Geodynamic evolution of the Tatra Mts. and the Pieniny Klippen Belt (Western Carpathians): problems and comments

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ABSTRACT:

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The geodynamic evolution of the Pieniny Klippen Belt (PKB) and the Tatra Mts. assumes that: The Oravic-Vahic Basin developed due to Jurassic rifting processes with thinned continental crust. The oblique rift without rift-related volcanism had probably a WSW-ENE course. Late Cretaceous thrust-folding of the Choč, Krížna and High-Tatric nappes took place underwater and at considerable overburden pressure (~6-7 km). The geometry of the structures was strongly disturbed by pressure solution processes leading to considerable mass loss. Nappe-folding in the PKB was connected with the slow and flat subduction of thinned continental crust of the Vahicum-Oravicum under the northern margin of the Central Carpathians Block. In the terminal phase, the northernmost units of the PKB were transported through gravitational sliding, forming numerous olistolites. In the Tatra Mts. and the PKB, the nappe thrust-folding was influenced by a strike-slip shear zone between the edge of the Central Carpathians and the PKB and caused e.g. the counter-clockwise rotation of the Tatra block and relative changing directions of thrusting. The consequence of Miocene oblique subduction and subsequent collision of the North-European continental crust with the Central Carpathian Block was the activation of NNW-SSE deep fault zones. With one of these – the Dunajec Fault – were connected *en echelon* shears trading on the andesite dykes swarm. Miocene collision caused the disintegration of the Central Carpathian Block into individual massifs and their rotational uplift. The value of rotation around the horizontal axis for the Tatra Massif is estimated at ~40º.

Key words: Rifting, Subduction, Folding, Thrust, Nappe, Tatra Mts., Pieniny Klippen Belt (PKB), Central Western Carpathians (CWC).

INTRODUCTION

The Tatra Mts. are the northernmost part of the Central Western Carpathians (CWC) and belong to the Alpine-Carpathian orogenic belt (Text-figs 1-3). They are part of the Tatric-Fatric-Veporic nappe system (ANDRUSOV 1968; MAHEĽ 1986; PLAŠIENKA & *al.* 1997). The Tatra Mts. belong to the so-called Tatricum, or Tatric superunit (PLAŠIENKA 2003a), situated between

the Pieniny Klippen Belt and separated from the Veporicum to the south by the Čertovica Line (ANDRUSOV 1968; ANDRUSOV & *al.* 1973; KOZUR & MOCK 1996).They are composed of a pre-Alpine crystalline basement and its sedimentary cover complexes.

The Pieniny Klippen Belt (PKB) represents a ca. 600-700 km long trace of a major suture located between the Central (in the south) and Outer Carpathians (in the north) (e.g. ANDRUSOV 1965;

SCHEIBNER 1968; MAHEĽ 1981; BIRKENMAJER 1976a, 1986; NEMČOK & al. 1998). The PKB (Text-figs 1-3) has a transitional position with respect to the Outer and Central Carpathians and forms a narrow zone with an intricate structure (PLAŠIENKA & al. 1997). In Poland, the PKB and the Tatra Mts. are separated by the Podhale Trough which belongs to a much larger structure called the Central Carpathian Palaeogene Basin – CCPB (MARSCHALKO 1968).

The Tatra Mts., as well as the PKB and the Outer Carpathians, were formed due to Cretaceous – Palaeogene orogenic processes, migrating from south to north. The main aim of this paper is the reconstruction of the geodynamic evolution of this area, not only from the point of view of plate tectonics theory, but additionally taking into account the most recent fieldwork, undertaken by me and many other workers. The second aim is the presentation of many important issues that are still controversial, and in need of revision.

GEOLOGICAL SETTING

After PLAŠIENKA (2003a), the so-called Tatric superunit is the lowermost basement/cover sheet of a tabular crustal scale body that probably overrides the South Penninic oceanic suture (TOMEK 1993), designated as the Vahic superunit (PLAŠIENKA 1995a,b, 2003a). Belonging to the Tatra Mts. a crystalline core is composed of two older structural elements forming its Variscan basement: 1) the metamorphic sequences of the Western Tatra Mts., and 2) the granitoid rocks of the High-Tatra Mts. (e.g. PUTIŠ 1992, JANÁK 1994). The crystalline core of the Tatra Mts. is overlain by Meso- and Cenozoic sedimentary rocks corresponding to the Austroalpine sedimentary basin (HÄUSLER & al. 1993; PLAŠIENKA & al. 1997). The Mesozoic sedimentary strata are composed of three groups of structural elements (KOTAŃSKI 1963a): 1) the High-Tatric autochthonous sedimentary cover; 2) two High-Tatric nappes: the Czerwone Wierchy Nappe (composed of two minor
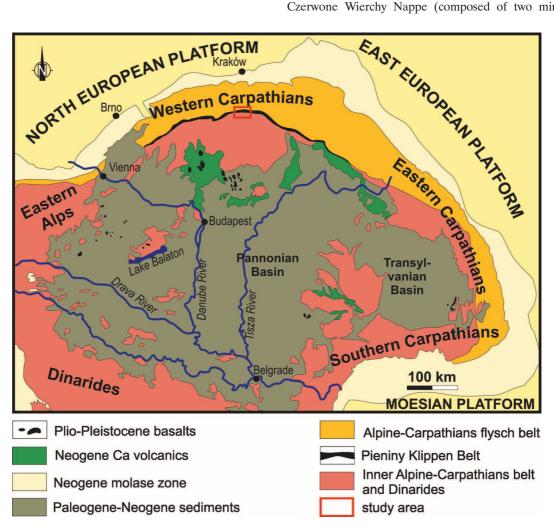

Fig. 1. Main tectonic units of the Alpine-Carpathian-Pannonian orogenic belt; after EMBEY-ISZTIN & al. (1993) and KOVÁČ & al. (1998), simplified
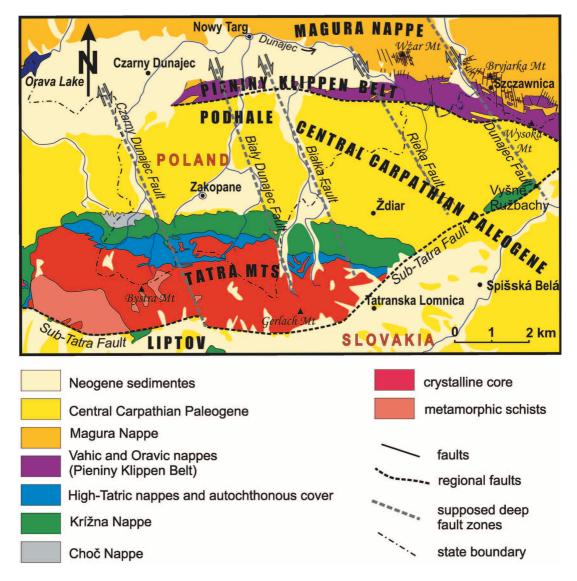
Fig. 2. Schematic geological map of the Tatra Mts. and PKB; compiled after FUSÁN & *al.* (1967), BAC-MOSZASZWILI & *al.* (1979), BIRKENMAJER (1979) and this study. Note that the northern boundary of the PKB is marked along the contact of the Klippen units and the Magura Nappe (Grajcarek Unit is not distinguished); to the east of Szczawnica the large-sized olistolite (Homole-Biała Woda block) is visible (see JUREWICZ 1997)

units: Zdziary and Organy) and the Giewont Nappe (which also comprises crystalline rocks); 3) Sub-Tatric nappes (Krížna and Choč). The Tatra Massif is overlapped by carbonate deposits of the so-called Nummulitic Eocene and a post-orogenic Palaeogene flysch sequence (i.e. BIEDA 1959, 1963). In morphology, the Tatra Massif appeared after the Miocene uplift and connected with it rotation around the W-E horizontal axis (SOKOŁOWSKI 1959; PIOTROWSKI 1978; BAC-MOSZASZWILI & *al.* 1984; JUREWICZ 2000a). The youngest sediments in the area of the Tatra Mts. are related to Pleistocene glaciations and Holocene erosion-accumulation processes.

The PKB, structurally and genetically, is a particular tectonotype linking two nappe-systems: the Palaeoalpine Central Carpathians and the Neoalpine Flysch Belt (MAHEL' 1989). Geometrically, the presumed basement of the PKB corresponds to the Briançonnais domain (TOMEK 1993; STAMPFLI 1993), which separates the northern and southern Penninic zones in the Western Alps. The subdivision of the CWC into tectonic units, their correlation and position has been changed several times (cf. MAHEL' 1981; PLAŠIENKA & *al.* 1997; PLAŠIENKA 1999; KOZUR & MOCK 1996). KOZUR & MOCK (1996) considered the Pieniny Basin to be a northern branch of the (southern) Penninic Ocean, opened
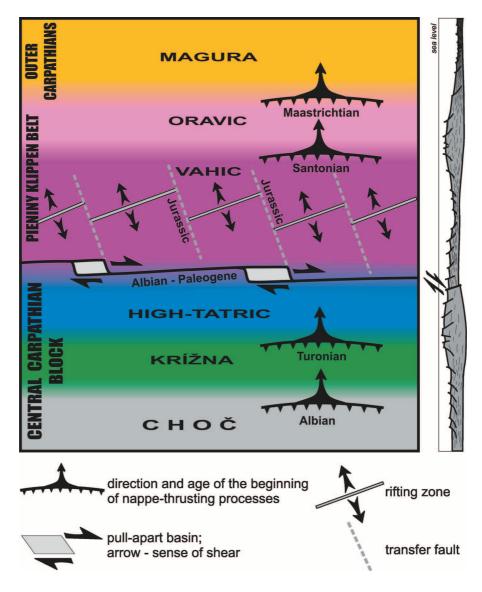
Fig. 3. Sedimentary zones at the boundary of the Central Carpathians, Pieniny Klippen Belt and Outer Carpathians during the Late Jurassic-Early Cretaceous. Right insert – schematic cross-section through the stretching and thinning lithospheric crust (see Text-fig. 5A for details); not to scale

already in Early Triassic times. PLAŠIENKA (1999) divided the Penninic basins into the Vahicum and Magura sub-basins, separated by the Oravicum (Czorsztyn) Ridge. BIRKENMAJER (1977, 1979, 1986) distinguished three main sedimentary zones within the PKB, which was at least 120-150 km wide: the northern (Czorsztyn) Ridge, the central furrow and the southern (Exotic=Andrusov) Ridge. According to the this author, the following struc-tural units may be distinguished within the PKB: A) the Klippen successions (in the investigated part of the PKB: Czorsztyn, Czertezik, Niedzica, Branisko, Pieniny and Haligovce), B) the Central Carpathian successions (Manín and High-Tatric), C) the Myjava successions, D) the Jarmuta cover, E) the Outer Carpathian succession

(Grajcarek Unit), and F) the Palaeogene cover. BIRKENMAJER (1986) assumed that after the basin for-mation stage connected with the defragmentation of the Triassic carbonate platform and expansion related to oceanic spreading in mid-Late Jurassic times (cf. PLAŠIENKA 2003b; GOLONKA & KROBICKI 2004), a pelag-ic state lasted till the Barremian-Aptian. The beginning of the compression stage was marked by the formation of pre-orogenic flysch about the Aptian-Albian boundary. During the late Sub-Hercynian stage, north-verging nappe-thrusting took place. After the sedimentation of Maastrichtian-Palaeocene molasse and flysch (Laramide stage), the retro-arc thrusted Grajcarek Unit was formed (BIRKENMAJER 1970, 1986). During the Palaeogene,

transgression took place in the PKB and flysch deposits appeared. Later, the Saavian phase produced the horst structure of the PKB, with deformations of fore-arc and back-arc vergencies. As a result of the Styrian phase, a transverse fault system had formed. Andesite intrusions are linked with the Saavian and Styrian faults (BIRKENMAJER & PÉSCKAY 1999, 2000a). Appearing in the western segment of the Polish part of the PKB, the strongly folded marine Miocene deposits are connected with the Orawa-Nowy Targ intra-mountain basin (CIESZKOWSKI 1992).

The Palaeogene of the Podhale Trough, located between the PKB and the Tatra Mts., belongs to the Central Carpathian Palaeogene Basin which, after TARI & al. (1993) and KÁZMÉR & al. (2003), is considered to be a fore-arc basin associated with B-type subduction of the European plate. The northern contact of the Podhale Trough with the PKB is tectonic in character (Text-fig. 2), although transgressive sediments of the Nummulitic

Eocene can be observed on the northern slopes of the Tatra Mts. In the south, the Tatra Massif contacts the Palaeogene flysch of the Liptov Trough along the so-called "Sub-Tatra Fault", a polygenetic and multiply acti-vated tectonic fault system, consisting of several seg-ments (UHLIG 1899; MAHEĽ 1986; SPERNER 1996; HRUŠECKÝ & al. 2002; SPERNER & al. 2002).

CRYSTALLINE CORE OF THE TATRA MTS.

The crystalline massif of the Western Tatra Mts. is composed of metamorphic rocks, mainly metagneisses, migmatites and mica-schists (metasedimentary rocks), as well as orthoamphibolites and orthogneisses. Two tec-tonic units can be distinguished within the crystalline core (JANÁK 1994; POLLER & al. 2000). The lower unit, composed of medium-grade metasedimentary rocks (mica-schists), is exposed in the Western Tatra Mts. only
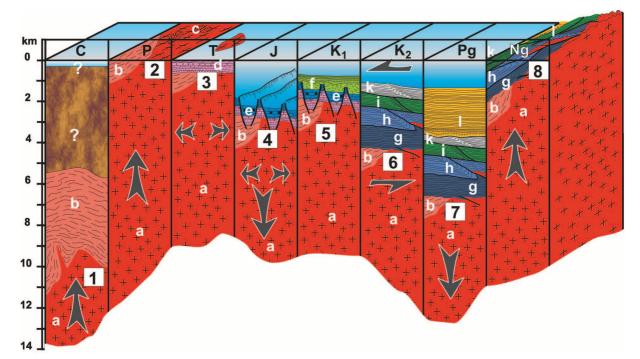

Fig. 4. Changes in the burial depth of the crystalline basement during the tectonic evolution of the Tatra Mts: **1** – age of intrusion and p-t condition: 310-290 Ma after Rb-Sr isochron data (BURCHART 1968); 330±3 Ma Ar/Ar dating in muscovite (MALUSKI & al. 1993); 500 MPa, 600-630°C after xenoliths in calc-silicate metamorphic rocks of the High-Tatra (JANÁK 1993); 341±5 Ma and 700-750°C – High-Tatra diorites and 314±4 Ma – High-Tatra granites after sin-gle zircon data (POLLER & TODT 2000). **2** – subaerial erosion. **3, 4, 5** – extension and normal faulting. **6** – Late Cretaceous thrusting and napping processes: 75±1 Ma – age related to the main period of shearing; 66.6±1.5 Ma – intense mylonitic events (MALUSKI & al. 1993); p ~145-170 MPa, t ~212-254°C (JUREWICZ & KOZŁOWSKI 2003); 7-8 km burial depth during the Late Senonian (KOVÁČ & al. 1994). **7** – plunging during the Palaeogene extension stage. **8** – rotational uplifting (in total ~40° northwards – JUREWICZ 2000a), exhumation and erosion; start of uplift: 36-10 Ma after fission track ages (BURCHART 1972); 70-50 Ma from the depths of 10-11 km (225°C) and 30-15 Ma from depths of 5 km (100°) (KOVÁČ & al. 1994); 11 Ma for the granitoids of the High-Tatra Mts. and 20-12 Ma for the crystalline core of the Western Tatra Mts. after apatite fission-track analysis (STRUZIK & al. 2003); a) granitoids, b) meta-morphic rocks, c) Carboniferous (?), d) Triassic sandstone, shale and carbonate, e) Jurassic carbonate and radiolarite, f) Cretaceous reef limestone and flysch, g) High-Tatric autochthonous cover, h) High-Tatric nappe, i) Krížna nappe, k) Choč nappe, l) Central Carpathian Palaeogene

(JANÁK 1994). The upper unit is divided into two parts. The lower comprises older granites (orthogneisses), paragneisses and amphibolites (JANÁK & *al.* 1996). The higher part of the upper unit contains migmatites (JANÁK & *al.* 1999). The upper unit was intruded by a granitoid plutonic body composed of leucogranite, tonalite and diorite (KOHÚT & JANÁK 1994).

The data referred to by POLLER & *al.* (1999, 2000) suggest multistage granitoid magmatism in the Western Carpathian Mts. At the beginning of Variscan continent collision (Laurussia and Gondwana), the High-Tatra diorites appeared. The age of the intrusion at 341±5 Ma is documented by U-Pb single zircon data of POLLER & TODT (2000). During the final stage of Variscan continent collision, the High-Tatra granites intruded (Text-fig. 4.1). These granites have an intrusion age of 314±4 Ma (POLLER & TODT 2000; cf. BURCHART 1968; MALUSKI & *al.* 1993; JANÁK 1994; GAWĘDA 1995).

In the geodynamic evolution of the Tatra Mts. during the Variscan stage the first tectonic deformation of the crystalline basement was connected with NW-SE thrusting of the upper unit onto the lower unit (FRITZ & *al.* 1992; JANÁK 1994). The second Variscan stage of tectonic deformation is connected with W-E extension. Both stages of deformation were achieved in ductile behavior (KOHÚT & JANÁK 1994). In the Late Permian, the Variscan orogenic belt collapsed (PLAŠIENKA & *al.* 1997).

BEGINNING OF BASIN FORMATION

In the Permian and Early Triassic, the area of the future Tatra Mts. was land (Text-fig 4.2) and, similarly to the platform area (German Basin), fluvial deposits were formed, both in the High-Tatra area and in the Krížna Basin (FUGLEWICZ 1980; cf. RONIEWICZ 1966). The oldest documented sedimentary rocks from the Tatra Mts. are the Verrucano-type conglomerates and shales from beneath the Jahňací Mt. (the so-called Koperšady Conglomerate), preserved only locally below Triassic deposits (PASSENDORFER 1949). After DŻUŁYŃSKI & GRADZIŃSKI (1960) and RONIEWICZ (1966), the Lower Triassic sandstones show radial transport directions of their clasts, to SW in the western part and to SE in the eastern part (see MIŠÍK & JABLONSKÝ 2000).

The Middle Triassic is the period of Tethys expansion and a transgressive event in the platform area (German Basin), where the land area of the Vindelico-Bohemian Massif was located (SZULC 2000). In turn, the Late Triassic was the interval when intense tectonic movements (linked with the old-Cimmerian phase) took place (Text-fig. 4.3). This resulted in the uplift of vast areas of present-day Europe. By the end of the Triassic, during

sedimentation of the Keuper (TURNAU-MORAWSKA 1953) and in the Rhaetian, continental conditions returned to the High-Tatric Basin, with the formation of fresh-water black shales with flora detritus (RACIBORSKI 1890), foot imprints of reptiles (MICHALÍK & *al.* 1976; NIEDŹWIEDZKI 2005) and sphaerolitic iron-ore nodules (RADWAŃSKI 1968; NEJBERT & JUREWICZ 2004).

A very important problem from the point of view of the course of sedimentation, tectonic movement, and rheological behavior is the origin and lithology of the so-called "brecciated Campilian" (Late Scythian) and the basal conglomerate at the base of the Anisian. After KOTAŃSKI (1959a, b, 1965), the Lower Anisian starts with breccias connected with Labine tectonic movements. During this time, periodical emersions of the basin floor took place, with the erosion locally reaching to the Seisian. Both the breccias and erosion could have resulted from extension and normal faulting linked with it, together with the rotation of blocks – Text-fig. 5Ba – (domino-like structures; cf. STEWART & ARGENT 2000; PLAŠIENKA 2003a), which can be observed on the Kominy Tylkowe Mt., where KOTAŃSKI (1959) ascertained the reduction in thickness of the Middle Triassic deposits (see below).

Intraformational breccias can be observed in many places. They are particularly clearly seen on weathered surfaces. The sedimentary character of some breccias does not raise any uncertainties. JAGLARZ & SZULC (2003; cf. MICHALÍK 1997) described numerous tsunamites in the High-Tatric Anisian dolomites indicating tectonic activity during sedimentation within the High-Tatric area, but they questioned the presence of the transgressional breccia of KOTAŃSKI (1959a, b) at the beginning of the Anisian. Most of the Anisian breccias comprise clasts of variable lithology. Field observations indicate the tectonic character of most breccias. Their correlation with certain lithostratigraphic unit by KOTAŃSKI (1959a, b) resulted from their predispositions conditioned by the specific rheological properties of the source rocks. They occur commonly in the so-called "brecciated Campilian" (Late Scythian) and at the base of the Anisian (KOTAŃSKI 1956, 1959a, b, 1961, 1965), which is composed of saline deposits and the "rauwacke" facies (PLAŠIENKA & SOTÁK 1996; MILOVSKÝ 1997; MILOVSKÝ & *al.* 2003). At present, it is considered that the evaporite-lubricated protoliths of rauwacke acted as a detachment horizon during thrusting and folding (WARREN 1999). Such breccias, comprising evaporitic rocks, form a tectonic *mélange* commonly occurring in the Northern Calcareous Alps (Haselgebirge – SPÖTL & HASENHÜTTL 1998) and Central Carpathians (PLAŠIENKA & SOTÁK 1996; MILOVSKÝ & *al.* 2003). In the Tatra Mts. area they typically occur along the thrust surfaces of the High-Tatric nappes (KOTAŃSKI 1959a; JUREWICZ 2003). Presumably cliff breccia and deposits interpreted as an Anisian basal conglomerate may be in many cases tectonic in character, and their for-

mation could be linked with processes resulting from changes of pore fluid pressure leading to hydraulic fracturing and pressure solution (JUREWICZ 2003; JUREWICZ & SŁABY 2004), and referred to as hydrotectonic phenomena (JAROSZEWSKI 1982; MILOVSKÝ & al. 2003; cf. KOPF 1982, 2003). In my opinion the main problem is the lack of macroscopic differences between sedimentary and tectonic breccias and, moreover, the possibility of interference of the tectonic breccia on the sedimentary one.

Triassic deposits in the PKB area are preserved fragmentarily. In the vicinity of the Haligovce Klippe crop out Scythian and Anisian limestones and dolomites showing lithofacies analogies with the High-Tatric and Sub-Tatric Triassic of the Tatra Mts. (BIRKENMAJER 1959c; KOTAŃSKI 1963c). Although in the High-Tatric Basin the Anisian developed as saccharoid dolomites begins with a basal breccia, in the PKB Basin (Haligovce series) the Anisian lies in sedimentary continuity on the Scythian (Campilian – KOTAŃSKI 1963c). According to BIRKENMAJER (1986, 1988), these deposits were formed on the northern slopes of the Exotic Cordilliera, which separated the PKB from the High-Tatric domain (Klape-Manín-Kostelec units), whereas Slovak authors (e.g. RAKÚS & MARSCHALKO 1997) point to their location in the most external part of the Tatricum. No Triassic deposits belonging to the Czorsztyn series are preserved in the area studied but clasts of Triassic rocks occur in the Bajocian-Bathonian crinoidal limestone. Here, pelagic sedimentation began with the end of the Early Jurassic, and changed into carbonate facies by the Middle Bajocian (BIRKENMAJER 1963, 1977; AUBRECHT & al. 1997).
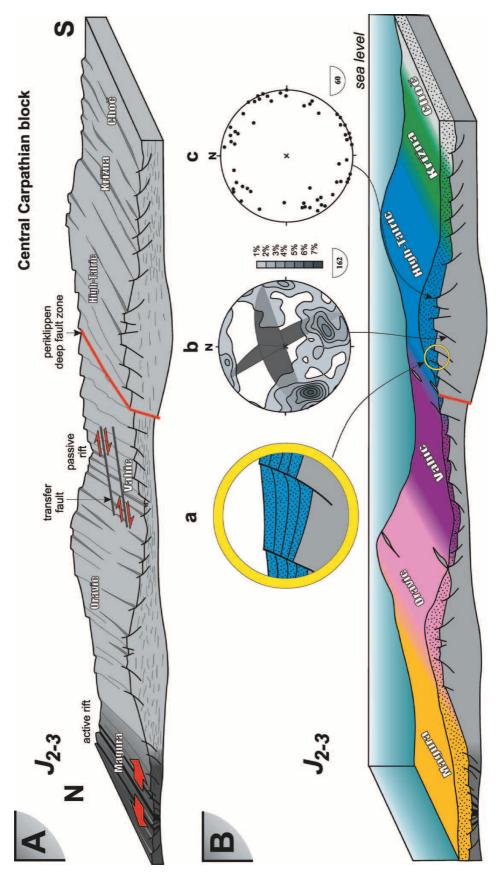
OCEANIC OR THINNED CONTINENTAL CRUST?

The first stage of sea expansion and transgression within the PKB Basin took place in Anisian-Norian times (BIRKENMAJER 1986, 1988; BIRKENMAJER & al. 1990) and was linked with the subsequent eastward lateral propagation of the Alpine Tethys rift (DUMONT & al. 1996). According to PLAŠIENKA (2003b), four principal rifting phases can be discerned within the Western Carpathians based on bathymetric evolution. The two Early Jurassic rifting phases, due to lithospheric stretching and breakdown of the epi-Variscan Triassic platform, were accompanied by crustal heating documented by a radiometrically dated thermal event in the Tatric basement around 200 Ma (MALUSKI & al. 1993; KRÁL' & al. 1997); however, no volcanic activity from that time can be observed in the sediments. The next two rifting phases resulted from the break-up of the South Penninic – Vahic Ocean in the Late Dogger and the North Penninic

– Magura Ocean in the Early Cretaceous (PLAŠIENKA 2003b) (Text-fig. 5A).

