 1 — continental crust, 2 — oceanic crust, 3 — Inner Carpathian and Pieniny Klippen Belt (PKB) units, 4 — Outer Carpathian accretionary wedge, 5 — Podhale Flysch, 6 — molasses.

lobe system. At the turn of Eocene the deposition of the Globigerina marls was preceded by the uplift of the basin floor (1500 m ?). This was a combined result of the folding which occurred in the S part of the basin and a global drop in sea level. This event was followed by a Late Eocene to Oligocene subsidence and a deposition in the marginal part of the basin, which probably persisted up to the Early Miocene. In the Silesian basin, three periods of subsidence took place: Late Jurassic/Early Cretaceous, Late Cretaceous–Paleocene and Late Oligocene/Early Miocene (Fig. 2). The first subsidence was initiated at the turn of Late Jurassic through the rifting of the southern part of the European Plate (Fig. 3). From Tithonian to Aptian, the rate of total subsidence and the rate of sedimentation reached 50 m/Ma and, during the Albian to Cenomanian, decreased to 5 m/Ma. A new period of subsidence was initiated during the Turonian and was acompanied by an extremely high rate of sedimentation (up to 485 m/Ma), which decreased to 59 m/Ma in the Paleocene. The Eocene period was dominated by deep water deposition and a low rate of sedimentation (15 m/Ma). At the turn of Eocene, the floor of the Silesian basin was uplifted and a deposition of the Globigerina marls followed by Menilite shales was initiated. The last period of subsidence began during the Late Oligocene and persisted to the Early Burdigalian (208 m/Ma). In the Sub-Silesian and Skole basins, total subsidence was a few times lower than in the Silesian Basin and did not exceed 1.5 km (Fig. 2). The subsidence plots of the Outer Carpathian basins revealed important differences between the Magura and Silesian-Sub-Silesian-Skole basins.

From the remnant ocean basin to the collision-related peripheral foreland basin

The Outer Carpathians are regarded as the remnant ocean basin which developed between the colliding European continent and the intra-oceanic arcs (Birkenmajer 1986; Ingersoll et al. 1995). During the Late Jurassic, the Western Carpathian domain was located SE of the European epicontinental sea and NW of the Neotethys Ocean tip. The Magura deep-sea basin, situated south of the European shelf, was separated from the Pieniny deep-sea basin by the Czorsztyn submerged ridge (Fig. 3). The Pieniny basin and the Inner Carpathian domain was separated by the Penninic (Vahic) oceanic rift. At the turn of the Jurassic, in the southern part of the European shelf, the paleorifts were floored by a thinned continental crust (Birkenmajer 1986; Săndulescu 1986). This was a rifted European margin which was incorporated into the Outer Carpathian basin (the Skole, Subsilesian — and Silesian areas). The Late Jurassic-Early Cretaceous subsidence of the Silesian basin was controlled by normal fault and post-rift thermal subsidence, which culminated with the Albian-Cenomian deep water expansion. The Late Cretaceous-Paleocene subsidence and deposition in the Silesian basin were connected with a southward subduction of the Silesian basin and accompanied by the uplift of the Silesian ridge. This event was probably a continuation of the pre-Late Albian subduction of the Outer Dacides (Săndulescu 1986). In the Magura basin the Paleocene subsidence was related to the uplift of the PKB. The moving load, in front of the Magura and PKB accretionary prism, caused a subsidence and a migration of depocenters to the north (Fig. 3). As a result, narrow and long submarine fans, supplied from the south-east, were formed. The northern part of the basin was dominated by a hemipelagic and pelagic deposition and affected by subsidence at the turn of the Eocene. After the Late Oligocene folding, the Magura Nappe was thrust northwards onto the terminal Krosno flysch basin (Oszczypko 1998), and during the Burdigalian its front reached the S part of the Silesian basin. This was followed by a progressive migration of axes of subsidence towards the north. During the course of the Burdigalian trangression, part of the Magura basin was flooded and the seaway connection with the Vienna basin via Orava was probably established (Oszczypko et al. in print). During the Ottnangian, the Late Krosno basin shifted towards NE and finally underwent dessication. This was followed by the Intra Burdigalian phase, where the marginal part of the Outer Carpathians was folded, overthrust and uplifted. The Carpathians overrode the platform and caused flexural depression — a peripheral foreland basin related to the moving Carpathian front (Oszczypko 1998).

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HOW TO LOOP CARPATHIANS — AN ATTEMPT TO RECONSTRUCT MESO-CENOZOIC PALINSPASTIC HISTORY OF THE CARPATHIAN OROCLINE

D. PLAŠIENKA[1] and M. KOVÁČ[2]

[1]Geological Institute, Slovak Academy of Sciences, Dúbravska cesta 9, 842 26 Bratislava, Slovak Republic; geolplas@savba.savba.sk
[2]Department of Geology and Paleontology, Comenius University, Mlynská dolina G, 842 15 Bratislava, Slovak Republic; kovacm@fns.uniba.sk

Key words: Carpathians, Mesozoic, Tertiary, plate motion, kinematic reconstruction.

Introduction

Being wrapped by the Tertiary Outer Carpathian arc, the Carpathian interior consists of several segments (terranes, microplates) showing varying affinities to the Alps, Balkans and Dinarides. Alternatively, they show time-dependent relationships either to the North European Platform, or to the "African", Adria-related mobile belt. Though, the final position of these segments is Late Tertiary in age, they were mostly disconnected and then partly welded together during the Mesozoic. We present a tentative reconstruction, in which the movement history inside the Carpathian loop is partitioned into several principal evolutionary periods.

Constraints

There is little agreement among the Carpathian geologists on the interrelations of individual Carpathian segments. Several possible solutions of how to assemble the Tertiary collage were published, but the main disagreement concerns the Early Mesozoic positions of these segments with respect either to Europe, or to Adria-Africa. Our approach tries to corroborate confusing paleogeographic criteria by geometric and kinematic principles. Possible driving forces are also considered. The basic constraints, which are integrated in our reconstruction, are as follows:

— relatively stable position of Europe in respect to drifting Adria, adopting the movement path of the latter as reconstructed e.g. by Savostin et al. (1986), or Dewey et al. (1989);

— more-or-less stable position of Moesia in respect to Europe starting since the Early Cretaceous, although, its later lateral shifting along the Teisseyre-Tornquist zone could take place;

— Cretaceous-Tertiary paleomagnetic data for large blocks (Alcapa and Tisza-Dacia) inside the Carpathian orocline that define their translation and rotation histories (e.g. Balla 1984; Pătrascu et al. 1994; Mauritsch & Márton 1995; Túnyi & Márton 1996; Kováč & Márton 1998);

— extent of Tertiary shortening of the Outer Carpathian Flysch belt revealed by the paleogeographic considerations and by cross-section balancing (e.g. Oszczypko 1992; Roca et al. 1995);

— evolution of Neogene Pannonian back-arc basin and associated volcanism (e.g. Csontos et al. 1992; Csontos 1995; Kováč et al. 1997);

— paleostress data for Tertiary history (e.g. Csontos et al. 1991; Márton & Fodor 1995; Marko et al. 1995);

— progradation trends of Cretaceous shortening events in different circum-Pannonian segments (e.g. Pamić et al. 1998; Săndulescu 1989; Tollmann 1989; Plašienka 1997);

— inferred ages of opening and closure of intervening oceanic domains and position of their present sutures based on stratigraphic, sedimentologic and structural data (e.g. Michalík & Kováč 1982; Kováč et al 1989; Michalík 1994).

Reconstruction

We present the results of our reconstruction in 5 pre-Tertiary and 6 Tertiary evolutionary stages that are briefly characterized as follows:

1. Late Triassic (225–220 Ma, Carnian). Narrow oceanic basins, collectively referred to as the Meliatic ocean, opened within the southern marginal, poorly consolidated epi-Variscan European areas in the Middle Triassic. Opening of the Meliatic ocean is attributed to the back-arc rifting and an extension triggered by the northward subduction of Paleotethys. The Austroalpine, Slovakocarpathian and Tisia domains were still firmly attached to the stable North European Platform. The Adriatic microplate with its northern, South Alpine-Transdanubian-Bükkian-Dinaridic margin and the Moesian-Getic segment were the terranes drifting southward.

2. Early Jurassic (200–195 Ma, Sinemurian). Southward, partly intraoceanic subduction of the Meliatic ocean commenced in the latest Triassic, but the contraction belt was restricted to only a narrow accretionary wedge rimming the northern Adriatic margin. In contrast, both the northern passive and the southern active margins of the Meliatic ocean suffered from an extensive rifting during the Early Jurassic. This is tentatively attributed to the back-arc rifting of the upper plate, and the passive rifting of the lower European plate due to the southward pull exerted by the negative buoyancy of the Meliatic slab, augmented by the eastward drift of Adria.

3. Late Jurassic (160–155 Ma, Oxfordian). The Meliata-Hallstatt ocean was gradually closing, but the Szarvaskő back-arc basin opened in its place. Within the lower plate, the Southern Penninic (Ligurian-Piemont-Vahic) ocean spread out after the latest Liassic break-up, which separated the Austroalpine-Slovakocarpathian-Tisia terranes from the North European Platform. An immature orogenic wedge nucleated along the Meliata-Hallstatt suture, associated with HP/LT metamorphism and blueschists exhumation during the initial collision and with synorogenic sedimentation dominated by olistostromes. However, the northern lower plate was still in tension. This is ascribed to the B-subduction and persisting southward slab-pull of the Meliatic oceanic lithosphere attached to the European lower plate. Some other fragments were separated from the southern passive European margin, as the Briançonnais, Hochstegen (Tauern) and Oravic (Pieniny Klippen Belt) continental ribbons. The Northern Penninic oceanic branches, as the Valais, Ybbsitz and Magura, opened during the Late Jurassic and Early Cretaceous.

4. Mid-Cretaceous (110–105 Ma, Early Albian). From the Early Jurassic onward, the movement vector of Adria has been reconstructed by Savostin et al. (1986). For the Jurassic and Early Cretaceous a systematic E- to SE-ward drift of Adria with respect to Europe is inferred. Due to persisting subcrustal compressive load exerted by the Meliatic slab, the Meliatic suture-collision wedge widened, progressively incorporating the Austroalpine-Slovakocarpathian system during the mid-Cretaceous times. The wedge growth was accompanied by the crustal thickening and by eo-Alpine metamorphism. However, tensional stresses still dominated in the western sector of the European lower plate.

5. Late Cretaceous (70–65 Ma, Maastrichtian). The general geodynamic situation fundamentally changed during the Late Cretaceous. Adria began to move northwards, bulldozing in its front all the previously shortened or stretched continental, as well as oceanic realms. The detached Meliatic slab drowned in the mantle and did not affect the crustal dynamics any more. The Austroalpine-Slovakocarpathian system was welded to Adria and to its northern pendants (Transdanubian, Bükk), but these became separated from the Adria by wedging of Tisia along a large-scale dextral strike-slip, a precursor of the Periadriatic and mid-Hungarian lines. At the same time, Tisia was welded with the Getic and Bukovinian segments to form the Tisza-Dacia microplate. The amalgamated Austroalpine, Slovakocarpathian, Transdanubian and Bükk segments constituted the Alcapa microplate. Movement of both the Alcapa and Tisza-Dacia microplates governed the following Tertiary evolution of the area. During the latest Cretaceous, consumption of the Southern Penninic oceanic crust by the A-type subduction started along the northern margin of the Alcapa. Consequently, the compressive stresses exerted by indenting Adria were effectively transmitted to far distances towards the north, where they caused basin inversions, wrenching and local thrusting in the complexes of the North European epi-Variscan platform (Saxon folding).

6. Eocene (45–34 Ma). The northward drift of the Alcapa and Tisza-Dacia microplates continued, consuming the remnant Southern Penninic oceanic basins. However, the Austroalpine system had already collided with the Hochstegen high and shortening shifted to the Rhenodanubian trough. North of the analogous Oravic high, situated in the mouth of the "Carpathian gulf", the large Magura oceanic trough was still opened. Inside the gulf, the Silesian–Krosno basins and the Moldavian flysch troughs were situated on a thinned continental and/or oceanic crust, separated from the Magura trough by the intraoceanic Silesian ridge. Later on, A-type subduction led to a final closing of the Rhenodanubian and Vahic troughs. The initial stage of the development of the Outer Western Carpathian accretionary wedge — stacking of the inner Magura units — was synchronous with the foundation of the Central Carpathian Paleogene forearc basin, filled with thick turbidite prisms. Southwards, the Eocene epi-continental sequences deposited in the Buda basin of Alcapa showing a retro-arc position, and in the Szolnok flysch trough located in front of the Tisza-Dacia microplate.

7. Oligocene (34–23 Ma). Adria vs. Europe convergence led to the collision of the Alps with the North European Platform and was associated with the displacement of totally unrooted Rhenodanubian trough sediments. In front of the Western Carpathians, there followed B-type subduction of the Magura flysch through basement below the overriding Alcapa microplate. Detached sediments became part of the Outer Carpathian accretionary prism. On the North European Platform margin, the molasse deposits accumulated in front of the Alps. Eastwards, the isolation of the Silesian–Krosno and Moldavian flysch basins led to a "menilite" anoxic event. Epicontinental seas flooded the Alcapa microplate, as well as the northern margin of the Tisza-Dacia microplate.

8. Early Miocene (23–17.5 Ma). After collision in the Alps, the Alcapa and Tisza-Dacia microplates continued their generally northeastward drift, passing the gate of the "Carpathian gulf". The Magura nappe stack was detached and thrust over the Silesian high. Afterwards, the Outer Carpathian flysch basins attained features of relatively deep-water foreland basins with turbiditic sedimentation in axial parts (residual flysch basins), while molasse deposits accumulated on the platform margins. The Eggenburgian marine transgression flooded the uplifted parts of the Outer Carpathian accretionary prism, frontal parts of the Central Carpathian, as well as the northern margin of the Tisza-Dacia microplate.

9. Early/Middle Miocene (17.5–15 Ma). The Karpatian to Early Badenian time is characterized by an extrusion of the Alcapa lithospheric fragment northeastwards as a result of Alpine collision and subduction pull in front of the Carpathian orogen. Passing the "Carpathian gulf" gate led to a counterclockwise rotation of the Alcapa microplate (at the end of Ottnangian ca 40–50^{o} and at the end of the Karpatian 30^{o}), due to a corner effect of the Bohemian massif. Active thrust front of the Carpathian accretionary wedge shifted to the outer edge of Silesian-Krosno nappe stack, which was thrust over the Carpathian foredeep. Extension of the overriding, now amalgamated microplates led to initial rifting in the back-arc region. Lithospheric thinning initiated the asthenosphere upwelling, crustal heating, partial melting and acid to intermediate calc-alkaline volcanism. At the boundary between the Eastern Alps and the Western Carpathians, the Vienna pull-apart basin opened within a NE-SW trending sinistral wrench corridor. The East Slovak pull-apart basin opened between the Western and Eastern Carpathians along a NW-SE trending dextral wrench zone.

