Geological Quarterly, 2009, **53** (1): 3–26

Cenozoic dynamic evolution of the Polish Platform

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Jarosiński M., Porawa P. and Ziegler P. A. (2009) — Cenozoic dynamic evolution of the Polish Platform. Geol. Quart., **53** (1): 3–26. Warszawa.

The Cenozoic tectonic evolution of the Polish Platform reflects repeated changes in loading conditions at the Alpine–Carpathian and Arctic–North Atlantic margins of the European continent. After the Late Cretaceous–Paleocene main phase of the Mid-Polish Basin inversion, a second phase of limited uplift of the Mid-Polish Swell occurred during the Middle–Late Eocene. End Eocene and Early Oligocene subsidence of narrow grabens on the Fore-Sudetic Monocline was coeval with normal faulting in the East Alpine foredeep basin and the development of the Central European rift system. At the same time the Outer Carpathian flysch basins were rearranged, presumably in response to the build-up of compressional stresses at crustal levels, whilst subsidence and erosion patterns changed in the Carpathian Foreland from being dominated by the NW–SE trending Mid-Polish Swell to being controlled by the development of the W–E trending Meta-Carpathian Swell. At the end of the Oligocene the Fore-Sudetic graben system propagated into the area of the Trans-European Suture Zone and the Sudetes and remained active during the Early and Middle Miocene. This was paralleled by intensified subduction activity and thrusting of the Carpathians and the development of their flexural foredeep basin. A short early Sarmatian episode of basement involving transpression along the SW margin of the Mid-Polish Swell correlates with the termination of north-directed nappe transport in the Outer Carpathians. This was followed by eastward migration of the subsidence centre of the Carpathian Foredeep Basin and the gradual termination of tectonic activity in the grabens of the Polish Lowlands. After a period of post-orogenic relaxation the present-day compressional stress regime built up during the Pliocene and Quaternary. Intensified ridge push forces exerted on the Arctic–North Atlantic passive margins contribute to this compressional stress field that is dominated by collision-related stresses reflecting continued indentation of the Adriatic Block. This sequence of events is interpreted in terms of changing tectonic loads in the Carpathians, Alps and at the NW passive margin of Europe. The complex and diachronous interaction of mechanically coupled and uncoupled plates along collision zones probably underlies the temporally varying response of the Carpathian Foreland that in addition was complicated by the heterogeneous structure of its lithosphere. Progressively increasing ridge push on the passive margin played a secondary role in the stress differentiation of the study area.

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Key words: Carpathians, foreland plate, Cenozoic, geodynamics, stress field.

INTRODUCTION

In this contribution we review the post-Paleocene evolution of the Polish Carpathian Foreland and compare it with loading condition of the European lithosphere at the Alpine–Carpathian collision zone and at the Arctic–North Atlantic passive margin. The area addressed includes both the proximal foreland of the orogen, namely the flexural foredeep of the Outer Carpathians, and the distal foreland, which is located beyond the flexural forebulge and extends northward into the Baltic Sea. During the time span considered, the evolution of the Carpathian Foreland was controlled by repeated changes in the stress field that can be related to the final stages of the Alpine–Carpathian collision and the opening of the Arctic–North Atlantic Ocean. In the following we will highlight how the segment of the European lithosphere that is located in front of the Carpathians responded to far-field changes in the tectonic setting of the European continent.

Stresses transmitted across collision zones into forelands can vary in time and space both in intensity and direction. Intense thrusting within an accretionary wedge can coexist with extension in the foreland as well as in the hinterland plate (e.g., Royden, 1993). Main stress generating forces, which play a dominant role in collision-related processes, can be generally divided into convergence driving and convergence resisting ones (Forsyth and Uyeda, 1975; Kusznir, 1991; Bott, 1993). Forces driving convergence comprise far field ridge push forces and pull of a subducting slab. Forces counteracting convergence are first of all friction due to basal drag and deformation, buoyant body forces related to partially subducted continental crust and tensional body forces inherent to the topography and density structure of mountain ranges (Molnar and Lyon-Caen, 1988). In a typical setting a balance between constituent forces evolves from a dominance of far field compression during the initiation and early stage of subduction, through growing slab pull forces due to the volume increase of the subducting oceanic slab, to the gradual building up of forces which resist convergence after continent-to-continent collision has occurred. In the complex interaction between these forces the rheology of the lithosphere and pre-existing tectonic structures play an important role.

The specific problem we address is the collisional interaction between an evolving orogenic wedge and its pro-wedge continental foreland lithosphere. Under such a collisional setting, the stress field in the foreland depends first of all on whether the upper plate orogenic wedge is mechanically coupled with the subducting lower plate or whether they are mechanically decoupled owing to e.g. sediment subduction (Ziegler *et al*., 2002). Under conditions of strong mechanical coupling of the upper and lower plate, compressional stresses can be effectively transmitted from the collision zone into the foreland (Ziegler *et al*., 2002). This requires strong friction between the colliding plates that is common during the final stage of continent/continent collision but can also occur during the initial collision of an orogenic wedge with a passive margin, as seen during the Paleocene collision of the East Alpine orogenic wedge with the European margin (Ziegler *et al*., 1998, 2002; Dèzes *et al*., 2004). Under conditions of mechanical decoupling of the colliding upper and lower plates, tensile stresses may be exerted on the foreland lithosphere by slab pull mechanisms. Related stresses increase with an increasing volume of the subducted oceanic lithosphere (Wortel *et al*., 1991) that may be associated with increasing subduction-related volcanism in the back-arc domain. It should be realized, however, that once subduction of continental foreland lithosphere has commenced, detachment of the subducted oceanic lithospheric slab can occur, triggering related volcanic activity that is only indirectly related to subduction processes (Von Blankenburg and Davies, 1995). Detachment of a subducted slab from the foreland lithosphere causes instant decay of slab pull forces and isostatic rebound of the foreland lithosphere and the orogenic wedge (Ziegler *et al*., 2002).

An additional source of tensional stresses is related to deflection of the foreland lithosphere in response to loads imposed on it by the orogenic wedge and the subducted lithospheric slab, controlling the development of foreland basins (Turcotte and Schubert, 1982; Stockmal and Beaumont, 1987; Royden, 1993; Andeweg and Cloetingh, 1998). Depending on the amplitude and wavelength of the deflection of the foreland lithosphere, the crust of foreland basins can be affected by an array of basin-parallel generally small-scale normal faults, which developed in response to bending stresses. Whilst the width of a foreland basin essentially depends on the elastic thickness of the foreland lithosphere and the magnitude of loads imposed on it, the build-up of collision-related compressional stresses can cause narrowing and deepening of a foreland basin, thus increasing flexural bending stresses (Ziegler *et al*., 2002). As tensional stresses inherent to the topography of orogens and the buoyancy of their crustal roots (Coblentz *et al*., 1994) die out at the margin of mountain ranges they do not significantly affect foreland domains (Bott, 1993; Bada *et al*., 2001). Therefore, their contribution to the stress field in front of an orogen can be neglected.

Under condition of essentially decoupled collision, major thin-skinned deformation of the sedimentary cover of the foreland can occur, as seen in the Canadian Cordillera (Price, 1981). This is associated with subduction of the crust and lithospheric mantle of the foreland plate. During the evolution of fore-arc foreland basins, and depending on the level of mechanical decoupling of the subducting lower plate with the evolving orogenic wedge, either slab-pull forces are exerted on the foreland or compressional stresses build up within it during the development of a thin-skinned fold-and-thrust belt. This can account for the development of different stress systems within the thrust belt and its underlying basement as well as in the undeformed sedimentary cover of the foreland and its basement.

Although collision-related forces at the Alpine–Carpathian active margin of the West and Central European Phanerozoic Platform are regarded as crucial for its state of stress, loads imposed on its other margins cannot be neglected. Whilst throughout Cenozoic times the East European Craton and the Fennoscandian Shield provided a stable buttress for the Phanerozoic European Platform (Jarosiński and Dąbrowski, 2006), the dynamic setting of the Atlantic margins underwent significant changes during the Paleogene. With the end Paleocene crustal separation between Greenland and Europe the long-standing tensional setting of the European Atlantic margins terminated (Ziegler, 1988) and became gradually replaced by a compressional one as ridge push forces gradually built up during the progressive opening of the Arctic–North Atlantic Ocean (Richardson and Cox, 1984). Whilst stresses related to the evolution of the Pyrenees (Verges and Garcia-Senez, 2001; Andeweg, 2002) and thermal asthenospheric instabilities related to volcanic activity associated with the European Cenozoic rift system (Prodehl *et al*., 1995; Wilson and Downes, 2006) played an important role in the Cenozoic evolution of the stress field in the Alpine Foreland (Dèzes *et al*., 2004) they are probably subordinate in the Carpathian Foreland (Ziegler and Dèzes, 2007).

The objective of this paper is to combine dispersed data pertaining to the Cenozoic evolution of the Carpathian Foreland that is indicative for stress-controlling mechanisms. We specifically attempt to take a broader area into account than is usually considered when trying to correlate tectonic processes of the orogen-foreland system of Poland. This approach provides for a better understanding of dynamic processes that controlled the evolution of the foreland, which in turn have implications for unraveling tectonic processes that played a role in the development of the orogenic wedge.

GENERAL STRUCTURAL SETTING

The area under consideration covers the foreland of the Northern Carpathians, the basement of which comprises a complex set of contrasting tectonic units, represented by the East European Craton (EEC) and the West and Central European

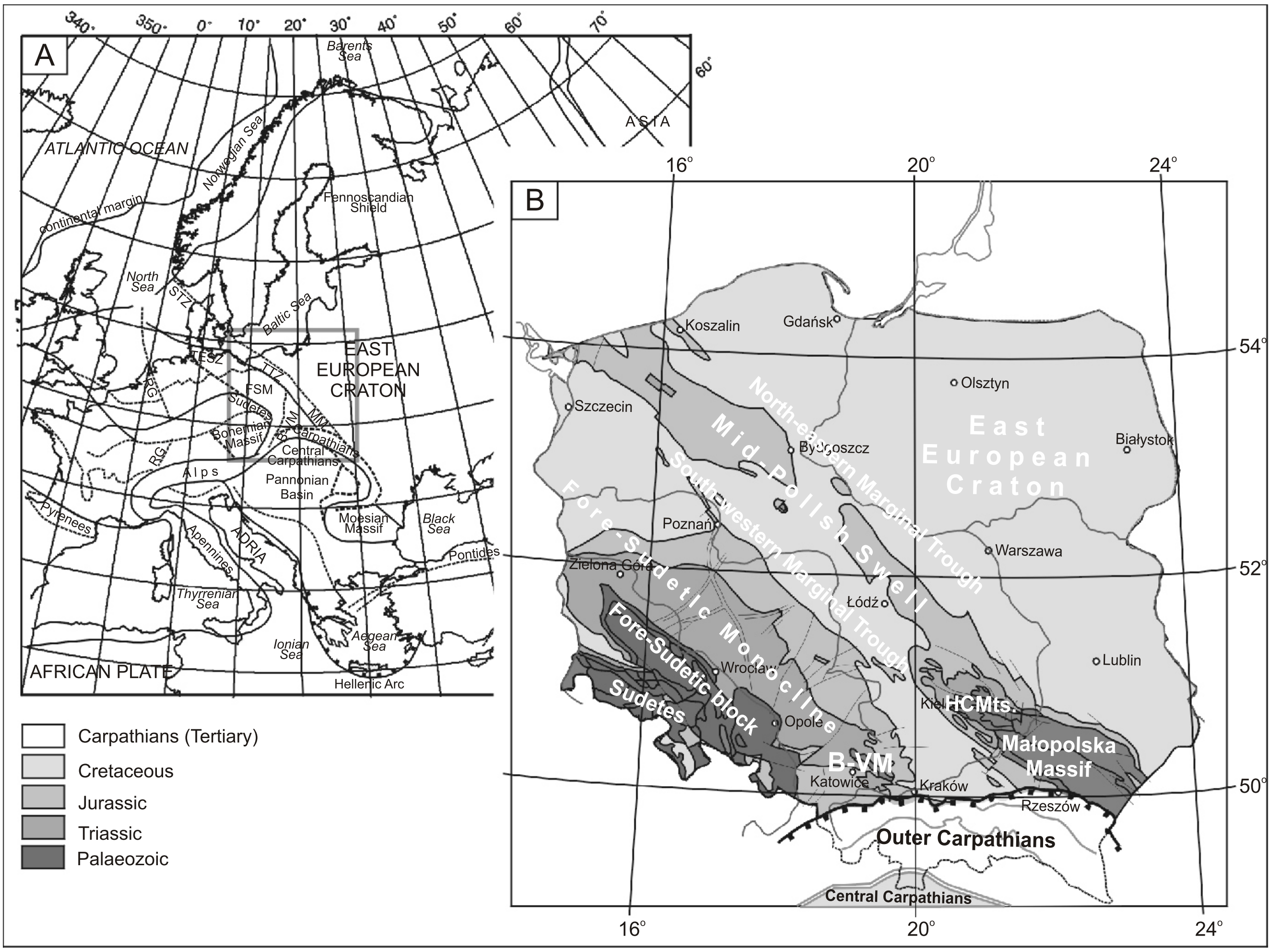

Fig. 1A — study area on the background of the tectonic map of Central Europe (modified after Berthelsen, 1992); B — tectonic setting of the Polish part of the Carpathian Foreland without Cenozoic cover (simplified after Dadlez *et al*., 2000)

B-VM — Bruno-Vistulicum Massif; FSM — Fore-Sudetic Monocline; HCMts. — Holy Cross Mountains; MM — Małopolska Massif; RG — Rhine Graben; STZ — Sorgenfrei-Tornquist Zone; TESZ — Trans-European Suture Zone; TTZ — Teisseyre-Tornquist Zone

Palaeozoic Platform (Fig. 1). The EEC is a mechanically strong and tectonically stable unit (Jarosiński *et al*., 2002) that was consolidated during the Middle Proterozoic (Bogdanova *et al*., 1996; Claesson *et al*., 2001) and is characterized by a relatively low present-day heat flow (Szewczyk and Gientka, 2009). The Palaeozoic Platform of Western and Central Europe developed during two major stages, involving the Caledonian accretion of Avalonia-type terranes and the Variscan accretion of the Armorican-type terranes (Ziegler, 1990; Pharaoh *et al*., 2006). A complex array of smaller terranes that is wedged in between the margin of the EEC and the Caledonian and Variscan terranes forms part of the Trans-European Suture Zone (TESZ) (Pożaryski, 1991; Belka *et al*., 2002; Nawrocki and Poprawa, 2006). A number of fault zones of the TESZ were selectively reactivated during Cenozoic times.