At the Triassic/Jurassic boundary, old-Cimmerian tectonic movements took place, which resulted in changes in the character of sedimentation, caused the unconformity between the Triassic and Jurassic, as well as the subdivision of the PKB Basin into particular sedimentary zones (BIRKENMAJER 1959b, c, 1979, 1986). During the Early Jurassic, the High-Tatric series of the Tatra Basin were subject to the greatest variability: the entire succession is preserved in the Kominy Tylkowe sedimentary zone, whereas in other High-Tatric sections there is a large number of stratigraphic gaps (KOTAŃSKI 1959a, b). The Kominy Tylkowe succession shows indirect evidence of synsedimentary normal faulting and blocks rotation. Listric normal faulting caused rotation of the hanging wall from horizontal to steeper inclinations and propagation of fault-related synclines (see STEWART & ARGENT 2000; KHALIL & MCCLAY 2002; IMBER & al. 2003). A similar situation, widespread in the Red Sea area (KHALIL & MCCLAY 2002), could be observed on the above-mentioned Kominy Tylkowe Mt., where KOTAŃSKI (1959a) discovered that the thickness of the Middle Triassic deposits in the Rzędy ridge is ca. 700 m, whereas in the Panienki ridge (within the same block but originally situated ca. 600 m to the NE) it is reduced to 350 m (Text-fig. 5Ba). The change of thickness of the Seisian deposits due to normal faulting within the Kominy Tylkowe area was also commented on by RUBINKIEWICZ & LUDWINIAK (2005); the value of the thickness reduction was estimated as from 250 m to 185 m.

During the Middle Jurassic, the High-Tatric sedimentation zone formed an isolated elevation surrounded by the Vahic Basin in the north and Fatric Basin in the south. The submersion of part of the High-Tatric area during the Bathonian may have been caused by tectonic block movements and accompanied by neptunian dykes (Text-fig. 5Bc; ŁUCZYŃSKI 2001a) and possibly by normal faulting within the granitoid core (Text-fig. 4.4, 5Bb; JUREWICZ 2002). The tectonic activity during the Early Berriasian may be evidenced by carbonate scarp breccias and basic volcanics (limburgite and tuffite interbeds at Osobita Mt.) (LEFELD & al. 1985; STANISZEWSKA & CIBOROWSKI 2000).

In the Krížna Basin, the maximum deepening took place in the Late Jurassic, which could be a reflection of regional depth changes in the ocean as supposed by LEFELD (1981) or postrift thermal subsidence giving rise to eupelagic sediments (PLAŠIENKA 2003). A similar situation occurred in the PKB Basin: the Pieniny zone attained its maximum depth during the Oxfordian (BIRKENMAJER & GĄSIOROWSKI 1961). It is not clear whether or not sedimentation took place on oceanic

crust in this interval. MIŠIK & *al.* (1981) suggest that the fragments of serpentinite, chromic and ferric spinels in the Cretaceous (Aptian) deposits were derived from eroded oceanic crust. The problem is where the oceanic crust was located. The presence of oceanic crust somewhere in the Vahic Basin may be indicated by an exotic block of basaltic rock, several metres in diameter, from the Upper Cretaceous-Palaeocene Jarmuta Formation (Biała Woda Valley), and dated as Aptian-Albian (BIRKENMAJER & PÉCSKAY 2000b; cf. BIRKENMAJER & WIESER 1990; BIRKENMAJER 1996). However, the presence of redeposited basalts testifies neither to sedimentation of the studied part of the PKB Basin taking place on oceanic crust, nor to active rifting processes.

According to PLAŠIENKA (2003b), the absence of rift-related volcanism and the persistence of an extensional tectonic regime for many tens of Ma indicate a passive rifting mode which generated lithospheric stretching and crustal thinning; the rifting-related horizontal tensile stresses were generated by the subducting slab in the Meliata oceanic crust. Thinning of the lithosphere could be possibly due to mechanical and thermal processes connected with an extensional regime and the presence of hot (possibly asthenospheric) upper mantle beneath the rift (see GOLONKA & BOCHAROVA 2000; ZIEGLER & CLOETINGH 2004). PLAŠIENKA (1991) assumed that crustal thinning could be explained by lateral material inhomogeneities in the lower crust. Crustal heating in the Tatric basement was radiometrically documented at about 200 Ma (MALUSKI & *al.* 1993; KRÁĽ & *al.* 1997). According to GHEBREAB (1998), thinning and breaching of the lithosphere are caused by a combination of normal faulting with plastic extension, facilitated by thermal and chemical reactions at the base of the lithosphere.

The lithostratigraphic columns of the Pieniny and Krížna successions are not typical of oceanic crust, although they might suggest its close vicinity. Middle and Upper Jurassic radiolarites (Sokolica and Czajakowa Radiolarite formations), up to 60 m thick, in the Pieniny Unit overlie carbonate deposits (Podzamcze Limestone Formation – BIRKENMAJER 1977). A similar situation is found in the Krížna Basin, where the total thickness of radiolarites reaches ca. 50 m. The total thickness of deposits formed during the Mesozoic within the Krížna

Basin is about 2000 m and does not exceed 800 m in the Pieniny-Kysuca Basin. Neither the lithological succession nor its thickness evidence the presence of oceanic crust, as was already pointed out many years ago by KOTAŃSKI (1963a), who suggested the lack of well-developed initial volcanism. The presence of exotic Triassic pelagic limestone pebbles indicated by BIRKENMAJER & *al.* (1990) is not necessarily evidence for Early Mesozoic oceanic-type rifting but only for the pelagic character of sedimentation and basin propagation (possibly due to thinning of continental crust). No basaltic dykes were noted, which in the rather narrow PKB (120-150 km – BIRKENMAJER 1979) would be registered at least by the presence of tuffites. At Poiana Botizei in the Romanian part of the PKB, fragments of ophiolites connected with pre-Callovian spreading of oceanic crust and autochthonous porphyric tuffites connected with island-arc volcanism were noted (BOMBIŢĂ & SAVU 1986). Basaltic rocks with pillow lavas covering a Middle Jurassic-lowermost Cretaceous carbonate sequence are exposed within the Ukrainian part of the PKB (KROBICKI & *al.* 2004). In the Polish part of the PKB similar evidence of volcanic activity was not noted; in turn, manifestations of extension and crustal thinning indicated by normal faulting could be observed, both in the Czorsztyn Basin of the PKB (BIRKENMAJER 1963; KUTEK & WIERZBOWSKI 1986; WIERZBOWSKI & *al.* 1999) as well as in the crystalline rocks of the Tatra Massif (JUREWICZ 2002; JUREWICZ & BAGIŃSKI 2005), and in its sedimentary cover (ŁUCZYŃSKI 2001a).

According to BIRKENMAJER (1986), the Czorsztyn Ridge was an "aseismic-type ridge" separated from the North European platform by normal faulting. Large-scale tectonic activity in the Czorsztyn Basin is testified by the presence of scarp breccias (BIRKENMAJER 1963; MIŠÍK & *al.* 1994), exotic material (MIŠÍK & AUBRECHT 1994), neptunian dykes (WIERZBOWSKI & *al.* 1994; AUBRECHT & *al.* 1997; AUBRECHT & TÚNYI 2001), sedimentary hiatuses (e.g. in the vicinity of Szlachtowa – from the Bajocian to the Albian – JUREWICZ 1997), as well as the character of the sedimentation. For instance, the Bajocian crinoid limestones were most probably formed not as beds but as heaps deposited below scarps and morphological margins (WIERZBOWSKI & *al.* 1999). It seems that extensional

Fig. 5. Stages of tectonic evolution of the Tatra Mts. and Pieniny Klippen Belt. A – Thinning of the crust due to extension and rifting processes causing lateral anisotropy (horizontal lines). Note the oblique boundary between the Vahic rifting basin and the Central Carpathian Block causing its further rotation and activation of the shear zone (see Text-fig. 4A). B – Sedimentation basins within the area near the boundary of the Central (northern part) and Outer Carpathians during the Late Jurassic and Early Cretaceous. Diagrams (lower hemisphere): a) scheme of listric normal faulting causing rotation of hanging wall from horizontal to steeper dip, creation of small depositional domains and propagation of fault-related synclines; b) stereoplot of shearing zones within the granitoid core of the Polish High-Tatra Mts. (see JUREWICZ 2002); c) poles to neptunian dykes within the High-Tatric nappe and autochthonous cover (after ŁUCZYŃSKI 2001a); b, c - position prior to ~45° westward horizontal and ~40° northward vertical rotation around horizontal axis (~90/0) of the Tatra block (JUREWICZ 2000a, b). Note deepening and unification of facies within the Vahic-Oravic basin. Not to scale

thinning of the continental crust took place in the basement of the Czorsztyn succession. Identical opinions were presented by GOLONKA & KROBICKI (2001), who indicated from palaeoclimatic data that the uplift of the Czorsztyn Ridge could have been a result of extension during the Jurassic supercontinental break-up. During the Late Cretaceous there was no expression of the Czorsztyn Ridge in the morphology of the basin floor: conditions of sedimentation and facies were more unified in the whole Vahic-Oravic basin (BIRKENMAJER 1977, 1986). This may have been the result of continuing strong crustal thinning at the basement of the Czorsztyn basin.

During extensional processes, normal faulting was responsible for graben–horst formation and morphological escarpments connected with them (see WEISSERT & BERNOULLI 1985), as well as low-angle detachments and horizontal anisotropy in the basin basement rocks that were utilized as *décollement* surfaces during later nappe-folding stages (see GHEBREAB 1998; GHEBREAB & TALBOT 2000). Normal faults near the axial part of the central rift could evolve from extension fractures and, having reached a critical depth, turned into normal faults (like in Ethiopia – see ACOCELLA & *al.* 2003). Morphological escarpments, which formed along the margins of these faults, could be the source of detrital material. Listric normal faults could sole out at the base of the sedimentary rocks, resulting in a mechanical discontinuity along the rheological boundary that would facilitate future thrust-faulting.

During the Jurassic, two episodes of basin deepening (down to several thousands of metres) took place in the Krížna Basin: one in the Bajocian and Bathonian, and the second in the Oxfordian (PASSENDORFER 1983). The expansion of the basin could have been accompanied by extension and normal faulting and domino-like structures (PROKEŠOVÁ 1994; PLAŠIENKA 2003a). The system of mylonitic and cataclastic zones in the granitoid core of the High-Tatra Mts. can most probably be linked with this stage (Text-fig. 5Bb). Simultaneously, neptunian dykes could develop (Text-fig. 5Bc) within the sedimentary rocks of the High-Tatric series (ŁUCZYŃSKI 2001a). Most of these dykes have an orientation similar to that of the normal faults within the granitoid core (Text-fig. 5Ba, b). Examples of synsedimentary activity have also been described from the Krížna Nappe: extensional faults enabling fluid migration and resulting in manganese-bearing sedimentation during the Early Toarcian (Siodło Fault – BAC-MOSZASZWILI 1998A; JACH & *al.* 2003).

The initial rifting took place in the axial deepest part of the basin ("ultra-Pieniny") during the mid-Late Jurassic; its rate rapidly decreased due to sea floor spreading within the Magura Basin (BIRKENMAJER 1986; WINKLER & ŚLĄCZKA 1994; GOLONKA & *al.* 2000). The

PKB Basin could have developed as a narrow rift zone, ~100 km wide, with intense normal faulting under increased heat flow condition (see CORTI & *al.* 2003). The lack of volcanic activity indicates "passive" rifting of ZIEGLER & CLOETINGH (2004). A comparison can be made with the Red Sea Basin, which started opening about 25 Ma ago, and in which basaltic rocks indicate that sea floor spreading began at only about 4-5 Ma ago (GHEBREAB 1998). In the considerably narrower PKB Basin, oceanic crust probably did not develop within its Polish part. On the other hand, oceanic crust could have existed in the more westerly part of the PKB Basin, as is directly testified by the eastward lateral propagation of the Alpine Tethys rift (DUMONT & *al.* 1996) and the wedge-like opening of the PKB Basin (from the west). The closest area with documented slices of amygdaloidal basalts is the Belice Unit in the Považský Inovec Mts. (SOTÁK & *al.* 1993). Indirect evidence of the lack of oceanic crust at the base of the PKB basin could be the long duration of thrust-folding and slow speed of subduction in comparison to the Outer Carpathians area (see below).

The passive rift could have been orientated obliquely to the southern margin of the PKB Basin. The rift orientation was ~SW-NE or ~WSW-ENE; the WSW-ENE orientation was documented after MARCHANT & STAMPFLI (1997), STAMPFLI & *al.* (1998), DUMONT & *al.* (1996) within the Ligurian-Piemont and Vahic oceans, and the SW-NE orientation was promoted after GOLONKA & KROBICKI (2001) and AUBRECHT & TÚNYI (2001) in the Pieniny-Magura ocean. According to SZULC (2000), crustal motion during the Triassic Tethys rift formation, was transmitted onto its northern periphery by reactivated Hercynian master faults. The NW-SE orientation of Hercynian deep fault zones was pointed out by ARTHAUD & MATTE (1977); faults of such orientation could have evolved as transfer faults within the Vahic-Oravic.

The presence of a counter-clockwise strike-slip zone along the southern margin of the Vahic Basin near the contact with the Central Carpathian Block was an effect of oblique subduction (TOMEK & HALL 1993; ZEYEN & *al.* 2002) in consequence of oblique extension. Similar pattern is observed within the Gulf of Aden (GHEBRAB 1998, CORTI & *al.* 2003) and in the Gulf of California (UMHOEFER 2000). According to WEISSERT & BERNOULLI (1985) the latter presents an actualistic example for the tectonic pattern of the Ligurian-Piemont Ocean. The result of polyphase activity of this shear zone was the existence of diachronous basins of pull-apart character, for example Manín or Kostelec (RAKÚS & MARSCHALKO 1997). This model explains the formation of the Myjava Basin of BIRKENMAJER (1986), DUDZIAK

(1990) and Kostka (1993), and of the Złatne Basin of Sikora (1971) and Golonka & al. (2000). These basins could have been formed through the same mechanism as that of the Gosau basins described by Wagreich (1995).

ANDRUSOV RIDGE AND MANÍN BASIN

In the opinion of Birkenmajer (1986, 1988), the Andrusov Ridge (earlier referred to as the Exotic Ridge or Exotic Cordillera) separated the PKB Basin and the High-Tatric domain *sensu lato* (Klape-Manín-Kostelec units) during the Cretaceous. At that time, the Andrusov Ridge was a source of clastic material and showed volcanic activity connected with the subduction of oceanic crust under the Central Carpathians (op. cit.). Initially, this structure was the passive northern margin of continental crust detached from the southern edge of the North European Plate (Birkenmajer 1988). The northern slope of this ridge was probably occupied by the Haligovce succession, and the southern slope by the Klape-Manín-Kostelec succession, transitional to the High-Tatric sedimentation zone (op. cit.*).* A detailed reconstruction of the Andrusov Ridge was made after Birkenmajer (1986, 1988) and Birkenmajer & al. (1990), based on the composition of exotic material. The ultramafic rocks (see Mišík & al. 1980) occurring within the Klippen units (Klape and Manín successions) were linked by Birkenmajer (1988) with obducted oceanic crust that was probably associated with the Andrusov Ridge. As an alternative possibility, the uplift of some segments of oceanic crust along some transform faults cutting into the northern slope of the Andrusov Cordillera was suggested.

Borza (1966), Marschalko (1975), Mišík & Sýkora (1981), and Mišík & al. (1991) noted the differentiated composition of the Exotic Cordillera, varying in its particular parts. Potfaj (1997) also pointed out the inhomogeneities of the clastic material source area, linked the individual stages of the cordillera and joined the evolutionary sequence of the Pieniny-Exotic-South-Magura cordilleras in one heritage chain.

The Manín Unit was initially included by Mahel' (1950) in the Manín-Inovec series. According to Kysela & al. (1982), the Manín Unit was located north of the Tatric and south of the PKB. Mahel' (1978) included the Manín "Series" in the Vysoká Nappe, which implied its position within the Fatricum. Plašienka (1995a) assigned the Manín Unit to the Fatricum, in a position lateral to the root zone of the Vysoká Nappe (northernmost Fatricum). Furthermore, he considered it a rather local succession without lateral continuation. Plašienka (1995b) linked the provenance of the exotic material

(e.g. the Jurassic blueschists in the Klape Unit noted by Dal Piaz & al. 1995) with the interior of the Carpathians and with the closing of the Meliata oceanic domain. During the mid-Cretaceous stage, compressional uplift of the Veporic-Gemeric-Meliatic pile took place, the top of which came to be exposed to intensive erosion and could be a source of exotic material (Plašienka 1997). Mišík (1996) questioned both the position of the Exotic Cordillera of Plašienka (1995a, b), and his concept of the palaeogeographic intra-Veporic position of the Klape and Manín basins. Rakús & Marschalko (1997) also did not accept Mahel's (1978) and Plašienka's (1995a) idea of the infra-Veporic placement of the Manín nappe. They discussed the palaeogeographic position of the Manín, Klape and other basins at the boundary of the Central and Outer Carpathians and considered that the Manín area must have been situated externally from the Tatric block. Rakús & Marschalko (1997) considered the Andrusov Ridge controversial, and linked the problems with the arrangement of the Western Carpathians region with the existence of a large strike-slip shear zone between the edge of the Central Carpathians and the Klippen Belt, which functioned from the Albian to the Palaeogene and caused the rotation of the CWC.

The reconstruction in Text-figs 5B and 6A has been based on the assumption that the source of exotic material could be linked with a deep fault zone along the northern margin of the Central Carpathian Block, which was activated repeatedly as a shear zone and a plate edge as well. Other sources of exotic material, ephemeral in character, could be connected with local and periodical emersions along listric normal faults causing blocks rotation. The appearance of exotic fragments could be linked with rotated blocks and local emersion of the basement, e.g. in the Bathonian pelagic deposits within the Niedzica series (Birkenmajer & al. 1960). Part of the exotic material could also have originated southwards, perhaps in the Meliata area.

"BLACK FLYSCH" – JURASSIC OR CRETACEOUS?

One of the greatest controversies in the investigations within the PKB was linked with the so-called "Sztolnia beds", also referred to as "black flysch", "black Cretaceous" or "Aalenian flysch". According to Birkenmajer & Pazdro (1968), Błaszyk (1968), Birkenmajer & Myczyński (1977), and Krawczyk & Słomka (1981, 1986), the black flysch is considered to be Aalenian in age, whereas Blaicher & Sikora (1969), Sikora (1971) and Książkiewicz (1972) included it in the Cretaceous. Neither of the opposing sides, however, considered a third possibility, namely that of similar

lithologies of flysch deposits in both the Jurassic and Cretaceous. One important piece of evidence may be the occurrence of the Bryjarka Formation in a borehole in Szczawnica (BIRKENMAJER & al. 1979). This formation does not crop out, has a similar lithology to the beds in dispute, and is of Cretaceous age. In my opinion, inclusion of the so-called "black flysch" in the Cretaceous would simplify the mapping, because in many places, i.e. in the vicinity of Jaworki in the Magura Unit close to the PKB (distinguished by BIRKENMAJER 1970, 1977, 1979, 1986 as the Grajcarek Unit, and by GOLONKA & al. 2000 as the Hulina Unit), or in the Szczawnica borehole succession (BIRKENMAJER & al. 1979) it contacts directly with the flysch of the Jarmuta Formation and other Cretaceous units. Controversies regarding the age of the "black flysch" still persist. OSZCZYPKO & al. (2004) determined the age of the "black flysch" as Albian-Cenomanian; on the other hand, BIRKENMAJER & GEDL (2004) suggested an Aalenian-Bajocian age for this deposit based on dinocyst data.

In my opinion, the problem with the "black flysch" lies most probably not in the issue of which age it is, but rather which flysch is Jurassic and which Cretaceous.

GEOTECTONIC SETTING AND AGE OF THE NAPPE-FOLDING

During the Late Cretaceous-Eocene, the Carpathian area was a part of a larger Alpine-Carpathian orogen formed by the south-eastward subduction of the Penninic Ocean (135–55 Ma) and later collision of the European and Adriatic continents (55-40 Ma; NEMČOK & NEMČOK 1994; NEMČOK & al. 1998).
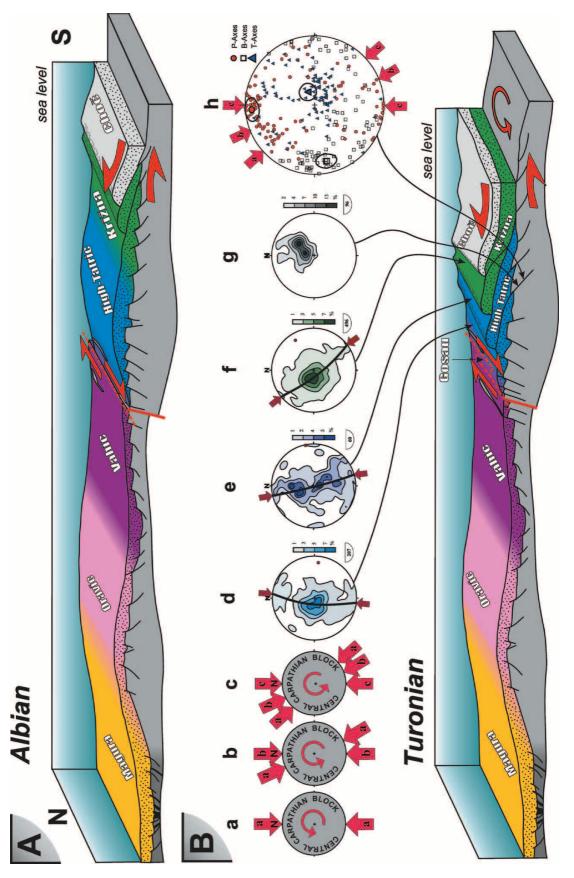
The northwards migration of nappe movement from the Meliata-Hallstatt to the Penninic-Vahic oceanic zones (PLAŠIENKA 1996) is well documented by the diachronous position of the pre-orogenic flysch: in the Krížna Basin in the Early Aptian; in the High-Tatric Basin (Text-fig. 4.5f) in the Early Albian (KRAJEWSKI 1980, 2003; LEFELD & al. 1985); in the Klippen Basin in the Coniacian (BIRKENMAJER 1977). Progressively younger ages of sediments were involved in the cover nappes towards the foreland: Tithonian in the Silicum,

Neocomian in the Hronicum, Albian in the south-eastern Fatricum, Cenomanian in the north-western Fatricum and south-eastern Tatricum, Early Turonian in the north-western Tatricum (e.g. NEMČOK & NEMČOK 1994), Early Santonian – in the Klape unit, Campanian in the Pieniny and Kysuca units (SCHEIBNER 1968), Maastrichtian in the Czorsztyn unit (BIRKENMAJER 1977; BIRKENMAJER & al. 1987). Compressional deformation progressed from the hinterland to the foreland lower plate (PLAŠIENKA 2003a). The age of thrusting and folding was progressively younger from south to north (Text-figs 3, 6). BEZÁK (1991) suggested a NW direction of thrusting of the Veporic Mesozoic cover and Gemericum over the north-western parts of the Veporicum during the Late Cretaceous. In the area more to the north-east, K-Ar dating indicates an Albian age for NW-SE thrust-fold shortening in the Krížna Nappe (NEMČOK & KANTOR 1990).

According to SCHEIBNER (1968) and BIRKENMAJER (1970, 1974), Late Cretaceous folding took place within the PKB in the Klape Unit in the Coniacian-Early Santonian; in the Pieniny and Branisko-Kysuca units in the Santonian-Campanian; in the Niedzica Unit in the Campanian-Maastrichtian (Text-figs 7-8). The folding continued within the peri-Magura Basin (i.e. in the Grajcarek Unit of BIRKENMAJER 1986) up to the Late Palaeocene (Text-fig. 8B), which is testified by deposits of the synorogenic Jarmuta Formation (BIRKENMAJER & al. 1987; BIRKENMAJER & DUDZIAK 1991).

During the Aptian-Cenomanian, thrust-napping processes occurred within the northern Veporicum and Fatricum (PLAŠIENKA & al. 1997). According to PLAŠIENKA & PROKEŠOVÁ (1996) and PLAŠIENKA & al. (1997), thinned continental crust of the Krížna Basin (Fatric) was underthrust along the north Veporic zone, and the Mesozoic deposits of the Krížna sedimentation zone were detached and glided gravitationally northwards to overlie the Tatric in the Late Turonian. The concept of gravitational slumps reaching back to the 1960s (KOTAŃSKI 1961, 1963a) was later questioned by BAC-MOSZASZWILI & al. (1984). The most important argument against this idea is clearly visible after reversing the Tatra block to a position prior to the Miocene rotational uplifting (GRECULA & ROTH 1978; PIOTROWSKI 1978;

Fig. 6. A – Beginning of thrusting and napping processes started from the Choč basin in the Albian. The oblique boundary between the Central Carpathian Block and the Vahic rifting basin caused activation of the shearing zone and rotation of the Central Carpathian Block. B – Scheme of geometry of the Late Cretaceous (Turonian) thrust-napping within the Tatra Mts. Changes of the direction of thrusting and rotation of the Central Carpathian Block (a-c) during Late Cretaceous nappe-thrusting documented on stereoplots of bedding within the High-Tatric autochthonous cover (d), High-Tatric Nappe (e) and Krížna Nappe (f); stereoplot of slickenside fault planes within the granitoid core of the High-Tatra Mts. (g); (d-g) after southward 40° rotation around 90/0 axis, Polish part of the Tatra Mts.; (h) reconstruction of palaeostress based on faults striation documenting the same process of rotation (TectonicsFP Programme, P-B-T method, dip-slip faults rotated about 40° southward to the position before Neogene uplifting; (a-f – after JUREWICZ 2000b; g, h – after JUREWICZ 2000a). Not to scale

Bac-Moszaszwili & al. 1984; Jurewicz 2000a). Both the surfaces of the nappe thrusts and the thrust fault planes within the granitoid core of the Tatra Mts that are geometrically linked with them (Jurewicz 2000a) then attain southerly dips (average 25-30° – Text-fig. 6Bg), which, according to Bac-Moszaszwili & al. (1984), points to their origin as shears. The next argument is the fact that the deformation structures of the High-Tatric and Krížna nappes originated in a compressional regime, in contrast to the structures observed in the PKB, where gravitational sliding was documented in Małe Pieniny, manifested by normal faulting and folding caused by morphological barriers, e.g. Czajakowa Skała fold – Jurewicz 1994, 1997). The processes taking place along the thrust surfaces in the Tatra Mts. also point rather to a compressional type of deformation (Jaroszewski 1982; Plašienka & Soták 1996; Jurewicz 2003; Milovský & al. 2003; Jurewicz & Słaby 2004).

