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Late Jurassic-Miocene evolution of the Outer Carpathian fold-and-thrust belt and its foredeep basin (Western Carpathians, Poland)

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The Outer Carpathian Basin domain developed in its initial stage as a Jurassic-Early Cretaceous rifted passive margin that faced the eastern parts of the oceanic Alpine Tethys. Following closure of this oceanic basin during the Late Cretaceous and collision of the Inner Western Carpathian orogenic wedge with the Outer Carpathian passive margin at the Cretaceous-Paleocene transition, the Outer Carpathian Basin domain was transformed into a foreland basin that was progressively scooped out by nappes and thrust sheets. In the pre- and syn-orogenic evolution of the Outer Carpathian basins the following prominent periods can be distinguished: (1) Middle Jurassic-Early Cretaceous syn-rift opening of basins followed by Early Cretaceous post-rift thermal subsidence, (2) latest Cretaceous-Paleocene syn-collisional inversion, (3) Late Paleocene to Middle Eocene flexural subsidence and (4) Late Eocene-Early Miocene synorogenic closure of the basins. In the Outer Carpathian domain driving forces of tectonic subsidence were syn-rift and thermal post-rift processes, as well as tectonic loads related to the emplacement of nappes and slab-pull. Similar to other orogenic belts, folding of the Outer Carpathians commenced in their internal parts and progressed in time towards the continental foreland. This process was initiated at the end of the Paleocene at the Pieniny Klippen Belt/Magura Basin boundary and was completed during early Burdigalian in the northern part of the Krosno Flysch Basin. During Early and Middle Miocene times the Polish Carpathian Foredeep developed as a peripheral foreland basin in front of the advancing Carpathian orogenic wedge. Subsidence of this basin was controlled both by tectonic and sedimentary loads. The Miocene convergence of the Carpathian wedge with the foreland resulted in outward migration of the foredeep depocenters and onlap of successively younger deposits onto the foreland.

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Key words: Outer Western Carpathians and their foredeep, rifting, inversion, subsidence, tectono-sedimentary evolution, peripheral foreland basin.

INTRODUCTION

The Polish Outer Carpathians (POC) form part of the arcuate Carpathian mountain system, that extends over more than 1300 km from the Danube valley in Austria to the Danube valley in Southern Romania. To the south-west, the Western Carpathians link up with the Eastern Alps, whilst the Southern Carpathians grade into the Balkan chain (Fig. 1). Traditionally, the Western Carpathians, based on their sedimentary and structural history, have been divided into two distinct ranges (Książkiewicz, 1977), namely the older Inner Carpathians and the younger Outer Carpathians that are separated by the Pieniny Klippen Belt (PKB). This belt, which is about 600 km long and 1–20 km wide and bounded by strike-slip faults, forms the Early/Middle Miocene suture between the European foreland and the ALKAPA Block of the Inner Carpathians (Birkenmajer, 1986; Csontos and Nagymarosy, 1998). The Polish Outer Carpathians involve several stacked nappes and thrust-sheets (also referred to as "units") that differ in their lithostratigraphic composition and structure (Figs. 2 and 3; Table 1). All Outer Carpathian nappes are flatly thrust over the Miocene fill of the Carpathian Foredeep.

The Polish Carpathian Foredeep is about 300 km long and up to 100 km wide and forms part of the large flexural foreland basin that extends along the front of the Western, Eastern and Southern Carpathians. To the west, the Carpathian Foredeep of Poland and Czechia links up with the Alpine Molasse Basin and to the south-east it grades into the Ukrainian foreland basin (Figs. 1 and 2). Similar to other classical foreland basins, the Carpathian Foredeep is asymmetric and predominantly filled with synorogenic Miocene clastic sediments that are up to 3 km thick near the Carpathian front in the Przemyśl area (Fig. 4, cross-section G–H). These molasse-type deposits rest uncon-

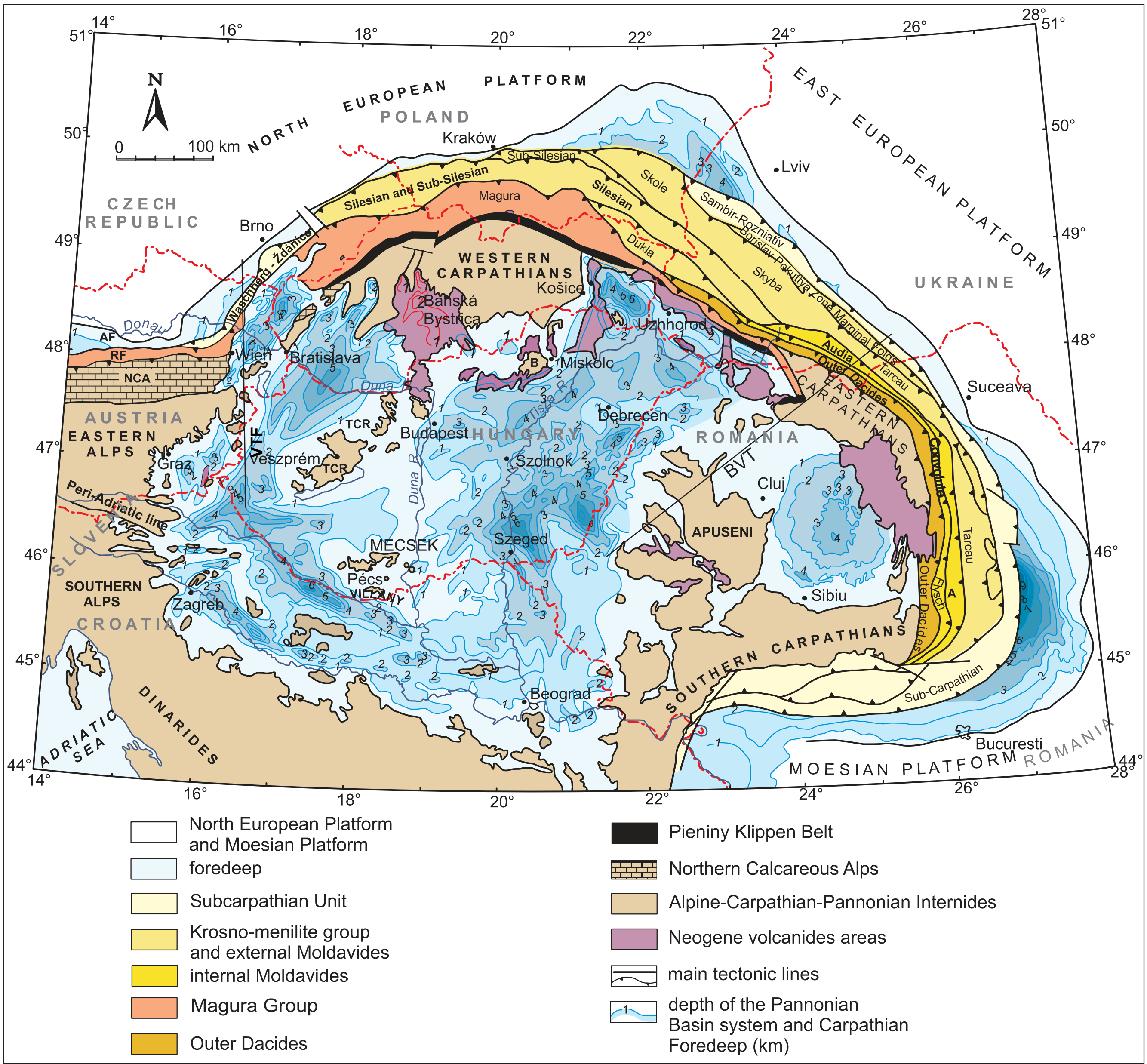


Fig. 1. Geological map of the East Alpine–Carpathian–Pannonian basin system (after Kovač *et al.*, 1998)

TCR — Trans-Danubian Ridge, B — Bükk Mts., NCA — Northern Calcareous Alps, RF — Rheno-Danubian Flysch, AF — Alpine Foredeep, VTF — Vienna Transform Fault, BVT — Bohdan Voda Transform Fault

formably on the basement and the sedimentary cover of the foreland platform.

The highly complex epi-Variscan platform and its cover (Figs. 3 and 4) form the basement of the Polish Outer Carpathians (POC) and the Polish Carpathian Foredeep (PCF) (Karnkowski, 1974; Oszczypko *et al.*, 1989, 2005*a*). This platform includes Brunovistulian Terrane and Małopolska Massif of presumably Gondwanan provenance. The Małopolska Massif was accreted to Baltica (East Europen Craton) probably during the Middle to Late Cambrian and was further deformed during the Caledonian Orogeny. The Brunovistulian Terrane was amalgamated with the Małopolska at the turn of Silurian and Devonian and is characterized by undeformed Lower Palaeozoic sediments (Belka *et al.*, 2002; Nawrocki and Poprawa, 2006). Lower Palaeozoic, Devonian and Carboniferous series cover both terranes. These were, however, partly disrupted during the late phases of the Variscan Orogeny and the Permo-Carboniferous phase of wrench faulting that accompanied the disintegration of the Variscan Orogen and the activation of the Teisseyre-Tornquist Zone (Ziegler, 1990). Mesozoic crustal extension, accompanying the development of the Tethyan rift system, controlled the subsidence of the Dyje/Thaya Depression in Southern Moravia and Lower Austria (2000 m of Jurassic and Upper Cretaceous deposits; Picha *et al.*, 2005) and the development of the Danish-Polish Trough that is superimposed on the Sorgenfrei-Tornquist and the Teis-

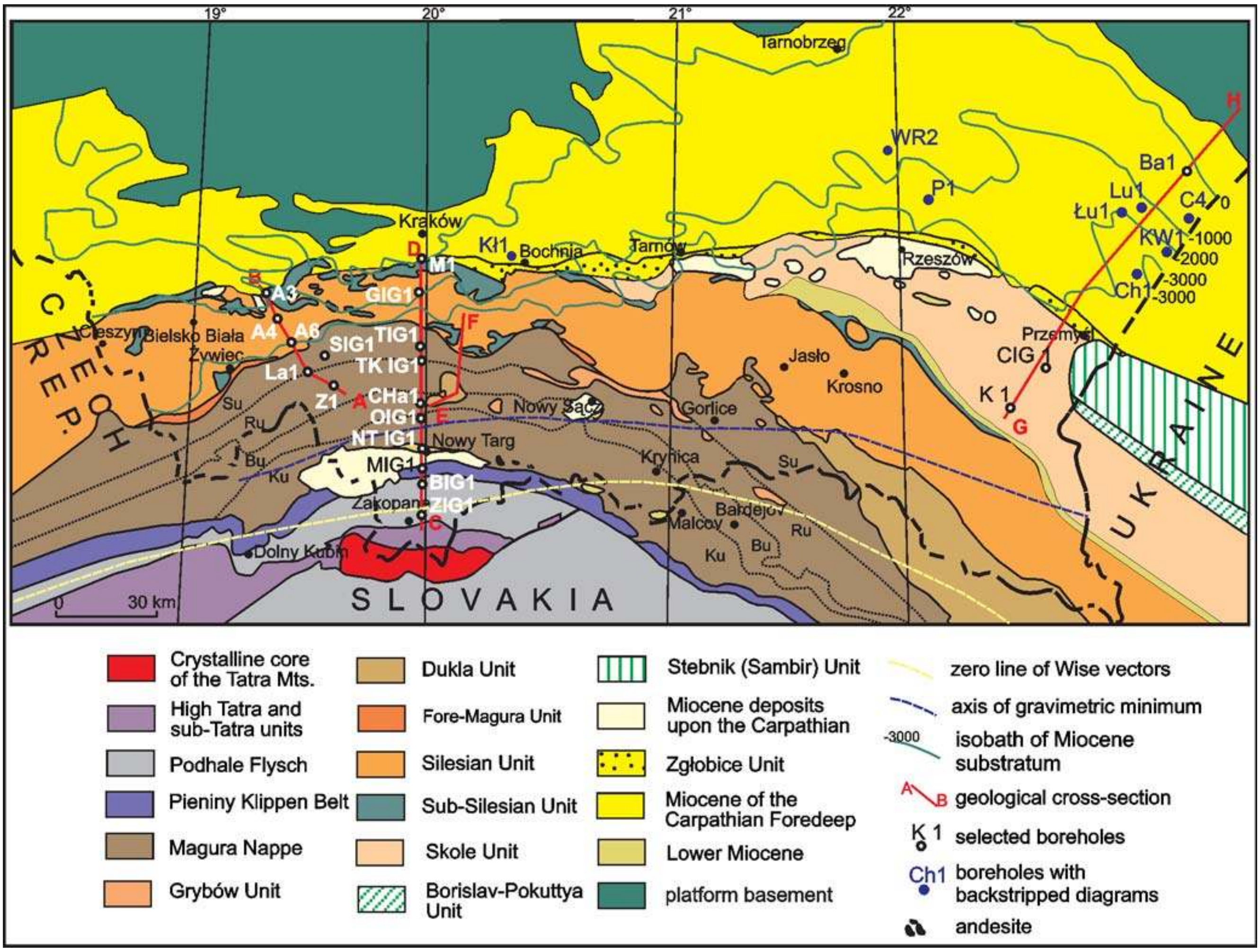


Fig. 2. Sketch map of the Polish Carpathians and their foredeep (based on Oszczypko, 1998; Oszczypko-Clowes, 2001)