10. Middle Miocene (15–11.5 Ma). Subduction of the Silesian-Krosno-Moldavian basin induced an island arc volcanic activity in the Transcarpathian depression, related to melting of the downgoing plate. Extension in the back-arc region was associated with voluminous, calc-alkaline volcanic activity of the areal type.

11. Late Miocene (11.5–7 Ma). Evolution of the Carpathian loop came to its final stage. Active subduction is reported only from the Eastern Carpathians during this time. Similarly as in the Western Carpathians, it is associated with calc-alkaline volcanism in the Transcarpathian and Transylvanian basins. At the beginning of this period, the Pannonian back-arc basin system was structurally rebuilt by wide rifting and passed into a post-rift thermal subsidence stage.

Conclusions

The results of our reconstruction should be considered as tentative subject of continuous complementation and strengthening by new data and constraints. One of the main conclusions to be emphasised is that the motion and deformation of crustal segments within the Carpathian collage seems to be driven, in addition to global plate movements, as was the Africa-Adria drift, also by the locally generated body forces. These were dominantly exerted on the adjacent plates by descending oceanic slabs.

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THE QUARTZ AND CALCITE X-RAY TEXTURE GONIOMETER PATTERNS FROM THE WESTERN CARPATHIANS CRETACEOUS DUCTILE SHEAR ZONES USED AS KINEMATIC INDICATORS

M. PUTIŠ[1], Z. NÉMETH[2], W. UNZOG[3]
and E. WALLBRECHER[3]

[1]Department of Mineralogy and Petrology, Faculty of Science, Comenius University, Mlynska dolina, 842 15 Bratislava, Slovak Republic; putis@fns. uniba. sk
[2]Geological Survey of Slovak Republic, Werferova 1, 040 11 Košice, Slovak Republic
[3]Institute of Geology and Paleontology, K.F. University of Graz, Heinrichstr. 26, A-8010 Graz, Austria

Key words: quartz, calcite, X-ray patterns, Western Carpathians, Cretaceous, shear zones.

Introduction

Major tectonic boundaries of the Western Carpathians as well as their continuation below the cover and tectonic nappe complexes, were characterized and discussed by Plašienka et al. (1997). The early Cretaceous thrust-faults (Fig. 1) enhanced the exhumation of ductilely deformed basement and cover rocks (Putiš 1991, 1992, 1994). Most of them are sinistral transpressive shear zones separating the thick-skinned crustal sheets. The North Veporic and South Veporic sheets are divided by the Pohorelá shear zone, while the South Veporic and Gemeric are separated by the Ľubeník shear zone. The latter shear zone also manifests a post-collisional, extensional collapse sliding. The third shear zone, called Čertov-

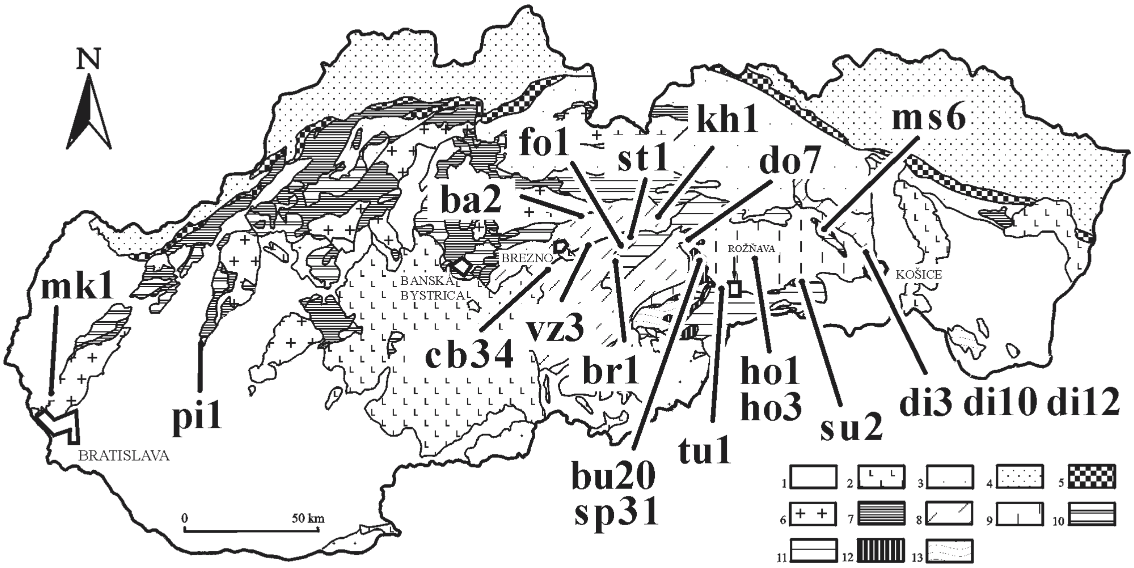

Fig. 1. Location of studied samples. 1 — Neogene sediments, 2 — Neogene volcanics, 3 — central Paleogene sediments, 4 — outer flysch sediments, 5 — Klippen belt, 6 — Tatric basement and cover, 7 — Fatric Krížna nappe, 8 — Veporic basement and cover, 9 — Gemeric basement and cover, 10 — Hronicum, 11 — Silicicum, 12 — Meliatic Permo-Triassic complex, 13 — Turnaicum and Zemplinicum.

ica shear zone, separates a zone of collision-related thick-skinned units (the Veporic) from a thin-skinned (the Supra-Tatric) unit. These zones and partial domains are well documented by the zoning of Cretaceous collision metamorphism in the central Western Carpathians (Korikovsky et al. 1997a,b) that reached a maximum in the lower amphibolite facies ($St_{Zn\ 2.42}$-Grt-Chl-Cld-Ky zone according to Korikovsky et al. 1989; Méres & Hovorka 1991; Krist et al. 1992, or Kováčik et al. 1997) with 8–10 kbar pressure conditions achieved in some basement fragments, later exhumed along the major mid-Cretaceous shear zones (Putiš 1991; Putiš et al. 1997; Dallmeyer et al. 1996).

We used the ductile shear zones as reliable kinematic indicators of the Alpine Western Carpathian tectogenesis. Moreover, we studied three samples of Early Paleozoic marbles (Gemeric Holec Beds) for comparison. Their recrystallization during Variscan orogeny is widely accepted.

Methods

Oriented samples cut parallel to the XZ plane were measured with a Siemens D500 reflection X-ray texture goniometer at the Institute of Geology and Paleontology of K.F. University of Graz and obtained data were used to calculate pole figure patterns (Figs. 2, 3).

Results

Malé Karpaty Mts.: the Borinka shear zone. Very low temperature (less than 300 ^{o}C) NW-vergent thrust of the Tatric basement/cover (Bratislava Nappe) over Jurassic limestones (analysed sample mk1-cc) of the Infratatric Borinka unit. Indistinct maxima and very fine-grained ultramylonitic matrix indicate a pressure-solution and/or superplastic flow instead of intracrystalline plasticity.

Považský Inovec Mts.: Hlohovec shear zone. Very low temperature, NW-vergent thrust of the Tatric basement/cover complex over the Jurassic limestones (sample pi1-cc) of the Infratatric unit in a small tectonic window. Flattened grains in fine calcite aggregate in marble show one broad distinct maximum coeval with the fine schistosity, but an ambiguous lineation.

Eastern Low Tatra Mts.: Čertovica shear zone. Very low temperature, NW-vergent thrust of the North-Veporic over the mid-Triassic limestones of the Tatric cover (sample ba2-cc) in the Bacúch tectonic window. Both microfabric and texture pattern are very similar to those from the Považský Inovec Mts.

Northwestern Vepor Mts.: Pohorelá sinistral transpressive shear zone. Low temperature (about 400 ^{o}C) top-to-the NE thrusting is indicated by quartz asymmetric patterns of the sheared basement rocks (samples vz3-qtz, cb34-qtz), with basal <a> and prism <a> slip dominating. The quartz microstructure is characterized by polygonal grains produced during dislocation recovery processes and during dynamic recrystallization through a subgrain rotation mechanism.

Northwestern Vepor Mts.: Kráľova Hoľa and Fabova Hoľa extension shear zones. The low temperature (400–500 ^{o}C) top-to-the SE sliding indicates a change from transpression to a transtension/extension regime, related to uplift. Exceptionally, prism <a> slip is dominated in the rod-shaped quartz from the basement phyllonite (sample kh1-qtz). In most cases a combined basal <a> and prism <a> slip is characteristic (sample br1-qtz, Lower Triassic quartzite). The calcite marble patterns (samples fo1-cc, st1-cc, mid-Triassic limestones) reflect flattened grains with distinct narrow e-lamellae, surrounded by dynamically recrystallized medium- to fine-grained aggregates.

Southeastern Vepor Mts.: Ľubeník extension shear zone in the Dobšiná half-window. The low temperature normal fault contact of the South-Veporic and Gemeric units exhibits deformed calcite fabrics (sample do7-cc, mid-Triassic limestone) which resemble those of the Fabova Hoľa Massif.

Spiš-Gemer Ore Mts.: Bôrka nappe. The soles of the outliers are documented in the *western Gemericum* (sample bu20-qtz, Permo-Scythian conglomerate; sample sp31-cc, Triassic limestone), in the vicinity of the glaucophanite belt in *southern Gemericum* (sample su2-cc, Triassic limestone), as well as below the Murovaná skala Mountain Hills in the *eastern Gemericum* (sample ms6-cc, Triassic limestone, Németh 1996). The samples show a uniform, low temperature, N-ward thrusting of the exhumed northern marginal Permian-Triassic sediments of the Meliata(-Hallstatt) ocean in a transpressional-transtensional regime.

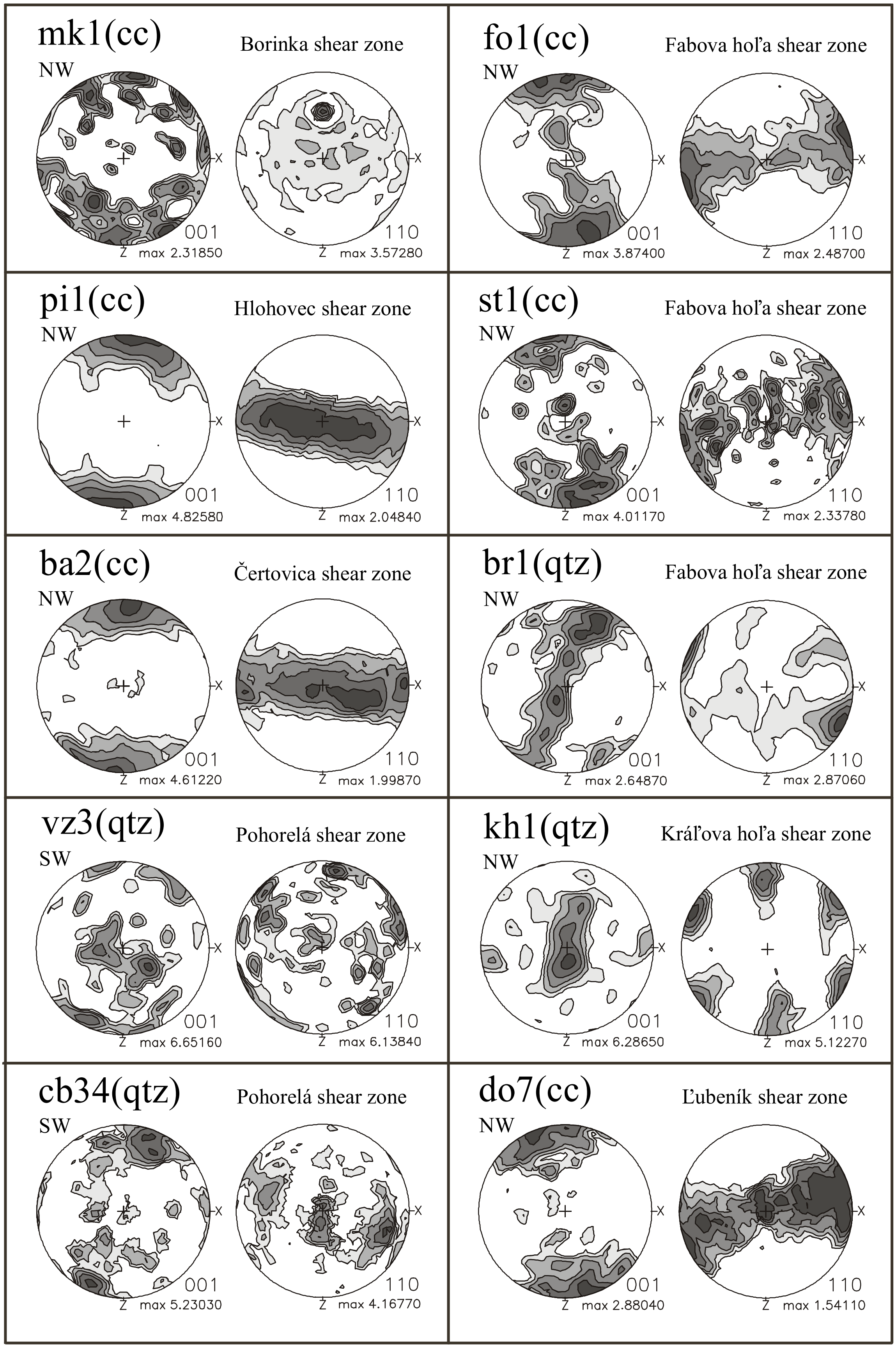

Fig. 2. (explanations in text).

