The structural position and tectonosedimentary evolution of the Polish Outer Carpathians

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A b s t r a ct. The sedimentary basins of the Outer Carpathians are regarded as the remnant oceanic basins that were transformed into the foreland basin. These basins developed between the colliding European continent and the intra-oceanic arcs. In the pre-orogenic and syn-orogenic evolution of the Carpathian basins the following prominent periods can be established: Middle Jurassic — Early Cretaceous opening of basins and post-rift subsidence, Late Cretaceous — Palaeocene inversion, Palaeocene to Middle Eocene subsidence, Late Eocene–Early Miocene synorogenic closing of the basins. In the Outer Carpathian sedimentary area the important driving forces of the tectonic subsidence were syn- and post-rift thermal processes as well as the emplacement of the nappe loads related to the subduction processes. Similar to the other orogenic belts, the Outer Carpathians were progressively folded towards the continental margin. This process was initiated at the end of the Palaeocene at the Pieniny Klippen Belt Magura Basin boundary and completed during Early Burdigalian in the northern part of the Krosno flysch basin.

Key words: *rifting, inversion, subsidence, tectono-sedimentary evolution, Outer Western Carpathians*

The Polish Carpathians are a part of the great arc of mountains, which stretches for more than 1300 km from the Vienna Forest to the Iron Gate on the Danube. In the west the Carpathians are linked with the Eastern Alps, while in the east they pass into the Balkan chain (Fig. 1). Traditionally, the Western Carpathians have always been subdivided into two distinct ranges. The Inner Carpathians are the older range and the Outer Carpathians are the younger (Książkiewicz, 1977). The Pieniny Klippen Belt (PKB) is situated between the Inner and Outer Carpathians. The belt is a Neogene suture zone about 600 km long and 1–20 km wide with a strike-slip boundary (Birkenmajer, 1986). The Outer Carpathians are built up of stacked nappes and thrust-sheets, which reveal a different lithostratigraphy and structure (Fig. 2). Traditionally, three groups of nappes could be distinguished (Książkiewicz, 1977). The Marginal Group consists mainly of folded Miocene rocks, which are well represented at the front of the Eastern Carpathians, whereas the Middle Group (Early/Middle Miocene accretionary wedge) consists of several nappes that form the corn of the Western and Eastern Carpathians. The Magura Group (Late Oligocene/Early Miocene accretionary wedge) is flatly overthrust onto the middle group which consists of several nappes: the Fore-Magura–Dukla group, Silesian, Sub-Silesian and Skole units (Fig. 3). In the Outer Carpathians the main decollement surfaces are located at different stratigraphic levels. The Magura Nappe was uprooted from its substratum at the base of the Turonian–Senonian variegated shales (Oszczypko, 1992), whereas the main decollement surfaces of the middle group are located in the Lower Cretaceous black shales, with the exception of the Fore-Magura group of units, which were detached at the Senonian base. All the Outer Carpathian nappes are flatly overthrust onto the Miocene deposits of the Carpathian Foredeep (Oszczypko, 1998; Oszczypko & Tomaś, 1985). However, along the frontal Carpathian thrust a narrow zone of folded Miocene (marginal group) deposits developed (Pouzdřany, Boryslav–Pokuttya, Stebnik (Sambir) and Zgłobice units). In Poland these are represented mainly by the Zgłobice and partially by the Stebnik units. The detachment levels of the folded Miocene units are usually connected with the Lower and Middle Miocene evaporites.

The basement of the Outer Carpathian is the epi-Variscan platform and its cover (Figs 3–4). The depth of the platform basement, known from boreholes, changes from a few hundred metres in the marginal part of the foredeep up to more than 7000 m beneath the Carpathians. The magneto-telluric soundings in the Polish Carpathians have revealed a high resistivity horizon, which is connected with the top of the consolidated-crystalline basement (Żytko, 1997). The top of magneto-telluric basement reaches a depth of about 3–5 km in the northern part of the Carpathians, drops to approximately 15–20 km at its deepest point and then peaks at 8–10 km in the southern part (Figs 5–6). The axis of the magneto-telluric low coincides, more or less, with the axis of regional gravimetric minimum. This was documented by the integrated geophysical modelling along the Rzeszów–Bardejov geotraverse (Fig. 6). South of Krosno this gravimetric low is a result of the combined effect of the thick Carpathian nappes, thick Early Miocene molasses, and possibly the Mesozoic and Paleogene deposits related to passive margin of the European Platform (Oszczypko, 1998; Oszczypko et al., 1998).

South of the gravimetric minimum and, more or less parallel to the PKB, a zone of zero values related to of the Wiese vectors was found in geomagnetic soundings (Jankowski et al., 1982). This zone is connected with a high conductivity body at a depth of 10–25 km and is located at the boundary between the North European Plate and the Central West Carpathian Block (Żytko, 1997). In the Polish Carpathians, the depth of the crust-mantle boundary ranges from 37–40 km at the front of the Carpathians and increases to 54 km towards the south before peaking along the PKB to 36–38 km (Fig. 5).

Main structural units and the problem of the SE prolongation of the Magura Nappe

Since the 1970s the principal structural units of the Outer Western Carpathians have been well correlated (see Żytko et al., 1989; Lexa et al., 2002). From the west of the Polish state boundary to the Valaške Mezerice area, where the Silesian Unit disappears, there is a direct continuity of

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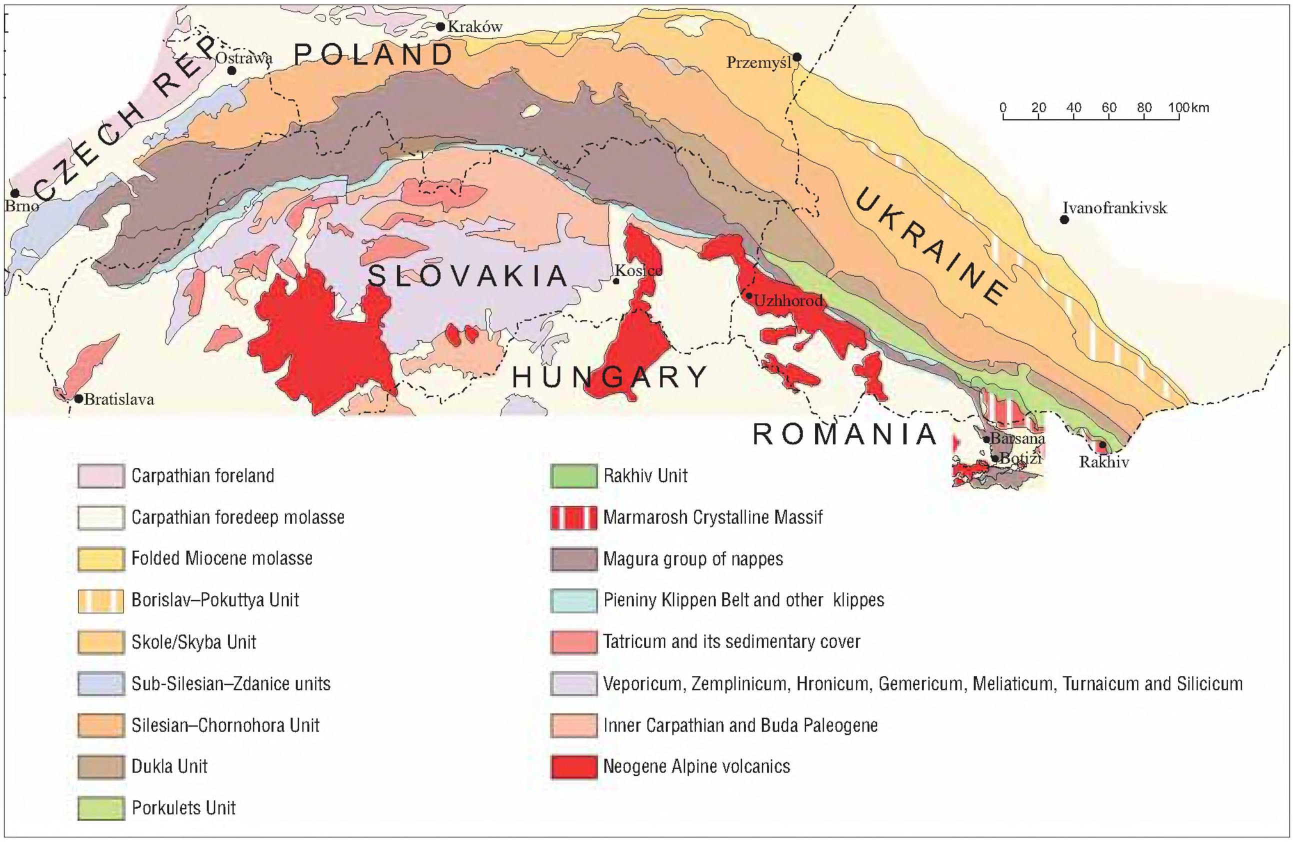


