Dynamics of the Polish and Eastern Slovakian parts of the Carpathian accretionary wedge: insights from palaeostress analyses

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Abstract: The arcuate Outer Carpathian accretionary wedge formed in front of the East Alpine–Carpathian–Pannonian (ALCAPA) megablock during the Eocene–Sarmatian. The wedge accreted sediments of the subducting remnants of the Carpathian Flysch Basin, a large oceanic tract left in front of the Alpine orogen. The palaeostress data for the orogenic hinterland (particularly the data related to the Early Miocene extension that was expanding towards the NE), combined with coeval subduction-related volcanism that was expanding towards the NE, indicate that the uneven roll-back of the subduction zone was the main mechanism controlling the development of the northern West Carpathian arc. The palaeostress data for the Tertiary accretionary wedge from the same time period are characterized by outward-fanning $\sigma_1$ trajectories that changed gradually during the wedge development. In contrast, the palaeostress data for the hinterland are characterized by preferred-directional stress events that changed abruptly during the wedge development. These palaeostress results are in accordance with the behaviour of the wedge and the hinterland, as the wedge behaved as a weak continuum and the hinterland behaved as a block mosaic with weak boundaries. The fault traces of the northern West Carpathian arc converge to both ends of the arc and suggest that the pre-existing basin was the factor that controlled the arc location. These fault trace patterns are asymmetric, indicating a slightly oblique overall convergence in a NE–SW direction. In accordance with this convergence, the palaeostress data for the accretionary wedge indicate that the western part of the wedge, which is characterized by NW–SE-oriented maximum principal compressional stress $\sigma_1$, was undergoing sinistral transpression. Meanwhile, the eastern part, which is characterized by NE–SW-oriented $\sigma_1$, was undergoing compression. Apparently, the dynamics of the accretionary wedge was further influenced by the shape of an elongated NE–SW-trending ALCAPA megablock, which was located behind the wedge and advanced in the direction of the general Early Miocene convergence during the most pronounced stages of the wedge development. This megablock served as the local indenter, as its strength surpassed that of the accretionary wedge located to its front. Further dynamic complexities were added because of the complex shape of the Magura Unit, which was located in the most proximal portion of the wedge and was stronger than the units in front of it. Wedge outcrops indicate that the large-scale shortening, which is characterized by the development of detachments and ramps, was preceded by an initial layer-parallel shortening. This is indicated by scaly fabrics and minor reverse faults that rotated into locked positions during the later accretion. Several outcrops with a wedge detachment fault indicate that there was a relatively low amount of friction during its development. The décollement zone is several hundred metres thick and shows evidence of transient fluid flow that was driven by pressure gradients. This is documented by frequent hydrofracturing, sandstone dykes and fibrous veins that opened against the weight of the whole wedge, all of which indicate cycles of higher pore fluid pressures that lowered the basal friction.

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Three major mechanisms have been proposed for the energy budget of orogens: (1) gravity spreading (e.g. Elliott 1976); (2) push from the rear (e.g. Chapl 1978; Davis et al. 1983; Stockmal 1983); (3) basal pull (Silver & Reed 1988; Barr & Dahlen 1989; Willett et al. 1993). Each mechanism generates a distinctive stress field, thereby implying different dynamics for the resultant orogenic belts, as documented by analogue material modelling (Macedo & Marshak 1999; Storti et al. 2000).

The field studies of arcuate orogens such as the Cretaceous–Quaternary South Carpathians, the Oligocene–Miocene Betic Chain, the Cretaceous–Palaeogene Sankiang–Yunnan–Malaya fold belt, the Miocene Mogok belt, the Cretaceous–Quaternary Hellenides and the Cretaceous–Quaternary Carpathian arc have indicated that certain combinations of major mechanisms can result in the development of arcuate orogenic belts. The development models of these belts vary between three distinct cases: (1) the oroclinal bending of the straight mountain belt (e.g. Ratschbacher et al. 1993a); (2) the gravitational spreading of the overthickened orogenic lithosphere (e.g. England 1983; Platt 1986; Dewey 1988; England & Houseman 1988; Coward 1994a); (3) the back-arc spreading of a thrust zone as a result of the uneven roll-back of a subduction zone (Angelier 1978; Burchfiel & Royden 1982).

The amount of curvature varies significantly between these arcuate orogenic belts (see Macedo & Marshak 1999), which probably indicates that not all of the three development cases can bring the orogen to its extreme curvature. The development case that results in extreme curvature can be found by studying an orogen that achieved extreme curvature, for example, the Carpathian arc in Central Europe, which has a curvature of 330° and was chosen as the study area for this paper.