Birkenmajer (1986) accepted the presence of a separate subduction zone under the Andrusov Cordillera and another one under the Czorsztyn Ridge: during the Campanian – Early Palaeocene the subduction zone jumped from the southern margin of the PKB Ocean to the northern margin of the Czorsztyn Ridge and began to consume the Magura Basin. The present day distance between the supposed suture zones, which were to originate in place of the former subduction zones, is ca. 20 km, which is less than the thickness of lithospheric plate. It should also be noted that when the existence of the Andrusov Ridge and ocean-type rift in the PKB Basin is rejected, the width of the sedimentary zone of the PKB, evaluated at 120-150 km by Birkenmajer (1979, 1986), becomes much smaller. It is difficult to find arguments confirming the presence of a subduction zone under the hypothetical and controversial Andrusov Ridge of Birkenmajer (1986). According to e.g. Soták (1992) and Oszczypko (1999), a subduction zone was located under the Czorsztyn Ridge and connected with the so-called Mesoalpine suture zone. This is testified by geophysical sections, e.g. worked out by Uchman (1973), Lefeld & Jankowski (1985) and Nemčok & al. (1998): the PKB is linked to a high conductivity zone and an elevated lithosphere/asthenosphere boundary south of the Klippen belt. Therefore, the problematic issue remains whether or not continental crust was present in the basement of the Czorsztyn Basin, and did subduction take place under the Czorsztyn Ridge. It is more probable that the northern margin of the Central Carpathian Block (as yet not divided into separate massifs) could have been a continental margin, under which subducted thinned crust layed at the base of the Vahic-Oravic Basin. In a structural sense, the boundary between the Central Carpathians and the PKB could have been manifested at this stage in sedimentation as a deep fault zone and a fault scarp providing a source of detrital material, and as a large shear zone (Text-figs 6-8), the sinistral activity of which was noted from the Albian by Rakús & Marschalko (1997). Structural analysis conducted by me indicated counter-clockwise rotation of the Carpathian Block and dextral shearing during Late Cretaceous – Palaeogene nappe-thrusting – Text-fig. 6B (Jurewicz 2000a, b).

"RUSINOWA CONGLOMERATE" – SENONIAN OR EOCENE?

In older literature, the thrust-folding in the Tatra Mts. was estimated to have taken place after the Turonian and before the Eocene (Andrusov 1965). By analogy with the Gosau Group, synorogenic sediments of the Malé Karpaty Mts. and from the vicinity of Šumiac in the Nížke Tatry Mts. (Andrusov & al. 1973) were related to the Turonian/Senonian boundary (Wagreich 1995; Plašienka 1995a, b, 1997; Linzer & al. 2002). Birkenmajer (2000) stated that the Rusinowa Polana conglomerates related to the basal unit of the "Palaeogene Podhale Succession" by Sokołowski & Jaczynowska (1979) and to the "Nummulitic Eocene" by Bieda (1963) and Bac-Moszaszwili & al. (1979), should be linked with the Upper Cretaceous Gosau Group of the Northern Calcareous Alps. According to Birkenmajer (2000), the Rusinowa Conglomerate Formation could be the equivalent of the Senonian Gosau Group conglomerate (Samuel 1977) resting on the Gemericum near the Dobšinska Ice Cave in Central Slovakia. Conclusions drawn from Birkenmajer (2000) are of crucial importance for the tectogenesis of the Tatra Mts., although they are based on selectively chosen prerequisites and not new facts. Not wanting to discuss the age of the so-called Rusinowa Polana Conglomerate, which at present should be regarded as undetermined, I would like to draw attention to the following aspects:

1) Assuming after Birkenmajer (2000) a dip of only 20° for the Zakopane beds in the Podhale Trough, it should be noted that the Rusinowa Polana conglomerate at present occurs ca. 250 m below the theoretical base of these beds. The present area of the Tatra Mts. would therefore have been a morphological depression already in the Late Cretaceous. If a larger dip is assumed (i.e. Passendorfer 1983 – 30-60°; Jurewicz 2000 – 40°), the depth of the Late Cretaceous erosional incision in relation to the Zakopane beds would increase by ca. 600-800 m.

2) The Rusinowa Polana conglomerate occurs in the western margin of the so-called Szeroka Jaworzyńska

Depression. Thus, if the Late Cretaceous erosion within the tectonic depression reached down to the Middle Triassic of the Krížna Nappe, then in the neighbouring – potentially more exposed to erosion – Koszysta elevation, granitoids would have appeared at the surface already in the Late Cretaceous. KOTAŃSKI (1961) and GŁAZEK (1963) assumed the possibility of deep erosion on the Koszysta elevation; however fission-track analysis of the Tatra Massif uplift age pointing to Late Miocene times (BURCHART 1972; KRÁĽ 1977) excludes this possibility. According to PASSENDORFER (1983), granite boulders are not observed even in Palaeogene sediments; only the Krížna and Choč Nappes were subject to erosion at that time. The argument that nappe-thrusting of the Krížna Nappe on the High-Tatric tectonic units (BAC-MOSZASZWILI & *al*. 1981) was preceded by erosion accompanied with karst-weathering or that the High-Tatric autochthonous cover was partly eroded during the Late Cretaceous (GŁAZEK 1963), should be rejected in the light of the documented processes of pressure solution accompanied by hydrothermal solutions taking place along thrust planes (JUREWICZ 2003; MILOVSKÝ & *al*. 2003; JUREWICZ & SŁABY 2004).

3) The thickness of the conglomerate estimated at 25 m and the small morphological differentiation (several tens of metres) observed in the field, allowed BIRKENMAJER (2000) to state that the Rusinowa Polana conglomerate lies at 100-150 m below the present-day surface level and to accept a retro-arc thrusting with a range of several hundreds of metres as well as to insert the thrust planes at ca. 500 m below the surface level. It was not the actual field exposures but their interpretations that were used by him as the basis of distinguishing a precisely timed separate tectonic stage (Laramide phase: Late Senonian – Early Palaeocene).

4) Determining the age of the Rusinowa Polana conglomerate as Late Cretaceous does not imply the acceptance of a separate folding phase in the PKB of retro-arc character, and linked to the Laramide phase, for several reasons:

a) the retro-arc thrusting in the PKB during the Laramide phase is questionable (JUREWICZ 1994, 1997);

b) accepting the assumed dip of the thrust planes at 70°N (BIRKENMAJER 2000), the gravitational mechanism of the southwards thrust of the nappe units on postorogenic sediments would require enormous structural reworking, regardlessof the extent of the Miocene rotation (20° – PIOTROWSKI 1978; 25° – BAC-MOSZASZWILI & *al*. 1984; or 40° – JUREWICZ 2000a).

c) it is much simpler to link the presence of reverse faults documented by SOCHACZEWSKI (1997) with the Miocene stage of the Podhale Trough formation and the resulting local compression in the internal bend of the "synclinorium".

The concept presented by BIRKENMAJER (2000) requires the following scenario of events. The Tatra Mts., folded underwater in the Turonian, must have been elevated enough to allow erosion preceding the sedimentation of the Rusinowa Conglomerate Formation to reach the Carpathian Keuper sediments of the Krížna Nappe in the Late Cretaceous, in consequence of which the granitoids on the Koszysta elevation would have been exposed to the surface. By the end of the Early Palaeocene the northern part of the Tatra Mts. must have been elevated enough to cause southward gravitational slumps. This scenario does not find confirmation in the fission-track analyses of the age of the Tatra Mts. uplift, which point to Late Miocene times (BURCHART 1972; KRÁĽ 1977), and not the Late Cretaceous. Accepting such a course of events in which nappe-thrusting would have taken place in one extremely fast event does not find confirmation in the character of deformations occurring along the thrust planes, which require long, repeatedly reactivated processes (JUREWICZ 2003).

The age of the Rusinowa Polana conglomerate remains undocumented. However, its Eocene age becomes more plausible through the discovery of tuff horizon within the conglomerates (A. SOCHACZEWSKI – personal communication) and its presumed equivalence with Eocene tuffite documented in the northern margin of the Tatra Mts. and in boreholes (GŁAZEK & *al*. 1998; KĘPIŃSKA & *al*. 2000). The tuffite in conglomerate of the Rusinowa Polana was indirectly evidenced earlier by SOKOŁOWSKI (1978), reported as a clay interbed, and interpreted subsequently as tuffite by GŁAZEK & *al*. (1998).

CONDITIONS OF DEFORMATION AND GEOMETRY OF NAPPE–THRUSTING IN THE TATRA MTS.
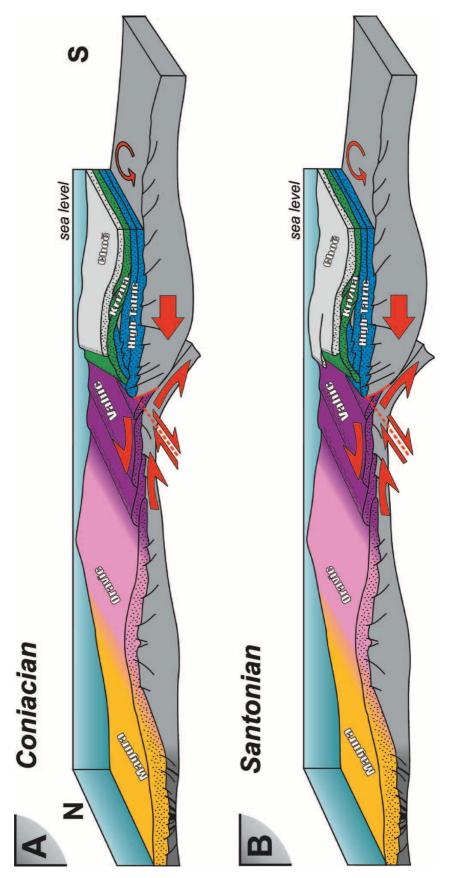
It can be assumed that the thinned continental crust at the basement of the Fatric-Tatric Basin has a lateral rheological heterogeneity (Text-fig. 5A) connected with lithosphere extension and rifting (cf. VAUCHEZ & *al*. 1998). This kind of inherited low-angle anisotropy favoured the formation of a detachment and major flat-lying thrust (see GHEBREAB 1998). Above the roof thrust, flat-lying, slightly disturbed strata could be present, and imbricate structures or a hinterland dipping duplex could be formed below (see BOYER & ELLIOT 1982). Such a simple pattern of folding, thrusting and formation of duplex was most probably disturbed initially by great –

both horizontal and vertical – lithological variability. Normal faults already generated during lithosphere extension and rifting of the sedimentary basin (PLAŠIENKA & *al.* 1997), could undergo activation as reverse faults due to later compression related to Alpine folding (PLAŠIENKA 2003a). When subduction completely consumed the basement of the southernmost sedimentary zones, the crystalline basement of the High-Tatric series also underwent compression and shearing processes (Text-figs 4.6, 6Bd, e). The fact that crystalline rocks also compose the nappes and that they were detached at depths of at least several or more kilometres, suggests that the detachment was preceded by compression, which resulted in reverse faults (e.g. BAC-MOSZASZWILI & *al.* 1984). Reverse faults could originate due to the change of the sense of movement of initially normal faults, which is evidenced e.g. by the angle of 60° between the surface of sedimentary contact of Seisian sandstones with the crystalline rocks and the Giewont Unit thrust (BAC-MOSZASZWILI & *al.* 1979). Most thrust surfaces and faults linked with Alpine folding are, however, characterised by flat and – befor rotational uplift – southern dip (Text-fig. 6Bg). Analysis of tectonic transport directions based on striae on such slickensides occurring within caps of crystalline rocks (so-called "Goryczkowa Island") was made by BURCHART (1963). Flat-dipping slickensides, geometrically linked with Alpine thrust folding from the granitoid core of the High-Tatra Mts. enabled the reconstruction of the Late Cretaceous stress field (Text-fig. 6Bh; JUREWICZ 2000a). The structural analysis of the crystalline core and the nappe units was preceded by reversal of the Tatra block to a position prior to the Miocene rotation linked with the uplift process (BURCHART 1972, KRÁĽ 1977, KOVÁČ & *al.* 1994). The rotation angle comprises the dip of the Eocene strata and the dip of the para-autochthonous sedimentary cover (PIOTROWSKI 1978; BAC-MOSZASZWILI & *al.* 1984), displacement on the sub-Tatric fault (see SPERNER 1996; KOHÚT & SHERLOCK 2003) as well as the gradient of erosion of the granitoid core (BURCHART 1972). The vertical rotation angle of 40° assumed herein was evaluated mainly on the basis of dips in the sedimentary cover of the granitoid core. The northern boundary between the Tatra Mts. and the Podhale Trough, with a parallel orientation 90/0 is taken as as the rotation axis (JUREWICZ 2000a, b). The reconstruction suggests that, at the assumed rotation angle in the stress field reconstructed by four methods in TectonicsFP soft-

ware, the main compression axis orientation is ca. ~355/10, axis *b* ~265/15, and axis *c* ~100/75 (JUREWICZ 2000a). The PTB method (Text-fig. 6Bh), calculating the orientation of the stress field for each fault plane/striae system, indicates that during nappe-thrusting either a relative change of the largest stress orientation (σ_1) from NW to N took place, or at a stable stress field counter-clockwise rotation of the basement occurred (Text-fig. 6Ba-c). Such rotations were indicated in different units of the Outer Carpathians (KRUCZYK & *al.* 1992; GRABOWSKI 1995a, b, 1997b; GRABOWSKI & NEMČOK 1999), along the boundary of Central and Outer Carpathians (BIRKENMAJER 1976, 1985) as well as in younger lithological complexes, i.e. in the Nummulitic Eocene (GRABOWSKI 1997a).

As for the relative direction of tectonic transport, and the resulting rotation direction, there is no doubt that the NW orientation is older than the N orientation (JUREWICZ 2000a, b). This can be inferred both from field observations of striae on slickensides within the granitoid core, as well as from diagram analysis (Text-fig. 6Bg). Older NW striae destroyed by younger, more northward directed striae can be observed on single slickenside planes. On the PTB diagram (Text-fig. 6Bh) the maximum is located at N, which probably directly results from the younger ages of the striae.

Other evidence of the counter-clockwise rotation of the basement during Alpine folding (Text-fig. 6Ba-c) comes from an analysis of the bedding, which points to the fact that higher, older units indicate at present a NW compression (Krížna and High-Tatric nappes – Text-fig. 6Be, f), whereas the autochthonous sedimentary cover shows a N orientation (Text-fig. 6Bd; JUREWICZ 2000b). Similarly, BURCHART (1963), based on analysis of striae in crystalline rocks of the so-called "Goryczkowa Island", indicated 340°-orientated tectonic transport of the Giewont Unit (High-Tatric Nappe).

The possibility of later block rotations in the basement of the Krížna Nappe has already been pointed out by BAC-MOSZASZWILI (1998a) and MARKO & *al.* (1995). In the opinion of the latter authors, this clockwise tendency of stress rotation or counter-clockwise rotation of the basement probably also persisted in the Western Carpathians during the Palaeogene-Neogene tectonic evolution (cf. JAROSIŃSKI 1998). These authors reveal the importance of rotation and translation movement of the rigid crustal blocks inside the wrench zone between the North-European and African plates. My own structural

Fig. 7. Coniacian-Santonian. Beginning of thrust napping processes within the Vahic area connected with activation of the subduction zone of Vahic thinned lithosphere under the Central Carpathian Block (A). Note that the shortening at the base of the Vahic basin caused the continuation of thrust and nappe-folding within the Tatra Mts. area. In some cases part of the Choč Nappe could be thrusted onto units of the Pieniny Klippen Belt (B). Well marked is the convergence of the subduction zone and oblique motion of the Vahic lithospheric plate under the Central Carpathian Block. Not to scale

investigations within the crystalline massif and sedimentary cover (JUREWICZ 2000a, b) have enabled me to evaluate the range of the counter-clockwise basement rotation during post-Turonian and pre-Eocene nappe-folding processes at ca. 45°.

Investigation of fluid inclusions in syn-kinematically grown quartz on slickensided faults within the granitoid core of the High-Tatra Mts. proved that syn-kinematic quartz slickenfibres crystallised at a pressure of about 1.45-1.7 kbar (145-170 MPa) and at a temperature of ~212-254°C (JUREWICZ & KOZŁOWSKI 2003). These values of pressure and temperature enable estimation of the depth of the deformation processes linked with Alpine thrusting at 6-7 km (Text-fig. 4.6) and the geothermal gradient for the Late Cretaceous at ca. 30°C/km (JUREWICZ & BAGIŃSKI 2004). Such high values of the geothermal gradient could result from the upwelling of the asthenosphere indicated e.g. by PLAŠIENKA (2003b), which could cause the crust to become elastic in the lowermost part and ensure that detachment could not take place deeper (cf. ORD & HOBBS 1989).

Geometric analysis of tectonic transport orientation and reconstruction of the stress field cannot be applied in relation to the nappe units and the autochthonous sedimentary cover because of the ductile type of deformation (JUREWICZ 2003). In the para-autochthonous sedimentary cover and in the nappe units, there are – contrary to in the PKB – only a few slickensides linked with the nappe-folding: these are generally on bedding surfaces, commonly of Seisian sandstones. In many cases, the contact between the nappes resembles a stylolite seam rather than a thrust zone, and is not a measure of the extent of displacement.

Nappe thrusts in the Tatra Mts. (Text-figs 4.6, 6) probably took place underwater at full saturation of the rocks with seawater, which greatly influenced lithostatic, hydrostatic and pore pressures values (see SIBSON 1996, 2000) and was responsible for the decrease in weight and internal friction angle at the soles of the thrusts. Processes taking place along the thrust planes (e.g. cataclasis, mylonitization and pressure solution) lead to considerable mass reduction and crucially disturbed the geometry of the thrusted and folded units, which had already lost their classic duplex character (JUREWICZ 2003; JUREWICZ & SŁABY 2004; cf. RING & al. 2001). According to ŁUCZYŃSKI (2001b), mass reduction caused by pressure solution and chemical compaction of the condensed Middle Jurassic deposits of the High-Tatric
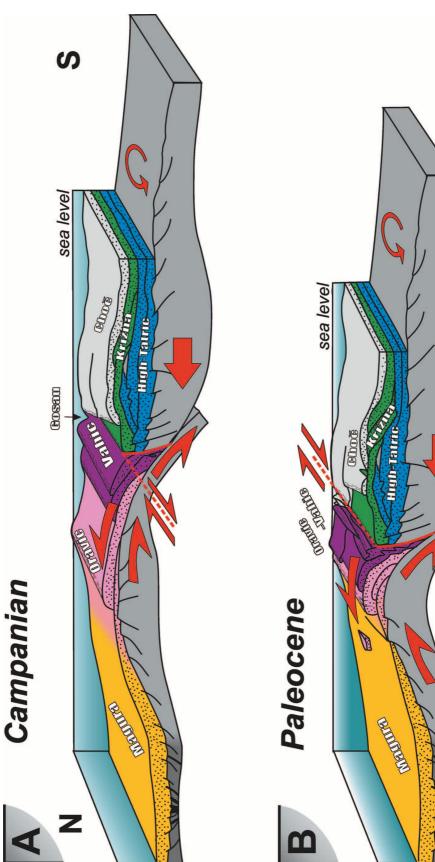
series could reach up to 50-70%. Along some tectonic surfaces, mass reduction could even be the main factor responsible for stress release ("ravenous" fault – JUREWICZ & SŁABY 2004).

In the process of tectonic transport, the key role was played by pulsate changes of pore pressure in the shear zones (JUREWICZ 2003). Increasing compression lead to an increase in pore pressure, which resulted in a reduction in effective stress, the formation of a shear surface (and, in the succeeding stages, its rejuvenation), and stress release by displacement. The newly opened shear zone thus became the pathway for migration of solutions transported according to the hydraulic gradient (see KNIPE 1989; SIBSON 1996), additionally supplied by released pore water. This zone acted as a channel carrying mechanically broken up and chemically dissolved rock material, which lead to a gradual but considerable mass reduction. With the destruction of rocks, solutions and suspensions appeared within the shear zone. Due to the rapid pressure fall caused by a displacement release of stress, gases were liberated and the angle of internal friction also decreased (JUREWICZ 2003). Pressures within the nappe thrusts evaluated by MILOVSKÝ & al. (2003) based on fluid inclusion investigations could vary between 0.2-5.4 kbar (20-540 MPa) and interpreted as a result of of hydrotectonic processes (see JAROSZEWSKI 1982; KOPF 1982, 2003). The presence of fluids in the sole of the nappe, and the consequent almost complete absence of friction, resulted in the displacement of the nappe over a "water pillow" like a hovercraft.

CONDITIONS OF DEFORMATION AND GEOMETRY OF THRUST-NAPPING IN THE PKB.

Folding in the PKB began later than in the Tatra Mts., i.e. in its southern part in the Coniacian (SCHEIBNER 1968; BIRKENMAJER 1970, 1974). It continued until the Late Palaeocene, i.e. until the end of Jarmuta Formation sedimentation in the Magura Basin (BIRKENMAJER & al. 1987; BIRKENMAJER & DUDZIAK 1991), gradually involving the succeeding marginal sedimentary zones to the north. The long duration of folding is connected with the slow-motion of subduction of the thinned continental crust from the base of the Vahic-Oravic Basin (Text-figs 7-8). In the last phase, due to gravitational sliding, the Klippen units were thrust onto the Magura foreland (Text-fig. 8B; see JUREWICZ 1994, 1997).

Fig. 8. A – Campanian – continuation of subduction under the Central Carpathian Block and thrust napping processes within the Vahic-Oravic zone. B – Palaeocene – thrusting and folding originated near the sea level. Due to favourable palaeomorphology, gravitational sliding and formation of olistolites took place (see JUREWICZ 1997). Shearing along the northern boundary of the Central Carpathian Block caused opening of pull-apart basins and sedimentation of Gosau-type conglomerates (cf. WAGREICH 1995). Not to scale

The character of the nappe thrusts in the Pieniny Mts. was completely different from that in the Tatra Mts. In the nappe units of the PKB more brittle deformations can be observed. The detachment of the nappe units in the PKB typically took place in rather susceptible shale-marly sediments (Krempachy Marl or Skrzypny Shale formations) underlying the limestone and limestone-radiolarian members of the Jurassic and Lower Cretaceous, and covered by marly-shale-sandstone deposits of the Upper Cretaceous. Shales and marls occurring at the bottom of the nappes are typically intensely tectonically disturbed and strongly slickensided. As a result of multiple tectonic activations, the older slickensides are crushed or folded (JUREWICZ 1994). The PKB nappes had a small thickness and mass, and, during transport underwent internal folding, strong faulting and brittle fracturing, which caused that a mass of calcite veins, and calcite coating numerous fault surfaces, prevail over the host rock. The conditions, under which such deformations are formed, have been described by KENNEDY & LOGAN (1998) from the southern Appalachians. Similarly, WOJTAL & MITRA (1986) and WOJTAL & PERSHING (1991) demonstrated that sheets were deformed primarily by sliding along discrete surfaces, resulting in numerous mesoscopic fault and fracture arrays within approximately 100 m of the thrust surface. These authors described the fault zone core as consisting of both hanging wall limestone and footwall shale cataclasites, and a neighbouring zone of fracture and fault arrays and folds in both the hanging wall and the footwall.

The thrusting of the nappe units was at least locally preceded by erosion, because in many cases (particularly in the eastern part of the PKB) the highest lithological units are absent (BIRKENMAJER 1979). In the more western part of the Polish section of the PKB, erosion preceding thrusting is evidenced by the synorogenic, partly fresh-water character of the Jarmuta Formation (BIRKENMAJER 1977, 1979) in the thrust foot, which encompasses numerous variably rounded grains both from the Klippen units and of exotic origin (KUTYBA 1986; BIRKENMAJER & WIESER 1990; BIRKENMAJER & SKUPIŃSKI 1990). KUTYBA (1986) stated that ca. 10-20% of the exotic rocks were composed of magmatic and metamorphic clasts. Particular attention is drawn to a block of basalt, several metres in diameter, from the vicinity of Biała Woda, and dated by K-Ar radiometry at ca. 140 Ma, indicating the Jurassic/Cretaceous boundary (BIRKENMAJER & WIESER 1990; BIRKENMAJER 1996), but according to recent investigations, Aptian-Albian (BIRKENMAJER & PÉCSKAY 2000b).

Differences in height favouring the formation of gravitational slumps are testified also by the olistolites that occur commonly in the Jarmuta Formation (e.g. in the vicinity of Szaflary – BIRKENMAJER 1979; NEMČOK & NEMČOK 1994; BAKOVÁ & SOTÁK 2000). Gravitational slumps have also been documented to the east of Szczawnica, where the so-called Homole block built of the plate-like Czorsztyn Unit and the thrusted and folded Niedzica Unit (BIRKENMAJER 1970), according to JUREWICZ (1994, 1997) slid to the Magura Basin, in which the sedimentation of the Jarmuta beds took place (Text-figs 2, 8B). The nappes bear traces of intense N-S stretching, manifested by the large number of normal faults (JUREWICZ 1994). There is also a distinct influence of morphology on the type of structures, e.g. the Czajakowa fold belonging to the Niedzica Nappe was formed due to the existence of a morphological ridge linked with the formation of a normal fault in the basement. Compartmental faults (see BROWN 1975) – partitions between domains of rocks in which a common magnitude of shortening was achieved in different ways – can also be observed (JUREWICZ 1994). The structural analysis of this part of the PKB throws light on the relation of its units to the northern Magura Basin with flysch sedimentation during the thrust folding. This does not mean, however, that the mechanism of gravitational slumping is responsible for the formation of the nappe structure of the PKB. Gravitational slumping of the nappes is plausible in the last phase of folding within the PKB, when the nappe front reached the northern Magura Basin and the differences in relative height enabled slumping.

Small exposures of the Klippen units within the Outer Carpathian flysch are known from Slovakia. In the opinion of POTFAJ (1997), the Klippen units have been dragged-out from the Klippen Belt and placed in their present position through sinistral strike-slip movement of the Carpathian block. It is obvious that during the folding stage the Czorsztyn Unit was detached from its basement, regardless of whether the basement was crystalline or sedimentary in character.