Fig. 1. Structural sketch-map of the Northern Carpathians — based on Lexa et al. (2000), Kuzovenko et al. (1996), and Aroldi (2001)

all main structural units (Fig. 1). Further to the SW the position of the Silesian Unit is occupied by a thin-skinned Zdanice–Sub-Silesian Unit. At the same time new and more external, allochthonous tectonic units (Pouzdrany Unit and then Waschberg zone), have appeared at the front of Sub-Silesian Unit. The correlation between the structural units of the middle group in the Polish and Ukrainian Carpathians has been discussed in detail by Żytko (1999). This particular correlation is more difficult because in the Ustrzyki Dolne area (Figs 1, 2), close to the Polish/Ukrainian boundary, the Sub-Silesian/Silesian overthrust is overlapped by the Lower Miocene Upper Krosno Beds. The eastern prolongation of the Sub-Silesian facies is sporadically marked by the occurrence of variegated marls (Rozluch and Holyatyn folds). The southern part of the Silesian Unit in Poland (i.e., Fore-Dukla Unit and Bystre thrust sheet) could be correlated with the Chornohora Unit. According to Żytko (1999), the SE prolongation of the Dukla Unit is related to the Porkulets (Burkut) Nappe, whereas Ukrainian authors (Shakin et al., 1976; Burov et al., 1986) link the northern boundary of the Dukla Unit to the Krasnoshora and Svidovets subunits. The southernmost units of the Ukrainian Carpathians belong to the Rakhiv and Kamianyj Potik units, which are correlated with the Ceahleu and Black Flysch units of the Romanian Eastern Carpathians, respectively. In the Western Outer Carpathians there are no equivalents of these units. The Magura Nappe is composed mainly of Upper Cretaceous to Eocene deposits. The oldest Jurassic–Early Cretaceous rocks are known from the peri-Pieniny Klippen Belt in Poland and few localities in Southern Moravia (Birkenmejer, 1977; Švabenicka et al., 1997), whereas the youngest deposits (Early Miocene) have been recently discovered in the Nowy Sącz area (Oszczypko et al., 1999; Oszczypko & Oszczypko-Clowes, 2002). The Magura Nappe is separated from the PKB by a subvertical Miocene strike-slip boundary, and flatly thrust at least 50 km towards the north over its foreland (Figs 2–6). This nappe has been subdivided into four structural subunits: Krynica, Bystrica, Rača and Siary (Fig. 2), which coincide to a large extent with the corresponding facies zones. On the west the Magura Nappe is linked with the Rheno–Danubian flysch of the Eastern Alps. Towards the east this nappe extends to Poland and runs through Eastern Slovakia before disappearing beneath the Miocene volcanic rocks, east of Uzhhorod (Trans-Carpathian Ukraine).

In the SE part of the Ukrainian and Romanian Carpathians, the zone of the Marmarosh (Maramures) Flysch has been distinguished (Smirnov, 1973; Sandulescu, 1988; Aroldi, 2001). Between the Latorica and Shopurka rivers this zone is bounded from NE by the Marmarosh Klippens and further to SE by the the Marmarosh Massif, which are thrust over the Lower Cretaceous flysch of the Rakhiv and Porkulets units (Fig. 7). In the Marmarosh Flysch zone two facies-tectonic units have been distinguished: the external Vezhany, and the internal Monastyrets units (Smirnov, 1973).

The basal part of the Vezhany Unit is built up of olistostrome, up to 100–200 m thick and is composed of Mesozoic carbonate rocks, serpentinites, basic volcanites, granitoids and metamorphic rocks. The olistostroma is followed by a 200 m thick sequence of the Upper Albian–Cenomanian grey and dark grey marly mudstones with intercalations of fine-grained, thin-bedded sandstones of the Soimul

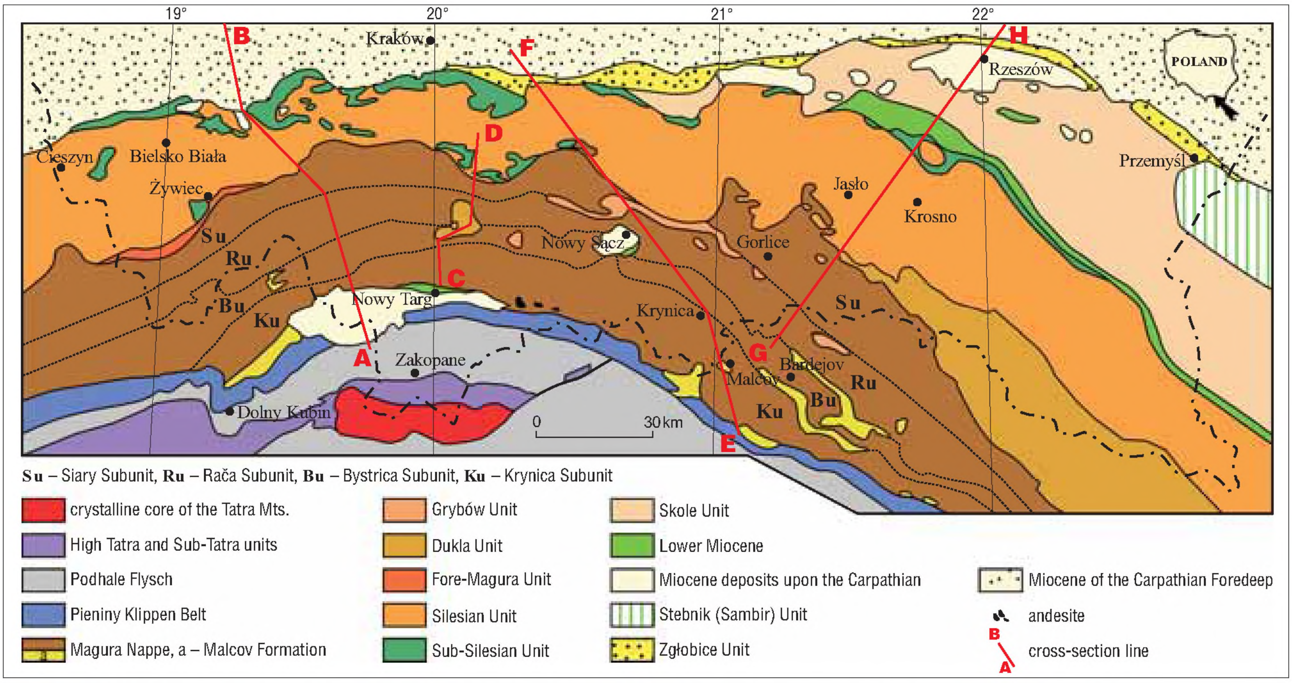


Fig. 2. Geological map of the Polish Carpathians (after Żytko et al., 1989 and Lexa et al., 2000 — supplemented)

Formation, around 180 m of the Turonian–Campanian pelagic red marls of the Puchov type and 30 m of the Maastrichtian thin-bedded flysch with intercalations of red shales of the "Jarmuta beds" (Dabagyan et al., 1989). The upper part of this sequence, 200–300 m thick, is composed of dark shaly flysch and thick-bedded sandstones of the Metove Beds (Eocene) with the Upper Eocene variegated marls at the top (Smirnov, 1973). Higher up in the section these beds are overlapped by black marls of the Luh Beds. In the Terebla River section, Oligocene (Rupelian) calcareous nanoplankton was recently discovered (Oszczypko & Oszczypko-Clowes, 2004). The Luh Beds resemble the Grybów and Dusyno bituminous marls known from the Fore-Magura units in Poland and Ukraine. In our opinion, the Vezhany succession could be regarded as an equivalent of the Jasło Unit and the North Fore-Magura thrust sheet in Poland (Oszczypko & Oszczypko-Clowes, 2004).