If the Carpathian arc was developed by oroclinal bending, it should have recorded a stress field characterized by outward-fanning $\sigma_1$ trajectories, which is a typical result for a push from behind. This stress record should be present in both the orogen and its foreland. If the arc was developed by spreading of the overthickened lithosphere, it should have recorded a gravity-dominated stress field that controls polydirectional material flow into low-energy areas. This stress record should be present in the orogen but not in its foreland. Again, this stress record should be present in the orogen but not in its foreland. The goal of this paper is to identify one of these three development models in the Carpathian arc. However, a suitable method of reaching the goal must first be discussed.

Typically, the arcuate belts have been studied using incremental strain data from regions such as the Western Himalayas (Treloar & Coward 1991) and the Western Alps (Spencer 1992; Coward 1994b). Some belts have been studied using palaeomagnetic methods; for example, the Betic Chain (e.g. Allerton 1994; Platt et al. 1995) and the Appalachians (Gray & Stamatakis 1997). Others have been studied using analogue material modelling; for example, the Appalachians, the Apennines, the Argentinean Andes, the Carpathians, the Hellenides, the Lesser Antilles arc system, the Southern Alps of New Zealand and the Sulaiman Range (Koons 1990; Ratschbacher et al. 1991; Marshak et al. 1992; Macedo & Marshak 1999; Storti et al. 2000; Marques & Cobbold 2002).

The interpretation of incremental strain data from ductile terrains (e.g. Shackleton & Ries 1984; Shackleton 1986; Ellis & Watkinson 1987) often experienced difficulties with overprinting, missing units and genetic problems that could be solved only with a very detailed study. A more suitable approach for this study utilizes palaeostress techniques, which gave successful results in the case of the Taiwan thrust belt (Angelier et al. 1986; Barrier & Angelier 1986), in which the interpreted stress field, characterized by outward-fanning $\sigma_1$ trajectories, was attributed to a push from behind, as documented by analogue material modelling (Huchon et al. 1986).

Typically, the interpretation of palaeomagnetic data was used to distinguish between rotational and non-rotational trend lines in the structural architectures of orogenic belts (e.g. Gray & Stamatakis 1997), so as to decide whether or not arcuate belts started with curved thrusts, and thereby differentiate between the three development cases.

The analogue material modelling was generally used to test whether the interpretations of palaeomagnetic and structural data are physically sound.

The best method for reaching the goal of this paper is a combination of palaeostress, palaeomagnetic and analogue material modelling methods. This choice of methods further justifies the choice of the study area, which contains the northern West Carpathians of Poland and eastern Slovakia (Fig. 1), for several reasons. First, the analogue material model for this arcuate orogen has already been made by
Fig. 1. Map of the Outer West Carpathian accretionary wedge, between longitude E19° and E23° and with locations of new kinematic studies. Numbers in bold indicate outcrops with polyphase structural data. The remaining outcrops have monophase data. More detailed descriptions of outcrop data can be found at http://www.geolsoc.org.uk/sup18268. A hard copy can be obtained from the Society Library. PS, Przemsyl Sigmoid. The inset shows the Carpathians. AFB, autochthonous Miocene molasse sediments of the foreland basin; OC, Outer Carpathians; IC, Inner Carpathians; EC, Eastern Carpathians; PKB, Pieniny Klippen Belt; A, Alps; PB, Pannonian Basin. The inset also shows the location of megablocks in the Inner Carpathians–Pannonian Basin. It should be noted that internal and external boundaries of the Krosno–Moldavian and Rhenodanubian–Magura flysch nappes converge at both ends of the West Carpathians located between the Alps and the East Carpathians.
Fig. 2. Palaeostress directions in the Carpathian–Pannonian region gathered from existing literature for:
(a) Eocene–Early Oligocene; (b) Late Oligocene–Early Miocene; (c) early Badenian; (d) middle Badenian;
(e) late Badenian–Sarmatian; (f) early Pannonian; (g) late Pannonian; (h) present. The directions in
(a)–(g) represent the average values of stress states calculated by stress inversion techniques in research papers,
or compiled in syntheses (Bada 1999; Fodor et al. 1999). The stress states were calculated in the Alps
(Decker & Jarník 1992; Decker et al. 1993, 1994; Kurz et al. 1993; Linzer et al. 1995, 1997; Nemes et al. 1995;
Nemes 1996; Peresson & Decker 1997a), in the broader Pannonian Basin region (Bergerat et al. 1983; Bergerat
1989; Csontos et al. 1991; Tari 1991; Csontos & Bergerat 1992; Fodor et al., 1990, 1995; Márton & Fodor
1995; Benkovics 1997; Bada 1999; Gýörfi et al. 1999), in the West Carpathians (Kováč et al. 1989a, 1990, 1993, 1995;
Nemčok et al. 1989, 1989a,b, 2000c; Fodor et al. 1990, 1995; Marko et al. 1990, 1991; Nemčok & Lexa 1990;
Nemčok & Nemčok 1990, 1994; Nemčok 1991, 1993; Kováč & Hók 1993; Ratschbacher et al. 1993b;
Vass et al. 1993; Fodor 1995; Sperner 1996; Sperner et al. 2002), in the East and South Carpathians
(Ratschbacher et al. 1993a; Hippolyte & Sandulescu 1996; Linzer 1996; Morley 1996; Moser & Frisch 1996;
Matenco et al. 1997; Zweigel 1997; Linzer et al. 1998) and in the Transylvanian Basin (Huismans et al. 1997).
Fig. 2. Continued
Fig. 2. Continued
The regional direction of compression is based on stress states calculated from thrust or strike-slip regimes and the regional direction of extension is taken from stress states calculated from normal fault regimes. The principal horizontal stress direction map for the present, (h), is modified from the World Stress Map Project (1997). Data come from earthquake focal mechanisms, overcoring and bore-hole breakout studies. The boundary conditions modelled by a finite-element approach (Bada 1999) that fit the data are shown. They indicate a fixed or slightly deforming Bohemian Massif, an active convergence in the Brasov region, a fixed Moesian Platform, an active deformation at the Dinaric front and a rotating Adriatic (Apulian) microplate.
M, Mecsek Mountains; R, Rechnitz metamorphic core complex; T, Transdanubian Central Range; TB, Transylvanian Basin; VB, Vienna Basin.
Fig. 2. Continued
Study area