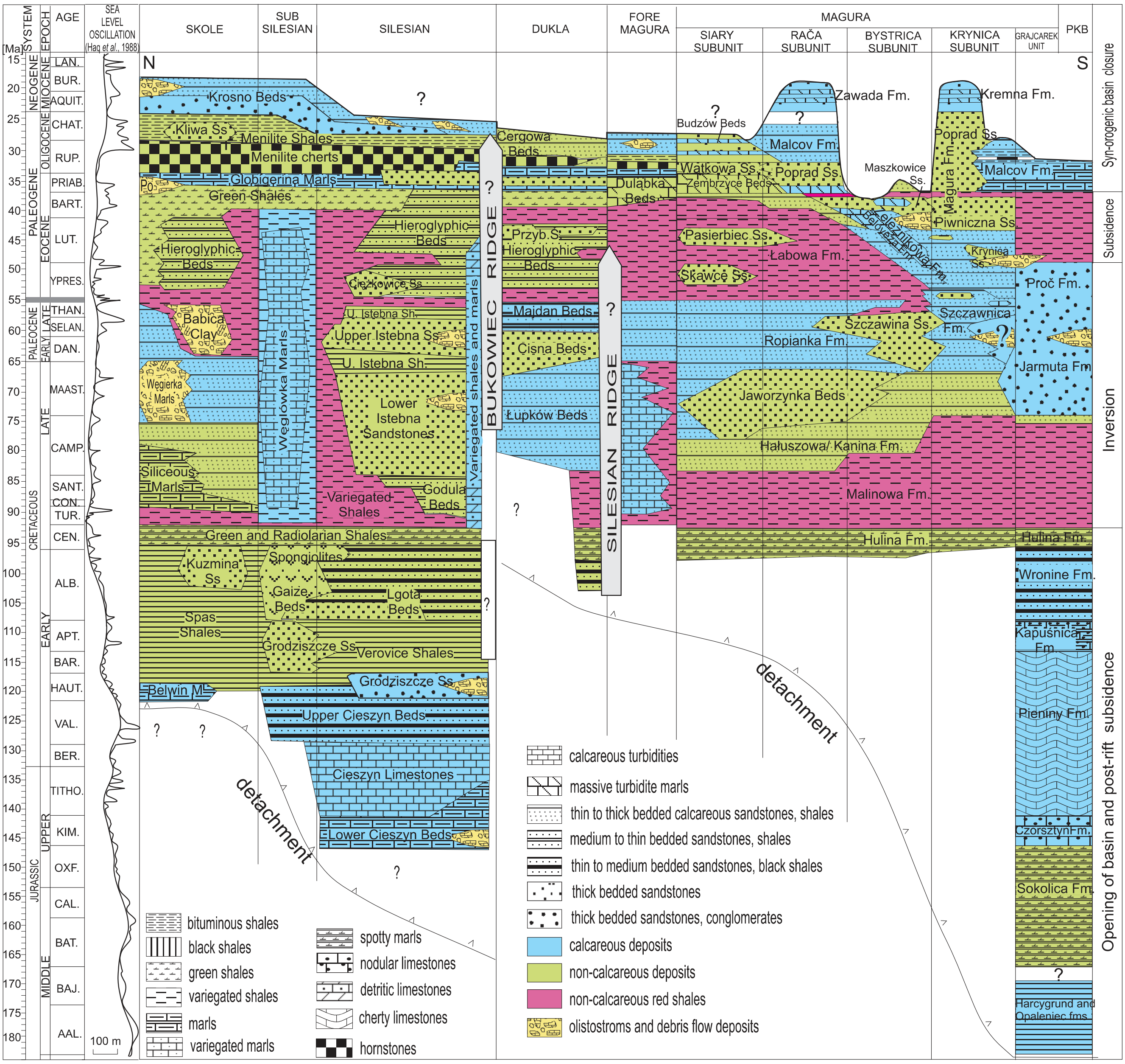
Su — Siary, Ru — Rača, Bu — Bystrica, Ku — Krynica Subunits of the Magura Nappe; boreholes: A 3 — Andrychów 3, A 4 — Andrychów 4, A 6 — Andrychów 6, La 1 — Łodygowice 1, Z 1 — Zawoja 1, SIG 1 — Sucha IG 1, ZIG 1 — Zakopane IG 1, BIG 1 — Bańska IG 1, MIG1 — Maruszyna IG 1, NT IG1 — Nowy Targ IG 1, OIG 1 — Obidowa IG 1, CHa1 — Chabówka 1, Tk IG 1 — Tokarnia IG 1, TIG 1 — Trzebunia IG 1, GIG 1 — Głogoczów IG 1, M 1 — Mogilany 1, Kł 1 — Kłaj 1, WR 2 — Wola Raniżowska 2, P 1 — Palikówka 1, K 1 — Kuźmina 1, CIG 1 — Cisowa IG 1, Ch 1 — Chotyniec 1, KW 1 — Kobylnica Wołoska 1, C 4 — Cetynia 4, Łu1 — Łukawiec 1, Lu2 — Lubaczów 2, Ba1 — Basznia 1; main groups of tectonic units of the Outer Western Carpathians after Książkiewicz (1977); Marginal Group (external): Borislav-Pokuttya, Stebnik (Sambir) and Zgłobice units; Middle Group (central): Grybów, Fore-Magura, Dukla, Silesian, Sub-Silesian, Skole units; and internal: Magura Group

seyre-Tornquist zones (Pożaryski and Brochwicz-Lewiński, 1978; Ziegler, 1990; Dadlez *et al.*, 1995). The Mid-Polish Trough extended south-eastward at least as far as the present-day Carpathian thrust front in Eastern Poland and the Western Ukraine (Pożaryski and Żytko, 1980; Kutek, 1994, 2001; Hakenberg and Świdrowska, 1997, 1998). Inversion of the Mid-Polish Trough during latest Cretaceous to Paleocene times can be correlated with collision events within the Alpine-Carpathian system (Kutek and Głazek, 1972; Krzywiec, 2002, 2006). Related deformation and uplift of the Carpathian foreland caused partial to total erosion of its Mesozoic sedimentary cover. The tectonic grain of the Carpathian foreland is related to this rifting and inversion tectonic activity that reflects repeated reactivation of the NW–SE striking Teisseyre-Tornquist Zone, which trends oblique to the POC. It is likely that during the Miocene development of the PCF reactivation of such inherited foreland structures influenced its geometry (Oszczypko *et al.*, 2005*a*).

The crust of the European foreland extends beneath the Carpathians at least as far as the PKB. Boreholes and seismic sections show that the depth to basement increases from a few hundred metres in the marginal part of the PCF to more than 7000 m beneath the POC (Figs. 3 and 4). Magneto-telluric sounding in the Polish Carpathians revealed a high resistivity horizon that is attributed to the top of the crystalline basement (Żytko, 1997). The top of the magneto-telluric basement reaches depths of about 3–5 km in the northern part of the Carpathians, descends to some 15–20 km at its deepest point (south of Krosno and north of Krynica) and rises to 8–10 km in their southern parts (north of PKB). The axis of the magneto-telluric basement low coincides, more or less, with the axis of a regional gravity minimum (Fig. 1, *cf.* Oszczypko, 1998, 2004).

Table 1

Lithostratigraphy of the Polish Outer Carpathians (compiled after Ślączka and Kaminski, 1998 and Oszczypko, 2004)



LAN. — Langhian, BUR. — Burdigalian, AQUIT. — Aquitanian, CHAT. — Chattian, RUP. — Rupelian, PRIAB. — Priabonian, BART. — Bartonian, LUT. — Lutetian, YPRES. — Ypresian, THAN. — Thanetian, SELAN. — Selandian, DAN. — Danian, MAAST. — Maastrichtian, CAMP. — Campanian, SANT. — Santonian, CON. — Coniacian, TUR. — Turonian, CEN. — Cenomanian, ALB. — Albian, APT. — Aptian, BAR. — Barremian, HAUT. — Hauterivian, VAL. — Valanginian, BER. — Berriassian, TITHO. — Tithonian, KIM. — Kimmeridgian, OXF. — Oxfordian, CAL. — Callovian, BAT. — Bathonian, BAJ. — Bajocian, AAL. — Aalenian; chronostratigraphic scale after Berggren *et al*. (1995)

South of this gravity minimum, geomagnetic sounding revealed a zone of zero values related to the Wise vectors (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 European foreland crust and the Central West-Carpathian Block (Żytko, 1997), also referred to as the ALCAPA Block (Csontos and Nagymarosy, 1998). In the Polish Carpathians, the depth of the crust-mantle boundary ranges between 30–40 km and generally decreases southward (see Guterch and Grad, 2006).

MAIN STRUCTURAL UNITS OF THE POLISH OUTER CARPATHIANS

Traditionally, three groups of structural units are distinguished in POC. These are the external Marginal, the central Middle, and the internal Magura Groups (Książkiewicz, 1977). The Marginal Group, which mainly involves folded Miocene rocks, forms in Poland a narrow zone along the Carpathian thrust front, consisting of the the Zgłobice and par-

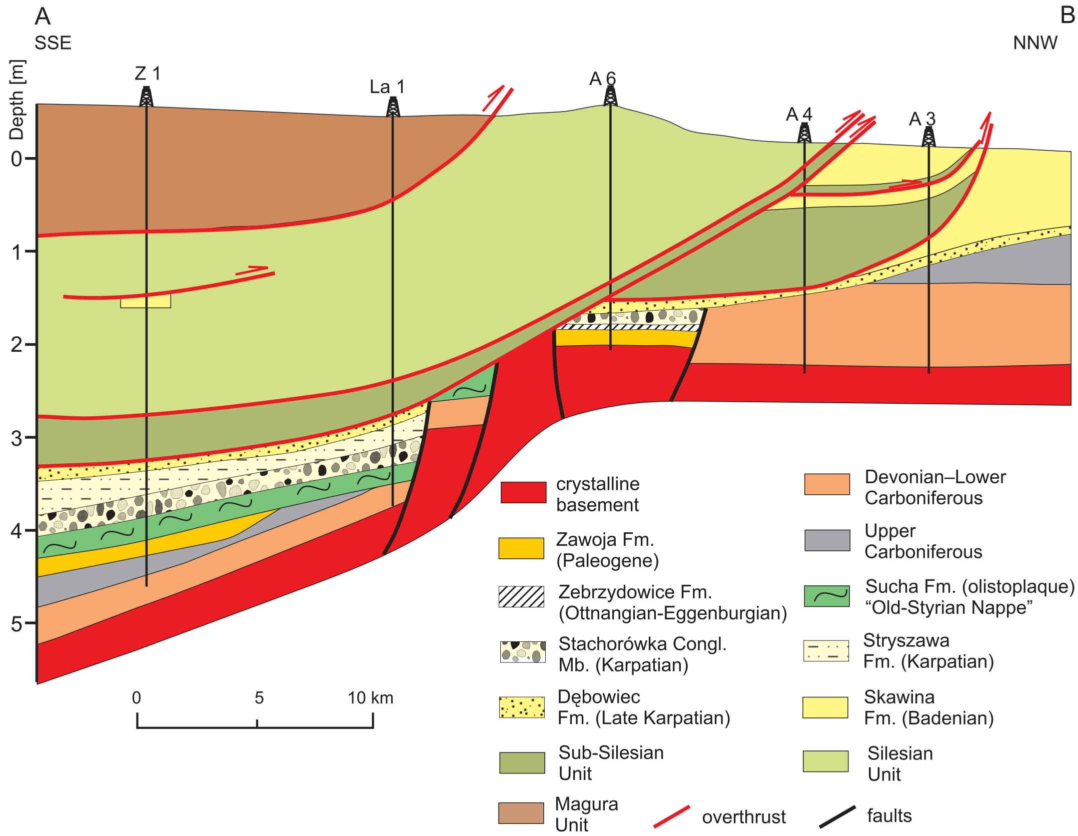


Fig. 3. Geological cross-section (A–B) Zawoja–Andrychów (modified after Oszczypko and Oszczypko-Clowes, 2003)

For location of cross-section and boreholes see Figure 2

tially the Stebnik structural units that are thrusted over the undeformed Miocene foredeep sequences (Oszczypko and Tomaś, 1985; Oszczypko, 1998; Oszczypko *et al.*, 2005*a*). In the Ukrainian Carpathians this zone widens out and involves the Boryslav-Pokuttya, Stebnik (Sambir) Nappes (Figs. 1 and 2). The Middle Group includes the Fore-Magura–Dukla, Silesian, Sub-Silesian and Skole Nappes, which involve the Early/Middle Miocene accretionary wedge and which are thrusted over the Marginal units. The nappes of the Middle Group form the core of the Western and Eastern Carpathians. The complex Magura Nappe, that involves a Late Oligocene/Early Miocene accretionary wedge, flatly overrides the nappes of the Middle Group (Figs. 3 and 4, cross-section C–D). The main detachment surfaces of the different Outer Carpathian Nappes are located at different stratigraphic levels (Table 1). For instance, the basal detachment surface of the Magura Nappe and the Fore-Magura group of units is located at the base of the Turonian-Senonian variegated shales (Oszczypko, 1992, 2004), whereas the main detachment surfaces of the Silesian, Sub-Silesian and Skole units (nappes) are located within Lower Cretaceous black shales. Whilst the basal detachment of the Marginal Boryslav-Pokuttya Nappe (Figs. 1 and 2) is located in Upper Cretaceous flysch, the Stebnik and the Zgłobice units were detached from their substratum in Lower and Middle Miocene shales, respectively (see also Bessereau *et al.*, 1996).

The Magura Nappe, the largest and innermost unit of POC, is mainly composed of Upper Cretaceous to Eocene sediments. The oldest Jurassic-Early Cretaceous rocks are known from the Peri-Pieniny Klippen Belt in Poland and a few localities in Southern Moravia (Birkenmajer, 1977; Švábenická *et al.*, 1997), whereas the youngest, Early Miocene deposits have recently been discovered in the area of Nowy Sącz (Oszczypko *et al.*, 1999; Oszczypko and Oszczypko-Clowes, 2002) and Peri-PKB area (Cieszkowski, 1992; Oszczypko *et al.*, 2005*c*). The Magura Nappe, which is delimited to the south against the PKB by a sub-vertical Miocene strike-slip fault, is flatly thrusted northward over the Fore-Magura group of units, and together with them upon the Silesian Nappe. The Magura Nappe accounts for at least 50 km of horizontal displacement (Figs. 2 and 4), of which more than 12 km were achieved in post-middle Badenian times (Oszczypko, 2004). Its northern limit is erosional and was shaped during the denivelation of the Magura forefield. A tectonic window zone is located 10–15 km south of the northern limit of the Magura Nappe. The largest is the Mszana Dolna tectonic window (Oszczypko-Clowes and Oszczypko, 2004), located in the middle segment of the POC (Fig. 4, cross-section E–F). South of the tectonic window zone the inclination of the Magura sole thrust increases with the thickness of the Magura Nappe attaining more than 4–5 km at the northern boundary of PKB (Fig. 4, cross-section C–D). This nappe can be subdivided into four structural subunits,

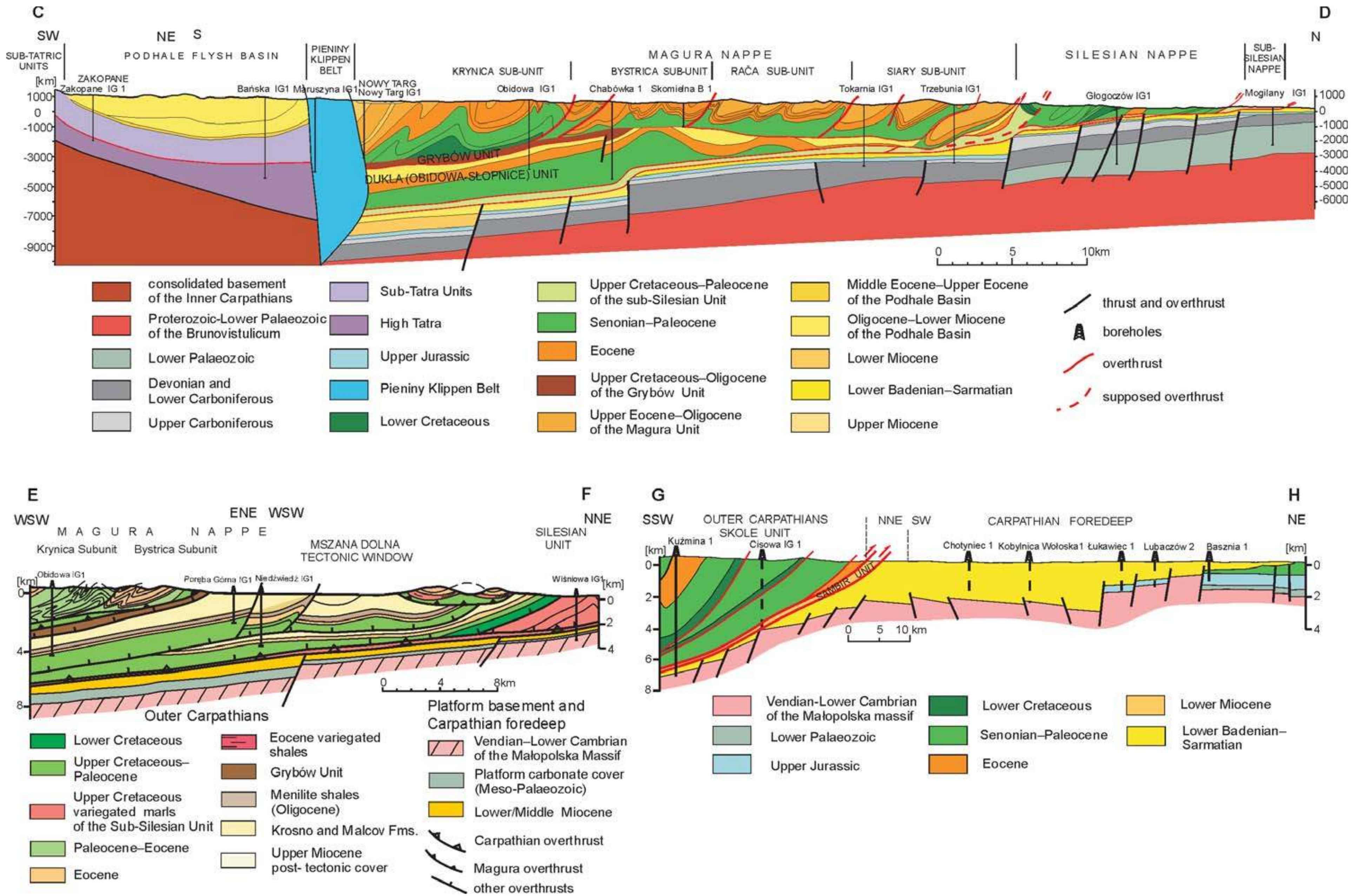


Fig. 4. Geological cross-sections: C–D — Zakopane–Kraków (modified after Sikora *et al*., 1980); E–F — Obidowa IG 1–Wiśniowa IG 1 (partly from Oszczypko, 2004); G–H — Kuźmina 1–Lubaczów (modified after Oszczypko *et al*., 2005*a*)

For location of cross-sections and boreholes see Figure 2 namely the Krynica, Bystrica, Rača and Siary subunits (Fig. 4). These differ in the facies development of their Late Cretaceous and Paleogene series (Table 1).