The Monastyrets Unit is composed of Coniacian–Palaeocene calcareous flysch with red shales (Kalyna beds, Vialov et al., 1988) at the base. These deposits are followed by thin-bedded flysch and variegated shales of the Shopurka Beds (Lower–Middle Eocene) and thick-bedded Drahovo Sandstones (Middle–Upper Eocene, see Smirnov, 1973; Andreyeva-Grigorovich et al., 1985). From the south this unit joins the PKB along the sub-vertical fault. Towards the NE it is thrust over the Vezhany Unit or directly onto the Rakhiv or Porkulets nappes.

In the Romanian Maramures equivalents of the Monastyrets Unit are known as the Leordina and Petrova units and are composed of Maastrichtian–Chattian deposits (Aroldi, 2001). South of the Bohdan Foda Fault position of the Petrova Unit is occupied by the the Wild Flysch Unit. According to Aroldi (2001) this unit is a SE prolongation of the Petrova Unit. All these units have been included by Aroldi (2001) to the Magura Group of units, which are flatly overthrust towards the NE and S onto the Paleogene–Lower Miocene deposits of the Borsa Beds. Between the Botiza–?Krichevo Unit (Late Cretaceous–Oligocene) and the Wild Flysch Unit, the Middle Jurassic–Oligocene Poiana Botizei Klippens are wedged. These klippes are regarded by Aroldi (2001) as the SE termination of the PKB, but according to Bombita et al. (1992) they represent the intra-Magura klippens (like Hluk Klippe in S Moravia).

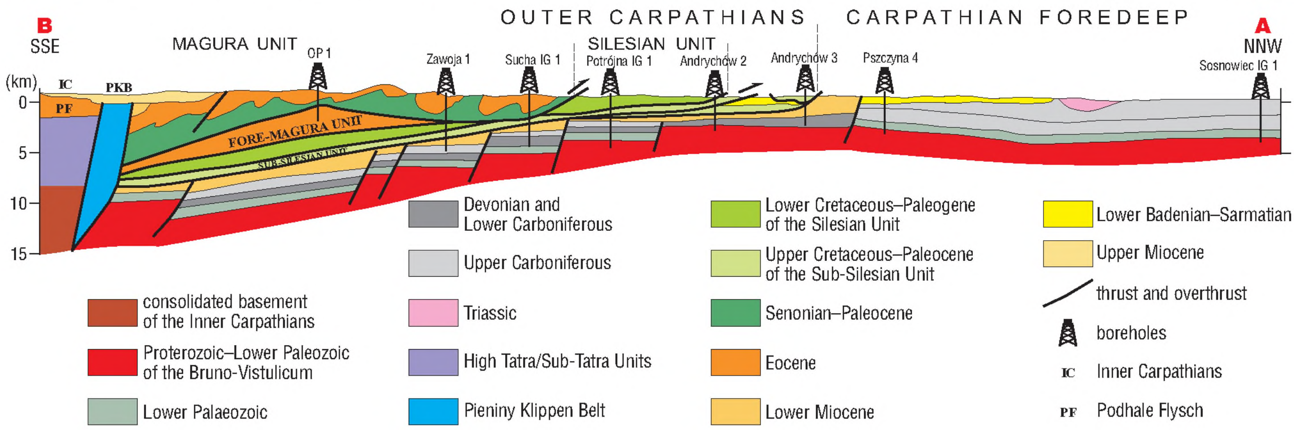


Fig. 3. Geological cross-section (A–B) Orawa-Sosnowiec (after Oszczypko et al., in print)

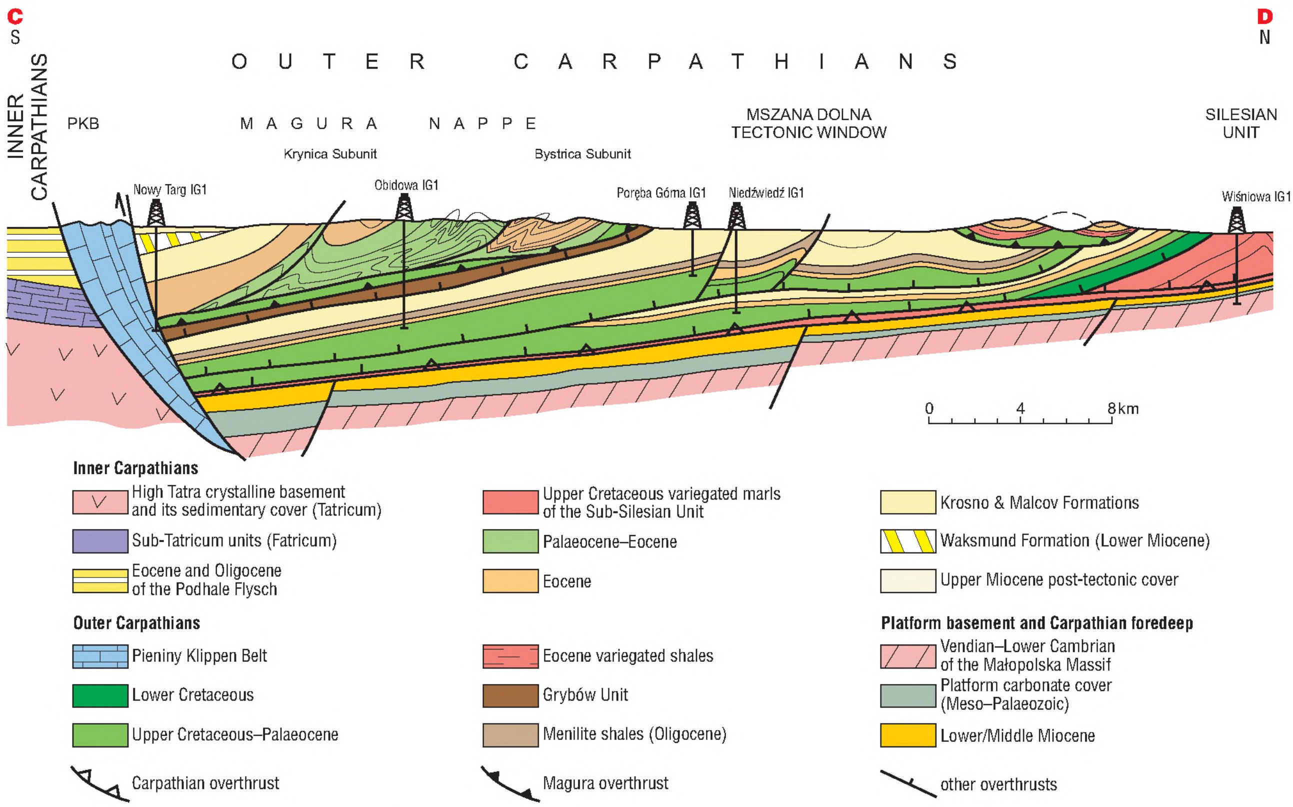


Fig. 4. Geological cross-section (C–D) Nowy Targ IG 1 — Wiśniowa IG 1

The Marmarosh Flysch of the Eastern Carpathians (Ukraine ad Romania) revealed several similarities to the Magura Nappe of the Western Carptahians in Slovakia and Poland. These nappes occupied the same geotectonic position and they are bounded from the north and south by the Fore-Magura group of units and PKB, respectively. The Magura and Marmarosh flysch successions revealed the same basin development trends, palaeocurent direction and location of source areas. Both these successions revealed a prominent, northward progradation of the Eocene/Oligocene thick-bedded muscovitic sandstones (see Żytko, 1999).

In the Marmarosh Flysch this is manifested by occurrence of the Secu Sandstones (Lutetian–Priabonian) in the Botiza Unit, Stramtura/Drahovo Sandstones (Priabonian) in the Petrova/Monastyrets Unit and the Voroniciu Sandstones (Rupelian–Chattian) in the Leordina Unit (Aroldi, 2001).