The specific character of the nappes in the PKB lies in the fact that they are rather thin, with thicknesses of 400-800 m, in comparison to 1100 m of the Czerwone Wierchy Unit in the Tatra Mts. (PASSENDORFER 1983) or several thousand metres in the Alps, and that lithologically they are generally bipartite. The nappes in the PKB comprise rather thin limestone complexes (e.g. in the Czorsztyn Unit limestones comprise 150-300 m as against 400-450 m of total thickness) which resulted in their low rigidity and ready break-up into blocks. The mosaic character of the geological map distinguishing the PKB from the surrounding structures is generally (besides the rheological conditions) a result of the small total thicknesses of the nappes and their lithological variability, favouring both folding and dismembering into

blocks. The Polish part of the PKB displays strong deformations, as a result of which it is difficult to find a single cubic metre of rock not cut by at least one fault. The bedding surfaces are also slickensided. Faults in the PKB units have a rather small extent (in the order of several metres) and small displacement (in the order of several or several tens of centimetres), what causes that despite their large number, in many places stratigraphic sections are only slightly disturbed.

POSITION OF THE SO-CALLED GRAJCAREK UNIT AND LARAMIDE RETRO-ARC THRUSTING

BIRKENMAJER (1986) assumed that after the stage of nappe-folding related with the late Subhercynian phase (Late Cretaceous-Early Palaeocene), and after the sedimentation of the Jarmuta Formation in the Early Palaeocene, retro-arc thrusting of the deposits from the peri-Klippen part of the Magura Basin onto the PKB took place. Deposits from the peri-Klippen part of the Magura Unit, incorporated into the PKB structure, are referred to the Grajcarek Unit (BIRKENMAJER 1970, 1979, 1986). This concept of BIRKENMAJER (1986) was a consequence of the assumption that, during the folding stage, the Czorsztyn Unit was autochthonous, as in the case of the High-Tatric Unit in the Tatra Mts. As a result, the boundary between the Pieniny Klippen Belt and the the Outer West Carpthians (OWC) lies at the stratigraphic boundary between the Jarmuta and Szczawnica formations within the Magura Unit. The idea of retro-arc thrusting of the Grajcarek Unit was contradicted by the results of deep borehole data in Hanušovce (East Slovakia), where units of the PKB were found thrust onto Magura flysch over a distance of some kilometres (LEŠKO & al. 1984). The present author (JUREWICZ 1994, 1997) also questioned the possibility of retro-arc thrusting because of the present topographic position of the Grajcarek Unit within the PKB (typically in deeply incised stream valleys, below the base of the Klippen units), the very short period of time available for structural transformation responsible for morphological inversion, and unfavourable morphologic conditions (deep Magura Basin in the north and eroded, gravitationally slumping Klippen units in the south). According to the present author, nappe-folding was progressive; it began in the south in the Coniacian (SCHEIBNER 1968) and persisted at least until the earliest Middle Palaeocene, i.e. until the beginning of sedimentation of the Szczawnica Formation (BIRKENMAJER & al. 1987). According to BIRKENMAJER & DUDZIAK (1991), the upper part of the Jarmuta Formation and the lower part of the Szczawnica Formation overlap in age (NP 7-9),

thus the nappe-folding processes could have continued until the Late Palaeocene. The two folding phases with different tectonic transport directions – late Subhercynian (progressive) and Laramide (retroarc) – are not marked here. There is also no evidence for distinguishing a separate Grajcarek Unit from deposits of the Magura Unit, because the northern margin of the former is of stratigraphic character (between the Jarmuta and Szczawnica formations) and is therefore difficult to recognise in the field and to accept as a boundary between two main structural units (PKB and OWC). The model presented by JUREWICZ (1997) distinctly simplifies the concept of the evolution and structure of the PKB, rejects one unit and one folding phase, and shifts the PKB boundary to the south, to the boundary recognisable in the field between the Klippen units and the Magura Unit. The Grajcarek Unit thus looses its autonomy and is incorporated within the Magura Unit, in which only the peri-Klippen deformation zone linked with nappe-folding in the PKB can be distinguished. Andesite dykes occur only to the north of the PKB, in the Magura Unit. Exposures, in which deposits of the Magura Unit appear within the PKB, should be interpreted as tectonic windows, whereas the Pieniny rocks within the Magura Unit should be treated as olistolites and olistostromes (Text-figs 2, 8B). The southernmost part of the Magura Basin has a longer sedimentary history, beginning in the Early Jurassic, and a facies character similar to that of the Klippen Basin (Branisko zone – BIRKENMAJER 1970, 1979, 1986). For this southernmost, peri-Klippen part of the Magura Basin, the term "Hulina Basin" *sensu* SIKORA (1971), recognised also by GOLONKA & al. (2000), and the term "Grajcarek Group" for the Lower and Upper Cretaceous lithostratigraphic units of formation rank in this area of the Magura Nappe, could be retained (BIRKENMAJER 1977; BIRKENMAJER & OSZCZYPKO 1989). GOLONKA & al. (2000) incorporated the southernmost part of the Magura basin (Hulina basin) into the PKB on the basis of sedimentological similarity. This view is controversial because, in a structural sense, the Magura and the PKB basins were two different areas during sedimentation, and they are also two different units at present (Text-figs 2-3). It is problematic whether the PKB is related only to the Central Carpathians. From the facies point of view, and in a sedimentological sense, the PKB is connected with the Central Carpathians; however, it is separated from them by a deep fault that became a shearing zone in the Albian (RAKÚS & MARSCHALKO 1997). Taking into account their structural evolution, the PKB are more related to the Outer Carpathians. The difference is the presence of the thinned continental crust at the base of the PKB instead of the oceanic floor of the Magura

Ocean. After MAHEĽ (1989), the PKB linked two areas of sedimentation and two systems of nappes (the Palaeoalpine Central Carpathian and Neoalpine Flysch Belt) into a uniform orogenic system. In the opinion of CSONTOS & VÖRÖS (2004), the Penninic-Vahic and Magura were two oceans separated by one or several minor continental fragments, like the Czorsztyn Ridge of the PKB, but where no such continental fragment were present, they formed one ocean. According to PLAŠIENKA (2003a, b), and in the geotectonic reconstruction of the present author, the PKB is an isolated basin with a different structure from that of the Outer and Central Carpathians (Text-fig. 3).

The PKB again underwent disintegration into blocks during nappe formation in the Outer Carpathians. The fragmentation of older nappes is responsible for the presence of characteristic klippes within the PKB and diapiric tectonics (BIRKENMAJER 1959). It had crucial structural consequences, because the nappe structure underwent destruction due to disturbance of the initial thrust planes, along which further tectonic transport was not possible, and the PKB behaved like a megabreccia zone (ANDRUSOV 1938).

MYJAVA BASIN PROBLEM

The Myjava succession is a strongly folded zone located along the southern border of the Pieniny Klippen Belt. This furrow-like zone, adjacent from the south to the Andrusov Ridge, is characterized by uninterrupted sedimentation during Coniacian-Oligocene times (SALAY 1987), i.e. during thrust- and nappe-folding within the PKB. In the Polish part of the PKB it is named the Maruszyna Slice, within which sedimentation continued from the Late Cretaceous to the Early Palaeogene (BIRKENMAJER & JEDNOROWSKA 1983; BIRKENMAJER 1986; DUDZIAK 1990; KOSTKA 1993). If the concept of RAKÚS & MARSCHALKO (1997), assuming the presence of local, laterally overlapping basins, located along a large strike-slip shear zone between the edge of the Central Carpathians and the PKB, and active from the Albian to Palaeogene, is to be accepted, then one such basin could be the pull-apart Myjava Basin (Text-figs 3, 8). A similar origin is suggested for sediments of the Gosau Group deposited within small basins, which originated along the northern margin of the Austroalpine units during the phase of strike-slip faulting (WAGREICH
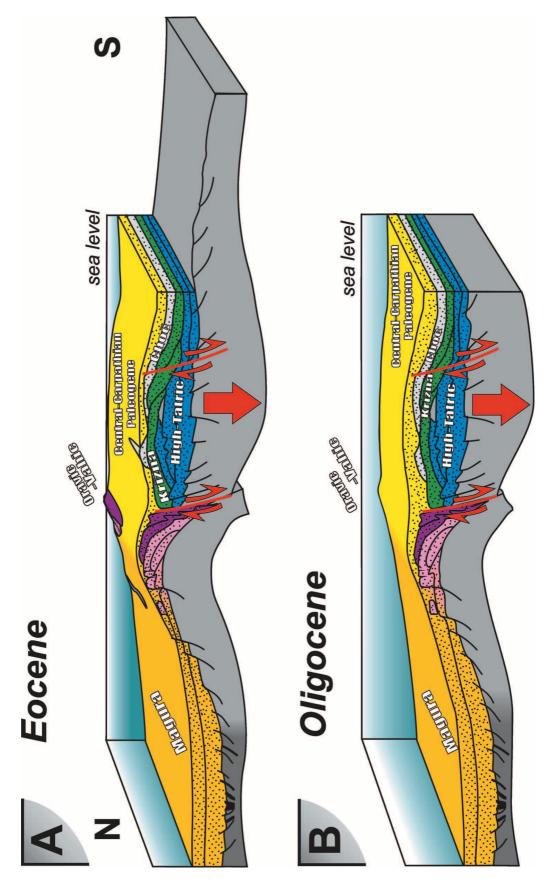
1995; LINZER & al. 2002). This similarity of the Myjava and Gosau basins was pointed out earlier by BIRKENMAJER (1979).

LATE PALAEOGENE TECTONIC STAGE

During the Late Cretaceous-Eocene, the future Carpathian orogen was part of the larger Alpine-Carpathian orogen formed by south-eastward subduction (135-55 Ma) of the Penninic Ocean and the ultimate collision (55-40 Ma) of the European and Adriatic continents (NEMČOK & al. 1998). SZAFIÁN & al. (1997) suggest that the Western Carpathians are at a different stage of evolution from that of the Southern Carpathians. The subducted oceanic slab under the Western Carpathians has already been assimilated with the asthenosphere, whereas a lithospheric slab is still present under the Southern Carpathians. This concept is compatible with the observation that the last major phase of crustal shortening terminated at the early Middle Miocene in the Western Carpathians, but continued throughout the Pliocene in the Southern Carpathians (SĂNDULESCU 1988).

As well as the Tatra Massif, the PKB and other structures developed in the course of Alpine nappe-folding underwent burial, which resulted in the formation of the Central Carpathian Palaeogene Basin (CCPB) (e.g. MARSCHALKO 1968). According to SOTÁK & al. (2001), the CCPB was formed as a marginal sea of the Peri-Tethyan basin and shows a fore-arc position developed in the proximal zone of the Outer Carpathian accretionary prism. It is not clear whether the Tatra Mts. were expressed in the morphology of the basin bottom or formed an island in the period preceding transgression and sedimentation. According to PASSENDORFER (1949, 1958) and PASSENDORFER & RONIEWICZ (1963), the Tatra Mts. could have been an island at least during the sedimentation of the Zakopane beds, whereas MARSCHALKO & RADOMSKI (1960) suggested that the axis of the CCPB passed through the Tatra Mts. WESTWALEWICZ-MOGILSKA (1986) postulated the presence of three independent sedimentary fans, in which the material came from the north and overpassed the PKB. WIECZOREK (1989) noted analogies of the Podhale Trough, belonging to the CCPB, with the Pyrenean Hecho Basin. He suggested that the clastic material was transported from the west to the east, and that its source

Fig. 9. Eocene-Oligocene – Central Carpathian Palaeogene sedimentation. In the beginning (A) – some parts of the bottom of the Podhale through and PKB (so-called "Poronin Island" – see OLSZEWSKA & WIECZOREK 1998) were the source of detrital material. During the Oligocene (B) continuation of plunging and unification of sedimentation took place. The southern boundary of the PKB and Tatra area is manifested as a synsedimentary fault (see KOHÚT & SHERLOCK 2003). Not to scale

was a canyon in the vicinity of Kubín (GROSS & *al.* 1984). MARSCHALKO (1968) pointed out the differences between the western and eastern part of the Central Carpathian Basin. JANOČKO & JACKO (1998) considered that the main factor governing deposition (Bartonian – Early Rupelian) in the Spišská Magura region was tectonics, which determined the shelf width, slope gradient and sediment input. According to BUČEK (2001), the different ages of the Borové Formation south of the Tatra Mts. are either a result of gradual flooding of the morphologically much dissected territory south of the Tatra Mts. or of distinct tectonic unrest. Larger foraminiferids of the Borové Formation (Bartonian) were repeatedly redeposited during sedimentation of the Šambron Member (BUČEK 2001). On the basis of palaeogeographic analysis, SOTÁK & *al.* (2001) postulated two systems of submarine fans with counter-directional palaeocurrent orientation. The large tectonic activity during sedimentation processes within the Liptov depression was pointed out by KOHÚT & SHERLOCK (2003), who determined the youngest ages of fault-related pseudotachylites on the base of laser microprobe ^{40}Ar/^{39}Ar analysis.

According to KOTAŃSKI (1963a), the transgression proceeded probably from the north (from the Magura Basin), leading to the formation of the so-called Tatra Eocene which, from the times of Uhlig and Limanowski has been referred to as the Nummulitic Eocene. In some exposures (e.g. the Pod Capkami Quarry), Eocene deposits display features of cliff breccia or (e.g. Hruby Regiel) deltaic deposits (PASSENDORFER 1958, 1983), which pass upwards into organodetrital carbonates with nummulites. Their age was estimated by foraminiferal analysis as Middle and Late Eocene (e.g. BIEDA 1959, 1963; ALEXANDROWICZ & GEROCH 1963; OLSZEWSKA & WIECZOREK 1998). The deposition of the Tatra Eocene (Borové Formation – GROSS & *al.* 1984) took place probably within an isolated carbonate platform (RONIEWICZ 1969). On the southern side of the Tatra Mts., in the Liptov Basin, carbonates of Choč type from the highest Mesozoic nappe of the Tatra Mts. play an important role in the Eocene conglomerates (JANOČKO & JACKO 1998; BUČEK 2001). According to GEDL (1998), the dinocyst and nummulitic assemblages indicate that oligotrophic conditions prevailed in the area of Nummulitic Eocene sedimentation. The flysch sequence occurring above the Nummulitic Eocene comprises (Polish and Slovak names – after GROSS & *al.* 1984): Szaflary beds – Borové Formation, Zakopane beds – Huty Formation, Chocho-łów beds – Zuberec Formation and Ostrysz beds – Biely Potok Formation, with a total thickness of ca. 4000 m (JANOČKO & JACKO 1998). Oligotrophic conditions prevailed during the sedimentation of the Zakopane beds.

There is, however, no evidence of a land area in the south, whereas the palynofacies of the Szaflary beds determined by GEDL (1998, 1999, 2000) point to a land area in the north. This was probably the PKB which, according to GEDL (2000), was the source area of clastic material during the sedimentation of the Szaflary and Zakopane beds.

In the vicinity of Sucha Woda the basal conglomerate in the Tatra Eocene contains tuffites, which were correlated with Carpathian tuffites and dated by GŁAZEK & *al.* (1998) at ca. 40 ±2 Ma (Priabonian). The Priabonian age of the pyroclastic series from the lowermost part of the Podhale Palaeogene was determined by KĘPIŃSKA & *al.* (2000) in the Biały Dunajec borehole. Six horizons with tuffites were situated at 2056-2089 m borehole depths, below typical flysch Lower Szaflary Beds and above the Middle Eocene carbonate conglomerates. Tuffites were deposited as a result of the activity of one magmatic centre located in the vicinity. The pyroclastic material found in the tuffites is similar to dacite or andesite, and corresponds to the tuffitic level of the so-called biotite sandstone connected with Early Priabonian stratovolcanic activity in the Little Hungarian Lowland (GŁAZEK & *al.* 1998; KĘPIŃSKA & *al.* 2000).

During the Eocene-Oligocene, the PKB collapsed and was buried under a blanket of flysch deposits (BIRKENMAJER 1986). It could represent an island arc or at least a system of submarine highs (Text-fig. 9). On the earlier napped and folded units of the PKB basal conglomerates and flysch deposits of transitional character appeared between the Magura Basins and the Central Carpathian Palaeogene (locally Late Palaeocene in ages – SAMUEL 1997).

EARLY NEOGENE TECTONIC EVENTS

The PKB Basin closed at the Cretaceous/Palaeocene boundary. During the Eocene and Early Miocene the Adria-ALCAPA microplates continued their northward movement, and their oblique collision with the North European plate (NEMČOK & *al.* 1998). During the Eocene the western part of the Alpine-Carpathian convergence underwent a change from subduction to collision. Towards the end of subduction, detachment of the subducting slab took place. The detachment propagated eastwards, accompanied by the break-off-related volcanism of Eocene-Eggenburgian age (NEMČOK & *al.* 1998). After SEGHEDI & *al.* (2004) subduction-related calc-alkaline magmatism (Text-fig. 1) was widespread in the Carpathian-Pannonian region between 11 and 0.2 Ma (see EMBEY-ISZTIN & *al.* 1993). The movement of microplates led to the development of the outer accrec-

tionary wedge, the formation of many flysch nappes and of a foredeep (NEMČOK & *al.* 1998; OSZCZYPKO 1998, 1999; GOLONKA & *al.* 2000; OSZCZYPKO & OSZCZYPKO-CLOWES 2003). The evolution of the foredeep basins of the Carpathian arc was marked by pronounced internal to external and lateral depocentre migration (MEULENKAMP & *al.* 1996). Orogenic shortening in the NE Outer Carpathians was about 260 km, at the rate of

~2.2 cm/year (BEHRMAN & *al.* 2000) and the speed of subduction was very high in comparison to the speed of subduction of the thinned continental crust of the Vahic-Oravic basin under the Central Carpathian Block. The last stage of nappe-thrusting in the Flysch Carpathians in the Polish part of the Western Carpathians took place in Badenian – Sarmatian times (OSZCZYPKO 1998; OSZCZYPKO & OSZCZYPKO-CLOWES 2003), whereas in
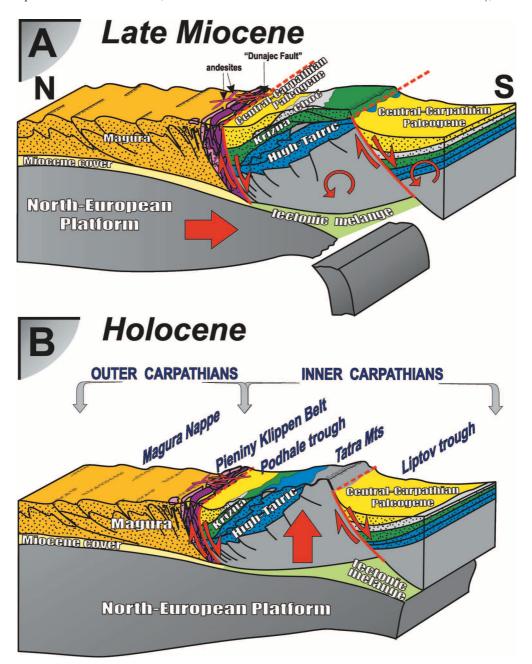

Fig. 10. A – Late Miocene collision of the Central Carpathian Block and North-European Platform after subduction of oceanic-type crust from the bottom of the Outer Carpathian flysch basin. B – present-day structural units of the northernmost Central Carpathians, near the boundary of the Outer Carpathians. Not to scale

the Eastern Carpathians the thrust movements persisted till the Pliocene (OSZCZYPKO & ŚLĄCZKA 1985; GOLONKA & *al.* 2000).

In the analysed area marine sedimentation (probably partly connected with the Central Carpathian Palaeogene of the Podhale Trough) still continued in a piggy-back basin in the vicinity of Nowy Targ in the Early and Middle Miocene (CIESZKOWSKI 1992). This was a result of modification of the southernmost part of the Magura Basin. Within the PKB area, these sediments are strongly folded. Lower Miocene deposits were also described from the Rača Subunit of the Magura Nappe (OSZCZYPKO & *al.* 1999; OSZCZYPKO-CLOWES 2001; OSZCZYPKO & OSZCZYPKO-CLOWES 2002). These facts indicate that within the Magura basin, partly folded in the Early Burdigalian, sedimentation processes continued during the Early Miocene.

Thrust movements were linked with the subduction of the Carpathian foreland plate i.e. the North European Platform (Text-fig. 10A). Based on the models of continental collision zones of ROYDEN (1993), KRZYWIEC & JOCHYM (1997) applied flexural modelling techniques to characterise the Miocene subduction zone of the Polish Carpathians. This implies that the Neogene subduction zone had retreating subduction boundaries. The main subduction-driving mechanism was postulated as a slab-pull mechanism, linked with the loading of oceanic crust or, less probably, with strongly thinned continental crust disappearing in the subduction zone. On the basis of gravity modelling in the SE Carpathians, SPERNER & *al.* (2004) presumed that the lack of ophiolites did not eliminate the presence of oceanic crust at the base of the Carpathians; the slab retreat produced a low-stress subduction zone, so that the entire oceanic lithosphere disappeared into the mantle.

The results of the magnetotelluric and gravity modelling of KRÓLIKOWSKI & PETECKI (2001) suggest that crustal structures in the western and eastern segments of the Polish Carpathians are different. The PKB in the Zakopane-Kraków section displays intensive resistivity variation in depth and horizontally (KLITYŃSKI & WÓJCICKI 2001). According to the magnetotelluric investigation of CZERWIŃSKI & STEFANIUK (2001) along the Zakopane-Kraków line, there is a high-resistivity block north of the PKB, which is probably overthrust onto a low-resistivity complex to the north. This is most probably the effect of the central overthrust within the North European Platform which originated during the last phase of its underthrusting, after subduction of oceanic crust from the base of the Outer Carpathian Basin. On the lithosphere thickness map of BIELIK & *al.* (2004), the thickening of the lithosphere in the central segment of the Carpathians arc, southward of the Tatra Mts., is very

clear. It is interpreted by the authors as the remnant of a subducted slab.

In the section of CZERWIŃSKI & STEFANIUK (2001), an elevation of high resistivity rock complexes can be observed south of the PKB (at a depth of about 3 km) and a depression of low resistivity rock formations further to the south (at a depth of about 6 km). This elevation may be correlated with the so-called "Poronin Island", which could occured during sedimentation of the Eocene conglomerate (OLSZEWSKA & WIECZOREK 1998). The nearby depression was connected with a pre- or synsedimentary extensional dip-slip fault in the Tatra Massif basement.

Geophysical sounding carried out during the POLONAISE CELEBRATION 2000 seismic experiment in relation to the Carpathian area is difficult to interpret because of the oblique course of the seismic profiles in relation to the structural orientation (GUTERCH & *al.* 2003). The CEL01 seismic profile located closest to the study area passes through the PKB in a place where the belt changes its course from parallel to NE-SW, which complicates the correlation of the deep structure with structures observed on the surface. In the vicinity of Chyżne, northwards of the PKB, the low-velocity zone (5.62 km/s) passes in a wedge down to a depth of ca. 10 km, whereas south of the klippen belt it is at a depth of ca. 3 km.

The geophysical investigations of LEFELD & JANKOWSKI (1985), based on the results of deep seismic soundings of UCHMAN (1973) and investigation of heat-flow data, Bouger reduced gravity anomalies and geomagnetic and geoelectrical anomalies, indicated the presence of two zones of deep fractures termed the "Peri-Carpathian fractures" zone *sensu* SIKORA (1976) and the "Pieninian fractures zone" (Pieninian lineament – NEY 1975). The Pieninian fracture zone crossed the PKB in the vicinity of Szczawnica. BOJDYS & LEMBERGER (1986), based on 3-D gravity modelling of the Earth's crust and upper mantle, consider that in the vicinity of Szczawnica the Peri-Pieniny fracture zone (cf. MAHEĽ 1980) meets the so-called "G-G" line (cf. Peri-Carpathian fracture zone of SIKORA 1976) which, according to these authors, is the northern margin of the unconsumed oceanic crust. According to this concept, westwards from the Szczawnica meridian, the entire consumption of oceanic crust from the base of the OWC took place, what resulted in continent/continent collision, the uplift of the Tatra Mts., and thickening of the crust in their vicinity. Between the Peri-Pieniny fracture zone and the "G-G" line the Moho lies at 54 km. A similar opinion was expressed by SZAFIÁN & *al.* (1997): the subducted oceanic slab under the Western Carpathians has already been assimilated into the asthenosphere,

whereas a crustal slab is still present under the Southern Carpathians.

With this stage of tectonic evolution in the PKB and Outer Carpathians are linked andesite dykes from the Szczawnica region (MAŁKOWSKI 1921; WOJCIECHOWSKI 1955; YOUSSEF 1978; BIRKENMAJER 1962, 1996; BIRKENMAJER & PÉCSKAY 1999, 2000a; BIRKENMAJER 2003). In relation to the orientation of the PKB, the dykes generally have a semi-parallel orientation, except one younger dyke of Mt. Wżar. BIRKENMAJER & al. (1987), BIRKENMAJER & PÉCSKAY (1999, 2000a) and BIRKENMAJER (2003) connect the occurrence of andesites with the so-called Pieniny Andesite Line (PAL), which runs WNW-ESE, obliquely traversing the northern boundary fault zone of the PKB and then extending north-westwards to the Odra Fault Line in Lower Silesia. Assuming the PKB structure after JUREWICZ (1997), that is without distinguishing the Grajcarek Unit of BIRKENMAJER (1979, 1986), the andesite dykes do not lie in the structure of the PKB but are entirely located northwards from it, within the Magura Nappe.