The study area (Fig. 1) covers the western portion of the Cretaceous–Quaternary Carpathian arc, which has a NE–SW trend near Vienna, an east–west trend in northern Slovakia and Poland, a NW–SE to north–south trend in Ukraine and NE Romania, and an east–west trend in SW Romania. The earthquake focal mechanisms in eastern Romania indicate the last active segment of the subduction zone. The subducting slab is subvertical as it sinks under its own weight.

The Carpathians comprise several zones (Fig. 1): (1) the foreland basin with molasse sediments; (2) the Tertiary accretionary wedge, which is called the Outer Carpathians or the Flysch Belt; (3) the Pieniny Klippen Belt, which occupies the contact between the wedge and the orogenic hinterland; (4) the orogenic hinterland, which is formed by the Inner Carpathians and the Pannonian Basin and is characterized by the occurrence of subduction-related volcanic rocks and intense crustal stretching.

The Miocene molasse sediments were deposited on the Variscan basement of the East European Platform (e.g. Oszczypko & Słaczka 1985, 1989; Kovač et al. 1989b), which dips under the Tertiary accretionary wedge. The wedge accreted Lower Cretaceous–Lower Miocene sediments (e.g. Poprawa & Nemčok 1989). It is traditionally divided into outer units, such as the Skole, the Subsilesian, the Silesian and the Dukla Units, comprising the Krosno–Moldavian flysch nappes, and the internal units of the Rhodanubian–Magura flysch nappes. An example of these internal units would be the Magura Unit, which has a different development history from the outer units, including a different detachment horizon, different timing of shortening and stronger mechanic stratigraphy. The Pieniny Klippen Belt (e.g. Andrusov et al. 1973) contains Triassic–Oligocene rocks as well as remnants of ophiolites. Apart from volcanic rocks (Pécskay et al. 1995), the orogenic hinterland is represented by Palaeogene, Neogene and Quaternary basins that have been superimposed on the older Inner Carpathians (e.g. Csontos et al. 1992). When the fill of these basins is stripped off (Fig. 1), the orogenic hinterland can be divided into two megablocks; the East Alpine–Carpathian–Pannonian (ALCAPA) megablock and the Tisza–Dacia megablock, which are characterized by separate movements during the development of the Carpathian arc (e.g. Csontos 1995).

Applied methods

Structural data (Fig. 1), including meso-scale folds, faults and extensional veins, have been used to gain a better understanding of the displacement history of each thrust sheet of the accretionary wedge, as well as to calculate palaeostress states. To exclude local complexities, such as local stress perturbations in fault zones, kinematic data were recorded away from the large-scale fault zones.