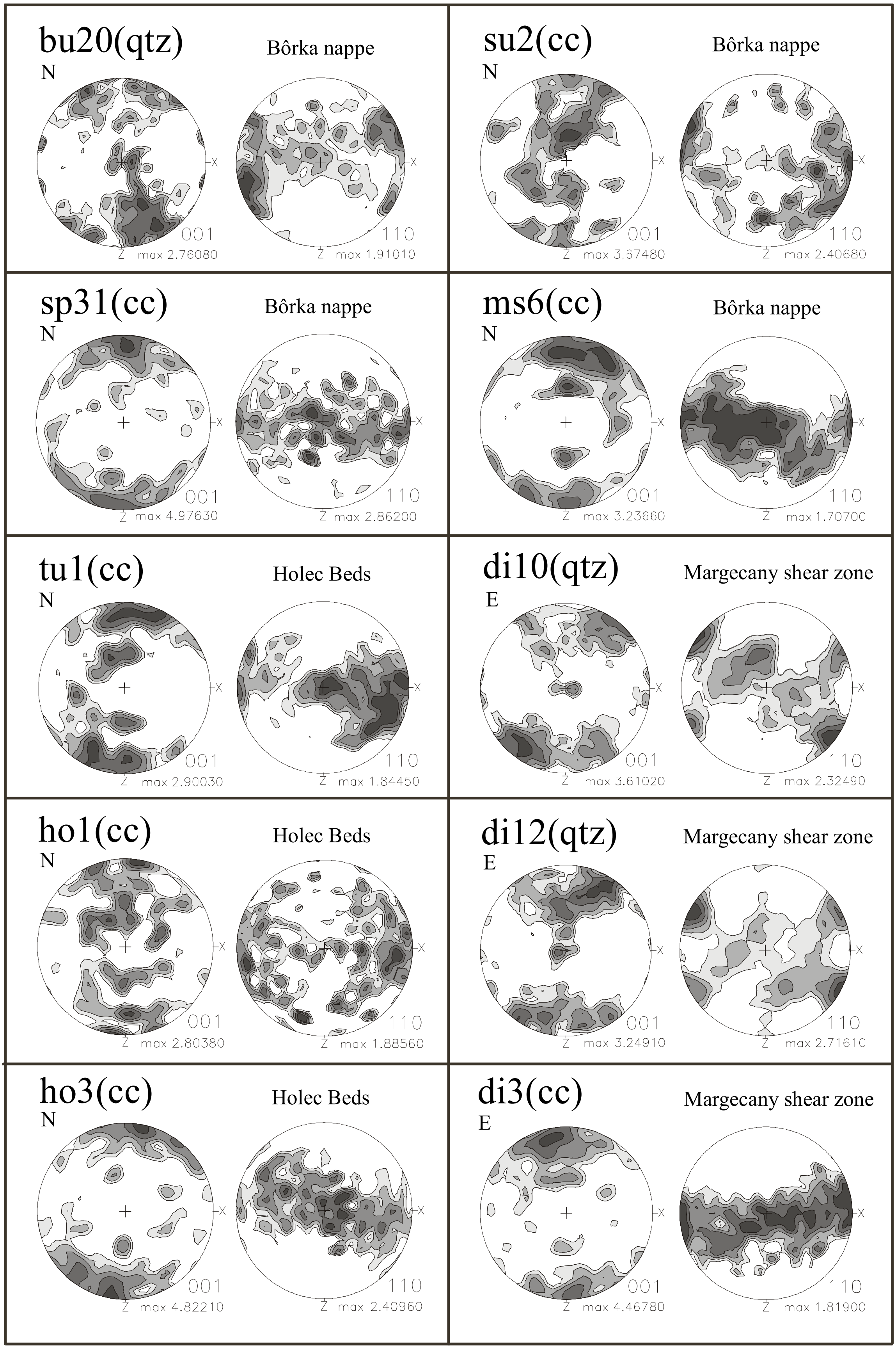

Fig. 3. (explanations in text).

Čierna hora Mts.: Margecany shear zone at the eastern contact zone of the Gemericum with the Veporicum. The quartzites contain dynamically recrystallized flattened grains. The calcite marbles show dynamically recrystallized layers alternating with layers of flattened, elongated twinned grains. Pole figures (samples di10-qtz, di12-qtz, Lower Triassic quartzite; sample di3-cc, mid-Triassic limestone) indicate a low temperature top-to-the W thrusting.

Spiš-Gemer Ore Mts.: Holec Beds marbles. These marbles are interpreted as Silurian carbonates (Grecula 1982; Bajaník & Vozárová eds. 1983) metamorphosed during the Variscan orogeny. The pole figures (samples tu1-cc, ho1-cc, ho3-cc) show remarkable similarities with those of the Triassic marbles, indicating their superimposed Alpine deformation.

Conclusions

The first quartz and calcite X-ray texture goniometer patterns from the Western Carpathian Cretaceous ductile shear zones indicate a mineral rheology caused by dislocation creep and/or mechanical twinning, thus reflecting the deep-crustal (thick-skinned) reactivation of the Veporic/Gemeric basement/cover- and Meliatic sedimentary-volcanic complexes. They precise the kinematics of the collisional nappe stacking, lateral transpression and extensional collapse sliding within the exhumed early Alpine (Cimmerian) orogen internides. In general, the textural patterns of the South-Veporic structural complex reflect hot conditions of the late Cretaceous extensional uplift and a normal faulting. The other textural patterns, from the North-Veporic, Gemeric and Meliatic units, are mainly the results of compressional early Cretaceous thrust tectonics.

On the other hand, very low temperatures of the Infratatric and Supratatric structural domains are only coeval with the calcite plasticity in the (thin-skinned) foreland. The textural patterns reflect the Cretaceous thrusting.

The rock-samples from the Gemeric Silurian Holec Beds, with assumed Variscan recrystallization are deformed under the same mechanisms as those from the Cretaceous ductile shear zones. Therefore we infer that their final deformation/recrystallization is of Alpine age.

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ARC FORMATION IN THE WESTERN ALPS AND CARPATHIANS: A COMPARISON

S.M. SCHMID

Geologisch-Paläontologisches Institut, Bernoullistr. 32, CH-4056 Basel, Switzerland; schmids@ubaclu.unibas.ch

Key words: Arc formation, orocline, indentation, collision, roll-back, Alps, Carpathians.

In map view the arc of the Western Alps and the arc of the Southern Carpathians, linking Carpathians and Balkan mountains, look strikingly similar. Both these arcs are convex towards the west and both surround a relatively stable block situated on the concave side (Adriatic microplate and Moesian platform, respectively). Western and Eastern Carpathians define a third arcuate structure convex towards the east. This contribution discusses the mechanism and timing of arc formation in the Alpine-Carpathian double loop, pointing out similarities and differences.

We first address the arc of the Northern and Eastern Carpathians which is by now relatively well understood. It is widely accepted that arc formation in the Western and the northern part of the Eastern Carpathians is primarily driven by subduction roll-back, contemporaneous with back-arc extension in the Pannonian basin during the Miocene. Migration of shortening within the Silesian-Moldavian flysch belt and migration of back-arc volcanism in time and space suggests that slab break-off may have laterally migrated (see Wortel & Spakman 1992) during the Miocene from west to east, being in its terminal stage in the seismically active Vrancea area, situated in front of the southernmost Eastern Carpathians. Many authors also agree that this roll-back is associated with the subduction of a last remnant of oceanic lithosphere adjacent to the western margin of the East European continent. Hence, closure of this oceanic remnant is not driven by the overall Africa-Europe plate convergence but rather by the negative buoyancy of the subducted slab, rollback inducing extension within the Tisia and Alcapa blocks and E-ward migration of their eastern margins. Consequently, the Tisia and Alcapa blocks, i.e. the more internal units of the Carpathians structured during earlier (pre-Miocene) phases of deformation, invade the former oceanic embayment within the East European continent. Contemporaneous with Miocene shortening in the arcuate Silesian-Moldavian flysch belt, the Tisia and Alcapa blocks not only undergo extension (Pannonian basin) but also significant rotations of opposite sign north and south of a very complex Mid-Hungarian belt, as is documented by paleomagnetic data. The significant clockwise rotation of the internal units of the Carpathians and Apuseni mountains south of the Mid-Hungarian belt, i.e. the Tisia block (60^{o} since the Oligocene and more than 90^{o} since the Cretaceous; Patrascu et al. 1994) is instrumental for understanding arc formation in the Southern Carpathians, by far not so well understood.

According to a working hypothesis recently proposed by Schmid et al. (1998) arc formation in the Southern Carpathians and the transitional area into the Balkans is essentially due to the corner effect of Moesia, which is part of the East-European foreland ever since it was welded to the East European craton across the Dobrogea mountains in Early Cretaceous times. According to this working hypothesis, large parts of the Tisia block (Bucovinian-Getic, Supragetic and Serbo-Macedonian units) were part of the so-called Rhodopean fragment (Burchfiel 1980) which extends into the Balkan and Rhodopean mountains and which was separated from the Moesian platform by the Severin ocean. During the Eocene, the western part of this Rhodopean fragment moved north, past the western edge of the Moesian platform. This differential northward movement is initially associated with dextral transpression within the future Southern Carpathian units situated in an area west of the Moesian platform at this time. As the Rhodopean fragment started to invade the oceanic embayment north of Moesia during the Late Eocene, it underwent some 60^{o} clockwise rotation, together with the rest of the Tisia block, the pole of rotation being situated within the Moesian platform. This rotation is associated with substantial orogen-parallel extension which led to core-complex formation in the Danubian units of the Southern Carpathians. This orogen-parallel extension was immediately followed by dextral strike slip along the curved Timok and Cerna faults during the Oligocene. Miocene E-W shortening in the Eastern Carpathians is associated with relatively minor dextral strike slip movements along an E-W oriented corridor within the present-day South Carparthians. In particular, the amount of Miocene N-S compression across the South Carpathians is neglegible, the foreland of the Getic depression representing an extensional rather than a compressional sedimentary basin. In summary, the arc of the Southern Carpathians is the result of true oroclinal bending of a pre-existent Cretaceous nappe stack around the western edge of Moesia. The former existence and consumption of an oceanic embayment in present-day Transylvania was instrumental for oroclinal bending in the South Carpathians, which already started during the Eocene.

While the Moesian platform remained firmly attached to the East European craton during orocline formation in the South Carpathians, the situation is different in the Western Alps, where the Adriatic microplate acted as an independently moving indenter. The formation of the Western Alpine arc was initiated by head-on Paleocene-Eocene collision of the north moving Adriatic micro plate in the — later to become — Central and Eastern Alps with the concave European plate boundary zone. This led to sinistral transpression and oblique collision in the future arc of the Western Alps until the Eocene. A recent re-interpretation of the ECORS-CROP transect of the Western Alps (Schmid & Kissling submitted) allowed to establish a kinematic model for ESE–WNW directed crustal shortening in the Western Alps which essentially is the result of post-collisional movements during the past 35 Ma. After 35 Ma ago the arc of the Western Alps formed as the result of WNW-directed movement and anti-clockwise rotation of the Adriatic microplate, leading to indentation of the Penninic realm by the Ivrea geophysical body. Dextrally transpressive movements with some 100 km diplacement along the Insubric (Tonale) and the Rhone-Simplon line were instrumental for Oligo-Miocene WNW-directed movement of the Adriatic microplate. As in the South Carpathians, arc formation was accompanied by substantial orogen-parallel extension which, in case of the Alps, led to the unroofing of the Lepontine metamorphic dome. In summary, arc formation in the Western Alps is the result of a change in plate motion of the Adriatic microplate during convergence and collision. In contrast to the internal Carpathians, arc formation in the Western Alps is not the result of oroclinal bending of a pre-existent (Cretaceous) orogen.

In all the three cases discussed above, arc formation is pre-determined by the geometry of former oceanic domains. However, while arc formation is the result of a two-stage "hard collision" in the Western Alps during the Tertiary, the Eastern and Southern Carpathian double loop results from subduction roll-back of a former oceanic domain towards the east, associated with the invasion of the Tisia block around the western corner of Moesia into this former oceanic realm and "soft collision" with the East European craton. Miocene shortening in the Silesian-Moldavian flysch belt only represents the last step of a longlasting process of true oroclinal bending of the more internal units belonging to the Rhodopean fragment in the South Carpathians, a process which already started during the Eocene.

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AN EARLY HISTORY OF THE OUTER CARPATHIAN BASIN

A. ŚLĄCZKA, N. OSZCZYPKO, E. MALATA and M. CIESZKOWSKI

Institute of Geological Sciences, Jagiellonian University, 30-063 Kraków, Oleandry 2a, Poland; slaczka@ing.uj.edu.pl

Key words: Outer Carpathians, Late Jurassic–Early Cretaceous, rifting, subsidence, flysch sedimentation.

Introduction

The Outer Carpathian Basin (OCB) was a part of the Alpine domain system (Fig. 1). It developed during the Late Jurassic-Early Cretaceous rifting and lasted till the Early Miocene tectogenesis. In a northerly direction the OCB consisted of the Magura, Dukla and Fore-Magura, Silesian, Subsilesian and Skole sub-basins. However, during the OCB evolution some of those sub-basins were temporarily united.

New data imply that the OCB could have started to evolve as early as the middle part of the Jurassic. The isotopic age of andesite blocks within the Żegocina tectonic window (Ślączka 1998) suggests their Middle Jurassic age. The U/Pb dating of idiomorphic crystals of zircon indicates the isotopic age of 144 +41/–45 Ma, whereas the K/Ar dating of biotite is 179 ± 4.7 Ma (Aalenian/Bajocian; M. Banas, pers. com). The extrusion can reflect earlier Alpine events of the southern margin of the North European Plate (NEP).

Results

The Late Jurassic through Early Cretaceous period of the OCB's evolution, when the continuous flysch sedimentation began, can be divided into four stages: 1 — distinct rifting of the southern part of the North European Plate (NEP); 2 — downwarping of this part of the NEP; 3 — compression and intensive sedimentation; 4 — second downwarping.

During the **first** stage (Kimeridgian–Valanginian) an incipient, relatively narrow basin developed, the flysch sedimentation was initiated, and the embryonic Silesian basin (Morava, Cieszyn-Bielsko area) was formed. That basin was probably prolongated into the Eastern Carpathians (Ukraine and Romania). These incipient basins were generally filled with dark marls followed by calcareous turbidites derived from the adjacent carbonate platforms. On the uplifted platform areas different shallow marine, mainly Stramberk type, or pelagic calpionella limestones were deposited. The regional subsidence was caused by the rifting of NEP accompanied by the extrusion of both basic lava (teschinites) in the Western Carpathians and diabase- melaphyre within the Black Flysch of the Eastern Carpathians (Săndulescu 1974). During that stage the subsidence reached as much as 69 m/Ma in the western part of the Polish Outer Carpathians. The foraminiferal assemblages, consisting mainly of calcareous benthic taxa showing some affinities with the Jurassic microfauna of the epicontinental sea, suggest the neritic/bathyal depth at the beginning of this stage. The increasing amount of the agglutinated taxa in time implies gradual deepening of the basin.

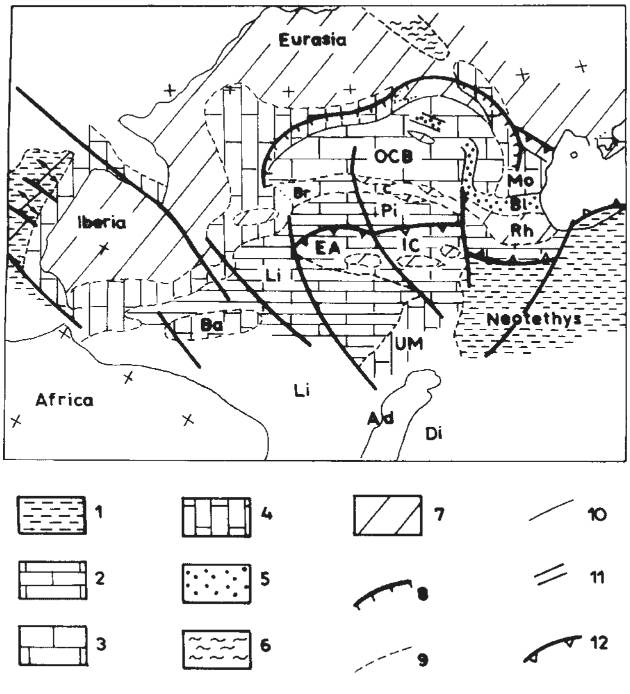

Fig. 1. Paleogeographic sketch of the Alpine realm during the Tithonian/Beriassian (based on J. Golonka 1997, partly changed). **1** — oceans, **2** — bathyal calcareous sediments, partly on oceanic crust, **3** — mainly neritic/bathyal calcareous sediments, **4** — mainly shallow water/neritic limestones, **5** — calcareous turbidites, **6** — brackish sediments, **7** — continents and islands, **8** — present day, outer boundary of the Alpine orogen, **9** — boundary of lithofacies, **10** — faults, **11** — rifts, **12** — contemporaneous overthrusts. **Mo** — Moesia, **Bl** — Balkan, **Rh** — Rhodopes, **C** — Czorsztyn ridge, **Pi** — Pieniny basin s.l., **IC** — Internal Carpathians, **EA** — Eastern Alps, **H** — Helveticum, **Br** - Briançonnes, **Li** — Ligurides, **Um** — Umbria, **Ad** — Adria, **Di** — Dinarides, **Ba** — Balearian.