Both on the 1 : 75 000 topographic map of UHLIG (1905) published at the beginning of the last century in the Atlas of Galicia, as well as on the satellite image, the NW-SE (~150°) lineament is clearly visible, running along the Dunajec River valley (Text-fig. 2). A dextral strike-slip fault, called the Dunajec Fault, can be recognised in the field, displacing the northern margin of the PKB on the eastern side by ca. 700 m south-eastwards (BIRKENMAJER 1979). In the Liptov area, the prolongation of this fault forms the eastern boundary of the Ružbachy Slice and is linked with the the presence of travertine. On the northern side of the PKB, along the *en echelon* faults surrounding the Dunajec Fault, occur andesites forming a swarm of dykes and sills (Text-figs 2, 10A). Northward of the PKB this fault could be responsible for the disturbance in the course of the isobaths of the Magura Nappe overthrust and for the appearance of the Szczawa tectonic window, which is clearly visible on the map of OSZCZYPKO-CLOWES & OSZCZYPKO (2004). Eastward of the Dunajec Fault, the Magura Nappe is thicker (op. cit.*),* whereas to the west, an elevation of some older structures can be observed (Grybów Unit from under the Magura Nappe – op. cit.). A similar situation appears in the southern extension of the Dunajec Fault, where the Ružbachy Slice (Krížna) emerges from under the Central Carpathian Palaeogene (Text-fig. 2).

Both on the topographic map of UHLIG (1905), as well as on the satellite image, there are arrays of NW-SE lineaments (Text-fig. 2): Poprad, Dunajec, Rieka, Białka, Biały Dunajec (Cicha Woda). The Cicha Woda Fault was

earlier described by BAC-MOSZASZWILI (1998b) and the Białka Fault by MASTELLA (1975) and MAHEĽ (1986). This structural pattern (~155-165°) is clearly visible on the radar images in the entire Polish Outer Carpathians (DOKTÓR & al. 2002) and marked on the West European Platform as an array of faults sub-parallel to the Teisseyre-Tornquist Zone (e.g. ŚWIDROWSKA & HAKENBERG 2000). On Mt. Wżar, the swarm of dykes longitudinal to the PKB is cut by an array of faults parallel to the Dunajec Fault (one of these faults was occupied by the youngest andesite dyke). Dykes in Krościenko, Szczawnica and on Jarmuta Mt. are strongly carbonatized/calcitized (BIRKENMAJER 2003) and sheared parallel to the Dunajec Fault.

The course of the Dunajec Fault has some connection with the parallel polygenic Kraków-Lubliniec strike-slip zone separating the Upper Silesian Block from the Małopolska Block, which could be a part of a larger fault zone – the Szczecin-Kraków-Prešov Zone (ŻABA 1996; JAROSIŃSKI 1998). This zone continues under the Flysch Carpathians and can be associated on the surface with the Poprad River valley where, as in the Szczawnica region, mineral waters as well as seismic epicentres occur (WIEJACZ 1994). The contact zone of the two blocks is controlled by a strike-slip fault (ŻABA 1996; JAROSIŃSKI 1998). According to JAROSIŃSKI (1998), the stress field in the basement of the Polish part of the Outer Carpathians could be induced by the northward motion of the Carpatho-Pannonian plate and regional European plate stresses. Earlier e.g. POSPÍŠIL & al. (1989) noted horizontal displacement along the SE branch of the PKB (south of the Polish boundary) and connected it with the dextral Močarany-Trebišov-Peri-Panonnian fault system, which is one boundary of the "leading wedge" of the microplate, the second margin of which is the Muráň tectonic system. POSPÍŠIL & FILO (1980) pointed out axes of regional gravity anomaly and magnetic anomalies along the later distinguished Močarany-Trebišov-Peri-Panonian fault system.

The fact that the Dunajec Fault cuts across the structure of the PKB and is accompanied by andesite intrusions, the magmatic chamber of which could be located at a depth of 10-12 km (BIRKENMAJER 2003), indicates that it might be connected with a deep fault zone. Its orientation is similar to that of the older transfer faults, which results from the structural pattern of the North-European Platform: the prevailing directions of the main faults during basin origination were inherited and activated during later subduction processes and nappe formation.

In the area where the mineral waters in the Polish Outer Carpathians appear (i.e. in the vicinity of Szczawnica), LEŚNIAK & al. (1997) ascertained mantle-

derived helium. It accompanies CO_2 exhalations and CO_2-rich carbonate waters, which escape through the 40-km thick continental crust (cf. OSZCZYPKO & ZUBER 2002). According to LEŚNIAK & al. (1997), the strong mantle helium signal could be connected with dyke formation episodes as a result of subduction and post-collisional volcanism (see TOKARSKI 1978; BIRKENMAJER 2003). PIN & al. (2004) remarked that the Sr-Nd characteristics of the andesites from the Szczawnica area were not typical of andesites produced above actively subducting oceanic lithosphere. Taking into account the radiometric age of the andesites (12.5-10.8 Ma – BIRKENMAJER & PÉCSKAY 1999, 2000a) and the *en echelon* position of the dyke array in relation to dextral shearing, the andesite dykes in my opinion are connected with a deep, mantle-rooted fault zone rather than with subduction processes. This zone could be the Dunajec Fault related to the repeatedly activated Kraków-Lubliniec strike-slip zone. Connecting the formation of the andesite dyke swarm with the opening of fractures surrounding the dextral slip zone separating the Małopolska and Upper Silesian Blocks and primarily activated in the Late Carboniferous (ŻABA 1996), allows determination of the absolute age of the displacement during the Neogene.

The array of NW-SE lineaments: Poprad, Dunajec, Cicha Woda (Text-fig. 2) can also be explained by the presence of dextral transtension along oblique faults due to the propagation of the arc-thrust generated in a convergent geodynamic setting. This type of extension was described by DOGLIONI (1995), who gave examples from Sicily and Barbados. A situation, in which rigid indentors act as bulldozers during plate collision (similarly to the Małopolska Block) is described by KEEP (2000) based on analogue models of continental collision, and a geological example from the Precambrian part of the Tasman orogenic belt and the contemporary collision at the leading edge of Australia.

Absolute age determinations of the andesites occurring in the direct vicinity of the PKB according to BIRKENMAJER (e.g. 1979, 1985, 1986, 2003) document Miocene compression connected with the stage of continent-continent collision (between the European Platform and the Central Carpathian-Pannonian Microplate). The evolution of the PKB between Early and Middle Miocene times was influenced by a transpressional regime that was responsible for numerous sigmoidal structures (KOVÁČ & HÓK 1996). According to the palaeomagnetic data of BAŽENOV & al. (1981), a clockwise rotation of the Central Carpathians in relation to the Outer Carpathians of the order of 65° took place during this interval, which these authors considered should have caused a left-lateral (sinistral) displacement by at least several tens of kilometres along the northern

margin of the PKB (BIRKENMAJER 1985). On the other hand, evidence from palaeomagnetic investigations from the Podhale area indicates a counter-clockwise rotation of the Central Palaeogene Flysch by ca. 60° (MÁRTON & al. 1999). Palaeomagnetic study of the Pieniny andesites also indicated their counter-clockwise rotation after the emplacement of the second phase dykes (MÁRTON & al. 2004). More recent palaeomagnetic papers are generally rather unanimous with regard to the counter-clockwise rotation of the basement in the Miocene, but with differences in respect of the angle of rotation. KOVÁČ & MÁRTON (1998) suggested a 40-50° sinistral rotation of the ALCAPA Microplate in the Early Miocene (Ottangian); FODOR (1995) accepted three rotation phases for the West Carpathians.: 1) Oligocene-Eggenburgian – 35°; 2) Ottangian-Carpathian – 20°; 3) Badenian – 25° (80° in total); whereas MÁRTON & al. (1999), on the basis of investigations from the Podhale flysch area, suggested a rotation of ca. 60°. Greater discrepancies occur in the older tectonic units and older time intervals (cf. KRUCZYK & al. 1992; GRABOWSKI & NEMČOK 1999; GRABOWSKI & al. 1999). The crucial problem with palaeomagnetic investigations lies in the correlation of the palaeomagnetic data with structural analysis and with explanation of a space problem (see KOVÁČ & MÁRTON 1998).

MIOCENE UPLIFT OF THE TATRA MTS.

The uplift of the Tatra Massif and the appearance of the mountain range as a morphological feature above the surrounding Orawa-Nowy Targ depression are linked with the Late Miocene uplift of the Central Carpathian massifs (Text-figs 4.7-4.8, 10). Apatite fission-track data (BURCHART 1972) indicate that the uplift of the Tatra Mts. from a depth of 5 km (100°C) took place about 26-10 Ma ago. According to KOVÁČ & al. (1994), the uplift of the Tatric pre-Alpine complexes started from depths of 10-11 km (225°C) about 70-50 Ma ago and reached the 5 km depth level (100°) 30-15 Ma ago. Fission-track data in relation to the stage of the Tatra uplift from 2 km (60°C) indicate an interval between 7-2 Ma ago (BAUMGART-KOTARBA & KRÁĽ 2002). The most recent apatite fission-track analysis of the uplift (STRUZIK & al. 2003) indicate dates of ca. 11 Ma for the granitoids of the High-Tatra Mts., 20-12 Ma for the crystalline core of the Western Tatra Mts. and 7.6 (±1.2) Ma for the flysch sandstones from Podhale. The latter authors, similarly to BAUMGART-KOTARBA & KRÁĽ (2002), pointed out the uneven uplift and the higher uplift of the High-Tatra Mts. Structural analysis of granitoids from the High-Tatra Mts. also indicates a western plunge of the B axis ~265/15 (Text-fig. 6Bh), obtained from stress field reconstruction

based on striae on slickensided faults (Jurewicz 2000b, 2002), which might be related to the asymmetric uplift of the Tatra Mts.

The uplift of the Tatra Block was rotational in character and accompanied by the formation of a Sub-Tatric Fault in the south, recognized already by Uhlig (1899), as well as by the folding of the Podhale flysch in the form of a synclinorium (Gołąb 1959; Mastella 1975). The rotation axis was horizontal, W-E orientated, parallel to the present-day northern margin of the Tatra Mts. Estimates of the the rotation angle, comprising: the dip of the Nummulitic Eocene and the Mesozoic sedimentary cover, the erosion gradient of the Tatra granitoid core and the displacement along the Sub-Tatric Fault differ depending on the author concerned: 20º (Piotrowski 1978); 30-35º (Bac-Moszaszwili 1995); and 40º (Jurewicz 2000a, b). According to Jurewicz (2000a), the rotation angle of 40º (Text-fig. 10) was evaluated on the basis of reconstruction of the stress field obtained from analysis of the striae on flat-dipping slickensides in the granitoid core of the High-Tatra Mts. (Text-fig. 6Bh).

According to some authors, i.e. Kotański (1961), Birkenmajer (1986, 2003), Sperner (1996), and Sperner & al. (2002), the Sub-Tatric Fault is a reverse fault; however, the concept of a normal dip-slip character of this fault prevails (Maheľ 1986; Bac-Moszaszwili 1993; Hrušecký & al. 2002). The reverse character of this fault – as calculated by Sperner (1996) – would be responsible for a displacement of the order of 29 km and should be associated with metamorphism. Such feature of this fault was questioned e.g. by Petrík & al. (2003) and Kohút & Sherlock (2003). Sperner & al. (2002) presented new thrust models for the High-Tatra Mts.: kink and listric, in which slip along the Sub-Tatric Fault was of the order of 18.6 km. The large-scale detachment character of the Sub-Tatric Fault is clearly visible in the reflection seismic profile of Hrušecký & al. (2002). The dating of movement on the Sub-Tatric Fault is inseparably connected with the Tatra Mts. uplift documented by fission-track data (see above). Kohút & Sherlock (2003) linked the commencement of its activity (36-28 Ma) with pseudotachylites, and showed its syn-sedimentary character (during the sedimentation of the Central Carpathian Palaeogene flysch sequence).

Relaxation and normal faulting are linked with the Tatra Mts. uplift; the existing discontinuities also underwent reactivation (Bac-Moszaszwili 1998a; Birkenmajer 1999; Jurewicz 2002). Based on correlation with the PKB, their age was determined by Birkenmajer (1999) as Middle Miocene (Sarmatian); their most common orientation (~35/60N) and the orientation of the extension axis within the range 106-120º were determined from striae analysis and reconstruction of

palaeostress in the granitoid core of the High-Tatra Mts. (Jurewicz 2002). Observations of striae as well as the geometric effect of displacement show that the dip-slip faults were sinistral faults (Jurewicz & Bagiński 2005). Bac-Moszaszwili (1998a) related this sense of movement to the presence of W-E dextral shearing and block rotation within the nappe of the Western Tatra Mts. Similar processes were observed in the internal zones of the south-western Alps, where widespread extension forming a dense fault network overprints the compressional structures. The strike-slip phase is connected with this extension and is compatible with dextral shearing along the longitudinal faults (Sue & Tricart 2003).

The extension phase and dextral shearing in the granitoid core of High-Tatra Mts. was documented as mineralisation (quartz, epidote, barite and carbonates) within older, usually mylonitic zones, and as striation on the wall rocks of the shear zones (Jurewicz & Bagiński 2004). The age of this shearing can be concluded from the fact that these faults occur in the nappe zones and pass into Oligocene deposits (Gedl 1999) of the Podhale Trough, e.g. in the region of Chochołowska (Passendorfer 1974), Mała Łąka (Bac-Moszaszwili 1971), and Olczyska valleys (Sokołowski 1959). According to Mastella & al. (1988), the faults should be linked with the Mid-Miocene compression of the Styrian phase, whereas Birkenmajer (1999) specified their age as Late Miocene (Sarmatian), based on investigations in the Biała Woda Valley and by analogy to the PKB.

In relation to the orientation of the dislocation zones cutting the nappe units and flysch of the Podhale Trough, the overthrust of the Palaeogene deposits on the Sub-Tatric units shown on the Geological Map of the Tatra Mts. (Bac-Moszaszwili & al. 1979) along the Mała Łąka Valley, should be revised. Geometrically it is difficult to explain this cartographic image as a thrust of younger deposits (Nummulitic Eocene) on older ones (Mesozoic nappes). It is, however, possible to assume the presence of a normal fault along the Małołącki Stream, moving down the hanging-wall with the Hruby Regiel Mt. Palaeogene conglomerates (see Sokołowski 1959; Mastella & Mizerski 1977; Bac-Moszaszwili 1998a). This type of dislocation is testified by loose tectonic breccia occured in the stream, mineralised by several-centimetres-thick milk calcite. A local depression in this area was documented by Głazek (2000) who connected red conglomerates lying on Jurassic limestone of the Choč Nappe described by Chowaniec & al. (1975) in the Hruby Regiel borehole with pre-Eocene karstification processes.

In the PKB, the Miocene tectonic event was connected by Birkenmajer (1986, 2003) with two phases, dated from the andesite intrusions as Sarmatian (Serravallian; Birkenmajer 1996; Birkenmajer & Pécskay 1999, 2000a). According to Birkenmajer (1979, 1986), the southern thrust of the PKB klippes is

related to this stage. This might be the effect of crushing the PKB between two lithological complexes characterised by different rheological properties: to the north, a ductile complex of Magura flysch; to the south, a ductile thin complex of the Podhale flysch lying on rigid crystalline rocks of the Tatra Massif covered with folded and napped sedimentary rocks. During compression accompanying the rotational uplift and formation of the Podhale Trough, this rigid Tatra block caused the "undercutting" of the deeper parts of the PKB, which is manifested closer to the surface by the presence of a Peri-Klippen flexure zone (MASTELLA 1975) and the south-vergence of the structure of the PKB.

"VIOLET VEINS" – PSEUDOTACHYLITES OR HYDROTHERMAL MINERALISATION?

The uplift of the Tatra Massif is associated with seismic activity which brought about the formation of pseudotachylites (PETRÍK & REICHWALDER 1996; GAWĘDA & PIWKOWSKI 2000; PETRÍK & al. 2003; KOHÚT & SHERLOCK 2002, 2003). Pseudotachylites from the Gerlach Mt. have been dated by direct $^{40}Ar/^{39}Ar$ laser-probe analyses, and documented as Oligocene (36-28 Ma) seismic events (KOHÚT & SHERLOCK 2002, 2003). The seismic activity in this interval is further confirmed by tuffite horizons appearing several times during sedimentation in the Podhale Trough, from the Upper Eocene conglomerates (GŁAZEK & al. 1998; KĘPIŃSKA & al. 2000), through the Zakopane beds (VAN COUVERING & al. 1981; WIESER 1973) and the Chochołów beds (HALICKI 1961). GAWĘDA & PIWKOWSKI (2000), who investigated pseudotachylites from the Kończysta and Pasternakowe Czuby also postulated that the so-called "violet veins" from the Mięguszowiecki Cirque are of similar origin. On earlier maps (JACZYNOWSKA 1980) these veins were named "titanite veins". My observations do not confirm the pseudotachylite origin of the "violet veins":

1. It is generally accepted that pseudotachylites developed as a result of Palaeogene earthquakes (KOHÚT & SHERLOCK 2002, 2003). In this case, stress release should have taken place along earlier discontinuity zones, in which friction was lower than in the surrounding rock and the chance of pseudotachylite formation was smaller. Pseudotachylites commonly originate due to frictional melting (HIROSE & SHIMAMOTO 2003; PETRÍK & al. 2003; KOHÚT & SHERLOCK 2003). LIN & al. (2003) distinguish two types of pseudotachylite veins in the shear zone: cataclasite-related and mylonite-related pseudotachylites. Most of the "violet veins" of the Tatra Mts. occur in the axial parts of 0.5-1.5 m wide, repeatedly reactivated

shear zones (JUREWICZ & BAGIŃSKI 2005), where the loose material of cataclastic zones dispersed rather than accumulated the destruction energy. A large rheological contrast between the shear zone and the surrounding rock should lead to displacement along the wall-rocks. Besides, the mylonitic zones were texturally predisposed to slip, and in loose cataclasite the main role in stress release was played by intergranular slips.

2. There is direct evidence of the influence of high temperature on the surrounding rock (biotite chloritisation, plagioclase sericitisation), as pointed out by GAWĘDA & PIWKOWSKI (2000), which might indicate a long period of high temperature influence (e.g. hydrothermal solutions) rather than melting of the rock through frictional heating.

3. The grains were also subject to high temperatures – their rounding can be observed with the naked eye. This was also shown by GAWĘDA & PIWKOWSKI (2000), who determined a correlation coefficient of R=0.55. This effect can be obtained as a result of the flow of hydrothermal solutions through mylonitic gouge: part of the grains undergoes complete dissolution or melting, part only at the margins. These fine grains derived from mylonites could have been transported as a suspension and could have penetrated the fractures beyond the mylonitic zones.

4. There is a lack of convincing glass: according to GAWĘDA & PIWKOWSKI (2000) the cement is ultramylonitic in character. Hydrothermal solution of the gouge and cataclasite grains caused the chemical composition of the hydrothermal solutions to approach that of the surrounding rocks, and fast cooling lead to the formation of "glass" with incompletely dissolved grains.

5. There is no clear explanation why the pseudotachylites, which should develop from melting of granite, should be enriched in ferruginous oxides (GAWĘDA & PIWKOWSKI 2000) and, in samples from the Mięguszowiecki Cirque, also in titanite. Microprobe analysis showed the presence of: quartz, K-feldspar, plagioclase, titanite, hematite, apatite, zircon, and uraninite. No micas were observed, which might be attributable to their complete decomposition (KOISAR & ZAWIDZKI 1972).

6. BSE microprobe images do not point to the detrital origin of hematite (as suggested by GAWĘDA & PIWKOWSKI 2000); they are rather euhedral microcrystals, typically not exceeding 5 μm. in size. KOISAR & ZAWIDZKI (1972) considered that the earlier mylonite, because of geochemical transformations caused by the circulation of oxidising solutions, was enriched in Fe^{3+} after oxidising Fe^{2+} from the decomposition of chlo-

rites, and impoverished in Mn and Ti in relation to the green mylonite surrounding the violet veins.

7. The violet veins contain large quantities of titanite, which, if Ti is substituted by Fe^{3+} or Al^{3+}, attains a dark brown colour; this variety is called grothite (see DEER & al. 1962). Titanite from the Mięguszowiecki Cirque contains ca. 1.5% by weight Fe and over 2% by weight Al, and its presence might influence the colour of the veins to a similar degree as does the hematite. Possibly this unique combination of cherry-coloured hematite and brown-coloured titanite in the variety with aluminium and iron is responsible for the unusual "violet" colour of the veins.

Summing up, the "violet veins" from the Mięguszowiecki Cirque could represent hydrothermal mineralisation cementing mineral grains of different sizes, adopting, as migration channels, the existing tectonic discontinuities, mainly mylonitic and cataclastic zones as well as the accompanying fractures. The violet colour of the veins is a result of the different proportions of the titanite variety grothite and hematite. During mineralisation the components of the mylonitic matrix formed a suspension in hydrothermal solutions and migrated in suspension beyond the shear zones to the surrounding fractures. Part of the grains could have undergone dissolution; however, these processes could have resulted from the temperature of the hydrothermal solutions, and not from frictional heating.

NEOTECTONIC ACTIVITY OF THE TATRA MTS., PKB AND RELATED AREAS

A crucial role in the geomorphological evolution of these areas during the Pleistocene-Holocene was played by glaciations, after the retreat of which isostatic movements was able to take place.

The neotectonic activity in the Tatra Mts., PKB and related areas has been documented by different methods. OSZAST (1973), BIRKENMAJER (1976b) and others conducted sedimentary, stratigraphic and tectonic research of the Neogene sediments. GRODZICKI (1979) investigated the nature of neotectonic movements on the basis of corrosional etching horizons in caves. BAUMGART-KOTARBA (1981, 2001) carried out morphometric analyses of river terraces in Podhale and Orawa. RĄCZKOWSKI & al. (1984) and ZUCHIEWICZ (e.g. 1995, 1998) analysed the variability of vertical movements on the basis of indirect evidence provided by geomorphological mapping, the construction of various morphometric maps based on mathematical transformation of the present-day topography, and the use of some mathematical techniques to model theoretical longitudinal

river profiles and valley network and to statistically process the drainage pattern parameters. HRADILEC & al. (1981) presented methods for determining recent movement and evaluated ca. 8 mm per year for the Tatra Mts. TOKARSKI & ŚWIERCZEWSKA (2003) analysed the fractured clasts in the Upper Badenian to Holocene gravels from the paraconglomerates within the Magura Nappe and concluded that the thrusts in the Outer Carpathians had been active in Quaternary times. BAUMGART-KOTARBA & KRÁĽ (2002) conducted fission-track analysis for determining the date of uplift of the Tatra Massif from a depth of ~2 km (60°C). CZARNECKA (1986, 2004) and MAKOWSKA & JAROSZEWSKI (1987) determined the neotectonic activity of the Tatra Mts. and PKB using precise levelling data. These data point not only to the presence of vertical movements, but also to their variability, i.e. uplift movements are observed in the southern part of the Podhale synclinorium and inwards in the Tatra Mts., which might suggest a continuing rotational uplift of the Tatra Massif (MAKOWSKA & JAROSZEWSKI 1987).

Positive vertical movements are noted also in the PKB, albeit to a variable extent (CZARNECKA 1986). Subsidence of the order of ~7 mm/10 years was observed in the region between Czorsztyn and Niedzica (CZARNECKA 2004). Analysis of horizontal movements by CZARNECKA (2004) between Wżar Mt. and Czorsztyn, as well as between Niedzica and Kacwin, indicated compression along the northern margin of the PKB at the contact with the Magura Nappe, and extension along the southern margin at the contact with the Podhale flysch. This is confirmed by observation of the seismic wave velocities, which fall from 3700 m/s to 2500 m/s at the southern margin of the PKB and rise again along the northern margin, possibly indicating an increase in stress (CZARNECKA 1986, 2004).

Possible evidence of neotectonic activity of the study area is provided by the earthquake (4.7 on the Richter scale) that took place at the end of December 2004. Its epicentre was located near the village of Czarny Dunajec and could perhaps be linked to movement on a NNW-SSE strike-slip fault marked on the map and termed the Czarny Dunajec Fault (Text-fig. 2).

CONCLUSIONS

1. As a result of Triassic basin expansion and Jurassic break-up of the southern margin of the North European plate and rifting processes, the development of the Oravic-Vahic Basin with thinned continental crust without rift-related volcanism took place. The oblique rift was probably orientated

WSW-ENE. Non-volcanic rifting resulted in the lack of oceanic crust in the sedimentary zone of the analysed part of the PKB Basin. Due to extensional tectonics, lithospheric stretching and crustal thinning took place as a result of normal listric faulting. These processes took place also within the previously continental crust of the Czorsztyn Ridge.

2. The ramp of listric faults that originated due to extension connected with passive rifting processes could have been the margin of sedimentary sub-basins within the Vahic-Oravic Basin (like the Niedzica or Czertezik basins) and a source of detrital material (also from the pre-Mesozoic basement). Listric normal faulting requires rotation from a horizontal to steeper dip of the hanging wall and propagation of fault-related synclines. These faults were responsible for the later rheological heterogeneity and, during the formation of nappes, could initiate the formation of thrusts and duplex ramps.

3. In the tectonic evolution of the PKB and the Tatra Mts. there is no need to assume the presence of the Andrusov Ridge of BIRKENMAJER (1986, 1988) between the Pieniny and Manín basins or in any other locality. The source of exotic material could have been the ramp of listric faults as well as the repeatedly tectonically reactivated northern margin of the Central Carpathians Block. Exotic blocks pointing to a rifting source-area could lie southwards of the Tatra Mts. and be related to the Meliata subduction zone.

4. Nappe-folding proceeded from the south, gradually engaging the more northward sedimentary zones. One of the last stages of Alpine folding and nappe formation due to basement shortening within the Central Carpathians was the underthrusting of the crystalline massif of the Tatra Mts. together with its sedimentary cover, under the previously arisen Krížna and Choč nappes, forming the High-Tatric nappes and folding of the autochthonous cover.

5. The Late Cretaceous (post-Turonian) thrust-folding of the Choč, Krížna and High-Tatric nappes took place underwater and at considerable overburden pressure (~6-7 km) and with a geothermal gradient of (~30°C/km). This lead to the formation of imbricate structures or a hinterland dipping duplex, the geometry of which was already strongly disturbed during its development by pressure solution processes leading to considerable mass loss. Due to the great thickness of the overthrust nappes and the submarine environment, there were no favourable conditions for subaerial weathering and karstification, or for gravitational sliding.