The stress calculation from meso-scale kinematic data was made using the computer programs of Sperner et al. (1993), Hardcastle & Hills (1991), Nemčok & Lisle (1995) and Nemčok et al. (1999a). The first program is based on the study by Turner (1953), who constructed compression (P) and extension (T) axes lying in the plane, comprising the fault-plane normal and striation vectors. The P axis is inclined at 45° to the fault plane and the T axis is perpendicular to the P axis. This method was used for datasets including fewer than four fault-striai readings and its broad-brush results are inferior to results calculated from well-populated datasets. The second to fourth programs were used for at least four faults with determined displacement sense. The second program was used only for monophase fault-striae data. These programs tested a large variety of stress configurations against the fault population from each location. Each stress tensor caused certain reactivation of faults (i.e. it resulted in calculated displacement vectors). Calculated displacement vectors were compared with the measured striations after each test cycle. The stress tensor compatible with the fault population was accepted as the stress configuration that was responsible for their activity. It has the form of a reduced stress tensor (Angelier 1989) and provides orientations of principal stresses and the ratio of their magnitudes.

Polyphase locations provided cross-cutting fault-striai data, allowing for the determination
of the relative sequence of tectonic events represented by calculated stress regimes. However, establishing a chronological sequence of stress regimes was difficult, because the accretionary wedge *sensu stricto* is deeply eroded and its outcrops range in stratigraphy from Valanginian to Oligocene, transitioning to Lower Miocene, as compared with the Eocene–Sarmatian development of the wedge (e.g. Nemčok et al. 2000c). A lack of Eggenburgian–Sarmatian outcrops, with the exception of upper Badenian location 106, proved to be the main obstacle in turning the relative stress regime sequence into a chronological one.

The palaeomagnetic declination data were taken from available literature (Koráb et al. 1981; Krs et al. 1982, 1991) and the measurements were made at nine locations in cooperation with E. Mártón (e.g. Mártón 1997; Mártón et al. 1997, 1999, 2000; Nemčok et al. 2000b; Zuchiewicz et al. 2000). New sampling was carried out to verify the quality of the previous work and to focus on missing key locations and stratigraphies. To exclude local complexities, such as complex palaeomagnetic signals, only massive shale and marl layers were used for palaeomagnetic sampling. The palaeomagnetic data provided constraints on the body rotation, allowing for a distinction between the rotated fault-striae data controlled by unperturbed stress regimes and the unrotated fault-striae data controlled by perturbed stress regimes (see Nemčok 1993; Mártón & Fodor 1995).

**Dynamic data from the Carpathian hinterland**

This section describes the existing data on the dynamics of the Inner Carpathians, which are located behind the Tertiary accretionary wedge (see Fig. 1 for location). Local geographical names are labelled in Figures 1 and 2. These data were taken from studies referred to in Figure 2, located on a set of maps in Figure 2 and synthesized in the following text.

The determined Late Eocene (40 Ma) stress field (Fig. 2) of the northwestern side of ALCAPA has a NW–SE compression and extension, respectively (e.g. Marko et al. 1990, 1991; Kováč & Hók 1993; Nemčok & Nemčok 1994; Fodor 1995). Syndepositional palaeostresses from the Priabonian sediments of central ALCAPA have a NW–SE and NE–SW compression and extension, respectively (e.g. Fodor et al. 1992, 1994; Kováč & Hók 1993; Bada 1999). Roughly, the east–west-oriented compression was determined from the Upper Eocene sediments in the southern part of ALCAPA (Fodor et al. 1999). Stress configurations derived from the northeastern part of ALCAPA have a north–south to NE–SW compression (Nemčok & Nemčok 1994). Determined stress tensors from the East Carpathians, the South Carpathians and the Transylvanian Basin are scarce and different workers have produced contradictory results for several reasons. First, the area contains the oldest portion of the stratigraphic section, which requires the largest amount of filtering to be done out of younger data. Next, the area has an incomplete stratigraphic section and a lack of robust datasets. Lastly, there has been an inappropriate use of palaeostress programs that were designed only for monophase data in simple structural settings (e.g. Angelier & Goguel 1979; Angelier et al. 1982; Sperner et al. 1993). Results include NE–SW (Moser & Frisch 1996; Matenco et al. 1997; Linzer et al. 1998) and east–west compressions in the South Carpathians, ENE–NE–directed compression in the northern Transylvanian Basin (Gyorfi et al. 1999) and north–south extension in the Transylvanian Basin (Huismans et al. 1997).