The Fore-Magura group of units consists of several nappes, of which the Dukla Nappe is the largest that is exposed in the eastern part of the POC (Fig. 2). In the Zakopane-Kraków geotraverse (Fig. 4, cross-section C–D), several boreholes reached this nappe at depths of a few kilometres.

The Silesian Nappe occupies the central position of the Western Outer Carpathians. In the western sector of the POC, the Silesian Nappe is composed mainly of Upper Cretaceous thick-bedded turbidites and attains a thickness of around 2 km. Eastward, the thickness of this nappe increases to 5–7 km (up to 3.5 km of the Oligocene Krosno Beds). The strongly deformed Sub-Silesian Nappe, composed mainly of Upper Cretaceous-Eocene pelagic variegated marls, is located along the northern margin of the Silesian Nappe. Beneath the latter, however, the Sub-Silesian Nappe is very thin and acts as a kind of tectonic lubricant for the higher units (Figs. 3 and 4).

The Skole Nappe occupies the outermost position in the POC and widens towards the east. This unit consists of several elongated thrust-sheets (Fig. 4, cross-section G–H), known in Polish as “skibas” (see Fig. 1, Skyba Nappe of the Ukrainian Carpathians).

A correlation of the principal structural units of the Outer Western Carpathians, that is still valid today, was already established in the 1970’s (see Żytko *et al.*, 1989; Lexa *et al.*, 2002).

The main structural units can be traced without interruption between 17 and 24° meridians, but towards the west and east the Silesian Unit disappears beneath the Magura and Audia units, respectively (Fig. 1). To the SW, in Czech Republic, the position of the Silesian Unit is occupied by the thin-skinned Ždánice–Sub-Silesian Unit (Picha *et al.*, 2005). At the same time the new and more external, allochthonous Pouzdrany Unit and ultimately the Waschberg Zone of Austria appear at the front of Sub-Silesian Unit. The correlation between the Middle Group of structural units in the Polish and Ukrainian Carpathians has been discussed in detail by Żytko (1999). Close to the Polish/Ukrainian boundary, the Marginal Cretaceous Zone of the Sub-Silesian/Silesian units plunges in the core of anticline composed of the Upper Oligocene-Lower Miocene Upper Krosno Beds of the Skole Unit (see Książkiewicz, 1977). The eastern prolongation of the Sub-Silesian facies is locally marked by the occurrence of variegated marls (Rozluch and Holyatyn folds). The southern part of the Silesian Unit of Poland, including the Fore-Dukla Unit and Bystre thrust sheet, can be correlated with the Chornohora Unit of the Ukraine, and the Audia Unit (Fig. 1) of the Romania (Ślączka *et al.*, 2005). According to Shakin *et al.* (1976) and Burov *et al.* (1986) the SE prolongation of the Dukla Unit corresponds to the Krasnoshora and Svidovets subunits (see also Ślączka *et al.*, 2005). The southernmost units of the Ukrainian Carpathians (the Porculets, Rakhiv and Kamyany Potik units, Oszczypko, 2004) can be correlated with the Ceahleu and Black Flysch units of the Romanian Eastern Carpathians, known as the Outer Dacides (Fig. 1). In the Western Outer Carpathians there are no equivalents of these units.

To the west, the Magura Nappe links up with the Rheno-Danubian Flysch of the Eastern Alps (Fig. 1). The Rheno-Danubian–Magura Nappe extends through Czech Republic, Poland and Eastern Slovakia before it disappears beneath Miocene volcanics east of Uzhhorod in the Trans-Carpathian Ukraine. In the SE part of the Ukrainian and in the Romanian Carpathians, the Magura Nappe corresponds to the Marmarosh Flysch Zone (Sandulescu, 1988; Oszczypko, 1992; Aroldi, 2001; Oszczypko *et al.*, 2005*b*) that has been subdivided into the external Vezhany and the internal Monastyrets units (Smirnov, 1973). Taking into account the internal position of the Marmarosh Flysch with respect to the Marmarosh Massif and similarities between the Vezhany and Fore-Magura successions, the Monastyrets/Petrova and Rača successions, and the Botiza and Krynica successions, it is likely that the Marmarosh Massif and the buried Silesian Ridge took in almost the same palaeogeographical position.

THE OUTER CARPATHIAN BASINS IN THE ALPINE-CARPATHIAN SYSTEM

The Alpine-Carpathian system forms part of the Alpine-Mediterranean system of orogenic belts that evolved in response to closure of the Western Tethyan system of oceanic basins during the Middle Cretaceous to Miocene convergence and collision of the African and European plates (Cavazza *et al.*, 2004). The domain of the Outer Carpathian basins evolved during the Jurassic and Cretaceous into a rifted passive margin that faced the eastern parts of the oceanic Alpine Tethys. Following closure of this oceanic basin during the Late Cretaceous and collision of the Inner Western Carpathian Orogenic Wedge (IWCW) with the Outer Carpathian passive margin at the Cretaceous-Paleocene transition, the latter was transformed into a flexural foreland basin that was progressively scooped out by nappes and thrust sheets during the Late Eocene and Miocene (Oszczypko, 1999).

The present-day configuration of the Outer Carpathian Flysch Belt resulted from late Early Miocene and Middle Miocene folding and thrusting (Oszczypko, 1997, 1998, 2004). In the Eastern Carpathians, these tectonic processes were completed during Pliocene times. The formation of the Outer Carpathians was strongly related to the eastward escape and rotation of the ALCAPA and Tisza Mega-Blocks.

During the last decade several attempts have been undertaken to reconstruct the syn-orogenic Cretaceous and Cenozoic evolution of the West Carpathian basins (e.g. Ziegler, 1990; Kovač *et al.*, 1998; Dercourt *et al.*, 2000; Golonka *et al.*, 2000, 2003, 2005; Plašienka, 2000; Haas and Csaba, 2004). All these reconstructions are very general and often contradict each other.

Recently Schmid *et al.* (2005) reviewed the most important tectonic units of the Alps, Dinarides and Carpathians and discussed the links between the different orogens of this system. Of special interest to the West Carpathian domain, which flanks the southern margin of the European Platform are:

1. The Miocene deformation of the internal foredeep, controlling development of Tarcau–Skole, Audia–Macla–Convolute Flysch, Silesian–Sub-Silesian and Dukla nappes.
2. The Europe-derived allochthons, including the Helvetic and Subpenninic units of the Alps, Danubian Nappes of the

South Carpathians, the Briançonnais terrane of the Alps and the Bucovinian–Gethic–Srednia Gora nappe system.

3. The ophiolite suture zones and accretionary prisms containing oceanic components related to the Ceahlau–Severin and Valais–Rhenodanubian (North Penninic) oceans, the Magura Flysch, the Pieniny Klippen Belt, the Piemont–Liguria-Krichevo–Szolniok–Sava Ocean, the Meliatia–Darno–Szarvarsko–western "Vardar"–Dinaridic–Mirdita ophiolites, the Jurassic accretionary prisms of the "Vardar" and Meliata oceans, and Transylvanian–South Apuseni–eastern Vardar obducted part of the Vardar Ocean.

4. The allochthonous Apulia-derived "ALCAPA Block" and the "Tisza Block" that has a mixed European and Apulian affinity.

The Outer Carpathian Flysch belt (Fig. 5) consist of an external unit, referred to as the Silesian-Krosno group or Moldavides (Sandulescu, 1988), and the internal Magura Nappe group. During the Late Jurassic–Early Cretaceous, the Outer Carpathian domain (Skole, Sub-Silesian and Silesian sub-basins and their eastern prolongation) formed part of the rifted European margin (Oszczypko *et al.*, 2003; Ślączka *et al.,* 2005; Golonka *et al.*, 2005). In Moravia and the Cieszyn-Żywiec area of Poland, Jurassic rifting was accompanied by volcanic activity (teschenite sills, dykes and local pillow lavas) that persisted until the end of the Hauterivian (Lucińska-Anczkiewicz *et al.*, 2000; Grabowski *et al.*, 2004) or Aptian (Ivan *et al.*, 1999). This rifted continental margin probably extended into the East Carpathian (basic effusives, Tithonian-Neocomian "Black Flysch" of the Kamyany Potic scale and Rakhiv "Sinaia" beds) and South Carpathian domains (Severin Zone, see Sandulescu, 1988; Ceahlau-Severin Ocean, Schmid *et al.*, 2005), and probably eastward into Black Sea area (Golonka, 2000; Golonka *et al.*, 2000, 2005; Senkovsky *et al.*, 2004). The relation between the Silesian/Sub-Silesian Basin and the Helvetic Shelf is not clear. Maybe both basins were in contact with each other along the Vienna Tranform Fault (VTF), known also as the West-Carpathian Transfer Zone (Picha *et al.*, 2005). In this case, it is possible that the Upper Cretaceous pelagic marls know from the Helvetic units extended eastward as

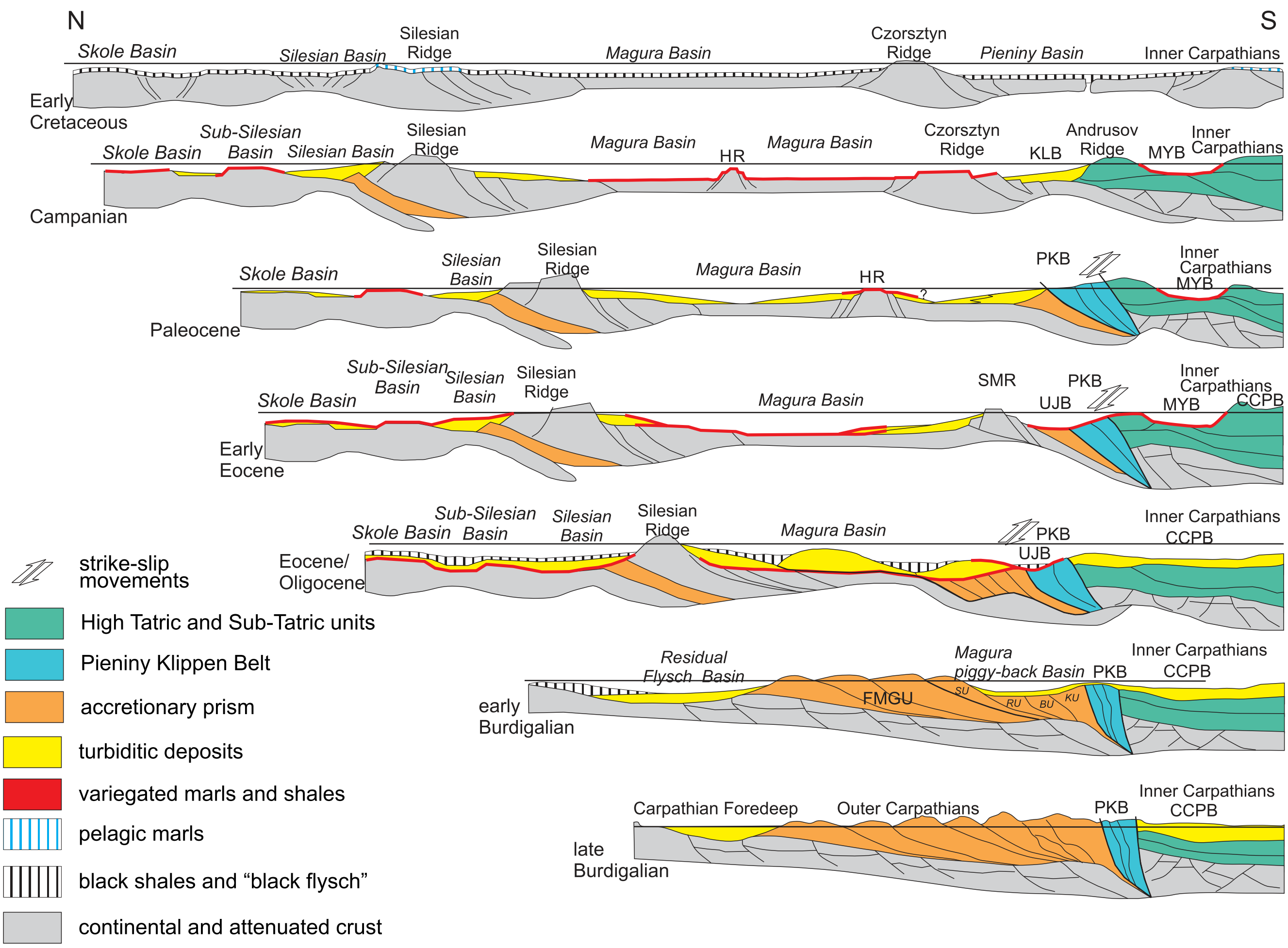


Fig. 5. Early Cretaceous–Early Miocene palinspastic evolutionary model for the Western Carpathians, not to scale (based on Oszczypko, 1999, supplemented)

CCPB — Central Carpathian Paleogene Basin, FMGU — Fore-Magura group of units, Su— Siary Subunit, Ru — Rača Subunit, Bu — Bystrica Subunit, Ku — Krynica Subunit, HR — Hluk Ridge, UJB — Ujak Basin, KLB — Klapa Basin, MYB — Myjava Basin, PKB — Pieniny Klippen Belt, SMR — South Magura Ridge the Sub-Silesian marls, whereas the Silesian rift and the Silesian Ridge terminated against the VTF.