The **second** stage (Hauterivian–Barremian — and Aptian in the western, marginal part of the OCB) was connected with general downwarping of the southern, inner part of the Silesian sub-basin, probably due to the cooling effect of the underlying lithosphere. A marine flooding was connected with this phenomenon and on the whole area of the OCB marine sedimentation took place. This stage is characterized by the occurrence of dark, silty, siliceous shales in the almost whole basin (Verovice Shales, Spas Shales) with dark calcareous shales (Hradiste shales). Along the northern margin of the Carpathian basin several small submarine fans (Hradiste sandstones) were locally formed. During the second stage the rate of subsidence decreased to 42 m/Ma in the western part and 12 m/Ma in the eastern part of the Polish Outer Carpathians. Only locally, where submarine fans were formed, rate of subsidence reached value of the 63 m/Ma. The depth of the basin deepened to bathyal/abbysal paleoenvironments.

The **third** stage began in the south-eastern part of the Carpathians during the Late Barremian-Aptian compressional events (Săndulescu 1984). That intensive folding, accompanied by the deposition of coarse clastic sediments, was completed by the Albian period. This diastrophism shows polarity and a westwards migration of an orogenic wave along the Carpathians. However, those compresional events can be traced along the Marmaros and Rachov units up to Svalava during the Aptian–Albian time (Kruglov 1986). It was also marked there by syn-orogenic facies (Soymul Beds) and in the more distal part by thick turbidite complexes (Biela Tisa and Upper Shipot beds). In the Western Carpathians this period of compression is manifested by the uplifting of the intra-basinal ridges within the OCB (Silesian and Subsilesian cordilleras), deposition of siliciclastic turbidites (Lower Lgota Beds, Gault flysch) and development of synsedimentary folds. However, the presence of cracks on the sea-bottom, in front of the compressed area, filled with overlying clastic material, could be a tensional effect of this diastrophism in the marginal part of the Polish Outer Carpathians. During the third period generally two lithofacies were distributed within the OCB however, that diferentation started already during the second stage. In the inner part of the Outer Carpathians the Albian deposits are represented by grey, often calcareous sandstones and marly shales (Biela Tisa and Suchov-Porkulec beds). The same deposits are reported from the southern part of the Magura Nappe in the Moravia (Švábenická et al. 1997). In the Magura Succession of the Pieniny Klippen Belt (Grajcarek Unit) the Albian deposits are represented by the radiolarian shales and spotty marls. Within the deposits of the outer part of the OCB black siliceous shales with intercalations of turbidites (Shipot Beds and their equivalents, Spas, Lgota beds and Gault flysch noticed in the northern part of the Magura Basin) prevailed. In the northern part of the basin locally grey, calcareous shales with gaize turbidites (Gaize Beds) were formed. The acceleration of the subsidence (31–63 m/Ma) is characteristic for this stage. The paleobathymetry of the basin was similar to the previous stage with the exception of the Subsilesian sub-basin where the basin depth corresponded to middle bathyal.

During the **fourth** stage (uppermost Albian–Cenomanian/Turonian) all the source areas of siliciclastic material ceased to be active and generally uniform pelagic sedimentation started: green, radiolarian shales with radiolarites in more outer part of the OCB, followed by red shales. At the end of this stage, during the Turonian a distinct change in structural plan of the OCB commenced. At the beginning of this stage the rate of sedimentation radically decreased to 4–6 m/ Ma, whereas the paleobathymetry reached abyssal depths.

Conclusions

The opening of the OCB at the end of Jurassic was connected with the rifting and submarine extrusion of basic lava. The upraise of the mantle caused subsidence of the OCB and development of normal fault system. It was followed by thermal subsidence and then by the strong compressional events in the Eastern Carpathians, terminating westwards. These stages of development of the OCB were clearly reflected in its depositional history.

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ALPINE PENNINICS IN THE EASTERN SLOVAKIA: FROM CRUSTAL UPDOMING TO BASIN DOWNFAULTING

J. SOTÁK[1], A. BIROŇ[1], I. DUNKL[2], R. PROKEŠOVÁ[3], J. MAGYAR[4], R. RUDINEC[4] and J. SPIŠIAK[1]

[1]Geological Institute, Slovak Academy of Sciences Bratislava, Severná 5, 974 01 Banská Bystrica, Slovak Republic; sotak@gu.bb.sanet.sk
[2]Institut für Geologie, Universität Tübingen, Sigwartstrasse 10, Tuebingen, Germany
[3]Faculty of Science, Matej Bel University, Tajovského 40, 974 01 Banská Bystrica, Slovak Republic
[4]Nafta Gbely a.s., PŤS Michalovce, Priemyselná 6, 071 01 Michalovce, Slovak Republic

Key words: East Slovak Basin, Penninics, underplating, exhumation, extensional detachment, core complex, basin subsidence.

Introduction

The Western Carpathians differ from the Alps inasmuch as they lack crustal-scale domes of the Penninic units, which were exhumed in response to the Alpine collision (Genser & Neubauer 1989; Ratschbacher et al. 1989; Severstone 1988; Behrmann 1988; Wallis & Behrmann 1996; Fügenschuh et al. 1997, etc.). Contrary to the Alps, the Penninic-like units in the Western Carpathians should be cored in the areas of extreme crustal extension. The easternmost exposure of the Penninic units occurs in the Rechnitz core complex, which was exhumed by the extensional detachment faulting (Tari et al. 1992). Similarly, the shallow-crustal elevation of the Penninic units was interpreted to participate in the basement structure of the Danube Basin (Lankreijer et al. 1995). The East Slovak Neogene Basin is also floored by the Penninic-like series of the Iňačovce-Krichevo Unit (Fig. 1). This unit appears to be a metamorphic core complex, exhumed jointly due to back-arc extension of the East Slovak Basin (Kováč et al. 1995). This paper provides a petrological, structural and geochronological data which allow us to interpret the processes of core complex updoming and basin downfaulting.

Metamorphic and structural geology

Mineral assemblages were used to estimate physical conditions of metamorphism in the Iňačovce-Krichevo Unit (Fig. 2). The peak metamorphic conditions were inferred from the assemblages: (1) biotite + actinolite and/or magnesioriebeckite + chlorite + titanite + epidote (metabasalts), (2) muscovite + quartz + pyrophyllite + paragonite + intermediate Na-K micas + chloritoid (Al-metapelites), and indicate metamorphic temperatures between 350 and $400\,^{o}C$. Co-existence of Na and Ca amphibols is considered here as a relic of an earlier, higher pressure metamorphic event (greenschist to blueschist transition zone, $p \approx 7$–8 kbar). During a decompressional phase of metamorphism normal greenschist assemblages formed (chloritoid in metapelites and biotite in metabasalts) at pressures < 5 kbar (Bucher & Frey 1994).

The youngest sediments of the Iňačovce-Krichevo Unit are Middle Eocene black phyllites and metasandstones. The phyllosilicate "crystallinity" and the coal rank indicate that their degree of metamorphism corresponds to higher anchizone, or to lower epizone, respectively ($IC = 0.31^{o}\Delta 2\Theta$, $ChC_{(002)} = 0.26^{o}\Delta 2\Theta$, $R_{o\,max} = 5.75$ %).

The Iňačovce-Krichevo Unit reveals a complex polydeformational history. Progressive deformation proceeded from (1) underthrusting — soft sediment deformation, stratal disruption and bou-

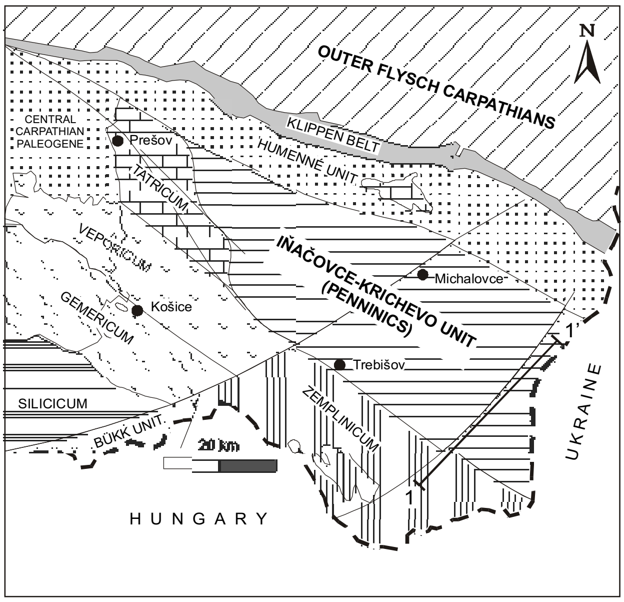

Fig. 1. Geological structure of the Pre-Neogene units in the eastern part of the Western Carpathians. Profile 1–1′ — interpreted seismic line.

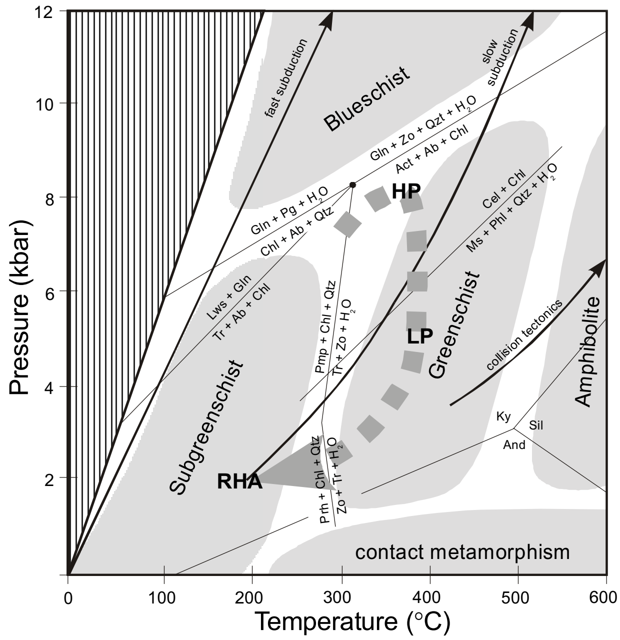

Fig. 2. Preliminary p-T path showing metamorphic evolution of the Iňačovce-Krichevo Unit. Metamorphic facies as well as model reactions are taken from Bucher & Frey (1994). HP — higher pressure event, LP — lower pressure event, PHA — post-metamorphic hydrothermal alteration.

dinage of high-competent layers, overpressured conditions, (2) underplating and deep tectonic burial — high flattening strain, F_1 foliation, synkinematic crystallization, intrafolial folding, diffusional mass transfer, crystalplastic deformation, (3) subcretion and intrawedge shortening — crenulation cleavage as F_2, transpositional foliation, high-strain zones, ultramylonites, δ-type porphyroclasts, open to tight F_2 folds, dynamic recrystallization, etc., and (4) updoming and extensional unroofing — shear bands, SC foliation fabric, kink-bands, en-echelon structures, extensive veining, cataclastic deformation, brecciation, normal faulting, etc.

Fission-track dating

The samples from the Iňačovce-Krichevo Unit gave significantly younger zircon FT ages than their sedimentation ages. This rather narrow range (with a mean of 20.1 ± 0.9 Ma) can be considered as a cooling age after a Neogene metamorphic event that caused total resetting of the zircon FT ages. The similarity between the white mica K/Ar ages and the range of zircon FT ages indicates a rapid cooling during Early Miocene. The Neogene synrift sediments that overlie the metamorphic basement did not suffer from significant post-depositional overprint and their zircon FT ages can be interpreted as a typical cooling ages of the source regions, but not the Iňačovce-Krichevo Unit.

Core complex history

The East Slovak core complex occurs in the area of strike-parallel wrench zone. Therefore, the exhumation of the Iňačovce-Krichevo Unit could be iniciated due to buoyancy and ductility of underplate rocks, updomed within the wrench zone. At that time (Upper Oligocene) the compressed wedge above the Iňačovce-Krichevo Unit became highly active as a sediment source for the Central Carpathian Paleogene fans. Since Early Miocene, the main factor controlling the exhumation was the extensional unroofing (Fig. 3). The extensional formation of the core complex structure is evidenced by

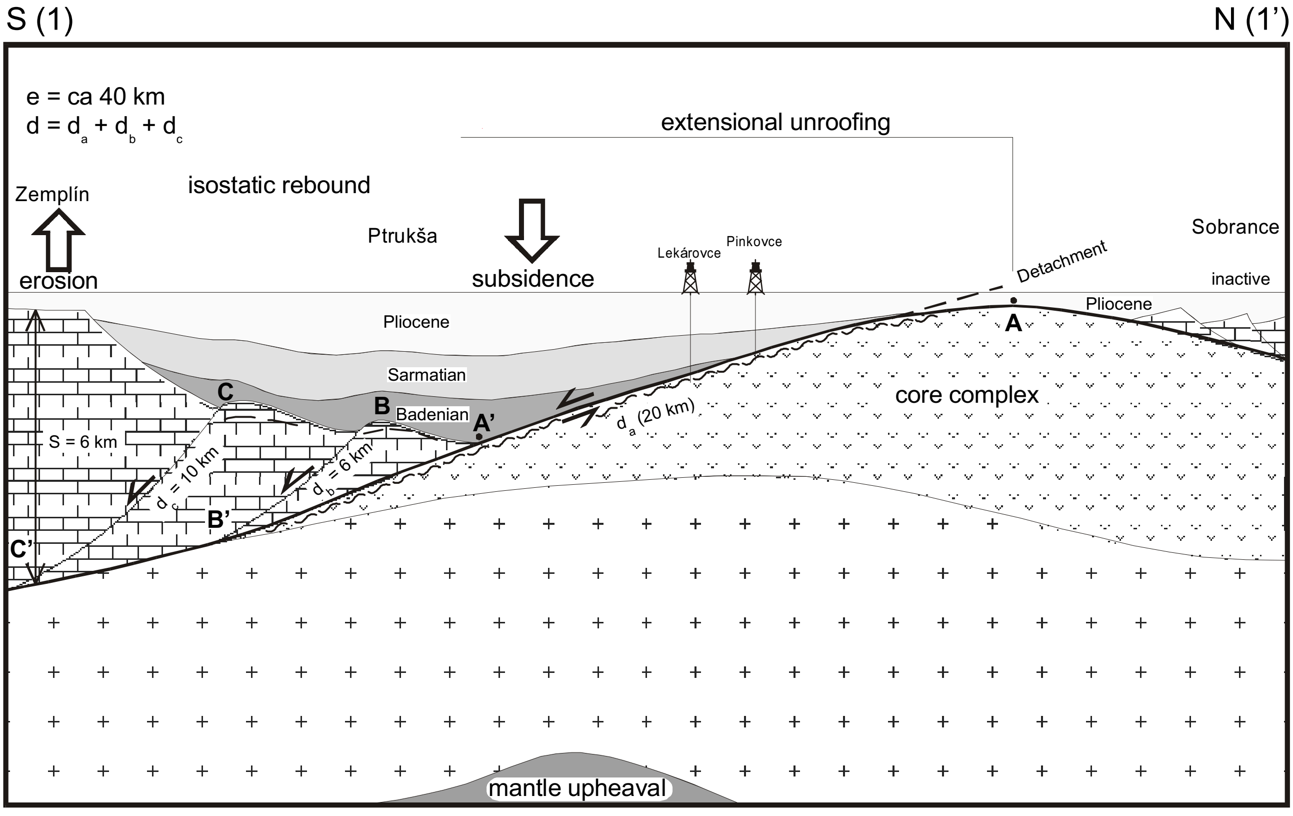

Fig. 3. Interpretation of basement structure of the East Slovak Basin derived from seismic profile PF 571A/82 and borehole data. The profile shows the core structure of the Iňačovce-Krichevo Unit (Penninics) exhumed along the detachment fault (marked by cataclasites), which also controlled the extensional unroofing, basin subsidence and roll-over growth of hangingwall elevations (Ptrukša Zone, Zemplín Island).

