Annales Societatis Geologorum Poloniae (2005), vol. 75: 211–248.

THE ORAVA DEEP DRILLING PROJECT AND POST-PALAEOGENE TECTONICS OF THE NORTHERN CARPATHIANS

Jan GOLONKA[1], Paweł ALEKSANDROWSKI[2], Roman AUBRECHT[3], Józef CHOWANIEC[4], Monika CHRUSTEK[1], Marek CIESZKOWSKI[5], Radosław FLOREK[6], Aleksandra GAWĘDA[7], Marek JAROSIŃSKI[8], Beata KĘPIŃSKA[9], Michał KROBICKI[1], Jerzy LEFELD[10], Marek LEWANDOWSKI[10, 14], František MARKO[3], Marek MICHALIK[5], Nestor OSZCZYPKO[5], Frank PICHA[11], Michal POTFAJ[12], Ewa SŁABY[13], Andrzej ŚLĄCZKA[5], Michał STEFANIUK[1], Alfred UCHMAN[5] & Andrzej ŻELAŹNIEWICZ[10]

[1] AGH University of Science and Technology, Al. Mickiewicza 30, 30-059 Kraków, Poland; e-mail: jan_golonka@yahoo.com
[2] University of Wrocław, Institute of Geological Sciences, ul. Cybulskiego 30, 50-205 Wrocław, Poland
[3] Department of Geology and Palaeontology, Comenius University, Mlynská dolina, 842 15 Bratislava, Slovak Republic
[4] Polish Geological Institute, Carpathian Branch, ul. Skrzatów 1, 31-560 Kraków, Poland
[5] Jagiellonian University, Institute of Geological Sciences, ul. Oleandry 2a, 30-063 Kraków, Poland
[6] Polish Oil and Gas Company, ul. Lubicz 25, 31-503 Kraków, Poland
[7] Silesian University, ul. Będzińska 60, 41-200 Sosnowiec, Poland
[8] Polish Geological Institute, ul. Rakowiecka 4, 00-975 Warszawa, Poland
[9] Mineral and Energy Economy Research Institute, ul. Wybickiego 7, 30-950 Warszawa, Poland
[10] Institute of Geological Sciences, Polish Academy of Sciences, ul. Twarda 51/55, 00-818 Warszawa, Poland
[11] 650 New Haven Court, 945 98 California, U.S.A.
[12] Geological Survey of Slovak Republic, Geologický Ústav Dionyza Štura, Mlynská dolina 1, 817 04 Bratislava, Slovak Republic
[13] University of Warsaw, Department of Geology, Al. Żwirki i Wigury 93, 02-089 Warszawa, Poland
[14] Institute of Geophysics, Polish Academy of Sciences, ul. Księcia Janusza 64, 01-452 Warszawa, Poland

Golonka, J., Aleksandrowski, P., Aubrecht, R., Chowaniec, J., Chrustek, M., Cieszkowski, M., Florek, R., Gawęda, A., Jarosiński, M., Kępińska, B., Krobicki, M., Lefeld, J., Lewandowski, M., Marko, F., Michalik, M., Oszczypko, N., Picha, F., Potfaj, M., Słaby, E., Ślączka, A., Stefaniuk, M., Uchman, A. & Żelaźniewicz, A., 2005. The Orava Deep Drilling Project and post-Palaeogene tectonics of the Northern Carpathians. *Annales Societatis Geologorum Poloniae*, 75: 211–248.

Abstract: This paper presents an insight into the geology of the area surrounding the ODDP proposed drilling site, and the structural development of the Carpathians in post-Palaeogene times. Since the deep drilling is proposed to be located in the Orava region of the Northern Carpathians, on the Polish-Slovak border, the structure and origin of the Neogene Orava Basin is also addressed in the paper.

The outline of geology of the Carpathian Mountains in Slovakia and Poland is presented. This outline includes the Inner Carpathian Tatra Mountains, the Inner Carpathian Palaeogene Basin, the Pieniny Klippen Belt, the Outer Carpathians, the deep structure below the Carpathian overthrust, the Orava Basin Neogene cover, the Neogene magmatism, faults and block rotations within the Inner and Outer Carpathians, and the Carpathian contemporary stress field.

The outline of geology is accompanied by the results of the most recent magnetotelluric survey and the detailed description of the post-Palaeogene plate tectonics of the circum-Carpathian region. The oblique collision of the Alcapa terrane with the North European plate led to the development of the accretionary wedge of the Outer Carpathians and foreland basin. The northward movement of the Alpine segment of the Carpathian-Alpine orogen had been stopped due to its collision with the Bohemian Massif. At the same time, the extruded Carpatho/Pannonian units were pushed to the open space, towards a bay of weak crust filled up by the Outer Carpathian flysch sediments. The separation of the Carpatho/Pannonian segment from the Alpine one and its propagation to the north was related to the development of the N–S dextral strike-slip faults. The formation of the West Carpathian thrusts was completed by the Miocene time. The thrust front was still progressing eastwards in the Eastern Carpathians. The Carpathian loop including the Pieniny Klippen Belt structure was formed. The Neogene evolution of the Carpathians resulted also in the formation of genetically different sedimentary basins. These basins were opened due to lithospheric extension, flexure, and strike-slip related processes. A possible asteno-sphere upwelling may have contributed to the origin of the Orava Basin, which represents a kind of a rift modified by strike-slip/pull-apart processes. In this way, a local extensional regime must have operated on a local scale in the Orava region, within the frame of an overall compressional stress field affecting the entire West Carpathians.

Nevertheless, many questions remain open. Without additional direct geological data, which can be achieved only by deep drilling under the Orava Deep Drilling Project, these questions cannot be fully and properly answered.

Key words: plate tectonics, neotectonics, Carpathians, Palaeogene, Neogene, continental deep drilling.

Manuscript received 29 March 2004, accepted 5 October 2005

INTRODUCTION

In December 1999 Poland was admitted to the International Continental Scientific Drilling Program (ICDP). In January 2003 the proposal to organize a workshop in Poland was submitted to ICDP. The aim of the workshop was to discuss the scientific value and strategy of the proposed deep drilling located in the Carpathian Mountains, on the Polish-Slovak border in Central Europe. This project was approved and the ICDP international workshop "*Orava Deep Drilling Project: Anatomy and evolution of the Europe/Africa collisional suture in a mantle plume-modified orogen*" was held on August 31- September 4, 2003 in Zakopane, Poland. The workshop was financed by ICDP and by the European Commission through the Polish Committee for Scientific Research and the Institute of Geophysics of the Polish Academy of Sciences. Sixty-six scientists from thirteen countries participated in this workshop. The discussion focused on identifying key knowledge gaps and project research goals in the following areas:

1. Structural position of the Carpathian - North European suture and its significance for the reconstruction of the Cenozoic Alpine system of Europe.
2. Relationship between the tectonic and geodynamic settings and magmatogenesis.
3. The nature of geophysical anomalies.
4. Geothermal issues.
5. Geodynamic reconstruction of the Mesozoic–Cenozoic basins.
6. Oil generation, migration and timing.
7. Regional heat-flow evolution.
8. Detection and studying of the Cadomian-Variscan basement.
9. Palaeostress evolution and its changes in horizontal and vertical sections.

The discussion highlighted, i.a. the question of the origin of the Orava Basin and the related issues of Cenozoic to Recent tectonic movements in the Carpathians. A special oral presentation of the planned goals of the ODDP was made during the 5th Conference "*Neotectonics of Poland*" in September 2003. This paper presents an insight into the geology of the area surrounding the ODDP proposed drilling site, and the structural development of the Carpathians in the post-Palaeogene time. Since the deep drilling is proposed to be located in the Orava region of the Carpathians, on the Polish-Slovak border, the structure and origin of the Neogene Orava Basin is also addressed in the paper.

OUTLINE OF GEOLOGY OF THE CARPATHIANS

The Carpathians define an extensive mountain arc, which stretches at a distance of more than 1,300 km, from Vienna area in Austria, to the Iron Gate on the Danube in Romania (Fig. 1). To the west, the Carpathians are linked with the Eastern Alps, whereas to the east they continue into the Balkan mountain chain. Traditionally, the Carpathians are subdivided into their western and eastern parts (e.g. Mahel', 1974). The West Carpathians consist of an older, internal orogenic zone known as the Inner or Central Carpathians, and the external, younger one, known as the Outer or Flysch Carpathians (e.g. Mahel', 1974; Książkiewicz, 1977; Ślączka & Kaminski, 1998; Ślączka *et al.*, 2005). The Inner Carpathians were folded during the Late Cretaceous and now are in a tectonic contact with the Outer Carpathian units across a transform fault zone represented by the Pieniny Klippen Belt (PKB) (Figs 2, 3, 4). An arcuate shape of the Carpathians is believed to be due to oroclinal bending, as indicated by palaeomagnetic data (cf. Kruczyk *et al.*, 1992)

Inner Carpathians close to the ODDP site

In the immediate vicinity of the Orava Basin, the Inner Carpathian Palaeozoic and Mesozoic rocks crop out in the Tatra Mountains. North of the Tatras, they are covered by the Central Carpathian Palaeogene (Figs 5, 6, 7) and known only from boreholes and geophysical data (Golonka & Lewandowski, eds., 2003)

Tatra Mountains

The Tatra Mts are the highest mountain range of the Carpathians, located in their western part. The Tatra Mts form an elevated asymmetric horst tilted northward, cut off from the south by a major Neogene–Quaternary normal fault (Gross *et al.*, 1993), and surrounded by sediments the Central Carpathian Palaeogene (Figs 5, 6). The uplift of the Tatras, dated using apatite fission tracks, took part probably during the Miocene (15–10 Ma) (Burchart, 1970; Kovač *et al.*, 1993).

The Tatra Mts consist of a crystalline core with an autochthonous Mesozoic sedimentary cover which are overlain by several thrust sheets and small nappes (Fig. 7). In the pre-thrusting depressions of the basement, the allochthonous units are imbricated, distinctly thicker, and more nu-

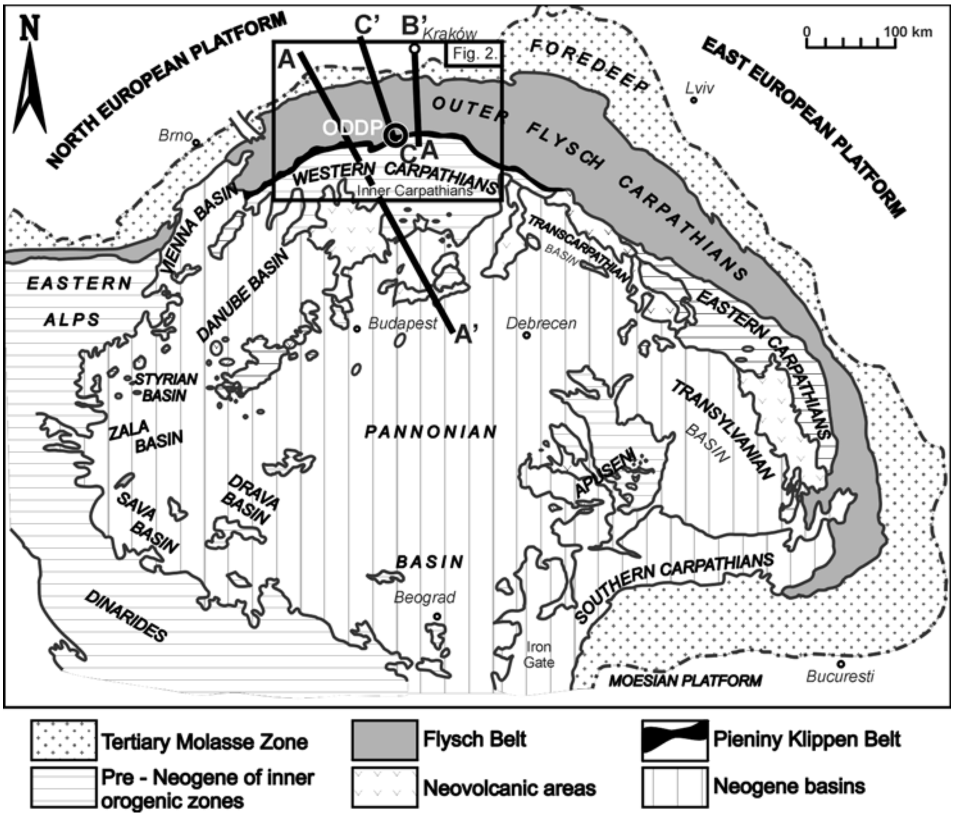

Fig. 1. Tectonic sketch map of the Alpine-Carpathian-Pannonian-Dinaride basin system (after Kováč *et al.*, 1998): A–A', B–B', C–C' – localization of cross-sections (Figs 3, 4, 14), ODDP – planned Orava Deep Drilling Project Well (International Continental Deep Drilling Program)

merous than those on relevant elevations. All these units are discordantly covered with a post-nappe transgressive succession of the Central Carpathian Palaeogene Basin. The crystalline basement consists of the Variscan polygenetic granitoid intrusion and the pre-Variscan to Variscan metamorphic envelope. The metamorphic complex experienced several tectono-metamorphic events, different on the southern and northern parts of the Tatra massif. The pre-Alpine crystalline basement exposed in the Tatra Mts is present also in several other units of the Carpathians. Rocks of the Tatra Mts were affected by successive Early Variscan to Alpine tectono-metamorphic events.

The northern part of the metamorphic envelope is composed of two units: the Upper Structural Unit and the underlying Lower Structural Unit, divided by a zone of ductile thrusting. Rocks of the Upper Structural Unit are generally migmatized; there occur both metapelitic gneisses and amphibolites (Burda & Gawęda, 1997; Burda & Gawęda, 1999; Gawęda *et al.*, 2000, 2001), metamorphosed in the upper amphibolite facies conditions. In turn, rocks of the Lower Structural Unit are metamorphosed in upper greenschist to lower amphibolite facies conditions (Gawęda *et al.*, 1998; Kozłowski & Gawęda, 1999). They all together form an inverted metamorphic zonation. The differences between both units are visible in mineral parageneses and chemistry as well as in the tectonic trends (Kozłowski & Gawęda, 1999).

The southern part of the metamorphic envelope shows also the inverted metamorphic zonation, but represents deeper portions of the crust. The Upper Structural Unit consists of migmatized paragneisses, orthogneisses and amphibolitized eclogites. Rocks of this unit were metamorphosed in the granulite facies, then subjected to regional hydratation and biotite blasthesis producing the tonalitic leucosomes, and then nearly isothermal decompression which caused cordierite formation and biotite dehydration-melting (Janak *et al.*, 1999). The intrusion of the Variscan Rohače granite led to migmatization in the surrounding metamorphic rocks in the southern envelope. The earliest granitoid body, at present orthogneiss, crops out in the westernmost part of the Tatra Mts, and the zircon dating suggest the 405 Ma age of intrusion with the superimposed metamorphism and gneisssification at 360 Ma (Poller *et al.*, 2001).

Granitoids of the Tatra Mts are represented by tonalities and granodiorites with subordinate amount of granites. Numerous secondary alterations (chloritization, albitization, sericitization, carbonatization, crystallization of epidote group minerals, breakdown of monazite) affected the granitoids (Michalik & Skublicki, 1999). Janak *et al.* (2001) consider amphibolitic lower crust as a source of granitic magma mixed with crustal melt generated in other sources. Structural data and interpretation of metamorphic processes indicate that the magma originated in a continental collision environment between 360 and 314 Ma (Poller *et al.*, 2001).

The Mesozoic sedimentary rocks (e.g., Kotański, 1979; Lefeld, 1985; Wieczorek, 2000) of the Tatra Mts belong to four main facies/palaeogeographic zones, i.e. to the High-Tatric and the Lower, Middle, and Upper Sub-Tatric zones

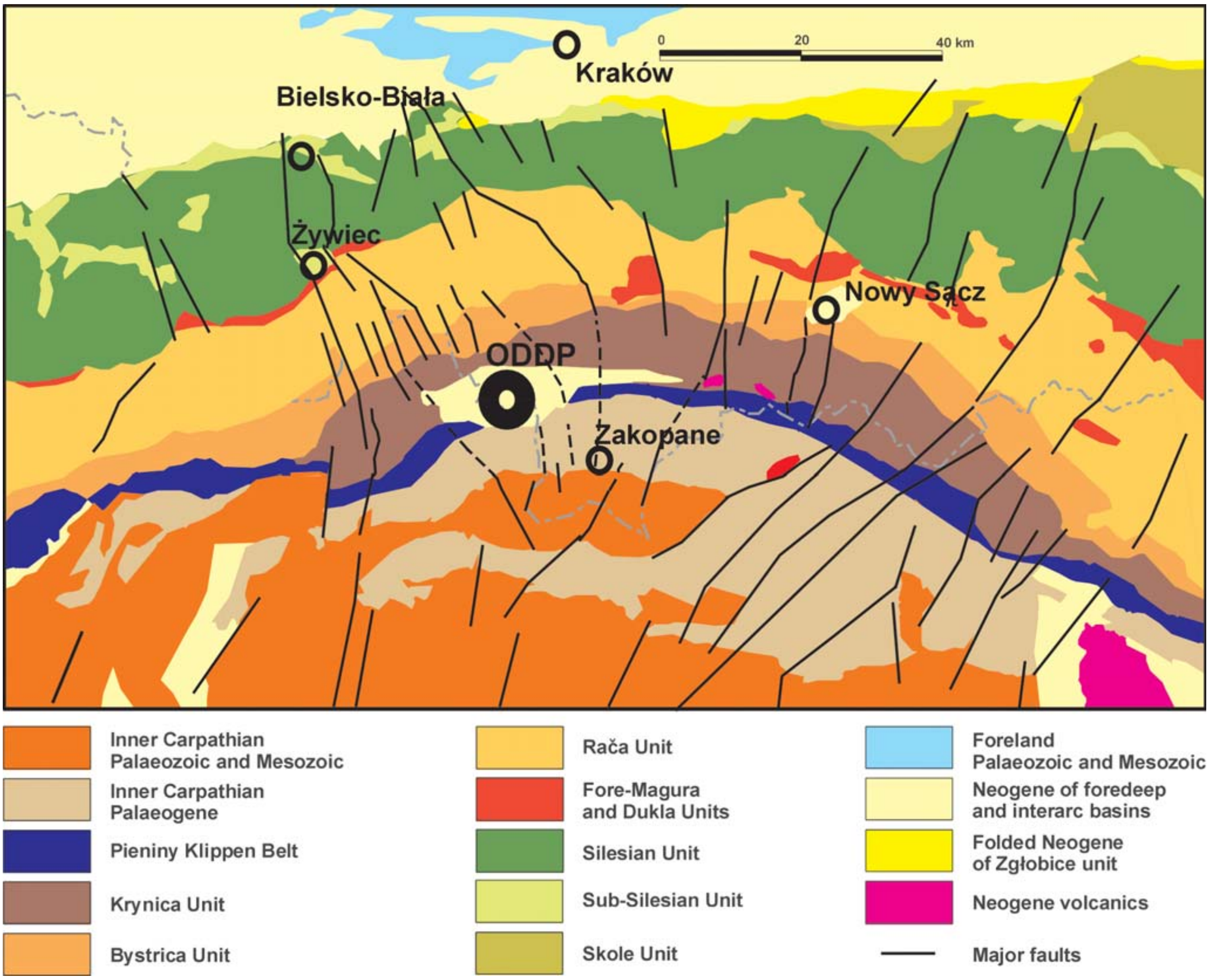

Fig. 2. Geological map of West Carpathians and adjacent areas with major West Carpathian faults perpendicular to the suture zones and general Outer Carpathian thrusts trends. Modified from Lexa *et al.* (2000) and Golonka *et al.* (2005)

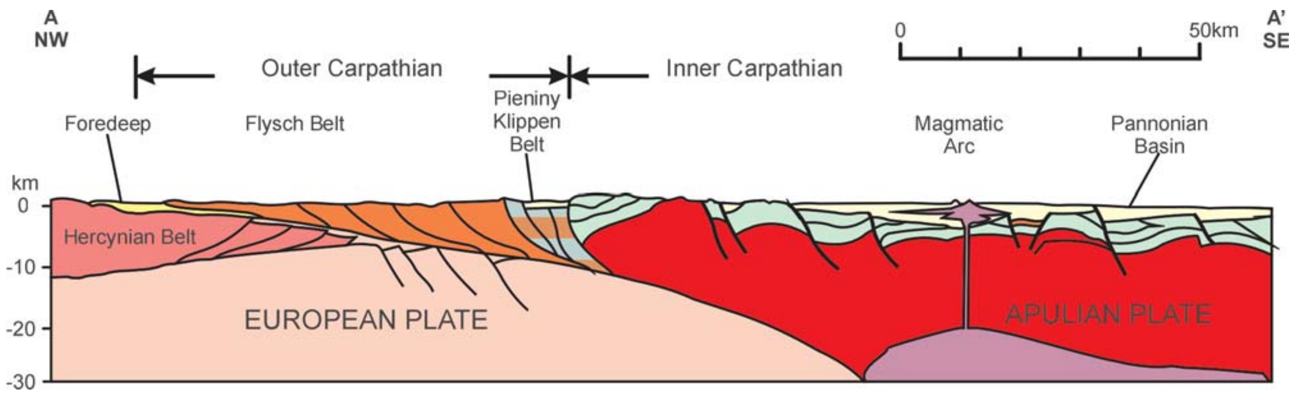

Fig. 3. Generalised cross-section A–A' across Carpathian-Pannonian region (after Picha, 1996). Cross-section location on Fig. 1

(Fig. 8). The High-Tatric zone (Kotański, 1961) includes the sedimentary cover of the crystalline core and the lower units of the overlying allochthon. Some of the allochthonous units contain also fragments of their crystalline (Lower Palaeozoic) basement overthrust together with sedimentary rocks. The High-Tatric zone and its basement form the lowermost structural element, i.e. the so-called Tatricum. The oldest rocks of the sedimentary cover are conglomerates, which crop out in a single locality (Koperšady), and are believed to be Permian in age. The High-Tatric zone is characterised by transitional Germanic-Alpine Triassic facies. The Lower Triassic is characterised by red bed sandstones, fol-

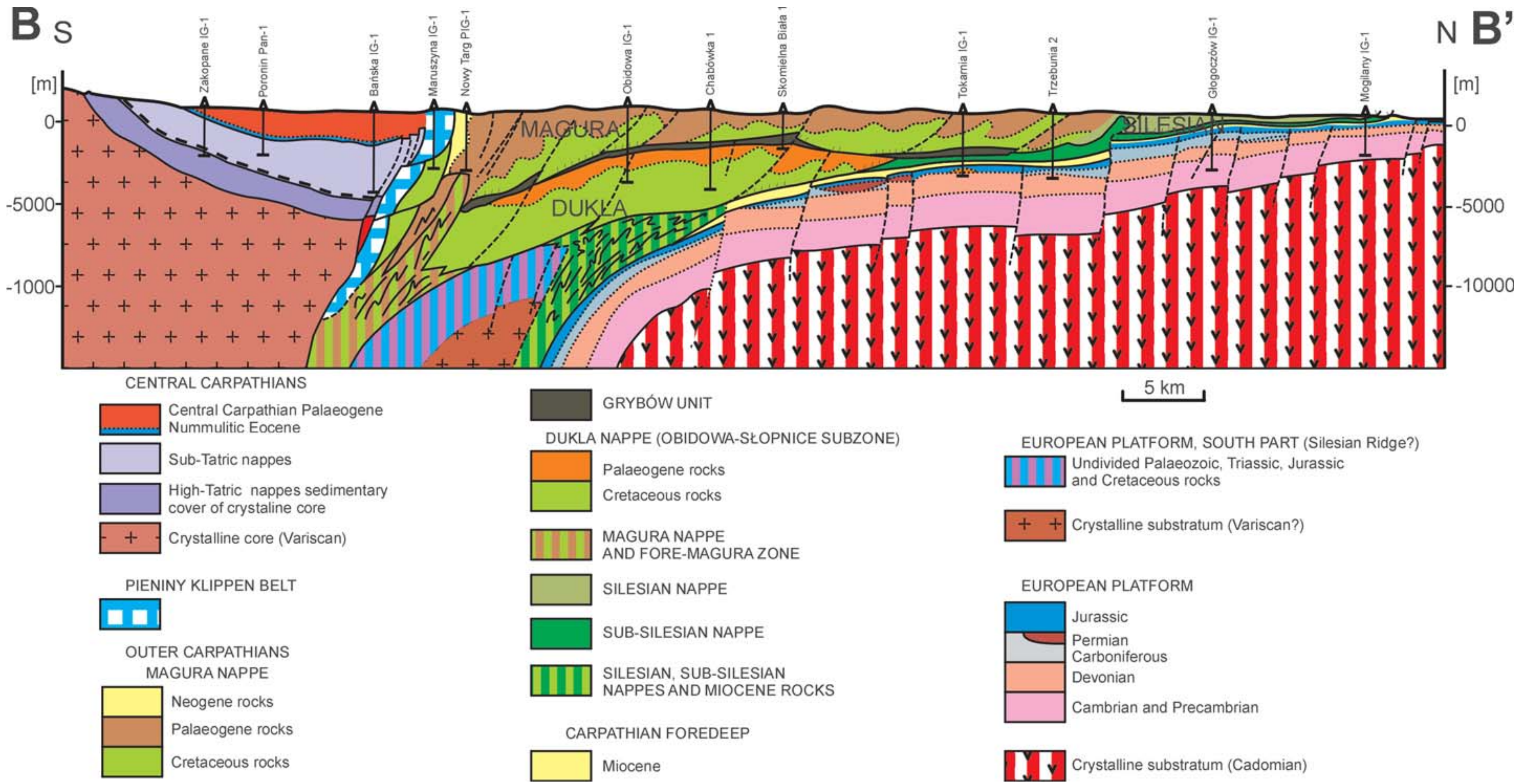

Fig. 4. Detail cross-section B–B' through the West Carpathians between the Inner Carpathian Tatra Mountains and the Carpathian Foredeep (Zakopane-Kraków line, based on multiple sources). Cross-section location on Fig. 1

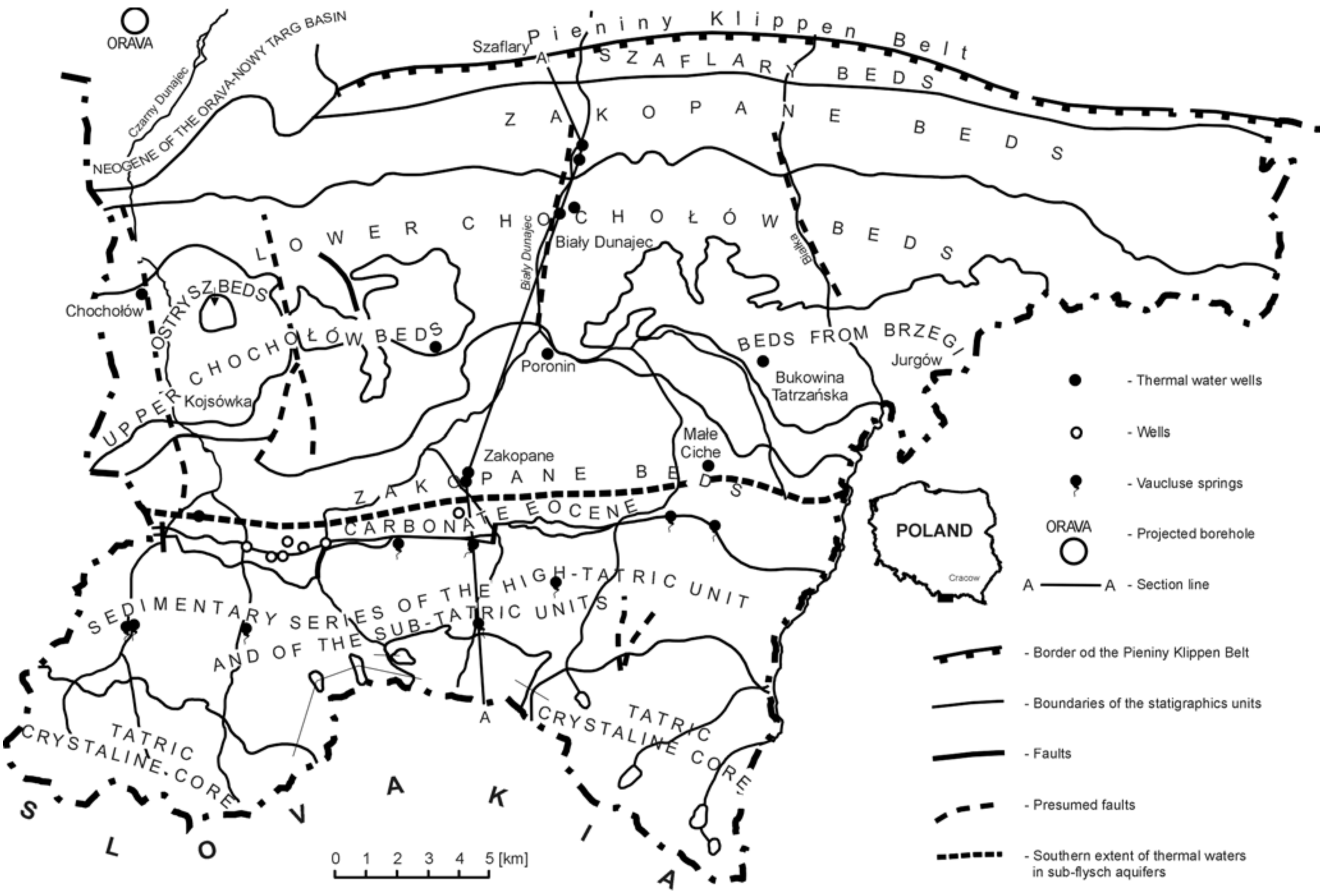

Fig. 5. Geological map of the Podhale and the Tatra Mts. (after Chowaniec & Kępińska, 2003)