6. A very important role during the tectonic displacement of the Choč, Krížna and High-Tatric nappes was played by the cyclic changes of pore pressure and hydrotectonic processes, leading to the formation of a "water pillow" within the shear zone at the base of the nappes, decreasing friction and facilitating movement. These changes were associated with mechanical fragmentation and disintegration, pressure solution and mass loss. Mass loss was one of the mechanisms of stress relaxation ("ravenous faults" – JUREWICZ & SŁABY 2004).

7. Nappe-folding in the PKB was connected with the subduction of thinned continental crust of the Vahicum-Oravicum under the northern margin of the Central Carpathian Block, rather than under the Czorsztyn Ridge (cf. BIRKENMAJER 1986). It proceeded from the south to the north, involving the marginal parts of the sedimentary basin and lasted from the Coniacian to the Middle Palaeocene (cf. SCHEIBNER 1968; BIRKENMAJER 1986; JUREWICZ 1997).

8. Nappe thrust-folding within the PKB originated at or above sea level; these processes were accompanied by erosion and the formation of the synorogenic, locally shallow-marine or fresh-water Jarmuta Formation, which can be correlated with the Gosau Group sediments (cf. WAGREICH 1995; PLAŠIENKA & al. 1997).

9. In the Tatra Mts. and the PKB, nappe thrust-folding was greatly influenced by the strike-slip shear zone between the edge of the Central Carpathians and the PKB (see RAKÚS & MARSHALKO 1997). Effects of the counter-clockwise rotation of the Tatra block during thrusting processes were the changing directions (from NW to N) of thrusting within the granitoid core and of the nappe-thrusting in the sedimentary cover (~45°). This shear zone may also be responsible for the origin of small pull-apart basins like the Myjava basin (Maruszyna Slice) which could have been formed through the same mechanism as the Gosau basins (WAGREICH 1995).

10. The character of deformation in the PKB was influenced by thin-skinned overthrusting (small thickness of the nappe units and their internal lithological and rheological variability). More brittle deformation conditions than within the High-Tatric and Krížna nappes lead to the formation of numerous slickensides. During later stages of nappe-folding, dismembering into blocks, diapiric tectonics and *mélange* took place.

11. In the terminal phase of nappe thrust-folding (until the Late Palaeocene), because of the favourable palaeomorphology, the northernmost units of the PKB (Czorsztyn and Niedzica units) were transported

through gravitational sliding, dismembered into slices with compartmental faults and thrust onto the Magura foreland, forming numerous olistolites and olistrostomes. The largest is observed in the Małe Pieniny (Homole-Biała Woda block – Jurewicz 1997).

12. As a result of the Miocene subduction of the North-European continental crust beneath the northern margin of the Central Carpathian massifs, the PKB suture zone was unrooted at a depth of about 6 km and the granitoid massif of Tatra Mts. was unrooted at a depth of about 10 km. Both were overthrusted onto the sedimentary rocks of the North-European Platform (Lefeld & Jankowski 1987; Bielik & al. 2005). The shearing was possibly due to the appearance of more brittle behaviour resulting from crust cooling and a decrease in the geothermal gradient (which reached 30°C/km during Late Cretaceous nappe-folding – Jurewicz & Bagiński 2005), due to subduction of cold continental crust.

13. A consequence of the oblique subduction of the North-European continental crust beneath the northern margin of the Central Carpathians was the reactivation of old deep fault zones. With one of these – the Dunajec Fault – were connected *en echelon* shears trading on the andesite dyke swarm, travertines and mineral waters.

14. Miocene continent/continent collision of the North-European Platform and Central Carpathian Block caused the disintegration of the marginal part of the so-called Central Carpathian block into individual massifs. The asymmetric character of these massifs (their northern dip) may result from the presence of horizontal shearing processes linked with the underthrusting of the North-European Platform, and consequently with the rotational uplift. The extent of rotation of the Tatra Massif, bounded to the south by the Sub-Tatric Fault and to the north by the Peri-Klippen flexure zone, is estimated at ~40°.

15. The Neogene uplift of the Tatra Massif was inhomogeneous (higher eastwards) and indicates a western plunge of the B axis ~265/15 (Jurewicz 2000a, b).

16. The apparent northern dip of the PKB and its southern retro-arc thrust is a result of compression between rock massifs with different rheological properties: the ductile flysch of the Magura nappe in the north and the ductile flysch of the Podhale Trough, with the rigid block of the Tatra Mts. at the base in the south.

17. Rotational uplift of the Tatra Massif still continues. This process and its influence on the geometry of the PKB (e.g. the south-vergence of the PKB) is confirmed by precise levelling data of Czarnecka (1986, 2004), and Makowska & Jaroszewski (1987).

Acknowledgements

REFERENCES

Acocella, V., Kormeb, T. & Salvinia, F. 2003 Formation of normal faults along the axial zone of the Ethiopian Rift. *Journal of structural Geology*, **25**, 503-513.

Alexandrowicz, S. & Geroch, S., 1963. Assotiation de petit Foraminifers dans l'Eocene de la Tatra. *Annales Societatis Geologorum Poloniae*, **33**, 219-228.

Andrusov, D. 1938. Étude géologique de la zone des Klippes internes des Carpathes Occidentales, 3e partie : Tectonique, 135 pp. *State Geological Survey*; Prague.

— 1965. Aperçu générale sur la géologie des Carpathes occidentales. *Bulletin de la Société Géologique de la France*, **7**, 1029-1062.

— 1968. Grundriss der Tectonik der nördlichen Karpaten, 188 pp. *Vydavateľstvo Slovenskej Académie Vied*; Bratislava.

Andrusov, D., Bystrický, J. & Fusán, O. 1973. Outline of the structure of the West Carpathian Mts. Guide book for geological excursion, X Congress of Carpatho-Balkan Geological Association, 44p. *GÚDŠ*; Bratislava.

Arthaud F. & Matte P. 1977. Late Palaeozoic strike-slip faulting in southern Europe and northren Africa: Result of a right-lateral shear zone between the Appalachians and the Urals. *Geological Society of America Bulletin*, **88**, 1305-1320.

Aubrecht, R., Mišik, M. & Sýkora, M. 1997. Jurassic synrift sedimentation on the Czorsztyn Swell of the Pieniny Klippen Belt in Western Slovakia. Alpine evolution of the Western Carpathians and related areas. Bratislava, Slovakia, September 11-14th 1997, pp. 53-64.

Aubrecht, R. & Túnyi, I. 2001. Original orientation of neptunian dykes in the Pieniny Klippen Belt (Western Carpathians): the first results. *Contribution to Geophysical and Geodesy,* **31**, 557-578.

Bac-Moszaszwili, M 1971. Tectonics of the Bobrowiec unit in the Western Tatra Mts. *Acta Geologica Polonica*, **21**, 279-317. [*In Polish with English summary*]

— 1993. Structure on the western termination of the Tatra massif. *Annales Societatis Geologorum Poloniae*, **63**, 167-193. [*In Polish with English summary*]

— 1995. Diversity of Neogene and Quaternary tectonic movements in Tatra Mountains. *Folia Quaternaria*, **66**, 131-144

— 1998a. Geology of the Subtatric units, Western Tatra Mts, Poland. *Studia Geologica Polonica*, **111**, 113-136.

— 1998b. Neogeńskie ruchy tektoniczne w Tatrach i na Podhalu. *In*: W. ZUCHIEWICZ (*Ed.*), III Ogólnopolska Konferencja Komisji Neotektoniki Komitetu Badań Czwartorzędu PAN, pp. 7-10. Kraków.

BAC-MOSZASZWILI, M., GAMKERLIDZE, I. P., JAROSZEWSKI, W., SCHROEDER, E., STOJANOV, S. & TZANKOV, T.V. 1981. Thrust zone of the Krížna Nappe at Stoły in Tatra Mts (Poland). *Studia Geologica Polonica,* **68**, 61-73.

BAC-MOSZASZWILI, M., BURCHART, J., GŁAZEK, J., IWANOW, A., JAROSZEWSKI, W., KOTAŃSKI, Z., LEFELD, J., MASTELLA, L., OZIMKOWSKI, W., RONIEWICZ, P., SKUPIŃSKI, & WESTFALEWICZ-MOGILSKA E. 1979: Geological map of the Polish Tatra Mts, 1:30000 scale. *Instytut Geologiczny*; Warszawa.

BAC-MOSZASZWILI, M., JAROSZEWSKI, W. & PASSENDORFER, E. 1984. On the tectonics of Czerwone Wierchy and Giewont area in the Tatra Mts., Poland. *Annales Societatis Geologorum Poloniae*, **52**, 67-88. [*In Polish with English summary*]

BAKOVA, L. & SOTÁK, J. 2000. Tectonosedimentary formation of the Pieniny Klippen Belt (Orava, Slovakia). *Slovak Geological Magazine*, **6**, 302-303.

BAŽENOV, M., BEGAN, A., BIRKENMAJER, K. & BURTMAN, V. 1981. Palaeomagnetic evidence of tectonic origin of the curvature of the West Carpathian arc. *Bulletin de l'Académie Polonaise de Sciences, Série des sciences de la Terre*, **28**, 281-290.

BAUMGART-KOTARBA, M. 1981. Tectonic movement on the eastern Podhale in the light of an analysis of Quaternary terraces of the Białka Tatrzańska valley and the lineaments from satellite image. *Przegląd Geograficzny*, **53**, 725-736. [*In Polish with English summary*]

— 2001. Continous tectonic evolution of the Orava Basin (Northern Carpathians) from Late Badenian to the present-day. *Geologica Carpathica*, **52**, 103- 110.

BAUMGART-KOTARBA, M. & KRÁĽ, J. 2002. Young tectonic uplift of the Tatra Mts (fission track data and geomorphological arguments). *Geologica Carpathica*, **53**, Special Issue – CD with extended abstracts.

BEHRMANN, J.H., STIASNY, S., MILICKA, J. & PERESZLENYI, M. 2000. Quantitative reconstruction of orogenic convergence in the northeast Carpathians. *Tectonophysics,* **319**, 111–127.

BEZÁK, V. 1991. Present knowledge of tectogenesis of Veporicum (West Carpathians). *Mineralia Slovca*, **23**, 381-383.

BIEDA, F. 1959. Nummulite fauna in the Tatra Eocene. *Biuletyn Instytut Geologicznego*, **141**, 5-13.

— 1963. Larger foraminifers of the Tatra Eocene. *Prace Instytutu Geologicznego*, **37**, 1-215.

BIELIK, M., ŠEFARA, J., KOVÁČ, M., BEZÁK, V. & PLAŠIENKA, D. 2004. The Western Carpathians–interaction of Hercynian and Alpine processes. *Tectonophysics,* **393**, 63-86.

BIRKENMAJER, K. 1959a. Diapiric tectonics in the Pieniny Klippen Belt (Carpathians) – *Bulletin de l'Acaddémie Polonaise des Sciences, Série des Sciences Chimiques, Géologiques et Géographiques*, **7**, 123-128.

— 1959b. A new Klippen Series in the Pieniny Mts., Carpathians – the Czertezik Series. *Acta Geologica Polonica*, **9**, 499-517.

— 1959c. Significance of the Haligovce Klippe for the geology of the Pieniny Klippen Belt, Carpathians. *Annales de la Société Géologique de Pologne*, **29**, 73-88.

— 1962. Remarks on geological form of the Mt Wżar andesites, Pieniny Mts, Carpathians. *Acta Geologica Polonica*, **12**, 201-213.

— 1963. Stratigraphy and palaeogeography of the Czorsztyn series (Pieniny Klippen Belt, Carpathians) Poland. *Studia Geologica Polonica*, **9**, 380 pp. [*In Polish, with English summary*]

— 1970. Pre-Eocene fold structures in the Pieniny Klippen Belt (Carpathians) of Poland. *Studia Geologica Polonica*, **31**, 1-77. [*In Polish with English summary*]

— 1974. Carpathian Mountains. *In*: A.M. SPENCER (*Ed.*). Mesozoic-Cenozoic orogenic belts – data for orogenic studies. *Geological Society Special Paper,* **4**, 127-157.

— 1976a. The Carpathian orogen and plate tectonics. *Publications of the Institute of Geophysics Polish Academy of Sciences*, A-2, **101**, 43-53.

— 1976b. Pleistocene tectonic deformations at Szaflary, West Carpathians, Poland. *Annales de la Société Géologique de Pologne*, **66**, 309-323.

— 1977. Jurassic and Cretaceous lithostratigraphic units of the Pieniny Klippen Belt, Carpathians, Poland. *Studia Geologica Polonica*, **45**, 1-158.

— 1979. Przewodnik Geologiczny po Pienińskim Pasie Skałkowym. 236 pp. *Wydawnictwa Geologiczne*; Warszawa.

— 1985. Major strike-slip faults of the Pieniny Klippen Belt and the Tertiary rotation of the Carpathians. *Publications of the Institute of Geophysics Polish Academy of Sciences*, A-16, **175**, 101-115.

— 1986. Stages of structural evolution of the Pieniny Klippen Belt, Carpathians. *Studia Geologica Polonica*, **88**, 7-32

— 1988. Exotic Andrusov ridge: its role in the plate-tectonic evolution of the West Carpathians foldbelt. *Studia Geologica Polonica*, **91**: 7-37.

— 1996. Mioceńskie intruzje andezytowe rejonu Pienin: ich formy geologiczne i rozmieszczenie w świetle badań geologicznych i magnetycznych. *Geologia, Kwartalnik AGH*, **22**, 15-25.

— 1999. Late Tertiary fault system of the Biała Woda Valley, Tatra Mountains, Carpathians. *Bulletin of the Polish Academy of Sciences, Earth Sciences*, **47**, 239-245.

— 2000. Gosau-type conglomerate in the Rusinowa Polana area, Polish Tatra Mts: its relation to the Lower Subtatric Nappe. *Bulletin of the Polish Academy of Sciences, Earth Sciences,* **48**, 117-133.

— 2003. Post-collisional Late Middle Miocene (Sarmatian) Pieniny Volcanic Arc, Western Carpathian. *Bulletin of the Polish Academy of Sciences, Earth Sciences,* **51**, 79-89.

BIRKENMAJER, K. & DUDZIAK, J. 1991. Middle to Late Palaeocene Nannoplankton Zones in the Jarmuta Formation, Pieniny Klippen Belt, Carpathians. *Bulletin of the Polish Academy of Sciences, Earth Sciences,* **39**, 47-52.

BIRKENMAJER, K., DUDZIAK J. & JEDNOROWSKA A. 1979. Subsurface geological structure of the northern boundary fault zone of the Pieniny Klippen Belt at Szczawnica, Carpathians. *Studia Geologica Polonica*, **61**, 7-36.

BIRKENMAJER, K., DUDZIAK, J., JEDNOROWSKA, A. & KUTYBA, J. 1987. Foraminiferal-nannoplankton evidence for Maastrichtian and Palaeocene ages of the Jarmuta Formation: its bearing on dating Laramian orogeny in Pieniny Klippen Belt, Carpathians, Poland). *Bulletin of the Polish Academy of Sciences, Earth Sciences,* **35**, 287-298.

BIRKENMAJER, K. & GĄSIOROWSKI, M. 1961. Sedimentary character of radiolarites in the Pieniny Klippen Belt, Carpathians. *Buletin de l'Académie Polonaise des Sciences, Série des Sciences chimiques, Géologiques et Géographiques*, **9**, 175-180.

BIRKENMAJER, K., GĄSIOROWSKI, M. & WIESER, T. 1960. Fragments of exotic rocks in the pelagic deposits of the Bathonian of the Niedzica Series, Pieniny Klippen Belt, Carpathians. *Annales de la Société Géologique de Pologne*, **30**, 31-57.

BIRKENMAJER, K. & GEDL, P. 2004. Dinocyst ages of some Jurassic strata, Grajcarek Unit at Sztolnia Creek, Pieniny Klippen Belt (Poland). *Studia Geologica Polonica*, **123**, 245-277.

BIRKENMAJER, K. & JEDNOROWSKA, A. 1979. Palaeocene foraminiferal assemblage from the Magura flysch at the contact with Pieniny Klippen Belt, at Jaworki (Carpathians, Poland. *Annales de la Société Géologique de Pologne*, **49**, 287-292.

— & — 1983. Upper Cretaceaus and Lower Paleogene deposits at Maruszyna, Pieniny Klippen Belt, Poland. *Studia Geologica Polonica*, **77**, 27-53.

BIRKENMAJER, K., KOZUR, H. & MOCK, R. 1990. Exotic Triassic pelagic limestone pebbles from the Pieniny Klippen Belt of Poland: a further evidence for early Mesozoic rifting in West Carpathians. *Annales Societatis Geologorum Poloniae*, **60**, 3-44.

BIRKENMAJER, K. & MYCZYŃSKI, R. 1977. Middle Jurassic deposits and fauna of the Magura succesion, near Szlachtowa, Pieniny Klippen Belt, Carpathians. *Acta Geologica Polonica*, **27**, 337-400. [*In Polish, English summary*]

BIRKENMAJER, K. OSZCZYPKO, N. 1989. Cretaceous and Palaeogene lithostratigraphic units of the Magura Nappe, Krynica Subunit, Carpathians. *Annales Societatis Geologorum Poloniae*, **66**, 1-15.

BIRKENMAJER, K. & PAZDRO O. 1968. On the so-called "Sztolnia beds" in the Pieniny Klippen Belt of Poland. *Acta Geologica Polonica*, **18**, 325-365.

BIRKENMAJER, K. & PÉCSKAY, Z. 1999. K-Ar dating of the Miocene andesite intrusions, Pieniny Mts, West Carpathians. *Bulletin of the Polish Academy of Sciences, Earth Sciences*, **47**, 155-169.

— & — 2000a. K-Ar dating of the Miocene andesite intrusions, Pieniny MTS, West Carpathians, Poland: a supplement. *Studia Geologica Polonica*, **117**, 7-25

— & — 2000b. Early Cretaceous K-Ar age of a large basalt olistolith at Biała Woda, Pieniny Klippen Belt, West Carpathians, Poland. *Studia Geologica Polonica,* **117**, 27-35.

BIRKENMAJER, K. & SKUPIŃSKI, A. 1990. On some volcanic and plutonic exotic rock fragments from the Upper Cretaceous of the Pieniny Klippen Belt, Carpathians, Poland. *Studia Geologica Polonica*, **97**, 69-78. [*In Polish with English summary*]

BIRKENMAJER, K. & WIESER, T. 1990. Exotic rock fragments from Upper Cretaceous deposits near Jaworki, Pieniny Klippen Belt, Carpathians, Poland. *Studia Geologica Polonica*, **97**, 7-67.

BLAICHER, J. & SIKORA, W. 1969. W sprawie ogniwa zwanego aalenem fliszowym w pienińskim pasie skałkowym Polski. *Kwartalnik Geologiczny*, **13**, 716-717.

BŁASZYK, J. 1968. Ostracods from the Sztolnia creek profile in the Pieniny Klippen Belt. *Acta Geologica Polonica*, **18**, 367-373.

BOYER, S.E. & ELLIOT, D. 1982, Thrust systems. *American Association of Petroleum Geologist Bulletin*, **66**, 1196-1230.

BOJDYS, G. & LEMBERGER, M. 1986. Three-dimensional gravity modelling of Earth's crust and upper mantle in the Polish Carpathians. *Annales Societatis Geologorum Poloniae*, **56**, 349-373.

BOMBIȚĂ, G. & SAVU, H. 1986. Sur les roches volcaniques associées aux klippes piénines de Poiana Botizii (Maramouresh Roumain. *Annales Societatis Geologorum Poloniae*, **56**, 337-348.

BORZA, K. 1966. Etude lithologique et pétrographique des conglomérates de la zone des klippes – pétrographie des galets de roches sedimentares. *Nauka o Zemi, Geologica*, **2**, 1-66.

BROWN, W.G. 1975. Casper Mountain area (Woyoming) structural model of Laramide deformation (abstract). *American Assotiation of Petroleum Geologist Bulletin*, **59**, p. 906.

BUČEK, S. 2001. Borové Formation of the Eocene age east of the Tatra Mts. *Slovak Geological Magazine* **7**, 45-52.

BURCHART, J. 1963: Remarks on the directions of the slickesides and fault striae in the crystalline rocks of the Goryczkowa "crystalline island" in the Tatra Mts. *Acta Geologica Polonica,* **13**, 27-40.

— 1968. Rubidium-strontium isochron ages of the crystaline core of the Tatra Mts., Poland. *American Journal of Science,* **266**, 895-907.

— 1972. Fission-track age determination of accessory apatite

from the Tatra Mts., Poland. *Earth and Planetary Science Letters*, **15**, 418-422.

CHOWANIEC, J., POPRAWA, D. & SOKOŁOWSKI, S. 1975. Wstępne wyniki badań geologicznych otworów Siwa Woda IG-1 i Hruby Regiel IG-2. *Geological Quarterly*, **19**, 949-950.

CIESZKOWSKI, M. 1992. Marine Miocene deposits near Nowy Targ, Magura nappe, Flysch Carpathians (south Poland). *Geologica Carpathica*, **43**, 339-346.

CORTI G., BONINI M., CONTICELLI, S., INNOCENTI, F., MANETTI, P. & SOKOUTIS, D. 2003. Analogue modelling of continental extension: a review focused on the relations between the patterns of deformation and the presence of magma. *Earth Science Reviews* **6,** 169–247

CSONTOS, L., VÖRÖS, A. 2004. Mesozoic plate tectonic recon-struction of the Carpathian region. *Palaeogeography, Palaeoclimatology, Palaeoecology,* **210**, 1– 56.

CZARNECKA, K. 1986. Uwarunkowania strukturalne współczes-nych ruchów tektonicznych pienińskiego pasa skałkowego w rejonie Czorsztyna. *Przegląd Geologiczny*, **10**, 556-560.

— 2004. Próby interpretacji wyników badania współczesnej dynamiki pienińskiego pasa skałkowego w okolicy Czorsztyna. *In*: K. CZARNECKI, (*Ed*.), Geodynamical studies of the Pieniny Klippen Belt – Czorsztyn region, pp. 83-99. *Politechnika Warszawska*; Warszawa. [*In Polish with English summary*]

CZERWIŃSKI, T. & STEFANIUK, M. 2001. Recognition of the Carpathians as a result of magnetotelluric investigations. *Slovak Geological Magazine*, **7**, 139-144.

DALLMEYER, R.D., NEUBAUER, F., PUTIŠ, M. 1993. ^{40}Ar – ^{39}Ar mineral age controls for the Pre-Alpine and Alpine tectonic evolution of nappe complexes in the Western Carpathians. PAEWCR Excursion Guide, Stará Lesná, pp. 13-20.

DAL PIAZ, G.V., MARTIN, S., VILLA, I.M., GOSSO, G. & MARSCHALKO, R. 1995. Late Jurassic blueschits facies peb-bles from the Western Carpathian orogenic wedge and paleostructural implications for Western Tethys evolution. *Tectonics*, **14**, 874-885.

DEER, W.A., HOWIE, R.A. & ZUSSMAN, J. 1962. Rock forming minerals. 333 pp. London.

DOKTÓR, S., GRANICZNY, M., KOWALSKI, Z. & WÓJCIK, A. 2002. Aplication of radar images interpretation for tectonic analysis of the Carpathians. *Przegląd Geologiczny*, **50**, 852-860.

DOGLIONI, C. 1995. Geological remarks on the relationships between extension and convergent geodynamic settings. *Tectonophysics*, **252**, 253-267.

DUDZIAK, J. 1990. Palaeogene stratigraphy of the Maruszyna Succession at Szaflary, Pieniny Klippen Belt (Carpathians), based on calcareous nannoplankton. *Studia Geologica Polonica*, **97**, 79-107. [*In Polish with English summary*]

DUMONT, T., WIECZOREK, J. & BOULLIN, J.P. 1996. Inverted Mesozoic rift structures in the Polish Western Carpathians (High-Tatric units). Comparision with similar

features in the Western Alps. *Eclogae Geologicae Helvetiae*, **89**, 181-202.

DŻUŁYŃSKI S. & GRADZIŃSKI R. 1960. Source of the Lower Triassic clastics in the Tatra Mts. *Buletin de l'Acaddémie Polonaise des Sciences, Série des Sciences Chimiques, Géo-logiques et Géographiques*, **8**, 45-48.

EMBEY-ISZTIN, A. DOWNESH., JAMES, D.E. UPTON, B.G.J., DOBOSI, G., INGRAM, G.A. HARMON, R.S. & SCHARBERT, H.G.1993. The petrogenesis of Pliocene alkaline volcanic rocks from Pannonian basin, eastern central Europe. *Journal of Petrology*, **34**, 317-343.

FODOR, L. 1995. From transpression to transtension: Oligocene-Miocene structural evolution of the Vienna Basin and the East Alpine-Western Carpathian junction. *Tectonophysics,* **242**, 151-182.

FRITZ, H., NEUBAUER, F., JANÁK, M. PUTIŠ, M. 1992. Variscan midcrustal thrusting in the Carpathians II: Kinematics and fabric evolution of the Western Tatra basement. Terra Abstracts, Suplement 2 to *Terra Nova* **4**, 24.

FUGLEWICZ, R.1980. Stratigraphy and palaeogeography of Lower Triassic in Poland on the basis of megaspores. *Acta Geologica Polonica*, **30**, 417-469.

FUSÁN, O., KODYM, O. MATĚJKA, A. & URBÁNEK, L. 1967. Geologická mapa ČSSR. 1:200 000 scale. *Ústředni Ústav Geologický*; Praha

GAWĘDA, A. 1995. Geochemistry and Rb/Sr isochron age of pegmatites from the Western Tatra Mts. *Geologica Carpathica*, **46**, 95-99.

GAWĘDA, A. & PIWKOWSKI, R. 2000. Struktura i geneza pseudo-tachylitów Tatr Wysokich. *Przegląd Geologiczny* **48**, 722-726.

GEDL, P. 1998. Trophic conditions of the Oligocene Podhale Flysh basin (Central Carpathians, Poland) in the dinocyst record. The 5th European palaeobotanical and Palynologi-cal conference, June 26-30, 1998, Cracow, Poland Abstracts: p. 60.