During the Permo-Mesozoic the Polish Basin developed on the basement of the TESZ. The main subsidence centre of this basin, which overlaps the Teisseyre-Tornquist Zone (TTZ), is referred to as the Mid-Polish Trough (e.g., Dadlez, 1989; Krzywiec, 2006*a*). This trough was inverted during the Late Cretaceous and Paleocene into the Mid-Polish Swell (MPS) (e.g., Dadlez, 1989; Ziegler, 1990; Krzywiec, 2006*b*; Fig. 1). Following erosion of the MPS and neighbouring areas, the northern parts of Poland formed a low lying stable platform that was only temporarily flooded by shallow marine transgressions during Eocene, Oligocene and Miocene times (Piwocki and Ziembińska-Tworzydło, 1997). Correspondingly, the northern part of Poland is referred to as the Polish Lowlands.

The southern margin of the Palaeozoic West and Central European Platform was affected by rifting during the Triassic, resulting in the opening of the oceanic Meliata Basin (e.g., Dercourt *et al*., 1990), during the Middle Jurassic, preceding opening of the South Penninic–Vahic Ocean (Schmid *et al*., 2004), and during the Early Cretaceous, controlling subsidence of the Western Outer Carpathian basins (Poprawa *et al*., 2002*a*; Słomka *et al*., 2002; Oszczypko, 2006). The West Carpathian passive margin underwent contraction during the Late Cretaceous and Paleocene (Andrusov *et al*., 1973; Dercourt *et al*., 1990; Poprawa *et al*., 2004) as evidenced by the onset of flysch-type sedimentation (Oszczypko, 2006). This was accompanied by the build-up of intraplate compressional stresses, which controlled the latest Cretaceous and Paleocene inversion of the Mid-Polish Trough (Krzywiec, 2006*b*). Intermittent crustal shortening persisted in the Western Carpathians during Paleogene and early Neogene times and ended during the Late Miocene. The Internal Western Carpathian Wedge (IWCW), located south of the Pieniny Klippen Belt suture, formed part of the ALCAPA (East Alpine–Carpathian–Pannonian) Block upper plate whilst the nappe systems of the Outer Carpathians were detached from the southward subducting lower plate European foreland lithosphere. The Outer Carpathians consist of several stacked nappe units, the most important of which are the Magura, Dukla, Silesian, Subsilesian and Skole units that mainly consist of flysch-type and pelagic sediments (Książkiewicz, 1977; Oszczypko, 2006). These essentially thin-skinned nappes, which account for a few hundred kilometres of shortening (Behrmann *et al*., 2000), override an autochthonous, Neogene foreland basin sequence, which rests unconformably on Mesozoic sediments that in turn unconformably overlay Palaeozoic series and/or the Precambrian basement (Oszczypko, 1997, 2006).

DATA USED

This paper is largely based on a compilation of published and unpublished data, partly collected by the authors. Constraints on Cenozoic sedimentary sequences and their distribution are provided by boreholes, which are widely distributed throughout the analysed area. Definition of the present-day stress field is based on borehole data and involved breakout and hydraulic fracturing analyses on almost a 100 wells (Jarosiński, 2005*a*, *b*).

In the foredeep basin of the Outer Carpathians a dense grid of good quality reflection-seismic profiles (Jarosiński, 1999*b*; Jarosiński and Krzywiec, 2000; Krzywiec, 2001), calibrated by numerous boreholes, provides close control on its tectonic evolution. To the north of the Carpathian Foredeep Basin, where seismic acquisition was focussed on deeper targets, such as Permian hydrocarbon traps below the Polish Basin, the thin Cenozoic sequence is poorly imaged on available profiles. In spite of this drawback, a few high-resolution shallow seismic profiles are available for the Damasławek and Kleszczów depressions (Krzywiec *et al*., 2000).

Another set of data used in the present study comes from structural analyses of outcrops. These are, however, not numerous owing to the thin, though almost continuous Quaternary cover of the Polish Lowlands. Therefore, analysed outcrops are limited to brown coal (Gotowała and Hałuszczak, 2002) and native sulfur (Jarosiński, 1992) opencast mines, as well as a small number of minor outcrops in the Carpathian Foredeep (Rauch, 1998; Krysiak, 2000; Rauch-Włodarska *et al*., 2006).

We also refer to the results of previous numerical modelling, including tectonic subsidence modelling of the Outer Carpathian flysch basins (Poprawa *et al*., 2002*a*, 2006) and the foredeep basin (Oszczypko, 1998, 2006). Moreover, we assessed the sources of the present-day stress field of Central Europe by comparing the World Stress Map database (Reinecker *et al*., 2003, Jarosiński, 2005*a*) with the results of finite element modelling (Jarosiński *et al*., 2006).

CENOZOIC PALAEOGEOGRAPHY AND SUBSIDENCE PATTERNS IN NORTHERN POLAND

The Paleogene and Neogene regional uplift and subsidence patterns of the Polish Platform north of the Outer Carpathian Basin provide valuable constraints for tectonic vertical crustal movements and controlling mechanisms. Below, we outline the regional subsidence trends whilst the subsidence of local grabens will be dealt with in the next paragraph. It should also be kept in mind that on the Polish Platform much of the Cenozoic time span is represented by hiatuses (Fig. 2).

In the Polish Basin, chalk sedimentation persisted into the Early Paleocene (Połońska, 1997), albeit with an increasing influx of clastic material that was mostly derived from eastern sources on the EEC and probably also from the axis of the growing Mid-Polish Swell (MPS). With a few local exceptions, Paleocene sediments are preserved only to the NE of the MPS in a shallow and wide depression where they attain thicknesses of about 30 m, whilst further east on the EEC they reach a maximum thickness of 100 m. In some parts of the MPS, remnants of Paleocene sediments rest unconformably on Late Cretaceous strata (Kramarska *et al*., 1999; Dadlez *et al*., 2000), thus testifying to an end-Cretaceous inversion phase for the MPS (Krzywiec, 2006*b*).

Throughout the Polish Basin, a major erosional gap spans Late Paleocene to early Middle Eocene times (Fig. 2; e.g., Piwocki, 2001), indicating that during the Late Paleocene a much wider area was uplifted than the inverted MPS only. On the Polish Platform, sedimentation resumed during the late Middle Eocene with marine transgressions advancing from the North Sea into NW Poland and from the Carpathian domain through Eastern Poland into the Baltic Sea area (Pożaryska and Odrzywolska-Bieńkowa, 1977). During Eocene times the MPS acted as a topographic barrier between faunal provinces (Fig. 3A). This is evidenced by the dominance of a Mediterranean (Tethyan) fauna in the depression that flanks the MPS to the NE, whilst a Boreal fauna dominates the depression on the SW flank the MPS (Pożaryska and Odrzywolska-Bieńkowa, 1977). Periodic mixing of both faunas indicates that the two faunal realms were intermittently connected with each other, perhaps across the relatively low relief MPS or either in the Baltic domain or in Southern Poland, from where Eocene deposits have been removed by erosion. The extent of Eocene sediments into Southern Poland is limited by erosion across the Meta-Carpathian Swell that was periodically uplifted during the Oligocene and Miocene. Middle and Late Eocene deposits, resting discordantly on older strata, are preserved in shallow depressions that flank the MPS (Fig. 3A) and in which they attain maximum thicknesses of up to 50 m in their northern parts and up to 100 m in their southern parts (Piwocki, 2004). The present-day distribution of Eocene deposits suggests that the MPS either underwent minor inversion during the Middle and Late Eocene or presented at these times an erosional high. However, the occurrence of Eocene deposits in marginal depressions flanking the MPS is in a favour of the first option.

Commencing with the Oligocene the subsidence pattern in the Polish Lowlands was no longer influenced by the NW–SE trending MPS. Instead, the E–W trending Meta-Carpathian

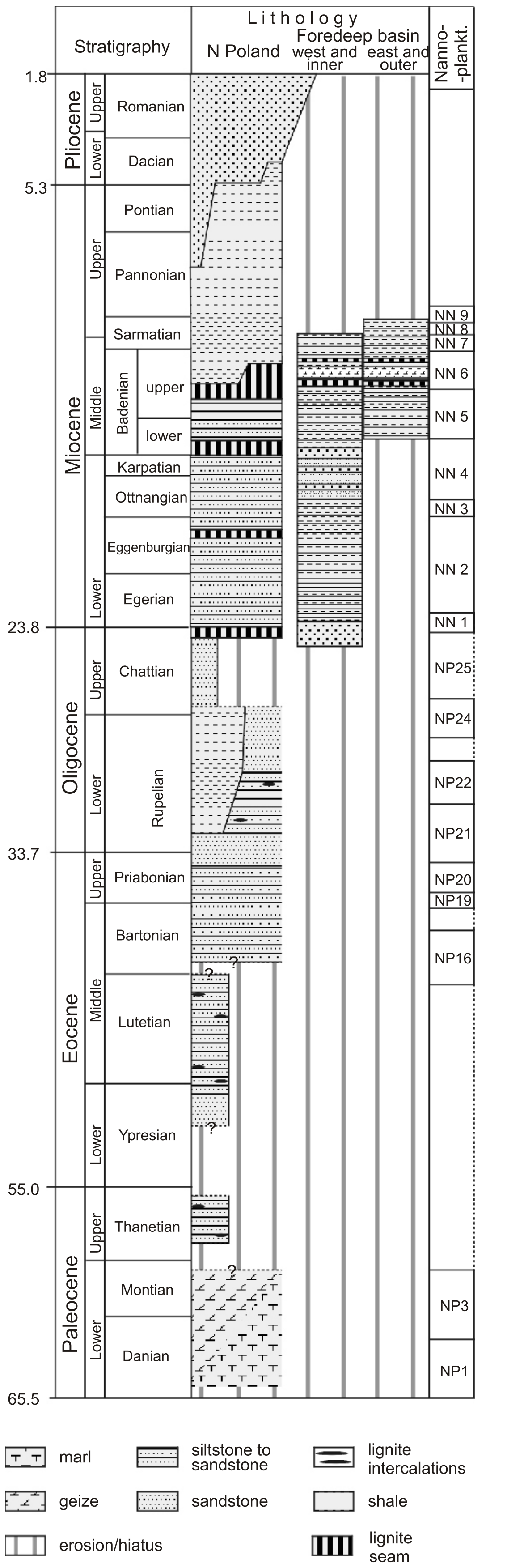

Fig. 2. Generalised Cenozoic chrono- and lithostratigraphic profile for the Polish Lowlands and Carpathian Foredeep compiled after Piwocki and Ziembińska-Tworzydło (1997), Oszczypko (1999), Piwocki and Kramarska (2004)

Swell started to control sedimentation as evidenced by the pinch out of Oligocene deposits on its northern slope (Fig. 3B). North of the Meta-Carpathian Swell, Oligocene sediments attain thicknesses of 60–100 m, rest conformably on Eocene ones and overstep the crest of the MPS where they are only slightly thinner (Piwocki, 2004). The absence of Oligocene marginal depressions flanking the MPS may indicate that its secondary inversion had ceased at the end of the Eocene (Fig. 3B).

During the Neogene the Carpathian Foredeep Basin and the Central European Basin formed part of two distinct faunal provinces. From the Miocene onward the flexural forebulge, which developed at the place of the Meta-Carpathian Swell, separated these basins, thus contributing toward the isolation of the exotic Paratethys faunal province and the resulting difficulties in precise stratigraphic correlation of Neogene series of the Carpathian and North Polish domains (e.g., Steininger and Rögel, 1984; Olszewska *et al*., 1996). In conjunction with uplift of the forebulge the southern segment of the MPS was subjected to erosion at the beginning of the Middle Miocene (Fig. 3C). In the Polish Lowlands Neogene deposits attain outside local grabens an average thickness of 150 m and reach a maximum of 350 m on the Fore-Sudetic Monocline (Piwocki *et al*., 2004). By contrast, the sedimentary fill of the Carpathian Foredeep Basin increases rapidly southward with Badenian–Sarmatian series exceeding a thickness of 2000 m (Oszczypko, 1997).

In the Polish Lowlands, the top part of the Cenozoic section is composed of Pleistocene glacial deposits, typically in the range of 100 to 200 m thick. These form an almost continuous cover as far south as the Carpathians and the Sudetes. On a regional scale the thickness of these deposits is controlled by the extent of glaciers rather than by tectonic subsidence (Lindner, 1992). Nevertheless, many local thickness variations have been attributed to the incision of erosional channels, glacitectonics and faulting (Zuchiewicz, 2000).

EOCENE–MIOCENE FAULTING IN THE POLISH LOWLANDS AND GRABEN DEVELOPMENT

Cenozoic sediments deposited north of the Carpathian Foredeep and the Sudetes are essentially undeformed, apart from active salt-induced structures and narrow grabens occurring in central and SW Poland (Fig. 4). These grabens, which are 1–2 km wide, several kilometres long and several hundred meters deep, have at Mesozoic levels the configuration of negative flower structures, suggestive of their transtensional origin (Widera *et al*., 2008). These structures apparently developed by reactivation of basement discontinuities, which had developed during the Triassic–Early Jurassic under a tensional setting and, comparable to the MPS, underwent partial inversion during the Late Cretaceous under a NE–SW directed transpressional regime (Deczkowski and Gajewska, 1980; Lamarche *et al*., 2002). Although these Cenozoic grabens account for only small amounts of extension they are important stress indicators.