The Magura Basin occupied a more internal position with respect to the Silesian-Krosno (Moldavides) Basin with the continental Silesian Ridge separating the two basins during the Late Cretaceous and Paleogene (Książkiewicz, 1962; Unrug, 1979; Oszczypko, 1992, 1999; Picha *et al*., 2005). Its possible (see Săndulescu, 1988; Oszczypko, 1992, 2004; Oszczypko *et al*., 2005*b*) that during the Late Cretaceous the Magura and Marmarosh Flysch was deposited in a continuous basin that was bounded to the north by the Silesian Ridge and the Marmarosh Ridge (Bucovinian–Getic–Srednia Gora nappe system, see Schmid *et al*., 2005).

The opening time of the Magura Basin is still under discussion. Traditionally an Early/Middle Jurassic age is accepted (see Birkenmajer, 1986; Oszczypko, 1992, 1999; Golonka *et al.*, 2000, 2003, 2005) that is essentially coeval with the opening of the S–Penninic–Piemont–Ligurian Ocean (Schmid *et al*., 2004*b*), that its supposed to find its prolongation in the Pieniny Ocean (Golonka *et al.*, 2000). The submerged continental (?) Czorsztyn (Oravicum) Ridge (Fig. 5) separated the Pieniny Ocean into a SE arm, referred to as the oceanic Vahicum Basin, and a NE arm that corresponds to the Magura deep-water basin that flanked the European Shelf to the south (Oszczypko, 1999). Whether, and to what extent, the Magura Basin was floored by oceanic crust is still a matter of dispute.

Alternatively, Plašienka (2002, 2003) suggests a Late Jurassic–Early Cretaceous opening for the Magura Basin. According to this scenario, the Early Cretaceous opening of the Magura Basin was accompanied (?) by thermal uplift of the Czorsztyn Ridge and post-rift thermal subsidence of the Magura Basin, resulting in uniform deposition of pelagic and hemipelagic shales below CCD (Calcium Carbonate Compensation Depth).

The latter scenario is compatible with the concepts of Schmid *et al.* (2005) that links opening of the Magura Basin with the Late Jurassic–Early Cretaceous opening of the oceanic Valais-Rhenodanubian (North Penninic) Basin. This northern branch of the Alpine Tethys, that is located north of the Iberia-Briançonnais (Mid-Penninic) Block, extended from the Pyrenees to the Western Carpathians. The Czorsztyn (Oravicum) Ridge that occupied a more internal position with respect to the Magura Basin can be regarded as an equivalent of the continental Briançonnais domain that was flanked to the south by the oceanic Ligurian–Piemont–Vahicum–Krichevo–Szolniok–Sava domain (see Schmid *et al.*, 2004*a*, *b*, 2005; Golonka *et al.*, 2005). This oceanic area was limited to the south by the Apulian Margin that is now represented by the Apulia-derived units. The Szolnok-Sava ophiolite suture zone is now partly incorporated into Mid-Hungarian fault system.

SOURCE AREAS

The Outer Carpathian sedimentary basin complex was supplied with clastics that were derived from external as well as internal source areas, with the latter being referred to as "cordilleras" (Książkiewicz, 1962). During the Miocene evolution of the Outer Carpathian fold-and-thrust belt, its sediments were detached from the basement, imbricated and stacked into nappes that were transported towards the foreland and account for a few hundred kilometres of shortening (Fig. 5). Our understanding of the geological structures controlling the Carpathian source areas is based on the investigations of sedimentary blocks and "exotic" pebbles that were transported into basinal areas by submarine gravity flows (see Książkiewicz, 1962).

The marginal Skole–Sub-Silesian and Silesian Flysch basins were supplied both from northern external and southern internal source areas. The northern source area can be regarded as an uplifted massif that was located along the southern margin of the European Platform. K/Ar isotopic dates obtained on crystalline exotic rocks derived from this northern source area suggest affinities to the low-grade metamorphic rocks of the Brunovistulicum and the Małopolska Massif that were deformed during the "Cadomian" and Sandomirian orogenies, respectively (Poprawa *et al.*, 2004).

In the Outer Carpathian sedimentary basin system the most important internal source area was the "Silesian Ridge (Cordillera)" (Figs. 5 and 6) that corresponded to the continental Silesian, Andrychów and Marmarosh ridges (Książkiewicz, 1965; Unrug, 1968; Golonka *et al.*, 2000, 2005; Picha *et al.*, 2005). According to Unrug (1968), the Silesian Ridge "paralleled the long axis of the flysch trough" and separated the northern Silesian Basin from the southern Magura Basin. Exotic rocks from the Fore-Magura units display a complex lithologic composition, and document changes in the sedimentary supply to this basin. During the Late Cretaceous-Paleocene the Fore-Magura succession (supplied from the north) formed a part of the Magura Basin (Fig. 6), whilst during the Late Eocene and Oligocene this succession was part of the Silesian Basin that was supplied from the south (see Unrug, 1968). Isotopic ages of "exotic" pebbles shed from the Silesian Ridge into the Silesian, Dukla and Magura (Rača Subunit) basins document a Variscan age of plutonic and metamorphic rocks (Poprawa *et al.*, 2004).

A further intrabasinal source area that is located on the boundary of the Silesian and Dukla basins is known as the Bukowiec Ridge (Ślączka, 2005). During the Late Cretaceous-Paleocene period the Bukowiec Ridge supplied the Dukla Basin, whereas during the Oligocene material derived from this source was deposited in the Silesian Basin. Exotic rocks contained in the Oligocene Krosno Beds of the Silesian Basin consist of a variety of metamorphic rocks (eg. phyllites, gneisses, amphibolites, quartzites and marbles) and blocks of Late Eocene shallow-marine limestones and marls (Ślączka, 2005). In this area a rounded block of anchi-metamorphic rocks recording Albian cooling has been found (Poprawa *et al.*, 2004). The age of this metamorphism is coeval with the Middle Cretaceous collision along the Outer/Medium Dacide (Marmarosh Massif) boundary in the Eastern Carpathians (Sandulescu, 1988).

During the Campanian, inversion-related uplift of the Silesian Ridge affected the northern part of the Magura Basin where it was accompanied by the onset of flysch deposition (Figs. 5 and 6). By contrast, along the southern margin of the Magura Basin the onset of flysch deposition occurred at the Maastrichtian-Paleocene transition as manifested by the conglomerates and olistoliths of the Jarmuta and Proč formations (Fig. 6, Table 1) (Birkenmajer and Oszczypko, 1989; Mišik *et al*., 1991*a*). Source areas for these clastics were uplifted exotic blocks, including internal elements of the PKB (Książkiewicz, 1977; Oszczypko *et al*., 2005*c*). This is attributed to the collision of the Inner Western Carpathian (ALCAPA) Block with the Czorsztyn (Oravicum) Ridge (see Plašienka, 2003) and/or the Andrusov Ridge (Birkenmajer, 1986, 1988).

During the Early Eocene, a deep-water submarine fan started to develop in the southern part of the Magura Basin, as evidenced by the occurrence of channel-lobe turbidites supplied from SE sources (Table 1, Krynica succession). The Eocene deposits of the Krynica Zone of the Magura Basin contain fragments of crystalline rocks, derived from a continental type of crust, and infrequent clasts of Mesozoic deep- and shallow-water limestones. Mišik *et al.* (1991*b*) suggested that this material was derived from "the basement of the Magura Basin", but differs from that of the Czorsztyn (Oravicum) Ridge, that was exhumed during the Early/Middle Eocene (Fig. 5). Alternatively, this clastic material may have been derived from an Inner Carpathian type source area, located on the SE margin of the basin (e.g. tip of the ALCAPA Block, see Plašienka, 2000).

THE EVOLUTION OF THE OUTER CARPATHIAN BASINS

The Outer Carpathians are composed of Late Jurassic to Early Miocene flysch-dominated series. The main structural units differ, however, in the facies development and thickness of their sedimentary sequences. The thickest sedimentary sequences occur in the Silesian Unit where they vary between 3000 m in the west and over 5000–7000 m in the east. The other units involve distinctly thinner sedimentary sequences that vary between 3000–3800 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.*, 2002). Taking facies distribution, sediment thickness and palaeocurrent directions into account (see Książkiewicz, 1962), only the Magura, Silesian and Skole basins can be considered as independent depositional entities (see Nemčok *et al.*, 2000). During the Late Cretaceous-Eocene period, the Sub-Silesian domain formed a sub-marine high that separated the Skole and Silesian basins (Fig. 5). The history of the Dukla domain, which played the role of a transfer zone between the Magura and Silesian basins, was more complex. According to reconstructions by Roure *et al.* (1993) and Behrmann *et al.* (2000), the Outer Carpathian Basin complex had during the Early Oligocene a width of at least 380 km along the Przemyśl–Hanušovce geotraverse, located near the Polish-Ukrainian border. This restoration does, however, not include the Silesian Ridge that was at least 20–50 km wide (Unrug, 1968) and separated the Magura and Silesian domains. Assuming for the Magura Basin a remnant width of about 100 km, this suggests that during Early

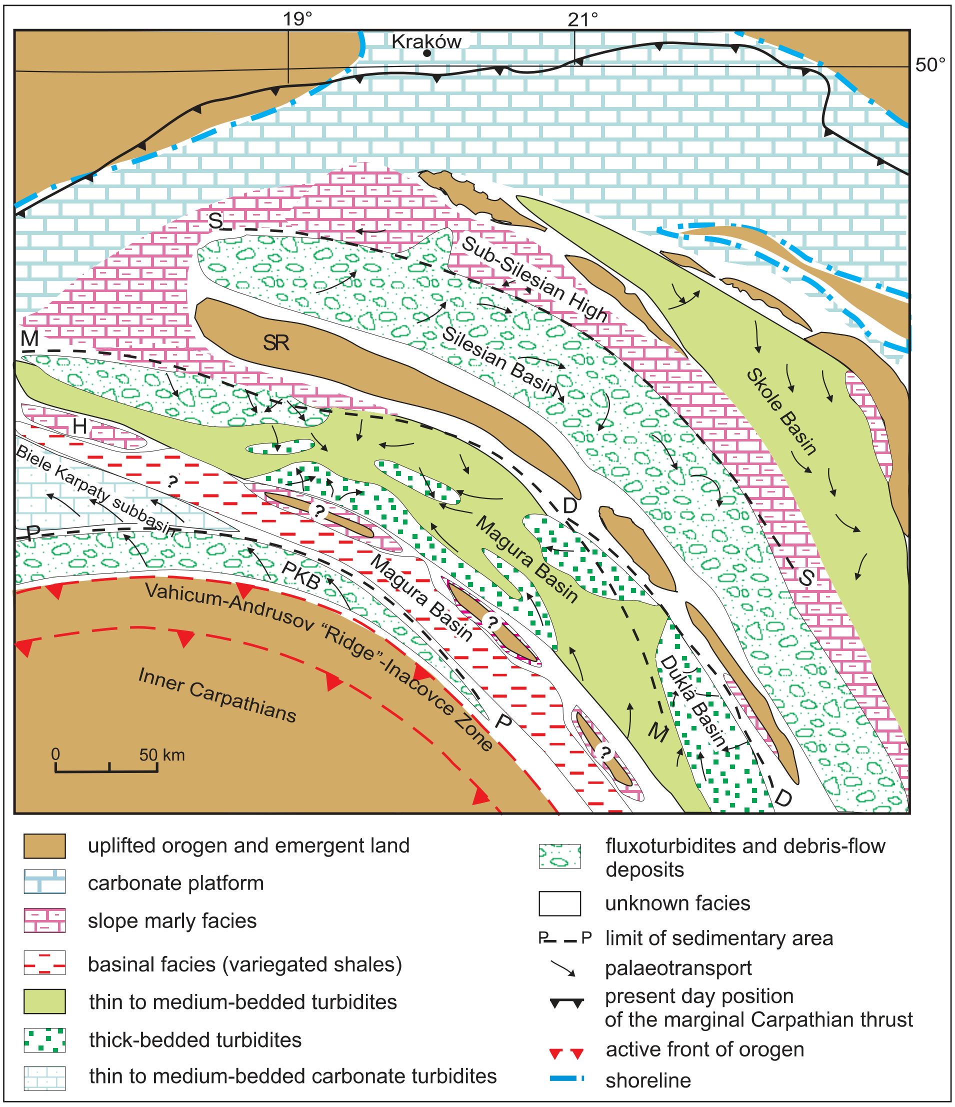


Fig. 6. Maastrichtian palaeogeographic and palinspastic map of the Northern Carpathians (partly after Książkiewicz, 1962; Oszczypko and Salata, 2005)

H — Hluk submerged ridge, SR — Silesian Ridge, PKB — Pieniny Klippen Belt

Oligocene times the entire Outer Carpathian domain was at least 500 km wide.

The traditional view that the Magura and Silesian basins paralleled each other (see Książkiewicz, 1962; Unrug, 1968, 1979; Birkenmajer, 1986) was recently questioned by Nemčok *et al.* (2000) who place the Magura Basin to the south-west of the Silesian Basin, whilst the present-day juxtaposition of these units resulted from Miocene eastwards transport of the Magura Nappe. This concept is, however, not compatible with the facies distribution and palaeocurent measurements in the Polish Outer Carpathians, nor with the transitional position of the Dukla succession between those of the Magura and Silesian basins (Fig. 6). The sedimentary succession of the Outer Carpathians (Table 1) comprises sequences that can be related both to the divergent and convergent evolutionary stages of the Tethyan-Alpine system (Picha *et al.*, 2005). Moreover, this succession reveals different mega-sequences, which reflect the main tectonic stages of basin development and global changes in relative sea level (Poprawa *et al.*, 2002; Oszczypko, 2004).

MIDDLE JURASSIC (150 MA)–EARLY CRETACEOUS (125 MA) OPENING OF BASINS AND POST-RIFT SUBSIDENCE

The Early/Middle Jurassic opening of the Magura Basin is rather speculative, because the Magura Nappe was detached from its substrate roughly at the base of the Upper Cretaceous sequence (Golonka *et al.*, 2000, 2003; Oszczypko *et al.*, 2003). However, a more or less complete Jurassic–Lower Cretaceous section attributed to the Magura Basin occurs only in the Pieniny Klippen Belt where it forms the Grajcarek Unit (Birkenmajer, 1986). These deposits consist of deep water, condensed pelagic limestones and radiolarites, whereas shallower facies are known from the Czorsztyn succession (Table 1).

During Kimmeridgian–Valanginian times, development of the Proto-Silesian Basin (Sileso-Moravian Beskides) commenced (Ślączka *et al.*, 1999; Golonka *et al.*, 2000, 2003, 2005). In this basin, which probably extended into the Eastern Carpathians (Ukraine and Romania, see Sandulescu, 1988), dark marls were deposited that are followed by calcareous turbidites derived from reef-fringed carbonate platforms (Cieszyn Limestones, see Słomka, 1986; Izotova and Popadyuk, 1996) that were dominated by shallow-water environments (Table 1). Regional subsidence of the Silesian Basin was controlled by rifting of the European Platform, and was accompanied by the extrusion of basic lavas (Birkenmajer, 1986; Sandulescu, 1988; Laschkevitch *et al.*, 1995; Lucińska-Anczkiewicz *et al.*, 2002; Grabowski *et al.*, 2004). During this stage the basin reached neritic to bathyal depths with subsidence rates attaining as much as 69 m/My (Ślączka *et al.*, 1999).