— 1999. The age of base and top of the Podhale Palaeogene flysch (Central Carpathians, Poland), based on dinocysts. *Bulletin of the Polish Academy of Sciences, Earth Sciences.*, **47**, 77-102.

— 2000. Biostratigraphy and palaeoenvironment of the Podhale Palaeogene (Inner Carpathians, Poland) in the light of palynological studies. Part I-II. *Studia Geologica Polonica*, **117**, 89-303.

GHEBREAB, W. 1998. Tectonics of the Red Sea region reassessed. *Earth Science Reviews*, **45**, 1-44.

GHEBREAB, W. & TALBOT C.J. 2000. Red Sea extension influ-enced by Pan-African tectonic grain in eastern Eritrea. *Journal of Structural Geology*, **22**, 931-946.

GŁAZEK, J. 1963. Les séries sédimentaires du versant nord de Wołoszyn, Hautes Tatras. *Acta Geologica Polonica*, **13**, 467-480. [*In Polish with French summary*]

— 2000. Pre-Eocene karst of the Tatra Mts. Materiały 34 Sympozjum Speleologicznego, 23-25. [*In Polish*]

GŁAZEK, J., PRZYBYCIN, A. & SOCHACZEWSKI, A. 1998. Tuffite between Upper Eocene conglomerates of the Tatra Mts. (Carpathians, Poland) and its stratigraphic importance. *Przegląd Geologiczny*, **46**, 622-630.

GOLONKA, J. & BOCHAROVA, N.Y. 2000. Hot spot activity and the break-up of Pangea. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **161**, 49–69.

GOLONKA, J. & KROBICKI M. 2001. Upwelling regime in the Carpathian Tethys: a Jurassic–Cretaceous palaeogeographic and paleoclimatic perspective. *Geological Quarterly* **45**, 15–32.

— & — 2004. Jurrasic Paleogeography of the Pieniny and Outer Carpathian Basins. *Rivista Italiana di Paleontologia e Stratigrafia*, **110**, 5-14.

GOLONKA, J., OSZCZYPKO, N. & ŚLĄCZKA, A. 2000. Late Carboniferous – Neogene geodynamic evolution and paleogeography of the circum Carpathian region and adjacent areas. *Annales Societatis Geologorum Poloniae*, **70**, 107-136.

GOŁĄB, J. 1959. On the geology of the Western Podhale flysch area. *Biuletyn Instytutu Geologicznego,* **149**, 225-240.

GRABOWSKI, J. 1995a. New paleomagnetic data from the Lower Sub-Tatric radiolarites, Upper Jurassic (Western Tatra Mts). *Geological Quarterly*, **39**, 61-74.

— 1995b. Paleomagnetic investigations of the High Tatric series in the Tatra Mts. *Przegląd Geologiczny*, **43**, 113-116.

— 1997a. Paleomagnetic results from the cover (High-Tatric) units and nummulitic Eocene in the Tatra Mts (Central West Carpathians, Poland) and their tectonic implications. *Annales Societatis Geologorum Poloniae*, **67**, 13-23.

— 1997b. New paleomagnetic data from Fatricum and Hronicum in the Tatra Mts (Poland) further evidences for Cretaceous remagnetization in the Central West Carpathians *Przegląd Geologiczny*, **45**, 1074.

GRABOWSKI, J., NARKIEWICZ, K. & POPRAWA P. 1999. First results of paleomagnetic and paleothermal (CAI) investigations of the highest Sub-Tatric units in the Polish Tatra Mts. *Przegląd Geologiczny* **47**, 153-158. [*In Polish with English summary*]

GRABOWSKI, J. & NEMČOK, M. 1999. Summary of paleomagnetic data from the Central West Carpathians of Poland and Slovakia: evidence for Late Crateceous – Early Tertiary transpression. *Physics and Chemistry of the Earth*, A, **24**, 681-685.

GRECULA, P. & ROTH, Z. 1978. Kinematic model of the West Carpathians. *Sbornik Geologickych Věd, Geologie,* **32**, 49-73.

GRODZICKI, J. 1979. Indication of neotectonic movements in the Czerwone Wierchy massif. *Przegląd Geologiczny*, **7**, 382-387. [*In Polish with English summary*]

GROSS, P. KÖHLER, E. & SAMUEL, O. 1984. Lithostratigraphic division of the Central-Carpathian Paleogene. *Geologické práce – Správy*, **81**, 103-117. [*In Slovak, English summary*]

GUTERCH, A., GRAD, M., KELLER, G.R., POSGAY, K., VOZÁR, J., ŠPIČÁK, A., BRÜCKL, E., HAJNAL, Z., THYBO, H. & SELVI, O.

2003. Celebration 2000 seismic experiment. *Studia Gephysica and Geodetica (Special Contribution),* **47**, 659-669.

HALICKI, B. 1961. Z badań nad fliszem podhalańskim i magurskim na Podhalu. *Acta Geologica Polonica*, **11**, 477-482.

HÄUSLER, H., PLAŠIENKA, D. & POLÁK, M. 1993. Comparision of Mesozoic succesions of the Central Eastern Alps and the Central Western Carpathians. *Jahrbuch der Geologischen Bundesanstalt*, **136**, 715-739.

HIROSE, T. & SHIMAMOTO, T. 2003. Fractal dimension of molten surfaces as a possible parameter to infer the slip-weakening distance of faults from natural pseudotachylytes. *Journal of Structural Geology*, **25**, 1569-1574.

HRADILEC, K., LOULOVA, A. & CHARAMZA, F. 1981. Method for determining recent movements of the Earth's crust in high mountain areas. *In*: Geophysical Syntheses in Czechoslovakia. *Veda*; Bratislava.

HRUŠECKÝ, I., POSPIŠIL, L. & KOHÚT, M. 2002. Geological interpretation of the reflection seismic profile 753/92. *In*: I. HRUŠECKÝ (*Ed.*), Hydrocarbon potential of the Eastern Slovakian Basin and Adjacent Areas. GSSR, Bratislava.

IMBER, J., CHILDS, C., NELL, P.A.R., WALSH, J.J., HODGETTS, D. & FLINT, S. 2003. Hanging wall fault kinematics and footwall collapse in listric growth fault systems. *Journal of Structural Geology*, **25**, 197-208.

JACH, R., DUDEK, T. & BARSKI, M. 2003. Toarcian manganese-bearing deposits in the Krížna Unit (Tatra Mts., Poland). A record of submarine hydrothermal activity. *In*: 22nd IAS Meeting of International Assotiation of Sedimentologist, Opatijia, September 17-19 2003, p.87.

JACZYNOWSKA, W. 1980. Geological map of the Polish Tatra Mts – sheet Mięguszowiecki, 1:10000 scale. *Instytut Geologiczny*; Warszawa.

JAGLARZ, P. & SZULC, J. 2003. Middle Triassic evolution of the Tatricum sedimentary basin: an attempt of sequence stratigraphy to the Wierchowa Unit in the Polish Tatra Mountains. *Annales Societatis Geologorum Poloniae*, **73**, 169-182.

JANÁK, M. 1993. Calc-silicate metamorphic rocks of the High Tatra crystalline basement. *Mineralia Slovaca*, **25**, 177-182.

— 1994. Variscan uplift of the crystalline basement, Tatra Mts, Central West Carpathians: evidence from ^{40}Ar/^{39}Ar laser probe dating of biotite and P-T paths. *Geologica Carpathica*, **45**, 239-300.

JANÁK, M., O'BRIEN, P.J., HURAI, V. & REUTEL, C. 1996. Metamorphic evolution and fluid composition of garnet-clinopyroxene amphibolites from the Tatra Mts., Western Carpathians. *Lithos*, **39**, 57-79.

JANÁK, M., HURAI, V., LUDHOVÁ, L., O'BRIEN, P.J. & HORN E. 1999. Dehydration melting and devolatilization during exhumation of high-grade metapelites: the Tatra Mts, Western Carpathians. *Journal of Metamorphic Geology*, **17**, 379-395.

JANOČKO, J. & JACKO, S. 1998. Marginal and deep-sea deposits

of Central-Carpathian Paleogene Basin, Spišská Magura region, Slovakia: implication for basin history. *Slovak Geological Magazine* 4, 281-292.

JAROSIŃSKI, M. 1998. Contemporary stresss field distortion in the Polish part of the Western Outer Carpathians and their basement. *Tectonophysics*, **297**, 91-119.

JAROSZEWSKI, W. 1982. Hydrotectonic phenomena at the base of the Krížna nappe, Tatra Mts. *In*: M. MAHEL' (*Ed.*), Alpine structural elements: Carpathian-Balkan-Caucasus-Pamir orogene zone, pp. 137-148. *Veda*; Bratislava.

JUREWICZ, E. 1994. Structral analysis of the Pieniny Klippen Belt at Jaworki, Carpathians, Poland. *Studia Geologica Polonica*, **106**, 7-87.

— 1997. The contact between the Pieniny Klippen belt and Magura Unit (the Małe Pieniny Mts.) *Geological Quarterly*, **41**, 315-326.

— 2000a. Tentative reconstructions of the stress axes from the thrust-folding stage in the Tatra Mts. on the basis of slickensides in the granitoid core, southern Poland. *Przegląd Geologiczny*, **48**, 239-246. [*In Polish with English summary*]

— 2000b: Tentative correlation of the results of structural analysis in the granitoid core and nappe units of the Tatra Mts., southern Poland. *Przegląd Geologiczny*, **48**, 1014-1018. [*In Polish with English summary*]

— 2002. Geometric analysis of steep dipping dislocations within the granitoid core in the Polish part of the Tatra Mts. *Annales Societatis Geologorum Poloniae*, **72**, 89-98.

— 2003. Multistage evolution of the shear zone at the base of the Giewont Unit Polish Tatra Mts. *Geologica Carpathica*, **54**, 337-351.

JUREWICZ, E. & BAGIŃSK,I B. 2001. Struktura i geneza pseudotachylitów Tatr Wysokich – polemika. *Przegląd Geologiczny*, **49**, 603-604.

— & — 2005. Deformation phases in the selected shear zones within the Tatra Mts granitoid core. *Geologica Carpathica*, **56**, 17-28.

JUREWICZ, E. & KOZŁOWSKI, A. 2003. Formation conditions of quartz mineralisation in the mylonitic zones and on the slickenside fault planes in the High Tatra granitoids. *Archiwum Mineralogiczne*, **54**, 65-75.

JUREWICZ, E. & SŁABY, E. 2004. The Zadnie Kamienne "ravenous" shear zone (High-Tatric Nappe) - conditions of deformation. *Geological Quarterly*, **48**, 371-382.

KEEP, M. 2000. Models of lithospheric-scale deformation during plate collision: effects of indentor shape and lithospheric thickness. *Tectonophysics*, **326**, 203-216

KENNEDY, L.A. & LOGAN, J.M. 1998. Microstructures of cataclasites in a limestone-on-shale thrust fault: implications for low-temperature recrystallization of calcite. *Tectonophysics*, **295**, 167–186.

KĘPIŃSKA, B., PAWLIKOWSKI, M. & NAGEL, J. 2000. Biotitic tuffites in the Eocene from the well Biały Dunajec PAN-1

(Podhale region, southern Poland). *Przegląd Geologiczny*, **48**, 1045-1052.

KHALIL, S.M. & MCCLAY, K.R. 2002. Extensional fault-related folding, northwestern Red Sea, Egypt. *Journal of Structural Geology*, **24**, 743-762

KLITYNSKI, W. & WÓJCICKI, A. 2001. Western Carpathians basement according to results of integrated geophysical investigation with the particular consideration of magnetotelluric data. *Slovak Geological Magazine* **7**, 155-161

KNIPE, R.J. 1989. Deformation mechanisms – recognition from natural tectonites. *Journal of Structural Geology,* **11**, 127-146.

KOHÚT, M. & JANÁK, M. 1994. Granitoids of the Tatra Mts., Western Carpathians: field relationship and petrogenetic implications. *Geologica Carpathica* **45**, 301-311.

KOHÚT, M. & SHERLOCK, S. 2002. Pseudotachylites from the High Tatra Mts. – evidences for late paleogene seismic/tectonic events. XVII. Congress of Carpathian-Balkan Geological Association, September 1st – 4th 2002, 1-4.

KOHÚT, M. & SHERLOCK, S.C. 2003. Laser microprobe Ar-40-Ar-39 analysis of pseudotachylyte and host-rocks from the Tatra Mountains, Slovakia: evidence for late Palaeogene seismic/tectonic activity, *Terra Nova*, **15**, 417-424

KOISAR, B. & ZAWIDZKI, P. 1972. Violet-coloured mylonite from the Tatra Mts. *Acta Geologica Polonica,* **22**, 127-142.

KOPF, R. W. 1982: Hydrotectonics: Principles and Relevance: U.S. Geological Survey Open-File Report, 82-307, 30 p.

— 2003. The hydrotectonic hypothesis: a tectonically-activated hydraulic system. Unpublished manuscript donated to the U.S. Geological Survey, 38 pp.

KOSTKA, A. 1993. The age and microfauna of the Maruszyna Succession (Upper Cretaceous-Palaeogene), Pieniny Klippen Belt, Carpathians, Poland. *Studia Geologica Polonica*, **102**, 7-134.

KOTAŃSKI, Z. 1956. High-Tatric Campillian in the Tatra Mts. *Acta Geologica Polonica*, **6**, 65-73. [*In Polish with English summary*]

— 1959a: Stratigraphical sections of the High-Tatric Series in the Tatra Mts. *Biuletyn Instytutu Geologicznego*, **139**, 1-139. [*In Polish with English summary*]

— 1959b: Contributions to the tectonics of the High-Tatric series. *Biuletyn Instytutu Geologicznego*, **149**, 159-174. [*In Polish with English summary*]

— 1961. Tectogénèse et reconstitution de la paléogéographie de la zone haut-tatrique dans les Tatras. *Acta Geologica Polonica*, **11**, 187-467. [*In Polish with French summary*]

— 1963a. On the character of the Western Carpathian Mesozoic geosyncline and the Podhale epimiogeosyncline. *Acta Geologica Polonica*, **13**, 13-25. [*In Polish with English summary*]

— 1963b. New elements in the structure of Czerwone Wierchy massif. *Acta Geologica Polonica*, **13**, 149-198.

— 1963c. On the Triassic of the Haligovce Klippen and the palaeogeographic position of the Haligovce series. *Acta Geologica Polonica*, **13**, 295-313.

— 1965. Analogies lithologiques entre le Trias de la Tatra et celui des Alpes orientales. *Acta Geologica Polonica*, **35**, 143-162.

KOVÁČ, P. & HÓK J. 1996. Tertiary development of western part of Klippen Belt. *Slovak Geological Magazine*, **2**, 137-149.

KOVÁČ, M., KRÁĽ, J., MÁRTON, E., PLAŠIENKA, D. & UHER P. 1994. Alpine uplift history of the Central Western Carpathians: geochronological, paleomagnetic sedimentary and structural data. *Geologica Carpathica*, **45**, 83-96.

KOVÁČ, M. & MÁRTON E. 1998. To rotate or to not rotate: Palinspastic reconstruction of the Carpatho-Pannonian area during the Miocene. *Slovak Geological Magazine* **4**, 75-85.

KOVÁČ, M., NAGYMAROSY, A., OSZCZYPKO, N., CSONTOS, L., ŚLĄCZKA A., MARUNTENAU, M., MATENCO, L. & MÁRTON, E. 1998. Palinspastic reconstruction of the Pannonian-Carpathian region during the Miocene. *In*: M. RAKÚS (*Ed*.), Geodynamic development of the Western Carpathians, 189-217. *GÚDŠ*; Bratislava

KOZUR, H. & MOCK, R. 1996. New paleogeographic and tectonic interpretations in Slovakian Carpathians and their implications for correlations with Eastern Alps. Part I: Central Western Carpathians. *Mineralia Slovaca* **28**, 151-174.

KRAJEWSKI, K. 1980. Towards sedimentological investigation of High-Tatric rocks in the Kościeliska Valley. *Przegląd Geologiczny* **11**, 636-637. [*In Polish with English summary*]

— 2003. Facies development and lithostratigraphy of the Hightatric mid-Cretaceous (Zabijak Formation) in the Polish Tatra Mountains. *Studia Geologica Polonica*, **121**, 81-158.

KRÁĽ, J. 1977. Fission track ages of apatites from some granitoid rocks in West Carpathians. *Geologick_ Zborník – Geologica Carpathica*, **28**, 269-276.

KRÁĽ, J., HESS, J.C., KOBER, B. & LIPPOLT, H. 1997. [207]Pb/[206]Pb and [40]Ar/[39]Ar age data from plutonic rocks of Strazˇowské vrchy Mts basement, Western Carpathians. In: *Mineralia Slovaca, Monography*, 253-260.

KRAWCZYK, A. & SŁOMKA, T. 1981. Trace-fossil Paleodiction from the Szlachtowa Formation, ?Upper Toarcian-Lower Aalenian, of the Pieniny Klippen Belt, Poland. *Studia Geologica Polonica*, **70**, 67-71. [*In Polish with English summary*]

KRAWCZYK, A. & SŁOMKA, T. 1986. Development and sedimentation of the Szlachtowa Formation (Jurassic flysch) east of Szczawnica (Grajcarek Unit, Pieniny Klippen Belt, Carpathians). *Studia Geologica Polonica*, **88**, 33-134. [*In Polish with English summary*]

KROBICKI, M., GOLONKA, J., LEWANDOWSKI, M., OSZCZYPKO, N., POPADYUK, I. & SŁABY, E. 2004. Volcanism of the Jurassic-Cretaceous triple-junction zone in the Eastern Carpathians. *Geolines*, **17**, 60-61.

KRÓLIKOWSKI, C. & PETECKI, Z. 2001. Recent results of the gravity and magnetotelluric modelling: lithosphere structure in the Polish Carpathians. *Slovak Geological Magazine*, **7**, 131-138.

KRUCZYK J., KĄDZIAŁKO-HOFMOKL, M., LEFELD, J., PAGÁČ, P. & TÚNYI, I. 1992. Paleomagnetism of Jurassic sediments as evidence for oroclinal bending of the Central West Carpathians. *Tectonophysics*, **206**, 315-324.

KRZYWIEC, P. & JOCHYM, P. 1997. Characteristics of the Miocene subduction zone of the Polish Carpathians: results of flexural modelling. *Przegląd Geologiczny*, **45**, 785-792. [*In Polish with English summary*]

KSIĄŻKIEWICZ, M. 1972. Budowa geologiczna Polski, IV Tektonika, 3 – Karpaty, 288 pp.*Wydawnictwa Geologiczne*; Warszawa.

KUTEK, J. & WIERZBOWSKI, A. 1986. A new account on the Upper Jurassic stratigraphy and amonites of the Czorsztyn succession, Pieniny Klippen Belt, Poland. *Acta Geologica Polonica,* **36**, 298-316.

KUTYBA, J. 1986. Dolina Grajcarka: formacja jarmucka jednostki Grajcarka, in: Przewodnik 57 Zjazdu PTG, Pieniny, 18-20 września 1986. Birkenmajer K. & Poprawa D. (*Eds*), Kraków.

KYSELA, J., MARSCHALKO, R. & SAMUEL, O. 1982. Litostratigrafická klasifikácia vrchnokriedových sedimentov manínskej jednotky. *Geologické Práce, Správy,* **78**, 143-167.

LEFELD, J. 1981. Upper Jurassic radiolarite-nodular limestone vertical symmetry in the Polish Central Carpathians as reflection of regional depth changes in the ocean. *Studia Geologica Polonica,* **68**, 89-96.

LEFELD, J., GAŚDZICKI, A., IWANOW, A. KRAJEWSKI, K. 1985. Jurrasic and Cretaceous lithostratigraphic units of the Tatra Mts. *Studia Geologica Polonica*, **84**, 1-86.

LEFELD, J. & JANKOWSKI, J. 1985. Model of deep structure of the Polish Central Carpathias. *Publications of the Institute of Geophysics Polish Academy of Sciences*, A-16. **175**, 593-614.

LEŠKO B., ĎURKOVIČ, T., GAŠPARIKOVÁ, V., SAMUEL, O. & SNOPKOVÁ, P. 1984. Geological significance of the Hanušovce 1 borehole. *Mineralia Slovaca*, 1**6**, 217-255. [*In Slovak with English summary*]

LEŚNIAK, P., SAKAI, H., ISHIBASHI, J. & WAKITA, H. 1997. Mantle helium signal in the West Carpathians, Poland. *Geochemical Journal*, **31**, 383-394.

LIN, A. SUN, Z. & YANG, Z. 2003. Multiple generation of pseudotachylyte in the brittle to ductile regimes, Qinling-Dabie Shan ultrahigh-pressure metamorphic complex, central China. *The Island Arc*, **12**, 423-435.

LINZER, H.G., DECKER, K., PERESSON, H., DELL'MOUR., R. & FRISCH, W. 2002. Balancing lateral orogenic float of the Eastern Alps. *Tectonophysics,* **354**, 211-237

ŁUCZYŃSKI, P. 2001a. Development history of Middle Jurassic neptunian dykes in the High-Tatric series, Tatra Mountains, Poland. *Acta Geologica Polonica*, **51**, 237-252.

— 2001b. Pressure-solution and compaction of condensed Middle Jurassic deposits, High Tatric series, Tatra Mts. *Geologica Carpathica*, **52**, 91-102.

MAHEĽ, M. 1950. Obalová séria Inovca. *Geologický Zborník*, **1**, 47-58.

— 1978. Manín tectonic unit relotions of the Klippen Belt and Central West Carpathians. *Geologický Zborník – Geologica Carpathica*, **29**, 197-214.

— 1980. Pribladlové pâsmo, charakteristika a význam. Mineralia Slovaca **13**, 289-306.

— 1981. Penninic units in the Western Carpathians from the point of the view of global tectonics. *Mineralia Slovaca*, **13**, 289-306.

— 1986. Geologická stavba Československých Karpat. (1) Paleoalpínske jednotky, pp. 503. *Veda*; Bratislava.

— 1989. The Klippen Belt from the aspect of geodynamic model. *Mineralia Slovaca*, **21**, 99-108. [*In Slovak with English summary*]

MAKOWSKA, A. & JAROSZEWSKI, W. 1987. On present vertical movement in Tatra Mts. and Podhale. *Przegląd Geologiczny* **10**, 506-512.

MALUSKI, H., RAJLICH, P. & MATTE, PH. 1993. $^{40}Ar – ^{39}Ar$ dating of the Central Carpathians Variscan basement and Alpine mylonitic overprinting. *Tectonophysics*, **223**, 313-337.

MAŁKOWSKI, S. 1921. Les andésites des environs de Piénines. *Prace Państwowego Instytutu Geologicznego*, **1**, 3-67.

MARCHANT, R.H. & STAMPFLI, G.M. 1997. Subduction zone in the Western Alps. *Tectonophysics*, **269**, 217-235.

MARKO, F., PLAŠIENKA, D. & FODOR, L. 1995. Meso-Cenozoic tectonic stress fields within the Alpine Carpathian transition zone: a review. *Geologica Carpathica*, **46**, 19-27.

MÁRTON, E., MASTELLA, L. & TOKARSKI A.K. 1999. Large counterclockwise rotation of the Central Paleogene Flysch – evidence from paleomagnetic investigations of the Podhale Flysch (Poland). *Physics and Chemistry of the Earth,* A, **24**, 645-649.E.

MÁRTON, A.K. TOKARSKI & D. HALÁSZ 2004. Late Miocene counterclockwise rotation of the Pieniny andesites at the contact of the Central and Outer Western Carpathians. *Geologica Carpathica*, **55**, 411-419.

MARSCHALKO, R. 1968. Facies, distribution, paleocurrents and paleotectonics of the Paleogene flysch of Central West Carpathians. *Geologický Zborník – Geologica Carpathica*, **19**, 69-94.

— 1975. Depositional environment of conglomerate as interpreted from sedimentological studies. *Nauka o Zemi, Geologica*, **9**, 1-145.

MARSCHALKO, R. & RADOMSKI, A. 1960. Preliminary results of investigations of current directions in the flysch basin of the Central Carpathians. *Annales Societatis Geologurum Poloniae*, **30**, 259-272.

MASTELLA, L. 1975. Flysch tectonic in the eastern part of the Podhale Basin (Carpathians, Poland. *Annales Societatis Geologorum Poloniae*, **45**, 361-401. [*In Polish with English summary*]

MASTELLA, L. & MIZERSKI, W.1977. Uwagi o budowie SW części Podhala. *Przegląd Geologiczny*, **10**, 494-498. Warszawa

MASTELLA, L., OZIMKOWSKI, W. & SZCZĘSNY, R. 1988. Tectonics of the northern Podhale Flysch. *Przegląd Geologiczny*, **36**, 566-572. [*In Polish with English summary*]

MEULENKAMP, J.E., KOVÁC, M. & CICHA, I. 1996. On Late Oligocene to Pliocene depocentre migrations and the evolution of the Carpathian-Pannonian system. *Tectonophysics*, **266**, 301-317.

MICHALÍK, J. 1997. Tsunamites in a storm-dominated Anisian carbonate ramp (Vysoká Formation, Malé Karpaty Mts., Western Carpathians). *Geologica Carpathica* **48**, 221-229.

MICHALÍK, J., PLANDEROVÁ, E. & SÝKORA, M. 1976. To the strathigraphic and paleogeographic position of the Tomanová Formation of the uppermost Triassic of the West Carpathians, *Geologický Zborník - Geologica Carpathica,* **27**, 299-318.

MILOVSKÝ, R. 1997. Rauhwackes and their importance in the thrust tectonics (Case studies from Muráň, Silica Drienok Choč and Krížna Nappe). *In*: Alpine evolution of the Western Carpathians and related areas, p. 32. *Dionýz Štúr Publishers*; Bratislava

MILOVSKÝ, R., HURAI, V., PLAŠIENKA, D. & BIROŇ, A. 2003. Hydrotectonic regime at soles of overthrust sheets: textural and fluid inclusion evidence from basal cataclasites of the Muráň nappe (Western Carpathians, Slovakia). *Geodinamica Acta*, **16**, 1-20.

MIŠÍK, M. & AUBRECHT, R. 1994. Source of rock fragments in the Jurassic crinoidal limestones of the Pieninicum (Klippen Belt, Western Carpathians). *Geologica Carpathica*, **45**, 159-170.