The oldest ENE–WSW and N–S trending grabens started to develop during the latest Late Eocene on the Fore-Sudetic Monocline to the SW of the Poznań–Oleśnica lineament

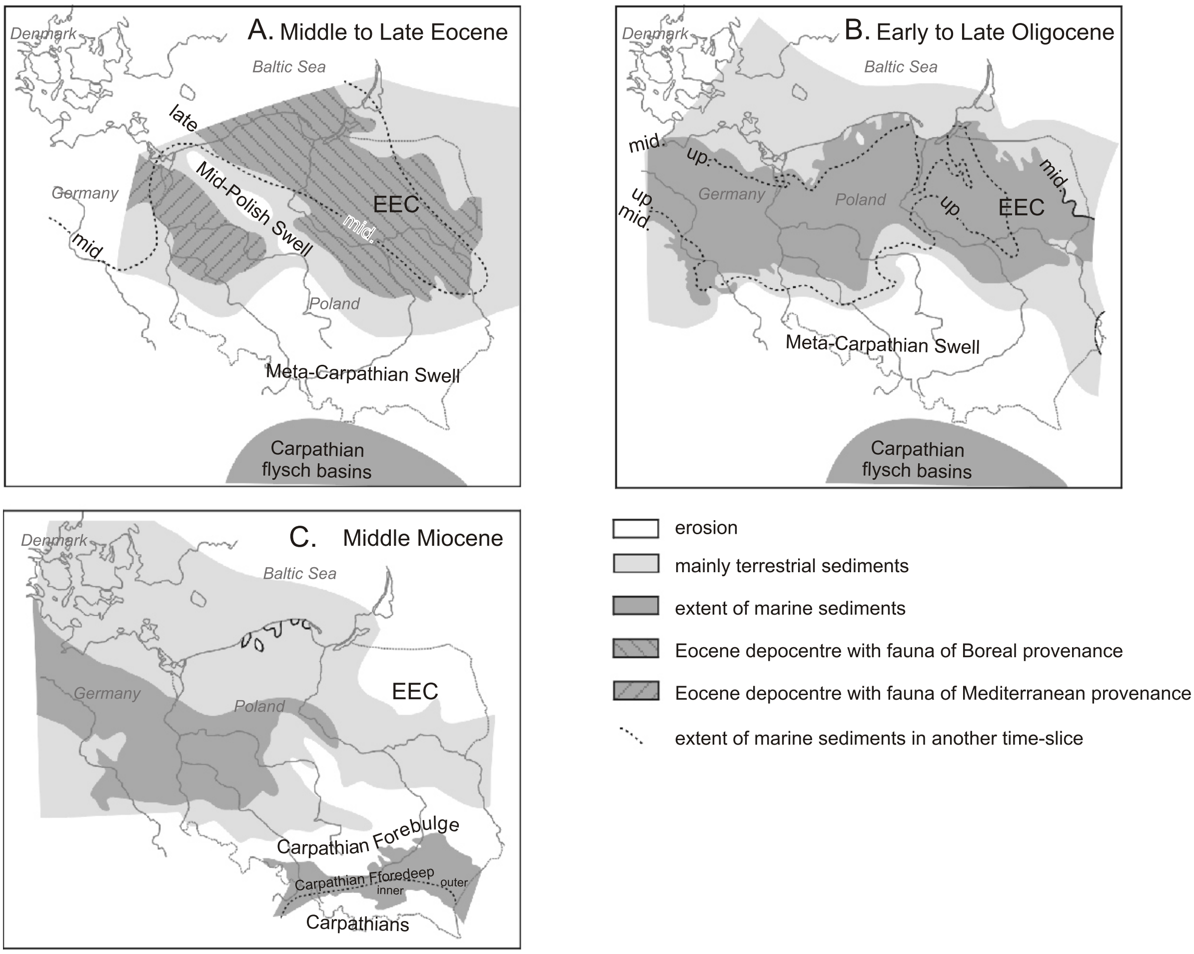

Fig. 3. A changes in Tertiary deposition patterns, compiled after Vinken (1988), Stankowski, (1996), Ważyńska (1998) and Piwocki (2004)

A — Eocene uplift and erosion of Mid-Polish Swell and subsidence of shallow marginal depressions; **B** — Oligocene uplift of Meta-Carpathian Swell and gentle subsidence of Polish Lowlands; **C** — Miocene subsidence of narrow Carpathian Foredeep Basin, coeval uplift of forebulge, residual basin in Polish Lowlands and erosion of the southern segment of the Mid-Polish Swell

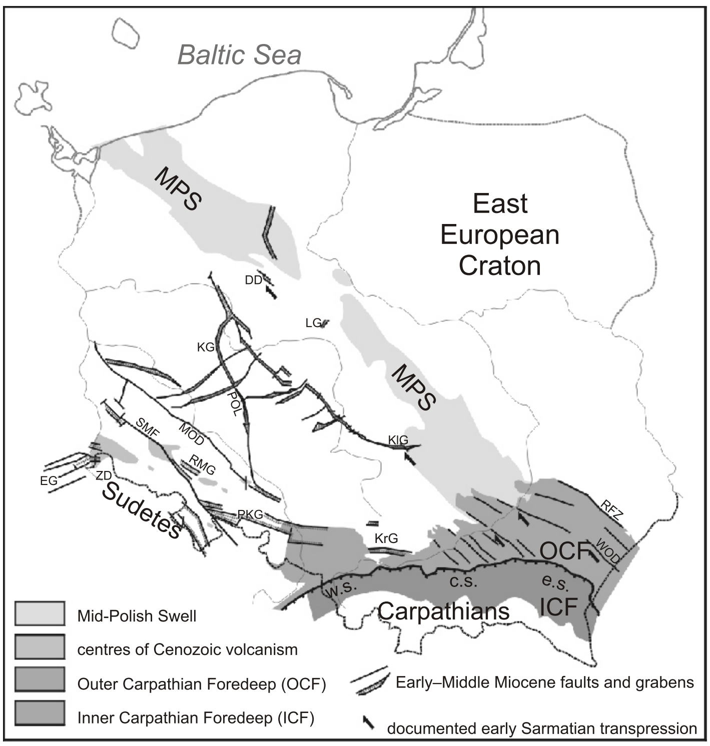

Fig. 4. Miocene tectonic features of the Carpathian Foreland, compiled after Dyjor (1983), Kasiński (1984), Oszczypko (1997), Jarosiński (1999*b*) and Krysiak (2000)

The Mid-Polish Swell (MPS) is limited to the sub-cropping Jurassic rocks below the Cenozoic cover. Oligocene-Pliocene basaltic volcanism is shown for the Lower Silesia. In the Carpathian Foredeep tectonic features are only shown for its outer parts. DD — Damasławek Depression; EG — Eger Graben; KG — Krzywin Graben; KlG — Kleszczów Graben; KrG — Krzeszowice Graben; LG — Lubstów Graben; MOD — Middle Odra Fault Zone; PKG — Paczków–Kędzierzyn Graben; POL — Poznań–Oleśnica lineament; RFZ — Roztocze Fault Zone; RMG — Roztoka–Mokrzeszów Graben; SMF — Sudetic Marginal Fault; WOD — Wielkie Oczy dyslocation; ZD — Zittau Depression; w.s., c.s., e.s. — western, central and eastern segments of the Western Outer Carpathians and their foredeep basin

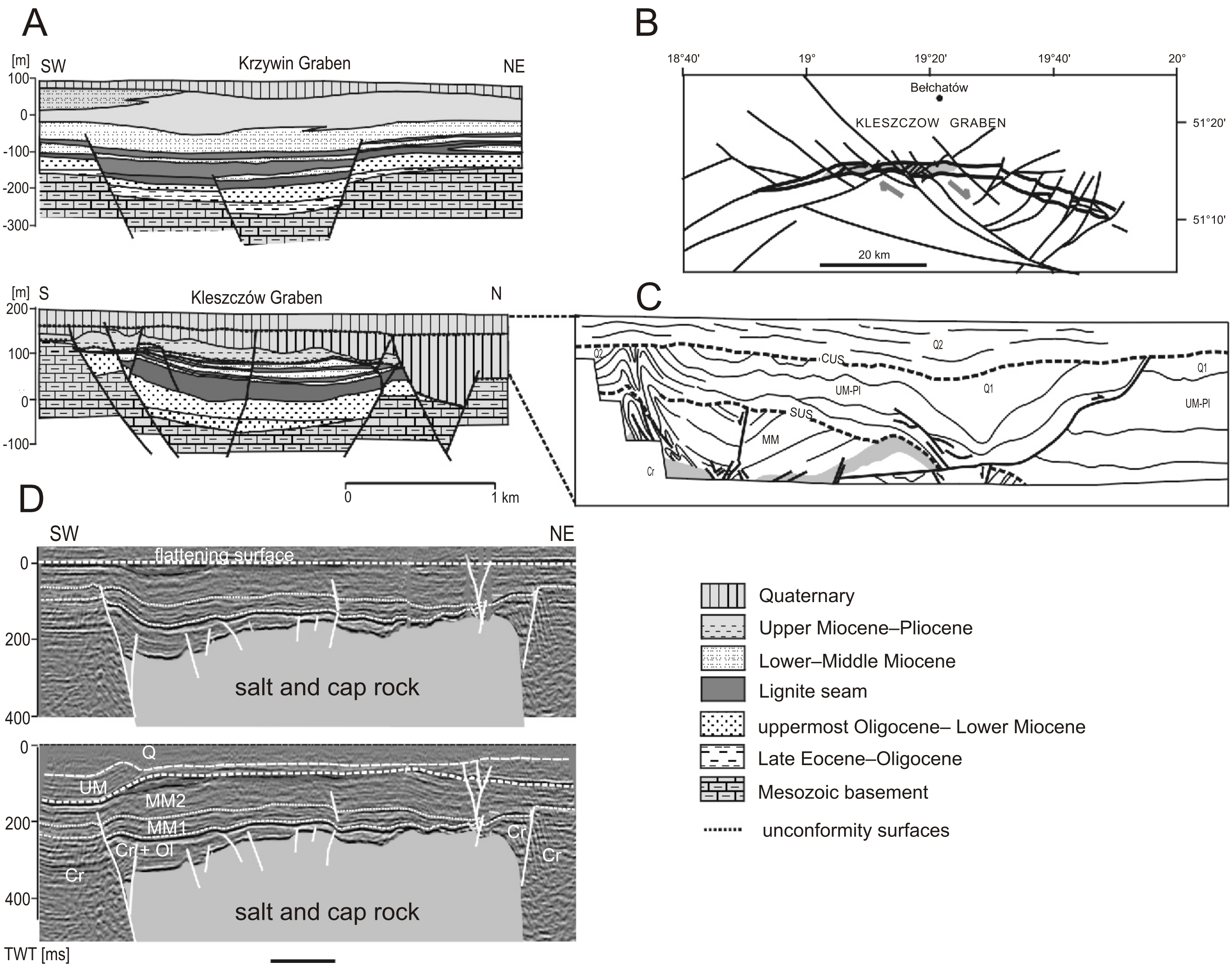

Fig. 5. Grabens in Polish Lowlands

A — Late Eocene–Late Miocene the Krzywin Graben, located in the centre of Poznan–Oleśnica graben system, and Late Oligocene–Late Miocene the Kleszczów Graben (after Kasiński, 1984; for location see Fig. 4); **B** — pattern of faults crossing the Kleszczów Graben with dextral transversal displacement during the Sarmatian and younger phases (after Gotowała and Hałuszczak, 2002; simplified); **C** — zoom on the top of sedimentary fill of Kleszczów Graben (after Hałuszczak and Gotowała, 1999) showing Sarmatian (SUS) and Cromerian (CUS) unconformity surfaces (x2 vertical exaggeration); Q1 and Q2 — Upper and Lower Quaternary, UM-Pl — Upper Miocene–Pliocene?, MM — Middle Miocene; **D** — interpretation of shallow seismic profile across the Miocene depression over the Damasławek salt diapir. Flattening at the top Middle Miocene surface shows thickness increase of the earlier sediments, which is less than the scale of subsidence due to compaction of lignite-bearing strata. Final structuring during Late Miocene and the Quaternary uplift. Cr — Cretaceous; Cr + Ol — Cretaceous and Oligocene sandstone, undifferentiated; MM1 and MM2 — Middle Miocene coal bearing sediments, lower and upper sequences, respectively; UM — Upper Miocene, Sarmatian clay complex; Q — Quaternary

(Deczkowski and Gajewska, 1980), whereas salt dome-related depressions of this age occur in the Szczecin Trough. For the Poznań–Oleśnica Graben system borehole data from the Krzywin Graben indicate an initial latest Eocene and Early Oligocene subsidence stage (Kasiński, 1984; Widera *et al*., 2008; Fig. 5A), during which 130 m thick marine-lagoonal sediments containing brown coal intercalations were deposited. During this stage the northern segment of this graben system subsided faster than its southern parts. After a Late Oligocene break in sedimentation, subsidence of these grabens resumed, as evidenced by the accumulation of 300 m thick Early Miocene limnic and fluvial sediments containing brown coal seams (Kasiński, 1984). An up to 100 m thick coal seam sequence was deposited during the Middle Miocene, whilst the centre of subsidence shifted to the southern segment of the graben system (Widera *et al*., 2008). Differential subsidence of the Krzywin Graben may have continued during the Late Miocene and Pliocene, as indicated by lateral thickness changes of the order of 20–50 m. Some of this subsidence can, however, be attributed to compaction of the older coal- and clay-rich graben fill. The compaction coefficient, which defines the ratio between the depositional and present thickness of sediments, is 2 and 2.5 for the first and the second lignite seam, respectively (Widera, 2002). As lignite compaction is fast during its early burial stages, this mechanism may underlay the observed Late Miocene graben subsidence.

East of the Poznań–Oleśnica lineament, similar grabens started to develop towards the end of the Oligocene or at the beginning of the Miocene in response to the reactivation of major tectonic lineaments forming part of the TESZ. In some of these grabens opencast brown coal mining permits to closely constraint their structural evolution.

The best-documented Kleszczów Graben developed at the intersection of tectonic structures within the TTZ to the NW of the Holy Cross Mts. (Fig. 5A, B). Late Oligocene–Early Miocene subsidence of this graben was driven by transtensional reactivation of conjugate WSW–ENE and NW–SE trending faults in response to WNW–ESE directed compression, enhanced by NNE–SSW directed extension (Gotowała and Hałuszczak, 2002). During the late Middle Miocene, purely normal fault stress conditions developed while the direction of extension shifted toward N–S. After this phase, subsidence of the graben ceased and accumulation of lignite-bearing deposits ended. Finally, during the Sarmatian the Kleszczów Graben was mildly inverted, involving the development of local thrust faults and uplift of a gentle anticline that was followed by erosion (Gotowała and Hałuszczak, 2002). Pannonian and early Pleistocene deposits rest discordantly on an erosional surface (Fig. 5C). Whilst their thickness increases into the graben, there is no evidence for syn-depositional faulting. Therefore, this thickness increase may reflect differential compaction of the graben fill with respect to graben shoulders, rather than a further extensional pulse. Finally, during the Pleistocene (Cromerian Stage) the central part of Kleszczów Graben was deformed in a transpressional mode creating a younger erosional surface (Fig. 5B). This may reflect transpressional reactivation of the main transversal fault zone that underlies the Dębina salt dome. Within the Kleszczów Graben Cenozoic series attain a thickness of up to 400 m; part of the underlying subsidence may, however, be attributed to withdrawal of the Zechstein salt from the rim syncline of an adjacent salt dome.