Development of the Outer Carpathian basins was controlled during Late Jurassic–Aptian times by normal faulting and syn-rift subsidence that was accompanied in the Western Carpathians by the extrusion of alkali-basalts ranging in age from Barremian to Aptian (Ivan *et al.*, 1999). This was followed by post-rift thermal subsidence, resulting in the Albian-Cenomanian expansion of deep-water facies (Nemčok *et al.*, 2001; Poprawa *et al.*, 2002). Hauterivian-Aptian series are characterized by the dark, silty and siliceous Verovice and Spas Shales, and by sedimentation rates decreasing from 40 to 12 m/My (Figs. 7 and 8) in the Silesian and Skole basins, respectively (Ślączka *et al.*, 1999; Poprawa *et al.*, 2002) in which water depths had increased to bathyal to abyssal conditions.

In the southeastern Outer Carpathians (Gethic and Marmarosh massifs and Ceahlau-Rakhiv units) compressional events occurred during the late Aptian to Albian (Sandulescu, 1988; Kruglov, 1989). These are manifested by intense folding, ending in Albian times, and the deposition of coarse clastic sediments, such as the uppermost Albian-Cenomanian Bucegi conglomerates and the Soymul Beds, and in the more distal part of the basin, the thick Biela Tisa and Upper Shipot turbidite complexes. In the Western Carpathians, this compressional episode is manifested by the uplift of intra-basinal ridges, the deposition of siliciclastic turbidites in the Silesian Basin (Lower Lgota Beds) and the Magura Basin (Gault Flysch) that was accompanied by an acceleration of sedimentation rates to 31–63 m/My, and the development of synsedimentary folds. Similar deposits are reported from the southern part of the Magura Nappe in Moravia (Švábenická *et al.*, 1997). In the Magura succession of the PKB, Albian-Cenomanian series are developed in "Black Flysch" facies (Oszczypko *et al.*, 2004; see also Birkenmajer and Gedl, 2004). With the latest Albian-Cenomanian deactivation of siliciclastic source areas, a uniform pelagic environment was gradually established in the Skole, Sub-Silesian/Silesian, Magura and PKB basins under which spotty marls, green radiolarian shales and radiolarites were deposited. The Cenomanian highstand in sea level resulted in the establishment of uniform sedimentary conditions in all Outer Carpathian basins and the deposition of green radiolarian shales (Cenomanian Key Horizon) that are followed by Turonian variegated shales. During the Cenomanian, sedimentation rates decreased sharply to 4–6 m/My, whereas water depths increased to abyssal conditions (Ślączka *et al.*, 1999; Poprawa *et al.*, 2002).

LATE CRETACEOUS (100 MA)–PALEOCENE (35 MA) INVERSION

During the Late Cretaceous progressive closure of the Pieniny Basin (Vahic Ocean) (Figs. 5 and 6) compressional stresses began to build up in the Outer Carpathian basins. In their western part the Silesian Ridge was reactivated and uplifted during the Turonian. This inversion pulse affected most of the Silesian, Sub-Silesian and Skole sub-basins (Fig. 7), and from the Campanian onward, also the northern part of the Magura Basin (Poprawa *et al.*, 2002; Oszczypko, 2004; Oszczypko *et al.*, 2005*d*). From uplifted and exposed ridges coarse clastics were shed into the adjacent basins. In the Skole and Sub-Silesian–Silesian basins this was accompanied by an increase in sedimentation rates up to 25 and 100 m/My, respectively (Poprawa *et al.*, 2002). In the western part of the Silesian Basin, sedimentation rates reached 400 m/My during the deposition of the Coniacian-Campanian (Fig. 8; see also Poprawa *et al.*, 2002; Oszczypko *et al.*, 2003).

During Maastrichtian-Paleocene times (Fig. 6), the Inner Western Carpathian Orogenic Wedge (IWCW) collided with the Czorsztyn Ridge (Plašienka, 2002, 2003; Oszczypko *et al.*, 2005*c*). In the southern part of the Magura Basin this is reflected the deposition of the up to 500 m thick Jarmuta Flysch

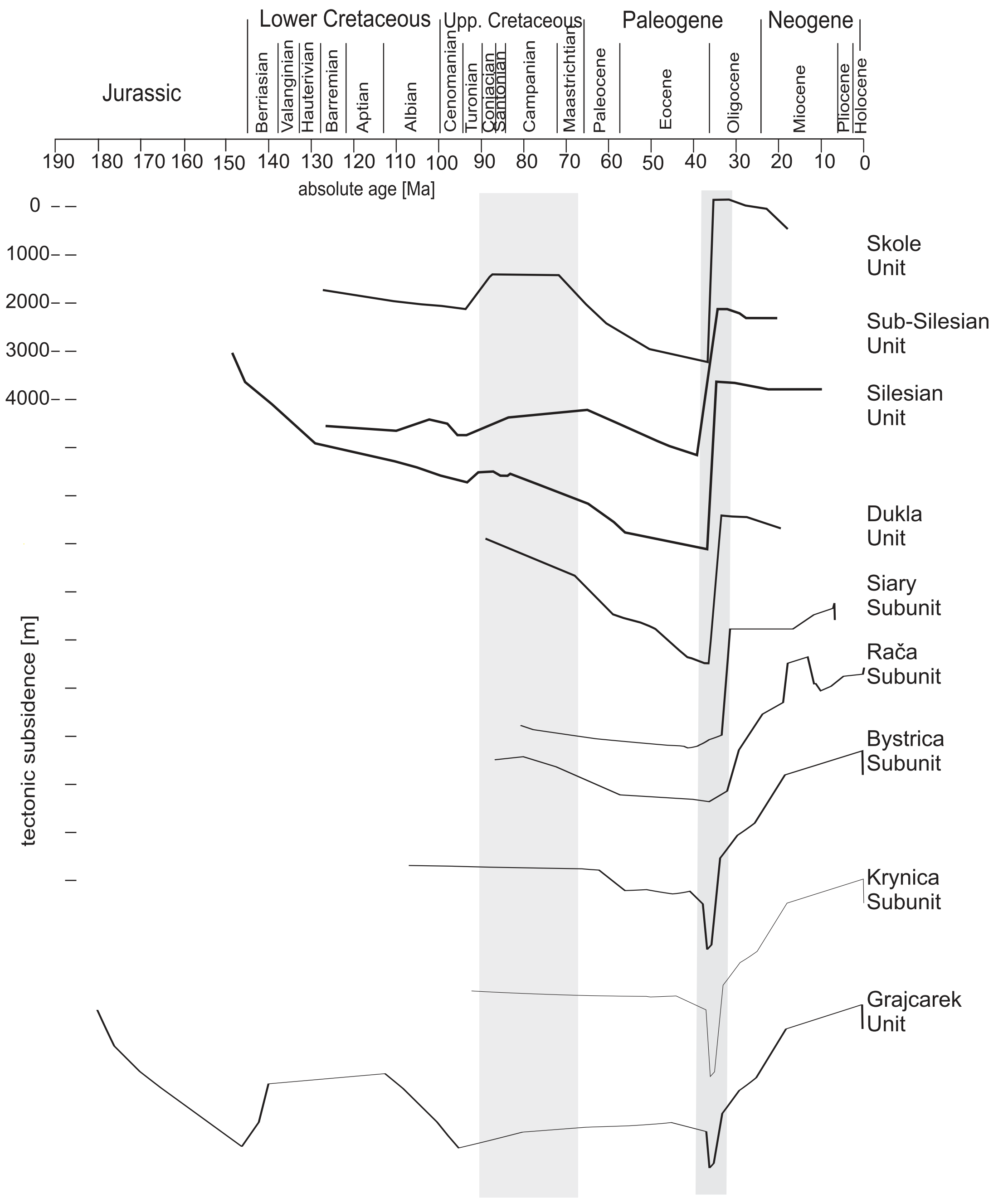


Fig. 7. Tectonic subsidence curves for selected synthetic profiles from the Polish Outer Carpathians (after Poprawa *et al*., 2002; Oszczypko *et al*., 2003)

Grey shaded bars indicate Late Cretaceous–Paleocene and Late Eocene–Early Oligocene tectonic uplift events

that is dated as Maastrichtian/Paleocene (Birkenmajer, 1977; Birkenmajer and Oszczypko, 1989). This formation is composed of thick- to medium-bedded turbidites containing conglomerates and breccias that consist of Jurassic and Cretaceous sedimentary components, as well as of exotic crystalline and basic volcanic rocks derived from the PKB and Andrusov Ridge (Mišik *et al*., 1991*a*; Birkenmajer and Wieser, 1992). Northward the upper part of this formation interfingers with the turbidites of the Szczawnica Formation (Paleocene–Lower Eocene, see Birkenmajer and Oszczypko, 1989). In the Jarmuta and Szczawnica formations, characterized by sedimentation rates of 20–50 m/My (Fig. 8, Grajcarek Unit), occur significant amounts of SE-supplied chrome spinels (Oszczypko and Salata, 2005), reflecting erosion of oceanic crust in the area of the Czorsztyn Ridge — Inner Carpathian collision zone. This oceanic crust presumably represented obducted Pieniny Basin (Vahic Ocean) floor (see Fig. 6).

Significantly, this collisional event, which was paralleled by major orogenic activity in the Inner Carpathian and Austro-Alpine domains (Książkiewicz, 1977; Sandulescu, 1988;

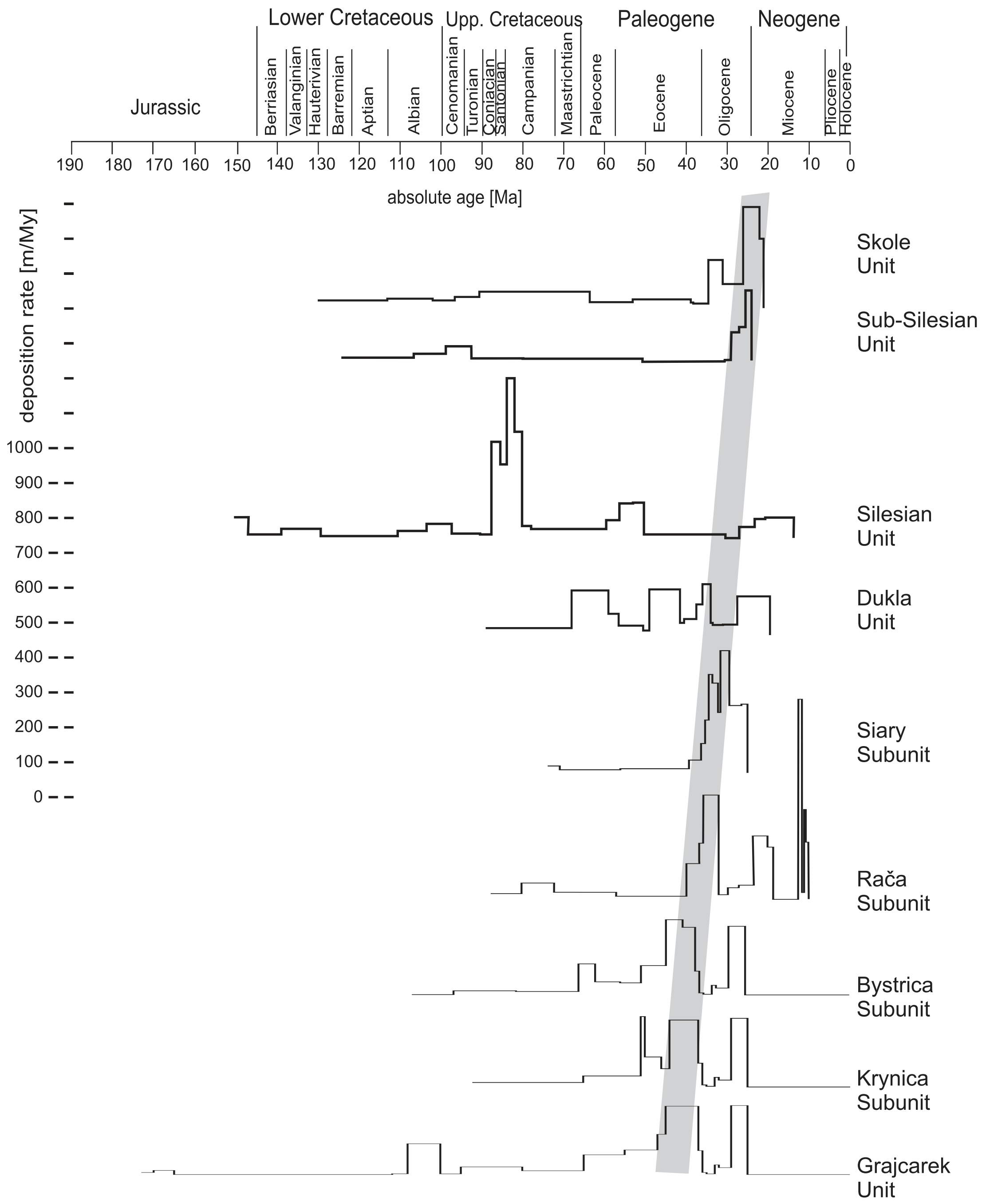


Fig. 8. Diagram of depositional rates versus time for selected synthetic profiles from the Polish Outer Carpathians (after Poprawa *et al.*, 2002; Oszczypko *et al.*, 2003)

Grey shaded bar indicates diachronous syn-orogenic deposition

Poprawa *et al.*, 2002; Schmid *et al.*, 2004*a*, *b*; Dèzes *et al.*, 2004; Picha *et al.*, 2005), was accompanied by the build-up of intraplate compressional stresses in the Outer Carpathian domain, as well as in its distal foreland.

During the Maastrichtian–Paleocene, coarse material derived from the Silesian Ridge was shed southward into the northern parts of the Magura Basin where the Solan (Švábenická *et al.*, 1997) and Jaworzynka beds and the Mutne Sandstones (Oszczypko *et al.*, 2005*d*) were deposited at sedimentation rates of 60–100 m/My (Fig. 8, Bystrica Subunit). Compressional uplift of the Silesian Ridge probably involved the inversion of pre-existing extensional structures (Roure *et al.*, 1993; Roca *et al.*, 1995; Krzywiec, 2002). Similarly, the sudden rise of the Sub-Silesian (Węglówka) High (Figs. 6 and 7, Sub-Silesian and Skole units), which was dominated by the deposition of pelagic variegated marls during Santonian–Eocene times, could also be related to the uplift of the Silesian Ridge.

In the distal foreland of the Outer Carpathian domain, the main inversion phase of Polish Trough is also dated as Maastrichtian/Paleocene (Ziegler, 1990; Dadlez *et al.*, 1995; Kutek, 2002; Krzywiec, 2006). This, in combination with inversion movements in the Outer Carpathian basins, reflects the build-up of major intraplate compressional stresses in the foreland of the IWCW. This must be attributed to strong mechanical coupling of the IWCW with Outer Carpathian lithosphere during their initial collisional phase (Ziegler *et al.*, 1998, 2002).

LATE PALEOCENE (58 MA) TO MIDDLE EOCENE SUBSIDENCE (37 MA)

Towards the end of the Paleocene compressional foreland stresses relaxed, as indicated by the termination of inversion movements in the Polish Trough (Ziegler, 1990; Dadlez *et al.*, 1995; Kutek, 2001) and the Outer Carpathian domain. This may be attributed to mechanical decoupling of the South-Magura Cordillera from its foreland lithosphere in response to sediments entering the subduction zone (Ziegler *et al.*, 1998, 2002).