According to Żytko (1999) the Petrova/Monastyrets, Botiza and Wild Flysch units of the Marmarosh Flysch could be the equivalents of the Rača, Bystrica and Krynica subunits of the Magura Nappe respectively. Taking into account facies prolongation of the Petrova/Monastyrets Unit into the Wild Flysch Unit (Aroldi, 2001) and the lack of Łącko Marls in the Botiza Unit, these correlations should be modified. It appears that there are no equivalents of the Bystrica succession in the Marmarosh Flysch, and the Botiza succession better fits the Krynica succession than that of the Wild Flysch (see Oszczypko & Oszczypko-Clowes, 2004).

In view of the internal position of the Marmarosh Flysch in relation to the Marmarosh Massif, as well as the above mentioned similarities between:

1) the Vezhany and Fore-Magura/Jasło successions,
2) the Monastyrets/Petrova and Rača and
3) the Botiza and Krynica successions, it is possible to conclude that the palaeogeographical positions of the Marmarosh Massif and the buried Silesian Ridge were almost the same (see Sandulescu, 1988; Oszczypko, 1992, 1999).

The Evolution of the Outer Carpathian basins

The Outer Carpathians are composed of Late Jurassic to Early Miocene mainly flysch deposits. The sedimentary sequences of the main tectonic units differ in the facies development as well as in the thickness. The thicker sedimentary cover belongs to the Silesian Unit, which varies from 3000 m, in its western part, to more than 5000 m in the east. The stratigraphic thickness of the other tectonic units is distinctively thinner and varies between 3000 and 3800 m in the Skole Unit, around 1000 m in the Sub-Silesian Unit, 2300–2500 m in the Dukla Unit and 2500–3500 m in the Magura Nappe (Poprawa et al., 2002a). Taking into account the distribution of facies, the thickness of the deposits and the palaeocurrent directions (see Książkiewicz, 1962) only the Magura, Silesian and Skole basins could be considered as independent sedimentary areas (see also Nemčok et al., 2000). During the Late Cretaceous–Eocene times, the Sub-Silesian depositional area formed a submarine high dividing the Skole and Silesian basins. The history of the Dukla sedimentary area, which played the role of a transfer zone between the Magura and Silesian basins, was more complex. According to the reconstructions of Roure et al. (1993) and Behrman et al. (2000), the Outer Carpathian basins during the Early Oligocene were at least 380 km wide across the Przemyśl–Hanušovce geotraverse. This restoration does not include the Silesian Ridge, at least 20–50 km wide (see Unrug, 1968), located between the Magura and Silesian depositional areas. This suggests

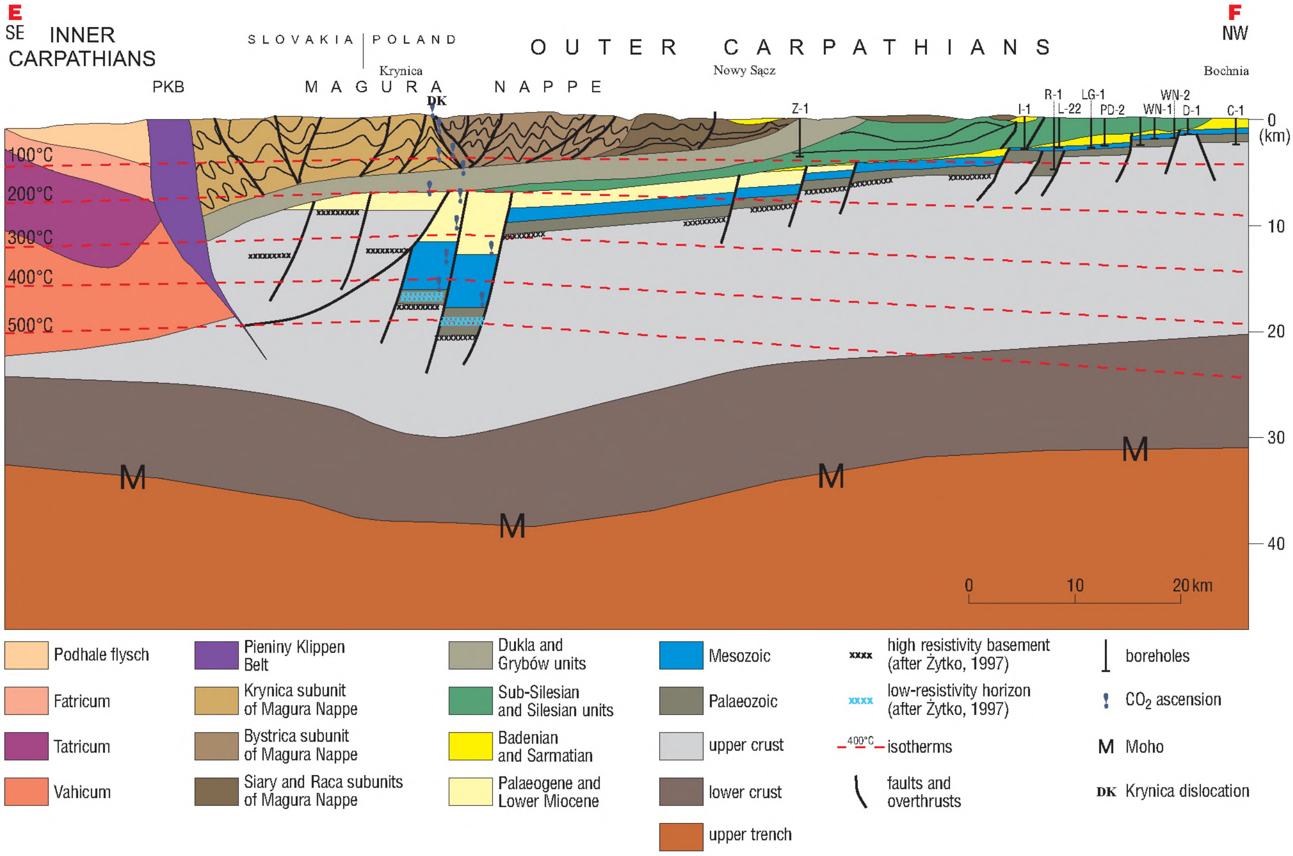


Fig. 5. Geological cross-section (E–F) Krynica–Bochnia (after Oszczypko & Zuber, 2002 — suplemented)

that the entire width of the Outer Carpathian domain reached at least 500 km.

Traditional opinions suggest that the Magura and Silesian basins were situated parallel to each other (see Książkiewicz, 1962; Unrug, 1968, 1979; Birkenmajer, 1986). This view was recently questioned by Nemčok et al. (2000) who placed the Magura depositional area as the south-western neighbour of the Silesian depositional area, whereas the present-day position of these units is a result of the Miocene eastwards escape of the Magura Nappe. This model does not fit the facies distribution in the Polish Outer Carpathians (Bieda et al., 1963), palaeocurent measurements, nor the transitional position of the Dukla succession, between the Magura and Silesian basins.

The sedimentary succession of the Outer Carpathians (Table 1) reveals three different megasequences of deposits, reflecting the main stages of the basins development (Poprawa et al., 2002a). The first (long lasting) and third (relatively short) periods were characterized by the unification of sedimentary conditions, whereas the intermediate periods were characterized by a maximal differentiation of sedimentary conditions.