On top of some Zechtein salt diapirs depressions began to subside at the same time as the grabens; this is suggestive of a common controlling mechanism. For example, on top of the Damasławek salt diapir, that is associated with a NW–SE trending salt swell within the TTZ, an oval-shape, fault-bounded depression began to subside during the Early Miocene in which brown coal bearing sediments were deposited (Fig. 5D; Jarosiński, 1999*a*; Krzywiec *et al*., 2000). During the Late Miocene this depression was inverted, involving the development of positive flower structures. Moreover, the Damasławek Depression records also minor Pleistocene compressional reactivation (Fig. 5D). Early and Middle Miocene subsidence of the Damasławek Depression and its Late Miocene and Pleistocene inversion are interpreted to reflect, respectively, transtensional and transpressional reactivation of the NW–SE trending sub-Zechstein fault that underlies the salt swell from which the Damasławek diapir rises (Jarosiński, 1999*a*).

CENOZOIC TECTONIC EVOLUTION OF THE SUDETIC AREA

The main phases of the Cenozoic evolution of the Sudetes and the Fore-Sudetic Block correlate with those evident in other parts of the Polish Lowland Platform. Nonetheless, more intense faulting with significant vertical offsets, as well as the occurrence of volcanism is specific for this area. During the Paleocene and the Eocene the area of the Sudets and Fore-Sudetic Block was uplifted and subjected to erosion. During the Oligocene and Miocene marine ingressions reached this area from the Central European Basin (Oberc, 1972). Only during Middle Miocene times was the eastern Sudetic Basin linked with the Carpathian Foredeep Basin.

Tertiary deposits in the Sudetes and on the Fore-Sudetic Block fill narrow subsidence centers, frequently limited to grabens (Dyjor, 1983). The largest of these grabens, the Paczków–Kędzierzyn and Roztoka–Mokrzeszów grabens, which are located along the Sudetic Marginal Fault Zone, began to subside during the latest Oligocene to Early Miocene (Dyjor, 1976; Fig. 4). During this initial subsidence stage, marine sedimentation in these grabens was controlled by W–E trending faults along which basalts erupted during the Early Miocene. During the Middle Miocene continental coal-bearing sediments accumulated in these grabens. During the Late Miocene, reactivation of NW–SE trending faults prevailed. A further graben, referred to as the Zittau Depression, developed at the junction between NW–SE trending set of the Sudetic faults and NE–SW trending faults in the continuation of the Eger Graben (Kasiński, 1984). This depression was filled by up to 400 m thick sediments during four depositional cycles (Kasiński, 2004). Sedimentation commenced in the Zittau Depression toward the end of Oligocene (cycle 1). Subsidence, accompanied by faulting increased during the Early Miocene depositional cycles 2 and 3, both of which terminate with thick lignite seams (Karoń, 2000). The lower Middle Miocene cycle 4 attains a maximum thickness of 250 m and contains distributed lignite seams. This sequence is disconformably covered by Pliocene coarse clastics lacking evidence for tectonically controlled subsidence.

A characteristic feature of the Sudetes and the Fore-Sudetic Block is the Oligocene, Miocene and Pliocene basaltic volcanism. This volcanic activity, which is widespread in the area of the Eger volcano-tectonic zone of the Bohemian Massif (Ulrych *et al*., 1999), did, however, not result in the extrusion of thick traps in the territory of Poland that covers the NE periphery of this volcanic province. The initial Oligocene phase of volcanism (30–26 Ma) was widely distributed over Lower Silesia (e.g., Birkenmajer *et al*., 2004). Basalts of the second Oligocene–Early Miocene cycle (26–18 Ma) have a more restricted distribution and are localized within the Sudetes and along their marginal fault. The third stage of volcanism straddling the Miocene/Pliocene boundary (5.5–3.8 Ma) is linked to the Central Sudetes volcanic province that remained active in the Czech Republic until the early Pleistocene (Ulrych *et al*., 1999; Birkenmajer *et al*., 2004). The origin of Cenozoic basaltic volcanism in the Sudetes is linked to upper mantle magma chambers (Wierzchołowski, 1993; Alibert *et al*., 1987) and partial melting induced an upwelling asthenospheric convective instability (Downes, 2001; Wilson and Downes, 2006).

An important tectonic reconstruction of the Sudetes occurred during the Late Miocene when the Sudetic Marginal Fault started to control sedimentation. This, according to Dyjor (1993), ended the development of the Meta-Carpathian Swell in the Lower Silesia region. Late Miocene reactivation of the Sudetes was followed during the Pliocene by a major volcanic event and uplift of the entire region, including elevation of the Sudetes with respect to the Fore-Sudetic Block by 1200–1500 m (Dyjor, 1983; Oberc, 1972). This uplift history of the Sudetes is compatible with the late stage evolution of the entire Bohemian Massif (Malkovsky, 1975; Badura *et al*., 2007; Ziegler and Dèzes, 2007).

EVOLUTION OF THE OUTER CARPATHIAN FLEXURAL FOREDEEP

The Polish Carpathian Foredeep Basin (CFB), which formed part of the Paratethys system of basins (Steininger and Rögel, 1984; Hámor, 1988), developed during the Early Miocene (Ottnangian) in front of the advancing Carpathian thrust belt and persisted until the late Sarmatian (Oszczypko and Ślączka, 1989; Oszczypko, 1997, 2006). The CFB can be divided into an internal part, which was overridden by the Outer Carpathian nappes, and an external part, which is located in front of them.

The internal CFB contains continental Early Miocene deposits, which are followed by marine Middle Miocene (up to early Badenian) sediments that attain a combined thickness of up to 1500 m in the western segment of the foredeep. The shift of the CFB from an internal to an external position occurred during the Middle Miocene and was preceded by an early Badenian subsidence interruption (Oszczypko, 1999). During early Badenian to Sarmatian times, the subsidence centre of the external CFB migrated progressively eastward where its sedimentary fill, including thick Sarmatian deposits, exceeds a thickness of 2000 m. Exploration for natural gas and sulfur in the external parts of the CFB (Fig. 6) has yielded detailed information on its structural and stratigraphic configuration that permits to reconstruct its evolution.

Prior to the subsidence of the external CFB, the area now occupied by it was exposed to erosion and may have formed part of the Meta-Carpathian Swell (Oszczypko and Ślączka, 1989; Krysiak, 2000). On the southern slope of this swell several hundred metres deep channels were incised into Mesozoic series (Fig. 7A). These channels are filled with Middle (?Early) Miocene deposits. This provides an upper limit to the timing of their development whilst the onset of their incision is more controversial. These channels transect the peneplain that developed after the latest Cretaceous and Paleocene inversion of the Mid-Polish Swell and cut variably into Mesozoic, Palaeozoic and Precambrian rocks. Large parts of this peneplain are still preserved below the Miocene sediments of the CFB, suggesting that the erosional event underlying the development of these channels was of short duration. Therefore, we hypothesize that these channels developed on the southern slope of flexural forebulge that formed during Late Oligocene–Miocene times in the place of the former Meta-Carpathian Swell, which according to our interpretation, developed in response to lithospheric buckling. This forebulge was flanked to the south by the internal CFB in which related erosion products were deposited. This is compatible with observations in the Kraków area, where erosional channels are exposed at the surface and

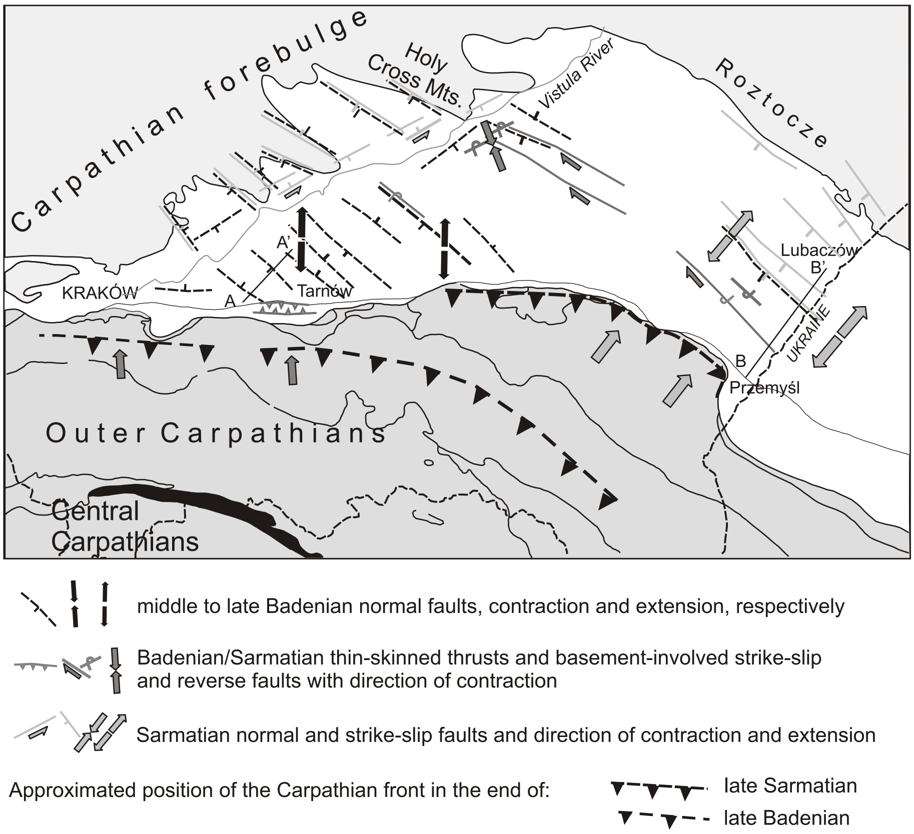

Fig. 6. Evolution of structures in the Outer Carpathian Foredeep Basin

Three tectonic phases are distinguished: (1) middle to late Badenian phase of extension due to plate flexure; (2) Badenian/Sarmatian episode of basement-involving contraction; (3) Sarmatian phase of extension due to plate flexure and stacking of the nappes. Possible positions of propagating front of folding in the Carpathians between these stages are shown

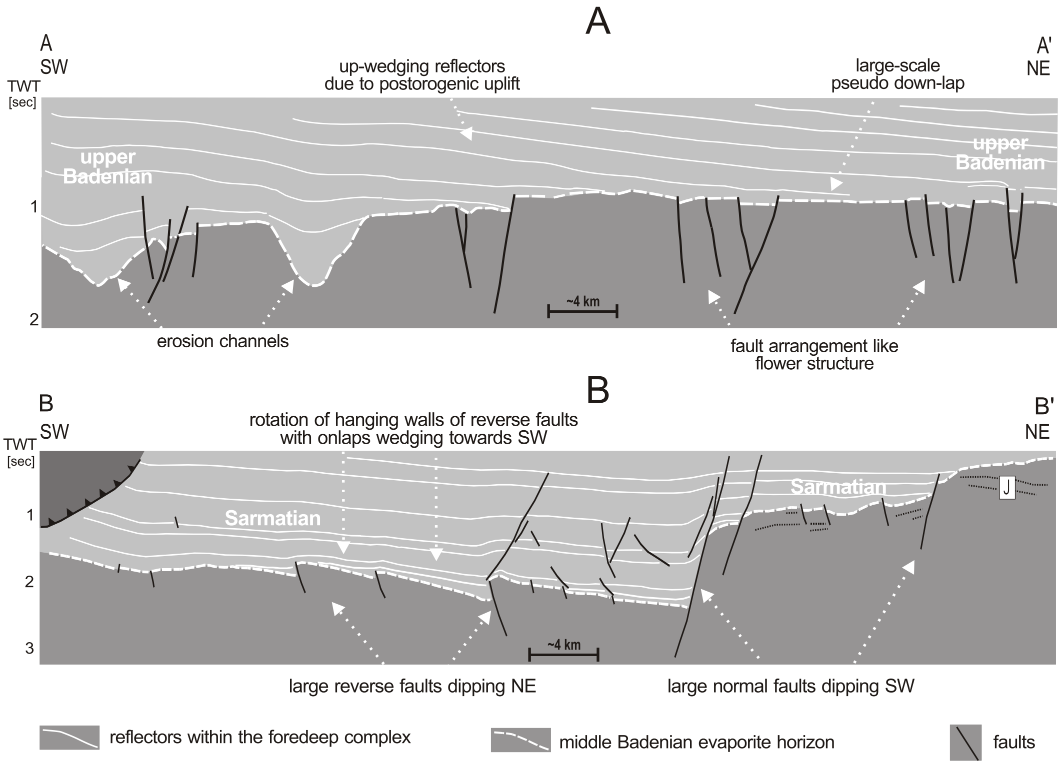

Fig. 7. Interpretation of seismic cross-sections through the Outer Carpathian Foredeep Basin (for location see Fig. 6)

A — subsidence in the central segment was governed by the plate flexure resulting in distributed normal faulting. Strata down-lapping towards the foreland were originally horizontal but were tilted during post orogenic uplift of the Carpathians; **B** — in the eastern segment (seismic interpretation after Krzywiec, 2001), sedimentation was accompanied by large-scale normal faulting resulting from flexural deformation of the foreland lithosphere. Sarmatian transgression was affected by block tilting involving reverse faulting along transpressional zones

the timing of their incision was estimated as being of Oligocene to Karpatian age (Felisiak, 1992). This coincides, at least partly, with the development of the inner CFB (Ney *et al.*, 1974; Oszczypko, 1998). However, it has to be kept in mind that similar erosional channels, evident in the Moravian part of the Carpathian Foredeep, appear to date back to the latest Cretaceous main inversion phase of the Bohemian Massif, since they are filled with Paleocene, Eocene and Early Oligocene clastics that were derived from the foreland; these are unconformably covered by Miocene sediments derived from the Carpathians (Brzobohaty *et al.*, 1996; Picha, 1999).