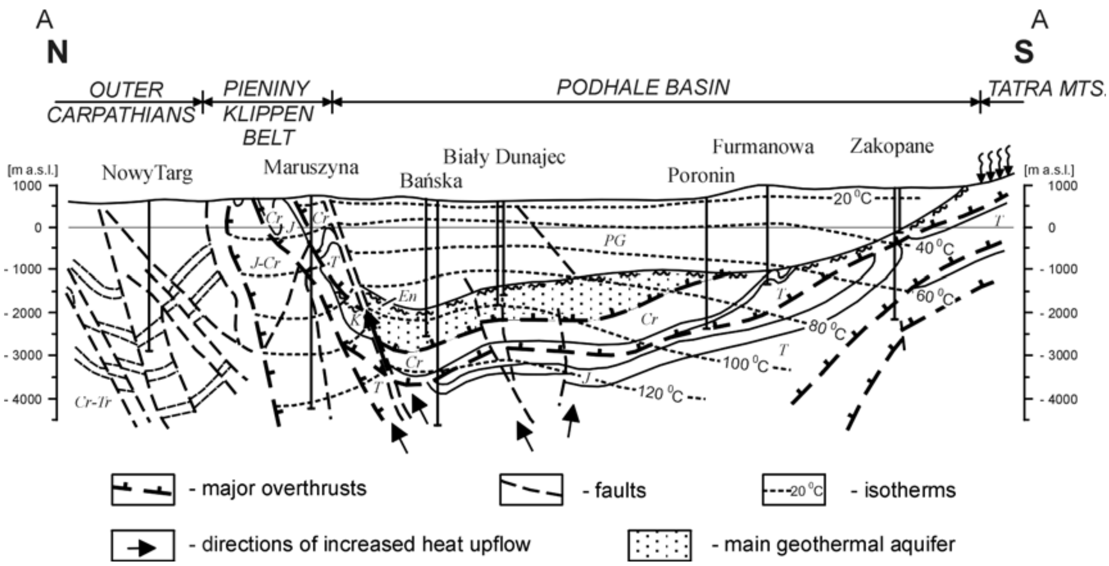

Fig. 6. Geological-thermal cross-section through the Podhale geothermal system; location: Fig. 6 (after Chowaniec & Kępińska, 2003). PG – Podhale Flysch, En – Middle Eocene Nummulitic carbonates, Tr – Tertiary, Cr – Cretaceous; J – Jurassic; T – Triassic

lowed by mudstones (Dżułyński & Gradziński, 1960; Roniewicz, 1966; Fejdiová, 1971; Mader, 1982), which are overlain by the Middle Triassic platform limestones and dolomites. The Upper Triassic is preserved only in the autochthonous cover and is very differentiated. It contains red beds of the Keuper type and coeval intertidal laminated dolomites, Rhaetian clastics, and shallow-marine fossiliferous limestones (Kotański, 1959, 1979). Lower Jurassic clastics and limestones occur in local troughs formed by block tectonics related to the rifting in the Western Tethys. The Middle Jurassic contains local crinoidal limestones, nodular limestones, commonly with stratigraphic gaps and conden-

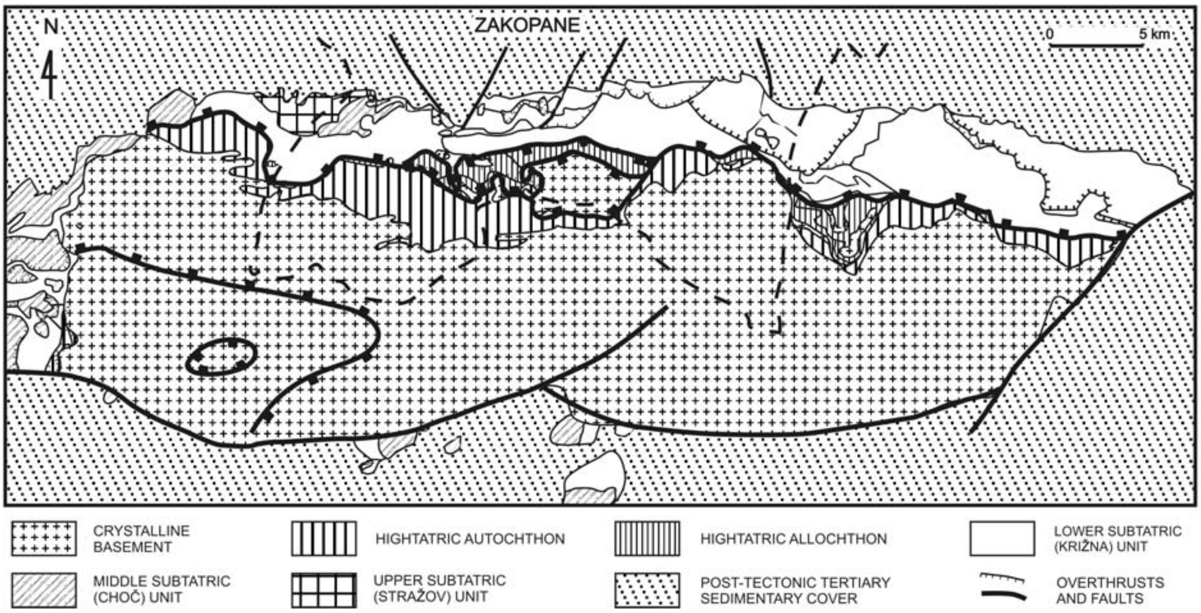

Fig. 7. Tectonic sketch map of the Tatra Mountains

sations, stromatolites, and iron crusts. The allochthonous High-Tatric units are typified by common stratigraphic gaps embracing the Upper Triassic–Lower Jurassic. The Upper Jurassic pelagic limestones display locally nodular structures. Locally (Osobitá Mt), shallow-marine crinoidal limestones with volcanic rocks (limburgite) occur. Shallow-marine platform limestones of the Urgonian-type (Schrattenkalk) facies typify the Lower Cretaceous. The Lower-Middle Albian occurs only locally as condensed deposits with glauconite and phosphates, indicating drowning of carbonate platform. During the terminal basin development (Middle Albian–Cenomanian), marls with turbidites indicate a deepening of facies (Lefeld, 1985).

Rocks of the Lower Sub-Tatric (Krížna), Middle Sub-Tatric (Choč), and the Upper Sub-Tatric (Strážov) units occur exclusively in thrust sheets, which overlie the High-Tatric units (Lefeld, 1999). The Lower Sub-Tatric units (part of the Fatricum structural elements) are characterised by transitional Germanic-Alpine Triassic facies, with the Lower Triassic red bed clastics, Middle Triassic platform dolomites, and the Carpathian Keuper and Rhaetian fossiliferous limestones in the Upper Triassic (Kotański, 1959, 1979). The Jurassic facies are characterised by gradual deepening from shallow marine-clastics, through spotty limestones and marlstones (Fleckenmergel facies), spiculites and radiolarites, to nodular and Maiolica limestones. Local shallowing is recorded in encrinites with manganese mineralisation (Toarcian) in the Western Tatra Mts. A condensation with large oncoids, iron crusts and stromatolites is present in the lower part of the Middle Jurassic sediments. Basinal marlstones and limestones dominated in the Lower Cretaceous sediments in the western part of the Tatra Mts, while in the eastern part (Belanské Tatry) massive carbonates occurred. The latter pinch out to the west (Lefeld, 1985; Bac-Moszaszwili, 1993; Wieczorek, 2000).

The Middle Sub-Tatric units (part of the Hronicum structural elements) comprise typical Alpine Triassic facies, including the Hauptdolomit, the Rhaetian Kössen facies, and the Lower Jurassic encrinites, spiculites, and Hierlatz-type limestones (Grabowski, 1967; Kotański, 1973; Iwanow & Wieczorek, 1987; Uchman, 1993). The Upper Sub-Tatric units (included also to the Hronicum), represented only by two small thrust sheets, are typified by the basinal Middle Triassic Reifling Limestone, Partnach Marl, and the shallow-marine Upper Triassic Wetterstein facies. All of the allochthonous units were thrust northward in the Late Cretaceous (Kotański, 1986a, b; Iwanow & Wieczorek, 1987; Jurewicz, 2002).

Central Carpathian Palaeogene Basin

The Podhale region is the northern part of the Inner Carpathian Palaeogene Basin, which includes Liptov, Orava and Spiš depressions (Figs 2, 5). It is built up of Palaeogene strata underlain by mostly calcareous Mesozoic rocks. The lithostratigraphic section of these deposits has been recognised by several deep boreholes (Figs 4, 5).

In subsequent years, the boreholes which were located as shown on Figure 5 provided very advantageous information. The results of investigations showed that the sub-Palaeogene substratum is an extension of geological-structural elements of the Tatra massif, to which the Sub-Tatric (Krížna, Choč) and High-Tatric nappes belong. Moreover, in logs of some deep drillings (Sokołowski, 1973; Chowaniec, 1989) the facies elements similar to certain rock types of the Pieniny series and deposits of uncertain affinity were found. After the retreat of the Late Cretaceous sea, a subsequent transgression took place in the Middle Eocene that resulted in the formation of conglomerates and limestones in the initial phase. These deposits form the basal member of the Podhale Palaeogene. Then, typical flysch de-

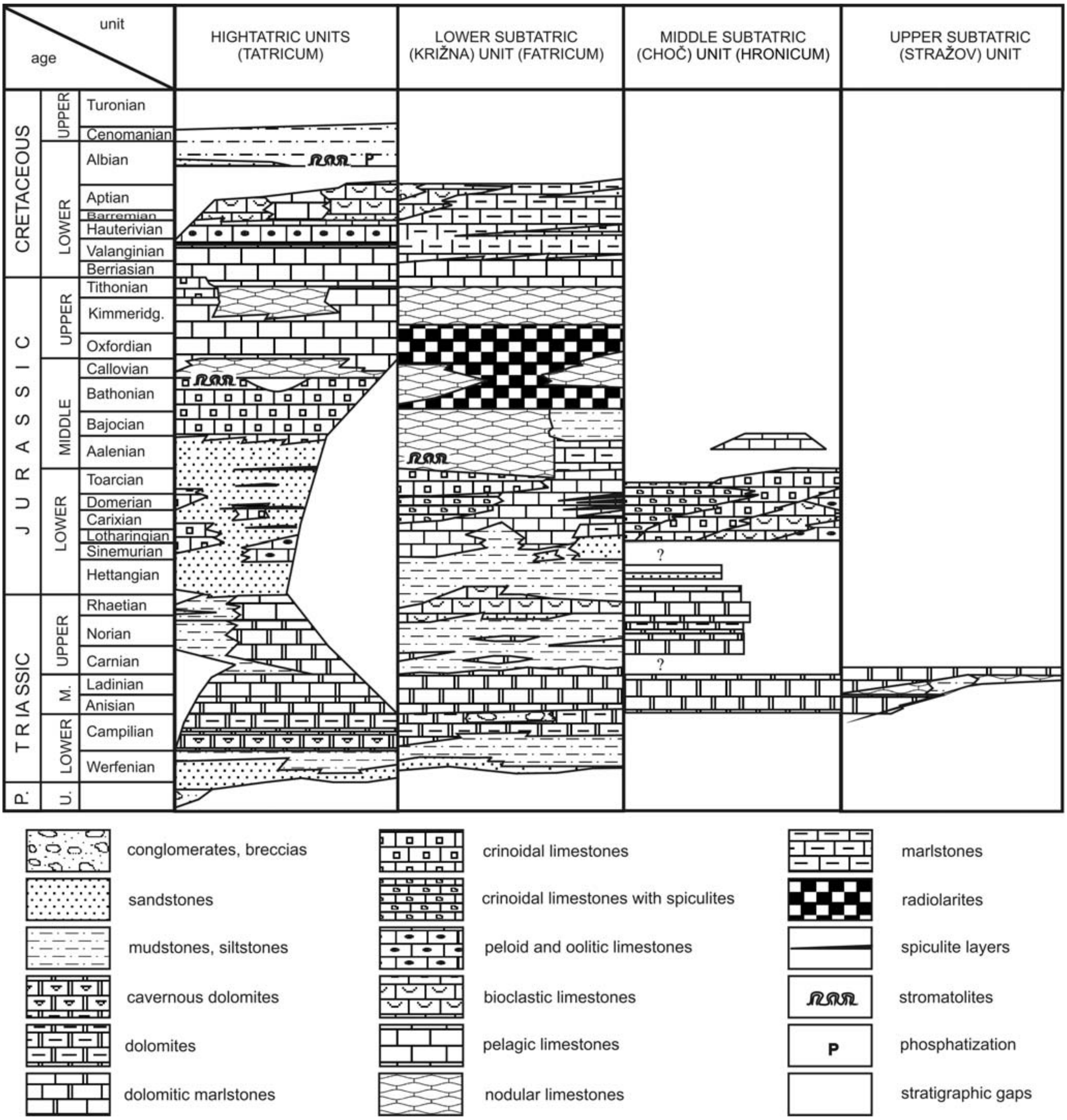

Fig. 8. Stratigraphic scheme of the Mesozoic sedimentary rocks of the Tatra Mountains

posits were formed. Sediments of the calcareous Eocene are known from numerous natural exposures situated at the outlets of valleys draining the Tatra massif and from drillings made in the Podhale Basin. Directly on the transgressive deposits of the calcareous Eocene there occur stratigraphically younger strata of the Palaeogene, i.e. the Podhale flysch. The largest thickness of the latter, ca. 3,000 m, was recorded in borehole Chochołów PIG-1. In the Slovak Orava, the Eocene sequence is known as the Podtatranská Group (Gross *et al.*, 1984, 1993), which is an equivalent of the Podhale flysch in Poland. The basal Borové Formation lies transgressively on the Mesozoic cover of the Malá Fatra, Tatra, and Choč Mts. The lithology of this formation is variable, being strongly dependent on the character of the substratum upon which it was deposited. It is composed of breccias, sandstones, and carbonates, sporadically with large foraminifers. The thickness of the entire formation varies from few centimetres to several tens of metres.

The Szaflary beds, occurring in the northern part of the basin, are generally assigned to the oldest flysch members (Kępińska, 1997; Chowaniec & Kępińska, 2003). Shaly flysch strata of the Zakopane beds, in turn, belong to the younger members. The Slovak equivalent of the Zakopane beds is the Huty Formation, which comprises mainly pelitic, sandstone (also some breccias) strata, only few centimetres thick, in contrast to several tens of centimetres to metres thick clayey beds. Localities exposing dark-brownish Menilite-like silty claystones occur at several places within the Huty Formation.

The Zuberec or Chochołów Formation, overlying the Zakopane beds, is of typical flysch facies, with variable sandstone/shale ratio. Fine-grained breccias and even slumped conglomerates are common. Submarine slumps are of sandy matrix with dispersed clasts of sandstones, siltstones, claystones, and limestones. The total thickness of this formation is about 500–900 m, and its age was determined as the Late Eocene to Early Oligocene (Gross *et al.*, 1993).

The uppermost formation in the Podtatranská Group in Slovakia is the Biely Potok Formation. Its equivalent in Poland is known as the Ostrysz beds, forming the culmination of Ostrysz Mt in the western Podhale. This formation consists of coarse-grained sandstones and subordinate claystone strata. The sandstones are mainly siliciclastic with clayey matrix, bearing only small percentage of carbonates. The thickness of the Biely Potok Formation is up to 700 m, and its age is Late Oligocene. At few places, the Pucov Conglomerates occur (e.g. south of the Oravský Podzámok). This member consists of blocky conglomerates bearing various Mesozoic carbonate clasts cemented with reddish sandy-pelitic matrix. Longitudinal, narrow bodies of conglomerates are incised into the Zuberec and Huty formations, and also to the Mesozoic substratum. The Pucov Conglomerate is interpreted as a channel fill supplied from the nearby southern source. The thickness of 170 m was documented by a borehole log (Gross *et al.*, 1993).

Janočko and Kováč (2003; see also references therein) suggested that the initial evolutionary stage of the basin was due to oblique convergence during the retreat of subduction boundary, which resulted in compressional regime in front of the advancing upper plate and extension in the plate's inner part. The opening of the Central Palaeogene Basin was related to this extension. The front of the Inner Carpathian plate served as a source area for sedimentation of the Szaflary beds. In the presently narrow zone along the Pieniny Klippen Belt, the Palaeogene strata were deformed into slices and folds. According to Marschalko (1968), the wide area (15–20 km?) of the northern rim of the Inner Carpathian Palaeogene is missing. It is that part, where the lateral input of the material from the source had produced proximal sediments. One has to bear in mind, however, that palaeomagnetic data point to significant (70–110°) counterclockwise rotations within the Carpathian Palaeogene Basin (see, e.g. Grabowski & Nemčok, 1999; Márton, 2003; Csontos & Vörös, 2004; Golonka, 2005). According to Golonka *et al.* (2005), an analysis of exotic clasts supports this rotation, indicating that neither the present-day Tatricum nor the Sub-Tatric (Križna and Choč) nappes were source areas for the Pieniny and Magura flysch during Palaeogene time. This question requires new independent research, but nevertheless it calls for a substantial correction in estimations of the genuine palaeogeographic location of alimentary areas. Perhaps the missing part of the Central Palaeogene Basin is located somewhere within the Tisza plate. The Transylvanian Palaeogene (e.g., Sandulescu *et al.*, 1981; Ciulavu & Bertotti; 1994; Meszaros, 1996) and certain parts of the Szolnok flysch Palaeogene sequences, situated in the marginal part of the Tisza unit (e.g. Nagymarosi & Báldi-Beke, 1993), display similarities with the Central Carpathian Palaeogene. This problem requires further investigations. In the Neogene, the Inner Carpathian plate rotation wiper effect led to significant deformation along the plate boundary, which resulted in a complex tectonic pattern along the present-day boundary between the Central Carpathian Palaeogene and the Pieniny Klippen Belt. At the same time, the Tatric horst was formed leading to tilting of Palaeogene strata from their initial position. An uplift of the Tatric massif brought about formation of fissures and cracks, as well as local folds and faults (sometimes of a regional extent) which are rooted in the Mesozoic rocks. The most important are the Jurgów-Trybsz, Biały and Czarny Dunajec, and Krowiarki (Prosečno) faults (Fig. 5). The present-day Podhale Basin is an asymmetric basin, delimited by the Tatras in the south, and by a steep fault along the Pieniny Klippen Belt in the north. According to Soták and Janočko (2001), the structural pattern of the Central Carpathian Palaeogene Basin includes basement-involving fault zones, like the Margecany and Muran faults. Extensional features, like half-grabens and listric and antitethic faults are to be found in the Hornad, Periklippen and Poprad Depressions, while structures related to retro-wedge thrusting, transform faulting, and strike-slip tectonics occur in the Šambron Zone (Mastella *et al.*, 1988; Kovač & Hók, 1993; Ratschbacher *et al.*, 1993; Nemčok & Nemčok, 1994; Marko, 1996; Sperner, 1996, 2002; Plašienka *et al.*, 1997; Soták & Janočko, 2001; Janočko & Kováč, 2003). Soták and Janočko (2001) assumed the following stages of basin development:

– initial faulting and alluvial fan deposition in an half-graben basin;

– carbonate factory on a shelf-margin basin;

– glacio-eustatic regression and semi-isolation in a restricted basin;

– progressive faulting and fault-controlled accumulation of radial fans in a tilted basin;

– highstand aggradation in a starved basin;

– Mid-Oligocene sea level lowering and retroarc backstepping of depocenters in a relic basin; and

– wedging of fans in an over-supplied basin.

The above-mentioned counterclockwise rotation of the Alcapa plate was compensated by dextral shearing in a transpressional zone between the Alcapa and North European plates (Ratschbacher *et al.*, 1993; Soták & Janočko, 2001). The present-day northern boundary was caused by amputation by a transform fault related to this rotation.

Pieniny Klippen Belt

The Pieniny Klippen Belt (PKB) is composed of several successions of mainly deep and shallow-water limestones, covering a time span from the Early Jurassic to Late Cretaceous (Andrusov, 1938, 1959; Andrusov *et al.*, 1973; Birkenmajer, 1958, 1977, 1986, 1988; Mišík, 1994; Golonka & Krobicki, 2001, 2004). This strongly tectonized structure is a terrain about 600 km long and 1–20 km wide, which stretches from Vienna in the west, to Romania in the east (Figs 1–4). The PKB is separated from the present-day Outer Carpathians by a Miocene subvertical strike-slip fault.

During the Jurassic and Cretaceous, the submarine Czorsztyn Ridge and surrounding zones within the Pieniny Klippen Belt Basin (PKBB) formed an elongated structure with domination of pelagic type of sedimentation (Birkenmajer, 1977, 1986; Mišík, 1994; Aubrecht *et al.*, 1997; Plašienka, 1999; Wierzbowski *et al.*, 1999; Golonka & Krobicki, 2001, 2004). The orientation of the PKBB was SW–NE (see, for instance, Aubrecht & Túnyi, 2001; Golonka & Krobicki, 2001, 2004). Its deepest part is documented by extremely deep-water Jurassic–Early Cretaceous pelagic limestones and radiolarites (Golonka & Sikora, 1981). Somewhat shallower sedimentary zones known as the Pieniny and Branisko (Kysuca) successions were located close to the central furrow. Transitional slope sequences between the deepest basinal units and ridge units are known as the Niedzica and Czertezik successions (Podbiel and Pruské successions), near the northern Czorsztyn Ridge, and the Haligovce-Nižná successions near the southern Exotic Andrusov Ridge (Birkenmajer, 1977, 1986, 1988; Aubrecht *et al.*, 1997). The strongly condensed Jurassic–Early Cretaceous pelagic cherty limestones (Maiolica-type facies) and radiolarites were also deposited in the northwestern (Magura) basin.

Generally, the Pieniny Klippen Belt Basin sedimentary history is tripartite (1–3): from (1) oxygen-reduced dark/black terrigenous deposits of the Early-early Middle Jurassic age (Fleckenkalk/Fleckenmergel facies), through (2) Middle Jurassic–earliest Cretaceous crinoidal, nodular (of the Ammonitico Rosso type) or cherty (of the Maiolica= Biancone type) limestones and radiolarites, up to (3) the Late Cretaceous pelagic marls (i.e. Scaglia Rossa = Couches Rouge = Capas Rojas) facies and/or flysch/flyschoidal series (Birkenmajer, 1986, 1988; Bąk, 2000; Golonka & Krobicki, 2004).

The oldest Jurassic rocks, known only from the Ukrainian and Slovak part of the PKB (e.g., Andrusov, 1938, 1959; Smirnov, 1973; Birkenmajer, 1977; Aubrecht *et al.*, 1997; Krobicki *et al.*, 2003; Golonka & Krobicki, 2004; and references therein), consist of different types of Gresten-like clastic sediments with intercalations of black fossiliferous limestones bearing brachiopods and grypheoids (Hettangian–Sinemurian). Spotty limestones and marls of oxygen-depleted, widespread Tethyan Fleckenkalk/ Fleckenmergel-type facies, and *Bositra* ("*Posidonia*") black shales with spherosiderites represent the Pliensbachian–Lower Bajocian (Birkenmajer, 1986; Golonka & Krobicki, 2004; and references therein). One of the most rapid changes of sedimentation/palaeoenvironments within this basin took place from the late Early Bajocian, when well-oxygenated multicoloured crinoidal limestones replaced in some zones dark and black sediments of the Early-early Middle Jurassic age (Birkenmajer, 1986; Aubrecht *et al.*, 1997; Wierzbowski *et al.*, 1999; Golonka & Krobicki, 2004). Sedimentation of younger (since the latest Bajocian), red, nodular Ammonitico Rosso-type limestones was an effect of the Meso-Cimmerian vertical movements which subsided the Czorsztyn Ridge (Golonka *et al.*, 2003). The Late Jurassic (Oxfordian–Kimmeridgian) history of the PKB reflects the strongest facies differentiation within sedimentary basin where mixed siliceous (radiolarites)-carbonate sedimentation took place. This may be at least partly attributed to radical and fast palaeogeographic evolution of the eastern segment of the Pieniny Basin, as indicated by recent palaeomagnetic results (Lewandowski *et al.*, 2003). The Upper Cretaceous pelagic deposits were dominated by Scaglia Rossa-type marls deposited during the latest, third episode of evolution of the Pieniny Klippen Belt Basin, when unification of sedimentary facies took place within all successions (Birkenmajer, 1977, 1986). Later flysch and/or flyschoidal deposition with several episodes of debris flows involving numerous exotic pebbles took place. The main "exotic source area" in the PKB was the so-called Exotic Andrusov Ridge – part of the Inner Carpathian plate uplifted during Albian–Late Cretaceous time (Birkenmajer, 1988; Golonka *et al.*, 2003).

The Pieniny Klippen Belt Basin was closed at the Cretaceous/Palaeogene transition as an effect of strong Late Cretaceous (Subhercynian and Laramian) thrust-folding (Birkenmajer, 1977, 1986). Simultaneously with this Laramian nappe folding, the uppermost Cretaceous (Maastrichtian) and Palaeogene flysch and molasse-type rocks with exotic material were deposited. These rocks covered, frequently with unconformity, several earlier folded klippen nappes. The second tectonic episode was connected with the strong Savian and Styrian (Early and Middle Miocene, respectively) compression, when the Cretaceous nappes, new Paleocene deposits of the PKB, and part of the Magura basin were refolded together (Birkenmajer, 1986). The PKB was formed as a melange in the suture zone between the Inner Carpathian-Alpine (Alcapa) terrane and the North European plate (Fig. 4). Part of allochthonous Outer Carpathian units and, perhaps, fragments of the basement were also located in this suture zone. Finally, with the eastward movement of the Alcapa plate, a system of strike-slip faults originated (Birkenmajer, 1983). The Middle Miocene (Sarmatian) post-orogenic volcanism represented by calc-alkaline andesite dykes and sills which cut mainly Paleocene flysch rocks of the Outer Carpathians (Magura Nappe), formed the so-called Pieniny Andesitic Line (Birkenmajer, 2003).

In the Orava segment (Fig. 11) of the PKB, all the main klippen sequences are preserved, including the Kysuce, Orava, Nižná, and Czorsztyn sequences (often categorized also as units) (Potfaj, 2003). The reduced Klape sequence is present as well, but it is not unambiguously recognised in the geological maps. The youngest sediments related directly to the classical klippen sequences are the Púchov marls and Jarmuta Formation, of Campanian–Maastrichtian age. In the Orava segment, we miss such strata as the Proč or žilina formations (Paleocene–Middle Eocene), though lately somewhat similar rocks were encountered at some localities (Nižná, Kňažia) (Potfaj, 2003). However, there are Malcov and Racibor formations that are incorporated to the klippen structure in the Oravská Magura Mts. Under certain conditions of the wider definition of the *klippe* phenomenon, we may consider also the entire crest of the Oravská Magura Mts as a large klippe, which was formed together with other parts of the PKB.