the cataclasites that developed along detachment faults. The youngest extensional detachment with cataclasites overprinted the contact between the basement core complex and the Neogene sediments (cataclastic breccias were misinterpreted sometimes as a basal clastic sediments). Therefore, the Neogene sediments appear to be detached during the core complex exhumation. This assumption is also supported by the FT results, obtained from a different type of zircon grains from Neogene syn-rift sediments that lack young FT annealing typical for the core complex associations. In the seismic profiles (e.g. PF 571A/82), this detachment is expressed as a basin/basement reflector which responds to the low-angle normal fault with roll-over growth of elevations in the Ptrukša Zone and Zemplín Unit. In this case, the East Slovak Basin was formed above the extensional detachment (master fault) as a consequence of core complex updoming accomplished by normal faulting and subsidence of the hangingwall.

Stratigraphic evidence and geochronological data for the Iňačovce-Krichevo Unit allow us to interpret the time-temperature path. The cored rocks of the Iňačovce-Krichevo Unit were brought from the metamorphic depth of about 15 km to shallow crustal level, or near to the surface (their material was recycled to form Merník conglomerates — Soták et al. 1990). Thus, the Iňačovce-Krichevo Unit complexes appear to have been cooled and exhumed rapidly. The vertical displacement of the core complex started in the Upper Oligocene, with a high uplift rate and approached the zircon FT blocking temperature during the Early Miocene (~20 Ma). If the time spanning 30–20 Ma is assumed as sufficient for the exhumation, the core complex reached the uplift rate of about 1.5 km/Ma. Although, at this rate the uplift was fast, it complied with other core complexes exhumed in the continent-continent collisional orogens (e.g. Boundy et al. 1996).

The cooling age of the Iňačovce-Krichevo Unit is most consistent with the zircon FT ages of the Rechnitz window. From this Penninic window, Dunkl & Demény (1991) reported the zircon FT data ranging from 15.1 to 18.5 Ma. Although this window is situated at the western margin of the Pannonian Basin System, the formation of both core was related to the same Early Miocene extensional period (Royden et al. 1983; Tari et al. 1992). Minor differences between the zircon FT age means for the windows could indicate some temporal shift of the main extension.

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QUARTZ MINERALIZATION IN THE MAGURA NAPPE (POLAND): A COMBINED MICROSTRUCTURAL AND MICROTHERMOMETRY APPROACH

A. ŚWIERCZEWSKA[1], V. HURAI[2], A.K. TOKARSKI[1] and R. KOPCIOWSKI[3]

[1]Polish Academy of Sciences, Institute of Geological Sciences, Research Centre in Kraków, Senacka 1, 31 002 Kraków, Poland; ndswierc@cyf-kr.edu.pl

[2]Geological Survey of Slovakia, Mlynská dolina 1, 817 04 Bratislava, Slovak Republic

[3]Polish Geological Survey, Skrzatów 1, 31 650 Kraków, Poland

Key words: Outer Carpathians, Magura nappe, quartz mineralization, fluid inclusions.

Introduction

This paper discusses origin of quartz mineralization hosted in Cretaceous and Tertiary sandstones of the Magura nappe: the innermost nappe in the Polish segment of the Outer Carpathians (Fig. 1) composed of (from the south to the north) the Krynica, Bystrica, Rača and Siary slices. Quartz mineralization has been studied along a cross-section situated in the central part of the Polish segment of the Magura nappe (Figs. 1, 2). This has been supplemented by study of this mineralization within the Oligocene strata of underlying Dukla nappe. The mineralization has been observed in sandstones matrix, joints and small-scale faults.

Origin of the quartz mineralization has been studied in the contex of structural development of the Magura nappe during the Tertiary times, which occurred in the three successive stages: (1) northwest-verging synsedimentary folding and thrusting, (2) northeast-verging thrusting accompanied by widespread strike-slip faulting and, (3) regional collapse (Tokarski & Świerczewska 1999; Tokarski et al. this volume; cf. Decker et al. 1997).

Observations

Two types of quartz mineralizations have been distinguished: (1) quartz overgrowths around detrital quartz grains, (2) quartz-calcite association filling joints and small-scale faults.

Quartz overgrowths

Quartz overgrowths have been observed in Eocene sandstones from the exposure (1) located in the outermost part of the Magura nappe (Fig. 2). Quartz grains of the sandstone framework are in

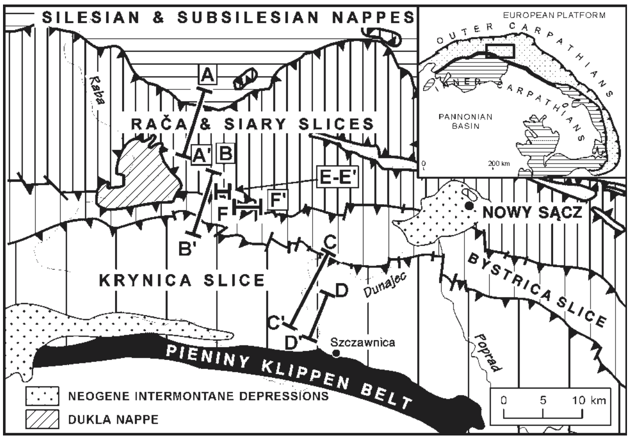

Fig. 1. Map with location of cross-sections (Fig. 2). Geology after Żytko et al. (1988). Inset shows location of the study area (boxed) within the Carpathian arc.

places armoured by newly formed, euhedral quartz from the side exposed in the cross-cutting crack (Fig. 3). The euhedral quartz overgrowths are cut by blocky calcite of a post-folding age (cf. Świerczewska & Tokarski 1998). Primary fluid inclusions are arranged along and within dust-rims, delineating boundaries between overgrowth and detrital quartz core. The dust-rims are composed of dark air bubbles, rock powder and mono- and two-phase aqueous inclusions (Fig. 3). Secondary aqueous inclusions are ubiquitous in the cracks healing the detrital quartz cores. These secondary, fluid inclusion-filled cracks always terminate inside detrital quartz core or at the dust-rim, but they never pass inside the quartz overgrowths. Monophase aqueous inclusions along dust-rims and in the secondary trails within quartz cores indicate preservation of high density, low temperature fluids (< 60 °C). Coexisting two-phase aqueous inclusions homogenized at 101–136 °C. Salinity of the aqueous phase is moderate, ranging between 1.2–5.1 wt. % NaCl eq.

Quartz-calcite association from joints and small-scale faults

Quartz mineralization, filling joints and small-scale faults, is abundant in the entire studied section within the Krynica and

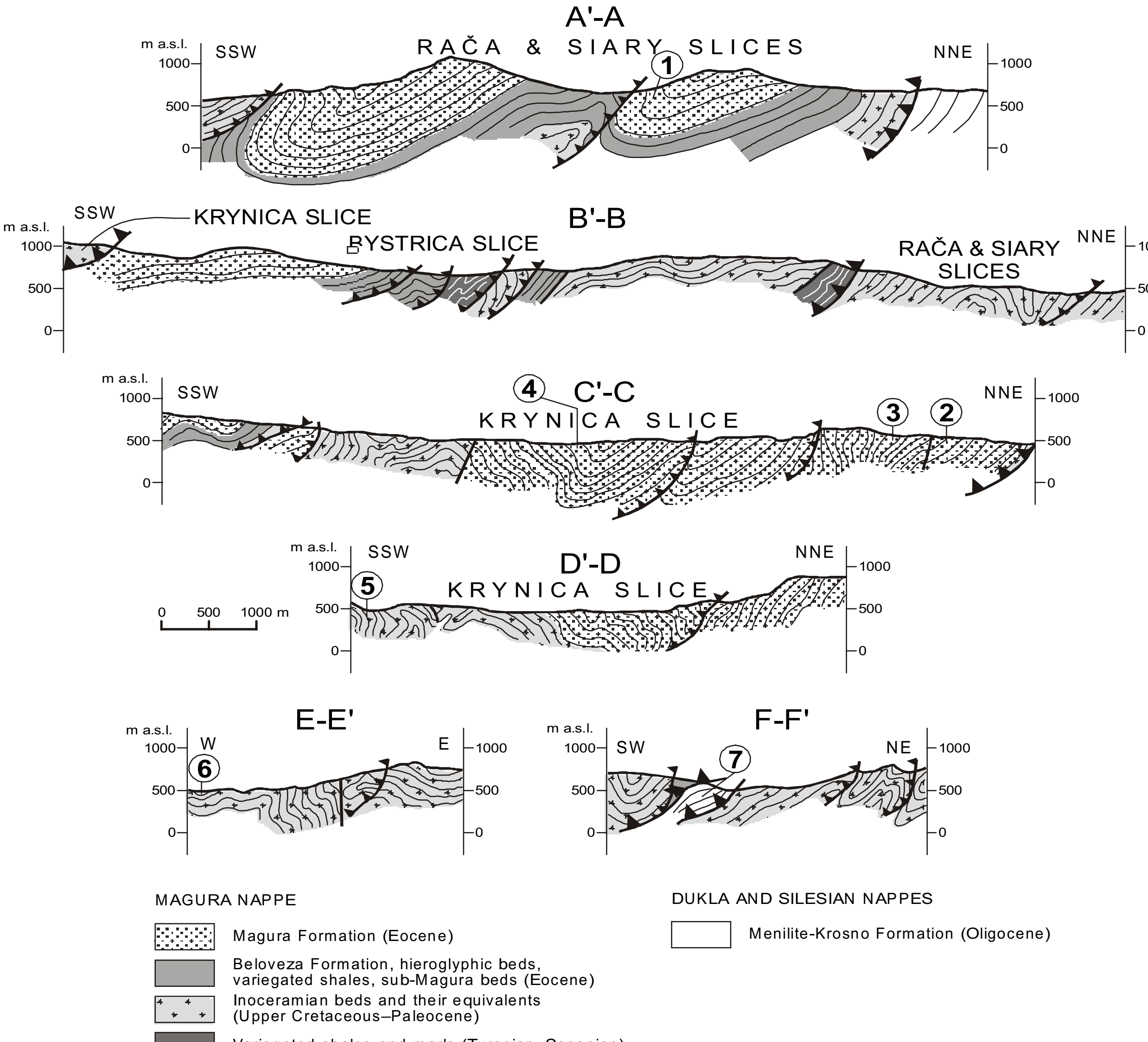

Fig. 2. Geological cross-sections showing structural position of discussed exposures (numbers in circles).

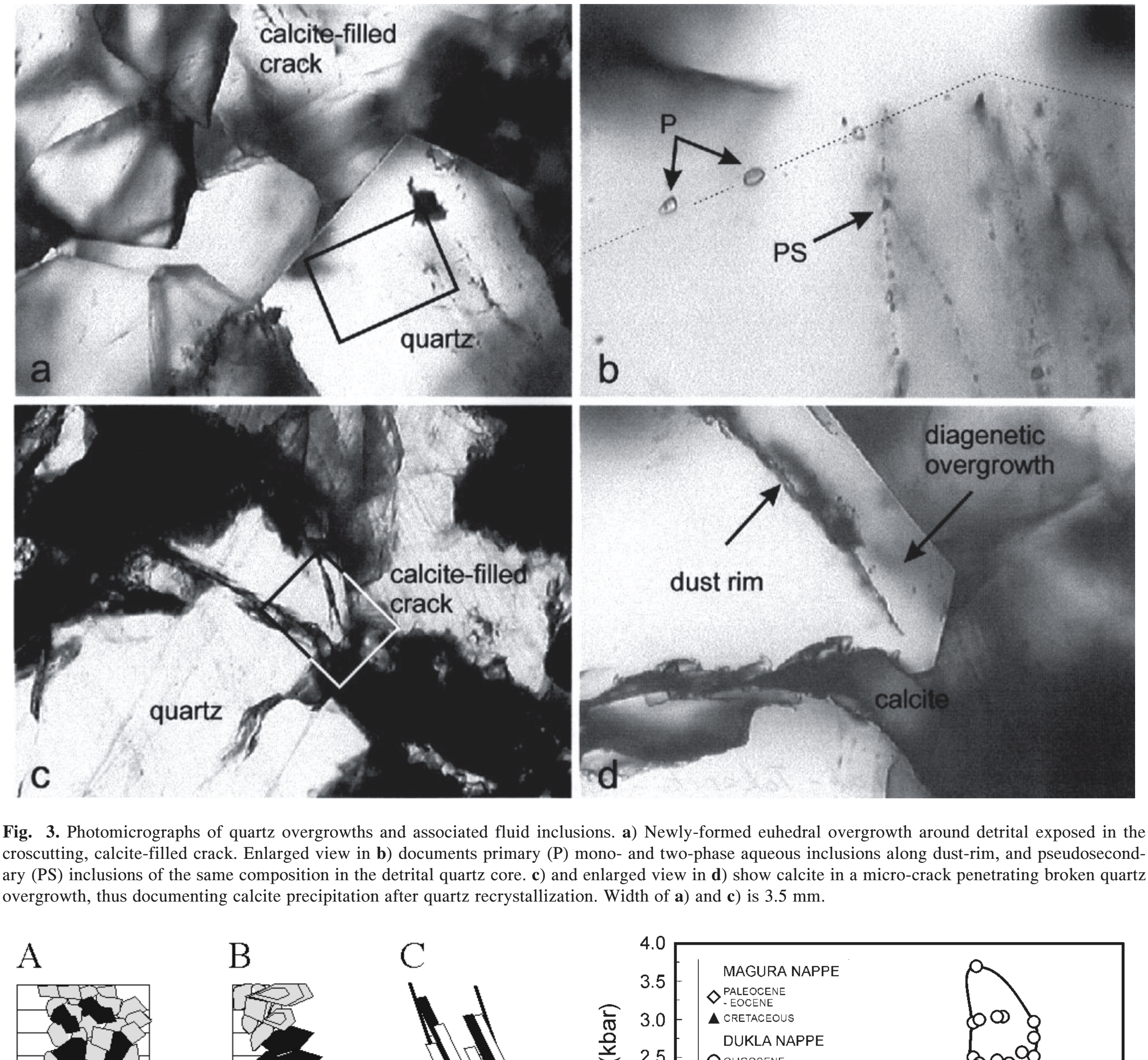

Fig. 3. Photomicrographs of quartz overgrowths and associated fluid inclusions. **a**) Newly-formed euhedral overgrowth around detrital exposed in the croscutting, calcite-filled crack. Enlarged view in **b**) documents primary (P) mono- and two-phase aqueous inclusions along dust-rim, and pseudosecondary (PS) inclusions of the same composition in the detrital quartz core. **c**) and enlarged view in **d**) show calcite in a micro-crack penetrating broken quartz overgrowth, thus documenting calcite precipitation after quartz recrystallization. Width of **a**) and **c**) is 3.5 mm.