Early Badenian subsidence and transgression of the external CFB of Poland (Oszczypko and Ślączka, 1989) was accompanied by only minor syn-flexural normal faulting (Krysiak, 1987, 2000; Jarosiński, 1992). During deposition of the middle Badenian evaporites only a low level of tectonic activity has been inferred from earthquake-related sedimentary structures (Peryt and Jasionowski, 1994). During the late Badenian, flexural subsidence rates increased as evidenced by a rapid deepening of the basin and its sediment starvation (Oszczypko, 1997), that was followed by its rapid infilling with late Badenian and Sarmatian orogen-derived sediments. This subsidence phase was accompanied by the development of system of syn-flexural synthetic and antithetic normal faults with throws of the order of 100 m, partly involving reactivation of pre-existing faults (Krysiak, 2000). These faults die out upward in upper Badenian siliciclastics (Fig. 7A). The lithospheric effective elastic thickness of the Małopolska Massif, which forms the basement of the central segment of the CFB, was estimated at 10–18 km. This relatively low figure is compatible with the results of elastic plate flexure models (Royden and Karner, 1984; Krzywiec and Jochym, 1997; Zoetemeijer *et al.*, 1999) and estimates of the depth to the neutral surface derived from structural and sedimentological data (Jarosiński and Krzywiec, 2000).

The first event of basement-involving contraction in the CFB is dated as straddling the Badenian–Sarmatian boundary (Jarosiński, 1992; Jarosiński and Krzywiec, 2000). Although reverse faulting is best documented in the eastern segment of the CFB (Krzywiec, 2001; Fig. 7B), minor structures occur also in the distal part of the central segment. NW–SE trending blind reverse faults and pop-up structures attain vertical offsets of up to 200 m with Sarmatian deposits discordantly covering slightly tilted hanging walls (Dziadzio, 2000; Fig. 7B). Similar structures were described from an opencast sulfur mine in the distal zone of the CFB where brachyanticlines, reverse faults and strike-slip faults die out in the basal part of the Sarmatian sequence (Krysiak, 1985, 2000; Jarosiński, 1992; Fig. 8). The geometry of these structures points to NNW–SSE directed contraction, which is subordinate to dextral transpression along WNW–ESE trending basement faults (Lamarche *et al.*, 2002).

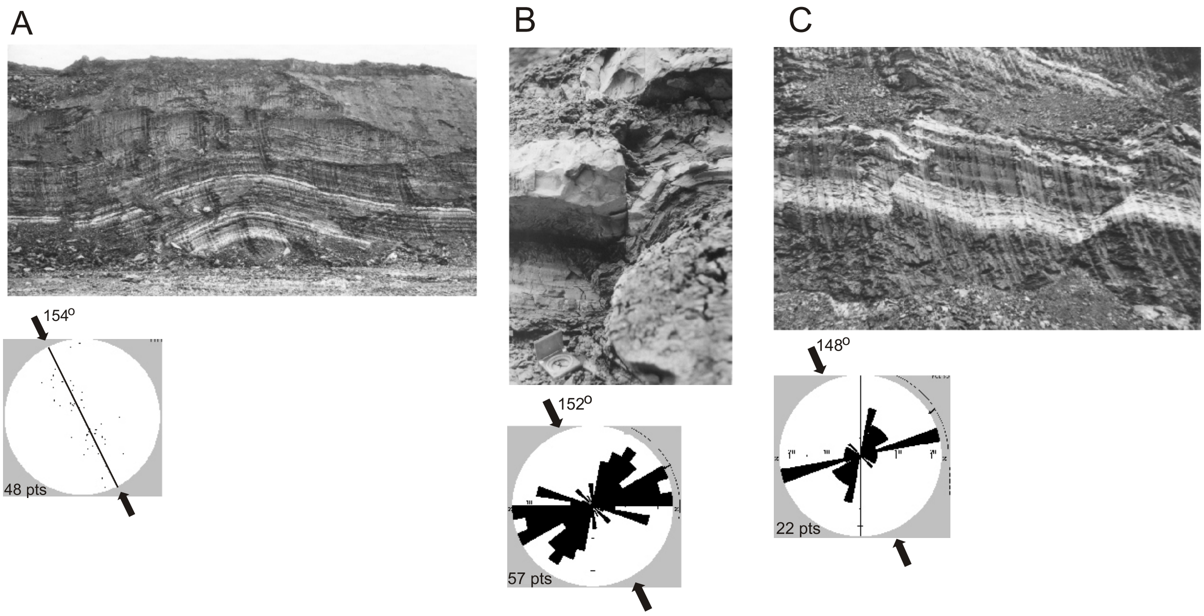

Fig. 8. Machów opencast sulfur mine showing wrench-induced deformation of upper Badenian–lower Sarmatian sediments above transpressional basement faults

A — brachyanticline; **B** — reverse fault; **C** — inverted normal fault; all features point to NNW–SSE contraction

This stage of transpressional foreland deformation was contemporaneous with the emplacement of the Carpathian thrust front at its present position in the central segment of the foredeep, as evidenced by the age of the corresponding triangle zone (Jarosiński and Krzywiec, 2000; Krzywiec *et al*., 2004).

After the short compressional pulse at the Badenian–Sarmatian transition, a new phase of enhanced basin subsidence commenced in the Sarmatian (Oszczypko, 1999), during which the depot centre of the CFB shifted from its central to its eastern segment. Subsidence of this segment was controlled by sharp deflection of the foreland lithosphere, causing the development of NW–SE striking syn-flexural synthetic normal faults with displacements ranging from 100 m up to 1 km (Fig. 7B; Jaroszewski, 1977; Krzywiec, 2001). These faults die out systematically upward in Sarmatian series and control the NE extent of the CFB (Fig. 7).

The sedimentary record of the CFB ends in the Sarmatian. Judging from the inclination and apparent large-scale northward down lap of the uppermost part of the foreland basin sedimentary fill (Krzywiec, 1997; Fig. 7) a several hundred metres thick sedimentary sequence was eroded from its top during Late Miocene and Pliocene times (Poprawa *et al*., 2002*b*). Late stage uplift of the CFB was accompanied by tectonic activity along its north margin (Krysiak, 1985, Jarosiński, 1992). Left-lateral transtensional movements along WSW–ENE-striking faults were inferred from outcrops (Osmólski *et al*., 1978; Jarosiński 1992; Krysiak, 2000) and further constrained by seismic profiles, which reveal NE–SW striking negative flower structures in Sarmatian series (Jarosiński, 1999*b*). Along the northern margin of the CFB the occurrence of local remnants of coarse terrestrial clastics in the vicinity of normal faults scarps (Krysiak, 2000; Rauch-Włodarska *et al*., 2006) points to the latest Miocene–Pliocene extensional stress regime.

NEOTECTONICS AND RECENT STRESS AND DEFORMATIONS

Quaternary phases of neotectonic activity are dated in the Polish Lowlands as Cromerian–Mindel, Mazowsze and post-Riss (Baraniecka, 1983). These phases are differentiated on the base of troughs, in which the thickness of Pleistocene strata increases to 50–200 m. The tectonic origin of these troughs is, however, poorly constrained. The main sets of Pleistocene troughs occur on the flanks of the Mid-Polish Swell (Baraniecka, 1995). Their en echelon arrangement suggests dextral motion along the NW–SE-striking TTZ. A compressional stress regime is evidenced by reverse faults, which cut Pleistocene sediments in the Kleszczów Graben (Hałuszczak *et al*., 1995), as well as in the Damasławek Depression that is located on top of a salt diapir (Fig. 5). In the Kleszczów Graben a significant unconformity developed at the top of the Cromerian interglacial sequence (*ca*. 0.4 Ma) in response to compressional deformation of the Dębina salt diapir (Hałuszczak, 2004).

Neotectonic faulting is also evident in the Sudetes (Fig. 9) where the Marginal Sudetic Fault Zone and subordinate W–E-trending faults have Quaternary vertical offsets in the range of 50–150 m (Dyjor, 1995; Badura *et al*., 2007). Neotectonic displacements with similar amplitudes are suggested for the Upper Silesian region north of the Carpathian Foredeep and along the Roztocze morphological swell (Harasimiuk and Henkiel, 1975; Laskowska-Wysoczańska, 1995; Lewandowski, 1995; Zuchiewicz *et al*., 2007).

These scattered examples of neotectonic activity (e.g., Liszkowski, 1982; Zuchiewicz, 1995; Zuchiewicz *et al*., 2007), combined with the results of borehole break-out analyses and earthquake focal mechanisms (Jarosiński, 2005*a*, 2006), permit

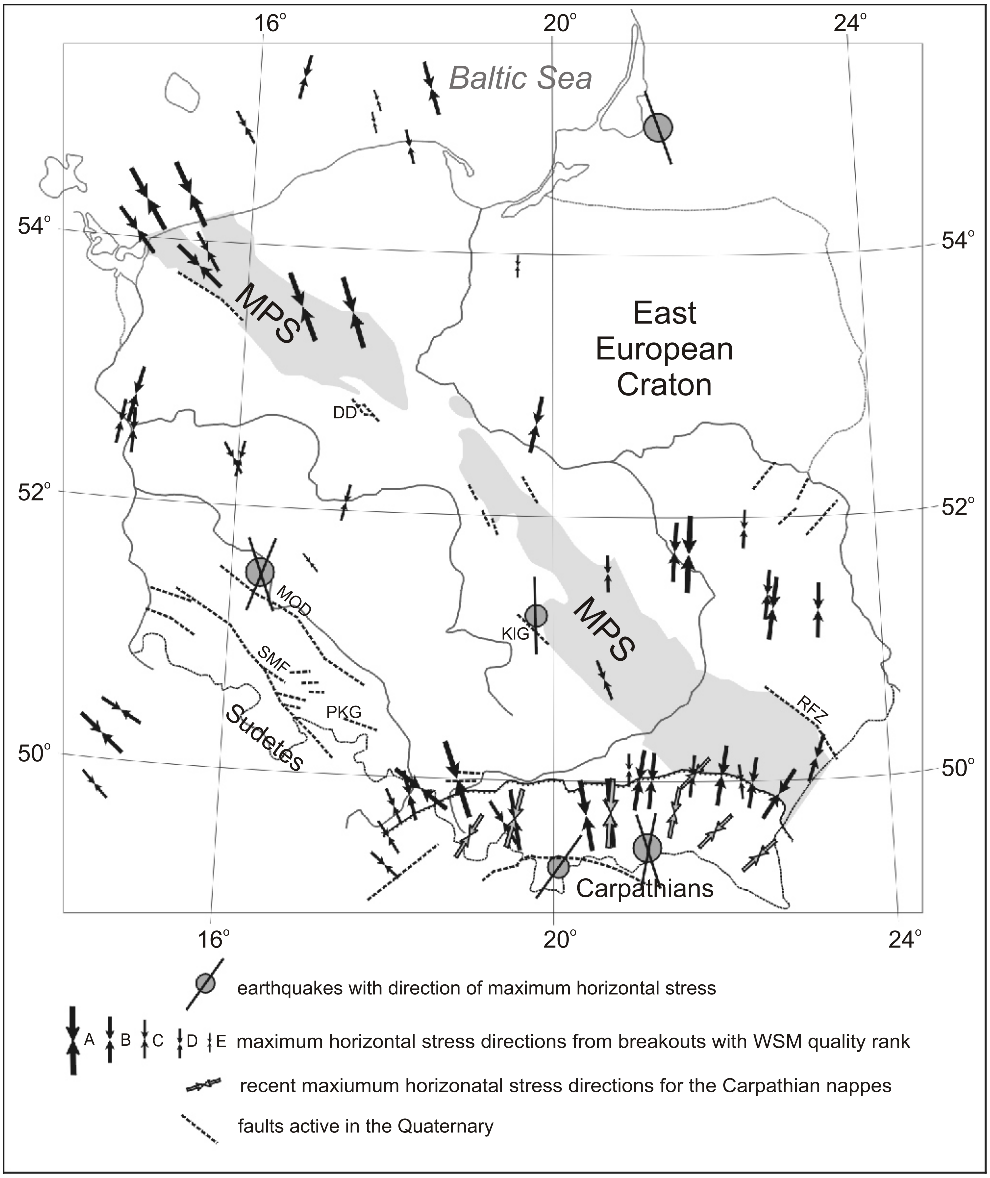

Fig. 9. Neotectonics and recent geodynamics of the Carpathians and their foreland

Stress directions from earthquake focal mechanism compiled after: Gibowicz *et al*. (1982), Gibowicz (1984), Wiejacz (1994). Stress directions from breakouts after Jarosiński (2005*a*). Neotectonically active faults are adopted after Dyjor (1995), Baraniecka (1995), Lewandowski (1995), Laskowska-Wysoczańska (1995). Explanations as in Figure 4

to reconstruct the present-day stress field of the Polish Lowlands and the Carpathian Foreland (Fig. 9).

In the basement of the Outer Carpathians and their foreland the trajectories of the maximum horizontal compressional stress axes describe a fan-like pattern that diverges from NW-directed in the western-most Carpathians and the Sudetes to NNE directed in the eastern Carpathians and the Małopolska Massif (Jarosiński, 1998). On the EEC and in the adjacent MPS compressional stress trajectories are NNE directed in Southern Poland and NNW directed in Northern Poland and the Baltic Sea. The stress field of Poland (Fig. 9) fits smoothly into the stress field of Western and Central Europe that is controlled by Alpine–Carpathian collisional forces and Arctic–North Atlantic ridge push (Gölke and Coblentz, 1996; Müller *et al*., 1997; Jarosiński *et al*., 2006). Within the TESZ the frequently observed rotations of stress trajectories between N–S and NW–SE in individual vertical boreholes as well as between boreholes is suggestive of a strike-slip regime (Jarosiński, 2005*a*). Particularly in the western segment of the Carpathians, stress trajectories in the basement are NNW to NW directed whilst in the nappes they are NNE directed, thus pointing to a decoupling of the thin-skinned nappes from their basement and strain partitioning between them.

Fault plane solutions of earthquakes and mining-induced tremors (Guterch and Lewandowska-Marciniak, 2002) suggest for the Carpathians in the vicinity of the Pieniny Klippen Belt (Wiejacz, 1994) either a strike-slip or thrust fault stress regime, and for Central Poland (Kleszczów Graben) and the southern part of the Fore-Sudetic Monocline a compressional stress regime (Gibowicz *et al.*, 1982; Gibowicz, 1984; Gotowała and Hałuszczak, 2000; Fig. 9). For SE Poland a strike-slip fault stress regime is indicated by the results of hydraulic fracturing tests in boreholes (Jarosiński, 2005*b*).