During the Late Paleocene, the IWCW had advanced northwards to the southern margin of the Magura Basin. Its load caused flexural subsidence of the PKB and deposition of deep-water facies (see Książkiewicz, 1977; Oszczypko, 2004). At the end of Paleocene, the Outer Carpathian basins began to subside regionally whilst sea levels rose (Poprawa *et al.* 2002). During the Eocene, wide connections were established between the Outer Carpathian basins and the World Ocean (Golonka *et al.*, 2000). This, in combination with regional subsidence, resulted in the unification of facies developments in the Outer Carpathian basins, including the position of the CCD level and low sedimentation rates. During Early to Middle Eocene times, this general trend persisted in the northern Skole, Sub-Silesian, Silesian and Dukla basins, as well as in the northern parts of the Magura Basin.

During the Early Eocene to Oligocene, the South-Magura Cordillera, bounding the Magura Basin to the south-east, supplied the southern and central parts of this basin with clastics. These are rich in exotic rock fragments, consisting mainly of granitoids, gneisses, phyllites and quartzites, with basic volcanic rocks and Mesozoic carbonates playing a subordinate role (Mišik *et al.* 1991*b*; Oszczypko, 1992). The carbonate fragments were derived from Jurassic-Lower Cretaceous deep-water series, as well as from shallow-water Triassic (Anisian), Kimmeridgian-upper Tithonian, Lower Cretaceous (Urgonian), Upper Cretaceous, Paleocene and lower Lutetian series (Oszczypko, 1975; Mišik *et al.* 1991*b*). As these exotic rocks differ substantially from those of the Paleocene/Lower Eocene Jarmuta and Proč formations, they may have possibly been derived from the substrate of the Magura that was exhumed during the Early/Middle Eocene, as advocated by Mišik *et al.* (1991*a*). Alternatively, and perhaps more likely, and as subduction of the Magura basement had already commenced during the Paleocene, elements of IWCW, including the ALCAPA and Tisza terranes (Inner Carpathian/Inner Dacide terrains, Oszczypko *et al.*, 2003; Oszczypko, 2004), were the source of these exotic rock fragments (Plašienka, 2000).

The migrating load of the IWCW, and its associated Magura and PKB accretionary wedge, caused flexural subsidence of the Outer Carpathian basins and a northward shift of depocentres. As a result, narrow and long submarine fans developed. The northern deepest part of the basin, largely located below the CCD, was dominated by basinal turbidites and hemipelagites. Sedimentation rates varied between 6–18 m/My on the abyssal plain and 103–160 m/My and 180–350 m/My in the outer and middle fan-lobe systems, respectively (Fig. 8; Oszczypko, 1999). During Late Eocene and Oligocene, the subsidence axis of the Magura Basin shifted northward into the area of the Rača and Siary subunits (Fig. 8).

SYNOROGENIC LATE EOCENE (37 MA)–EARLY MIOCENE (18 MA) CLOSING OF THE OUTER CARPATHIAN BASINS

The Late Eocene was a time of major changes in the Outer Carpathian Basin system (Table 1) during which a uniform pelagic depositional regime was established except for the northern parts of the Magura Basin in which thick-bedded turbidites accumulated (Fig. 8, Rača and Siary subunits). During the Late Eocene–Early Oligocene, the remnant Outer Carpathian Basin was transformed into a flexural foreland basin (Oszczypko, 1999). Deposition of deep-water basinal turbidites gave way to pelagic *Globigerina* Marls that were followed by the Early Oligocene organic-rich Menilite Shale, deposited under anoxic bottom water conditions (Bessereau *et al.*, 1996). The Menilite Shales, that attain thicknesses of 100 m in the Silesian and 900 m in the Dukla Basin, grade diachronously upwards into the Late Oligocene–Early Miocene Krosno Flysch. The combined Menilite-Krosno Formation exhibits major thickness variations. It reaches 2500 m in the Dukla Basin, 4000 m in the Silesian Basin and 2100 m in the Skole Basin, but is missing on the intervening cordilleras that formed submarine highs during the Menilite Shale deposition but later on acted as sediment sources (Bessereau *et al.*, 1996). According to Senkovsky *et al.* (2004), deposition of the black Menilite Shales was related to upwelling currents along the flanks of the Silesian Ridge. This resulted in the development of an oxygene minimum layer at outer shelf- basin slope depths above the submarine fans. At the Early to Late Oligocene transition, the drastic (175 m) glacio-eustatic sea level fall (Haq *et al.*, 1988) may have contributed to the onset of the Krosno Flysch deposition (Table 1).

Oligocene subsidence and structuring of the Outer Carpathian basins was paralleled by the development of an accretionary wedge in the southern part of the Magura Basin (Krynica Zone) in response to subduction of the foreland lithosphere beneath the Pieniny Klippen Belt/Central Carpathian Block (Oszczypko, 1992, 1999).

In the Magura Basin, Oligocene series, characterized by high sedimentation rates (300–350 m/Ma), are developed in three interfingering lithofacies. These are the Malcov and Poprad sandstones of the southern and central parts of the basin that were derived from a SE source, and the glauconitic Wątkowa Sandstones of the northern part of the basin that were derived from the Silesian Ridge to the NE (Table 1).

Following its Late Oligocene folding, the Magura Nappe was thrusted northwards (Fig. 5) onto the terminal Krosno

Flysch Basin (Oszczypko-Clowes and Oszczypko, 2004) with its front reaching the southern parts of the Silesian Basin during the Burdigalian. In the remnant Outer Carpathian basins this was accompanied by a last, Late Oligocene-Early Miocene minor subsidence event, which can be attributed to thrust-loading of the foreland lithosphere by the advancing orogenic wedge (Poprawa *et al.*, 2002). This was paralleled by a progressive northward migration of the depocentre axis, accompanied by an increase in sedimentation rates, during the Rupelian in the northern part of Magura Basin to the Early Miocene in the SE part of Skole Basin (Fig. 8). The restored width of the early Burdigalian Outer Carpathian Basin probably reached at least 150 km. During the early Burdigalian sea level high-stand, the Magura piggyback basin developed that was linked via Orava by a seaway to the Vienna Basin (Fig. 9, see Oszczypko *et al.*, 1999; Oszczypko-Clowes, 2001; Oszczypko and Oszczypko-Clowes, 2002). During Ottnangian, the Krosno Flysch Basin shifted towards NE (Zdanice Unit, Boryslav-Pokuttya and Marginal Fold Units) and underwent desiccation (evaporites of the Vorotyscha Formation in the Ukraine and Salt Formation in Romania).

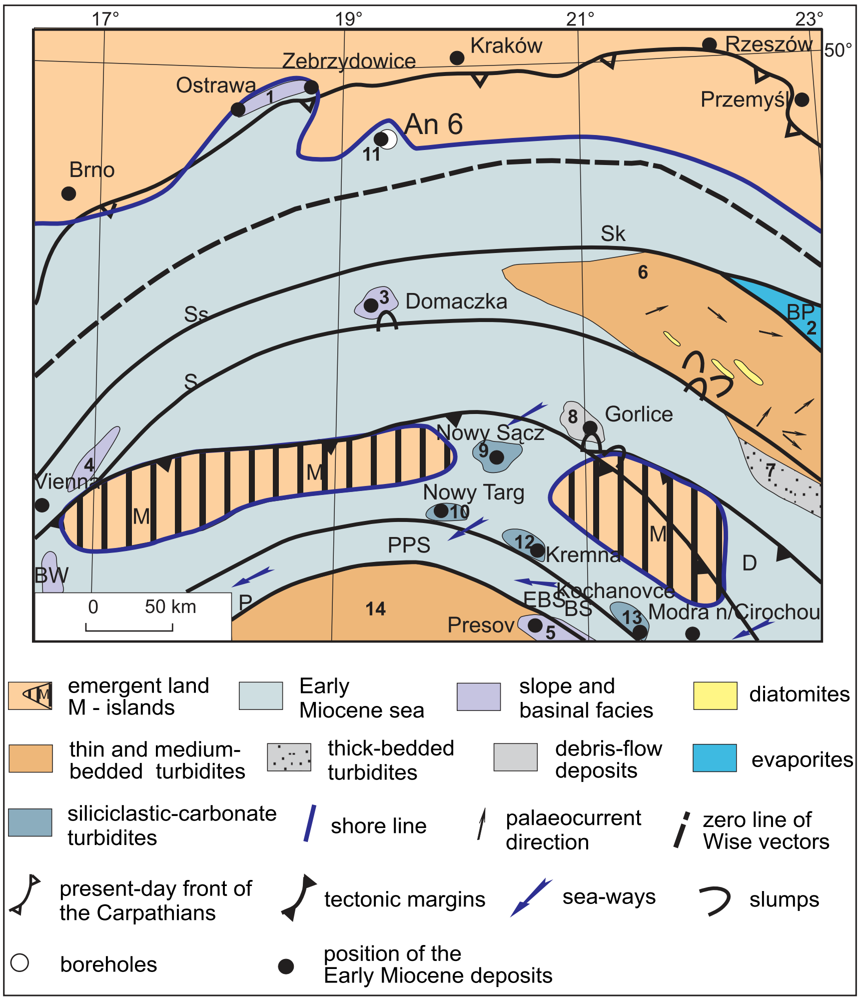


Fig. 9. Early Miocene (NN2–NN3?) palaeogeography of the Northern Carpathian Basin (after Kovač *et al.*, 1998, Oszczypko and Oszczypko-Clowes, 2003, supplemented)

Lithostratigraphic subdivisions: 1 — Zebrzydowice Fm., 2 — Vorotyscha Fm., 3 — Domaczka Fm., 4 — Sakvice Fm., 5 — Prešov Fm., 6 — shaly facies of Krosno Beds, 7 — sandstone facies of Krosno Beds, 8 — Gorlice Beds, 9 — Zawada Fm., 10 — Waksmund Fm., 11 — Andrychów Fm., 12 — Kremna Fm., 13 — Kochanovce Fm., 14 — Chochołów and Ostrysz beds; BW — Vienna Basin, PPS — Pieniny Klippen Belt, D — Dukla Unit, Ss — Skole Basin, BP — Borislav-Pokuttya Basin, EBS — East Slovakian Basin, S — Silesian Basin, Sk — Skok Basin

EARLY MIOCENE COMPRESSION (17 MA)

The Early Miocene folding, thrusting and inversion of the Outer Carpathians was traditionally referred to the Early Miocene Savian and Styrian orogenic phases. However, taking into account that the youngest flysch deposits of the marginal Zdanice, Skole-Skiba and Boryslav-Pokuttya units, as well as of the Magura Nappe belong to the Ottnangian-late early Burdigalian NN3 and NN4 calcareous nannoplankton zones, the onset of this deformation phase cannot be older than about 17 Ma. This corresponds with the intra-Burdigalian (Ottnangian) compressional tectonic event, referred to as the early Styrian phase (see Oszczypko, 1997, 1998; Oszczypko *et al.*, 2005*a*).

The Intra-Burdigalian folding and the uplift of the Outer Carpathians can be related to the north-eastward translation of ALCAPA and Tisza-Dacia microplates in response to roll-back of the Carpathian subduction slab and increasing collisional coupling of the Adriatic and European plate in the Alpine domain (Fodor *et al.*, 1999; Schmid *et al.*, 2004*b*, 2005). In the Outer Carpathians, this was accompanied by north- and north-east-directed nappe transport and the development of the peripheral flexural Carpathian Foredeep along the advancing orogenic front (Oszczypko, 1998; Kovač *et al.*, 1998). At the turn of the Ottnangian, after the first thrusting stage, the front of the Outer Carpathians was located about 50 km south of its present-day position (Oszczypko and Tomaś, 1985; Oszczypko, 1997; Oszczypko and Oszczypko-Clowes, 2003). The load of the Carpathian accretionary wedge caused flexural subsidence of the foreland platform, involving tensional reactivation of pre-existing crustal discontinuities (Figs. 3 and 4), and the development of a moat-like foredeep that was filled by coarse clastics (Oszczypko, 1998). 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 that progressively filled in 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 sedimentary cover. These overthrusts are known as the "Old Styrian Nappes" (Jurkova, 1971) or as the Sucha and Zamarski formations flysch olistoplaque (Fig. 3, see Buła and Jura, 1983; Oszczypko and Tomaś, 1985; Moryc, 1989; Oszczypko, 1998). In the Cieszyn area, these thrusts reached before the early Badenian more or less their present-day position (Oszczypko and Oszczypko-Clowes, 2003).

POLISH CARPATHIAN FOREDEEP

The Polish Carpathian Foredeep (PCF) developed during the Early and Middle Miocene as a peripheral flexural foreland basin in front of the advancing Carpathian front (Oszczypko, 1982, 1997, 1998; Oszczypko *et al.*, 2005*a*). This basin can be subdivided into an inner and an outer part. The inner foredeep is located south of the Carpathian frontal thrust and contains up to 1500 m of Lower to Middle Miocene autochthonous deposits that are overridden by thrust sheets (Fig. 3). The outer foredeep is filled by Middle Miocene (Badenian and Sarmatian) marine deposits, which range in thickness from a few hundred metres in its northern, marginal parts to as much as 3 500 m in its southeastern parts adjacent to the Carpathian thrust front (Figs. 2 and 4, cross-section G–H).

Whereas Lower Miocene strata are mainly terrestrial in origin, Badenian and Sarmatian deposits are developed in marine facies. Traditionally, the Central Paratethys stage boundary between the Badenian and Sarmatian is dated as 13.5 Ma and is located at the top of Zone NN5 (see Rögl, 1996, 1999). In this paper we place this boundary at the top of Zone NN6, dated as 11.8 Ma (Marunteanu, 1999; Oszczypko *et al.*, 2005*a*).

As the Miocene deposits of the Polish Carpathian Foredeep (PCF) are poorly exposed, its structure and stratigraphy are based on boreholes (about 2000) and seismic sections.

STAGES OF STRUCTURAL AND DEPOSITIONAL DEVELOPMENT

In the western part of the Polish Outer Carpathians (Cieszyn-Bielsko area), several boreholes have penetrated autochthonous Late Oligocene to Early Miocene deposits beneath the Carpathian Nappes (Fig. 3). The basal portion of this sequence is composed of the fan-delta pebbly mudstones and conglomerates of the Zawoja Formation (Oligocene-Egerian). In the Andrychów 6 borehole these conglomerates pass upwards into dark marine mudstones of the Egerian age (nannoplankton NN1 Zone, see Oszczypko and Oszczypko-Clowes, 2003). Northward, these shelf mudstones are followed by the marine Eggenburgian (Ottnangian) grey-greenish marine clays of the Zebrzydowice Fm. (Garecka *et al.* 1996), that reflect progressive flooding of the foreland basin Jurkova *et al.*,1983). At the same times this marine basin was probably linked with the Outer Carpathian residual flysch basin (Figs. 9 and 10).