Middle Jurassic–Early Cretaceous opening of basins and post-rift subsidence (125–150 My)

The Outer Carpathian basins can be regarded as remnant ocean basins, which developed between the colliding European continent and the intra-oceanic arcs (Oszczypko, 1999). The Early/Middle Jurassic opening of the Magura Basin was probably coeval with the timing of the Ligurian–Penninic Ocean opening and its supposed prolongation towards the east (Oszczypko, 1992; Golonka et al., 2000, 2003). This oceanic domain was divided by the submerged Czorsztyn Ridge into the NE and SE arms. The Czorsztyn Ridge and the Inner Carpathian domain were separated by the SE arm of the Pieniny Ocean, known also as the Vahicum Oceanic Rift (south Penninic domain), whereas NE arm was occupied by the Magura deep-sea basin situated south of the European shelf, an equivalent of the north-Penninic (Valais) domain (see also, Plasienka, 2003). This stage of the Magura Basin evolution is rather speculative, because the Magura Nappe was uprooted roughly at the base of the Upper Cretaceous sequence. The Jurassic–Lower Cretaceous deposits of the Magura Basin were probably represented by deep water, condensed pelagic limestones and radiolarites. At the end of the Jurassic in the southern part of the European shelf, the palaeorifts were floored by a thinned continental crust (Birkenmajer, 1988; Sandulescu, 1988). This rifted European margin was incorporated into the Outer Carpathian Basin (Skole, Sub-Silesian/Silesian basins). The rifting process was accompanied by a volcanic activity (teschenite sills, dykes, and local pillow lavas), which persisted up to the end of Hauterivian (Lucińska-Anczkiewicz et al., 2002; Grabowski et al., 2004). This part of the rifted continental margin probably extended in the Eastern Carpathian (basic effusives —Tithonian–Hauterivian), see Lashkevich et al. (1995) of the “Black Flysch”, Kamyany Potic, and Rakhiv beds) as well as to the Southern Carpathian (Sandulescu, 1988). During the initial stage of development, the Silesian Basin was filled with calcareous flysch followed by siliciclastic flysch and pelagic shales. The Early Cretaceous–Cenomanian deposition took place during relatively

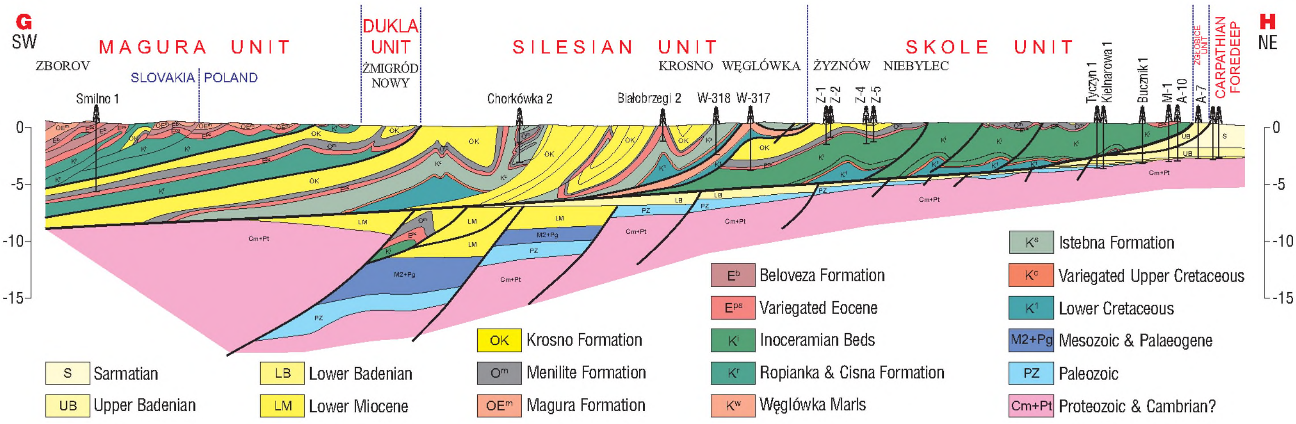


Fig. 6. Integrated geological-geophysical-sections (G–H) Smilno–Rzeszów (after Oszczypko et al., 1998 — suplemented)

low sea levels (Table 1) and was characterized by a low and decreasing rate of sedimentation from 40–20 m/My to 115–40 m/My (Figs 8, 9) for the Skole and Silesian basins respectively (Poprawa et al., 2002a). The Late Jurassic–Hauterivian deposition of the Silesian Basin was controlled by normal fault and syn-rift subsidence, and later (Barremian–Cenomanian) by a post-rift thermal subsidence, which culminated with the Albian–Cenomanian expansion of deep-water facies (Poprawa et al., 2002a, b; Nemčok et al., 2001). The Cenomanian high stand of the sea level resulted in unification of the sedimentary condition in all Outer Carpathian basins, and deposition of the green radiolarian shales (Cenomanian Key Horizon) followed by the Turonian variegated shales (Table 1).

Late Cretaceous–Palaeocene inversion (35 My)

During the Turonian in the central part of the Outer Carpathian domain, the Silesian Ridge was restructured and uplifted (Fig. 8). The inversion affected most of Silesian, Sub-Silesian and Skole sub-basins. Since the Campanian, an inversion effect is also visible in the northern part of the Magura Basin. The amplitude of the Silesian Ridge uplift reached several hundreds meters (Poprawa et al., 2002a). This was accompanied by an increase in the rates of deposition to 25–55 m/My and 50–100 m/My (Fig. 9) in the Skole and Sub-Silesian–Silesian basins, respectively (Poprawa et al., 2002a). A maximal increase in sedimentary rates took place in the western part of the Silesian area, up to 400 m/My in the Godula Beds (Poprawa et al., 2002 a, Oszczypko et al., 2003). In the Magura Basin, during the Maastrichtian–Palaeocene, coarse material derived from the Silesian Ridge supplied deposition of the Solan Beds (Švabenicka et al., 1997), Jaworzynka Beds and Mutne Sandstones (deposition rate 60 do 100 m/My, Fig. 9). The uplift of the Silesian Ridge was coeval with a regional uplifting in the southern margin of continental Europe from the Carpathian and Alpine foreland to Spain (Poprawa et al., 2002a). This was caused by regional, early orogenic compression in the Inner Carpathians and the Northern Alps (see Książkiewicz, 1977; Sandulescu, 1988; Poprawa et al., 2002a) and the rift development in the Biscay Bay (Golonka & Bocharova, 2000). In the Northern Carpathians the development of the Silesian Ridge was probably related to the inversion of pre-existing extensional structures (Roure et al., 1993; Roca et al., 1995; Kuśmierek, 1990; Krzywiec, 2002). The development of the Węglówka High, dominated by deposition of pelagic variegated marls, could also be associated with the uplift of the Silesian Ridge. The Weglówka High (like peripheral bulge) separated Silesian and Skole basins during the Santonian–Eocene time. The suggested shortening of the Silesian Basin (Oszczypko, 1999) can be also regarded as a westwards continuation of the pre-Late Albian subduction of the Outer Dacides (Sandulescu, 1988).

In the southern (peri-PKB) part of the Magura the coarse clastic deposition began with the Jarmuta Formation (Maastrichtian/Palaeocene, see Birkenmajer, 1977; Birkenmajer & Oszczypko, 1989), which was up to 500 m thick. This formation is composed of thick- to medium-bedded turbidites, contains conglomerates and sedimentary breccias composed of the Jurassic–Cretaceous sedimentary rocks, and exotic crystalline and basic volcanic rocks (Birkenmajer, 1977; Birkenmajer & Wieser; 1992; Mišik et al., 1991). Towards the north, the upper portion of this formation alternates with medium-bedded, calcareous turbidites of the Szczawnica Formation (Palaeocene–Lower Eocene, Birkenmajer & Oszczypko, 1989). In the Jarmuta and Szczawnica formations (rate of deposition 20–50 m/My, Fig. 9) significant amounts of SE–supplied chromian spinels have been found (Oszczypko & Salata, 2004). The Jarmuta formation is regarded as the synorogenic wild flysch, derived both from the erosion of the PKB as well as of the Andrusov Exotic Ridge (Birkenmajer, 1986, 1988; Birkenmajer & Wieser, 1992). These deposits probably reflect the collision of the Inner Western Carpathian Orogenic Wedge (IWCW) with the Czorsztyn Ridge (Plasienka, 2002, 2003).

Palaeocene to Middle Eocene subsidence (25 My)

At the end of Palaeocene the Carpathian basins were affected by general subsidence and the rise in sea level (Poprawa et al., 2002a, b). During the Eocene, a wide connection of the Outer Carpathian basins and the world ocean was established (Golonka et al., 2000). This resulted in unification of facies, including the position of the CCD level and low sedimentation rates. This general trend dominated during the Early to Middle Eocene time in the northern basins (Skole, Sub-Silesian, Silesian and Dukla ones) as well as in the northern part of the Magura Basin.