GEODYNAMICS OF THE CARPATHIAN FORELAND *VERSUS* FAR FIELD TECTONIC LOADS: A DISCUSSION

FACTORS CONTROLLING THE FORELAND STRESS FIELD

We interpret that the present-day first-order stress pattern of Poland is controlled by forces related to the collisional interaction of the Carpathian and East Alpine orogens with the European Platform, as well as by ridge-push forces, which are exerted by the still active spreading axes of the North Atlantic, Norwegian–Greenland Sea and the Eurasian Basin (Gölke and Coblentz, 1996; Müller *et al.*, 1997; Heidbach *et al.*, 2007). Intraplate stress sources, such as lateral changes in the density and/or thickness of the continental crust, thermal loads of mantle plumes, as well as topography and glacial rebound appear to play at best a subordinate role in the present-day stress pattern of the Polish Platform. Nevertheless, Late Cretaceous and Cenozoic intraplate deformations recorded in the Polish Lowlands, the Carpathian Foreland and the Outer Carpathian Basin testify to repeated stress field changes, both in magnitude and orientation, reflecting changes in the interaction of the evolving Carpathian and East Alpine orogens with the European Platform as well as in the level far-field ridge push forces.

CARPATHIAN OROGEN

Collisional interaction of the Carpathian Orogen with its foreland exerted intermittently tectonic loads on the Bruno-Vistulian and Małopolska massifs. Palinspastic restorations of the Outer Carpathian nappe stack indicate that it developed out of a 300–350 km wide continental shelf and slope that was subdivided by the Silesian Ridge into the proximal Silesian–Skole Basin complex and the distal Magura Basin (Behrmann *et al.*, 2000). Reconstruction of tectonic processes operating at the Carpathian margin of European Platform is difficult and ambiguous due to its subsequent collisional deformations. However, the different stages of tectonic subsidence and uplift of the individual basins are rather well constrained (Poprawa *et al.*, 2002*a*; Poprawa and Malata, 2006; Oszczypko, 2006) and were used as rough indicators for changes in the geotectonic regime.

The Carpathian Basin system was probably floored by highly attenuated continental crust and bounden to the south by the oceanic Vahic (Pieniny) Basin (Oszczypko, 2006). Closure of the Vahic Ocean commenced during the Late Cretaceous (Schmid *et al.*, 2008) and gave rise to the build-up of compressional stresses in the North Carpathian passive margin controlling Turonian and early Senonian initial inversion movements in its Silesian and Skole sub-basins (Poprawa *et al.*, 2002*a*) as well as late Senonian–Early Paleocene inversion of the Mid-Polish Trough in the foreland (Krzywiec, 2006*b*). Towards the end of the Paleocene, compressional foreland stresses relaxed, as evidenced by the termination of inversion movements in the MPS and accelerated subsidence of the Outer Carpathian domain.

During Late Paleocene to Middle Eocene times of relatively slow convergence rates the load of the advancing Inner West-Carpathian orogenic wedge (IWCW) and the associated Pieniny Klippen Belt and Magura Ridge accretionary wedge caused flexural subsidence of the Outer Carpathian Basin complex and a progressive northward shift of the Magura Basin depocentre (Poprawa *et al.*, 2002*a*; Oszczypko, 2006). During the Early and Middle Eocene the Silesian and Skole sub-basins subsided under decreasing sedimentation rates to deeper water conditions whilst in the Magura Basin clastic supply derived from the southward adjacent IWCW increased.

During the latest Eocene and earliest Oligocene the Silesian and Skole sub-basins were tectonically rapidly uplifted, as evidenced by a significant reduction of water depths that cannot be explained by sediment accumulation nor by a eustatic sea level change (Poprawa *et al.*, 2002*a*). This uplift reflects a renewed build-up of compressional stresses in the foreland of the advancing IWCW and initiation of folding in the Magura Basin (Tokarski and Świerczewska, 1998).

Eocene/Oligocene uplift of the Outer Carpathian Silesian and Skole domains was followed by their subsidence that relates to Late Oligocene–Early Miocene deposition of the thick Krosno flysch, folding and northward thrusting of the Magura nappe and subduction of the foreland lithosphere. By early Burdigalian times the thin-skinned Magura nappe had reached the southern margin of the Silesian sub-basin (Poprawa *et al.*, 2002*a*; Oszczypko, 2006). Commencing in the Burdigalian, the sedimentary fill of the Outer Carpathian Basin was scooped out by a system of major NE-verging thin-skinned thrust sheets while its basement was subducted beneath the rapidly advancing IWCW. Motion of the latter was controlled by the eastward lateral extrusion of the ALCAPA Block under the impact of the Adriatic indenter and by roll back of the Carpathian subduction slab (Wortel and Spakman, 2000; Schmid *et al.*, 2008). Mechanical decoupling of the IWCW from the subducting foreland lithosphere allowed for the development of an extensional stress regime in the autochthonous basement of the Carpathians during the Miocene stacking of their flysch nappes. Subduction-related magmatic activity in the internal Carpathians indicates that detachment of the subducted European foreland slab had commenced around 20–17.5 Ma in the westernmost parts of the Carpathians, from where it propagated between 14–9 Ma eastward through the Polish sector into the Ukraine (Nemčok *et al.*, 1998; Wortel and Spakman, 2000; Harangi *et al.*, 2006). With this, slab-pull forces exerted on the Polish Carpathian Foreland decayed and isostatic uplift of the orogenic wedge and the proximal parts of the CFB commenced. Moreover, Late Miocene slab detachment allowed for a renewed build-up of compressional stresses in the Carpathian basement.

EAST-ALPINE OROGEN

The East-Alpine Orogen intermittently exerted tectonic loads on the analyzed area via the Bohemian Massif. Geophysical data indicate that the European foreland crust extends from the thrust front of the Eastern Alps by about 125 km southward beneath them (Schmid *et al*., 2004, 2008). During the Maastrichtian–Paleocene collision of the Austro-Alpine orogenic wedge with this part of the European passive margin, related intraplate compressional stresses caused a profound disruption of its sedimentary cover, upthrusting of basement blocks in the Bohemian Massif and inversion of Mesozoic tensional basins, such as the Lower Saxony, Altmark–Brandenburg and North Danish basins, at distances of up to 1500 km to the NW of the contemporary collision front (Ziegler, 1990; Ziegler *et al*., 1995, 1998; Ziegler and Dèzes, 2007). The main inversion phase of the MPS coincides with this collisional event (Krzywiec, 2006*b*).

In the East-Alpine Foreland, collision-related intraplate stresses relaxed at the transition to the Eocene, as indicated by the termination of inversion movements and the development of a broad flexural foreland basin in response to thrust and slab loading by the East-Alpine orogenic wedge (Wagner, 1996, 1998). By late Eocene times the subducted oceanic Penninic slab was detached from the foreland lithosphere (Dèzes *et al*., 2004, 2005). During the Oligocene–Early Miocene emplacement of the Austro-Alpine nappe stack, involving about 250 km of shortening (Wagner, 1996; Schmid and Kissling, 2000), continental European lithosphere was subducted southward. This was accompanied by rapid subsidence and narrowing of the foreland basin involving syn-flexural normal faulting (Wagner, 1998). The absence of contemporaneous compressional intraplate deformations in the European foreland is indicative for its mechanical decoupling from the East-Alpine orogenic wedge, presumably in response to sediment subduction. During the Eocene and Oligocene the inversion-induced topographic relief of the Bohemian Massif was degraded to a peneplain. Oligocene plume-related magmatic activity abated in the Eger volcano-tectonic zone prior to the accumulation of up to 500 m of latest Oligocene to Burdigalian sediments under a mildly tensional setting (Malkovsky, 1975, 1979). By about 20 Ma the East-Alpine thrust front had reached its present position and eastward expulsion of the ALCAPA Block commenced owing to continued N-ward movement of the Adriatic indenter (Ratschbacher *et al*., 1991; Peresson and Decker, 1997; Schmid *et al*., 2004). In the area of the Eastern Alps, related wrench movements caused detachment of the subducted south-dipping continental slab. This was followed by northward subduction of Adriatic lithosphere (Schmid *et al*., 2004; Kissling *et al*., 2006). As a result, mechanical coupling between the East-Alpine orogenic wedge and its European foreland increased. This is evidenced by regional uplift of the northern parts of the Bohemian Massif around 18 Ma in response to lithospheric folding (Ziegler *et al*., 2002; Ziegler and Dèzes, 2007). Moreover, from 15–13 Ma onward compressional reactivation of fault systems of the Bohemian Massif resulted in progressive uplift of its marginal blocks, such as the Sudetes and the Moravo-Silesian Block, disruption of the pre-existing peneplain and a resurgence of volcanic activity spanning 11.4–3.95 Ma (Ziegler and Dèzes, 2007).

ARCTIC–NORTH ATLANTIC PASSIVE MARGINS AND SEA-FLOOR SPREADING AXES

Passive margins flanking extra-Alpine Europe developed during the Cretaceous and Cenozoic by northward propagation of the Central Atlantic sea-floor spreading axis in response to clock-wise rotational westward drift of Laurentia relative to Eurasia. Stepwise opening of the North Atlantic, Norwegian–Greenland Sea and the Eurasian Basin, involving repeated changes in the pattern of sea-floor spreading axes and plate boundaries, was preceded by a long history of Mesozoic rifting, during which large areas around the future passive margins were subjected to tensional stresses (Ziegler, 1988, 1989; Torsvik *et al*., 2002).

Opening of the Atlantic Ocean to the north of the Azores fracture zone commenced during the Early Cretaceous (*ca*. 120 Ma) whilst north of the Charlie Gibbs fracture zone sea-floor spreading began in the Labrador Sea during the Campanian (*ca*. 80 Ma) or Paleocene (62 Ma). Between Greenland and Europe and in the Eurasian Basin sea-floor spreading commenced during the earliest Eocene (54 Ma). Separation between NE Greenland and the Svalbard became only effective during the Early Miocene (20–15 Ma) when the Knipovich Ridge that links the Mohns and Nansen spreading axes started to develop (Ziegler, 1988; Lundin, 2002; Torsvik *et al*., 2002; Engen *et al*., 2008). Crustal separation between Europe and Greenland was preceded by the impingement of the Iceland plume around 62 Ma. This plume affected prior to the onset of sea-floor spreading at 54 Ma an area with a radius of 1100 km that was centered on Iceland. By contrast opening of the Eurasian Basin and the North Atlantic between the Azores and Charlie Gibbs fracture zones was preceded by only minor magmatic activity (Ziegler, 1988). An increase in ridge push force is evidenced by the development of local compressional structures on the shelves of the British Isles and Norway during the Late Eocene–Early Oligocene reorganization of sea-floor spreading axes and again during the Miocene (Ziegler *et al*., 1995; Doré and Lundin, 1996; Mosar *et al*., 2002).

Loads exerted on passive margins are, by definition, stable as they are determined by slowly increasing ridge push, resulting from persisting sea-floor spreading. Estimates of variations in the magnitude of ridge push forces as a function of time (Parsons and Sclater, 1977; Wortel, 1980; Bott, 1991), point to a systematic increase of North Atlantic push forces on the North Sea segment of the passive margin, amounting to *ca*. 0.5×10^{12} N/m at the end of the Eocene, 1.5×10^{12} N/m in the Miocene and 2.0×10^{12} N/m at present-day (Gölke and Coblentz, 1996; Andeweg, 2002). Similar ridge push force build-up rates were obtained for the Australian–Antarctic ridge, where during the last 20 Ma a 0.8×10^{12} N/m increase was calculated (Dyksterhuis and Müller, 2004).

Counteracting deviatoric tension inherent to a continental margin with a thick crust and a lithospheric mantle characterized by a lower density than that of the oceanic lithosphere (Bott, 1991) is in the range of -1.0×10^{12} N/m (Kusznir, 1991). Taking this into account for the Eocene–Oligocene development of the passive Norwegian margin, minor tensional stresses may have propagated into the continent. This does, however, not exclude local compressional stress concentrations at the passive margin. Ridge push forces built up during the Miocene, contributing to intra-continental compression, which grew in time to the present-day magnitude of 0.8–1.4 $\times 10^{12}$ N/m when calibrated to the gravitational potential energy at sea level (Jarosiński *et al*., 2006).

Tensional stresses emanating from passive margins may render the respective plate interior more sensitive to changes in tectonic loads at its collisional margin. Moreover, relaxed passive margins combined with compression at active margins, promote extensional and strike-slip deformation of continental intraplate domains. By contrast, compressional stresses exerted on the passive margins rendered the respective plate interior to behave more stiffly and promotes intraplate thrusting, folding or transpression in response to strong indenting at its active margin.

RESPONSE OF CARPATHIAN FORELAND TO FAR-FIELD STRESSES

Tectonic loads exerted on the active Alpine–Carpathian margin and on the passive Arctic–North Atlantic margin of the European Platform evolved independently, giving rise to different combinations of stress-generating factors. Below, we attempt to correlate the timing of geodynamic events in the northern Carpathian Foreland with the possible state of loading at these margin segments and to assess their relative importance.

The Late Cretaceous and Paleocene uplift of the Mid-Polish Swell predated the Cenozoic evolution of the Carpathian Foreland that is the main subject of the following discussion. This major inversion of Permo-Mesozoic extensional basins in Western and Central Europe is generally regarded as relating to the closure of the Penninic Ocean and the collision of the Austro-Alpine orogenic wedge with the continental Briançonnais terrane in the Western and Central Alps and with the European passive margin in the Eastern Alps (Ziegler *et al*., 1995, 1998; Dèzes *et al*., 2004; Ziegler and Dèzes, 2007). The case of the Polish Basin was to some extent more complex, as it was subjected to compression from both the SW and S, namely from the East Alpine and the Carpathian collision zone, respectively. A genetic relation between the collision of the IWCW with the foreland plate and the inversion of the Polish Trough, at least of its southern segment, is suggested by the correlation of the Late Cretaceous and Paleocene deformation phases of these domains (Poprawa and Malata, 2006).

During the regional Late Paleocene–Early Eocene erosional phase the relief of the inverted MPT may have been degraded to a peneplain. In the Carpathian Foreland there are no structural data available on the stress regime during this post-inversion stage. Plate boundary conditions indicate relaxation of collision-related compressional stress that were projected into the European foreland by the East Alpine and Carpathian orogenic wedges and relaxation of tensional stresses at the North Atlantic margin in conjunction with crustal separation between Greenland and Europe at the Paleocene–Eocene transition. From the above, relaxed stress conditions are expected for the Late Paleocene–Early Eocene evolution of the Carpathian Foreland and Polish Lowlands.