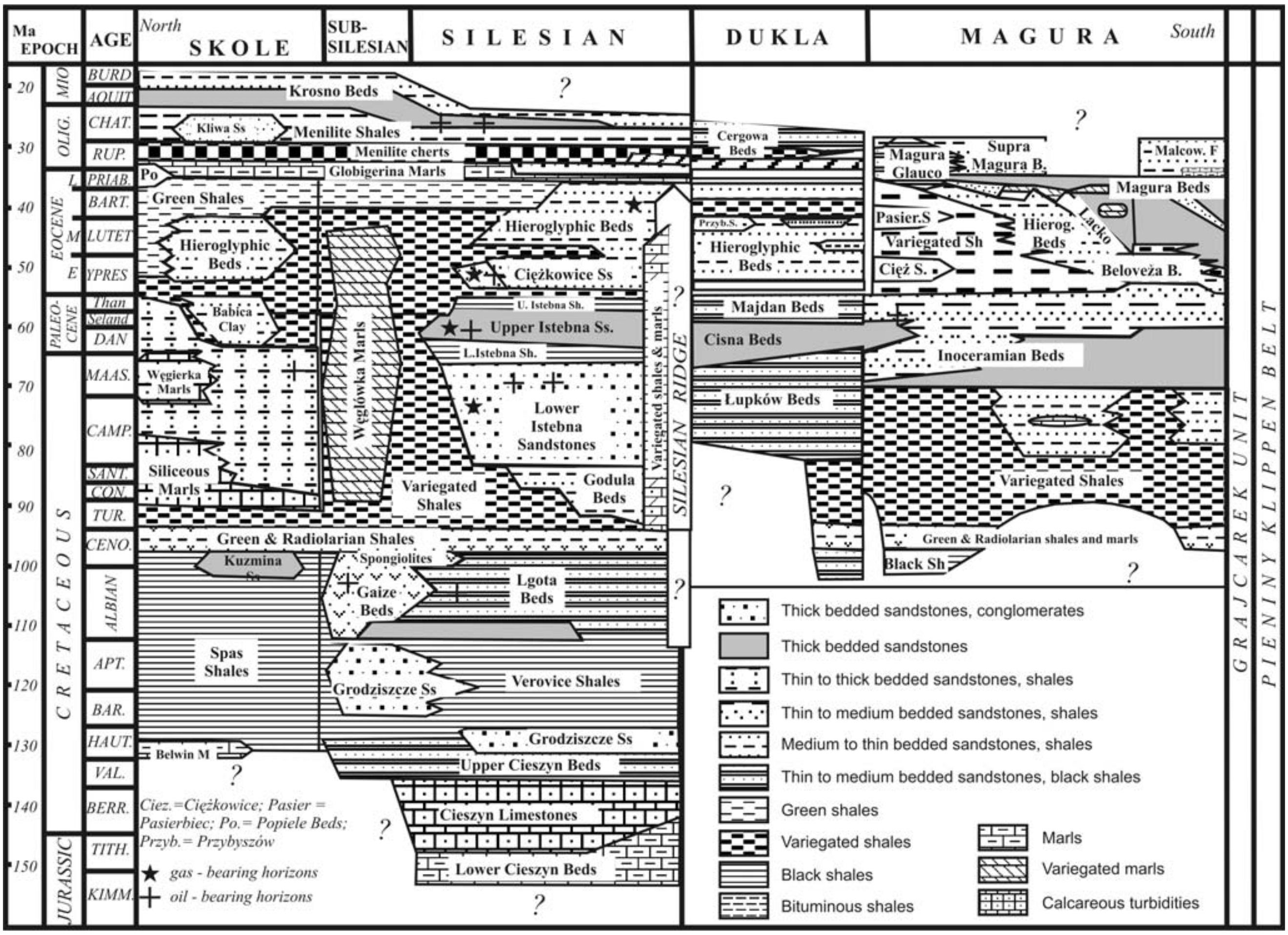

Fig. 9 Lithology and chronostratigraphy of the Polish Outer Carpathians (after Ślączka & Kaminski, 1998; modified)

The relationship between the PKB and Magura Nappe changes along the strike of the klippen belt. In the Vah and Orava River valleys, these two units are divided by a Miocene subvertical strike-slip fault, and both units are involved in a complex flower structure. The present-day confines of the PKB are strictly tectonic. They may be characterised as (sub)vertical faults and shear zones, along which strong reduction in space of the original sedimentary basins occurred. The NE–SW orientated faults accompanying the PKB (Figs 2–4) are of strike-slip character, as indicated by the presence of flower structures on the contact zone between the Magura Nappe and the PKB, and by structural asymmetry of the Inner Carpathian Palaeogene Basin.

The tectonic character of the Polish segment of PKB is differentiated, showing both strike-slip and thrust components (e.g., Książkiewicz, 1977; Golonka & Rączkowski, 1984; Birkenmajer, 1986; Ratschbacher *et al.*, 1993; Nemčok & Nemčok, 1994; Jurewicz, 1994, 1997). In general, the subvertically arranged Jurassic–Lower Cretaceous basinal facies display the tectonics of a diapir originated in a strike-slip zone between two plates. The ridge facies are often uprooted and show thrust or even nappe character. The Niedzica Succession is thrust over the Czorsztyn Succession, while the Czorsztyn Succession is displaced and thrust over the Grajcarek Unit (e.g., Książkiewicz, 1977; Golonka & Rączkowski, 1984; Jurewicz, 1994, 1997). The Grajcarek Unit is often thrust over the Krynica Subunit of the Magura Nappe. The Upper Cretaceous–Palaeogene flysch sequences of the Złatne furrow (Golonka & Sikora, 1981) are frequently thrust over various slope and ridge sequences. In the East Slovak sector, the back-thrusts of the Magura Nappe onto PKB, as well as the PKB onto the Central Carpathian Palaeogene are commonly accepted (e.g., Lexa *et al.*, 2000). Like in the Polish sector, a mixture of thrust and strike-slip components is present, but the degree of dispersion of the Mesozoic klippen inside the Jaworki-Proč Formation is higher in the eastern, Slovak sector of the PKB. The PKB tectonic components of different age, strike-slip, thrust, as well as toe-thrusts and olistostromes are mixed together and contribute to the present-day melange character of the PKB, where individual tectonic units are difficult to distinguish.

Outer Carpathians

The Outer Carpathians (Figs 1–4, 9) are composed of a stack of nappes and thrust sheets spreading along the Carpathian arc, which are mainly built up of up to six kilometres thick continuous flysch sequences, representing the Jurassic through Early Miocene time span. All the Outer Carpathian nappes are overthrust by at least 70 km onto the southern part of the North European plate, covered by autochthonous

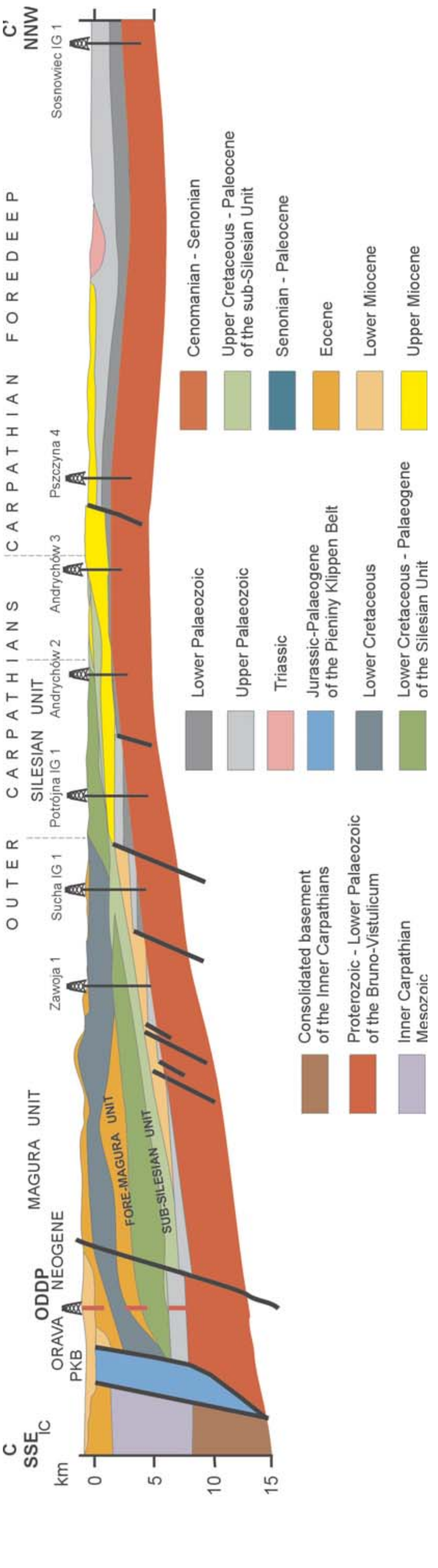

Fig. 10. Cross-section C–C' trough the West Carpathians between the Pieniny Klippen Belt in Orava at the Polish-Slovak border and the Carpathian Foreland in Upper Silesia (Chyżne-Sosnowiec line; after Oszczypko, 1998; modified). Cross-section location on Fig. 1

Miocene deposits of the Carpathian Foredeep (Książkiewicz, 1977; Pescatore & Ślączka, 1984, Picha *et al.*, 2005; Ślączka *et al.*, 2005) (Figs 10, 11). Well-log and seismic data indicate that the size of the Carpathian overthrust was at least 60 km. During overthrusting, the northern Carpathian nappes became uprooted from the basement and only their basinal parts were preserved (Figs 3, 4). According to Picha *et al.* (2005), the extent of shortening and the character of original depositional sites of various thrust sheets are little known and remain subjects of alternative geodynamic reconstructions (see, e.g. Książkiewicz, 1977; Pescatore & Ślączka, 1984; Cieszkowski *et al.*, 1985; Kruglov, 1989; Roure *et al.*, 1993; Ellouz & Roca, 1994; Matenco *et al.*, 1997; Plašienka *et al.*, 1997; Zuchiewicz 1998; Oszczypko, 1999; Plašienka, 1999; Matenco & Bertotti, 2000; Behrmann *et al.*, 2000; Golonka *et al.*, 2000, 2003, 2005; Golonka, 2004; Picha *et al.*, 2005; Ślączka *et al.*, 2005). According to Zuchiewicz (1998), the rates of thrusting in the Polish Outer West Carpathians were 19–23 mm/yr in the late Burdigalian, 21–23 mm/yr in the Langhian, 17–18 mm/yr in the Middle Serravallian, and 7–10 mm/yr in the Late Serravallian. The average rate during the 10 m.y. time span was estimated at 6–7 mm/yr. Behrmann *et al.* (2000) conclude about 260 km of shortening in the NE Outer Carpathians. Picha *et al.* (2005; and references therein) estimate the shortening of the Outer Carpathians in Moravia at about 160 km, or at ca. 9.4 mm/yr from the middle Oligocene to the early Badenian (Serravallian). Their estimation is based on balanced cross-sections (see also Nemčok *et al.*, 2000). The nappe succession from the highest to the lowest ones includes the Magura Nappe, Fore-Magura group of nappes, Silesian Nappe, Subsilesian Nappe, and Skole (Skiba) Nappe. A narrow zone of folded Miocene deposits was developed along the frontal Carpathian thrust. This is represented by the Zgłobice Unit in the Northern Carpathians, and its equivalent Subcarpathian (Borislav or Sambor-Rožniatov in Ukraine) Unit in the Ukrainian and Romanian parts of the Eastern Carpathians.

Magura Nappe

The Magura Nappe is the innermost and largest tectonic unit of the Western Carpathians (Matějka & Roth, 1950; Oszczypko 1992; Picha *et al.*, 2005; Ślączka *et al.*, 2005) which is thrust over various units of the Fore-Magura group of nappes and the Silesian Nappe. The substratum of the Magura Nappe is exposed in several tectonic windows and has also been found in several deep wells in Poland and Slovakia (e.g. Bystra IG-1, Zawoja 1, Oravska Polhora 1, Tokarnia IG-1, Sucha Beskidzka IG-1, Obidowa IG-1, Chabówka 1, Słopnice 1 and 20, Leśniówka 1). To the south, it is in tectonic contact with the Pieniny Klippen Belt that separates it from the Inner Carpathians. The oldest Jurassic–Lower Cretaceous rocks are only found in that part of the Magura basin which was incorporated into the PKB (i.e. the Grajcarek Unit; cf. Birkenmajer, 1977). The Albian/Cenomanian spotty shales remain in the southern margin of the Mszana Dolna tectonic window (Birkenmajer & Oszczypko, 1989; Cieszkowski *et al.*, 1989; Malata *et al.*, 1996). More recently, Hauterivian–Albian deposits have been recognised in a few localities in Southern Moravia

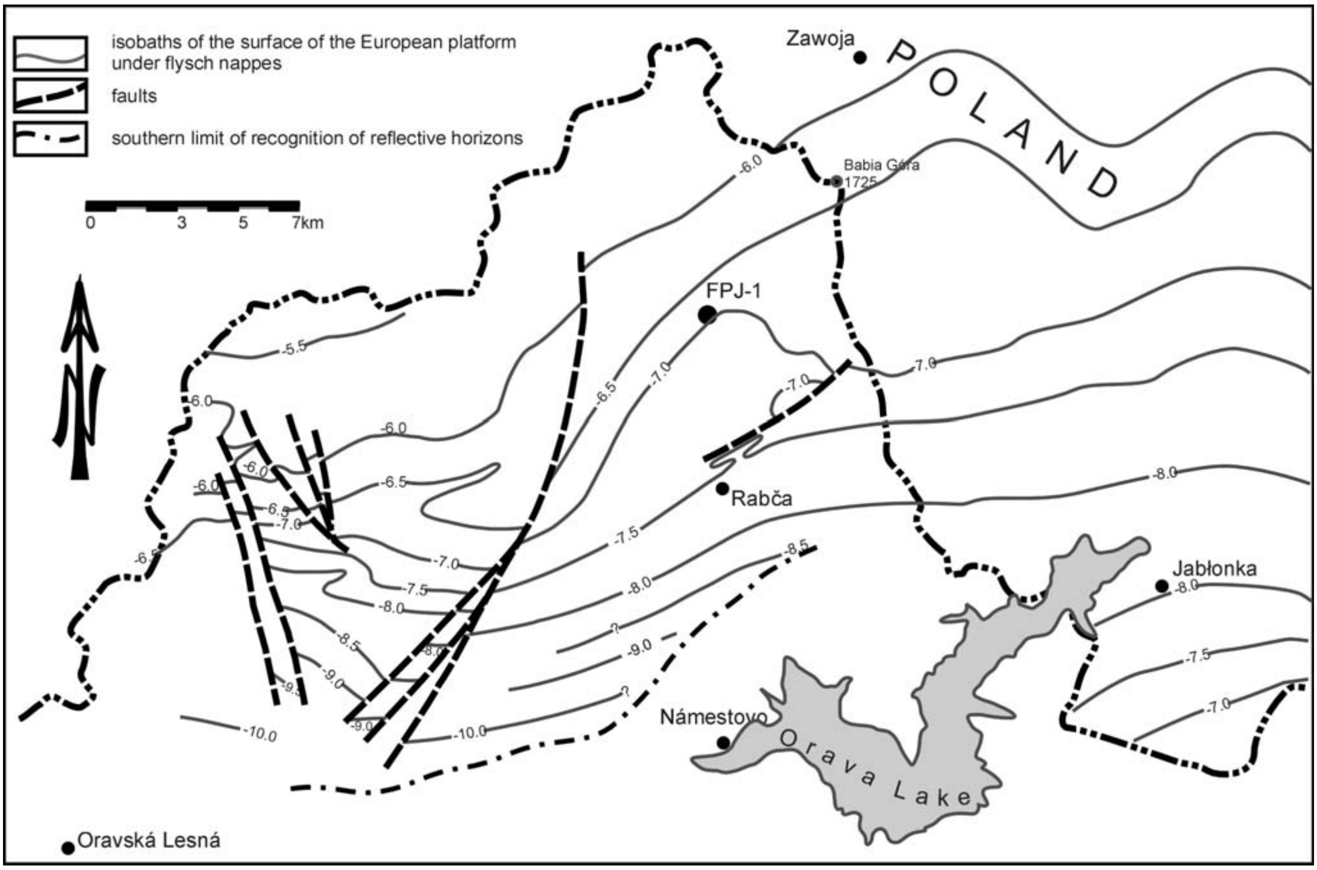

Fig. 11. Depth of the autochthonous platform under the flysch nappes in the Orava region

(Švabenická *et al.*, 1997). The Upper Cretaceous–Palaeogene deposits of the Magura Nappe may be subdivided into the Campanian/Maastrichtian–Paleocene, and Lower-Upper Eocene turbiditic complexes. Each of them begins with pelitic basinal deposits (variegated shales) which pass into thin-and medium-bedded turbidites with intercalations of allodapic limestones/marls, and then into thick-bedded ones. Finally, there come thin-bedded turbidites (Oszczypko, 1992).

The Magura Nappe is flatly thrust over its foreland, built up of the Fore-Magura group of nappes, and over the Silesian Nappe. The amplitude of the overthrust is at least 50 km, and the post-Middle Badenian thrust displacement exceeds 12 km (Oszczypko, 1999; Oszczypko & Zuchiewicz, 2000). The northern limit of the nappe has an erosional character, whereas the southern one coincides with a more or less vertical strike-slip fault along the northern boundary of the PKB. The thrust developed mainly within the ductile Upper Cretaceous variegated shales. The sub-thrust morphology of the Magura foreland is very distinctive. The shape of the northern limit of the Magura Nappe and the distribution of tectonic windows inside the nappe are connected with denivelations of the Magura basement. As a rule, the "embayments" of the marginal thrust are related to transversal bulges in the Magura basement, whereas the "peninsulas" are located upon basement depressions (Oszczypko, 2001). At a distance of 10–15 km south of the northern limit of the nappe, a zone of tectonic windows connected with the uplifted Fore-Magura basement is located (e.g. Sól, Sopotnia Mała, Mszana Dolna, Szczawa, Klęczany-Limanowa, Ropa, Ujście Gorlickie, and Świątkowa tectonic windows). The biggest one is the Mszana Dolna tectonic window, situated in the middle part of the Polish Carpathians. This window developed as a duplex structure during the Middle Miocene thrusting of the Magura Nappe (Oszczypko, 2001). South of the zone of tectonic windows, inclination of the Magura thrust surface increases, and at the northern boundary of the PKB the thickness of the nappe exceeds 5 km.

The Magura Nappe has been subdivided into four structural subunits (thrust sheets): Oravská Magura-Krynica, Bystrica (Nowy Sącz), Rača, and Siary. These subunits coincide, to a large extent, with the corresponding facies zones (Matějka & Roth, 1950; Koszarski *et al.*, 1974; Golonka, 1981; Cieszkowski *et al.*, 1985; Ślączka *et al.*, 2005). In the area surrounding the Mszana Dolna and Szczawa tectonic windows, the basal part of the nappe built up of Upper Cretaceous–Paleocene flysch rocks is strongly deformed. In the Lower to Upper Eocene flysch of the Rača and Krynica subunits, broad, W–E trending synclines and narrow anticlines dominate. The southern limbs of synclines are often reduced. In the Bystrica (Nowy Sącz) Subunit, subvertical thrust sheets are common. Both the northern limbs of anticlines and southern limbs of synclines are tectonically reduced and usually overturned. The Magura flysch in the Orava region (Fig. 4) is folded into several major folds of WSW–ENE to W–E trend, i.e. parallel to the trace of the Magura frontal thrust some 20–30 km ahead. These folds are, therefore, termed here longitudinal folds F_L. These are north-vergent, overturned fault-propagation folds, exposing local imbricated thrust planes along their inverted limbs. On a regional scale, upon the F_L folds superimposed are younger and smaller "diagonal" folds F_D, of the East-Carpathian, NW–SE trend, and local "transverse" folds parallel and related to major transverse faults (Aleksandrowski, 1985, 1989). The buckle fold interference pattern closely resembles that produced experimentally by, for instance, Ghosh and Ramberg (1968) and Skjernaa (1975). The folds of all generations are accompanied by regionally persistent joint sets, which are symmetrically orientated and genetically related to folding. Most minor to major high-angle normal faults cross-cutting the fold structure of the Magura Nappe (Fig. 2) seem to have initiated on earlier joint surfaces. The sequential development of minor tectonic structures over the Polish segment of Orava reflects successive stages of thrusting of the Outer Carpathian nappe pile onto its foredeep, accompanied by a gradual dextral rotation of regional tectonic compression trajectories during Miocene to Pliocene times (Aleksandrowski, 1989).

Fore-Magura Zone

The Fore-Magura Zone (Książkiewicz, 1956) includes a group of tectonic units which are folded and thrust one upon another. The Dukla Nappe (Ślączka, 1970) is the largest and most important unit of the Fore-Magura Zone. It crops out on the surface in the eastern sector of the Polish and Slovak Outer Carpathians and in Ukraine. The Dukla Nappe is stretching from the Polish to Ukrainian Carpathians. In its SE part, the nappe consists of several imbricated, thrust-faulted folds showing NW–SE orientation and maximum elevation in the eastern part of the nappe. The fold axes plunge gradually towards the northwest, and eventually the entire nappe disappears below the Magura Nappe. Well-bore data (Zboj 1, Smilno 1) show that the Dukla Nappe extends under the Magura Nappe far to the south. South-east of the Slovak-Ukraine border, where Magura Nappe disappears, the inner part of the Dukla Nappe is hidden below the Porkulec Nappe. From the more external Silesian Nappe and/or Zboj Unit, the Dukla Nappe is separated by a thrust plane which is more distinct in the eastern part than in the western one. Data from deep boreholes Jaśliska 2 and Wetlina 3 indicate that the thrust plane in the Polish part of the nappe is very steep. However, data from deep borehole Zboj 1 show that the thrust exceeds 15 km and that the thrust plane beneath the more internal part of the Dukla Nappe becomes more flat. Both in the Polish and Slovak parts, the two subunits can be distinguished in the Dukla Nappe, namely the internal and external ones. Folds within the internal subunit are generally gently dipping towards the southwest and are characterised by low-dipping overthrusts, whereas within the external subunit the folds are steep and often with a reversed (southwestern) vergence. The internal subunit disappears on the border between Slovakia and Ukraine; however, it cannot be excluded that this subunit continues into the Porkulec Nappe.

Beneath the Dukla Nappe, a separate tectonic unit, the Zboj Unit, was described from borehole Zboj 1 situated in eastern Slovakia. Only a fragment of a limb of an anticline does represent this unit. Its internal structures and relation to the more outer tectonic units, especially the Silesian Nappe, are unknown.

Towards the west, following the facies changes, the Dukla Nappe possibly passes into the Obidowa-Słopnice (Cieszkowski *et al.*, 1981a,b; Cieszkowski, 1985, 2001) and Grybów nappes. Both have been encountered in several deep wells below the Magura Nappe; for instance, in the Rabka - Nowy Targ (Obidowa IG-1, Chabówka 1) and Limanowa - Słopnice areas (Słopnice 1, Słopnice 20, Leśniówka 1, and others). The sedimentary sequences of these nappes are represented by the Upper Cretaceous–Paleocene deposits developed as thin- and medium-bedded shale-sandstone flysch (Inoceramian, Łupków, and Majdan beds), replaced in part by thick-bedded turbidites (Cisna and Bukowiec Wielki beds). These are covered by the Eocene thin-bedded flysch (Hieroglyphic beds) with local fans of the thick-bedded Przybyszów Sandstones. Upper Palaeogene strata are represented by the Menilite (Mszanka Sandstones, Jawornik Marls, Cergowa Sandstones, Menilite Shales) and Krosno beds. In the Obidowa-Słopnice facies zone, the uppermost Eocene and Lower Oligocene deposits are developed as very special facies with coarse sandstones and fine conglomerates (Zboj Sandstones), and silicified sandstones with intercalations of black mudstones (Rdzawka beds). The Grybów Nappe is exposed in several tectonic windows within the Magura Nappe, starting from the Mszana Dolna tectonic window in Poland to the Smilno tectonic window in Slovakia. The latter was also encountered in several boreholes below the Magura Nappe. This unit is strongly folded, with several disharmonic thrust-faulted folds. The Grybów Nappe is thrust over the Obidowa-Słopnice Nappe or the Silesian Nappe. The Obidowa-Słopnice Nappe is present in several boreholes between Obidowa and Słopnice (e.g. Obidowa IG 1, Chabówka 1). The strata of this unit are gently dipping towards the south, usually without any intense tectonic deformations, except in the higher part.

The innermost unit of the Fore-Magura Zone, called the Jasło Nappe (Koszarski, 1999), has been distinguished close to the cities of Gorlice and Jasło. It was separated in the Harklowa and Łużna "peninsulas" from the Magura Nappe. The Jasło Nappe is flatly thrust over the Silesian Nappe with complicated internal structures. However, there is an opinion (Jankowski, *pers. comm.*) that strata regarded as belonging to the Jasło Nappe represent olistostromes within the youngest deposits of the Silesian Unit.

In front of the Magura Nappe, two more outer units are located. These are the Fore-Magura Nappe *s.s.* (e.g., Książkiewicz, 1977) and Michalczowa Unit (Cieszkowski, 1992). The Fore-Magura Nappe *s.s.* occupies the most western position, near the town of Żywiec. It consists of two narrow, asymmetrical anticlines accompanied by thrust faults, the inner anticline being strongly deformed and disharmonic. Towards the east, this unit disappears completely. The Upper Cretaceous, Paleocene and Eocene rocks include strata analogous to those of the Magura or Dukla nappes, i.e. the Inoceramian beds, Variegated Shales, and Hieroglyphic beds. The Uppermost Eocene–Oligocene deposits are represented by typical facies of the Menilite and Krosno beds, as well as by other rocks, like Duląbka beds, Grybów Marls, Cergowa beds, and Michalczowa beds. Within the Fore-Magura Unit *s.s*., the Upper Cretaceous, Paleocene, and Eocene deposits are represented by grey and variegated marls. Up to the Late Cretaceous, strata of the Fore-Magura Zone had been deposited in an area connected with the Magura Basin. According to Golonka (2005), the new Fore-Magura pull-apart basin was formed in the Late Cretaceous during strike-slip reorganization of the Outer Carpathian sedimentary realm, caused by important strike-slip movements. It was separated from the Magura basin by the Fore-Magura Ridge, and from the Silesian basin by the newly re-organised Silesian Ridge. The Dukla, Grybów-Obidowa-Słopnice, and Fore-Magura *s.s*. sub-basins were arranged in *en echelon* pattern. The Silesian Ridge (Cordillera) had separated the Dukla and Silesian basins until Oligocene time, when a wide connection between these basins was opened. This connection is indicated by the presence of facies typical for the Dukla succession in the innermost part of the Silesian Nappe (cf. Cieszkowski, 1992, 2001; Ślączka *et al.*, 2005). Today, units of the Fore-Magura Zone are tectonically covered by the Magura Nappe which was shifted from south, and all together are thrust over the Silesian Nappe. Some of these units, i.e. the Grybów and Słopnice-Obidowa ones, display a general affinity and probably belonged to a bigger nappe, which became divided into separate units during the Neogene folding.

Silesian Nappe

The Silesian Nappe occupies the central part of the Outer Carpathians, pinching out below the most internal nappes. Sedimentary facies of the Silesian Nappe represent a continuous succession of deposits of the Late Jurassic through Early Miocene age. The Late Jurassic and Early Cretaceous sedimentation is represented by carbonate deposits of the Cieszyn beds (marls, calciturbidites), passing up into sandstones and conglomerates of the Grodziszcze beds. Up the section, these strata are covered by black shales of the Veřovice beds and quartzitic sandstone-dominated flysch of the Lgota beds. During the Late Cretaceous and Paleocene, sedimentation of sandy, often thick-bedded turbidites that represent the Godula and Istebna beds took place in the Silesian Basin. Their complete thickness is estimated in the western sector of the Polish Carpathians at about 4,500 m. The sedimentation of thick-bedded sandstones (Ciężkowice Sandstones) or variegated shales lasted until the Middle Eocene, and was later replaced by thin-bedded, shale-sandstone flysch of the Hieroglyphic beds. During the Oligocene, the Menilite and Krosno beds were deposited. Then, in the southern part of the Silesian Basin numerous olistostromes were formed at certain levels (Cieszkowski *et al.*, 2003). Deposition of the Krosno beds was completed in the Early Miocene.