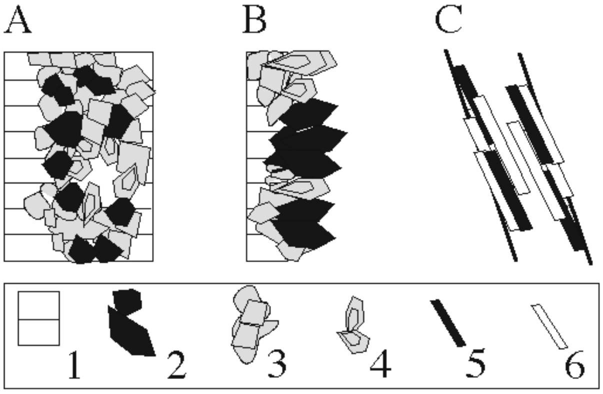

Fig. 4. Schematic outline of the spatial and temporal relationships between calcite and quartz infillings of joints and faults. A, B — composite veins filling joints; C — fault seal; 1 — early columnar calcite; 2, 3 — blocky quartz and calcite association (2 — quartz, 3 — calcite); 4 — late drusy calcite; 5 — fibrous quartz; 6 — fibrous calcite.

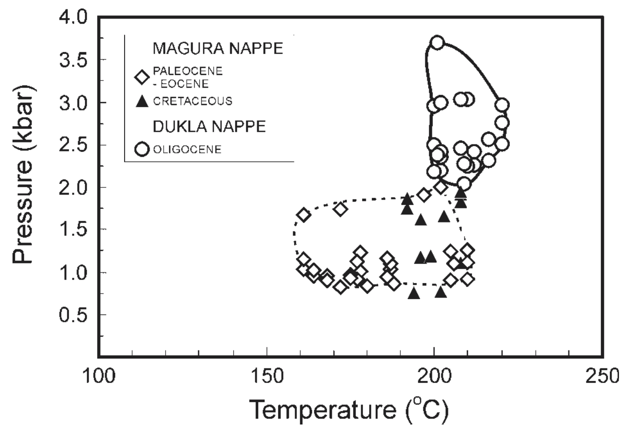

Fig. 5. Formation PT conditions of blocky and drusy quartz from joints.

Bystrica slices of the Magura nappe and in the underlying Dukla nappe. The mineralization has been observed in exposures of the Eocene (2–4) Paleocene (5) and Upper Cretaceous (6) sandstones of the Magura nappe as well as in exposure of Oligocene sandstones (7) of the Dukla nappe (Fig. 2).

Joints are filled by blocky and drusy quartz and calcite forming veins up to 2 cm thick, whereas fault surfaces are covered by fibrous quartz and calcite seals up to 5 mm thick (Fig. 4). The mineralization is more abundant in the proximity of exposure-scale faults. In some joints, the quartz-calcite association post-dates columnar calcite and is followed by still later blocky and/or drusy calcite (Fig. 4A, B).

Primary fluid inclusions, up to several hundreds micrometers in diameter, are abundant in the blocky and drusy quartz. Variable phase ratios in the inclusions point to a heterogeneous trapping of immiscible methane and aqueous liquids. Small amounts (up to 5 mol. %) of carbon dioxide and higher hydrocarbons are also present in the methane rich-phase. Coeval calcite occasionally contains CH_4-rich inclusions along with dominant aqueous inclusions.

Salinities of the aqueous inclusions are very consistent (1.2–1.4 wt. % NaCl eq.) in Oligocene sandstones of the Dukla nappe and in Cretaceous sandstones of the Magura nappe, i.e. at the bottom of the section. In contrast, wider salinity fluctuations (0–3.1 wt. % NaCl) have been reported in the upper part of the section i.e. in Paleocene-Eocene sandstones of the Magura nappe.

Formation PT conditions of the blocky and drusy quartz have been derived directly from microthermometry data on coexisting methane and aqueous fluid inclusions. Minimum homogenization temperature of aqueous inclusions was considered to be representative of actual trapping temperature, while pressure corresponding to the given temperature was derived from isochore of coexisting methane inclusion. Estimated formation PT conditions for the drusy and blocky quartz are illustrated in Fig. 5. Formation temperatures during precipitation of the blocky and drusy quartz generally tend to fall from 200–220 ^{o}C in the Dukla nappe, to 160–210 ^{o}C in the Magura nappe. Fluid pressures decrease from 2.1–3.7 kbar in the Dukla nappe to 0.75–2 kbar in the Magura nappe.

Discussion

Low crystallisation temperatures of quartz overgrowths and the secondary inclusions in detrital quartz cores can be unequivocally correlated with diagenesis and burial. Cross-cutting relationships indicate quartz precipitation during the synsedimentary folding (cf. Świerczewska & Tokarski 1998).

The high temperatures (160–220 ^{o}C) quartz from joints and small-scale faults can be linked neither with burial nor with a metamorphism. In the studied section, host rocks have been affected by advanced diagenesis but not by anchimetamorphism (Tokarski & Świerczewska 1998). This suggests an external source for the hot, quartz-forming fluids, percolating the Magura nappe. Vertical PT variations attest to a fluid ascent from beneath of the Magura nappe. Fluid flow was accompanied by migration of methane, possibly liberated due to thermal cracking of kerogene or crude oil (cf. Hurai et al. 1995b). Strong salinity fluctuations and its decrease to essentially zero wt. % NaCl eq. in the upper part of the studied section may be interpreted in terms of an influx of freshwater, probably of meteoric origin. This is confirmed by low $^{87}Sr/^{86}Sr$ isotopic ratio for coeval calcite (cf. Świerczewska et al. 1998).

The drusy quartz can be correlated with "marmarosh diamonds", which are particularly abundant in the Ukrainian segment of the Outer Carpathians (e.g. Tokarski 1905; Vityk et al. 1995, 1996), as well as in the Central Carpathian Paleogene (Hurai et al. 1995a).

The high temperature quartz mineralization is more abundant in proximity to normal faults, implying migration of fluids along these faults. Similarly, epigenetic quartz mineralization of the Central Carpathian Paleogene basin increases with the diminishing distance from the Klippen Belt: a tectonic boundary dividing Outher and Central Carpathians (Hurai et al. 1995b). Cross-cutting relationships show that the high tempoerature quartz mineralization in the Magura nappe was formed during regional collapse (cf. Świerczewska et al. 1998).

Assuming lithostatic load and average density of 2.5 g/cm^3, depths of the quartz formation derived from fluid inclusion microthermometry data should equal to 8–15 km in the Dukla nappe and 3–8 km in the Magura nappe. Such immense depths of burial are not, however, supported by recent tectonic models. It seems to be thus more likely that the high temperature quartz has precipitated under regime of a fluid overpressure, which could locally result in a hydraulic fracturing of the host rocks.

Conclusions

1. Newly formed overgrowths around detrital quartz grains can be correlated with diagenesis and burial.
2. High temperature quartz mineralization of the Magura nappe and underlying Dukla nappe was formed during a thermal event coincidental with regional collapse of the region.
3. This thermal event was accompanied by circulation of hot, CH_4-bearing aqueous fluids, ascending along normal faults.
4. In the upper part of the studied tectonic pile, the quartz-forming fluids have been diluted by freshwater, probably of meteoric origin. At the base of the Magura nappe, significant portion of the aqueous phase might represent marine and/or sedimentary (connate) waters.
5. Fluid pressures during the regional collapse have been higher than those corresponding to a lithostatic load.

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THE INFLUENCE OF EARLY JOINTS ON STRUCTURAL DEVELOPMENT OF THRUST-AND-FOLD BELTS: A CASE STUDY FROM THE OUTER CARPATHIANS (POLAND)

A.K. TOKARSKI[1], W. ZUCHIEWICZ[2]
and A. ŚWIERCZEWSKA[1]

[1]Polish Academy of Sciences, Institute of geological Sciences, Research Centre in Kraków, Senacka 1, 31-002 Kraków, Poland; ndtokars@cyf-kr.edu.pl
[2]Institute of Geological Sciences, Jagiellonian University, Oleandry 2a, 30-063 Kraków, Poland

Key words: joints, thrust-and-fold belts, Outer Carpathians.

Introduction

Joints are the most frequent tectonic structures. Moreover, in some areas, joints are early, synsedimentary features. It follows that in these areas joints record the stress arrangement which occurred during sedimentation of the host strata. In distinction to other early tectonic features (deformation bands, vein structures, hydroplastic faults), joints can easily be transformed into different structures (faults, tension gashes) due to changing stress arrangement.

We will demonstrate, that joints are capable of recording the complete structural evolution of fold-and-thrust belts, using the Polish segment of the Outer Carpathians as an example. Moreover, we will show, that the early synsedimentary joints had overwhelming control on the fault formation during subsequent structural evolution.

Regional setting

This study has been focused on joints in Tertiary sandstones within the Polish segment of the Outer Carpathians (Fig. 1). This

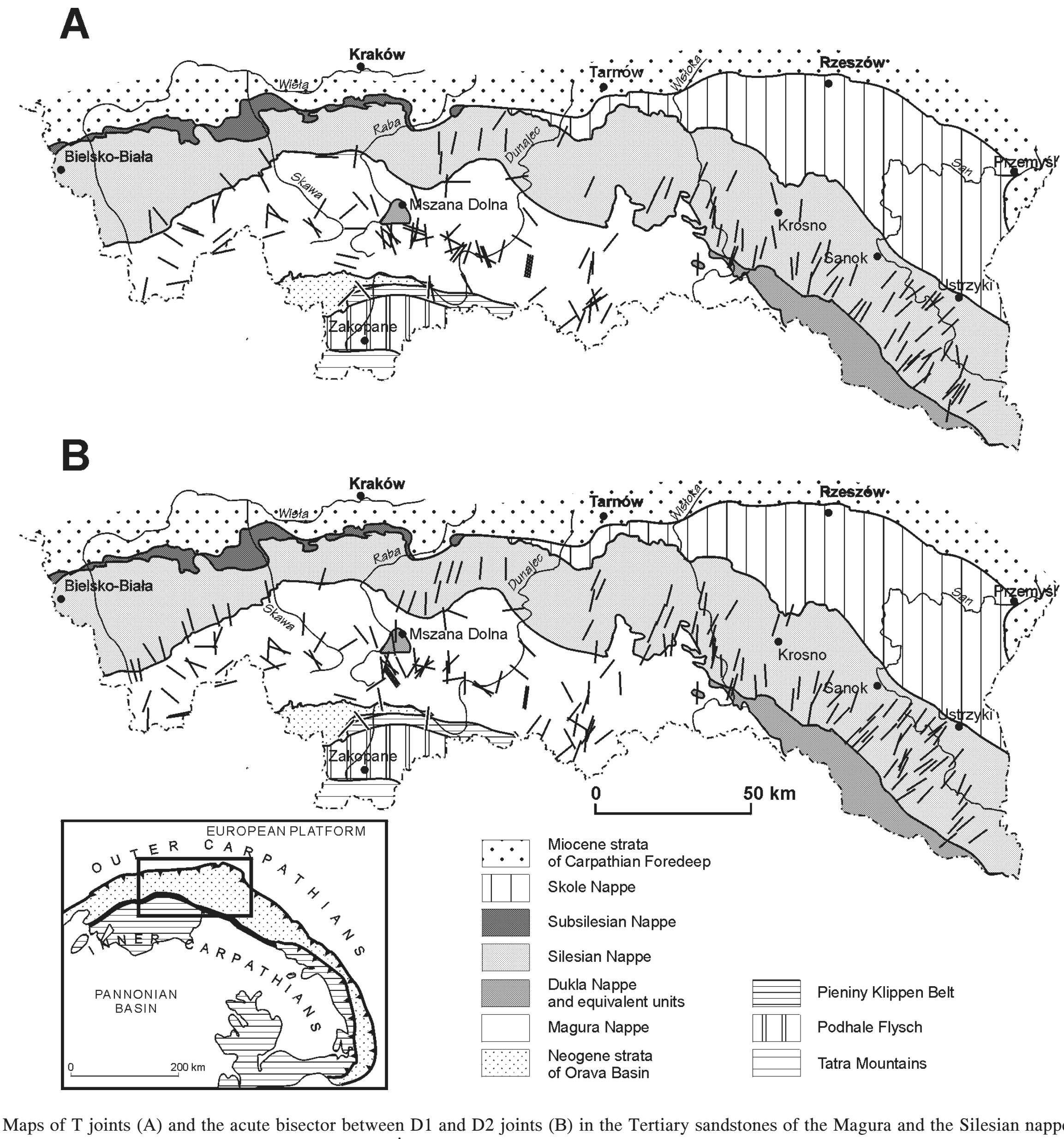

Fig. 1. Maps of T joints (A) and the acute bisector between D1 and D2 joints (B) in the Tertiary sandstones of the Magura and the Silesian nappes. Location of the study area is shown in the inset. Geology after Żytko et al. (1988).

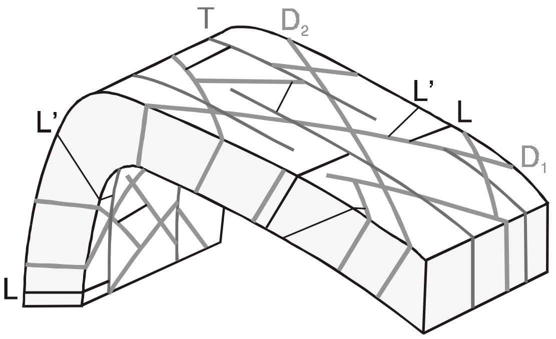

Fig. 2. Distribution of joints in folded strata.