During the Palaeocene, the IWCW reached the southern margin of the Magura Basin. Its load caused subsidence, collapse of the PKB and southwards shift of the Magura Basin margin. This explains the deposition of the deep-water facies in the PKB (see Leško & Samuel, 1968; Bystricka et al., 1970; Książkiewicz, 1977), and allows us to explain the transfer of clastic material to Magura Basin via PKB, from the source area located in the SE part of the Inner Carpathian domain. This also enables the explanation

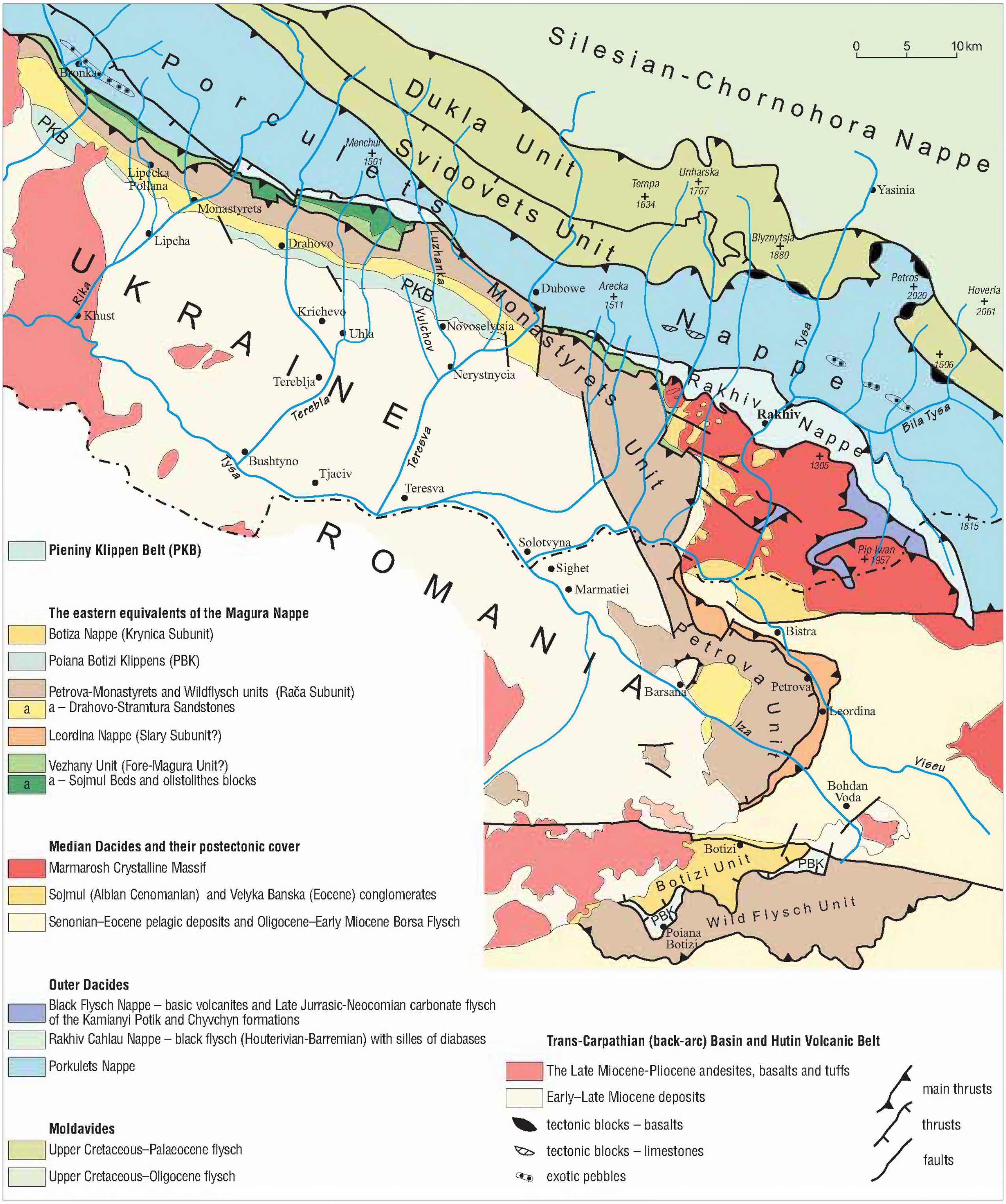


Fig. 7. Geological sketch-map map of the SE part of the Ukrainian Carpathians and adjacent part of the Romanian Maramures — based on Shakin et al. (1976), Burov et al. (1986) and Aroldi (2001)

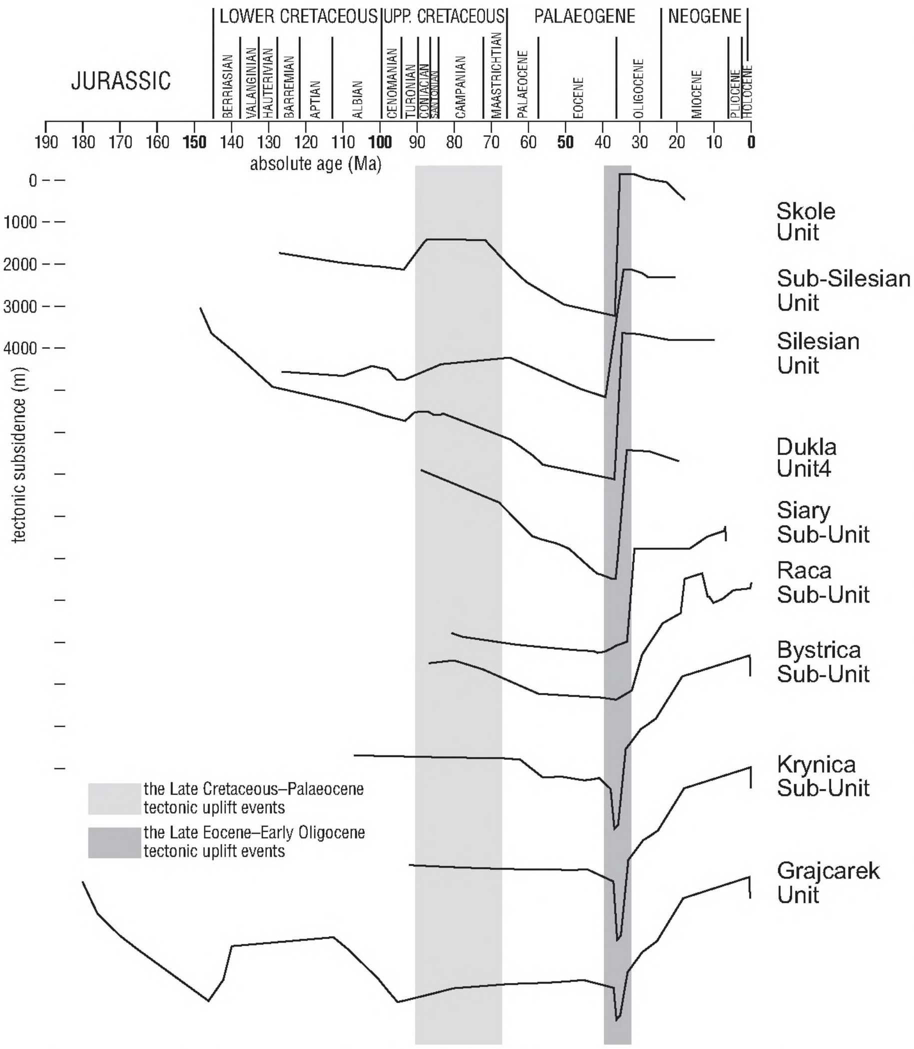


Fig. 8. Tectonic subsidence curves for selected synthethic profiles from the Polish Outer Carpathians (after Poprawa et al., 2002a; Oszczypko et al., 2003)

of provenance of the huge amount of crystalline clasts derived to the Palaeocene/Eocene deposits of the Magura Basin.

The migrating load of the Magura and PKB accretionary wedge caused further subsidence and a shift of depocentres to the north. As a result, narrow and long submarine fans developed. The northern deepest part of the basin, often located below the CCD was dominated by basinal turbidites and hemipelagites. The rate of sedimentation varied from 6–18 m/My on the abyssal plain to 103–160 m/My in the outer fan and between 180 and 350 m/My (Fig. 9) in the area affected by the middle fan-lobe systems (Oszczypko, 1999). The total amount of these deposits, can be estimated at least 3750–4500 km^3 (250–300 km x 15 km x 1 km). These were supplied from the southeast, probably from the Inner Carpathian/Inner Dacide terrains (Oszczypko et al., 2003). During the Late Eocene and Oligocene, the axes of subsidence of the Magura Basin shifted to the north towards the Rača and Siary sedimentary areas (Fig. 8).