The Silesian Nappe stretches from Moravia (Czech Republic) to Ukraine where it looses its individuality. In the western segment of the Polish Carpathians, the Silesian Nappe is flatly overthrust onto the substratum. Within the Silesian Nappe there are several tectonic windows, where the Subsilesian Nappe is exposed. Towards the east, the thrust plane gradually plunges and the character of tectonic structures within the Silesian Nappe changes. In the western part, the structures are generally shallow and gently folded, whereas towards the east they pass into long, narrow, steeply dipping, imbricated folds. The southern part of the Silesian Nappe is hidden beneath the Magura and Dukla - Fore-Magura nappes.

West of the Soła River, the Silesian Nappe near the western border of Poland, is composed of two subunits: the Cieszyn Subunit, built up of strongly folded Lower Cretaceous strata, and the Godula Subunit, built up of Upper Cretaceous and Palaeogene deposits which dip monoclinally southwards. This part of the Silesian Nappe is cut by several transverse faults. According to Picha *et al.* (2005), the Silesian Nappe comprises both massive, several-kilometres-thick, competent strata of the Upper Cretaceous Godula and Istebna flysch formations, and the predominantly incompetent Upper Jurassic and Lower Cretaceous strata. During deformation and tectonic transport, these two lithologically different sets of strata were locally decoupled, deformed, and thrust disharmonically. The competent younger strata of the Silesian Nappe were locally thrust over the older incompetent members of the unit, thus invoking the idea of an existence of separate nappes formed during the subsequent stages of deformation. Between the Olza and Wisła Rivers (Książkiewicz, 1977), the Cieszyn Subunit consists of five thrust-sheets (partial nappes) thrust one upon another. East of Bielsko, near the Soła River, the Cieszyn Subunit becomes narrower and folded into several small anticlines. The tectonic unconformity between the complex of the Cieszyn beds and that of higher-situated beds is less evident here. Between the Soła and Skawa rivers, the Cieszyn beds occur only in small shreds at the bottom of the Godula Subunit. Farther to the east, the Cieszyn and Godula subunits merge together, and the Silesian Nappe is built up of several gently folded structures. According to Książkiewicz (1977), the imbricated folds gradually become more and more marked eastwards, and east of the Dunajec River, the Silesian Nappe consists of numerous folds. The Stróże, Jankowa and Ciężkowice folds are most important ones, with which small oil fields are associated. Towards the east, the Stróże Fold passes into the broad Gorlice Fold where one of the oldest oil fields in the Carpathians exists. Within the culmination of the next fold to the north, the Ciężkowice-Biecz Fold, small oil fields were found. The eastern part of the Silesian Nappe, east of the Wisłok River, is plunging towards the south-east and is represented by a synclinorium (Central Carpathian Synclinorium) which is mainly built up of Oligocene strata (Wdowiarz, 1985). The Central Carpathian Synclinorium is composed of several long, narrow, imbricated, thrust-faulted folds, which are often disharmonic. These folds are cut by several transverse faults that divide them into separate blocks. The folds display several along-strike axial culminations, where along the northern and southern margins of the synclinorium Cretaceous and Eocene strata are exposed. Several folds and thrust folds were distinguished within the Central Carpathian Synclinorium, including the Folusz - Bukowica - Fore-Dukla Zone, Zboiska, Lubatówka - Iwonicz Spa - Tokarnia, Osobnica - Bóbrka Rogi -Suche Rzeki, Lubienka - Mokre - Zatwarnica, Roztoki - Potok - Turaszówka - Krościenko - Tarnawa -

Wielopole - Czarna, Sanok - Zmiennica -Strachocina - Czarnorzeki, and Ustianowa - Międzybrodzie - Grabownica folds.

Subsilesian Nappe

The Subsilesian Nappe underlies tectonically the Silesian Nappe. In the western sector of the West Carpathians, both nappes are thrust over Miocene molasses of the Carpathian Foredeep, and in the eastern sector they are thrust over the Skole Nappe. The Silesian Basin and Subsilesian sedimentary area have been connected during their sedimentation period. Toward the north and northeast, the Cretaceous and Palaeogene clastic deposits, typical for the Silesian Basin, were gradually replaced in the Subsilesian sedimentary area first by variegated shales, and eventually by marls of the Subsilesian emerged ridge. In Palaeogene time, sedimentation of variegated deposits continued up to the end of the Middle Eocene. Younger deposits represent those facies which are common for the Silesian Basin. Deposits of the Subsilesian Nappe crop out on the surface in a narrow zone in front of the Silesian thrust and are exposed in several tectonic windows. Between the Soła and Wisła rivers, the Subsilesian Nappe occurs in the form of tiny fragments of variegated shales and Upper Cretaceous and Eocene marls, sometimes with the Menilite and Krosno beds (Książkiewicz, 1977). The Subsilesian Nappe has also been drilled in many boreholes between Bielsko, Cieszyn and Ustroń, beneath the Silesian Nappe. This unit also appears in the Żywiec window (Książkiewicz, 1977). The strata of the Subsilesian Nappe are intensely folded and arranged in mostly N–S-orientated slices, steeply dipping to the west, and form a diapiric anticlinal uplift. Farther eastwards, several tectonic windows occur under the Silesian and Skole nappes. These windows comprise rocks of the Subsilesian Nappe. In the frontal part of the Silesian Nappe, north of Krosno, the Subsilesian Nappe is exposed in the Węglówka tectonic half-window. Deep wells connected with the Węglówka oil field show that this window is built up of a refolded thrust-faulted anticline. The Subsilesian Nappe is steeply overthrust onto the Skole Nappe. Farther to the east, the Subsilesian Nappe forms once more a narrow zone in front of the Silesian Nappe. Near the town of Ustrzyki Dolne, the Subsilesian Nappe disappears from the surface, and the frontal part of the Silesian Nappe becomes a thrust-faulted fold and eventually joins with the Skole Nappe. There is also a possibility that tectonic continuation of the Subsilesian Nappe is the Rosluch slice in Ukraine.

In the western part of the Outer Carpathians, near the town of Andrychów, several huge blocks composed mainly of Jurassic limestones occur along the Silesian Nappe. These were regarded as tectonic klippen that were sheared off during the movements of the Silesian Nappe (Książkiewicz, 1977). However, new pieces of evidence suggest that these are olistoliths embedded in the uppermost part of the Krosno beds of the Subsilesian Nappe (Ślączka *et al.*, 2005). It is possible that the Andrychów and Subsilesian Upper Cretaceous and Palaeogene rocks were deposited within the same ridge area. The Andrychów facies represent the central, partially emerged part of the ridge, while the Subsilesian ones – a much broader slope area.

The Skole Nappe

The Skole Nappe (cf. Kotlarczyk, 1985) occupies a large area in the northeastern part of the Polish Outer Carpathians. Towards the east, in Ukraine, it is wider but towards the west its width diminishes until it eventually disappears from the surface, plunging beneath the Silesian and Subsilesian nappes. The Skole Nappe consists of several narrow elongated thrust folds. There is predominance of the Oligocene Menilite and Krosno beds cropping out on the surface in the inner zone of this unit, while the outer zone is mainly built up of Cretaceous strata. In the Skole Basin, sedimentation started not later than in the Hauterivian. The Early Cretaceous is represented by shales and marls (the Belwin Marls) and black shales (the Spas beds). At the beginning of the Late Cretaceous, Cenomanian radiolarites followed by red shales were deposited. Higher up, in the Upper Cretaceous–Paleocene section there arrive Siliceous Marls, Inoceramian beds, and other episodic deposits (Węgierka Marls, Babica Clays). The Eocene is represented by the variegated shales, Hieroglyphic beds, Green Shales, and locally widespread Popiele beds. The Oligocene Menilite beds include intercalations of very characteristic Kliwa sandstones. The Menilite beds pass upward into the Krosno beds that terminated flysch sedimentation in the Early Miocene.

In the Polish part of the Outer Carpathians there are no traces of the most external unit, known from the Ukrainian Carpathians as the Borislav-Pokuty Nappe. Its occurrence beneath the Skole Nappe in the Polish Outer Carpathians has been inferred, but evidence from deep boreholes (e.g. Paszowa 1, Cisowa IG-1) suggest that the Borislav-Pokuty Nappe does not continue to the west of the Polish-Ukrainian border (Żytko, 1997, 1999).

Deep structure below the Carpathian overthrust

The deep structure of the Polish Outer Carpathians and their basement, that is the southern continuation of the North European Platform, has been recognised by deep boreholes as well as by magnetotelluric, gravimetric, magnetic, geomagnetic, and deep seismic sounding profiles (Ślączka, 1976, 1996a, b; Oszczypko *et al.*, 1989, 2005; Picha, 1996; Guterch & Grad, 2001, Picha *et al.*, 2005; Ślączka *et al.*, 2005; and references therein). Tens of deep (up to 9,000 m) boreholes, which reached the Carpathian substratum, were drilled along the Outer Carpathians. They allowed for the recognition of the deep structures of the Carpathians, the depth of the Carpathian thrust plane, its minimum range, as well as character of the substratum. First of all, they proved the thin-skin character of the Outer Carpathians orogen, which is thrust over autochthonous Miocene deposits covering the eastern and western parts of the North European Platform. They also documented the occurrence of several uprooted nappes thrust upon each other and the existence of new tectonic and lithostratigraphic units that had not been known from the surface data (Figs 12–14).

Generally, the thrust plane of the Carpathians dips gently to the south in the western part (see Oszczypko & Tomaś, 1985; Oszczypko *et al.*, 1989, 2005). Bystra IG 1 borehole, located 30 km southward of the northern margin of the

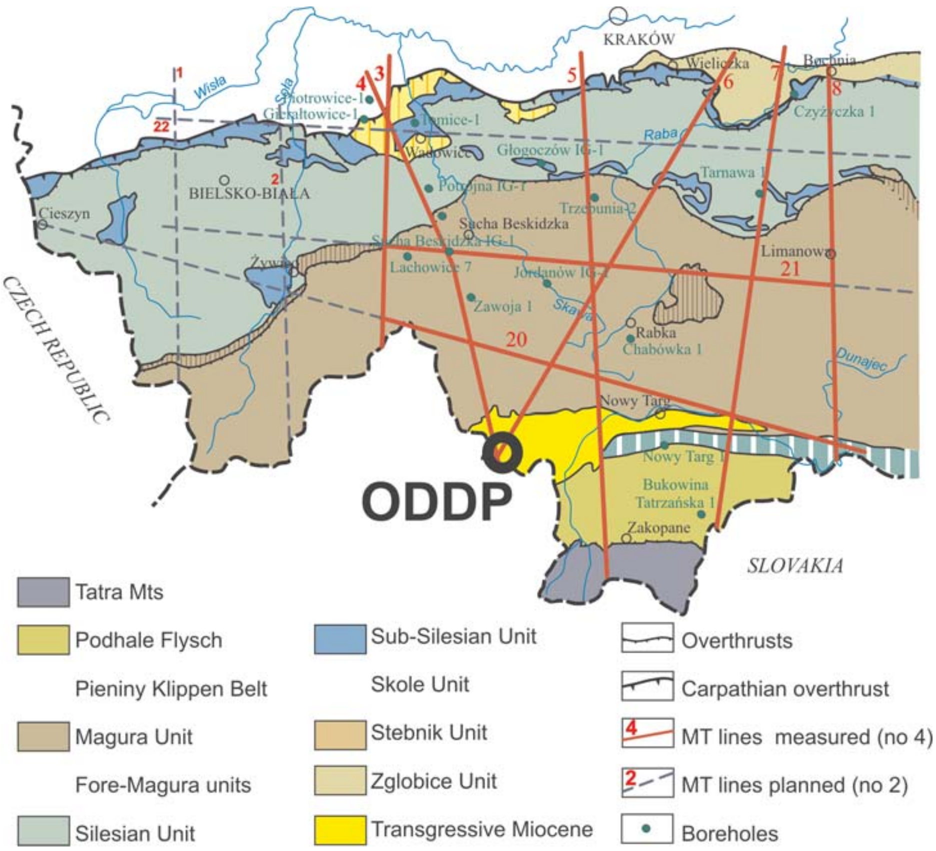

Fig. 12. Map of the Polish West Carpathians west of the Dunajec River showing the location of magnetotelluric soundings lines. Line 5 depicted on Fig. 13, line 6 on Fig. 14, line 7 on Fig. 15. ODDP – planned Orava Deep Drilling Project Well (International Continental Deep Drilling Program)

Carpathians, crossed this plane at a depth of –3,131 m below the sea level. Towards the east, this plane plunges gradually. West of Kraków, borehole Zawoja 1, also situated 30 km south of the Carpathian margin, crossed the plane at a depth of –3,225 m b.s.l. In the eastern part of the Polish Carpathians (Oszczypko & Tomaś, 1985), borehole Brzozowa 1, situated 10 km south of the margin, reached the substratum at a depth of –2,575 m b.s.l. Borehole Szufnarowa 1, 15 km away from the margin, penetrated the plane at a depth of –3,455 m b.s.l., and borehole Kuźmina 1, 25 km south of the margin, reached the plane at a depth of –6,885 m b.s.l. Seismic data provided comparable values of the depth to the thrust plane. These data were obtained from hundreds of reflection and refraction profiles crossing the Outer Carpathians, especially in their outer part.

The surface of the Outer Carpathian overthrust is of regular shape and rather gently inclined. Faults older than the Carpathian overthrust were recognised in the platform basement beneath the Outer Carpathians (Oszczypko & Tomaś, 1985). Beneath the western part of the Outer Carpathians, a system of NE–SW trending normal faults progressively lowered the platform basement from a depth of 2 km down to 10 km. All these faults are blind and do not cut the Carpathian sole thrust. West of Tarnów, the NW–SE trending normal faults dominate beneath the Carpathians and folded Miocene units. The same applies to the well documented Przemyśl gas area, where fault amplitudes tend to increase eastward to 4 km.

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

Two basement blocks occur within the Precambrian basement of the Carpathians: the Upper Silesian Block on the west and the Małopolska Block on the east, separated by the Kraków-Lubliniec Fault, taken as a terrane boundary. Refraction seismic data (Guterch & Grad, 2001; Bielik *et al.*, 2004) have proved that the crustal structure of both blocks is different. In the Upper Silesian Block, the Precambrian basement is entirely concealed not only by the Carpathian thrust-belt, but also by the Palaeozoic platform strata and foreland deposits of the Variscan orogen. In the Małopolska Block, no high-grade crystalline basement was found in the area penetrated by boreholes. Precambrian rocks are mostly represented by siltstones with mudstone, sandstone and conglomerate interbeds which are typical of flysch strata supplied from a recycled orogen, probably located to the south (Jachowicz *et al.*, 2002). The distal turbi-

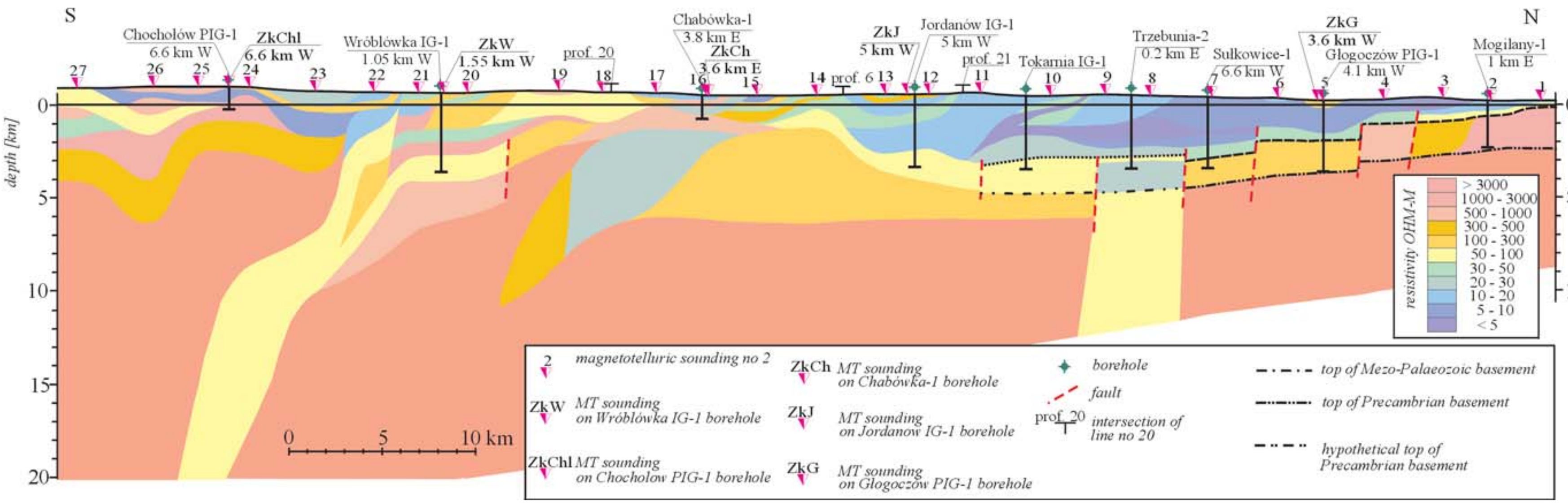

Fig 13. Magnetotelluric soundings profile along line 5 (location on Fig. 12)

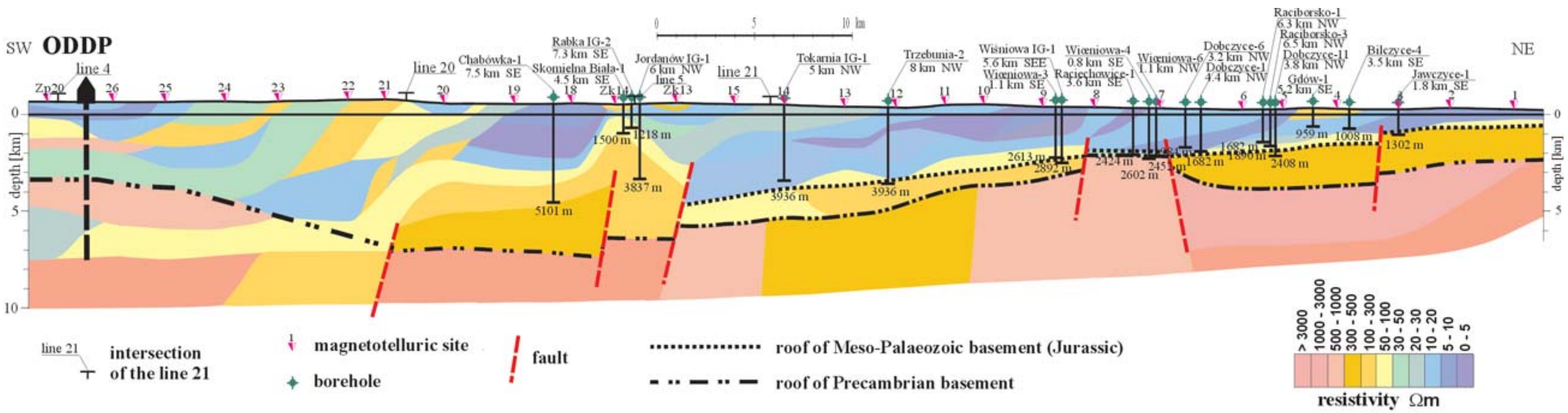

Fig. 14. Magnetotelluric soundings profile along line 6 – Chyżne-Niepołomice line (after Czerwiński *et al.*, 2002; modified) location on Fig. 12). ODDP – planned Orava Deep Drilling Project Well (International Continental Deep Drilling Program)

dite sequence was metamorphosed up to the lower greenschist facies in a WNW-trending, ca. 50 km wide belt, which is flanked on either side by unmetamorphosed rocks yielding Vendian to early Cambrian acritarchs (Moryc & Jachowicz, 2000; M. Jachowicz, *pers. comm.*). Strongly folded metamorphic rocks of the Upper Silesian Block (Bruno-Vistulicum) are discordantly covered by generally flat-laying Palaeozoic rocks. The lower part of Cambrian strata was drilled by several boreholes north-east and east of the Lachowice-Goczałkowice line, up to the Rzeszotary horst. East of that horst, the Cambrian rocks occur only locally. The Silurian and Ordovician rocks were not yet encountered on the platform below the Western Carpathians, and Devonian strata rest directly on either eroded Cambrian rocks or the crystalline basement. These Devonian rocks were encountered north of Szczyrk and north of Babia Góra Mt. Their southeastern extent is not known. The Late Devonian limestones pass upwards into those of the Lower Carboniferous. The Carboniferous strata, representing the SE continuation of the Upper Silesian Coal Basin, were found as far SE as Sucha Beskidzka, and their farther continuation is unknown. However, the occurrence of Carboniferous clasts in the Carpathian flysch as exotics suggests that these rocks continue up to the northern margin of the Carpathian basin. Remnants of the Lower Triassic deposits are preserved only in a local syncline near the town of Sucha Beskidzka, but their extent widens towards the east. The Jurassic strata below the Carpathian overthrust are only known from boreholes situated east of the Zator-Jordanów line. However, the presence of Jurassic blocks in the Carpathian flysch derived from the northern margin of the Carpathian basin implies that farther towards the south remnants of Jurassic strata may be preserved. Data from borehole Zawoja 1 suggest that in the southern part of the platform Palaeogene deposits occur (Oszczypko, 1998) and that their thickness can increase towards the south.

The position of the crust-mantle boundary (Moho) has been recognised along several seismic profiles (Guterch & Grad, 2001; Bielik *et al.*, 2004). The depth to the Moho discontinuity ranges from 30–40 km at the front of the central part of the Polish Outer Carpathians and increases to 50 km south of Nowy Sącz. South of the Pieniny Klippen Belt, these values decrease to 36–37 km. The consolidated basement beneath the Inner Carpathians is situated at depths of 10–18 km. The depth to the cratonic basement in the southern part of the North European plate below the Outer Carpathian allochthonous nappes, according to the results of deep seismic (CELEBRATION Profile 9), magnetotelluric, and magnetic soundings (e.g. Żytko, 1997; see also chapter "Magnetotelluric survey"), is below 6–8 km (Figs 4, 10, 15), while the basement depth calculated from the platform bending model is 10 km. The axis of the basement depression is situated along the Námestovo-Nowy Targ-Krynica line. The results of gravimetric studies show a distinct gravity minimum (up to 60 mGal) along the Flysch Carpathians. The axis of this anomaly runs in the west approximately along the northern boundary of the PKB, and east of the town of Nowy Targ it is shifted towards the NW. Geomagnetic soundings revealed the presence of the zone of zero values of the Wiese's vectors (see Jankowski *et al.*, 1985; Ślączka *et al.*, 2005), which in the west runs south of the Pieniny Klippen Belt. This zone is connected with a high conductivity body, 2.5–6 km thick and situated at a depth of 15–30 km, and probably indicates the position of the southern extent of the North European Platform and its contact with the Inner Carpathian basement (Cieszkowski *et al.*, 1981a; Oszczypko *et al.*, 1989, 2005). According to Picha *et al.* (2005), geological interpretation of regional seismic lines across the Outer Carpathians in Moravia and Slovakia shows that the European platform continues uninterrupted to the vicinity of the Pieniny Klippen Belt, where it is intersected by normal and reverse faults, which do not continue into the thrust belt. These faults mark the break between the thick platform-type crust and lithosphere of the European plate, and the rifted attenuated crust and lithosphere of the European continental margins. During the compressional orogeny, the edges of the thick European platform acted as a buttress, causing the piling of the rootless slices of the flysch belt and the Pieniny Klippen Belt in a low-gravity zone. Given the significant component of strike-slip motion in the Pieniny Klippen Belt, it is likely that the pieces of the European lithosphere juxtaposed on both sides of the klippen belt had moved laterally. Their present fit may thus differ from the original one, especially prior to the late orogenic northeastward translation of the Western Carpathians. Nemčok *et al.* (1998, 2000) excluded the existence of a crustal root along the western part of the Carpathian arc (Picha *et al.*, 2005) and proposed that the end of the subduction of the remnant oceanic flysch basin and the beginning of the collision were accompanied by the detachment of the subducting plate and by the occurrence of the break-off-related volcanism. Such an interpretation assumes the existence of a large oceanic domain in the Outer Carpathian Flysch Basin. The extent of the oceanic crust and lithosphere is limited to the Pieniny/Magura Basin and not necessarily under the other (Silesian, Subsilesian and Skole) basins. The subduction of the Pieniny/Magura Basin would thus have not compensated for the large shortening in the Outer Carpathian belt. Neither would the interpretation proposed by Nemčok *et al.* (1998) satisfactorily explain the fate of the large portions of the continental lithosphere, which originally underlined the Outer Carpathian depositional system. It looks like the geodynamic reconstruction of the Carpathian region remains a challenging task, whose satisfactory solution will require additional geological and geophysical studies, as well as deep drilling tackling both the specific local problems and the wider regional solutions involving the entire Alpine Carpathian system of Europe.

During the Miocene, on the southern part of the North European Plate, the Carpathian Foredeep basin filled up by the Lower Miocene molasse developed. The Middle Miocene strata were developed mainly farther to the north. The overthrust of the Outer Carpathian accretionary wedge onto the Miocene molasse deposits has been very well documented in the Outer Carpathians by deep boreholes, located as far as 30–40 km south of the present-day Carpathian frontal thrust (Wdowiarz, 1976; Oszczypko & Tomaś, 1985). According to Oszczyko *et al.* (2005; and references therein), the flexure of the foreland lithospheric plate beneath the orogenic belt contributed to the formation of the Carpathian

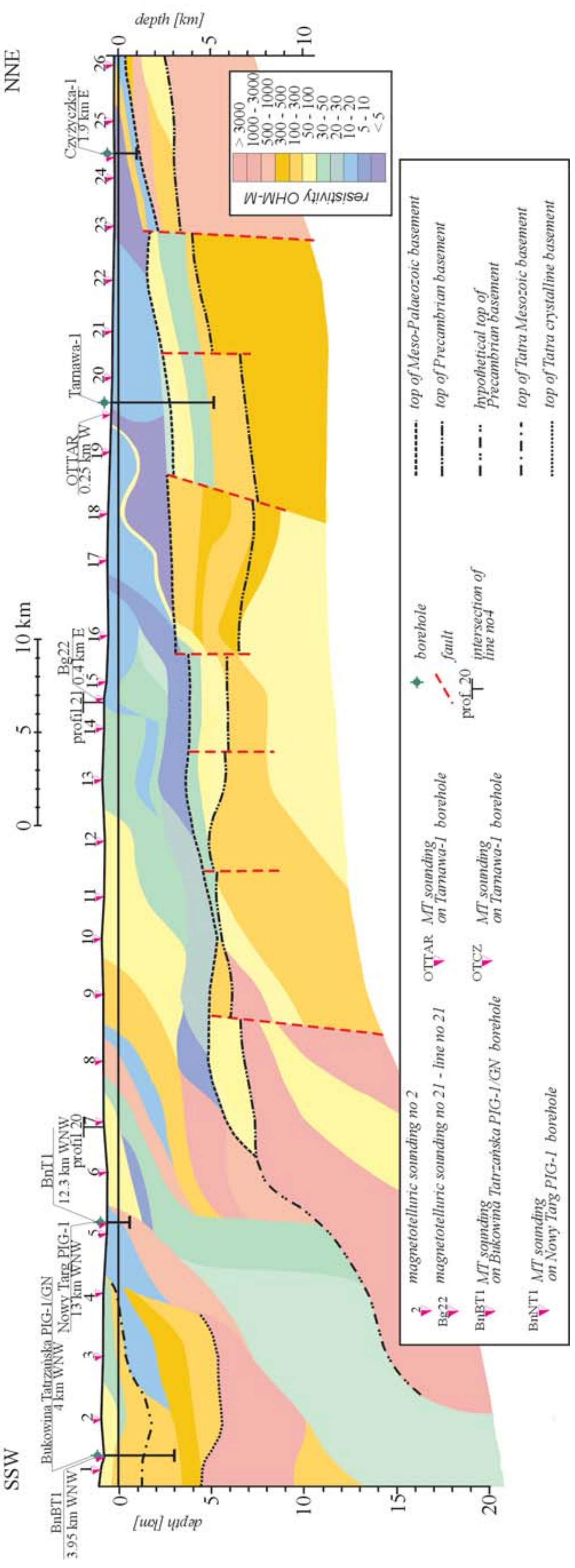

Fig 15. Magnetotelluric soundings profile along line 7 (location on Fig. 12)

Foredeep Basin (see also Royden & Karner, 1984; Royden, 1988; Krzywiec & Jochym, 1997; Zoetemeijer *et al.*, 1999).