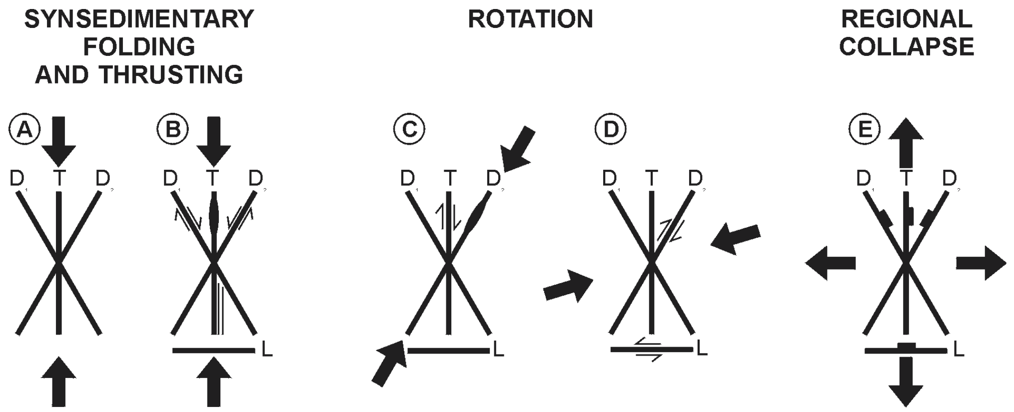

Fig. 3. Diagram illustrating the activity of small-scale tectonic features during Tertiary structural development of the Polish segment of the Outer Carpathians. For explanations see text.

is a north-verging thrust-and-fold belt composed largely of Lower Cretaceous to Lower Miocene flysch. During the Tertiary period the belt underwent three successive stages of deformation (Tokarski & Świerczewska 1999). At first, the belt was formed as an accretionary prism related to southward directed subduction (Tomek & Hall 1993) which resulted in north-verging synsedimentary folding and thrusting (Tokarski et al. 1995; Świerczewska & Tokarski 1998; Tokarski & Świerczewska 1998). This was followed either by major clockwise rotation of the regional stress field (Aleksandrowski 1985; Decker et al. 1997) or major anticlockwise rotation of the belt (Márton & Tokarski unpublished data) resulting in fold-parallel strike-slip faulting within the whole belt (cf. Decker et al. 1997) and refoldind confined to the western part of the Magura nappe (Aleksandrowski 1985). Regional collapse marked by normal faulting followed (Decker et al. 1997).

Exposure-scale patterns of joints

In this paper we define a joint as a fracture on which we have not observed any offset (cf. Dunne & Hancock 1994). Joints have been studied in 240 exposures within the Magura and the Silesian nappes. In particular exposures we have observed 2–5 sets of joints (Fig. 2) comprising cross-fold joints (T and D) and fold-parallel joints (L and L´). Cross-cutting relationships indicate that the fold-parallel joints usually post-date the cross-fold joints. However, the exact age of the fold-parallel joints is not clear. We believe that they were formed during the synsedimentary folding.

The cross-fold joints are normal to the bedding, which means that they were formed before folding (Mastella et al. 1997; Tokarski & Świerczewska 1998). Moreover, in the study area, the folding started during deposition of the host strata (Świerczewska & Tokarski 1998), which means that the cross-fold joints are synsedimentary features. In the exposures in which sandstones are cut by clastic dykes, the dykes are usually oriented parallel to the cross-fold joints (Tokarski & Świerczewska 1998). This confirms the early age of the joints. In the study area, the architecture of cross-fold joints is controlled by the thickness of the host sandstones. Within thin-bedded sandstones only D joints occur or predominate, whereas in thick-bedded sandstones only T joints occur or predominate.

Map-scale patterns of joints

Within the Tertiary sandstones of the Silesian and the Magura nappes, both T-joints and the acute bisector between D joints are oriented perpendicular to the map-scale fold trends (Fig. 1). However, within the Magura nappe there are numerous local anomalies. We interpret these anomalies as resulting from refolding (in the western part of the Magura nappe) or from drag on cross-fold strike-slip faults or from rotations of blocks bounded by flat-dipping shear zones.

Joints during structural evolution

(1) Subduction-related, synsedimentary folding and thrusting

Sedimentation of the Tertiary strata occurred in a stress regime in which the maximum stress axis was horizontal and normal to the present-day map-scale fold axes (Rubinkiewicz 1998; Świerczewska & Tokarski 1998). In this stress arrangement, T and D joints were formed (Fig. 3A). Numerous D joints were transformed into strike-slip faults of two conjugated sets (Fig. 3B). Right-lateral faults were formed on the D1 joints and left-lateral faults on the D2 joints. Except for the fold-parallel south- and north verging thrusts, no other faults were formed during this stage of the structural evolution. No strike-slip faulting occurred in thick-bedded sandstones devoid of the D joints. Instead, in these sandstones T joints in places formed high density zones. Some of the T joints were transformed into tension gashes. We believe that fold-parallel joints were also formed during the folding.

(2) Rotation

During rotation (Fig. 3C), at first some T joints were reactivated as right-lateral strike-slip faults, while some D2 joints were transformed into tension gashes. After that (Fig. 3D), some D2 joints and some of the D2 joints related left-lateral strike-slip faults were reactivated as right-lateral strike-slip faults. Close to completion of the rotation, the fold-parallel joints (L, L´) were reactivated as left-lateral strike-slip faults (cf. Decker et al. 1997). Very few faults unrelated to preexisting joints and faults were formed during the rotation.

(3) Regional collapse

Completion of the rotation, was followed by a collapse marked by normal faulting. At that time, numerous preexisting joints and faults of all sets were reactivated as normal faults (Fig. 3E). Very few faults unrelated to preexisting joints and faults were during the regional collapse.

Summary

Joints were formed during the onset of the structural evolution, their architecture controlled by the thickness of the host sandstone beds. Later the joints were reactivated several times as faults and tension gashes. On the other hand, very few faults and tension gashes unrelated to preexisting joints and faults were formed during the whole span of the structural evolution. It follows that this evolution was largelly controlled by early synsedimentary joints.

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OROGENS, FORELANDS AND INTRA-FOLDBELT NUCLEI — NEW APPROACHES AND SPACE-SAVING INTERPRETATION

G. ZOLNAI

11. Av. des Sayettes, 64000 Pau, France

Key words: Hercynian and Alpine orogenies, Pannonian Basin, compressive folding, basement tectonics, wrenching, inversion tectonics, gravitational nappes.

The South European realm, including the Alpine-Carpathian foldbelt, has known a long geological evolution. The late-Paleozoic Hercynian (Variscan) orogeny resulted in a continuous basement-complex the effects of which were evident throughout the Alpine cycle, which was composed by two major periods: an extensional (transtensional) phase which lasted from the Permian to mid-Cretaceous, and a compressional (transpressional) phase from the Late Cretaceous to present day. First the Paleozoic basement was fractured under sinistral trans-tensional conditions. This resulted in a complex chessboard (Figs. 1, 2), with high-lying cratonic masses and deep pull-apart basins (subsequently merged into furrows under continued wrenching) in between. The long straight structural lines (such as Balaton line, the Darno, Muran etc. faults) seem to be a parts of the Paleozoic heritage, while the arcuate features (e.g. flysh furrows) may be linked to the ongoing passive-

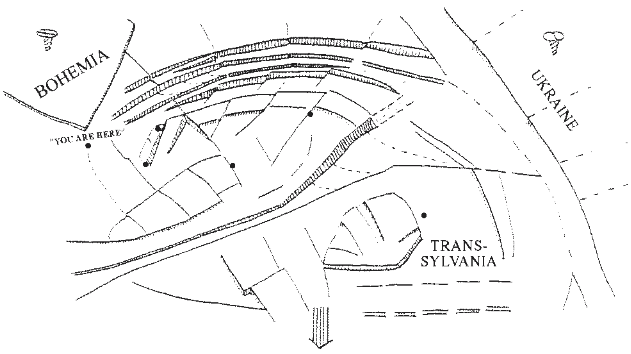

Fig. 1. Bird's eye sketch of the main basement units in the Carpatho-Pannonian area (after full extension).

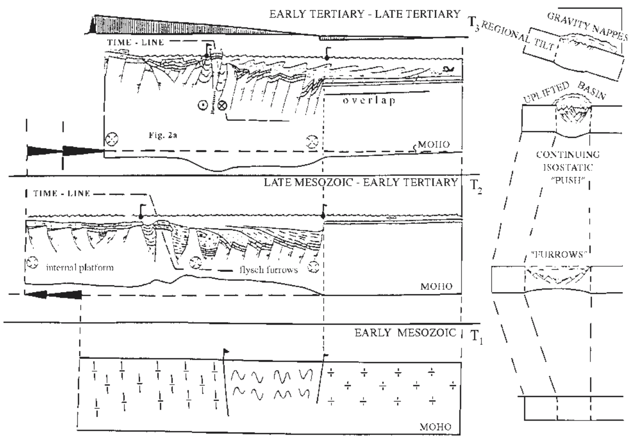

Fig. 2. Evolution model; Northern Carpathians.

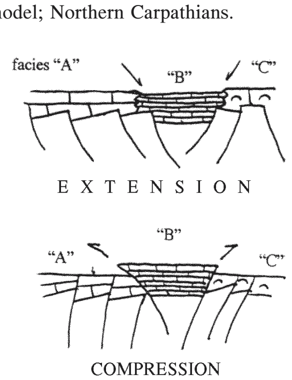

Fig. 2/a. Structural inversion in platform areas.

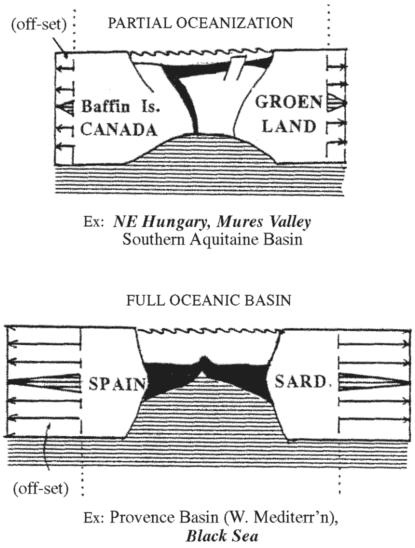

Fig. 3. "Partial" versus "full" oceanization.

margin extension. The new paleogeographical pattern has been compressed again under sinistral transpression. Folding of the area was intense, with local overthrusting, structural inversions and imbricates in the internal platforms. The deepest external furrows, which had their basement the most intensely thinned (streched), absorbed the greatest part of the crustal shortening as well. Fish-scale (imbricate) stacking of the basement "slabs" caused the thickening of the basement-complex, followed by uplifting of the furrows (structural inversion). This resulted in the spilling of the thick, yet unconsolidated sedimentary fill (flyshes) onto the foreland platform (gravity nappes: "Camembert Tectonics"). The orogenic shortening renewed the wrenching, which helped to emplace the major uplifts within the mobile belts (Mount Blanc, Tatras, Southern Carpathians, also the Klippenbelt and similar uplifts in SE Apuseni Mts). Uplifts-and-basins were formed in the intramountain domains ("nuclei") and the foreland cratonic margins (s. as: Malé Karpaty, Bakony, Apuseni Mountains, Vosges, etc). Collision-subduction proper occurred only along the southern margin of the Alpine system (Sputhurn Spain–Crete–Cyprus–Persia); the intensity of convergence decreased progressively to nil to the North (Bohemia, Ukrainian).

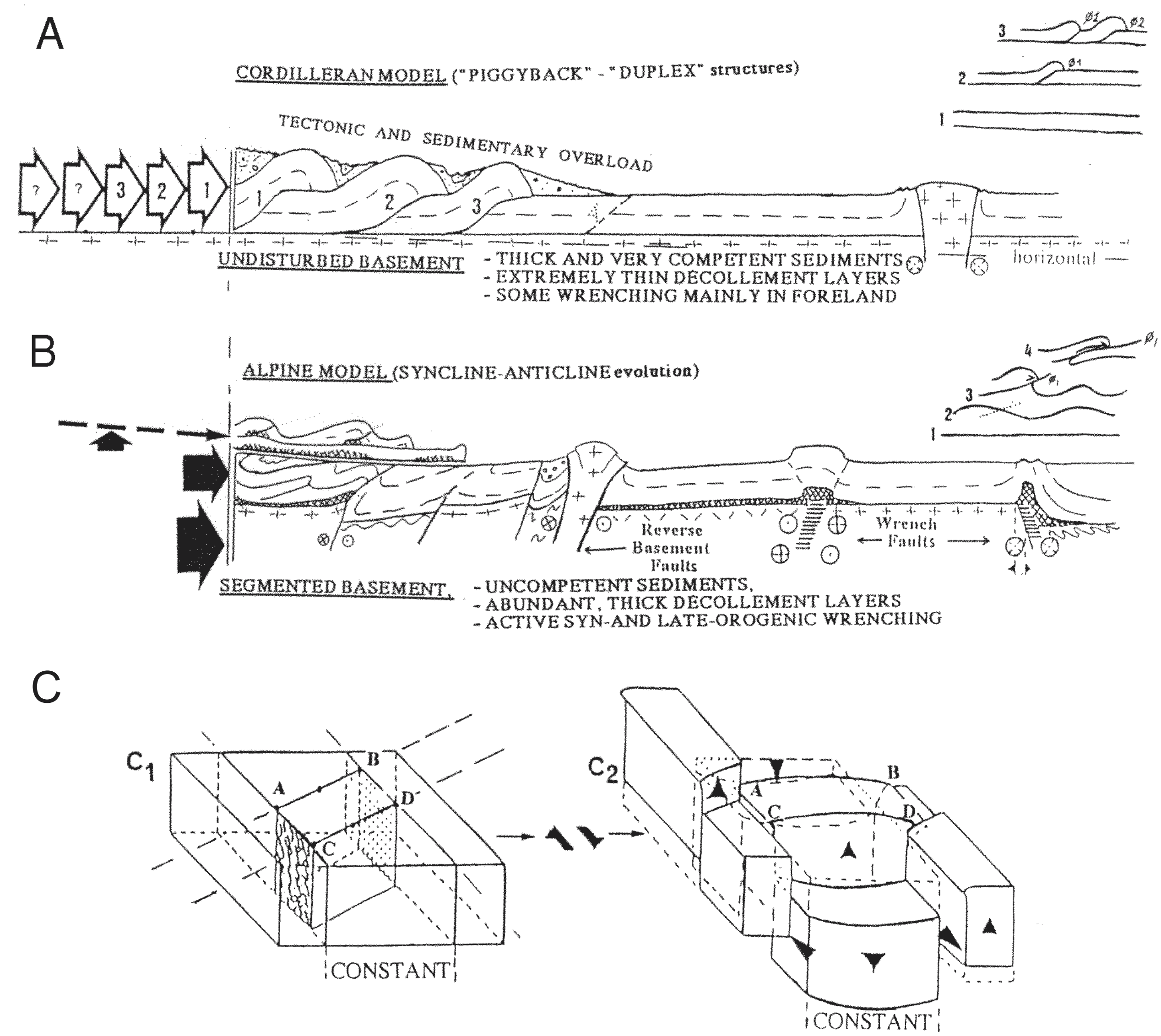

Fig. 4. Text see on next page.