Synorogenic Late Eocene–Early Miocene closing of the basins (15 My)

In the Outer Carpathian sedimentary area, the Late Eocene brought about drastic changes of depositional conditions (Table 1); this was accompanied by the transformation of the Outer Carpathian remnant oceanic basins into a foreland basin (Oszczypko, 1999). This resulted in the replacement of deep-water deposits (variegated shales and basinal turbidites) by pelagic Globigerina Marls, and this was followed by Oligocene bituminous–menilite shales, deposited in the restricted basin. The Late Eocene event in the Western Carpathians was probably caused both by the global glacioeustatic fall of sea level (at least 100 m) (Van Couvering et al., 1981; Haq et al., 1988), as well as by the tectonics. The latter was related to the closure of the Neotethys in the course of the Alpine–Himalayan orogenesis (Golonka et al., 2000). This was contemporaneous with the main collision phases in the Alpine belt and final stage of development of accretionary wedge in the southern part the Magura Basin (Krynica Zone), caused by subduction of the Magura Basin beneath the Pieniny Klippen Belt/Central Carpathian Block (Oszczypko, 1992, 1999).

During the Priabonian and Rupelian, a prominent uplift (Fig. 8) in the Outer Carpathian Basin was recorded (Oszczypko, 1999; Poprawa et al., 2002a). After the Late Oligocene folding, the Magura Nappe was thrust northwards onto the terminal Krosno flysch basin and during Burdigalian its front reached the S part of the Silesian Basin. This was followed by the last, minor subsidence event (Late Oligocene–Early Miocene) in the Outer Carpathian basins, which partially could be related to loading of the plate by accretionary wedge (Poprawa et al., 2002a). This subsidence was accompanied by a progressive migration of axes of depocentres towards the north, and increase of deposition rates from 350 m/My in the Rupelian (northern part of Magura Basin) to 600 m/My (Fig. 9) at the end of the Oligocene (SE part of the Silesian Basin). The restored width of the Early Burdigalian basin probably reached at least 150 km. During the Early Burdigalian sea level high stand, the Magura piggy-back basin developed and the sea-way connection to the Vienna Basin via Orava was established (Oszczypko et al., 1999; Oszczypko-Clowes, 2001; Oszczypko & Oszczypko-Clowes, 2002). During the Ottnangian, the Krosno flysch basin shifted towards NE (Zdanice Unit, Boryslav–Pokuttya and Marginal Fold units) and underwent desiccation (evaporates of the Vorotysche Formation in the Ukraine and Salt Formation in Romania).

The Outer Carpathian residual Krosno flysch basin was finally closed by the intra-Burdigalian folding and the upli-

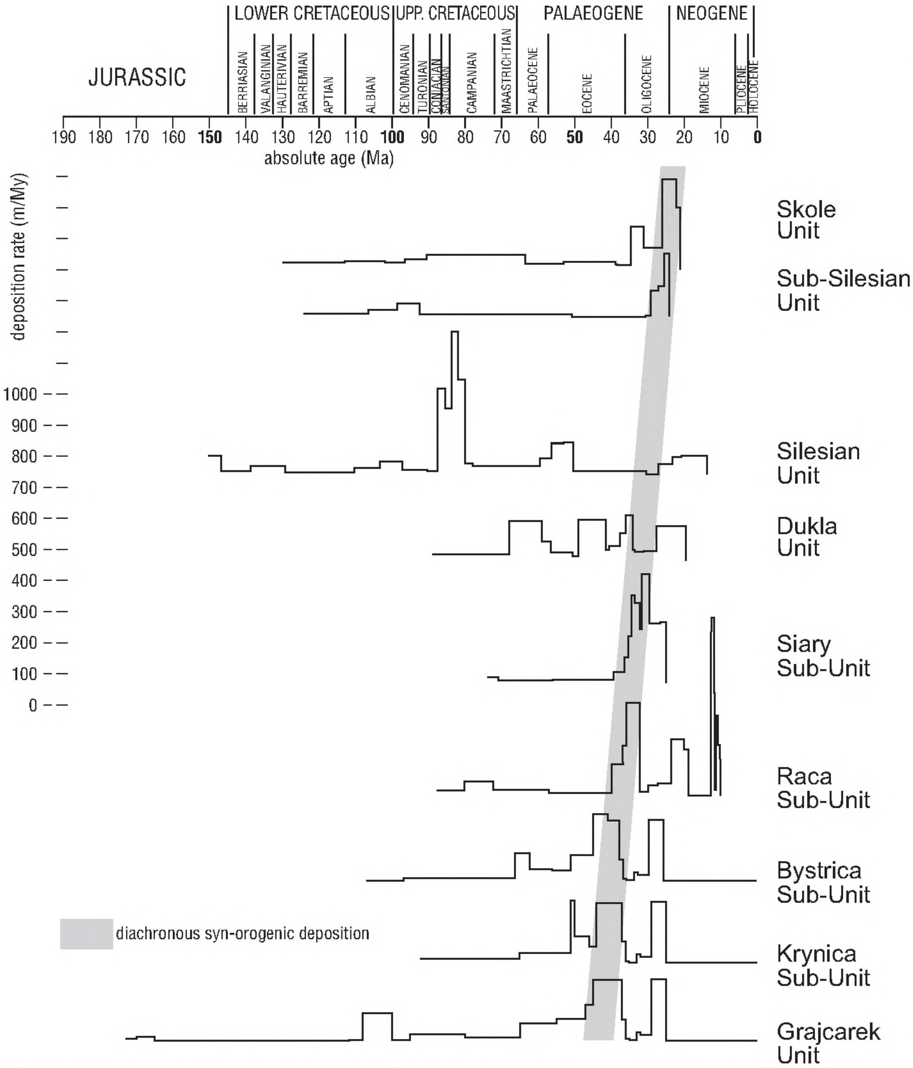


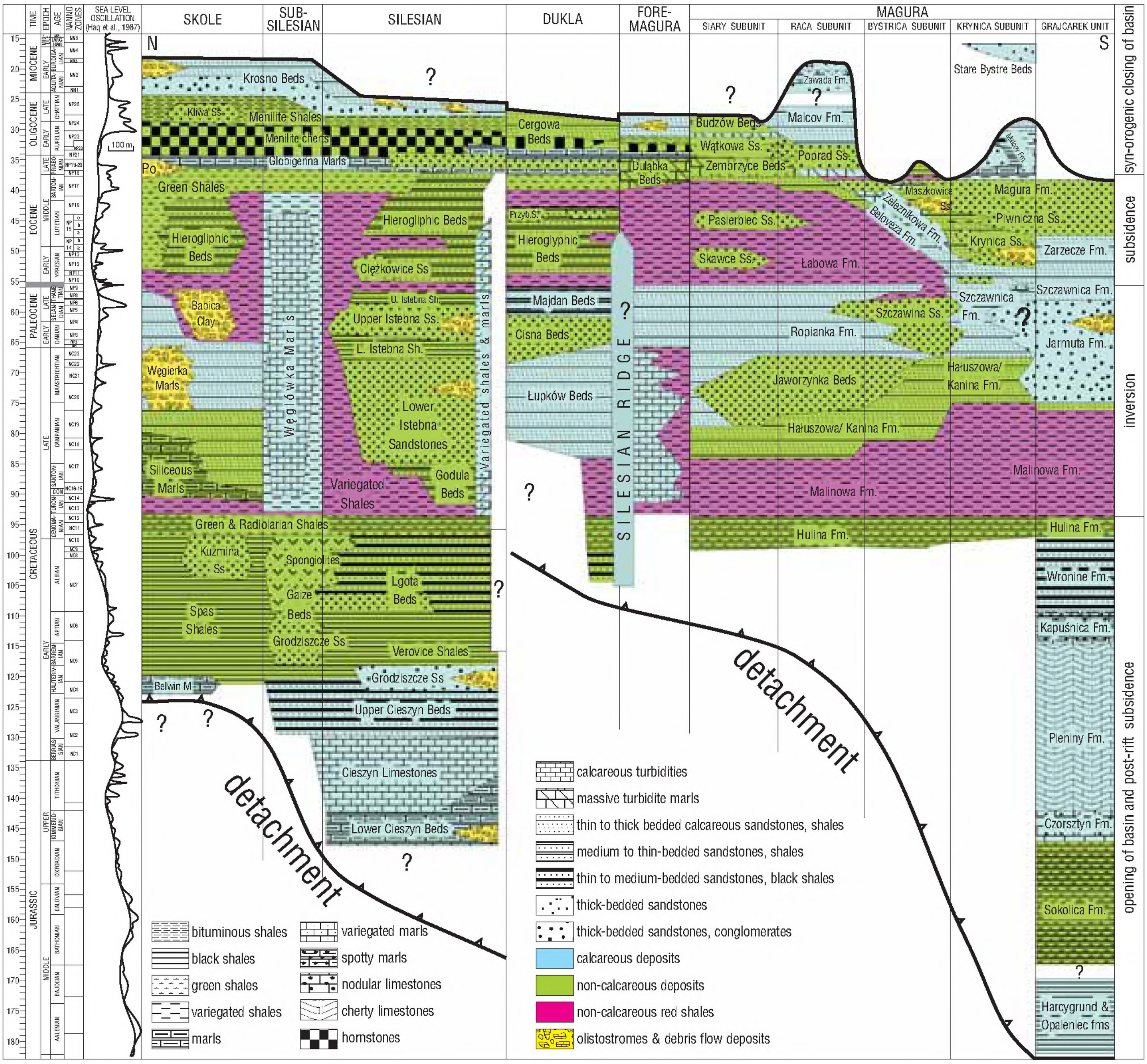
Fig. 9. Diagram of deposition rates versus time for selected synthethic profiles from the Polish Outer Carpathians (after Poprawa et al., 2002b; Oszczypko et al., 2003)