Andesitic Magmatism

The Neogene andesites, both in the Pieniny Mts and in flysch rocks of the Outer Carpathians close to the Pieniny Klippen Belt (Fig. 16), are relatively numerous but their volume is low. According to Małkowski (1958), total aerial extent of andesite outcrops is ca. 1 km^2. The Neogene volcanic activity in Carpathian–Pannonian region is widespread. Spatial distribution, temporal relationships, and geochemical evolution of magmas contribute to interpretation of the geodynamic development of this area (e.g. Kováč *et al.*, 1997). According to Lexa *et al.* (1993), Neogene volcanic rocks in the Western Carpathians can be divided into four groups: (1) dacite and rhyolite areal-type volcanism, (2) areal-type of andesitic volcanism, (3) basalt-andesitic to andesitic volcanism, and (4) alkali-basaltic to basanitic volcanism.

According to Lexa *et al.* (1993) and Kováč *et al.* (1997), the dacite and rhyolite areal-type volcanism is related to initial stage of back-arc extension; the areal-type of andesitic volcanism is comparable to that of continental island arc (continental margin); the basalt-andesitic to andesitic volcanism represent mature island arc environment; and the alkali-basaltic to basanitic volcanism can be considered as an effect of partial melting in the uprising diapirs in the mantle in extensional conditions, and of emplacement of relatively primitive magmas (without crustal contamination) in the upper crust. The origin of most of andesites is related to processes operating in subduction zones.

According to Birkenmajer and Pécskay (1999), andesitic rocks in the Pieniny region are products of hybridization of primary mantle-derived magma over subducted slab of the North European plate. These authors pointed out that andesitic intrusions in the Pieniny Mts mark the so-called Pieniny Andesite Line, the continuation of which matches the Odra Fault Zone in Lower Silesia, where numerous Neogene alkaline silica-undersaturated mafic rock (included into Central European Volcanic Province) are present. Andesites occur in the form of dykes and sills. Numerous petrographical varieties were distinguished, based mainly on the composition of phenocryst assemblage. The age of andesites is within the range of 22.6 to 10.9 Ma (K-Ar determinations by Birkenmajer & Pécskay, 1999, 2000). Baumgart-Kotarba (2001) argued for a relationship among the andesitic subvolcanic activity, uplift of the Tatra Mountains, and the opening of the Orava Basin.

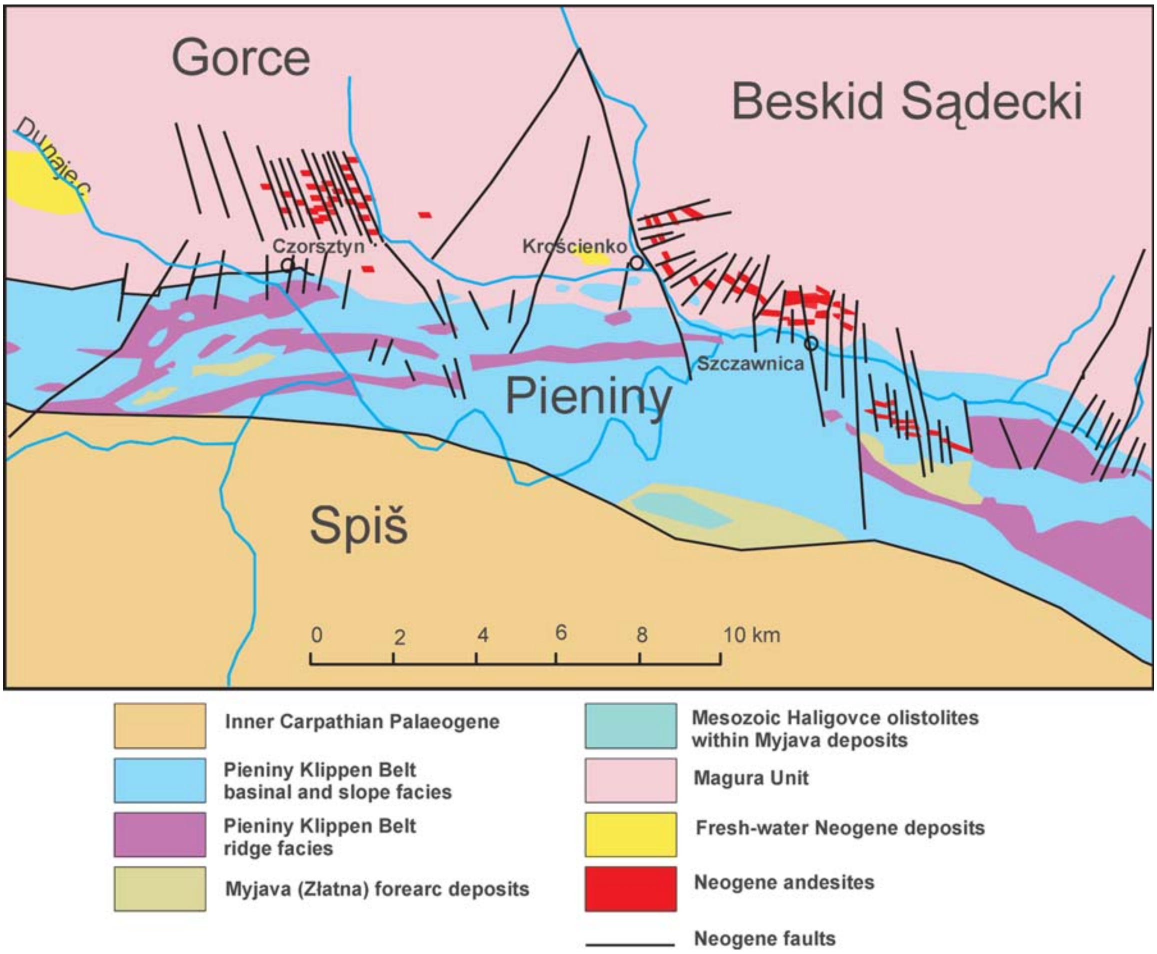

Fig. 16. Distribution of the Neogene andesite intrusions in the Pieniny Klippen Belt and Magura Nappe in the vicinity of Czorsztyn, Krościenko and Szczawnica (after Birkenmajer & Pécskay, 2000; modified). The depicted faults are perpendicular to the suture zones and general Outer Carpathian thrusts trends. These faults display strike-slip, extensional or mixed character

Orava Neogene cover

All the above-mentioned structures were covered during the Neogene by at least 600–900 m of sand–silt and clay, which were deposited in the Orava-Nowy Targ Basin. Some thin layers of coal intercalate faintly cemented silty clays. The age of the oldest sediments was established to the Karpatian–Badenian. These occur only in a sunken block in the central part of the Orava-Nowy Targ Basin, close to the Suchá Hora area, where the OH-1 borehole was drilled (Gross *et al.*, 1993) without reaching the base of the Neogene sequence. The Karpatian–Badenian deposits have been also found in the Polish sector by Cieszkowski (1995). According to palynological analysis by Oszast and Stuchlik (1977), the fresh-water sedimentation started in Late Badenian (Serravallian) time. The dip of the sedimentary pile is subhorizontal (10–15°), only at the edge of the basin rising to 20–25°. A borehole in the vicinity of Czarny Dunajec (Watycha, 1977) revealed 920 m of Miocene and Pliocene deposits.

The origin of the Orava-Nowy Targ Basin was interpreted as a result of flexural warping caused by the reverse overthrust of the Magura Nappe onto the Pieniny Klippen Belt (Roth *et al.*, 1963) and as such it could be classified as a retroarc-type basin. Modern interpretations, however, prefer to explain it as a reaction to strike-slip movements in the basement (pull-apart basin) (Royden, 1985, Baumgart-Kotarba, 1996, 2001; Pomianowski, 2003). This idea is also supported by relative steepness of the basin burial curves (Nagy *et al.*, 1996). The present-day elevation of the Babia Góra and Pilsko mountain blocks and the morphotectonic depression between these two are most probably related to the extensional regime enhanced by erosion. Recent vertical movements in the depression area are up to + 0.5 mm per year (Vanko, 1988; Vass, 1998). According to Vass (1998), the most recent geomorphological depression is a consequence of lithological contrast between the soft sedimentary infill and harder rocks surrounding the depression. This is only partially true because the hard sandstones exist also in the morphotectonic depressions in the northwestern Orava in Slovakia and in the area between Jabłonka and Rabka, north of the Orava Basin (Lexa *et al.*, 2000). According to Zuchiewicz *et al.* (2002), the Orava Basin is bordered by sets of normal faults (see also Pomianowski, 2003; and references therein). At least some of the faults were still active during Quaternary time (Baumgart-Kotarba, 1996, 2001; Zuchiewicz *et al.*, 2002). The studies of the 1995 earthquake (Baumgart-Kotarba, 2001; and references therein) show good agreement of the focal model with the tendencies of vertical crustal movements. The faults bordering the depression are steep to vertical (Baumgart-Kotarba, 2001; Pomianowski, 2003), indicating the mixed origin – strike slip changed into extensional. The thickness of Neogene deposits near the boundary fault in the Czarny Dunajec well (Watycha, 1977) indicates a synsedimentary character of this fault. The Domański Wierch Conglomerates (Tokarski & Zuchiewicz, 1998; Zuchiewicz *et al.*, 2002; and references therein) of Sarmatian to Pliocene age (Oszast & Stuchlik, 1977) display fractures formed in a strike-slip regime. Probably, the oblique faults described by Pomianowski (2003) are entering the periphery of the Orava Basin. Preliminary results of an analysis of both digital elevation model and shaded relief map (Chrustek, 2005) have shown abrupt depressions and steeply dipping slopes of linear arrangement. These are clearly visible in the southern part of the Orava Basin. The steep slopes of the Oravska Magura and Skorušina Mountains point to displacement in the southern fringe of the Orava Basin. The region situated in front of the Oravska Magura and Skorušina Mountains is strongly depressed and resembles a downfaulted block. The rectilinear contour pattern allows one to distinguish several linear structures on a topographic map. Lineaments distinguished on the basis of digital elevation model tend to cluster along W–E, SW–NE, NNW–SSE, and SSW–NNE orientations. Morpholineamets detected on contour and shaded relief maps, as well as on the 3D model coincide with faults which are buried under sedimentary infill of the Orava Basin.

The relationship between andesitic volcanism and opening of the Orava Basin indicates a possibility of the rifting thermal regime. Possible occurrence of volcanic bodies under the basin infill has been indicated by magnetotelluric surveys (Fig. 14).

Approximately 300 km SW of the Orava Basin, the pull-apart Vienna Basin was opened during Badenian (Serravallian) time due to the activity of NE–SW-trending sinistral, crustal faults (Royden *et al.*, 1982). Structural setting of the Orava Basin, however, is somewhat more complicated. The Orava Neogene sediments cover all the major faults of the Pieniny Klippen Belt and Magura Nappe, preventing their direct recognition and study. The age of the oldest sediments within the Orava Basin (Baumgart-Kotarba, 2001) is Late Badenian (Serravallian). The NNW–SSE trending (at the present-day coordinates) strike-slip tear faults in the Orava segment of the Outer Western Carpathians were reactivated during late Miocene times in an extensional regime as normal faults. The depocenter of the Orava Neogene is located in the Czarny Dunajec area, in the western part of the basin. It is hypothesized to have reflected synorogenic sedimentation along an extensional border fault of presumable N–S orientation. Later tectonic events seem to have also reactivated the WSW–ENE to W–E trending strike-slip and normal faults, producing the present-day Orava-Nowy Targ Basin which is filled with Quaternary deposits resting on top of the Miocene–Pliocene strata. Throughout the Late Miocene–Pleistocene time span, complex tectonic processes occurred in the Orava Basin. The extensional period was followed by a compressional one. The extensional regime led to subsidence in the Orava Basin. In the western part of the basin, these processes were active in the Miocene, but the Pliocene compression caused uplift of this area together with adjacent parts of the Magura Nappe (Baumgart-Kotarba, 2001), while the zone of active subsidence became shifted to the east. The most intensive subsidence occurred during the late Pliocene and Pleistocene in the northeastern part of the Orava Basin, where the Wróblówka graben was formed. This graben is bordered by normal faults (Pomianowski, 2003) and filled with thick Quaternary sediments (Baumgart-Kotarba, 2001). The compression hardly extended to this part of the Orava Basin (Baumgart-Kotarba, 1992), while in the south the Domański Wierch, Podhale, and Pieniny Klippen Belt became uplifted, probably due to reactivation of the strike-slip fault (Baumgart-Kotarba, 2001) which limits the southern part of the Orava Basin and produces recent earthquakes.

According to Golonka (2003), a possible astenosphere upwelling may have contributed to the origin of the Orava Basin, which represents a kind of a rift modified by strike-slip/pull-apart processes. In this way, a local extensional regime must have operated on a local scale of the Orava region, within the frame of an overall compressional stress field affecting the entire West Carpathians; a situation apparently uncommon within the Alpine system of Europe, of as yet not fully recognised impact on the orogen evolution (Golonka & Lewandowski, eds., 2003).

As was mentioned above, many questions remain open. We think that without additional direct geological data, which can be achieved only by deep drilling under the Orava Deep Drilling Project (International Continental Deep Drilling Program), these questions cannot be fully and properly answered.

PODHALE THERMAL CONDITIONS

The terrestrial heat flow values amount to 55–60 mW/m^2. In particular, these values have been determined at 55.6 mW/m^2 and 60.2 mW/m^2 in Zakopane IG-1 and Bańska IG-1 boreholes, respectively (Chowaniec & Kępińska, 2003). The average geothermal gradient varies between 1.9 and 2.3°C/100 m. In some parts of the main geothermal aquifer, positive thermal anomalies were detected: the temperatures at depths of 2–3 km amount to over 80–90°C, i.e. markedly higher than those resulting from the geothermal gradient (Fig. 6). Apart from heat convection within the aquifer, this fact can be explained by an increased upflow of heat and/or hotter fluids from the deeper part of the system along discontinuity planes (Kępińska, 1997). In particular, it refers to the northern part of the system which borders the Pieniny Klippen Belt. The role of this zone in creating thermal conditions of the Podhale system is certainly very important, but so far, insufficiently understood. The up-to-date results of geophysical exploration have shown that the zero induction anomaly is ascribed to it. This suggests the occurrence of geothermal fluids at great depths (6–16 km) and increased upflow of heat and hot fluids (Jankowski *et al.*, 1982), as well as their inflow into the geothermal aquifers. This supposition has been supported by the surface thermal anomalies recorded on the tectonic contact between the Podhale Basin and the Pieniny Klippen Belt, 2–3°C above the average background values (Pomianowski, 1988) (Fig. 6). Moreover, other surface studies of regional dislocation zones in the Podhale area have shown them as a privileged paths of increased heat transport (Kępińska, 1997). Since some of them are still tectonically active, they can play important role in the discussed heat transport, too (Chowaniec & Kępińska, 2003).

MAGNETOTELLURIC SURVEY

In the early 1960s, a first-order magnetotelluric (MT) anomaly north of the Tatra Mts, running parallel to the Carpathian axis, was recognised (Jankowski, 1967). However, the nature of this anomaly still remains a puzzle. Essentially, two competing hypothesis do exist: the origin of MT anomaly in the Carpathians has been related to the presence of either graphite or mineralized waters at depth (Jankowski *et al.*, 1982, 1985, 1991; Żytko, 1997). Moreover, the concept of a thick complex of silty-clayey sediments is considered as the source of the conductivity anomaly (Stanley, 1988; Czerwiński & Stefaniuk, 2001, Czerwiński *et al.*, 2002).

As a result of previous studies and an international MT survey across the Tatra Mts employing combined magnetotelluric and geomagnetic methods, the models of resistivity distribution and deep lithosphere structure were obtained (Lefeld & Jankowski, 1985; Ernst *et al.*, 1997; Bielik *et al.*, 2004). Due to strong man-made electromagnetic noise as well as complex near-surface geology of the area, resulting in strong disturbances of the telluric field, the western part of the Polish Carpathians is not a favourable site for magnetotelluric survey. Therefore, the available previous MT data may be erroneous. Since 1997, the Geophysical Exploration Company has been applying the MT-1 system, which enables such problems to be avoided. The MT soundings were conducted along a number of profiles in the western part of the Polish Carpathians under *The Project of Magnetotelluric Survey in the Carpathians* (Figs 12–15). Two profiles (Figs 13, 15) crossed the Pieniny Klippen Belt and ended near the northern margin of the Tatra Mts. Two other profiles reached the Pieniny Klippen Belt. Profile 6 (Fig. 14) runs from the planned Orava Deep Drilling Project well site near the Polish-Slovak border to Niepołomice east of Kraków. Results of the interpretation of new MT sounding data allowed the structural model of the Carpathian basement in the deep front of the Pieniny Klippen Belt to be obtained, as well as the relation between the Inner Carpathians and Outer Carpathians to be recognised (Czerwiński & Stefaniuk, 2001; Czerwiński *et al.*, 2002). The presence of a low-resistivity complex beneath the Tatra Mts confirms an allochthonous character of the Tatra massif (Ernst *et al.*, 1997). A block of high-resistivity rocks occurs in front of the Pieniny Klippen Belt, probably representing an uplifted part of the basement. However, it cannot be excluded that the high resistivity of the medium is connected with magmatic intrusions.

The Carpathian conductivity distribution anomaly is one of the best recognised conductivity anomalies in the world (Jankowski & Praus, 2003). It was discovered as a second in Europe, after the North German-Polish anomaly. Its discovery was based on the results of the so-called geomagnetic sounding, and later confirmed by magnetotelluric sounding. The length of the anomaly is of the order of thousands of kilometres. The distribution of magnetic induction vectors for one-hour periods is very regular and can be approximated by a two-dimensional structure. The axis of the anomaly, i.e., the line separating regions in which the vectors have opposite directions, is very distinctly marked. This line runs nearby the Pieniny Klippen Belt. In some places the two lines overlap, but usually they are some tens of kilometres apart. In practice, the geomagnetic measurements in the Carpathians were carried out by geophysicists from the countries in which the Carpathians are situated. Interpretations of various research groups exhibited considerable differences. The models can be divided into two groups. The first includes those models which assume the existence of a few kilometres thick, near-surface layer of sedimentary rocks, underlain by a highly-resistive rock complex bearing very well-conducting rocks at a depth of some 30 km. The other group contains models that explain the observed features exclusively by the induction in that sedimentary rock complex, in which the conductivity near the surface is lower, and the highly-conductive rocks are situated at a depth of the order of 7–17 km.

The well-conducting rock complex is composed of graphitized or porous rocks filled-in with mineralized water. Alternatively, shale complexes are regarded as a source of the conductivity anomaly. In most cases, the resistivity of clayey sediments is several Ω m. Downhole data from a few boreholes drilled in the eastern part of the Polish Carpathians have proven that resistivity of Lower Palaeozoic shales falls in the same range (Czerwiński & Stefaniuk, 2001; Czerwiński *et al.*, 2002). It is likely that due to relatively low heat flow in the northern Carpathians, the rocks have not been metamorphosed (Stanley, 1989) and preserved their high conductivity. In our opinion, the question which of these hypotheses is more likely, remains open. We think that without additional direct geological data, which can be achieved only by deep drilling, this controversy cannot be solved.

FAULTS AND BLOCK ROTATIONS

The Carpathian nappes are cut by several major faults of different origin. Some of these faults are local, some form huge systems more than one hundred kilometres long. These systems affected all of the Outer Carpathian nappes, the Pieniny Klippen Belt, and the Inner Carpathians (Figs 2, 16). According to Marko (2003), the WSW–ENE mesoscale dextral strike-slip faults exist along the boundaries and within the Pieniny Klippen Belt in Slovakia, documenting the presence of dextral shearing between the Alcapa and North European plate during the Early Miocene.

During the Early-Middle Miocene, the Outer Carpathian nappes became fully detached from the basement and thrust northwestward in the west and northward onto the North European plate in the western part of the Polish Carpathians (Cieszkowski, 2003). The bearing of nappes and major structures is SW–NE in the western part (Moravia and Slovakia), turning gradually eastwards into W–E (Golonka, 1981). The major faults dissect nappes perpendicularly and/or obliquely. The main faults are orientated N–S and NNW–SSE, and originated mostly as strike-slip faults. The age of some of these faults is younger than the formation of the Pieniny Klippen Belt. The major right-lateral strike-slip faults displaced the PKB in the Zazriva area (Potfaj, 2003). This fault system extends northwards toward the Żywiec tectonic window and the Carpathian border near Bielsko-Biała, and divides the Silesian Nappe into the present-day morphotectonic units – Beskid Śląski and Beskid Mały ranges (Golonka, 1981). Another major fault system runs along the Skawa River. It displaces the Silesian and Subsilesian Nappes as well as the Carpathian front in the Wadowice area. Several faults are orientated NNW–SSE, while others strike N–S and NW–SE in the Beskid Żywiecki Mts, south of the Magura Nappe's front. These faults displace the thrust front of the Bystrica Unit. These nappe and thrust displacements indicate that the age of the faults is younger or contemporaneous to the main Badenian (Serravallian) phase of formation of the Outer Carpathian fold-and-thrust system.

Strike-slip motions led to rotation of blocks along the faults planes. These rotations are especially well visible in the Beskid Wysoki Mts, where the huge rigid bodies of sandstones exist (Golonka, 1981). The Upper Eocene Magura (or Poprad, Kýčera, Babia Góra in Slovakia) thick-bedded sandstones are up to 1,500 m thick in the Babia Góra and Pilsko mountains. The sandstones cap less competent flysch rocks of the Late Cretaceous to Eocene age. These rocks often form diapirs between the Magura (or Poprad, Kýčera, Babia Góra in Slovakia) sandstones that build the rotated blocks. The Upper Cretaceous red Cebula shales reach the surface in one of such diapiric slices which is situated southwest of the rotated sandstone block of the Pilsko Mt. The so-called Sopotnia Mała Window also represents part of the diapir related to a rotated block. The centre of the diapir is built up of the Upper Cretaceous Cieszyn beds and Palaeogene Krosno beds representing the Silesian or Fore-Magura units, originally situated below the Magura Nappe. Strike-slip related block rotations are responsible for palaeomagnetically detected changes in the Outer Carpathian flysch. Márton (2003) recognised two phases with a total of 90° rotation in the western part of the Carpathians. The rotations are also visible at the contact between the Inner and Outer West Carpathians, both within the andesites (Márton *et al.*, 2004) and along faults that border the Orava Basin (Baumgart-Kotarba *et al.*, 2004).

CONTEMPORARY STRESS FIELD OF THE WESTERN OUTER CARPATHIANS AND THEIR BASEMENT FROM BOREHOLE BREAKOUTS

The present-day stress has been investigated in the Polish part of the Outer Carpathians by means of borehole breakout analysis. The main structural elements that play dominant role in recent geodynamics of the Outer Carpathians are the Alcapa tectonic block (Central Carpathians) and the North European plate, which is covered in part by the Carpathian accretionary wedge. The autochthonous basement of the Outer Carpathians and the foreland plate is divided into the Upper Silesian Massif and the Małopolska Massif. Remnants of these massifs are believed to continue southwards as far as the PKB suture. Considerable differences in stress distribution between the domains of the Upper Silesian and Małopolska massifs have been detected in the Outer Carpathians.

The most prominent feature of the stress field in the Upper Silesian Massif domain is systematic, counterclockwise S_{Hmax} rotation with increasing depth. Moderate and poor quality data for the flysch nappes indicate the NNE–SSW mean direction of S_{Hmax}. Good quality data for the autochthonous basement show remarkably different stress orientation. Herein, S_{Hmax} rotates from NNW–SSE to NW–SE in the deepest well sections. A shift of stress direction across the Carpathian floor thrust was directly documented only in one borehole (Lachowice-7), where maximum range of S_{Hmax} rotation from the nappes to the metamorphic basement reaches 60°. Nevertheless, consistent stress rotation is observed in most of the wells in the Upper Silesian Massif domain. In the Małopolska Massif domain of the Outer Carpathians, stress direction is more stable. In the autochthonous basement below the front of the orogen, S_{Hmax} is limited to a narrow range of azimuths N0-20°E. Within the flysch nappes, data of lower quality indicate S_{Hmax} distortion from NNE–SSW in the central segment, to ENE–WSW in the easternmost segment of the Outer Carpathians.

South of the PKB, scarce breakouts were found in one borehole (PGP-2) only. These suggest the NNE–SSW direction of S_{Hmax}, which is consistent with both the left-lateral strike-slip motion along the Domański Wierch fault (Baumgart-Kotarba, 2001), and recent direction of convergence between the Central and Outer Carpathians (Hefty, 1998).

The hitherto collected data point to present-day compressive reactivation of the Carpathians. The Alcapa block, advancing towards the NNE, exerts thin-skinned compression in the flysch nappes of the Outer Carpathians. The Alcapa push seems to involve also the autochthonous basement of the Małopolska Massif domain, since analogous S_{Hmax} orientations were documented under the front of accretionary wedge and in the foreland. Therefore, a resistive contact between the Alcapa and Małopolska Massif can be anticipated. On the contrary, the contact between overriding and subducting plates in the Upper Silesian Massif segment of the Outer Carpathians appears to be weak, as the Alcapa push does not affect the basement. In this domain, S_{Hmax} rotations can be interpreted in terms of stress partitioning due to interference of two stress generating factors, including: the SE-orientated Mid-Atlantic ridge push that might propagate from the Bohemian Massif to the basement of the Upper Silesian Massif, and the Alcapa push component that governs stresses within the nappes. Accommodation of left-lateral strike-slip along the SSW–NNE orientated fault zones can also be taken into account.

NEOGENE TECTONIC EVOLUTION OF THE NORTHERN CARPATHIANS

The dominant tectonic phase that affected the Outer Carpathians took place in the Miocene. The Miocene tectonic mobility occurred during the collision between the overriding Alcapa block and the North European plate (Cieszkowski, 2003). The boundary between these plates is the Pieniny Klippen Belt. As a result of the intense Neogene orogeny, the sedimentary infill of the Outer Carpathian basins became folded and detached from its substratum, and several uprooted nappes were created reflecting the original configuration of these basins. During folding and thrusting, the main nappes were in part differentiated and subdivided to smaller tectonic units. The Outer Carpathian nappes, thrust one upon another, are all together overthrust onto the North European plate. The Carpathian Foredeep was formed in front of the steeply northward-advancing nappes (Golonka & Lewandowski, eds., 2003).

Chattian–Burdigalian (Egerian–Early Karpatian)

This was the time of the major Alpine orogenic phase, the formation of mountains in the Alpine-Carpathian area, the Mediterranean, Central Asia and the Himalayas (Figs 17, 17a, 18). The morphostructural character of the Carpathians is a product of Tertiary evolution, finished up by the Neogene (Neoalpine) collisional events, which imprinted the present-day shape to the Carpathian orogenic belt (Golonka *et al.*, 2005; and references therein). Acceleration of ocean floor spreading in the Atlantic Ocean was the engine of the Alpine orogenesis and led to the convergence of the Afro-Arabian lithospheric plate including its promontory and the North European plate. The space between these two megablocks where Apulia and Alcapa were located was broken up into several translating and rotating microplates, reflecting squeezing caused by northward propagation of the African plate. Within the structure of the Carpathian tectogene, several typical distinct morphotectonic features can be distinguished, such as: the arc-shaped orogenic belt, core mountains alternating with intramontane basins, well-developed Miocene volcanic arc, structures generated by intra-orogen rotations, the Pieniny Klippen Belt (Andrusov, 1938) structure, fan-like structures and intra-orogen bending. All these phenomena are believed to be products of the youngest, Neogene (Neoalpine) tectonic period. Faulting played a tremendous role during the Neogene tectonic evolution. The dense and regular fault network is one of the characteristic features of the Carpathians. Brittle faults, mainly strike-slip ones, in combination with other dynamic tectonic boundaries allowed for propagation of individual detached blocks to the realm of the future Carpathian region (Fig. 18).