MIDDLE AND LATE EOCENE UPLIFT STAGE OF THE MID-POLISH SWELL

In the Polish Lowlands, sedimentation resumed during the Middle and Late Eocene in gentle depressions, which flanked the MPS (Fig. 3). Development of these marginal depressions and apparent gentle uplift of the MPS can be interpreted as reflecting a second, albeit minor phase of compressional reactivation of the MPS. This suspected secondary inversion pulse is only documented in the northern parts of the MPS as in its southern parts corresponding sediments are missing owing to erosion across the Meta-Carpathian Swell that was uplifted during the Oligocene and Miocene.

Interpreting this late stage topography of the MPS as an effect of intraplate compression meets, however, with the difficulty as there is no evidence for a Middle to Late Eocene renewed build-up of intraplate compressional stresses in the foreland of the Carpathians and the Eastern Alps (Ziegler, 1990; Ziegler and Dèzes, 2007). Alternatively, the postulated gentle Eocene uplift of the MPS could be attributed to a rapid release of compressional stresses after the strong Paleocene phase of basin inversion, according to a mechanism proposed by Nielsen *et al*. (2005). Yet, also for this model we are unable to show evidence for Eocene contraction that is followed by a relaxation event. Thus, for the time being we have to leave the question open whether there is a tectonic control on the Middle–Late Eocene subsidence and erosion pattern in the Carpathian Foreland or whether these are effects of remnant topography resulting from the Paleocene inversion of the MPS.

LATEST EOCENE–EARLY OLIGOCENE RELAXATION OF THE FORE-SUDETIC MONOCLINE AND COMPRESSIONAL BUCKLING IN FRONT OF THE CARPATHIANS

On the Fore-Sudetic Monocline narrow and shallow grabens started to develop during the latest Eocene and Early Oligocene, involving transtensional reactivation of pre-existing basement discontinuities. From a regional point of view, development of these grabens coincides with the activation of the European Cenozoic Rift System (ECRIS) that extends from the coastal Netherlands into the western Mediterranean (Ziegler, 1992, 1994; Prodehl *et al*., 1995). The probably "plume"-related Oligocene basaltic volcanism of the Bohemian Massif (Wilson and Downes, 1991, 2006; Ulrych *et al*., 1999; Ziegler and Dèzes, 2007), which extends eastward as far as Lower Silesia, did, however, not reach the Fore-Sudetic Monocline.

The hypothesis that the end Eocene–Early Oligocene extensional phase of the Fore-Sudetic Monocline may be related with the activation of the ECRIS in the Alpine Foreland is difficult to defend. Significantly, tensional subsidence of the shallow Eger Graben commenced only during the latest Oligocene and continued during the Early Miocene. There is no evidence for Late Eocene and Oligocene extension in the Bohemian Massif. Oligocene magmatic activity preceded subsidence of the Eger Graben (see Ziegler and Dèzes, 2007). Subsidence of minor grabens on the Fore-Sudetic Monocline involved, nevertheless, transtensional reactivation of pre-existing basement discontinuities, and as such reflects that also in the Polish part of the European foreland collision-related intraplate stresses began to build up during the activation of the ECRIS.

At the Eocene–Oligocene transition syn-flexural subsidence of the Molasse Basin accelerated. This was accompanied by the development of an array of essentially basin-parallel synthetic and antithetic normal faults with throws of the order of a few 100 m (Wagner, 1996, 1998). As by end Eocene times the south-dipping subducted slab beneath the Central and Eastern Alps had already been detached from the European lithosphere (Dèzes *et al*., 2005; Ziegler and Dèzes, 2007), these nor-

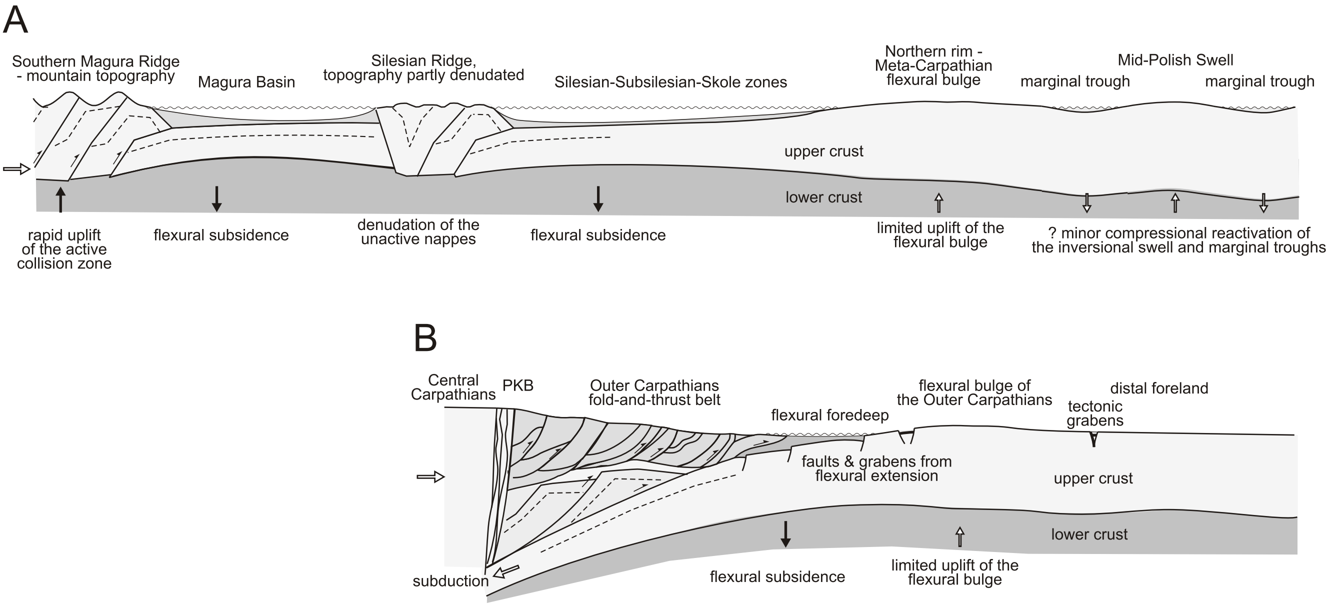

Fig. 10. Schematic cross-sections through the Western Outer Carpathians and their foreland, illustrating general tectonic relations between the two domains

A — Eocene (after Poprawa and Malata, 2006, modified and supplemented): thick-skinned contraction and collision south of the Magura Ridge Basin, caused flexure of the Magura Basin, inversion of the Silesian Ridge and buckling of the Silesian–Skole Basins and uplift of the Meta-Carpathian Swell; **B** — Middle Miocene: southward subduction of the foreland lithosphere and thin-skinned northward thrusting of the Outer Carpathian fold-and-thrust belt, accompanied by development of the flexural foredeep and forebulge involving normal faulting; PKB — Pieniny Klippen Belt

mal faults probably developed exclusively in response to thrust-loaded deflection of the foreland lithosphere. These syn-flexural faults young northward in tandem with the northward shift of the axis of the Molasse Basin in front of the advancing Alpine nappes.

The extensional event on the Fore-Sudetic Monocline at the Eocene–Oligocene transition coincides in the Polish Lowlands with a change from a subsidence pattern controlled by the NW–SE trending Mid-Polish Swell to one controlled by the W–E trending Meta-Carpathian Swell. We postulate that the related uplift of the Meta-Carpathian Swell and gentle subsidence of the Polish Lowlands involved long-wavelength folding of the lithosphere (e.g., Cloetingh *et al*., 2002) in response to the build-up of compressional stresses in the Carpathian Foreland (Fig. 10A). This is compatible with the Late Eocene–Early Oligocene, compressional deformation of the Outer Carpathian system of basins (Poprawa *et al*., 2002*a*). The Oligocene termination of graben subsidence on the Fore-Sudetic Monocline may be attributed to a relaxation of these foreland stresses.

EARLY MIOCENE TO BADENIAN FORELAND EXTENSION

In the Polish Lowlands the second phase of graben subsidence commenced during the latest Oligocene. Compared to the first graben-forming phase a much wider area was affected by the second tensional phase covering apart from the EEC the entire Carpathian Foreland. During the Early to Middle Miocene all grabens in the Polish Lowlands and the Sudetes, as well as in the Carpathian Foredeep remained tectonically active.

The latest Oligocene–Early and Middle Miocene subsidence of grabens in the distal foreland was coeval with the development of the Carpathian flexural foreland basin (Oszczypko, 2006), subduction-related volcanic activity in the Pannonian region (Pécskay *et al*., 1995; Nemčok *et al*., 1998; Harangi *et al*., 2006) and intense thin-skinned thrusting in the Outer Carpathians (Picha, 1999; Oszczypko, 2006). This points to an acceleration of subduction processes in response to E-ward expulsion of the ALCAPA Block, involving sediment subduction, and a decoupling of the IWCW from the foreland lithosphere (Fig. 11C). Mechanical decoupling of the colliding upper and lower plate can account for normal faulting in the foreland lithosphere in response to its thrust- and slab-loaded deflection triggering the flexural mechanisms of foredeep subsidence and forebulge uplift (Fig. 10B). This "soft-collision" concept is compatible with the "slab-retreat" model advocated by Royden (1993).

Within the Eastern Alps, E-ward expulsion of the ALCAPA Block commenced during the Early Miocene and was followed by two-stage counter-clockwise rotation of ALCAPA Block, namely during the Ottnangian and in the Karpatian–lower Badenian (Márton and Fodor, 1995). This lateral escape tectonics implies major transpressional movements between the colliding ALCAPA Block and the Carpathian Foreland lithosphere (Schmid *et al*., 2008). The shallow Eger Graben, which is considered as forming part of the ECRIS, subsided during the Late Oligocene and Early Miocene prior to its Middle Miocene uplift (Malkovsky, 1987) that involved lithospheric folding in response to the build-up of compressional foreland stresses (Ziegler and Dèzes, 2007). This reflects Early Miocene increasing mechanical coupling between the East Alpine orogenic wedge and its foreland.

On the other hand, with progressive opening of the Arctic–North Atlantic Ocean ridge push forces exerted on the Eu-

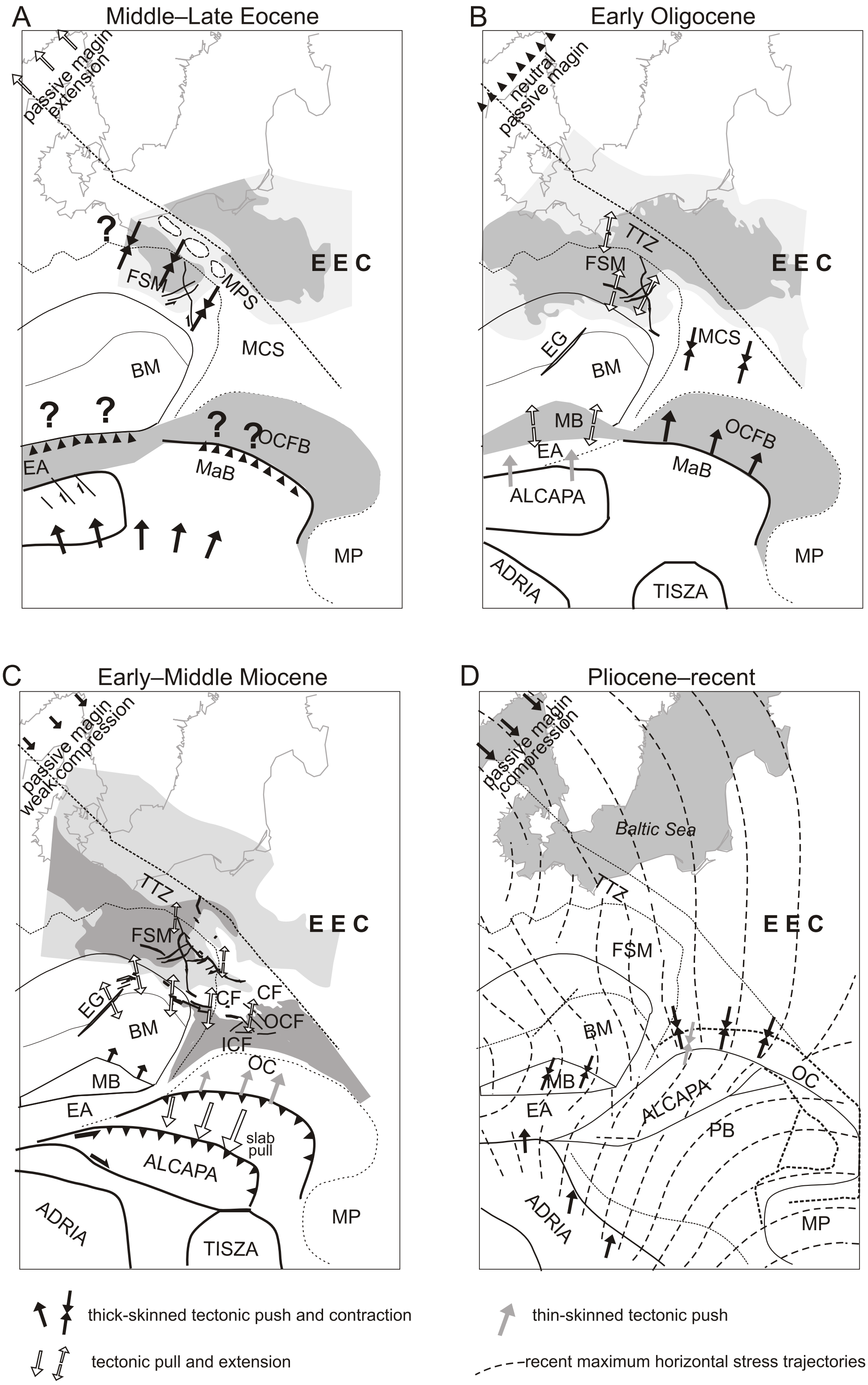

Fig. 11. Sketch showing correlation of stress regime evolution in the northern foreland of the Carpathians with changes of tectonic loads at the segments of the Alps, the Carpathian and the passive margin

Subsidence pattern, marked in grey, is taken after **A** — during the Eocene minor inversion stage of the Mid-Polish Swell there is no evidence for strong compression at any of the analyzed plate boundary; **B** — Oligocene uncoupled collision in the Eastern Alps resulted in flexural subsidence of the Molasse Basin and extension in the western part of Poland. Basement-involving compression in the Carpathian flysch domain causing buckling of the Carpathian Foreland; **C** — latest Oligocene–earliest Miocene eastward extrusion of the central Eastern Alps. Position of ALCAPA Block after extrusion is adopted after Márton and Fodor (1995). Successive subduction and uncoupled collision in the Carpathians produced far-ranging pull to the foreland plate; **D** — recent compressive stress regime is forced by the push of indenting Adria micro-plate, sustained by the Atlantic ridge push from the side of NW European passive margin. Recent stress trajectories after Jarosiński *et al*. (2006); BM — Bohemian Massif; CF — Carpathian forebulge; EA — Eastern Alps; EEC — East European Craton; FSM — Fore-Sudetic Monocline; MaB — Magura Basin; MB — Molasse Basin; MCS — Meta-Carpathian Swell; MP — Moesian Platform; MPS — Mid-Polish Swell; OC — Outer Carpathians; OCFB — Outer Carpathian flysch basins; PB — Pannonian Basin; other explanations as in Figures 1 and 4 ropean passive margin gradually increased during the Miocene, as evidenced by the development of inversion structures on the Mid-Norway and Shetland shelves (Lundin and Doré, 2002). Moreover, compressional stresses built up also in the West European foreland as indicated by Late Oligocene and Early Miocene inversion of e.g. the Channel and Western Approaches areas (see Ziegler, 1990; Ziegler *et al*., 1995) and Burdigalian lithospheric folding controlling uplift of the Vosges-Black Forest arch that extends into the Bohemian Massif and the Massif Central (see discussions in Dèzes *et al*., 2004; Ziegler and Dèzes, 2007).