MIŠÍK, M. & JABLONSKÝ, J. 2000. Lower Triassic quartzites of the Western Carpathians: transport directions, source of clastics. *Geologica Carpathica*, **51**, 251-264.

MIŠÍK, M., JABLONSKÝ, J., FEJDI, P. & SÝKORA, M. 1980. Chromian and ferrian spineles from Cretaceous sediments of the West Carpathians. *Mineralia Slovaca*, **12**, 209-228.

MIŠÍK, M. & SÝKORA, M. 1981. Der pieninische exotische Rücken, rekonstruiet aus Geröllen karbonatischer Gesteins Kretazischer Konglomerate der Klippenzone und der Manín-Einheit. *Západne Karpaty, Geológia*, **7**, 7-111.

MIŠÍK, M., SÝKORA, M., MOCK, R. & JABLONSKÝ, J. 1991. Paleogene Proč conglomerates of the Klippen belt in the West Carpathians, material from neopienninic exotic ridge. *Acta Geologica et Geographica Universitatis Comenianae,* **46**, 9-101.

MIŠÍK M., SÝKORA, M. & AUBRECHT, R. 1994. Middle Jurassic scarp breccias with clefts filled by Oxfordian and Valanginian-Hauterivian sediments, Krasín near Dolná Súča (Pieniny Klippen Belt). *Geologica Carpathica*, **45**, 343-

356.

NEJBERT, K. & JUREWICZ, E. 2004. Mineralogy and petrography of the ferruginous concretions from siliciclastic Rhaetian deposits, Tatra Mountains, Poland. *Mineralogical Society of Poland Special Papers*, **24**, 299-302.

NEMČOK, M. & KANTOR, J. 1990. Pohybová Štúdia vybranej oblasti jednotky Velkého Boku. *Regionálna Geológia Západných Karpát: Správy o vyskumoch*. GÚDŠ, Bratislava, pp.75-83.

NEMČOK, M. & NEMČOK, J. 1994. Late Cretaceous deformation of the Pieniny Klippen Belt, West Carpathians. *Tectonophysics*, **239**, 81–109.

NEMČOK, M., POSPÍŠIL, L., LEXA, J. & DONELICK, R.A. 1998. Tertiary subduction and slab break-off model of the Carpathian-Pannonian region. *Tectonophysics,* **295**, 307-340.

NEY, R. 1975. Tectogenesis of the Carpathians in the light of new tectonics of the Earth's globe. *Publications of the Institute of Geophysics Polish Academy of Sciences*, **82**, 95-110.

NIEDŹWIEDZKI, G. 2005. A new find of dinosaur footprints in the Upper Triassic of the Tatra Mountains, southern Poland. *Przegląg Geologiczny*, **53**, 410–413. [*In Polish with English summary*]

OLSZEWSKA, B. & WIECZOREK, J. 1998. The Paleogene of the Podhale Basin (Polish Central Carpathians) – micropaleontological perspective. *Przegląd Geologiczny*, **46**, 721-728.

ORD, A. & HOBBS, B.E. 1989. Strenght of the continental crust, detachment zones and the development of plastic instabilites. *Tectonophysics*, **158**, 269-289.

OSZAST, J. 1973. The Pliocene profile of Domański Wierch near Czarny Dunajec in the light of palynological investigations, Western Carpathians, Poland. *Acta Paleobotanica*, **14**, 1-42.

OSZCZYPKO, N. 1998. The Western Carpathian Foredeep — development of the foreland basin in front of the accretionary wedge and its burial history (Poland). *Geologica Carpathica,* **49,** 415-431.

— 1999. From remnant ocean basin to collision-related foreland basin – a tentative history of the Outer Western Carpathians. *Geologica Carpathica*, **50**, Spec. Issue, pp. 161-165.

OSZCZYPKO, N., ANDREYEVA-GRIGOROVICH, A.S., MALATA, E. & OSZCZYPKO-CLOWES, M. 1999. The Lower Miocene deposits of the Rača Subunit near Nowy Sącz (Magura Nappe, Polish Outer Carpathians). *Geologica Carpathica*, **50**, 419-433.

OSZCZYPKO, N., MALATA, E., ŠVÁBENICKÁ, L., GOLONKA, J. & MARKO, F. 2004. Jurassic-Cretaceous controversies in the Western Carpathian Flysch: the "black flysch" case study. *Cretaceous Research*, **25**, 89-113.

OSZCZYPKO, N. & OSZCZYPKO-CLOWES, M. 2002. Newly discovered Early Miocene deposits in the Nowy Sącz area (Magura Nappe, Polish Outer Carpathians). *Geological Quarterly*, **46**, 15-30.

OSZCZYPKO, N. & OSZCZYPKO-CLOWES, M. 2003. The Aquitanian marine deposits in the basement of Polish Western Carpathians and its palaeogeographical and palaeotectonic implications. *Acta Geologica Polonica*, **53**, 101-122.

OSZCZYPKO, N. & ŚLĄCZKA, A. 1985. An attempt to palinspastic reconstruction of Neogene basins in the Carpathian Foredeep. *Annales Societatis Geologorum Poloniae,* **55**, 55-75.

OSZCZYPKO, N. & ZUBER, A. 2002. Geological and isotopic evidence of diagenetic waters in the Polish Flysch Carpathians. *Geologica Carpathica*, **53**, 257-262.

OSZCZYPKO-CLOWES, M. 2001. The nannofossils biostratigraphy of the youngest deposits of the Magura nappe (East of the Skawa River, Polish flysch Carpathians) and their palaeoenviromental conditions. *Annales Societatis Geologorum Poloniae*, **71**, 139-188.

OSZCZYPKO-CLOWES, M. & OSZCZYPKO, N. 2004. The position and age of the youngest deposits in the Mszana Dolna and Szczawa tectonic windows (Magura Nappe, Western Carpathians, Poland). *Acta Geologica Polonica*, **54**, 339-367.

PASSENDORFER, E. 1949. Matériaux pour la connaissance de la géologie des Tatras. Sur le conglomérat de Koperszady. *Annales de la Société Géologique de Pologne*, **19**, 401-417.

— 1958. About sedimentation of the Eocene in the Tatra Mts. *Acta Geologica Polonica,* **8**, 451-475. [*In Polish with English summary*]

— 1974. Geological structure of the Tatra Valleys. *Czasopismo Geograficzne*, **45**, 9-29.

— 1983. Jak powstały Tatry? 286 pp. *Wydawnictwa Geologiczne*; Warszawa.

PASSENDORFER, E. & RONIEWICZ, P. 1963. Additional notes on an Eocene Island in the Tatra Mts. *Acta Geologica Polonica,* **13**, 1-12.

PETRÍK, I. & REICHWALDER, P. 1996. Pseudotachylyte from the High Tatra Mts. *Mineralogical Society of Poland, Special Papers,* **7**, 61-63.

PETRÍK, I., NABELEK, P., JANÁK, M. & PLAŠIENKA, D. 2003. Condition of formation and crystallization kinetics of highly oxidized pseudotachylytes from the High Tatras (Slovakia). *Journal of Petrology*, **44**, 901-927.

PIN, C., BOUVIER, A. & ALEKSANDROWSKI, P. 2004. Major trace element and Sr-Nd Isotope Data on Neogene andesitic rocks from the Pieniny Klippen Belt (southern Poland) and geodynamic inferences. *Mineralogical Society of Poland – Special Paper*, **24**, 323-328.

PIOTROWSKI, J. 1978. Mesostructural analysis of the main tectonic units of the Tatra Mts. *Studia Geologica Polonica*, **55**, 1-80. [*In Polish with English summary*]

PLAŠIENKA, D. 1991: Mesozoic tectonic evolution of the epi-Variscan continental crust of the Central Western Carpathians – a tentative model. *Mineralia Slovaca*, **23**, 447-457.

— 1995a. Pasive and active margin history of the northern Tatricum (Western Carpathians, Slovakia). *Geologische Rundschau*, **84**, 748-760.

— 1995b. Mesozoic evolution of Tatric units in the Malé Karpaty and Považský Inovec Mts: implication for the position of the Klape and related units in Western Slovakia. *Geologica Carpathica*, **46**, 106- 112.

— 1996. Mid-Cretaceous (120-80 Ma) orogenic processe in the Central Western Carpathians: brief review and interpretation of data. *Slovak Geological Magazine*, **3-4**, 319-324.

— 1997. Cretaceous tectonochronology of the Central Western Carpathians, Slovakia. *Geologica Carpathica*, **48**, 99-111.

— 1999. Tectonochronology and paleotectonic evolution of the Central Western Carpathians during the Jurassic and Cretaceous. 127 pp. *Veda*; Bratislava. [*In Slovak with English summary*]

— 2003a. Development of basement-involved fold and thrust structures exemplified by the Tatric-Fatric-Veporic nappe system of the Western Carpathians (Slovakia). *Geodynamica Acta*, **16**, 21-38

— 2003b. Dynamics of Mesozoic pre-orogenic rifting in the Western Carpathias. *In*: 5th Workshop of Alpine Geological Studies. *Mitteilungen der Österreichischen Geologischen Gesesellschaft*, **94**, 79-98. [for 2001]

PLAŠIENKA, D., GRECULA, P, PUTIŠ, M., KOVÁČ, M. & HOVORKA, D. 1997. Evolution and structure of the Western Carpathians: an overview. *Mineralia Slovaca*, *In*: GRECULA P., PUTIŠ M. & HOVORKA D. (*Eds*), Geological evolution of the Western Carpathians: 1-24.

PLAŠIENKA, D. & PROKEŠOVÁ, R. 1996. Towards an evolutionary tectonic model of the Krížna cover nappe (Western Carpathians, Slovakia). *Slovak Geological Magazine,* **3-4**, 279-286.

PLAŠIENKA, D. & SOTÁK, J. 1996. Rauhwackized carbonate tectonic breccias in the West Carpathian nappe edifice: introductory remarks and preliminary results. *Slovak Geological Magazine*, **3-4**, 287-291.

POLLER, U. & TODT, W. 2000. U-Pb single zircon data of granitoids from the High Tatra Mountains (Slovakia): implications for the geodynamic evolution. *Transactions of the Royal Society of Edinburgh: Earth Sciences,* **91**, 235-243.

POLLER, U., JANÁK, M., KOHÚT, M. & TODT, W. 2000. Early Variscan magmatism in the Western Carpathians: U-Pb zircon data from granitoids and orthogneisses of the Tatra Mountains (Slovakia). *International Journal of Earth Science,* **89**, 336-349.

POLLER, U., TODT, W., JANÁK, M., KOHÚT, M. 1999. The relationship between the Variscides and the Western Carpathians basement: New Sr, Nd and Pb-Pb isotope data from the Tatra Mountains. *Geologica Carpathica,* **50**, 131-133.

POSPÍŠIL, L., BEZÁK, V., NEMČOK, J., FERANEC, J. VASS, D. & OBERNAUER, D. 1989. The Muráň tectonic system as example of horizontal displacement in the West Carpathians. *Mineralia Slovaca*, **4**, 305-322.

POSPÍŠIL, L. & FILO, M. 1980. The West Carpathian central gravity minimum and its interpretation. *Mineralia Slovaca*, **2**, 149-164.

POTFAJ, M. 1997. Relation of the Klippen Belt and the Flysch zone in the territory of western Slovakia. *In*: Alpine evolution of the Western Carpathians and related areas, pp. 119-126. *Dionýz Štúr Publishers*; Bratislava

PROKEŠOVÁ, R. 1994. Structural analysis of the Krížna nappe in its near-root and superficial position. *Mineralia Slovaca*, **26**, 254-347. [*In Slovak with English summary*]

PUTIŠ, M. 1992. Variscan and Alpidic nappe structures of the Western Carpathian crystalline basement. *Geologica Carpathica*, **43**, 369-380.

RACIBORSKI, M. 1980. Flora retycka w Tatrach (in Polish only). *Rozprawy Akademii Umiejętności*, **21**, 243-260.

RADWAŃSKI, A. 1968. Petrographical and sedimentological studies of the High-Tatric Rhaetic in the Tatra Mountains. *Studia Geologica Polonica*, **25**, 1-114. [*In Polish with English summary*]

RAKÚS, M. & MARSCHALKO, R. 1997. Position of the Manín, Drietoma and Klape units at the boundary of the Central and Outer Carpathians. *In*: Alpine evolution of the Western Carpathians and related areas, pp. 79-97. *Dionýz Štúr Publishers*; Bratislava.

RĄCZKOWSKI, W., WÓJCIK, A. & ZUCHIEWICZ, W. 1984. Late Neogene-Quaternary tectonics of the Polish Carpathians in the light of neotectonic mapping. *Tectonophysics,* **108**, 51-69.

RING, U., BRANDON, M.T. & RAMTHUN, A. 2001: Solution-mass-transfer deformation adjacent to the Glarus Thrust, with implications for the tectonic evolution of the Alpine wedge in eastern Switzerland. *Journal of Structural Geology*, **23**, 1491-1505.

RONIEWICZ, P. 1966. Clastic deposits of the Lower Werfenian (Seis) in the Tatra Mts. *Acta Geologica Polonica*, **16**, 1-90. [*In Polish with English summary*]

— 1969. Sedimentation of the Nummulite Eocene in the Tatra. *Acta Geologica Polonica,* **19**, 503-608. [*In Polish with English summary*]

ROYDEN, L.H. 1993. The tectonic expression slab pull at continental convergent boundaries. *Tectonics* **12**, 303–325.

RUBINKIEWICZ, J. & LUDWINIAK, M. in press. Fracture and fault development in the Werfenian quartzite sandstone – case study from the High-Tatric autochtonous cover (Tatra Mts., Poland). *Annales Societatis Geologorum Poloniae.*

SĂNDULEŞCOU, M. 1988. Cenozoic tectonic history of the Carpathians. *American Association of Petroleum Geologist Memoirs*, **45**, 17–25

SAMUEL, O. 1977. Find of foraminifers from variegated beds near Dobšinská Ľadová Jaskyňa ice cave) and their stratigraphic interpretation. *Geologické Prace, Správy*, **67**, 93-103.

SAMUEL, O. 1997. Survey of Upper Cretaceous and Paleogene biostratigraphy of the Klippen and Periklippen Belts in the middle Váh valley. *In*: Alpine evolution of the Western

Carpathians and related areas, pp. 99-117. *Dionýz Štúr Publishers*; Bratislava.

SCHEIBNER, E. 1968. The Klippen Belt of Carpathians. *In*: M. MAHEĽ, T. BUDAY & *al*. (*Eds*), Regional geology of Czechoslovakia, II: The West Carpathians, pp. 304-371. *Academia*; Praha.

SEGHEDI, I., DOWNES, H., VASELLI, O., SZAKÁCS, A., BALOGH, K. & PÉCSKAY, Z. 2004. Post-collisional Tertiary–Quaternary mafic alkalic magmatism in the Carpathian–Pannonian region: a review. *Tectonophysics*, **393**, 43-62.

SIBSON, R.H. 1996. Structural permeability of fluid-driven fault-fracture meshes. *Journal of Structural Geology*, **18**,1031-1042.

SIBSON, R.H. 2000. Fluid involvement in normal faulting. *Journal of Geodynamics* **29**, 469-499.

SIKORA, W. 1971. Esquisse de la tectogénèse de la zone des Klippes des Pieniny en Pologne d'après de nouvelles données géolgiques. *Rocznik Polskiego Towarzystwa Geologicznego*, **4**, 221-239.

— 1976. On lineaments found in the Carpathians. *Annales Societatis Geologorum Poloniae,* **46**, 3-37.

SOCHACZEWSKI, A. 1997. Profil Złotego Potoku. Przewodnik 67 Zjazdu PTG, Zakopane 2-4 października 1997. *PIG*; Warszawa.

SOKOŁOWSKI, S. 1959. Geological map of the Nummulitic Eocene Region – Northern Margin of the Polish Tatra. *Biuletyn Instytutu Geologicznego*, **149**, 197-212. [*In Polish with English summary*]

SOKOŁOWSKI, S. 1978. Geologia strefy reglowej Tatr Polskich między Kopami Sołtysimi a Białką. *Prace Muzeum Ziemi*, **28**, 35-61.

SOKOŁOWSKI, S. & JACZYNOWSKA W. 1979. Geological map of the Polish Tatra Mts. – sheet Kopy Sołtysie, 1:10 000 scale. *Wydawnictwa Geologiczne*; Warszawa.

SOTÁK, J. 1992. Evolution of the Western Carpathia suture zone – Principal geotectonic events. *Geologica Carpathica*, **43**, 355-362.

SOTÁK, J. PERESZLÉNYI, M., MARSCHALKO, R., MILIČKA, J., & STAREK, D. 2001. Sedimentology and hydrocarbon habitat of the submarine-fan deposits of the Central Carpathian Paleogene Basin (NE Slovakia). *Marine and Petroleum Geology*, **18**, 87-114.

SOTÁK, J., PLAŠIENKA, D., SPIŠIAK, J. & UHER, P. 1993. Neptunian carbonate dykes hosted by basic volcanic rocks in the Považský Inovec Mts. – Western Carpathians. *Mineralia Slovaca*, **25**, 193-201. [*In Slovak with English summary*]

SPERNER, B. 1996. Computer programs for the kinematic analysis of brittle deformation structures and the Tertiary evolution of the Western Carpathians (Slovakia). *Tübingen Geowissenschaftliche Arbeiten, Reihe A,* **27,** 1-81.

SPERNER, B., IOANE, D. & LILLIE, R.J. 2004. Slab behaviour and its surface expression: new insights from gravity modelling in the SE-Carpathians. *Tectonophysics*, **382**, 51-84.

SPERNER, B., RATSBACHER, L. & NŮMCOK, M. 2002. Interplay between subduction retreat and lateral extrusion: Tectonics of the Western Carpathoans. *Tectonics*, **21**, 1051-1075.

SPÖTL, C. & HASENHÜTTL, C. 1998: Thermal history of the evaporatic Haselgebirge melange in the Northern Calcareous Alps (Austria). *Geologische Rundschau* **87**, 449-460.

STAMPFLI, G.M. 1993. Le Briançonnais, terrain exotique dans les Alpes? *Eclogae Geologicae Helvetiae*, **86**, 1-45.

STAMPFLI, G.M., MOSAR, J., MARQUER, D., MARCHANT, R., BAUDIN, T. & BOREL, G. 1998. Subduction and obduction processes in the Swiss Alps. *Tectonophysics* **296**, 159–204.

STANISZEWSKA, A. & CIBOROWSKI, T. 2000. Lower Cretaceous breccia from autochthonous High-Tatric Succesion in Western Tatra Mts, southern Poland. *Przegląd Geologiczny*, **48**, 264-250. [*In Polish with English summary*]

STAREK, D., ANDREYEVA-GRIGOROVICH, A.S. & SOTÁK, J. 2000. Suprafan deposits of the Biely Potok Formation in the Orava region: sedimentary facies and nannoplankton Distribution. *Slovak Geological Magazine*, **6**, 188-190

STEWART, S.A & ARGENT, J.D. 2000. Relationship between polarity of extensional fault arrays and presence of detachments. *Journal of Structural Geology*, **22**, 693-711.

STRUZIK, A., ZATTIN, M. & ANCZKIEWICZ, R. 2002. Apatite fission track analyses from the Polish Western Carpathians. *Geolines,* **14**, 87-89.

SUE, C. & TRICART, P. 2003. Neogene to ongoing normal faulting in the Central Western Alps: A major evolution of the late Alpine tectonics. *Tectonics*, **22**, 105

SZAFIÁN, P., HORVÁTH, F., CLOETINGH, S. 1997. Gravity constraints on the crustal structure and slab evolution along a transcarpathian transect. *Tectonophysics*, **272**, 233-247.

SZULC, J. 2000. Middle Triassic evolution of the northern Peri-Tethys area as influenced by early opening of the Tethys Ocean. *Annales Societatis Geologorum Poloniae*, **70**, 1-48.

ŚWIDROWSKA, J. & HAKENBERG, M. 2000. Palaeotectonic conditions of Cretaceous basin development in the southeastern segment of the Mid-Polish Trough. *In*: CRASQUIN-SOLEAU S. & BARRIER É. (*Eds*), *Mémoires du Museum National d'Histoire Naturelle*, **182**, 239-256.

TARI, G., BÁLDI, T. & BÁLDI-BEKE, M. 1993. Paleogene retroarc flexural basin beneath the Neogene Pannonian Basin: a geodynamic model. *Tectonophysics*, **226**, 455.

TOKARSKI, A. 1978. Orogenesis and morphogenesis of Outer Carpathians and plate tectonics. *Studia Geomorphologica Carpatho-Balcanica*, **12**, 29-43.

TOKARSKI, A. & ŚWIERCZEWSKA, A. 2003. Miocene to Holocene tectonic activity of the Magura Nappe – ongoing orogeny? VI Karpackie Warsztaty Tektoniczne, 28.02-01.03.2003, Kraków, 40-41.

TOMEK, Č. 1993. Deep crustal structure beneath the Central and Inner West Carpathians. *Tectonophysics*, **226**, 417-431.

TOMEK, Č. & HALL, J. 1993. Subducted continental margin

imaged in the Carpathians of Czechoslovakia. *Geology*, **21**, 535-538.

Turnau-Morawska M. 1953. Kajper tatrzański, jego petrografia i sedymentologia. *Acta Geologica Polonica*, **3**, 33-102.

Uchman, J. 1973. Results of deep seismic soundings along international Profil V. *Publications of the Institute of Geophysics Polish Academy of Sciences*, **60**, 47-52.

Uhlig, V. 1899. Die Geologie des Tatragebirges. Theil II-IV. *Denkschriften Mathematisch-Naturwissenschafthichen Classe, Academie der Wissenschaften, Wien*, **68**, 43-140.

— 1905. Atlas geologiczny Galicyi, Szczawnica (1:75 000 scale), Zeszyt **24**. *Wydawnictwo Komisji Fizjograficznej Akademii Umiejętności*; Kraków.

Umhoefer, P.J. 2000. Where are the missing faults in translated terranes? *Tectonophysics,* **326**, 23–35.

Van Couvering, J.A., Aubry, M.P., Berggren, W.W., Bujak, J.P., Naeser, C.W. & Wieser, T. 1981. The terminal Eocene event and the Polish connection. *Palaeogeography, Palaeoclimatology, Palaeoecology,* **36**, 321-362.

Vauchez, A., Tomasi, A. & Barruol, G. 1998. Rheological heterogeneity, mechanical anisotropy and the deformation of the continental lithosphere. *Tectonophysics,* **296**, 61-86.

Wagreich, W. 1995. Subduction tectonic erosion and Late Cretaceous subsidence along the northern Austroalpine margin (Eastern Alps, Austria). *Tectonophysics*, **242**, 63-78.

Warren, J. 1999. Evaporites. Their evolution and economics, 422 pp. *Blackwell Science*; London – Tokio.

Weissert, H. J. & Bernoulli, D. 1985. A transform margin in the Mesozoic Tethys: evidence from the Swiss Alps. *Geologische Rundschau*, **74**, 665-679.

Westwalewicz-Mogilska, E. 1986. Nowe spojrzenie na genezę osadów fliszu podhalańskiego. *Przegląd Geologiczny*, **12**, 690-698. [*In Polish with English summary*]

Wieczorek, J. 1989. Model Hecho dla fliszu podhalańskiego? *Przegląd Geologiczny,* **9**, 419- 423.

Wiejacz, P. 1994. An attempt to determine tectonic stress patterns in Poland. *Acta Geophysica Polonica,* **3**, 169-176.

Wierzbowski, A. 1994. Late Middle Jurassic to earliest Cretaceous stratigraphy and microfacies of the Czorsztyn Succesion in the Spisz area, Pieniny Klippen Belt, Poland. *Acta Geologica Polonica*, **44**, 223-249.

Wierzbowski, A., Jaworska, M. & Krobicki, M. 1999. Jurassic (upper Bajocian-lowest Oxfordian) ammonitico rosso facies in the Pieniny Klippen Belt, Carpathians, Poland: its, fauna, age, microfacies and sedimentary environment. *Studia Geologica Polonica*, **115**, 7-74.

Wieser, T. 1973. Petrographical characteristics of the Podhale flysch in the Zakopane IG 1 borehole. *Biuletyn Instytutu Geologicznego*, **265**, 135-156. [*In Polish with English summary*]

Winkler, W. & Ślączka, A. 1994. A Late Cretaceous to Paleogene geodynamic model for the Western Carpathians in Poland. *Geologica Carpathica*, **45**, 71-82.

Wojciechowski, J. 1955. Ore veins near Szczawnica, Pieniny Mts. – Western Carpathians. *Biuletyn Instytutu Geologicznego*, **101**, 1-82. [*In Polish with English summary*]

Wojtal, S.F. & Mitra, G. 1986. Strain hardening and strain softening in fault zones from foreland thrusts. *Geological Society of America Bulletin*, **97**, 674–687.

Wojtal, S.F. & Pershing, J. 1991. Paleostresses associated with faults of large offset. *Journal of Structural Geology* **13**, 49–62.

Youssef, M.M. 1978. Large scale geological survey of the Mt Wżar andesites, Pieniny Mts, Poland. *Studia Geologica Polonica*, **66**, 1-30.

Zeyen, H., Dererova J. & Bielik M. 2002. Determination of the continental lithospheric thermal structure in the Western Carpathians: integrated modelling of surface heat flow, gravity anomalies and topography. *Physics of the Earth and Planetary Interiors*, **134**, 89-104.

Ziegler, P.A. & Cloetingh S. 2004. Dynamic processes controlling evolution of rifted basins. *Earth Science Reviews,* **64**, 1–50.

Zuchiewicz, W. 1995. Selected aspects of neotectonics of the Polish Carpathians. *Folia Qaternaria* **66**, 145-204.

— 1998. Quaternary tectonics of the Outer West Carpathians. *Tectonophysics*, **297**, 121-132.

Żaba, J. 1996. Late Carboniferous strikeslip activity at the boundary zone of Upper Silesia and Małopolska Blocks. *Przegląd Geologiczny*, **44**, 173-180. [*In Polish*]