Convergence continued in the area between Africa and Eurasia during the Chattian–Aquitanian time. The movement of Corsica and Sardinia caused the plates to push eastwards in the future, resulting in deformation of the Alpine-Carpathian system. This deformation reached as far as the Romanian Carpathians (Royden, 1988; Ellouz & Roca, 1994) and continued throughout the Neogene. The Apulia and the Alpine-Carpathian terranes were moving northwards, colliding with the European plate, until 17 Ma (Decker & Peresson, 1996). The crust of the northward propagating Apulia converged with that of the Bohemian Massif. Terranes in-between Apulia and the stable European plate were extremely shortened and uplifted. For example, the lowermost crystalline basement was exhumed in the Tauern Window within the Eastern Alpine plate. The eastward movement caused by the Sardinia-Corsica pushing force was combined with the process of lateral extrusion and gravitational spreading of the elastic Carpatho/Pannonian crustal units to the east (Ratschbacher *et al.*, 1991, 1993). This so-called tectonic escape to the still free space within the Carpatho/Pannonian realm was controlled by strike-slip faults operating as border faults of crustal wedges moving to the east (Fig. 1). The already consolidated palaeo-mesoalpine structures were reworked during this process. Within active strike-slip shear zones in the regime of dextral transpression (Marko *et al.*, 1991; Golonka *et al.*, 2005), Early Miocene wrench furrows were formed at the western and eastern part of Slovakia (Vass, 1998). These E–W trending narrow and long furrows led to the development of basins filled up by poorly sorted and shortly transported sediments of the Eggenburgian–Early Karpatian age. After basin formation, the dextral transpression within strike-slip corridors was converted to a sinistral one, and the basins rotated in the ENE–WSW direction. This rotation did not exceed 15° in the counterclockwise direction.

The Apulia and Alpine-Carpathian terrane collision with the European plate also caused the foreland to propagate northward. The north to NNW-vergent thrust system of the Eastern Alps was formed. Oblique collision between the North European plate and the overriding Western Carpathian terranes led to the development of the outer accretionary wedge, as well as the formation of flysch nappes and the foredeep (Kováč *et al.*, 1993, 1998; Ślączka, 1996a, b; Oszczypko *et al.*, 2005). These nappes were detached from their original basement and thrust over the Palaeozoic–Tertiary deposits of the North European Platform (Figs 3, 4, 11). This process was completed in the Vienna Basin area and then progressed northeastwards (Oszczypko, 1997; Golonka *et al.*, 2005; Oszczypko *et al.*, 2005; Picha *et al.*, 2005). After the Late Oligocene folding, the Magura Nappe was thrust northward in the direction of the terminal Krosno flysch basin (Oszczypko, 1998). This synorogenic basin with flysch sedimentation of the Krosno beds was formed during the Oligocene as a continuation of the older Fore-Magura (with the outer part of Magura), Dukla, Silesian-Subsilesian, and Skole-Tarcau Outer Carpathian units. Initial folding in this zone also occurred.

During the Early Burdigalian, the front of the Magura Nappe reached the southern part of the Silesian Basin (Fig. 17) (Oszczypko, 1998). This was followed by a progressive northward migration of the axis of subsidence. During the course of the Burdigalian transgression, part of the Magura Basin was flooded and a narrow seaway connection with the Vienna Basin via Orava was probably established (Poprawa *et al.*, 2002; Oszczypko *et al.*, 2003). The thinned continental crust of the residual flysch basin was underthrust beneath

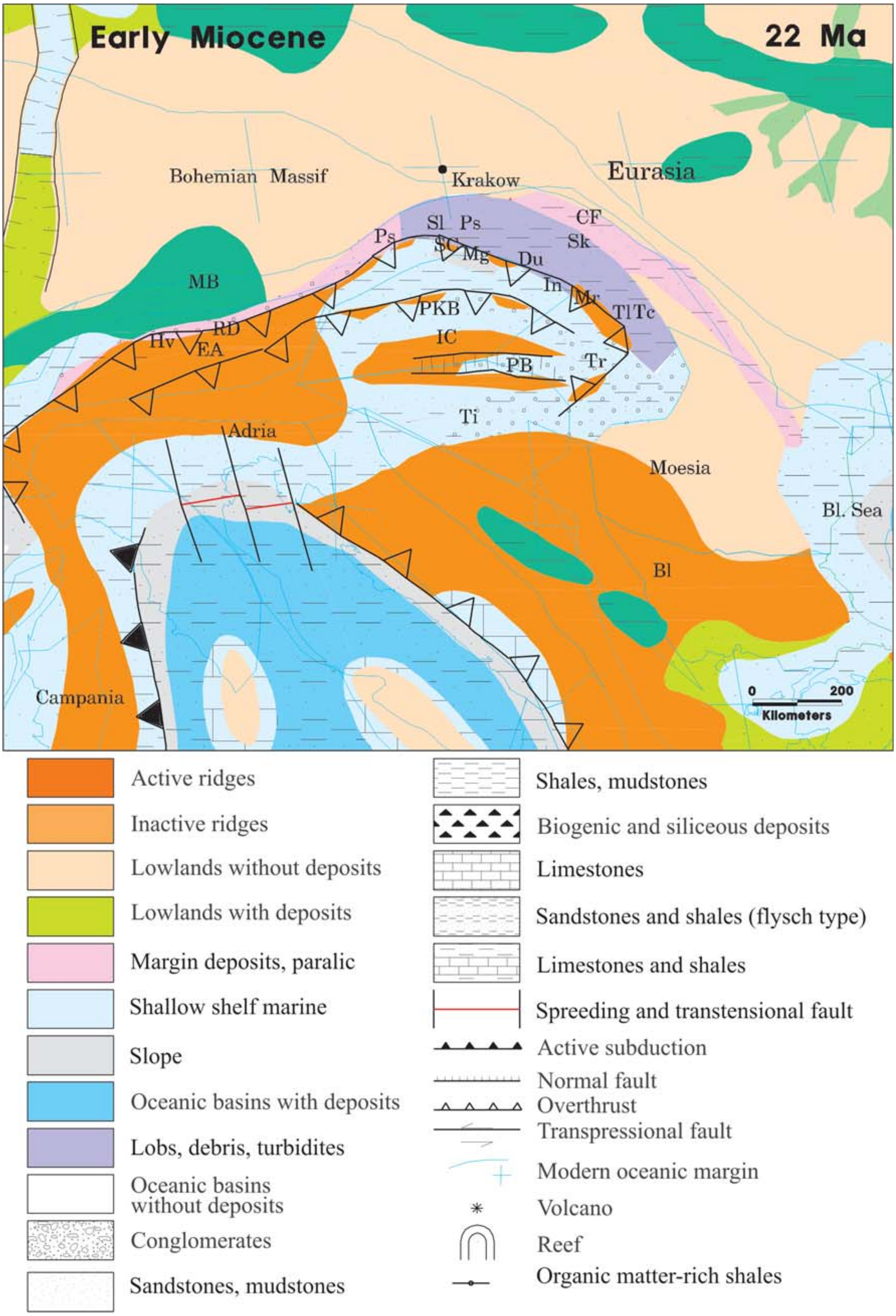

Fig. 17. Palaeoenvironment and lithofacies of the circum-Carpathian area during the Early Miocene (Chattian–Aquitanian); plates position at 22 Ma (modified from Golonka *et al.*, 2000, 2003). Abbreviations: Bl – Balkan foldbelt, CF – Carpathian Foredeep, Du – Dukla Basin, EA – Eastern Alps, Hv – Helvetic shelf, IC – Inner Carpathians, In – Inacovce-Krichevo zone, MB – Molasse Basin, Mg – Magura Basin, Mr – Marmarosh Massif, PB – Pannonian Basin, PKB – Pieniny Klippen Belt, Ps – Sub-Silesian Ridge and slope zone, RD – Rheno-Danubian, SC – Silesian Ridge (Cordillera), Si – Silesian Basin, Sk – Skole Basin, Tc – Tarcau Basin, Ti – Tisa plate, Tl – Teleajen Basin, Tr – Transylvanian Basin

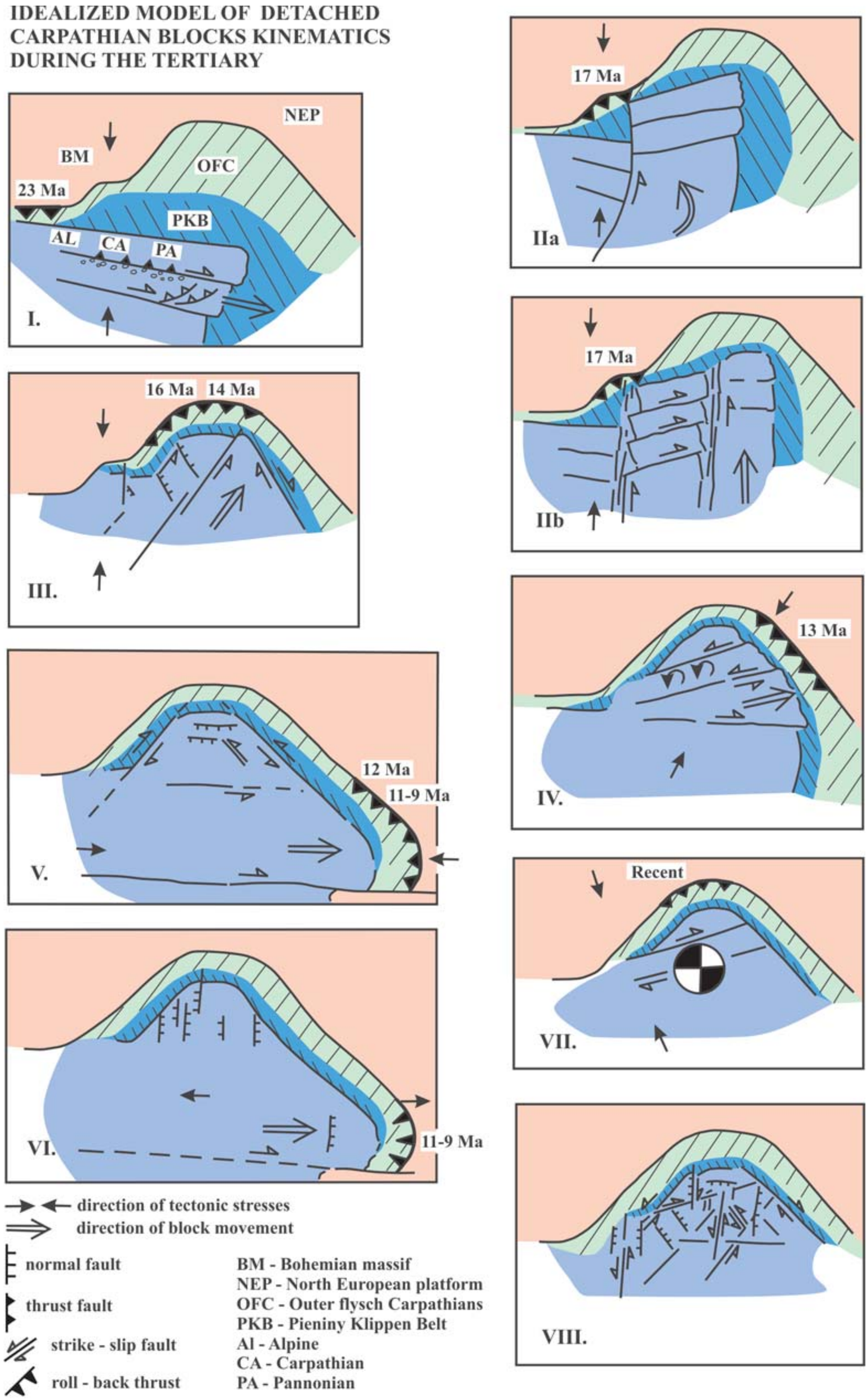

Fig. 18. Idealized model of detached Carpathian blocks kinematics during the Palaeogene and Neogene (after Golonka *et al.*, 2005)

the overriding Carpathian orogen. This underthrusting was connected with the intra-Burdigalian folding and uplift of the Outer Carpathians. The formation of the West Carpathian thrusts was completed (Kováč *et al.*, 1993, 1998; Ślączka, 1996a, b; Oszczypko, 1997). The Carpathians continued overriding the Eurasian platform and caused flexural depression – a peripheral foreland basin related to the prograding Carpathian front (Oszczypko, 1998). The thrust front was still migrating eastwards in the Eastern Carpathians.

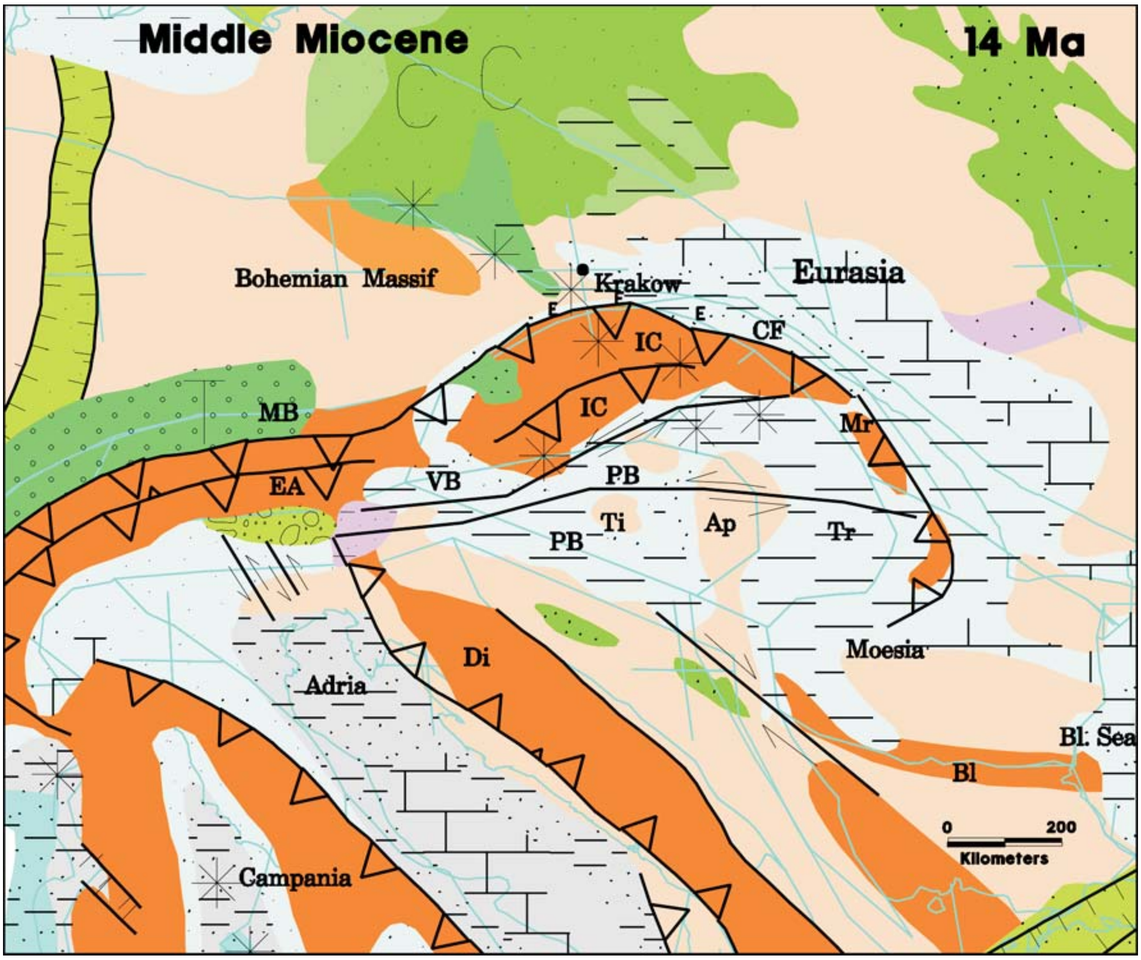

Fig. 19. Palaeoenvironment and lithofacies of the circum-Carpathian area during the Middle Miocene (Burdigalian– Serravallian); plates position at 14 Ma (modified from Golonka *et al.*, 2000, 2005). Abbreviations: Ap – Apuseni, Bl – Balkan foldbelt, CF – Carpathian Foredeep, EA – Eastern Alps, IC – Inner Carpathians, MB – Molasse Basin, Mr – Marmarosh Massif, PB – Pannonian Basin, Ti – Tisa plate, Tl – Teleajen Basin, Tr – Transylvanian Basin, VB – Vienna Basin

Late Burdigalian–Langhian (Late Karpatian–Early Badenian)

The northward movement of the Alpine segment of the Carpathian-Alpine orogen was stopped due to a collision with the Bohemian Massif (Figs 17–19). At the same time, the extruded Carpatho/Pannonian units were pushed to the open space, towards a bay of weak crust filled up by the Outer Carpathian flysch sediments. The separation of the Carpatho/Pannonian segment from the Alpine one and its propagation to the north was related to the development of N–S dextral strike-slip faults. These faults are known from the Outer Carpathians and inferred for the basement of the Vienna Basin below Neogene sediments (Krs & Roth, 1979; Balla, 1987). The formation of a spectacular oroclinal loop began at that time (Fig. 18). The loop has been traditionally regarded as a ductile structure due to the bend shape of the Alpine units. A gradual formation of the Carpathian loop due to independent propagation of detached crustal segments to the Carpathian gulf, rimmed by the Bohemian Massif, Tornquist-Teisseyre line and Moesia, is also possible.

The Carpathian segment of Alcapa moved to the north during the Late Karpatian. The detached slab of the crust, underneath the northward progressing continental Carpatho/Pannonian blocks (Doglioni *et al.*, 1991), was partially melted and produced early volcanic activity of the Western Carpathians. Distribution of sedimentation depocenters in the Vienna Basin shows a dramatic change of the structural plan during the Late Karpatian. Structural records indicate the change of compression direction from N–S to NNE–SSW, which resulted in a change of fault-controlled sedimentation. The NE–SW-orientated basins overprinted the older, E–W-trending ones. The NE–SW faults were activated as dominant normal and sinistral strike-slip ones. These faults controlled sedimentation in the Vienna Basin and simultaneously allowed for northeastward escape of the detached blocks after their collision with the platform to the north. During the Late Karpatian, the first core mountains with exhumed crystalline basement emerged, as indicated by fission track ages (Král, 1977; Kováč *et al.*, 1994) and the perifault debris sedimentation along the uplifted mountains (Vass, 1998). The ENE–WSW orientated wrench corridor was active as a sinistral transtensional wrench zone. The former Early Miocene wrench furrows were broken up, and separated remnants were rotated approximately 30°–40° counterclockwise (Kováč & Tunyi, 1995). These rotations were coeval with those described in the area the between the Carpathians and the Pannonian Basin (Pelso unit; Márton & Fodor, 1995).

Serravallian (Badenian)–Recent

During the Badenian (Serravallian) time (Figs 18, 19), the Vienna Basin was opened due to pull-apart activity of major, NE–SW-orientated sinistral strike-slip faults (Royden *et al.*, 1982; Royden, 1985). The Badenian depocenters showed a N–S trend due to normal faults which accommodated large magnitudes of strike-slip translations. Nevertheless, the Badenian Vienna Basin had not all attributes of a pull-apart basin. The western part of the basin lies over the accretionary flysch wedge, somehow in a piggy-back position, while the eastern part lies over the Alpine-Carpathian units deformed during the Mesozoic time (internides). The Vienna Basin displays some attributes of a fore-arc basin related to the Neogene volcanic arc.

The Carpathian foreland basin continued its development partly on top of the thrust front, with mainly terrestrial deposits forming the clastic wedge. This clastic wedge along the Carpathians could be compared to the lower fresh-water molasse of the Alpine Foreland Basin. During the Serravallian, the marine transgression flooded the foreland basin and adjacent platform.

Crustal extension of the internal zone of the Alps started in the Early Miocene, during the continued thrusting (Decker & Peresson, 1996). While transtensional fore-arc basins were formed at the northern edge of the Carpathian internides, Early to Middle Miocene extension and back-arc type rifting resulted in the formation of an intramontane Pannonian Basin (Royden, 1988; Decker & Peresson, 1996). The formation of the Pannonian Basin is related to thermal lithospheric stretching and volcanism provoked by heat of the rising Pannonian astenolith, which was formed in the Pannonian area.

According to palaeomagnetic data, the last counterclockwise block rotations ceased in the western part of the Western Carpathians in the Middle Badenian (Serravallian), while the last rotations in the eastern part of the Western Carpathians were recorded in the Tortonian (Sarmatian; Orlický, 1996). The NE–SW direction of compression and movement of the Carpatho/Pannonian lithospheric blocks is typical for the Serravallian time. During this period, a substantial part of volcanic activity had been accomplished and the Carpatho-Pannonian volcanic arc was formed. A huge bulk of volcanism was a result of the melting of subducted oceanic/quasi-oceanic crustal slab, as well as further rising of the astenolith in the Pannonian region (Póka, 1988; Lexa *et al.*, 1993). Crustal stretching due to astenospheric heat was accommodated by extensional normal faulting (Tari *et al.*, 1992; Horváth & Cloething, 1996; Tari & Horvath, 2005), marking the advanced stage of development of back-arc basins (Pannonian and Danube basins), under conditions of thermal crustal extension.

After the Serravallian, the sea retreated from the Carpathian peripheral foreland basin. This was followed by the last overthrust of the Carpathians toward their present-day position (Oszczypko, 1998). The Carpathian lithospheric block propagated to the east until the collision with the East-European Platform around 9 Ma ago. This collision produced a E–W compressional event (Decker & Peresson, 1996), which even affected such distal terranes as the Western Carpathians (Fig. 18). The E–W normal faults were reactivated within this stress period and controlled formation of the E–W trending horst and depression morphostructures. During the Tortonian–Gelasian time, the Carpathian thrusting progressed east and southeastwards, showing a strong element of translation (Royden, 1988; Ellouz & Roca, 1994; Linzer, 1996; Golonka *et al.*, 2005). The thrusting was completed during the Pliocene–Quaternary in the Moldavides in Romania. The eastward progression of the orogen was related to the opening of the Tyrrhenian Sea (Golonka *et al.*, 2000). The extrusion caused by the collision of Apulian plate with Europe could have also played a role in the eastward movement of the orogen (Decker & Peresson, 1996).

The SW–NE direction of the compression and the movement of the Carpatho/Pannonian lithospheric blocks is typical for the Tortonian time. An active front of the orogen was removed far from the Western Carpathian area. The NE–SW compression was gradually inverted to extension. The E–W extension was induced in the Carpathian realm due to the westward movement of the subducting plate (Royden *et al.*, 1982; Doglioni *et al.*, 1991) and related rollback effect. Numerous new N–S-orientated faults developed and many old strike-slip faults were reactivated as normal faults within this youngest tensional event. Normal faults played a dominant role as sedimentation controlling structures. These faults are the most numerous and conspicuous brittle features within the recent architecture of the Western Carpathians.

Lacustrine modest sedimentation is typical for the Pliocene time. An important regional tectonic stress, generated by the subduction/collision of large lithospheric plates ceased. The Carpathian loop, including the Pieniny Klippen Belt, was formed. Structural records within the Western Carpathians and Pannonian area indicate that approximately N–S (NNW–SSE) orientated compression affected the orogen during the last period of the Carpathian evolution (Csontos *et al.*, 1991; Becker, 1993; Hók *et al.*, 1995) and survived until recent time. An analysis of focal mechanisms of earthquakes in seismogenic fault zones supports interpretation of such a recent compression.

CONCLUSIONS: THE PROPOSED LOCATION OF THE ODDP DRILLING

A significant progress in the investigations of geodynamics and plate tectonics of the Carpathian orogen was achieved in the last years. This progress that was summarized in the present paper enabled us to formulate new scientific ideas and highlight problems that could be solved by the scientific drilling. These questions and problems can be grouped into several disciplinary topics, which combine both regional and continental-scale issues. As was mentioned in the Introduction chapter, many questions remain open. We think that without additional direct geological data, which can be achieved only by deep drilling under the

Orava Deep Drilling Project Well (*International Continental Deep Drilling Program*), these questions cannot be fully and properly answered. The summary of nine topics listed at the beginning of this paper is as follows:

TOPIC 1. Structural position of the Pieniny Klippen Belt (PKB) within the Carpathian Fold Belt (CFB) and its meaning for a reconstruction of the Cenozoic Alpine system of Europe. The PKB, which is one of the most complicated and enigmatic structures within the CFB is bounded by probably first-order strike-slip faults both from the North and the South. In-depth structure and geometry of the PKB is unknown. In order to solve fundamental obstacles in the reconstruction of CFB evolution, the following points need to be addressed:

(1) structural relations between the PKB and adjacent formations;

(2) nature of the contact zones, magnitude and relative displacement between contacting units;

(3) models' verification, identification of the subduction type (A or B), and direction within the CFB;

(4) geometry of the PKB suture at depth;

(5) total thickness of the Lower Miocene deposits in the context of the evolution of the foreland basin;

(6) identification of the tectonic units under the Magura nappe's overthrust;

(7) palaeogeographic disposition of the PKB Basin.

TOPIC 2. Relationship between geotectonic and geodynamic setting and magmatogenesis. The Flysch Belt, adjacent to the PKB from the North (Fig. 2), contains also andesitic Neogene (Sarmatian) volcanic rocks, whose origin is interpreted as a combined effect of the mantle uplift and subduction processes. In effect, extensional regime is exerted within generally compressional stress field, a situation unique within the Alpine system of Europe, with yet not fully recognised impact for the orogenic evolution. A temporal and spatial definition of the igneous activity, together with a precise geochemical and petrological characterization of whole rocks and mineral phases is of fundamental importance for solving the relative contribution of the asthenosphere and lithosphere to the magmatic products emplaced during the Alpine orogenesis. Worth to note is the presence in the study area of OIB-like (OIB = Ocean Island Basalt) mantle melts with extreme geochemical and isotopic features (e.g., extremely unradiogenic $^{87}Sr/^{86}Sr$, coupled with radiogenic $^{143}Nd/^{144}Nd$ and $^{206}Pb/^{204}Pb$, and $^{208}Pb/^{204}Pb$) among the Cenozoic European Volcanic Province (CEVP) products.

TOPIC 3. Nature of the geophysical anomalies. Since the late 1960s, a first-order magnetotelluric (M-T) anomaly north of the Tatra Mts, running parallel to the Carpathian axis, has been recognised. However, the nature of this anomaly still remains a puzzle. Essentially, two competing hypothesis do exist: the origin of M-T anomaly in the Carpathians has been related to the presence of either graphite or mineralized waters at depth. The origin of the Carpathian negative gravity anomaly, the axis of which is oblique with respect to the PKB, is also unclear.

TOPIC 4. Geothermal issues. This topic has a significant socio-economic implication and is closely related to the TOPIC 3. If MT anomaly is caused by low-resistivity brines, then ODDP may turn out very important for the discovery of new renewal energy resources.

TOPIC 5. Geodynamic reconstruction of the Mesozoic–Cenozoic basins. Particularly important are time and tectonic contexts of the development of the CFB basins, followed by palinspastic restoration and palaeogeographic reconstructions showing a provenance of these basins with respect to other crustal blocks, currently incorporated into the European Alpine fold belt. A better correlation of the main structural units of the Western and Eastern Carpathians is one of the main goals. Verification requires the occurrence and age of the supposed oceanic floor in these basins to be recognised. Controversial stratigraphic-facies themes (especially correlation of several flysch successions of the Outer Carpathians), as well as relation of subduction processes to the transform strike-slip movements need to be clarified.

TOPIC 6. Oil generation, migration and timing. Identification of petroleum systems of the Western Carpathians and their basement, and formulation of a general model for the other oil and gas reservoirs in the thrustbelt setting is scheduled. Rock magnetic and palaeomagnetic studies may help in dating of the hydrocarbon migration.

TOPIC 7. Regional heat-flow evolution. The border of the European hotspot field is getting through the Carpathian area. This border, together with the entire Central and Western Europe, is marked by Neogene volcanism and high heat flow. Within the Carpathians, the heat flow peak is now located in the Pannonian Basin. Notwithstanding, the Neogene volcanic rocks of the Transcarpathian Ukraine are placed on the Alcapa plate, while the PKB andesites and the Silesian basalts are located entirely within the European plate and cut obliquely the Outer Carpathian nappes. This distribution may point to the ancient position of the centre of the heat flow anomaly: the shift towards the Pannonian Basin could be due to the northerly drift of the European plate, overriding a thermal anomaly. Palaeotemperature indicators within the CFB may help in solving this topic.

TOPIC 8. Identification and definition of the Cadomian-Hercynian basement structure of the Carpathians. Palaeogeographic reconstructions of the basement terranes, reconstruction of the Hercynides during the circum-Carpathian geodynamic evolution, and mutual relations of the fore-Alpine terranes will be addressed in detail.