The Egerian-Ottnangian marine sedimentation period (Oszczypko and Oszczypko-Clowes, 2003) was followed by

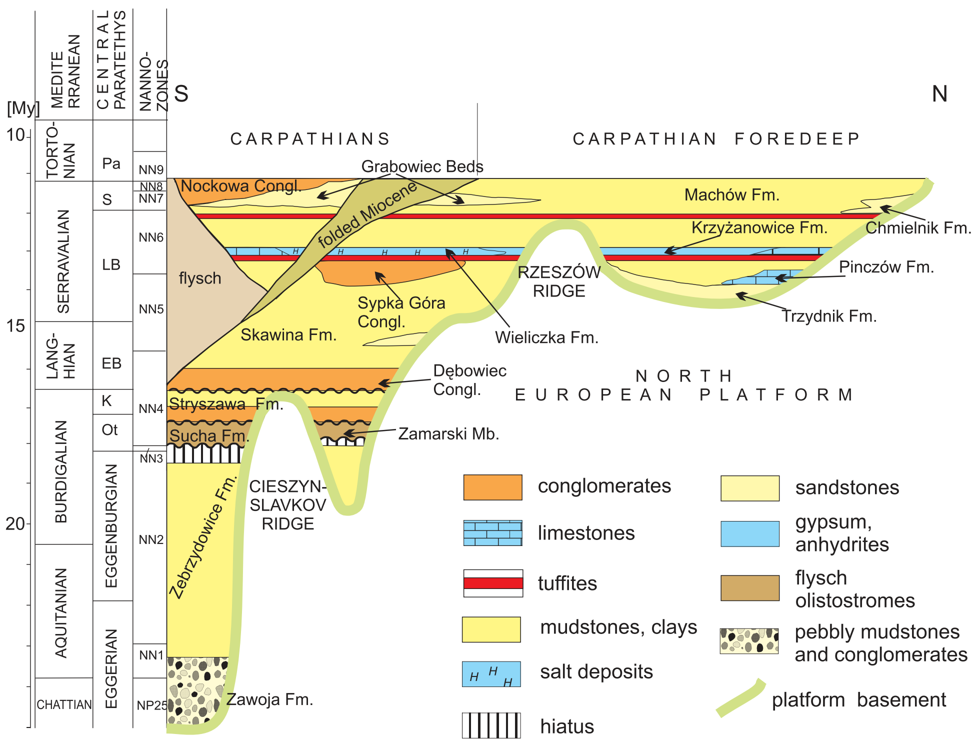


Fig. 10. Lithostratigraphic model of the Miocene deposits in the Polish Carpathian Foredeep against the Mediterranean and Central Paratethys chronostratigraphy (modified after Oszczypko, 1999)

Ot — Ottnangian, K — Karpatian, EB — early Badenian, LB — late Badenian, S — Sarmatian, Pa — Pannonian an intra-Burdigalian (late Ottnangian) uplift, folding and thrusting of the Outer Carpathians onto the foreland platform. This was accompanied by the development of large-scale slides (olistoplaques and gravitational nappes) along the frontal parts of the Sub-Silesian Nappe (Fig. 10). By late Ottnangian times the deformation front of the orogen was located about 50 km south of its present-day position in the eastern parts of the PCF whilst in the Cieszyn area gravitational nappes had more or less reached the present-day position of the Carpathian front (Fig. 11).

In the Zawoja 1 borehole (Fig. 3), the Oligocene-Egerian fan-delta conglomerates are overlapped by an up to 370 m thick olistoplaque (Sucha Fm., Ślączka, 1977; Buła and Jura, 1983; Moryc, 1989; Oszczypko, 1998) that consists of Lower Cretaceous to Paleocene flysch, derived from the Sub-Silesian and Silesian units that is contained in an Early Miocene matrix yielding Ottnangian-Karpatian nannoplankton (NN 4) (Garecka *et al.*, 1996).

In the westernmost part of the PCF (Cieszyn area), the Early Miocene Zebrzydowice Formation is overlain by the 25–150 m thick flysch olistoplaque of the Zamarski Formation (Buła and Jura, 1983) that extends over at least 50 km^2 and consists of elements of the Sub-Silesian Unit. This olistoplaque is also known as the ”Old Styrian overthrust” from the marginal part of the Moravo-Silesian Carpathians (Jurkova, 1971).

Emplacement of the Zamarski olistoplaque was followed by the Karpatian period of intense subsidence of the inner foredeep that probably affected also the frontal part of the Carpathian Nappes. This foredeep was filled with the coarse alluvial fan deposits of the Stryszawa Formation that were derived from both the Carpathians and emerging parts of the foreland platform (Oszczypko, 1997, 1998; Oszczypko *et al.*, 2005*a*). The Stryszawa Formation, attains thicknesses of up to 566 m (Figs. 3 and 10; Ślączka, 1977; Moryc, 1989; Oszczypko, 1998; Oszczypko *et al.*, 2005*a*). Its basal part consists of conglomerates containing components derived from the Carpathian Flysch as well as the Palaeozoic basement of the foreland. These conglomerates grade upward into sandstones, the carbonate and gypsum matrix of which yielded Ottnangian-Karpatian nannoplankton (NN 4, Garecka *et al.*, 1996), as well as reworked Lower Cretaceous to Early Miocene foraminifera. The youngest recycled microfauna found in the Stryszawa Formation are of Eggenburgian-Ottnangian age and occur also in the youngest strata of the Outer Carpathians (Oszczypko, 1997).

The Early Miocene inner PCF Basin was at least 40 km wide and extended to the south-east into the Sambir Basin of the Ukraine (Oszczypko *et al.*, 2005*a*). Deposition of the Stryszawa Formation was followed by late Karpatian erosion, which was caused by the uplift of the peripheral Cieszyn-Slavkov Palaeoridge (Oszczypko and Tomaś, 1985; Oszczypko, 1997; Oszczypko and Lucińska-Anczkiewicz, 2000). In Southern Moravia, this erosional period can be correlated with the discordance below the terminal Karpatian strata (Jiřiček, 1995). Erosion along the northern flank of Cieszyn-Slavkov Palaeoridge was coupled with the development of normal fault bounded W–E and NW–SE trending grabens (e.g. Bludowice-Skoczów palaeovalley; Oszczypko and Lucińska-Anczkiewicz, 2000). During the Badenian the axes of the extensional grabens migrated towards the NE.

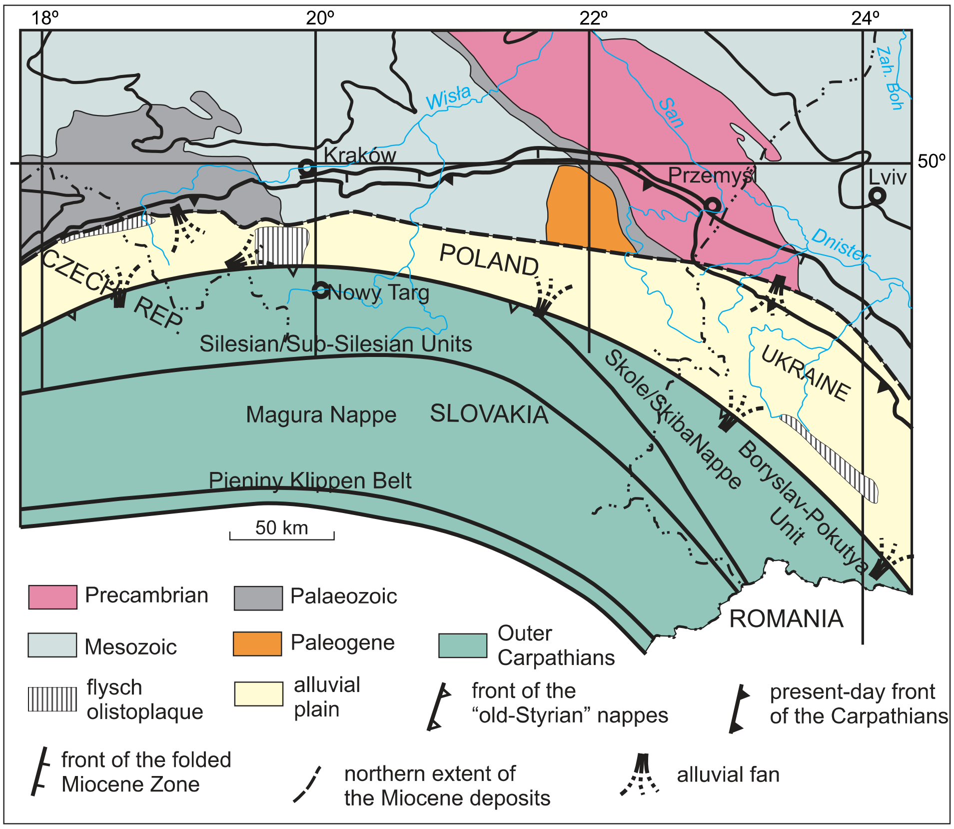


Fig. 11. Karpatian palinspastic palaeogeography of the Polish Carpathian Foredeep (after Oszczypko *et al.*, 2005*a*, simplified)

During the late Karpatian-early Badenian, these subsiding grabens were successively filled in with slope deposits (blocks of Carboniferous rocks), and the near-shore Dębowiec Conglomerate. The latter is 40–100 m thick and is composed of Upper Carboniferous clasts (Ney, 1968; Ney *et al.*, 1974; Oszczypko, 1998; Oszczypko *et al.*, 2005*a*). In the external parts of the PCF the transgressive conglomerates rest directly on the basement of the foreland (Fig. 3). This marine transgression invaded both the foreland and the Carpathians (Fig. 12). The Dębowiec conglomerates grade upwards into the marine clayey-sandy sediments of the Skawina Formation (Fig. 10) that attains a thickness of up to 1000 m in the western, internal parts of the basin, whereas elsewhere it rarely exceeds 30–40 m (Ney *et al.*, 1974). Deposition of the Skawina began in the internal parts of the

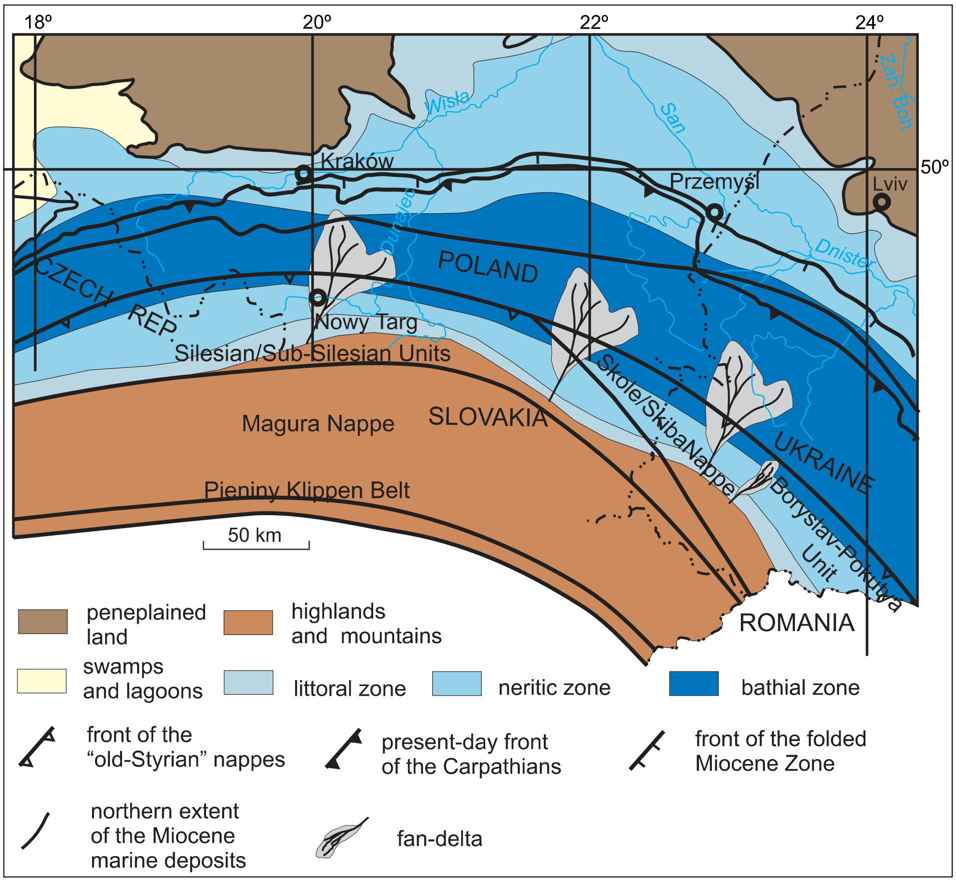


Fig. 12. Langhian (early Badenian) palinspastic palaeogeography of the Polish Carpathian Foredeep (after Oszczypko *et al.*, 2005*a*, simplified)

foredeep with the *Praeorbulina glomerosa* Zone (N8), and in its external parts with the *Orbulina suturalis* Zone (N9 or N10) (Garecka *et al.*, 1996; Oszczypko, 1997, 1998). According to nannoplankton studies, the Skawina Formation spans the NN 5 Zone and extends into the NN 6 Zone (Andreyeva-Grigorovich *et al.*, 1997, 2003). A tuffite from the uppermost part of the Skawina Formation has yielded a radiometric age of 12.5±0.9 Ma (Bukowski and Szaran, 1997).

The late Badenian drop in sea level and climatic cooling initiated a salinity crisis in the Carpathian foreland basin (see Oszczypko, 1998; Andreyeva-Grigorovich *et al.*, 2003; Bąbel, 2004). These Badenian evaporites belong to the lower part of the NN6 Zone (Peryt, 1997; Peryt *et al.*, 1998) and consist of rock salt, claystones, anhydrites, gypsum and marls. Between Wieliczka and Tarnów (Fig. 2) salts attain a thickness of 70–110 m (Garlicki, 1968; Bukowski and Szaran, 1997) decreasing towards the east to a few dozen metres, whereas the thickness of gypsum and anhydrites commonly varies between 10 and 30 m (Bąbel, 2004, 2005).

The shallow (stable shelf) parts of the basin were dominated by sulphate facies, whereas its deeper parts, located along the Carpathian front, were occupied by chloride-sulphate facies (Garlicki, 1979; Kasprzyk, 1993, 1999, 2005; Bąbel, 1999, 2004). According to Bąbel (2004), water depths in "the gypsum sub-basin were very shallow, zero to several metres" whilst the "halite sub-basin was estimated as less than 30–40 m deep". After evaporite deposition, parts of the outer foredeep were uplifted and subjected to erosion (e.g. Rzeszów Palaeoridge).

The evaporites are overlain a sandy-silty series that are attributed to the late Badenian and Sarmatian (NN6/7 to NN 8/9 Zone, see Gaździcka, 1994; Andreyeva-Grigorovich *et al.*, 2003). These deposits range in thickness between a few hundred metres in the northern parts of the basin to up to 3000 metres in the south see Krzywiec (1997, 2001). In the Kraków-Bochnia region, tuffite intercalations just above the Badenian evaporites yielded a radiometric age of ±12 Ma (Van Couvering *et al.*, 1981). In the northern, marginal part of the PCF, early Sarmatian littoral carbonates and clastic deposits are well preserved (Oszczypko *et al.*, 2005*a*).