The determined Oligocene (35–30 Ma) stress field (Fig. 2) of the northwestern side of ALCAPA has a NW–SE and a NE–SW compression and extension, respectively (e.g. Marko et al. 1990, 1991; Kováč & Hók 1993; Nemčok & Nemčok 1994; Fodor 1995). Roughly, east–west compression was inferred from the dextral offset of the Oligocene basins along the southeastern side of ALCAPA (Csontos et al. 1992). A NW–SE- and NE–SW-oriented compression and extension was calculated (Fodor et al. 1992, 1994; Kováč & Hók 1993; Bada 1999) from the central parts of ALCAPA. Stress configurations from the northeastern parts of ALCAPA have a north–south to NE–SW compression (Nemčok & Nemčok 1994). Stress tensors from the South Carpathians have NE–SW compression (Moser & Frisch 1996; Linzer et al. 1998), a NW–SE first-order extension (Matenco et al. 1997) or east–west-oriented compression (Ratschbacher et al. 1993a). Stress tensors from the northern Transylvanian Basin have NNW–SSE compression (Gyorfi et al. 1999) or NW–SE compression (Huismans et al. 1997).

The reconstructed Early Miocene (22–20 Ma) stress field (Fig. 2) of the northern and central parts of ALCAPA has a NW–SE and NE–SW compression and extension, respectively (e.g. Kováč et al. 1989; Nemčok et al. 1989a; Fodor et al. 1990, 1992, 1994; Kováč & Hók 1993; Vass et al. 1993; Sperner 1996). Inferred and calculated stresses of the southeastern and northeastern sides of ALCAPA have a west–east and north–south to NE–SW compression, respectively (Csontos et al. 1992; Nemčok & Nemčok 2000).
basins from the dextral offset of the Early Miocene (Kováč et al. 1995). Stresses were inferred from the dextral offset of the Early Miocene basins by Csontos et al. (1992). In southern and central ALCAPA, the sinistral transpressive activity of faults, driven by NW–NNW-directed compression, finished by Otnangian times (Bada 1999). Stress tensors from the South Carpathians have a NE–SW to NNE–SSW compression (Moser & Frisch 1996; Linzer et al. 1998), a NW–SE first-order extension (Matenco et al. 1997) or an east–west-oriented compression (Ratschbacher et al. 1993a). Stress tensors from the northern Transylvanian Basin have NNW–SSE compression (Györfi et al. 1999) or NNE–SSW compression (Huismans et al. 1997).

Rift initiation inside the Pannonian Basin occurred in Eocene–Oligocene time (Pécskay et al. 1995; Horváth & Tari 1999). The Pannonian Basin region is characterized by NE–SW to ENE–WSW extension since late Oligocene times (e.g. Fodor 1995). Pull-apart basins opened in the intra-Carpathian region during the Karpatian–middle Badenian (17.5–15 Ma) (Fig. 2) and were controlled by a north–south and west–east compression and extension, respectively (e.g. Bergerat 1989; Nemčok et al. 1989; Csontos et al. 1991). The Rechnthor metamorphic core complex was uplifted in Karpatian times as a result of to major NE–SW and minor NW–SE extension dafced by dension tracks (Dunkl & Demény 1997), which drove low-angle normal faults (Tari et al. 1999). The Otnangian–Karpatian (19–16.5 Ma) regime, with NW–SE compression in the area of the Transdanubian Central Range, changed to the Badenian–Sarmatian (16.5–11.5 Ma) stress regime with north–south compression and the Pannonian (11.5–6.2 Ma) stress regime with NNE–SW compression (Csontos et al. 1991; Tari 1991). North–south compression changed to NE–SW compression in various West Carpathian blocks during the middle–late Badenian (15.5–13.6 Ma) (Nemčok et al. 1989; Fodor et al. 1990; Nemčok & Lexa 1990). This stress rotation finished in central ALCAPA during the Sarmatian (13.6–11.5 Ma) (Bada 1999). Otnangian–early Badenian sinistral transtension, driven by NW–SE compression and NE–SW extension, was reported from central ALCAPA (Márton & Fodor 1995). Karpatian–Badenian transtension driven by north–south compression and west–east extension was present in southern ALCAPA (Benkovics 1997). NW–SE compression was determined for flysch units along western ALCAPA, changing direction from NW to north, then to NE during the Karpatian–Badenian (e.g. Nemčok 1993; Fodor et al. 1995). It was calculated that the extension in ALCAPA advanced from its centre towards its margins during the Karpatian–middle Badenian (Nemčok et al. 1998a).