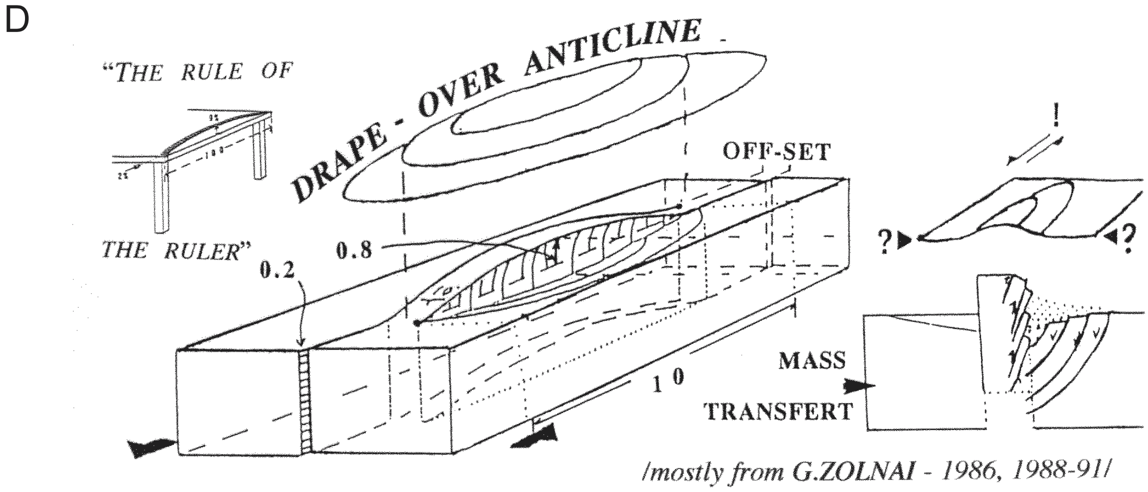

Fig. 4. "Cordilleran" versus "Alpine" deformation. **A**. Cordilleran model. **B**. Alpine model. **C**. "Paired uplifts and basins" due to slight wrenching within a constant-width wrench corridor. Note the overhangs (reverse faults) often present along the edges of the uplifted blocks. In the case of simple wrenching or translation, the mass balance (between "up" and "down" blocks) may remain unchanged. Examples: Central Pannonian wrench corridor; also California, Anadarko basin and Ancestral Rockies Foldbelt, SW USA. **D**. Vaulted basement uplifts. (0.8 km vertical closure) due to very small (0.2 km off-set) basement-wrenching, in presence of a fault-relay (10 km long "basement slab": "the rule of the ruler"). Superficial drape-folds may appear in the meatime, without any compression "perpendicular-to-the-fold-axis". Note adjoining, genetically related downwraps ("sags"). Examples: Malé Karpaty Mts. also Central Montana uplift, Uinta Mts., (West-Central USA), "Basque transverse zone" (Westernmost Pyrenees, SW. France), Vosges uplift-Rhine Graben-Black Forest "structural compound".

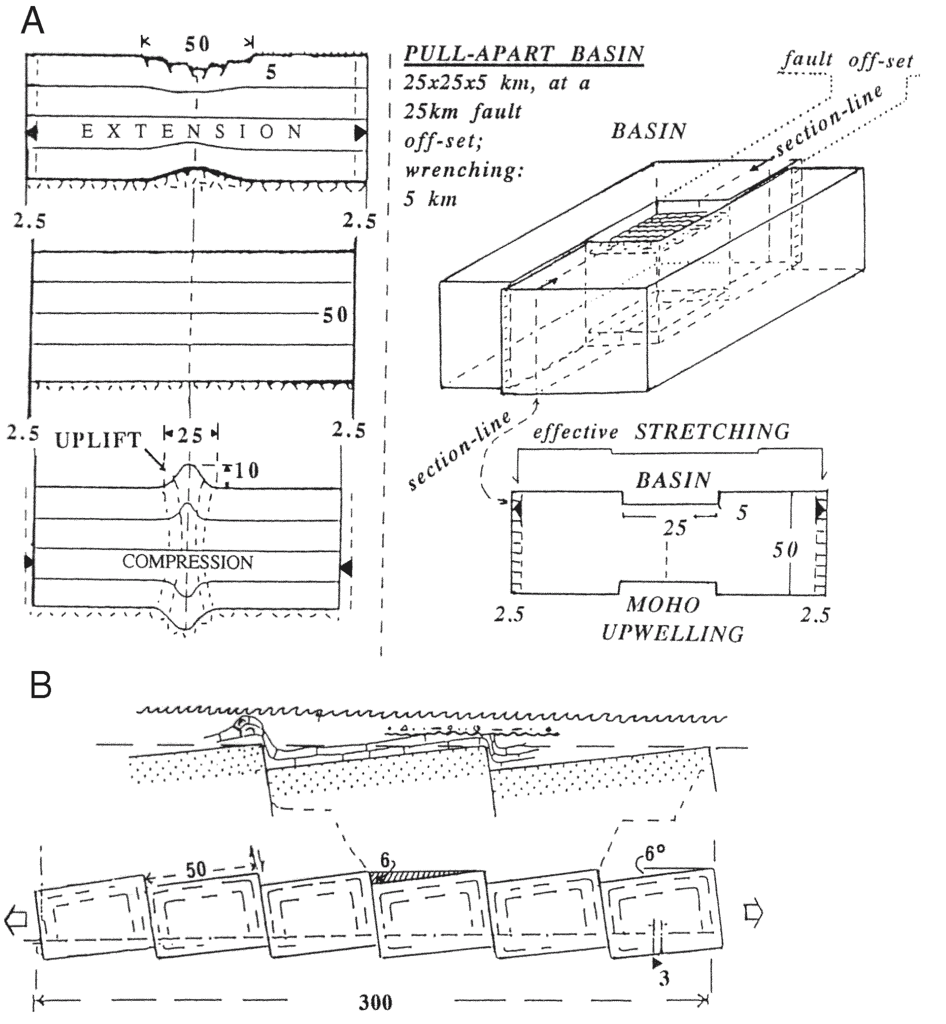

Fig. 5. Mechanical models: (all values in km). **A**. "Mass balance diagram" demonstrating how small-scale extension (stretching: upper part) or compression (lower figure) can generate sizeable basins or uplift, in presence of a thick basement. Examples: Vienna and East Slovak Basins, Tatras; also southern Aquitaine basin, Mt. Blanc, High Pyreneas etc. uplifts. **B**. Block tilting. Followed by erosion and transgression generates phenomena often (mis)taken for signs of compressive folding ("orogenic phase"). The over-all extension here is only 1 % ± (3 km/300 km); each basin may be 6 km deep. Examples: Sümeg area W. Hungary, also: SW Sardinia, Rainbow basin, W Canada, North-Pyrenean foldbelt/Albian.

In this interpretation basement-tectonics substitute in places hypothetical large-scale overthrust napping. Only "partial oceanization" (Fig. 3) is used here, since observed in distinct, limited areas (western Aquitaine, central Alpine, eastern Pannonian?, Transylvanian?, basins: Jurassic). The Black sea and Vardar zone may nevertheless be true "ocean-windows" or "ocean-arms". Adjustment of the loosened (yet para-autochthonous) south–European basement-block during the Tertiary caused both the renewed but limited "true oceanization" (Tyrrhenian and Provence basins), and some major but local overthrust-napping (Central Alps), neither of which should be extended (used as "model") over the whole Alpine realm. Due to the presence of a fragmented, but continuous basement and the preponderance of soft to medium-hard sediments, compressive folding mainly occurred in the "syncline-anticline mode" as opposed to the "piggyback" model (Fig. 4). "Tectonic units" including basement imbricates (rather than "nappe units") were formed and uplifted within and around the intra-orogenic "nucleus" of the Pannonian Basin (Mecsek, Bakony, Apuseni Mountains). Volcanism, both basic (pillow lavas, ophites and ophites) as well as intermediate-acid, was related to the renewal of deep fractures (incl. furrows) and reflects the presence of (thick) continental crust underneath. Very deep-seated yet local seismicity in the Easternmost Carpathians (Vrancea area) can be accounted for by effects of the convergence of still active crustal faults (forcing "chunks" of crust to a depth) and should not be extrapolated over the whole Carpathian arc, as proof of regional subduction-system.

The mechanical model used here (Fig. 5): pull-aparts, arching, uplifts-and-basins, tilting and overthrust imbricates ("fishscales") all imply limited crustal movements (ten-km order), either in extensional or compressive mode. Moderate-displacement strike-slipping (of km-magnitude) is also typical of basement-dominant areas. Restitutions thus become more compatible with the given limited space. They necesitate no huge transfers nor the stacking of multiple, albeit thin and flat-lying, overthrust nappes. On the contrary, fan-like large-scale overthrust systems or semi-circular, centripetal subduction of thick crustal segments would, in excess of their inherent physical unlikeliness, cause insolvable regional volume-problems. Such processes would involve stacking of basement-slabs, necessarily followed by large-scale isostatic uplifting, and subsequently cause erosion of the sediments still present.

Wandering of terranes and rotation ("waltzing") of interior blocks could have occurred during the Paleozoic evolution but such large-scale displacement are unlikely during the Alpine evolution (continuity of the basement fault-pattern). The opening-and-closing of large oceans (Meliata) complicate, without conclusive evidence and to no avail, the paleotectonic evolution. The relative thinness of the calcareous sediments filling the basins up to sea-level, also pleads against the presence of very deep (> 4 km) truly oceanic domains.

Being in an Orogenic System should not mean the use of all orogenic phenomena at hand. The systematic use of the best-known, since best exposed overthrust models (High Central Alps, American Cordilleras) seems risky and misleading in the Carpatho-Balkan domain, for the lack of good outcrops, and/or that of sufficient deep evidence (mines, drilling). Moreover, the simplest possible interpretation always has the best chance to be closest to reality.

The reinterpretation of the area with the above "**alternative hypotheses**" may need much further work, to be done by those who know all the details, of the entire structural province. It could, in the end, help to eliminate major discrepancies, inherent to the use of far-fetched models, irrelevant in this area. It should, thus, result in simpler, yet more realistic "*in situ*" solutions, better applicable also in everyday exploration.

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TERTIARY STRESS FIELD IN THE POLISH OUTER CARPATHIANS: INSIGHTS FROM THE STUDY OF JOINTS

W. ZUCHIEWICZ[1] and A.K. TOKARSKI[2]

[1]Institute of Geological Sciences, Jagiellonian University, Oleandry 2a, 30063 Kraków, Poland; witold@ing.uj.edu.pl

[2]Institute of Geological Sciences, Polish Academy of Sciences, Senacka 1-3, 31002 Kraków, Poland; ndtokars@cyf-kr.edu.pl

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Introduction

We present the results of regional mapping of joints which has been going on since 1995 within the Polish segment of the Outer Carpathians, and has already been completed for the Tertiary strata of the Silesian and Magura nappes. In this paper we follow the terminology summarized by Dunne & Hancock (1994), and use the word "joint" as a field term.

The Polish segment of the Outer Carpathians is a north-verging fold-and-thrust belt, composed mostly of Lower Cretaceous to Lower Miocene flysch strata. The belt comprises several nappes. Two of them, the Silesian and Magura nappes, extend along the whole belt. During Tertiary times, the structural development of these nappes was different: the strata of the Silesian nappe were folded only once, during Late Oligocene to Miocene times (Roca et al. 1995, and the references therein), whereas the strata of the Magura nappe were folded twice (Aleksandrowski 1985). The first episode of folding started during the Eocene and was completed during the Eocene (Tokarski et al. 1995); the second episode occurred during Miocene times (Aleksandrowski 1985). The map-scale fold pattern is also different for each nappe. Within the Silesian nappe, the map-scale fold axes display two orientations. East of Tarnów, the axes trend NW whereas west of Tarnów they trend W-E. In contrast, within the Magura nappe, the F_1 map-scale fold axes form an arc, within which the fold axes trend from NNW in the east to NE in the west. The F_2 map-scale fold axes trend NNW. The latter can be easily discerned in the western portion of the Magura nappe only. East of Kraków, these folds become less discernible, and east of Tarnów they merge completely with the F_1 folds.

Material and methods

Cross-fold joints have been studied in sandstones at 240 stations located within the Paleocene to Lower Miocene strata of the Silesian nappe (116) and within the Paleocene to Oligocene strata of the Magura nappe (124). Within the Magura nappe, the joints have been studied only within the map-scale F_1 folds.

The cross-fold joint system comprises: (1) a single set of T joints striking subperpendicular (80–90^o) to map-scale fold axes, and (2) two sets of D joints (D_1 and D_2) striking at high angles (60–80^o) to these axes. All joints are normal to the bedding. Within thin-bedded sandstones only D joints occur or predominate, in medium-bedded sandstones both T and D joints occur in roughly equal proportions, whereas in thick-bedded sandstones only T joints occur or predominate decisively.

Both in the Silesian and Magura nappes, the T joints are oriented roughly parallel to the acute bisector between the D_1 and D_2 joints. No more than one system of cross-fold joints has been observed at each of the studied stations. Within the Magura nappe, the orientation of T joints and the acute bisector between D_1 and D_2 joints show local deviations from the normal position to the F_1 map-scale fold axes. These deviations occur at stations situated close to map-scale faults. The largest deviations are to be found in the westernmost segment of the Magura nappe and in a structural depression north of the Orawa Basin.

Discussion

Within both studied nappes, the cross-fold joints display a consistent pattern. Both the T joints and the acute bisector between the D_1 and D_2 joints maintain orientation normal to the map-scale fold axes, except for the westernmost part of the Magura nappe. This indicates interrelation beween the joints and map-scale folds. The T joints are extension joints, whereas the D joints are hybrid and shear joints that form two conjugate sets (cf. Mastella et al. 1997). In the Outer Carpathians, the D joints are of pre-folding and/or early folding age (Mastella et al. 1997). Frequent replacement of T joints by D_1 or D_2 joints and vice versa indicates that they were formed contemporaneously. Furthermore, within the Magura nappe, at least the Eocene strata were folded syndepositionally (Tokarski et al. 1995). It appears that the cross-fold joints record a stress regime which occurred during the deposition of the strata involved. It was a strike-slip regime with σ_1 oriented parallel both to the acute bisector between the D_1 and D_2 joints and to the T joints. Similar development of the cross-fold joints in the whole succession discussed indicates that this stress regime remained constant within both studied nappes during Paleogene times.

The spatial arrangement of the reconstructed orientation of cross-fold joint-related σ_1 within the Magura nappe is less consistent than that in the Silesian nappe, possibly due to post-folding rotations of fault-bounded blocks in the westernmost portion of the nappe. Moreover, within the western portion of the Magura nappe, two systems of cross-fold joints occur in places (Aleksandrowski 1985), whereas the single system of cross-fold joints occurs throughout the Silesian nappe. These differences point to variations in structural development between these nappes during Neogene times.

Conclusions

1. The Paleogene stress history of the Silesian and Magura nappes was dominated by strike-slip regime with a constant orientation of σ_1.

2. The post-Paleogene structural development of the Silesian nappe was different from that of the Magura nappe. Within the latter, the Paleogene joint pattern was disturbed by the F_2 folding and post-folding cross-fold strike-slip faults.

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