TOPIC 9. Palaeostress evolution and its changes in the horizontal and vertical sections. Reconstruction of palaeostress fields associated with the Western Carpathian development, geodynamic and palinspastic reconstruction of the Outer Western Carpathians and the Pieniny Klippen Belt during the Miocene should be considered. The present-day stress field changes in a vertical section throughout the Outer Western Carpathians, and differences between surface measurements, measurements in the flysch zone nappe pile, and in the slopes of the European Platform are evident.

Based on the current stage of knowledge, if the basement of the Alpine orogen is to be reached and the rocks responsible for MT anomaly would also be recognised, the main drilling should reach 8,000 m. Tentatively, we presume that the ODDP should be located in the vicinity of Chyżne village, near the Polish-Slovak border, to serve both Polish and Slovak scientists and to utilize the international cooperation. It would be also situated along the CELEBRATION 2000 deep seismic profile in the near vicinity of the deep geological cross-sections Kraków-Zakopane and Andrychów-Chyżne, utilizing a series of the Polish Geological Institute and Polish Oil Industry wells.

Acknowledgements

This work was supported in part by the grants from the AGH University of Science and Technology (grant 11.11.140.159), the University of Wrocław (grants: 1017/S/ING, 2022/W/ING1017/S/ING, and 2022/W/ING), and the Polish Committee for Scientific Research (KBN grants nos. 3 P04D 020 22, 4T12B 025 28, and 9T12B 03217). The Zakopane workshop, when many of the presented ideas were born, was supported by the International Continental Deep Drilling Program. An abridged version of this paper was presented at the 5th National Conference *"Neotectonics of Poland"*, which was organized in Kraków, 26-27 Sept. 2003, by the Commission for Neotectonics of the Committee for Quaternary Studies of the Polish Academy of Sciences, the Institute of Geological Studies of the Jagiellonian University, and the Galicia Tectonic Group.

The authors would like to express their gratitude to Piotr Krzywiec and Dušan Plašienka for constructive remarks, which really helped to improve the original mauscipt, and to Witold Zuchiewicz for his tedious and patient editorial work.

REFERENCES

Aleksandrowski, P., 1985. Structure of the Mt. Babia Góra region, Magura nappe, Western Outer Carpathians: an interference of West and East Carpathian fold trends. (In Polish, English summary). *Annales Societatis Geologorum Poloniae*, 55: 375–422.

Aleksandrowski, P., 1989. Structural geology of the Magura nappe in the Mt. Babia Góra region, Western Outer Carpathians. (In Polish, English summary). *Studia Geologica Polonica*, 96: 1–140.

Andrusov, D., 1938. Geologický výskum vnitřního bradlového pásma v Západních Karpatech. III. Tektonika. (In Czech). *Rozpravy Štátnego geologickego Ústavu ČSR*, 9: 1–135.

Andrusov, D., 1959. *Geológia československých Karpát II*. (In Czech). SAV, Bratislava, 376 pp.

Andrusov, D., Bystrický, J. & Fusán, O., 1973. *Outline of the structure of the West Carpathians*. *Guide book, X Congress CBGA*, GÚDŠ, Bratislava, 44 pp.

Aubrecht, R., Mišík, M. & Sýkora, M., 1997. Jurassic synrift sedimentation on the Czorsztyn Swell of the Pieniny Klippen Belt in Western Slovakia. In: Plašienka, D., Hók, J. Vozár, J. & Elečko, M. (eds), *Alpine evolution of the Western Carpathians*. Geological Survey of Slovak Republic, Bratislava: 53–64.

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

Bac-Moszaszwili, M., 1993. Structure of the western termination of the Tatra massif. (In Polish, English summary). *Annales Societatis Geologorum Poloniae*, 63: 167–193.

Balla, Z., l987. Tertiary paleomagnetic data for the Carpatho-Pannonian region in the light of Miocene rotation kinematics. *Tectonophysics*, l39: 67–98.

Baumgart-Kotarba, M., 1992. The geomorphological evolution of the intramontane Orava Basin associated with neotectonic movements, Polish Carpathians. *Studia Geomorphologica Carpatho-Balcanica*, 25: 3–28.

Baumgart-Kotarba, M., 1996. On origin and age of the Orava Basin, West Carpathians. *Studia Geomorphologica Carpatho-Balcanica*, 30: 101–116.

Baumgart-Kotarba, M., 2001. Continuous tectonic evolution of the Orava basin from Late Badenian to the present-day. *Geologica Carpathica,* 52: 103–110.

Baumgart-Kotarba, M., Marcak, H. & Márton, E., 2004. Rotation along the transverse transforming Orava strike-slip fault: based on geomorphological, geophysical and paleomagnetic data (Western Carpathians). *Geologica Carpathica*, 55: 219–226.

Bąk, K., 2000. Biostratigraphy of deep-water agglutinated Foraminifera in Scaglia Rossa-type deposits, the Pieniny Klippen Belt, Carpathians, Poland. In: Hart,. M., Smart, C. & Kaminski, M. A. (eds), *Proceedings of the Fifth International Workshop on Agglutinated Foraminifera*. Grzybowski Foundation Special Publication, 7: 15–41.

Becker, A., 1993. Contemporary state of stress and neotectonic deformation in the Carpathian-Pannonian region. *Terra Nova* 5: 375–388.

Behrmann, J. H., Stiasny, S., Milicka, J. & Pereszlenyi, M., 2000. Quantitative reconstruction of orogenic convergence in the northern Carpathians. *Tectonophysics*, 319: 111–127.

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., 1958. Submarine erosional breaks and Late Jurassic synorogenic movements in the Pieniny Klippen Belt geosyncline. *Bull. Acad. Polon. Sci., Sér., sci. chim., géol. et géogr.*, 8: 551–558.

Birkenmajer, K., 1977. Jurassic and Cretaceous lithostratigraphic units of the Pieniny Klippen Belt, Carpathians, Poland. *Studia Geologica Polonica*, 45: 1–158.

Birkenmajer, K., 1983. Strike-slip faults in the northern boundary zone of the Pieniny Klippen Belt, Carpathians. (In Polish, English summary). *Studia Geologica Polonica*, 77: 89–112.

Birkenmajer, K., 1986. Stages of structural evolution of the Pieniny Klippen Belt, Carpathians. *Studia Geologica Polonica*, 88: 7–32.

Birkenmajer, K., 1988. Exotic Andrusov Ridge: its role in plate-tectonic evolution of the West Carpathian Foldbelt. *Studia Geologica Polonica*, 91: 7–37.

Birkenmajer, K., 2003. Post-collisional Late Middle Miocene (Sarmatian) Pieniny Volcanic Arc, Western Carpathians. *Bulletin of the Polish Academy of Sciences, Earth Sciences*, 51: 79–89.

Birkenmajer, K. & Oszczypko, N., 1989. Cretaceous and Palaeogene lithostratigraphic units of the Magura Nappe, Krynica Subunit, Carpathians. *Annales Societatis Geologorum Poloniae*, 59: 145–181.

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.

Birkenmajer, K. & Pécskay, Z., 2000. K-Ar dating of the Miocene andesite intrusions, Pieniny Mts, West Carpathians, Poland: a supplement. *Studia Geologica Polonica*, 117: 7–25.

Burchart, J., 1970. Rocks of the Goryczkowa "Crystalline Island" in the Tatra Mountains. *Studia Geologica Polonica*, 32: 1–160.

Burda, J. & Gawęda, A., 1997. Mass-balance calculations in migmatites from the Upper Kościeliska Valley (The Western Tatra Mts. S-Poland). *Mineralogia Polonica*, 28: 53–68.

Burda, J. & Gawęda, A., 1999. Petrogenesis of migmatites from the Upper Kościeliska Valley (Western Tatra Mountains). *Archiwum Mineralogiczne*, 52: 163–194.

Chowaniec, J., 1989. *Hydrogeologiczne warunki zasilania i przepływu wód podziemnych w utworach trzeciorzędowych Podhala między Zakopanem a Białym*. (In Polish). Unpublished Ph. D. thesis. Centralne Archiwum Geologiczne Państwowego Instytutu Geologicznego, Warszawa, 144 ms. pp.

Chowaniec, J. & Kępińska, B., 2003. Podhale geothermal system – selected issues. In: Golonka, J. & Lewandowski, M., (eds), *Geology, geophysics, geothermics and deep structure of the West Carpathians and their basement*. Publications of the Institute of Geophysics, Polish Academy of Sciences, Monographic Volume, 28 (363): 13–23.

Chrustek, M., 2005. Morphotectonics of the Orava Basin, Western Carpathians. *GeoLines*, 19: 26–27.

Cieszkowski, M., 1985. Stop 21: Obidowa. In: Birkenmajer, K. (ed.), *Main geotraverse of the Polish Carpathians (Cracow-Zakopane), Guide to excursion 2, Carpatho-Balkan Geological Association, 13 Congress, Cracow, Poland*. Geological Institute, Warszawa: 54–58.

Cieszkowski, M., 1992. Michalczowa zone: a new unit of the Fore-Magura Zone, Outer Carpathians, South Poland. (In Polish, English summary). *Geologia, Kwartalnik AGH,*, 18: 1–125.

Cieszkowski, M., 1995. Marine Miocene deposits close to Nowy Targ and their importance for determining age of the Orava – Nowy Targ basin, south Poland. (In Polish, English summary). *Geologia, Kwartalnik AGH*, 21: 153–168.

Cieszkowski, M., 2001. Fore-Magura Zone of the Outer Carpathians in Poland. *Biuletyn Państwowgo Instytutu Geologicznego*, 396: 32–33.

Cieszkowski, M., 2003. The Outer Carpathians thrustbelt. In: Golonka, J. & Lewandowski, M. (eds), *Geology, geophysics, geothermics and deep structure of the West Carpathians and their basement*. Publications of the Institute of Geophysics, Polish Academy of Sciences, Monographic Volume, 28 (363): 107–110.

Cieszkowski, M., Durkovič, T., Jawor, E., Korab, T. & Sikora, W., 1981a. A new tectonic unit in the Polish and Slovak Flysch Carpathians. *Abstracts*, *Carpatho-Balkan Geological Association, 12 Congress, Bucharest, 1981*. Bucharest: 118–120.

Cieszkowski, M., Durkovič, T., Jawor, E., Korab, T. & Sikora, W., 1981b. Geological interpretation of the Obidowa-Słopnice tectonic unit in the Polish and Slovak Flysch Carpathians. *Abstracts, Carpatho-Balkan Geological Association, 12 Congress, Bucharest, 1981*. Bucharest: 272–274.

Cieszkowski, M., Golonka, J., Polak, A., Rajchel, J. & Ślączka, A., 2003. Olistostroms in the deep-water flysch deposits in the Polish Outer Carpathians. In: Vlahovic, I. (ed.), *Abstract Book. 22nd IAS Meeting of Sedimentology – Opatija 2003*. Opatija, Croatia, p. 66.

Cieszkowski, M., Oszczypko, N. & Zuchiewicz, W., 1989. Upper Cretaceous siliciclastic-carbonate turbidites at Szczawa, Magura Nappe, West Carpathians, Poland. *Bulletin of the Polish Academy of Sciences, Earth Sciences*, 37: 231–245.

Cieszkowski, M., Ślączka, A. & Wdowiarz, S., 1985. New data on structure of the Flysch Carpathians. *Przegląd Geologiczny*, 33: 313–333.

Ciulavu, D., & Bertotti, G., 1994. The Transylvanian Basin and its Upper Cretaceous Substratum: ALCAPA II. *Romanian Journal of Tectonics and Regional Geology*, 75: 59–65.

Csontos, L., Tari, G., Bergerat, F. & Fodor, L., 1991. Evolution of the stress fields in the Carpatho-Pannonian area during the Neogene. *Tectonophysics*, 199: 73–91.

Csontos, L. & Vörös, A., 2004. Mesozoic plate tectonic reconstruction of the Carpathian region. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 210: 1–56.

Czerwiński, T., Florek, R. & Klityński, W., 2002. Geophysical and geological interpretation of regional magnetotelluric survey in the Western Polish Carpathians. *Naukovij Visnik Ivano-Frankivs'kogo Natsional'nogo Technichnogo Universitetu Nafti i Gazu*, 3 (4): 97–103.

Czerwiński, T. & Stefaniuk, M., 2001. Recognition of geological structure of the Carpathians as a result of magnetotelluric investigations. *Slovak Geological Magazine*, 7: 139–144.

Decker, K. & Peresson, H., 1996. Tertiary kinematics in the Alpine-Carpathian-Pannonian system: links between thrusting, transform faulting and crustal extension. In: Wessely, G. & Liebl, W., (eds), *Oil and gas in Alpidic thrustbelts and basins of Central and Eastern Europe.* Special Publications of the European Association of Geoscientists and Engineers 5. Geological Society, London: 17–21.

Doglioni, C., Moretti, I. & Roure, F., 1991. Basal lithospheric detachment, eastward mantle flow and Mediterranean geodynamics: a discussion. *Journal of Geodynamics*, 13: 47–65.

Dżułyński, S. & Gradziński, R., 1960. Source of Lower Triassic clastics in the Tatra Mts. *Bulletin of the Polish Academy of Sciences, Earth Sciences*, 8: 45–48.

Ellouz, N. & Roca, E., 1994. Palinspastic Reconstruction of the Carpathians and Adjacent Areas since the Cretaceous: a Quantitative Approach. In: Roure, F. (ed.), *Peri-Tethyan Platform*. Éditions Technip, Paris: 51–77.

Ernst, T., Jankowski, J., Semenov, V. J., Adam, A., Hvozdara, M., Jóźwiak, W., Lefeld, J., Pawliszyn, J., Szarka, L. & Wesztergom, V., 1997. Electromagnetic soundings across the Tatra Mountains. *Acta Geophysica Polonica*, 45: 33–43.

Fejdiová, O., 1971. Sedimentary petrology of the Lower Triassic quartzites from the Vysoké Tatry and Malé Karpaty envelope unit. *Geologický Zborník – Geologica Carpathica*, 28: 49–77.

Gawęda, A., Goławska, B., Jędrysek, M., Leichman, J., Paulo, A. & Włodyka, R., 2001. Carbonate mineralisation in the Tatra Mts. crystalline basement. *Mineralogical Society of Poland, Special Papers*, 18: 39–42.

Gawęda, A., Kozłowski, K. & Piotrowska, K. 1998. Tectonic development of the crystalline basement of the Polish part of the Western Tatra Mts. *Acta Universitatis Carolinae, Geologica,* 42: 252.

Gawęda, A., Winchester, J. A., Kozłowski, K., Narębski, W. & Holland, J. G., 2000. Geochemistry and paleotectonic setting of amphibolites from the Western Tatra Mts., southern Poland. *Geological Journal*, 35: 69–85.

Ghosh, S. K. & Ramberg, H., 1968. Buckling experiments on intersecting fold patterns. *Tectonophysics*, 5: 89–105.

Golonka, J., 1981. *Objaśnienia do Mapy geologicznej Polski 1:200 000, arkusz Bielsko-Biała*. (In Polish). Wydawnictwa Geologiczne, Warszawa, 80 pp.

Golonka, J., 2003. Plate tectonics of the circum-Carpathian area and the ultra deep drilling proposal. In: Golonka, J. & Lewandowski, M., (eds), *Geology, geophysics, geothermics and deep structure of the West Carpathians and their basement*. Publications of the Institute of Geophysics, Polish Academy of Sciences, Monographic Volume, 28 (363): 113–125.

Golonka, J., 2004. Plate tectonic evolution of the southern margin of Eurasia in the Mesozoic and Cenozoic. *Tectonophysics*, 381: 235–273.

Golonka, J., 2005. Mesozoic plate tectonics of the Inner Carpathians – rotational approach. *GeoLines*, 19: 42–43.

Golonka, J., Gahagan, L., Krobicki, M., Marko, F., Oszczypko, N.

& Ślączka, A., 2005. Plate Tectonic Evolution and Paleogeography of the Circum-Carpathian Region. In: Picha, F. & Golonka, J. (eds), *The Carpathians and their foreland: Geology and hydrocarbon resources*. American Association of Petroleum Geologists, Memoir, 84: 1–60.

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

Golonka, J. & Krobicki, M., 2004. Jurassic paleogeography of the Pieniny and Outer Carpathian basins. *Rivista Italiana di Paleontologia e Stratigrafia*, 110: 5–14.

Golonka, J., Krobicki, M., Oszczypko, N., Ślączka, A. & Słomka, T., 2003. Geodynamic evolution and palaeogeography of the Polish Carpathians and adjacent areas during Neo-Cimmerian and preceding events (latest Triassic–earliest Cretaceous). In: McCann, T. & Saintot, A. (eds), *Tracing Tectonic Deformation Using the Sedimentary Record*. Geological Society, London, Special Publications, 208: 138–158.

Golonka, J., Krobicki, M. & Tomaś, A., 2005. Paleogeografia Karpat centralnych i wewnętrznych – ujęcie rotacyjne. (In Polish). In: Cieszkowski, M. & Golonka, J. (eds), *Wapienie organogeniczne i organodetrytyczne w Karpatach zewnętrznych i ich znaczenie dla rekonstrukcji paleogeograficznych*. Seminarium naukowe, 21 kwietnia 2005, ING UJ, Kraków: 41–45.

Golonka, J. & Lewandowski, M. (eds), 2003. *Geology, geophysics, geothermics and deep structure of the West Carpathians and their basement*. Publications of the Institute of Geophysics, Polish Academy of Sciences, Monographic Volume, 28 (363), 184 pp.

Golonka, J., Oszczypko, N. & Ślączka, A., 2000. Late Carboniferous–Neogene geodynamic evolution and palaeogeography of the circum-Carpathian region and adjacent areas. *Annales Societatis Geologorum Poloniae*, 70: 107–136.

Golonka, J. & Rączkowski, W., 1984. *Objaśnienia do Szczegółowej mapy geologicznej Polski 1: 50 000, arkusz Piwniczna*. (In Polish). Wydawnictwa Geologiczne, Warszawa, 85 pp.

Golonka, J. & Sikora, W., 1981. Microfacies of the Jurassic and Lower Cretaceous sedimentarily thinned deposits of the Pieniny Klippen Belt in Poland. (In Polish, English summary). *Biuletyn Instytutu Geologicznego*, 31: 7–37.

Grabowski, J. & Nemčok, M., 1999. Summary of Paleomagnetic Data from the Central West Carpathians of Poland and Slovakia: Evidence for the Late Cretaceous–Early Tertiary Transpression. *Physics and Chemistry of the Earth*, 24: 681–685.

Grabowski, P., 1967. Budowa geologiczna choczańskich łusek Upłazu, Kończystej i Bramy Kantaka na wschód od Doliny Kościeliskiej. (In Polish, English summary). *Acta Geologica Polonica*, 17: 672–696.

Gross, P., Köhler, E., Mello, J., Haško, J., Halouzka, R., Nagy, A., Kováč, P., Filo, I., Havrila, M., Maglay, J., Salaj, J., Franko, O., Zakovič, M., Pospíšil, L., Bystrická, H., Samuel, O., Snopková, P., Potfaj, M., žecová, K. & Borová, D., 1993. *Geology of southern and eastern Orava*. (In Slovak, English summary). Geologický Ústav Dionyza Štura, Bratislava: 1–319.

Gross, P., Köhler, E. & Samuel, O., 1984. A new lithostratigraphical division of the Inner-Carpathian Paleogene. (In Slovak, English summary). *Geologicke práce – Správy*, 81: 103–117.

Guterch, A. & Grad, M. (reporters), 2001. New deep seismic studies of the lithosphere in Polonaise'97 and Celebration 2000 seismic experiments. *Biuletyn Państwowego Instytutu Geologicznego*, 396: 61.

Hefty, J., 1998. Estimation of site velocities from CEGRN GPS campaigns referred to CERGOP reference frame. *Publications of the Warsaw University of Technology, Institute of Geodesy and Geodetic Astronomy*, 9 (30): 67–79.

Horváth, F. & Cloetingh, S., 1996. Stress-induced late-stage subsidence anomalies in the Pannonian basin. *Tectonophysics*, 266: 287–300.

Hók, J., Šimon, L., Kováč, P., Elečko, M., Vass, D., Verbich, F. & Halmo, J., 1995. Tectonics of the Hornonitrianska kotlina Basin during Tertiary. *Geologica Carpathica*, 46: 191–196.

Iwanow, A. & Wieczorek, J., 1987. Problem najwyższych jednostek tektonicznych w Tatrach. (In Polish, English summary). *Przegląd Geologiczny*, 35 (10): 525–528.

Jachowicz, M., Bobiński, W., Buła, Z., Habryn, R., Markowiak, M., Żaba, J., Żelaźniewicz, A., 2002. *Geneza i pozycja stratygraficzna podkambryjskich i podordowickich anchimetamorficznych skał w południowej Polsce – przedpole orogenu neoproterozoicznego?* Raport końcowy: grant KBN 9T12B 03217. (In Polish). Unpublished report, pp. 1–118.

Janak, M., Hurai, V., Ludhova, L., O'Brien, P. J. & Horn, E. E., 1999. Dehydration melting and devolatilization during exhumation of high grade metapelites: the Tatra Mountains, Western Carpathians. *Journal of Metamorphic Geology*, 17: 379–395.

Janak, M., Kohut, M. & Poller, U., 2001. The Tatra Mountains: an example of lower-crustal melting and multistage granite evolution during Variscan orogeny. In: Petrik, I., Kohut, M. & Broška, I. (eds), *Granitic plutonism of the Western Carpathians: Characteristics and evolution*. Geological Institute, Slovak Academy Sciences, Geological Survey of Slovak Republic, Bratislava: 80–92.

Jankowski, J., 1967. The marginal structures of the East European Platform in Poland on basis of data of geomagnetic field variations. *Publications of the Institute of Geophysics, Polish Academy of Sciences*, 4: 93–102.

Jankowski J., Ney, R. & Praus, O., 1982. Do geothermal waters occur underneath the whole arc of the North-East Carpathians? (In Polish, English summary). *Przegląd Geologiczny*, 4: 165–169.

Jankowski, J., Pawliszyn, J., Jóźwiak, W. & Ernst, T., 1991. Synthesis of electric conductivity surveys performed on the Polish part of the Carpathians with geomagnetic and magnetotelluric sounding methods. *Publications of the Institute of Geophysics, Polish Academy of Sciences*, A-19 (236): 183–214.

Jankowski, J. & Praus, O., 2003. Carpathian anomaly of electrical conductivity: history of its discovery and its interpretation. In: Golonka, J. & Lewandowski, M. (eds), *Geology, geophysics, geothermics and deep structure of the West Carpathians and their basement*. Publications of the Institute of Geophysics, Polish Academy of Sciences, Monographic Volume, 28 (363): 127–128.

Jankowski, J., Tarłowski, Z., Praus, O., Pečova, J. & Peter, V., 1985. The results of deep geomagnetic sounding in the West Carpatians. *Geophysical Journal, Russian Academy of Sciences*, 80: 561–574.

Janočko, J. & Kováč, M., 2003. Geotectonic background of origin of Western Carpathian sedimentary basins. In: Janočko, J. & Elečko, M. (eds), *Tectono-sedimentary evolution of Western Carpathian Tertiary Basins*. Mineralia Slovaca, 36: 169–180.

Jurewicz, E., 1994. Structural analysis of the Pieniny Klippen Belt at Jaworki, Carpathians, Poland. (In Polish, English summary). *Studia Geologica Polonica*, 106: 7–87.

Jurewicz, E., 1997. The contact between the Pieniny Klippen Belt and Magura Unit (the Małe Pieniny Mts.). *Geological Quarterly*, 41: 315–326.

Jurewicz, E., 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.

Kępińska, B., 1997. Model geologiczno-geotermalny niecki podhalańskiej. (In Polish). *Studia, Rozprawy, Monografie*. *Wydawnictwo CPPGSMiE PAN, Kraków*, 48: 5–11.

Koszarski, A., 1999. The Jasło Nappe *versus* the Gorlice beds. A controversy at the front of the Magura Nappe in the Jasło-Gorlice region. (Polish Middle Carpathians). *Geologica Carpathica*, 50: 153–154.

Koszarski, L., Sikora, W. & Wdowiarz, S., 1974. The Flysch Carpathians. Polish Carpathians. In: Mahel, M. (ed.), *Tectonics of the Carpathian-Balkan Regions*. Geologický Ústav Dionyza Štura, Bratislava: 180–197.

Kotański, Z., 1959. Stratigraphy, sedimentology and paleogeography of the High-Tatric Triassic in the Tatra Mts. *Acta Geologica Polonica*, 9: 113–145.

Kotański, Z., 1961. Tektogeneza i rekonstrukcja paleogeografii pasma wierchowego w Tatrach. (In Polish, French summary). *Acta Geologica Polonica*, 11: 187–476.

Kotański, Z., 1973. Upper and Middle Subtatric nappe in the Tatra Mts. *Bulletin of the Polish Academy of Sciences, Earth Sciences*, 21: 75–83.

Kotański, Z., 1979. Trias tatrzański. (In Polish). *Przegląd Geologiczny*, 7: 369–377.

Kotański, Z., 1986a. Jeszcze raz o płaszczowinie strażowskiej, cz. 1. (In Polish). *Przegląd Geologiczny*, 10: 547–552.

Kotański, Z., 1986b. Jeszcze raz o płaszczowinie strażowskiej, cz. 2. (In Polish). *Przegląd Geologiczny*, 11: 621–628.

Kotlarczyk, J. (ed.), 1985. *Geotraverse Kraków - Baranów - Rzeszów - Przemyśl - Ustrzyki Dolne - Komańcza - Dukla. Guide to Excursion 4*. Carpatho-Balkan Geological Association XIII Congress, Cracow, Poland 1985. Geological Institute, Warszawa, 172 pp.

Kováč, M., Bielik, M., Pereszlényi, M., Šefara, I., Túnyi, I. & Vass, D., 1997. The Western Carpathian intramontane basins. In: Grecula, P., Hovorka, D. & Putiš, M. (eds), *Geological evolution of the Western Carpathians*. Mineralia Slovaca – Monograph, Bratislava: 43–64.

Kováč, M. & Hók, J., 1993. The Central Slovak Fault System: Field evidence of a strike-slip. *Geologica Carpathica*, 44: 155–160.

Kováč, M., Král, 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., Nagymarosy, A., Oszczypko, N., Ślączka, A., Csontos, L., Marunteanu, M., Matenco, L. & Márton, M., 1998. Palinspastic reconstruction of the Carpathian-Pannonian region during the Miocene. In: Rakús, M. (ed.), *Geodynamic development of the Western Carpathians*. Geological Survey of Slovak Republic, Dionýz Štúr Publishers, Bratislava: 189–217.

Kováč, M., Nagymarosy, A. Soták, J. & Šutovská, K., 1993. Late Tertiary Paleogeographic Evolution of the Western Carpathians. *Tectonophysics*, 226: 401–415.

Kováč, M. & Túnyi, I., 1995. Interpretácia paleomagnetických údajov zo západnej časti Centrálnych Západných Karpát. (In Slovak). *Mineralia Slovaca*, 27: 2l3–220.

Kozłowski, K. & Gawęda, A., 1999. Pre-Variscan metamorphic evolution of the crystalline basement of the Polish part of the Western Tatra Mountains. (In Polish, English summary). In: Jankowski, A. T. & Szczypek, T. (eds)*, Memorial of 25 years of geographical center at the University of Silesia, Sosnowiec*. University of Silesia, Special Issue, Katowice: 91–99.

Král, J., 1977. Fission track ages of apatites from some granitoid rocks in West Carpathians. *Geologica Carpathica*, 28: 269–276.

Krobicki, M., Kruglov, S. S., Matyja, B. A., Wierzbowski, A., Aubrecht, R., Bubniak, A. & Bubniak, I., 2003. Relation between Jurassic klippen successions in the Polish and Ukrainian parts of the Pieniny Klippen Belt. *Mineralia Slovaca*, 35: 56–58.

Krs, M. & Roth, Z., 1979. Insubrický karpatský trežohorný blokový systém, jeho vznik a rozloženie. (In Slovak). *Geologica Carpathica*, 30: 3–17.

Kruczyk, J., Kądziałko-Hofmokl, M., Lefeld, J. Pagac, P. & Túnyi, I., 1992. Paleomagnetism of Jurassic sediments as evidence for oroclinal bending of the Inner West Carpathians. *Tectonophysics*, 206: 315–324.