We postulate that the Carpathian slab pull forces may be responsible for the transtensional reactivation of pre-existing basement discontinuities controlling development of minor grabens in the Polish Lowlands and the Sudetes. Moreover, flexural bending stresses probably contributed to tensional faulting in the foredeep basin and possibly also on the forebulge where plate curvature reached a maximum (Lamarche *et al*., 2002).

LATEST BADENIAN–EARLIEST SARMATIAN COMPRESSIONAL EPISODE

Thin-skinned, north-directed thrusting terminated in the central frontal segment of the Carpathians at the beginning of Sarmatian, whilst in the eastward-adjacent segment Sarmatian sediments are involved in the frontal Stebnik and Sambir units of the Carpathians (Oszczypko, 2006). Thus, folding ended after the Sarmatian around 10.5 Ma when uplift of the foreland basin commenced. According to results of structural analyses this was associated with a change from tension to contraction in the basement of the Carpathian Foreland (Jarosiński, 1999*b*; Jarosiński and Krzywiec, 2000; Lamarche *et al*., 2002). This short episode of transpression is well documented in the eastern and distal part of the foredeep, but not in the front of the orogen where bending-related tension was highest and probably effectively counteracted contraction.

Although stratigraphic correlations between the Paratethys domain and the Polish Lowlands are not precise, inversion of the Damasławek Depression and the Kleszczów Graben, both located at the SW border of the MPS, approximately correlate with the Sarmatian termination of strike-slip movements in the Carpathian Foredeep. The above-mentioned compressional deformations are located along the SW border of the TTZ and its continuation beneath the foredeep basin (see Fig. 4), indicating minor transpressional reactivation of this basement-involving structure. Such a tectonic regime might be attributed to the final episode of thrusting in the central segment of Outer Carpathians and N–S convergence of ALCAPA Block with the Polish Carpathian Foreland lithosphere.

SARMATIAN EXTENSIONAL PHASE IN THE FOREDEEP BASIN

The short foreland compressional event at the beginning of Sarmatian was followed by an E-ward migration of the foredeep basin depocentre and NE-directed thrusting in the eastern segment of the Outer Carpathians (Oszczypko, 1998, 2006). These rapid changes are thought to coincide with the migration of slab break-off across the Western Carpathians (Wortel and Spakman, 2000; Sperner *et al*., 2001). Lateral propagation of slab detachment caused an increase in slab pull forces and their progressive eastward migration as well as rapid changes in direction of contraction in the Outer Carpathians and an abrupt change from north-directed convergence to sinistral translation between the Inner and Outer Carpathians.

The Sarmatian stage of foredeep subsidence was accompanied by major normal faulting in its eastern segment. Here, faults step regularly down towards the orogen, suggestive of increasing deflection of the foreland lithosphere under the load of the accretionary wedge and the laterally already detached slab. A secondary reason for this step-like fault structure is the proximity to the rigid, thick and deflection resisting lithosphere of the EEC. Large scale normal faulting (see Fig. 7B) intensifies further in the Ukrainian segment of the foredeep where the Carpathians parallel the SE-ward continuation of the Polish Trough (Sovchik and Vul, 1996). Whole crustal failure of the foreland lithosphere under slab-pull and Carpathian loading stresses probably involved the reactivation of pre-existing crustal discontinuities that marked the boundary between the rheologically contrasting lithosphere of the old and strong EEC and the younger and weaker Małopolska Massif.

POST-OROGENIC RELAXATION

Although the grabens in the Polish Lowlands continued to subside during the Late Miocene this is rather attributed to mechanical compaction of the thick lignite seams than to continued tectonic activity. With the Late Miocene final detachment of the subducted slab beneath the West Carpathians, slab-pull forces decayed, the thrust belt became inactive and the foreland lithosphere unflexed, as evidenced by uplift and erosion of the Carpathians and their foredeep basin (Cloetingh *et al*., 2006; Oszczypko, 2006).

Similar to the Eger Graben, volcanic activity accompanied the Pliocene–Pleistocene uplift of the Sudetes (Wilson and Downes, 1991; Ulrych *et al*., 1999; Birkenmajer *et al*., 2004). During the Late Miocene and Pliocene differential uplift of the marginal blocks of the Bohemian Massif involved transpressional reactivation of pre-existing crustal discontinuities in response to the build-up of compressional stresses at crustal levels, reflecting increasing coupling between the East Alpine orogenic wedge and its foreland. This is interpreted as resulting from the onset of northward subduction of Adriatic continental lithosphere under the Eastern Alps after the subducted continental European slab had been detached from the lithosphere around 20 Ma (Schmid *et al*., 2004; Kissling *et al*., 2006; Ziegler and Dèzes, 2007).

NEOTECTONIC AND RECENT COMPRESSION

Borehole breakout data (Jarosiński, 2005*a*) and limited seismic events indicate that the present-day stress field of Poland is characterized by N- to NNW-directed trajectories of the horizontal maximum compressional stress axes and a prevailing strike-slip stress regime (Jarosiński, 2006). When this stress field had developed is, however, uncertain. Neotectonic remobilization of the Carpathian foreland is dated as Pleistocene. Earliest structural evidences for a late event of increased compressional deformations come from the Pleistocene (*ca*. 0.4 Ma; Hałuszczak *et al*., 1995).

In the context of still on-going convergence of Africa–Arabia with Europe (Jiménez-Munt *et al*., 2003) and according to results of numerical modeling (Bada *et al*., 1998; Jarosiński *et al*., 2006), the present-day stress field in the Carpathian Foreland is first of all controlled by compressional stresses related to continuing indentation of the rigid Adriatic Block. These stresses propagate through the Pannonian region, across the Carpathian suture into the Polish Platform (Fig. 11D; Jarosiński, 2006). This suggests that following Middle Miocene slab detachment beneath the Western Carpathians and Late Miocene isostatic rebound of the foreland lithosphere, the IWCW became mechanically strongly coupled with the foreland. In the Pannonian Basin the build up of a compressional stress regime, driving late-stage inversion of tensional basins, commenced at the end of the Miocene and continued during the Pliocene to recent (Horváth and Tari, 1999; Fodor *et al.*, 2005).

On the other hand, NW–SE directed compressional stresses are exerted on the Carpathian Foreland by the Arctic–North Atlantic Ocean, in which sea-floor spreading had commenced around 54 Ma ago (Gölke and Coblentz, 1996). This ridge push compression governs stress direction in the Baltic area (Jarosiński, 2005*a*). Compressional stresses related to continued collisional interaction of the Alpine Orogen with the European foreland (Dèzes *et al*., 2004; Ziegler and Dèzes, 2007), controlling accelerated Plio-Pleistocene subsidence of the North Sea–North German Basin (Cloetingh *et al*., 2008), hardly affect the Polish Lowlands. In the SW part of Poland stress directions vary considerably and range between NW–SE to NNE–SSW, with stress regimes varying between normal and strike-slip faulting (Jarosiński, 2006). Strain differences between Eastern and Western Poland are accommodated along the TTZ where stress rotations are common.

CONCLUSIONS

The Cenozoic evolution of the Polish Platform can be interpreted in terms of changing tectonic loading conditions along the margins of the European Platform. Three margin segments have been identified as being crucial for palaeostress conditions on the Polish Platform, namely the Carpathian and the East Alpine active margins and the Arctic–North Atlantic passive margin. The EEC played the role of a stable buttress during the Cenozoic evolution of the Polish Platform. Intermittent compressional and tensional stresses emanating from the Alpine and Carpathian collision zones in the south and south-west interfered with systematically growing ridge-push forces of the Arctic–North Atlantic sea-floor in the northwest and north. The stress distribution in the Carpathian Foreland was furthermore influenced by the complex tectonic structure of the TTZ and the high rheological contrast between the strong Precambrian EEC and the weak Palaeozoic platform. In this tectonic setting, western Poland was more influenced by stresses emanating from the Alpine active margin, while the eastern part of Poland was controlled by Carpathian-related loads. In the northern parts of the Polish Platform, stress directions were more sensitive to loads on the Arctic–North Atlantic passive margin.

With the Late Paleocene termination of compressional inversion of the Mid-Polish Swell, controlling stresses relaxed and the Carpathian Foreland was subjected to erosion. During the Eocene, minor uplift of the Mid-Polish Swell was accompanied by subsidence of its marginal troughs. Driving mechanisms controlling this suspected secondary inversion phase are, however, not yet understood. Although the geometry of the Mid-Polish depocentres and uplifted areas is suggestive of continued basin inversion in response to SW–NE directed compressional stresses, there is no evidence in the foreland of the Eastern Alps for an Eocene compressional stress regime, comparable to the one that had prevailed during the Paleocene.

By contrast to the foreland of the Eastern Alps, compressional stress built up in the Carpathian Foreland at the end of the Eocene. Reconfiguration of the Outer Carpathian flysch basins at the Eocene–Oligocene transition is interpreted as resulting from basement involving compressional deformations, reflecting increasing collisional coupling of the IWCW with the foreland lithosphere. At the same time subsidence patterns changed in the Polish Lowlands from being controlled during the Eocene by the NW–SE trending Mid-Polish Swell to Oligocene controls exerted by uplift of the W–E trending Meta-Carpathian Swell. Development of this swell is thought to reflect long-wavelength buckling of the lithosphere in response to the build-up of N-directed compressional intraplate stresses emanating from the Carpathian collision zone. Under this stress regime pre-existing crustal discontinuities on the Fore-Sudetic Monocline were transtensionally reactivated during the latest Eocene, controlling the subsidence of a system of narrow grabens.

At the end of Oligocene an extensional stress regime was established in the Carpathian Foreland, as evidenced by the subsidence of narrow grabens. This extensional phase, which persisted during the Lower and Middle Miocene, coincided with the subsidence of the flexural Carpathian Foredeep Basin and the main thrusting phase of the Outer Carpathians. These phenomena can be explained in terms of slab pull forces exerted by the subducting foreland lithospheric slab, which dipped beneath the IWCW. Rollback of this slab accompanied lateral extrusion of the ALCAPA Block from the Eastern Alps and its collisional interaction with the segment of the European margin that was occupied by the Carpathian flysch basin complex. Extension of the Carpathian Foreland was interrupted by the short early Sarmatian transpressional episode, which caused selective inversion of normal faults and grabens along the SW border of TTZ. This episode was coeval with the termination of N–S directed thrusting in the Outer Carpathians and eastward migration of the Carpathian Foredeep depocentre. Both phenomena are interpreted as an effect of slab detachment propagating during the Middle Miocene eastward across the Western Carpathians and related migration of the point, at which maximum slab-pull forces were exerted on the lithosphere. Early and Middle Miocene extension in the Carpathian Foreland paralleled by major thin-skinned contraction in the orogen, reflects mechanical decoupling of the IWCW from the subducting foreland lithosphere. In the Polish Lowlands the related extensional stresses exceeded apparently compressional stresses transmitted from the Arctic–North Atlantic passive margin and the Eastern Alps.

Upon Late Miocene final detachment of the subducted slab beneath the Western Carpathians, their isostatic uplift commenced and tensional stresses relaxed in the foreland. The present-day compressional stress regime of the Polish Platform results from continued counter-clockwise rotational indentation of the Adriatic Block combined with far-field Arctic–North Atlantic ridge push forces. Effective propagation of collision-related stresses into Central Europe reflects strong mechanical coupling of Inner West Carpathian and East Alpine orogenic wedges with the foreland lithosphere. The stress field of the Sudetes and Fore-Sudetic Monocline is characterized by the interference of compressional stresses emanating from the Eastern Alps and the Carpathians.

Cenozoic diachronic collision within the Alps and Carpathians caused strain partitioning between the western and eastern parts of Polish Lowlands. For example, Early Miocene compression in the Alpine Foreland, controlling uplift of the Bohemian Massif, was paralleled by extension in the Carpathian Foreland. The present-day compressional state on the Alpine, Carpathian and Arctic–North Atlantic margins of the European Platform causes puzzling effects of lateral stress changes and strain compensation between different tectonic units. From Paleocene to recent times stress conditions on the NW passive margin of Europe changed from rift-related extension to compression induced by ridge push. Systematic growth of ridge push forces through time played a secondary role in the deformation pattern of the Polish Platform.

In view of the complex evolution of plate boundary loads on the different margins of the European Platform, the conceptual model presented in this paper should be considered as a first attempt to elucidate the dynamics of the observed intracontinental deformations.

Acknowledgements. For inspiring discussions we are grateful to our colleges from Polish Geological Institute M. Piwocki, Z. Krysiak and J. Kasiński. This study was funded by Grant no. 2 P04D 05428 of the Polish Committee for Scientific Research. This paper is a contribution to the international TOPO-WECEP Project (see http://www. topo-wecep.eu).

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