Along the southern margin of the PCF, development of the folded Miocene units (Figs. 1 and 2) was strongly influenced by the configuration of the frontal Carpathian thrust fault and the depth to basement (Oszczypko and Tomaś, 1985; Oszczypko and Ślączka, 1985). The Stebnik (Sambir) Unit, that occurs along the front and beneath of the Skole Nappe SE of Przemyśl near the Polish-Ukrainian border (Ney, 1968), involves Early and Middle Miocene as young as Sarmatian (Andreyeva-Grigorovich *et al.*, 1997; Garecka and Olszewska, 1997). Between Przemyśl and Kraków, an up to 10 km wide zone of folded Badenian and Sarmatian strata occurs along the Carpathian frontal thrust (Zgłobice Unit; Kotlarczyk, 1985; Krzywiec *et al.*, 2004) that can be regarded as a prolongation of the Sambir Unit (Oszczypko *et al.*, 2005). Badenian and Sarmatian strata are also preserved as erosional outliers in the Polish Outer Carpathians. The southernmost occurrence of the late Badenian/Sarmatian marine sediments is known from the Nowy Sącz Basin (Fig. 2; Oszczypko *et al.*, 1992). In the Cieszyn-Wadowice area, early and late Badenian deposits are involved in the sub-Silesian and Silesian units. The Zawoja 1 borehole penetrated parautochthonous, supposedly lower Badenian sediments beneath the Magura Nappe (Fig. 3).

During the subsequent Intra-Badenian compressional event, major shortening occurred in the Carpathian Nappes (see Oszczypko, 1997, 1998; Kovač *et al.*, 1998). This is documented by a displacement of at least 12 km of the Magura and Fore-Magura units relative to the Silesian Unit, as well as of the Silesian Unit relative to the Sub-Silesian Unit, and a tectonic duplication of the Sub-Silesian Unit. Finally, the present-day position of the Carpathian Nappes was reached in post-Sarmatian times (Wójcik and Jugowiec, 1998; Oszczypko, 1998).

SUBSIDENCE DEVELOPMENT

Development of foredeep basins is generally regarded as resulting from flexural deformation of the lithosphere in response to its loading by a growing orogenic wedge (Price, 1973; Beaumont, 1981). The Carpathian Foredeep Basin is a typical slab-loaded fore-arc foreland basin (Ziegler *et al.*, 2002) that developed in front of the advancing Carpathians. The arcuate shape of the Carpathians and their foredeep (Fig. 1) was primarily controlled by the configuration of the margin of the European foreland plate (Oszczypko *et al.*, 2005*a*). The present width of the Carpathian outer foredeep varies between 30–40 km in its western segment, 10 km in the Kraków area, up to 90 km on the Rzeszów meridian, and around 50 km along the Polish-Ukrainian state boundary (Fig. 2). Significant narrowing of the foredeep in the Kraków vicinity is related to NW–SE trending foreland structures (Ney, 1968), whereas to the east erosional remnants of marine Miocene deposits can be found beyond the present-day margin of the foredeep, suggesting that its depositional margin was located further to the north.

Towards the east, prominent changes occur in configuration of foredeep, particularly concerning the change of its SW–NE trend in Czechia and Western Poland to NW–SE in Eastern Poland and the Ukraine (Fig. 1). In the area of the Polish-Ukrainian border, the Carpathian Foredeep is superimposed on the NW–SE trending Teisseyre-Tornquist Zone, a major crustal boundary between the East European Craton and the West- and Central European Palaeozoic Platform (Oszczypko *et al.*, 2005*a*).

On burial diagrams (Fig. 13) three periods of intense foreland subsidence are evident, namely dur-

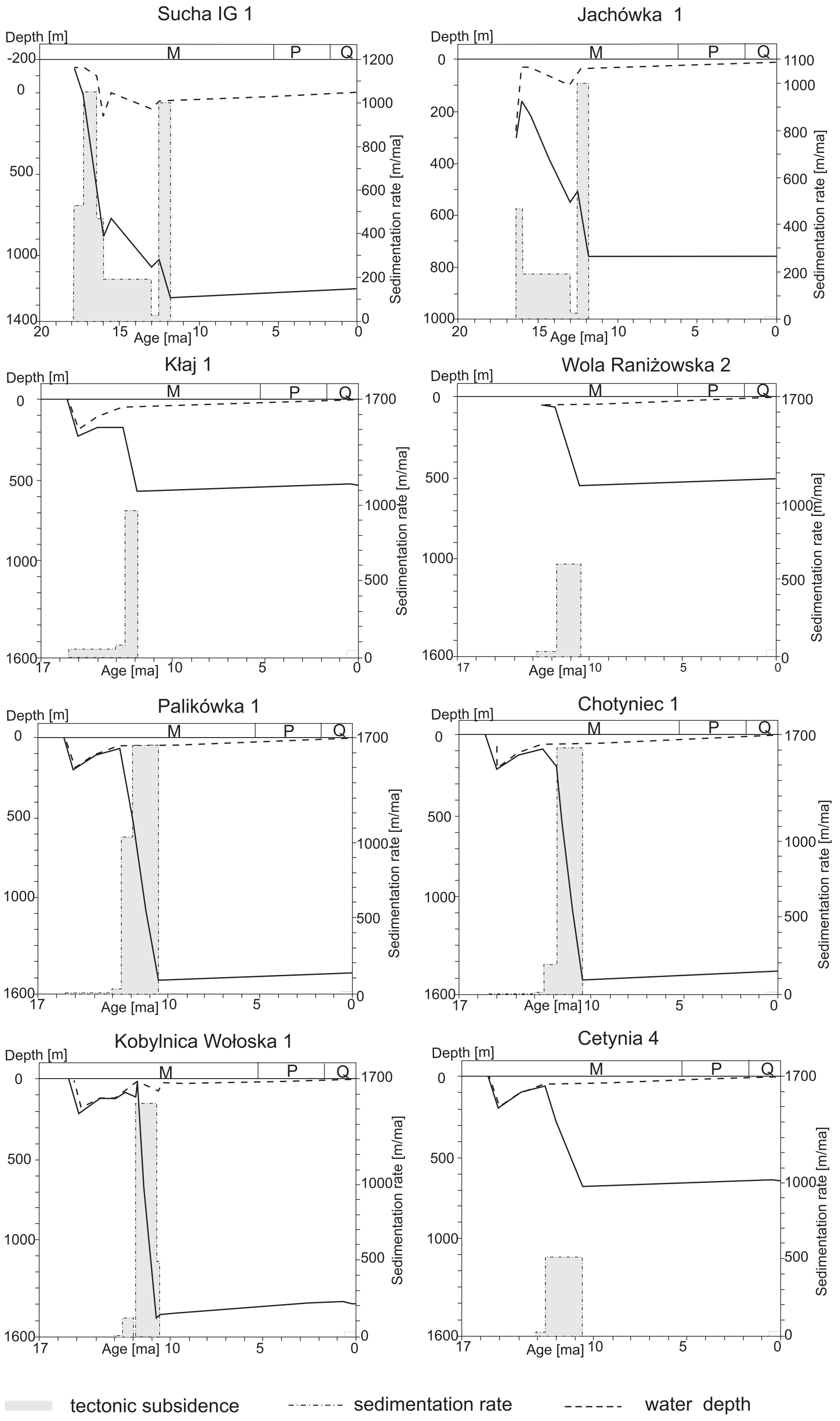


Fig. 13. Backstripped burial diagrams for selected boreholes from the middle and eastern part of Polish Carpathian Foredeep (after Oszczypko, 1999)

M — Miocene, P — Pliocene, Q — Quaternary; for location of boreholes see Figure 2 ing the Early Miocene, early Badenian, and late Badenian to Sarmatian times. During the initial stage of inner foredeep development in the Czech and Polish segments (Ottnangian-Karpatian), subsidence rates (1000–1400 m/My) and sedimentation rates were in balance, as evidence by the accumulation of terrestrial and shallow marine series (Vass and Cech, 1983; Meulenkamp *et al.*, 1996; Oszczypko, 1997, 1998).

During the late Karpatian/early Badenian, SW–NE and NW–SE trending troughs developed in the front of the "Old Styrian" thrust. Starting with the early Badenian (Langhian) marine transgression, subsidence rates exceeded sedimentation rates, resulting in the establishment of marine environments that prevailed during Badenian and Sarmatian times. The early Badenian sedimentation rates reached in the axial part of the basin 250 to 500 m/My in Moravia (Vass and Cech, 1983; Meulenkamp *et al.*, 1996) and 200 m/My in Poland (Oszczypko, 1997, 1998), whilst its northern stable shelf subsided very slowly with sedimentation rates oscillating between a few dozens and 50 m/My (Fig. 13). During the early Badenian, water depths varied in this basin between upper bathyal in its axial parts and neritic to littoral in its northern and southern (Carpathian) marginal parts.

Since the Serravalian (*ca*. 15 Ma), a gradual decrease in water depths can be observed in the Carpathian foreland basin that coincided with a eustatic gradual fall in sea level. This resulted in a partial isolation of the basin and the onset of the Badenian (*ca*. 13.0–12.5 Ma) salinity crisis. For this period, subsidence curves reveal a tendency of progressive uplift of the basin whilst sedimentation rates varied between a few dozen metres/My in areas of sulphate facies and up to some 50 m/My in areas of chloride facies. These are very rough approximations, as episodes of chemical precipitation can be of very short duration (25–35 Ky, see Garlicki, 1968; Petrichenko *et al.*, 1997) with sedimentation rates being one order higher, up to 500 m/My. At the end of evaporite deposition (*ca*. 12.5 Ma), when the basin was the shallowest (Fig. 13, see also Kasprzyk, 1993; Gonera, 1994; Czepiec, 1996), tectonic uplift of the foreland resulted in the development of a regional unconformity, and the erosion of up to 50–100 m of evaporitic and sub-evaporitic deposits (Rzeszów Palaeoridge, see Komorowska-Błaszczyńska, 1965). This erosional surface has also been reported from the Ukrainian part of the foredeep (see Peryt and Peryt, 1994; Panow and Plotnikow, 1996; Andreyeva-Grigorovich *et al.*, 1997).

A last, but very intense phase of foredeep subsidence commenced around 12.5 Ma during the late Badenian and ended around 10.5 Ma during the Sarmatian (Fig. 13). This subsidence phase started with a transgression during which outer and inner neritic conditions were established in the inner and outer parts of the foredeep, respectively (Gonera, 1994; Czepiec, 1996).

In the eastern segment of the foredeep, the basin axis shifted during the late Badenian 15 km northward with respect to its early Badenian position (migration rate 3.75 m/My), reaching approximately the present-day position of the Carpathians deformation front. The highest subsidence rate of up to 2000 m/My was determined in the SE part of PCF in the Przemyśl area (Oszczypko, 1997, 1998). Towards the NW, in the Rzeszów area, subsidence rates decreased, whereas sedimentation rates oscillated around 1500 m/My (Fig. 13). Slightly lower subsidence and sedimentation rates are observed in the Tarnów and Bochnia areas. In areas that were effected by post-evaporite erosion (e.g. Rzeszów Island), late Badenian deposits transgressed over the Precambrian and Palaeozoic basement. Towards the NE margin of the PCF, late Badenian subsidence and sedimentation rates decreased to 100–200 m/My. Although the PCF basin continued to subside during the Badenian-Sarmatian transition, its depocentre shifted 40–50 km towards the NE whilst is axis rotated clock-wise by up to 20° (Oszczypko and Żytko, 1987; Oszczypko *et al.*, 2005). The zone of maximum Sarmatian subsidence coincides with the so-called "Wielkie Oczy Graben". Total subsidence in this zone varied between 1500 m in its NE part and up to 3000 m in its SE part (Fig. 13), whereas tectonic subsidence reached a maximum of 1500 m and decreased to a few hundred metres towards the northern margin of PCF. The high Sarmatian subsidence rates were compensated by high sedimentation rates, reaching 1700–2400 m/My. During the Sarmatian water depths oscillated between outer neritic and littoral (Czepiec, 1996).

The late Badenian-early Sarmatian pulse of accelerated subsidence of the Polish-Ukrainian foredeep, which followed the intra-Badenian pulse of major shortening in the Carpathian nappe stack, can be attributed to increased slab loading, preceding slab detachment around 10.5 Ma when uplift and erosion of the basin commenced (Fig. 13). As such, this subsidence pulse represented the final stage in the evolution of the Polish-Ukrainian Carpathian foreland basin (Bubniak.*et al*., 2001).

Although development of foredeep basins is basically controlled by deflection of the underthrusted foreland lithosphere under the load of orogenic belts, the loads exerted by water and sediments are additional important driving mechanisms responsible for creation of accommodation space in such basins (Allen and Allen, 1992; Oszczypko *et al.*, 2005*a*).

The observed clear relationship between periods of Carpathian Nappe emplacement and subsidence of PCF suggests that tectonic loading of the foreland was a significant driving force of subsidence. Each advance of the Carpathian deformation front initiated a new subsidence phase of the foreland basin. During Early–Middle Miocene times, the loading effect of the thickening Carpathian orogenic wedge on the foreland plate increased and caused progressively increasing total subsidence of the foredeep. However, Royden and Karner (1984), Royden (1988, 1993) suggested that this supracrustal load was inadequate to explain the observed deflection of the Carpathian foreland plate and postulated the presence of an additional subsurface "load" exerted by the subducted lithospheric slab (Royden and Burchfiel, 1989).

Flexural modelling studies of the Polish and Ukrainian Carpathians (Royden and Burchfiel, 1989; Krzywiec and Jochym, 1996, 1997; Zoetemeijer *et al.*, 1999) suggest that such deep subsurface loads played an important role in the development of the observed present-day flexure of the foreland lithospheric plate. According to other authors, this extra "load" should be taken into account only during the early collisional history (see discussion in Miall, 1995). However, it must be stressed that the temporal interaction between thrust-loaded and slab pull-induced subsidence was a direct consequence of southward subduction of the foreland lithosphere and roll back of the subducted slab, at least until its detachment.

Based on the nature and distribution of magmatic activity in the Inner Carpathian domain and tomographic data, slab detachment commenced around 19–17.5 Ma in the westernmost Carpathians and propagated around the Carpathian bend to Romania where it is currently still in progress (Nemčok *et al.*, 1998; Mason *et al.*, 1998; Wortel and Spakman, 2000; Wenzel *et al.*, 2002). Under the weight of the laterally already detached slab, slab-pull forces presumably increased in the adjacent area where the subducted slab was still attached to the foreland lithosphere, only to cease when the slab window propagated through and beyond it (Wortel and Spakman, 2000). Magmatic activity in the Polish-Ukrainian Carpathians suggests that in this area slab detachment became effective in the course of Middle to Late Miocene (14–9 Ma; Nemčok *et al.*, 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 rift-related opening of basins followed by their post-rift subsidence;

— Late Cretaceous–Paleocene basin inversion;

— Paleocene to Middle Eocene subsidence;

— synorogenic Late Eocene–Early Miocene closing of the basins;

— Middle Miocene compression.

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 tectonic subsidence were syn- and post-rift thermal processes, as well as loads related to the nappe emplacement and subduction processes.

3. Similarly to the other orogenic belts, folding and thrusting of the Outer Carpathians progressively advanced towards the foreland. This process was initiated at the end of the Paleocene at the PKB/Magura Basin boundary and was completed during the early Burdigalian in the northern part the Krosno Flysch Basin.

4. During Early to Middle Miocene times the Carpathian Foredeep of Poland developed as a peripheral foreland basin related to the advancing Carpathian front. Subsidence of this basin was controlled by the tectonic loads the orogenic wedge and the subducting lithospheric slab as well as by sediment loading.

5. During the Early–Middle Miocene time the loading effect of the thickening Carpathian accretionary wedge on the foreland plate increased and caused a progressive increase in total subsidence, except during the "middle Badenian" salinity crisis.

6. Miocene convergence of the Carpathian wedge with the foreland caused outward migration of the depocentre of the foredeep and in its distal parts onlap of successively younger deposits onto the foreland.

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