Kruglov, S. S., 1989. Geodynamics of the Ukrainian Carpathians. *Geologica Carpathica*, 40: 101–123.

Krzywiec, P. & Jochym, P., 1997. Characteristics of Miocene subduction of the Polish Carpathians: Results of flexural modelling. (In Polish, English summary). *Przegląd Geologiczny*, 45 (8): 785–792.

Książkiewicz, M., 1956. Geology of the Northern Carpathians. *Geologische Rundschau*, 45: 396–411.

Książkiewicz, M., 1977. The Tectonics of the Carpathians. In: Pożaryski, W. (ed.), *Geology of Poland, IV. Tectonics*. Wydawnictwa Geologiczne, Warszawa: 476–620.

Lefeld, J. (ed.), 1985. Jurassic and Cretaceous lithostratigraphic units in the Tatra Mountains. *Studia Geologica Polonica*, 84: 82–93.

Lefeld, J., 1999. Tectonics of the Subtatric Units, Eastern Tatra Mts. *Studia Geologica Polonica*, 115: 139–166.

Lefeld, J. & Jankowski, J., 1985. Model of deep structure of the Polish Inner Carpathians. *Publications of the Institute of Geophysics, Polish Academy of Sciences*, A-16 (175): 71–100.

Lewandowski, M., Krobicki, M., Matyja, B. A. & Wierzbowski, A., 2003. Palaeogeography and dynamics of the Eastern Pieniny Klippen Belt (Carpathians, Southwestern Ukraine) on the basis of palaeomagnetic data. In: Golonka, J. & Lewandowski, M. (eds), *Geology, geophysics, geothermics and deep structure of the West Carpathians and their basement*. Publications of the Institute of Geophysics, Polish Academy of Sciences, Monographic Volume, 28 (363): 137–142.

Lexa, J., Bzák, V., Elečko, M., Mello, J., Polák, M., Potfaj, M. & Vozár, J. (eds), 2000. *Geological map of the Western Carpathians and adjacent areas, 1: 500,000*. Geol. Survey of Slovak Republic, Bratislava.

Lexa, J., Konečný, V., Kaličiak, M. & Hojstričová, V. 1993. Distribúcia vulkanitov karpato-panónskeho regiónu v priestore a čase. (In Slovak). In: Rakús, M. (ed.), *Geodynamický model a hlbinná stavba Západných Karpát. Konf. – symp. – sem*. Geologický Ústav Dionyza Štura, Bratislava: 57–69.

Linzer, H.-G., 1996. Kinematics of retreating subduction along the Carpathian arc, Romania. *Geology*, 24: 167–170.

Mader, D., 1982. *Evolution of Palaeoecology and Palaeoenvironment of Permian and Triassic Fluvial Basins in Europe. Volume 1. Western and Eastern Europe*. Gustav Fischer Verlag, Stuttgart-New York, 738 pp.

Mahel, M., 1974. The Inner Carpathians. In: Mahel, M. (ed.), *Tectonics of the Carpathian-Balkan Regions*. Geologický Ústav Dionyza Štura, Bratislava: 91–133.

Malata, E., Malata, T. & Oszczypko, N., 1996. Litho- and biostratigraphy of the Magura Nappe in the eastern part of the Beskid Wyspowy Range (Polish Western Carpathians). *Annales Societatis Geologorum Poloniae*, 66: 269–283.

Małkowski, S., 1958. Przejawy wulkanizmu w dziejach geologicznych okolic Pienin. (In Polish, English summary). *Prace Muzeum Ziemi*, 1: 105–113.

Marko, F., 1996. Dynamic analysis of faulting in the Central-Carpathian Paleogene Basin, based on structural investigations in the Levočske Vrchy Mts. In: Sotak, J. (ed.), *Geological assessment of the Levočske vrchy Mts. – research report*. Geofond, Bratislava: 399–462.

Marko, F., 2003. Slickensides and related paleostress field in the western part of the Pieniny Klippen Belt. In: Golonka, J. & Lewandowski, M. (eds), *Geology, geophysics, geothermics and deep structure of the West Carpathians and their basement*. Publications of the Institute of Geophysics, Polish Academy of Sciences, Monographic Volume, 28 (363): 145–146.

Marko, F., Fodor, L. & Kováč, M., 1991. Miocene strike-slip faulting and block rotation in Brezovské Karpaty Mts. (Western Carpathians). *Mineralia Slovaca*, 23: l89–200.

Marschalko, R., 1968. Facies distributions, paleocurrents and paleotectonics of the Paleogene flysch of Central West-Carpathians. *Geologica Carpathica*, 19: 69–94.

Márton, E., 2003. Western Carpathians: Tertiary paleomagnetism and rotations. In: Golonka, J. & Lewandowski, M. (eds), *Geology, geophysics, geothermics and deep structure of the West Carpathians and their basement*. Publications of the Institute of Geophysics, Polish Academy of Sciences, Monographic Volume, 28 (363): 147–148.

Márton, E. & Fodor, L., 1995. Combination of paleomagnetic and stress data – a case study from North Hungary. *Tectonophysics*, 242: 99–114.

Márton, E., Tokarski, A. K. & Halasz, D., 2004. Late Miocene counterclockwise rotation of the Pieniny andesites at the contact of the Inner and Outer Western Carpathians. *Geologica Carpathica*, 55: 411–419.

Mastella, L., Ozimkowski, W. & Szczęsny, R., 1988. Tectonics of the northwestern Podhale flysch. (In Polish, English summary). *Przegląd Geologiczny*, 36: 566–572.

Matějka, A. & Roth, Z., 1950. Remark on the main tectonic units of the Magura Flysch in Czechoslovakia. (In Czech, English summary). *Vestnik Štatneho Geologickeho Ústavu*, 25: 301–308.

Matenco, L. & Bertotti, G., 2000. Tertiary tectonic evolution of the external East Carpathians (Romania). *Tectonophysics*, 316: 255–286.

Matenco, L., Zoetemeijer, R., Cloetingh, S. & Dinu, C., 1997. Lateral variations in mechanical properties of the Romanian external Carpathians: inferences of flexure and gravity modelling. *Tectonophysics*, 282: 147–166.

Meszaros, N., 1996. Stratigrafia regiunii Turnu Rosu-Porcesti. (In Romanian). *Convergente transilvane*, 4: 42–45, Sibiu.

Michalik, M. & Skublicki, Ł., 1999. Phosphate accessory minerals in High Tatra granitoids. *Geologica Carpathica*, 50, *Special Issue*: 123–125.

Mišík, M., 1994. The Czorsztyn submarine ridge (Jurassic–Lower Cretaceous, Pieniny Klippen Belt): an example of a pelagic swell. *Mitteilungen der Österreichische Geologische Gesellschaft*, 86: 133–140.

Moryc, W. & Jachowicz, M., 2000. Precambrian deposits in the Bochnia-Tarnów-Dębica area (southern Poland). (In Polish, English summary). *Przegląd Geologiczny*, 48: 601–606.

Nagy, A., Vass, D., Petrik, F. & Pereszlényi, M., 1996. Tectogenesis of the Orava Depression in the light of latest biostratigraphic investigations and organic matter alteration studies. *Slovak Geological Magazine*, 1/96: 49–58.

Nagymarosy, A. & Báldi-Beke, M., 1993. The Szolnok unit and its probabale paleogeographic position. *Tectonophysics*, 226: 457–470.

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

Nemčok, M., Nemčok, J., Wojtaszek, M., Ludhova, L., Klecker, R. A., Sercombe, W. J., Coward, M. P. & Keith, F. K., Jr., 2000. Results of 2D balancing along 20° and 21°30’ longitude and pseudo-3D in the Smilno tectonic window: implications for shortening mechanisms of the West Carpathian accretionary wedge. *Geologica Carpathica*, 51: 281–300.

Nemčok, M., Pospišil, L., Lexa, J. & Donelick, R. A., 1998. Tertiary subduction and slab break-off model of the Carpathian-Pannonian region. *Tectonophysics*, 295: 307–340.

Orlický, O.,1996. Paleomagnetism of neovolcanics of the East-Slovak lowlands and Zemplínske vrchy Mts.: A study of the tectonics applying the paleomagnetic data (W. Carpathians). *Geologica Carpathica*, 47: 13–20.

Oszast, J. & Stuchlik, L.,1977. The Neogene vegetation of the Podhale (West Carpathians, Poland). (In Polish, English summary). *Acta Palaeobotanica*, 18: 45–86.

Oszczypko, N., 1992. Late Cretaceous through Paleogene evolution of Magura Basin. *Geologica Carpathica*, 43: 333–338.

Oszczypko, N., 1997. The early-middle Miocene Carpathian peripheral foreland basin (Western Carpathians, Poland). In: Krobicki, M. & Zuchiewicz, W., (eds), *Dynamics of the Pannonian-Carpathian-Dinaride system; PANCARDI 97. Przegląd Gelogiczny*, 45: 1054–1063.

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.

Oszczypko, N., 1999. From remnant oceanic basin to collision-related foreland basin – a tentative history of the Outer Western Carpathians. *Geologica Carpathica*, 40: 161–163.

Oszczypko, N., 2001. Magura Unit – general geology. In: Birkenmajer, K. & Krobicki, M. (eds), *Carpathian palaeogeography and geodynamics: a multidisciplinary approach: Field trip C*. 12th Meeting of the Association of European Geological Societies, Kraków 2001. Państwowy Instytut Geologiczny, Kraków: 173–177.

Oszczypko, N., Golonka, J., Malata, T., Poprawa, P. Słomka, T. & Uchman, A., 2003. Tectonostratigraphic evolution of the Outer Carpathian basins (Western Carpathians, Poland). *Mineralia Slovaca*, 35: 17–20.

Oszczypko, N., Krzywiec, P., Popadyuk, I. & Peryt, T., 2005. Carpathian Foredeep Basin (Poland and Ukraine): Sedimentary, Structural, and Geodynamic Evolution. In: Picha, F. & Golonka, J. (eds), *The Carpathians and their foreland: Geology and hydrocarbon resources. American Association of Petroleum Geologists, Memoir*, 84: 1–58.

Oszczypko, N., & Tomaś, A., 1985. Tectonic evolution of marginal part of the Polish Flysch Carpathians in the Middle Miocene. *Kwartalnik Geologiczny*, 29: 109–128.

Oszczypko, N., Zając, R., Garlicka, I., Menčik, E., Dvorak, J. & Matejovská, O., 1989. Geological map of the substratum of the Tertiary of the Western Outer Carpathians and their Foreland. In: Poprawa, D. & Nemčok, J. (eds), *Geological Atlas of the Western Outer Carpathians and their Foreland, 1:500,000*. Państwowy Instytut Geologiczny, Warszawa.

Oszczypko, N. & Zuchiewicz, W., 2000. Jointing in Eocene flysch strata of the mid-eastern Magura Nappe, Polish outer Carpathians: implications for the timing of deformation. *Slovak Geological Magazin*, 6: 441–456.

Pescatore, T. & Ślączka, A., 1984. Evolution models of two flysch basins: the Northern Carpathians and the Southern Apennines. *Tectonophysics*, 106: 49–70.

Picha, F. J., 1996. Exploring for hydrocarbons under thrust belts: a challenging new frontier in the Carpathians and elsewhere. *American Association of Petroleum Geologists, Bulletin*, 89: 1547–1564.

Picha, F., Stráník, Z. & Krejčí, O., 2005. Geology and Hydrocarbon Resources of the Outer West Carpathians and their Foreland, Czech Republic. In: Picha, F. & Golonka, J. (eds), *The Carpathians and their foreland: Geology and hydrocarbon resources*. *American Association of Petroleum Geologists, Memoir*, 84: 1–123.

Plašienka, A., 1999. *Tectonochrology and paleotectonic model of the Jurassic–Cretaceous evolution of the Central Western Carpathians*. (In Slovak, English summary). Veda, Bratislava, 127 pp.

Plašienka, D., Grecula, P., Putiš, M., Kováč, M. & Hovorka, D., 1997. Evolution and structure of the Western Carpathians: an overview. In: Grecula, P., Hovorka, D. & Putiš, M. (eds), *Geological evolution of the Western Carpathians*. Mineralia Slovaca, Monograph, Bratislava: 1–24.

Poller, U., Todt, W., Kohut, M. & Janak, M., 2001. Nd, Sr, Pb isotope study of the Western Carpathians: implications for the Paleozoic evolution. *Schweizerische Mineralische Petrographische Mitteilungen*, 81: 159–174.

Pomianowski, P., 1988. Thermal anomalies above the contact zone of the Pieniny Klippen Belt and the Podhale Flysch. (In Polish, English summary). *Przegląd Geologiczny*, 36 (2): 94–97.

Pomianowski, P., 2003. Tectonics of the Orava-Nowy Targ Basin – results of the combined analysis of the gravity and geoelectrical data. (In Polish, English summary). *Przegląd Geologiczny*, 51: 498–506.

Poprawa, P., Malata, T. & Oszczypko, N., 2002. Tectonic evolution of the Polish part of Outer Carpathian's sedimentary basins – constraints from subsidence analysis. (In Polish, English summary). *Przegląd Geologiczny*, 50: 1092–1108.

Potfaj, M., 2003. Geology of the Slovakian part of the Orava – and overview. In: Golonka, J. & Lewandowski, M. (eds), *Geology, geophysics, geothermics and deep structure of the West Carpathians and their basement*. Publications of the Institute of Geophysics, Polish Academy of Sciences, Monographic Volume, 28 (363): 51–56.

Póka, T., 1988. Neogene and Quaternary Volcanism of the Carpathian-Pannonian Region: Changes in Chemical Composition and its Relationship to Basin Formation. In: Royden, L. & Horváth, F., (eds), *The Pannonian Basin: A study in basin evolution*. *American Association of Petroleum Geologists, Memoir*, 45: 257–278.

Ratschbacher, L., Frisch, W., Linzer, H.-G., Sperner, B., Meschede, M., Decker, K., Nemčok, M., Nemčok, J. & Grygar, R., 1993. The Pieniny Klippen Belt in the Western Carpathians of northeastern Slovakia: structural evidence for transpression. *Tectonophysics*, 226: 471–483.

Ratschbacher, L., Merle, G., Davy, P. & Cobbold, P., 1991. Lateral extrusion in the Eastern Alps. Part I: Boundary conditions and experiments scaled for gravity. *Tectonics*, 10: 245–266.

Roniewicz, P., 1966. Lower Werfenian (Seisian) clastics in the Tatra mountains. (In Polish, English summary). *Acta Geologica Polonica*, 16: 1–90.

Roth, Z., Benešová, E., Čechovič, V., Eliáš, M., Hanzlíková, E., Chmelík, F., Matějka, A. & Pícha, F., 1963. *Vysvetlivky k prehladnej geologickej mape ČSSR 1: 200 000, M-34-XX Trstená*. (In Slovak). Geofond, Bratislava, 59 pp.

Roure, F., Roca, E. & Sassi, W., 1993. The Neogene evolution of the outer Carpathian flysch units (Poland, Ukraine and Romania): kinematics of a foreland/fold-and-thrust belt system. *Sedimentary Geology*, 86: 177–201.

Royden, L. H., 1985. Vienna Basin: a thin skinned pull-apart basin. In: Biddle, K. & Christie Blick, N. (eds), *Strike-slip deformation, basin formation, and sedimentation*. *Society of Economic Paleontologists and Mineralogists, Special Publication*, 37: 319–338.

Royden, L. H., 1988. Late Cenozoic tectonics of the Pannonian basin system. In: Royden, L. H. & Horváth, F., (eds), *The Pannonian Basin: A study in basin evolution*. *American Association of Petroleum Geologists, Memoir*, 45: 27–48.

Royden, L. H., Horvath, F. & Burchfiel, B. C., 1982. Transform faulting, extension and subduction in the Carpathian Pannonian region. *Geological Society of America Bulletin*, 93: 717–725.

Royden, L. H. & Karner, G. D., 1984. Flexure of lithosphere beneath Apennine and Carpathian foredeep basins: Evidence for insufficient topographic load. *American Association of Petroleum Geologists, Bulletin*, 68: 704–712.

Sandulescu, M., Stefanescu, M., Butac, A., Patrut, I. & Zaharescu, P. M., 1981. *Genetical and structural relations between Flysch and Molasse (The East Carpathians Model)*. Institute of Geology and Geophysics, Romania, Bucharest, 95 pp.

Skjernaa, L., 1975. Experiments on superimposed buckle folding. *Tectonophysics*, 27: 255–270.

Smirnov, S. E., 1973. *Paleogene of the Marmarosh and Pieniny zones of the Ukrainian Carpathians*. (In Russian) Nedra Press, Moskva, 120 pp.

Sokołowski, S., 1973. Geologia paleogenu i mezozoicznego podłoża południowego skrzydła niecki podhalańskiej w profilu głębokiego wiercenia w Zakopanem. (In Polish, English summary). *Biuletyn Instytutu Geologicznego*, 265: 5–74.

Soták, J. & Janočko, J., 2001. Central-Carpathian Paleogene Basin –an outline to sedimentology, sequence stratigraphy and basin history. In: Janočko, J. & Soták, J. (eds), *Sedimentary sequences and depositional systems of the Central-Carpathian Paleogene Basin. Guidebook to IAS field trip, Slovakia 2001*. Cassovia Print, Košice: 1–32.

Sperner, B., 1996. Computer programs for the kinematic analysis of brittle deformation structures and the Tertiary tectonic evolution of the Western Carpathians (Slovakia). *Tübinger Geowissenschaftliche Arbeiten*, 27: 1–120.

Sperner, B., Ratschbacher, L. & Nemčok, M., 2002. Interplay between subduction retreat and lateral extrusion: Tectonics of the Western Carpathians. *Tectonics*, 21: 1051.

Stanley, W. D., 1988. Comparison of geoelectric/tectonic models for suture zones in western U.S. and Eastern Europe. Are black shales a possible source of high conductivities?. Physics of Earth and Planetary Interiors, 53: 228–239.

Ślączka, A., 1970. Geology of the Dukla Unit. (In Polish, English summary). *Prace Państwowego Instytutu Geologicznego*, 63: 1–97.

Ślączka, A., 1976. New data on structure of the basement of the Carpathian south of Wadowice (in Polish with English summary). *Annales Societatis Geologorum Poloniae*, 46: 337–350.

Ślączka, A., 1996a. Oil and gas in the northern Carpathians. In: Wessely, G. & Liebl, W. (eds), *Oil and gas in Alpidic thrustbelts and basins of Central and Eastern Europe*. Special Publications of the European Association of Geoscientists and Engineers, 5. Geological Society, London: 187–195.

Ślączka, A., 1996b. Oil and gas in the Ukrainian part of the Carpathians and their foredeep. In: Wessely, G. & Liebl, W. (eds), *Oil and gas in Alpidic thrustbelts and basins of Central and Eastern Europe*. Special Publications of the European Association of Geoscientists and Engineers, 5. Geological Society, London: 17–21.

Ślączka, A. & Kaminski, M. A., 1998. A Guidebook to excursions in the Polish Carpathians: Field Trips for Geoscientists. *Grzybowski Foundation, Special Publication*, 6: 1–173.

Ślączka, A., Kruglow, S., Golonka, J., Oszczypko, N. & Popadyuk, I., 2005. The General Geology of the Outer Carpathians, Poland, Slovakia, and Ukraine. In: Picha, F. & Golonka, J. (eds), *The Carpathians and their foreland: Geology and hydrocarbon resources. American Association of Petroleum Geologists, Memoir*, 84: 1–70.

Švabenická, L., Bubik, M., Krejči, O. & Stranik, Z., 1997. Stratigraphy of Cretaceous sediments of the Magura Group of Nappes in Moravia (Czech Republic). *Geologica Carpathica*, 48: 179–191.

Tari, G., &. Horvath, F., 2005. Eoalpine evolution of the Pannonian Basin: a brief oveview. In: Picha, F. & Golonka, J. (eds), *The Carpathians and their foreland: Geology and hydrocarbon resources. American Association of Petroleum Geologists, Memoir*, 84: 1–14.

Tari, G., Horváth, F. & Rumpler, J., 1992. Styles of extension in the Pannonian basin. *Tectonophysics*, 208: 203–216.

Tokarski, A. K. & Zuchiewicz, W., 1998. Fractured clasts in the Domański Wierch series: Contribution to structural evolution of the Orava Basin (Carpathians, Poland) during Neogene through Quaternary times. (In Polish, English summary). *Przegląd Geologiczny,* 46: 62–66.

Tomek, C. & Hall, J., 1993. Subducted continental margin imaged in the Carpathians of Czechoslovakia. *Geology*, 21: 535–538.

Uchman, A., 1993. Lower Jurassic carbonate sedimentation controlled by tilted blocks in the Choč Unit in the Tatra Mts., Poland. *Zentralblatt für Geologie und Paläontologie*, I, 7/8: 875–883, Stuttgart.

Vanko, J., 1988. Mapa recentných vertikálnych pohybov Západných Karpát na Slovensku v epochu 1952–1977. (In Slovak). *Geodetický a kartografický obzor*, 34 (76): 216–222, Bratislava.

Vass, D., 1998. Neogene geodynamic development of the Carpathian arc and associated basins. In: Rakús, M. (ed.), *Geodynamic development of the Western Carpathians*. Geological Survey of Slovak Republic, Dionýz Štúr Publishers, Bratislava: 155–158.

Watycha, L., 1977. *Objaśnienia do Szczegółowej Mapy Geologicznej Polski, 1: 50 000, arkusz Czarny Dunajec*. (In Polish). Wydawnictwa Geologiczne, Warszawa, 105 pp.

Wdowiarz, S., 1976, On the relation of the Carpathians to the Carpathian Foredeep in Poland. (In Polish, English summary). *Przegląd Geologiczny*, 6: 350–357.

Wdowiarz, S., 1985. On some aspects of the geological structure and oil and gas accumulation of the Carpathian Synclinorium. (In Polish, English summary). *Biuletyn Instytutu Geologicznego*, 350: 5–52.

Wieczorek, J., 2000. Mesozoic evolution of the Tatra Mountains (Carpathians). *Mitteilungen der Gesellschaft der Geologie und Bergbaustudenten in Wien*, 44: 241–262.

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.

Zoetemeijer, R., Tomek, C. & Cloething, S., 1999. Flexural expression of European continental lithosphere under the Western Outer Carpathians. *Tectonics*, 18: 843–861.

Zuchiewicz, W., 1998. Quaternary tectonics of the Outer West Carpathians, Poland. *Tectonophysics*, 297: 121–132.

Zuchiewicz, W., Tokarski, A. K, Jarosiński, M. & Márton, E., 2002. Late Miocene to present day structural development of the Polish segment of the Outer Carpathians. *EGU Stephen Mueller Special Publication Series*, 3: 185–202.

Żytko, K., 1997. Electrical conductivity anomaly of the northern Carpathians and the deep structure of the orogen. *Annales Societatis Geologorum Poloniae*, 67: 25–44.

Żytko, K., 1999. Symmetrical pattern of the late Alpine features of the northern Carpathian basement, their foreland and hinterland; orogen and craton suture. (In Polish, English summary). *Prace Państwowego Instytutu Geologicznego*, 168: 165–194.

Streszczenie

PROBLEMATYKA GŁĘBOKIEGO WIERCENIA NA ORAWIE A POPALEOGEŃSKA TEKTONIKA KARPAT PÓŁNOCNYCH

Jan Golonka, Paweł Aleksandrowski, Roman Aubrecht, Józef Chowaniec, Monika Chrustek, Marek Cieszkowski, Radosław Florek, Aleksandra Gawęda, Marek Jarosiński, Beata Kępińska, Michał Krobicki, Jerzy Lefeld, Marek Lewandowski, František Marko, Marek Michalik, Nestor Oszczypko, Frank Picha, Michal Potfaj, Ewa Słaby, Andrzej Ślączka, Michał Stefaniuk, Alfred Uchman & Andrzej Żelaźniewicz

W grudniu 1999 Polska dołączyła do programu wierceń kontynentalnych – *International Continental Scientific Drilling Program* (ICDP). W ramach tego programu jest przygotowywany projekt głębokiego wiercenia w strefie kontaktu teranu Karpat wewnętrznych i płyty północnoeuropejskiej. Praca przedstawia zarys geologii Karpat na terenie Polski i Słowacji, ze szczególnym uwzględnieniem Tatr, paleogenu wewnątrzkarpackiego, pienińskiego pasa skałkowego, zachodnich Karpat zewnętrznych, podłoża nasunięcia karpackiego na południe od Krakowa, neogeńskiego wulkanizmu i budowy geologicznej niecki orawskiej.

Wiercenie "Orawa" byłoby usytuowane w rejonie Jabłonki-Chyżnego na linii przekroju sejsmicznego CELEBRATION CEL01, jak również w niedalekim sąsiedztwie głębokiego przekroju geologicznego Kraków-Zakopane i na linii przekroju Andrychów-Chyżne. Przekroje Kraków-Zakopane i Andrychów-Chyżne wykorzystują szereg wierceń Państwowego Instytutu Geologicznego i PGNiG, a także badania sejsmiczne i magnetotelluryczne. Usytuowanie wiercenia w rejonie przygranicznym pozwoli na międzynarodową współpracę z geologami i geofizykami słowackimi.

Wiercenie to ma na celu wyjaśnienie szeregu problemów badawczych. Jednym z nich jest zagadnienie młodych i współczesnych ruchów tektonicznych w Karpatach. Przez obszar karpacki przebiega granica europejskiego pola plam gorąca, wyznaczona neogeńskim wulkanizmem oraz rozkładem strumienia cieplnego. Na obszarze pomiędzy Górną Orawą a Górnym Śląskiem, linia graniczna łącząca neogeńskie wulkanity Zakarpacia z andezytami rejonu przypienińskiego i bazaltami Dolnego Śląska przecina skośnie nasunięcia jednostek fliszowych Karpat Zewnętrznych. Równocześnie w rejonie Orawy do pienińskiego pasa skałkowego skośnie dochodzi oś karpackiej, ujemnej anomalii grawimetrycznej, a podłoże skonsolidowane występuje na głębokości nie większej niż 6–9 km, a więc w zasięgu głębokiego wiercenia, co sugerują wyniki badań megnetotellurycznych (Żytko, 1999) i magnetycznych. Podniesienie to, przy generalnym zapadaniu podłoża platformy europejskiej pod Karpaty ku południowi, może być spowodowane warunkami geotermicznymi, na skutek podnoszenia się astenosfery i występowania pióropuszy płaszcza. Pióropusze te mogą być niezależne od karpackiej kompresji i subdukcji. Z pióropuszami tymi łączy się lokalna i regionalna ekstensja w warunkach megaregionalnej kompresji. Zjawiska tego rodzaju nie są jeszcze dokładnie poznane, aczkolwiek występują w kilku miejscach na świecie (np. Panteleria na Morzu Śródziemnym). Opracowanie zagadnienia roli pióropuszy płaszcza i określenie ich relacji do kolizji i subdukcji mają zasięg globalny, a ich wyjaśnienie w rejonie karpackim pozwoli na stworzenie uniwersalnego modelu ewolucji orogenów. Nie jest wykluczone, że mamy do czynienia z orogenezą "modyfikowaną" przez pióropusz płaszcza.

Powstanie niecki Orawy i Podhala mogłoby więc mieć związek z riftingiem spowodowanym wpływem pióropuszy płaszcza na pograniczu dwóch płyt. Ryft ten jest obrzeżony między innymi wyniesieniami Babiej Góry i Orawskiej Magury. Z ryftem może być związany wulkanizm ukryty pod neogeńskimi utworami niecki orawskiej, a widoczny jako wysokooporowe ciała na profilach megnetotellurycznych. Tektonikę tego obszaru komplikuje występowanie uskoków przesuwczych o różnym przebiegu i orientacji i związane z nimi tworzenie się basenów międzyprzesuwczych typu *pull-apart.*

Proponowane wiercenie przyczyniłoby się do uzyskania odpowiedzi na postawione wyżej problemy. Dla określenia dokładnej lokalizacji wiercenia i jego właściwej interpretacji geologicznej konieczne będzie wykonanie dodatkowych prac geofizycznych. Płytka sejsmika wyjaśniłaby zasięg utworów neogeńskich i pozycję pienińskiego pasa skałkowego pod utworami neogenu, zaś głęboka sejsmika, a zwłaszcza zdjęcie 3-D, przyczyniłaby się do lepszego rozpoznania tektoniki wgłębnej.