fting of the Outer Carpathians, which was connected with the collision between the European Plate and overriding Alcapa and Tisza–Dacia microplates. This was accompanied by the north and northeast overthrust and the formation of the flexural depression of the Carpathian Foredeep — related to the moving orogenic front (Oszczypko, 1998).

Early/Late Miocene folding, thrusting and development of the Carpathian foredeep basin (8 My)

The terminal flysch deposition in the Krosno residual basin and the Magura piggy-back-basin was followed by the Intra-Burdigalian (Late Ottnangian) folding, uplift and overthrust of the Outer Carpathians onto the foreland platform (Oszczypko, 1998; Kovač et al., 1998). At the turn of the Ottnangian, the front of the Outer Carpathians was located about 50 km south of the present-day position (Oszczypko & Tomaś, 1985; Oszczypko, 1997; Oszczypko & Oszczypko-Clowes, 2003). The load of the growing Carpathian accretionary wedge caused a bending of the platform basement and the development of the moat-like flexural depression (inner foredeep, see Oszczypko 1998), which was filled by coarse clastic deposits. This was accompanied by the development of large scale slides along the frontal part of the Sub-Silesian Nappe. These slides form olistoplaques and gravitational nappes, which progressively overthrust the subsiding area. In NE Moravia and S Silesia the thin-skinned Sub-Silesian and Silesian nappes overrode the platform basement and its Paleogene/Early Miocene cover. These overthrusts are known as the “Old Styrian Nappes” (Jurkova, 1971) or as the Sucha and Zamarski formations (flysch olistoplaque, see Buła & Jura, 1981; Oszczypko & Tomaś, 1985; Moryc, 1989; Oszczypko, 1998). In the Cieszyn area, this overthrust reached more or less the present-day position of the Carpathians (Oszczypko & Oszczypko-Clowes, 2003). The overthrust developed in terrestrial conditions. This is documented by alluvial origin conglomerates of the Stryszawa Fm. type, which were found at the base of the overthrust in some boreholes (Bielowicko IG 1, Zawoja 1). The olistoplaque formation was postdated by the Karpatian period of intensive subsidence and the deposition in the inner foredeep, which was filled with coarse clastic sediments of the Stryszawa Fm. (Oszczypko, 1997, 1998). The subsidence and the deposition also probably affected the frontal part of the Carpathian Nappes. The Stryszawa Formation was deposited by the alluvial fan, which was supplied by material derived from the erosion of both the Carpathians, as well as the emerged platform. The youngest recycled microfauna found in the Stryszawa Formation belong to the Eggenburgian–Ottnangian N5–N6 zone (Oszczypko, 1997). The same origin could also be suggested by the calcareous nannoplankton of NN 4 Zone found in the Stryszawa Formation (Garecka & al., 1996). These foraminifers and calcareous nannoplankton can be found both in the youngest strata of the Outer Carpathians as well as in the Zebrzydowice Formation. The deposition of the Stryszawa Formation was followed by Late Karpatian erosion, which was caused by the uplift of the peripheral bulge (Cieszyn–Slavkov Palaeo-Ridge, see Oszczypko & Tomaś, 1985, Oszczypko, 1997, Oszczypko & Lucińska-Anczkiewicz, 2000). In Southern Moravia this period of erosion could be correlated with the discordance below the terminal Karpatian strata (Jiriček, 1995). Simultaneously, the erosion on the northern flank of Cieszyn Slavkov Palaeo-Ridge resulted in the beginning of the development of the W–E and NW–SE trending graben (e.g., Bludovice–Sko-

Table 1. Litostratigraphy of the Polish Outer Carpathians [after Ślączka & Kaminski (1998) and Oszczypko & Oszczypko-Clowes (2002); supplemented, time scale after Berggren et al. (1995) & Gradstein & Ogg (1996)]



czów Palaeo-valley) that was bounded by normal faults (Oszczypko & Lucińska-Anczkiewicz, 2000). During the Late Karpatian–Early Badenian these subsiding grabens were successively filled with slope deposits (blocks of Carboniferous rocks), the near-shore Dębowiec Conglomerate, and were finally flooded by relatively deep sea (marly mudstones of the Skawina Formation.). This marine transgression invaded both the foreland plate and the Carpathians. During the Badenian the axes of the extensional grabens migrated towards the NE (Zawada and Krzeszowice grabens). The Late Badenian drop of sea level and climatic cooling initiated a salinity crisis in the Carpathian foreland basin (see Oszczypko 1998; Andreyeva-Grigorovich et al., 2003). The shallow (stable shelf) part of the evaporate basin was dominated by sulfate facies, whereas the deeper part, located along the Carpathian front, was occupied by chloride-sulfate facies. After the evaporate deposition the basement of the outer foredeep was uplifted and a part of the foredeep was affected by erosion (e.g., Rzeszów Palaeo–Ridge). This event was followed by a telescopic shortening of the Carpathian nappes (Intra-Badenian compressive event, see Oszczypko, 1997, 1998, Kovač et al., 1998). This is documented, at least, by a 12-km-long shift of the Magura and Fore-Magura units against the Silesian Unit, as well as the Silesian Unit against the Sub-Silesian Unit and the tectonic reduplication of the Sub-Silesian Unit. Finally, the present-day position of the Carpathian nappes was reached during the post-Sarmatian time (Wójcik & Jugowiec, 1998; Oszczypko, 1998).

Conclusions

1. In the pre-orogenic and syn-orogenic evolution of the Outer Carpathian domain the following main tectonic events took place: Middle Jurassic–Early Cretaceous opening of basin and post-rift subsidence, Late Cretaceous–Palaeocene inversion, Palaeocene to Middle Eocene subsidence, synergetic Late Eocene–Early Miocene closing of the basins.

The total subsidence in the Silesian Basin was two times higher than in the Magura Basin and more than three times higher than in Sub-Silesian and Skole basins.

2. The important driving forces of the tectonic subsidence were syn- and post-rift thermal processes, as well as the emplacement of the nappe loads related to the subduction processes.

3. Similarly to the other orogenic belts, the Outer Carpathians were progressively folded towards the continental margin. This process was initiated at the end of the Palaeocene at the PKB/Magura Basin boundary and completed during the Early Burdigalian in the northern part the Krosno flysch basin.

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