The northern East Carpathians were characterized by west to WNW compression and the southern East Carpathians by NW to NNW compression during the Early Miocene–Sarmatian (e.g. Zweigel 1997).

The Badenian–Pannonian period (Fig. 2) was the time of the intra-Carpathian basin and stress rearrangement, during which a very prominent west–east to WNW–ESE regional extension started occurring in different times and places during the end-Badenian–Sarmatian (e.g. Bergerat 1989; Fodor et al. 1990; Nemčok & Lexa 1990; Csontos et al. 1991; Bada 1999). A comparison of the calculated ratios of principal stress magnitudes of the progressively younger stresses from the Middle Slovakian volcanic region indicates a progression towards a \( \sigma_3 = \sigma_2 > \sigma_1 \) type of stress field (Nemčok et al. 2000a). Karpatian–early Badenian sinistral strike-slip faults in the Vienna Basin area were reactivated as normal faults during the Middle Miocene (e.g. Nemčok et al. 1989; Nemčok 1993; Fodor 1995).

The Middle Miocene stress field determined in the South Carpathians had a NW–SE compression (e.g. Ratschbacher et al. 1993a; Girbacea 1997; Matenco et al. 1997). The Middle Miocene compression in the East Carpathians varied between NE–SW and NW–SE orientation in the northern and southern parts, respectively (Linzer 1996; Girbacea 1997; Zweigel 1997). The Transylvanian Basin was controlled by east–west extension during the Sarmatian (Huismans et al. 1997).

The boundary between ALCAPA and Tisza–Dacia has been controlled by NE–SW compression and NW–SE extension since the late Badenian (Györfi et al. 1999).

From the end of the Badenian to the Pannonian (13.6–11.5 Ma) (Fig. 2), the intra-Carpathian basins experienced a regionally consistent east–west to ESE–WNW extension (e.g. Bergerat 1989; Nemčok 1993). The South Carpathians were characterized by north–south to NNE–SSW compression (e.g. Ratschbacher et al. 1993a) or NW–SE compression (Girbacea 1997).

 Later, during the Pannonian (11.5–6.2 Ma), the Carpathian stress configuration changed to a stress configuration similar to the present-day one (Fig. 2), as determined in the Vienna Basin, southern and northern Hungary, and central Slovakia (Fodor et al. 1990; Csontos et al. 1991; Benkovics 1997; Nemčok et al. 1998a). Since the early Pannonian, the time period was characterized by a short-term intra-Pannonian Basin inversion event and Pliocene–Quaternary
inversion (Bada 1999 and references therein). The timing of the short-term event is largely mismatched; for example: (1) Sarmatian–Pontian (9–5.3 Ma) for the entire region, which was driven by east–west compression (Peresson & Decker 1997b) and was proven by numerical modelling to be physically unrealistic (see Bada 1999); (2) early Pannonian for the Central Hungarian Fault Zone (which forms the southeastern ALCAPA boundary) or Mecsek Mts, which and was driven by north–south compression (Benkovics 1997; Csontos & Nagymarosy 1999); (3) late Pannonian for the Middle Slovakian basins, which was driven by north–south compression (Nemčok et al. 2000a).

The described calculated Pannonian stress regimes suffer from the sparsity of outcrops. The Post-Pannonian stress regimes inferred from 2D seismic data suffer from a lack of 3D interpretation.

Between the East and South Carpathians, an early Pannonian NW–SE compression later replaced by north–south compression was determined (e.g. Hippolyte & Sandulescu 1996).

To summarize this section, the Cenozoic dynamics of the Carpathian hinterland was characterized by distinct stress events, which frequently resulted from the interaction between microplates advancing behind the developing accretionary wedge.

Kinematic data from the accretionary wedge

This section describes the new data on the development of the Outer Carpathians, gathered from locations situated in the Tertiary accretionary wedge (see Fig. 1 for location). The data comprise folds and extensional veins, which are described later along with the $\sigma_1$ and $\sigma_3$ stress orientations derived from them. They also include fault-striae data from 169 locations shown in Figure 3. Figure 3a–e shows the fault-striae populations, where: (1) the whole dataset contains fewer than four fault-striae readings with similar controlling stress regimes, as determined by the method of Sperner et al. (1993); (2) the whole dataset suitable for other stress methods contains fewer than four outliers; (3) the whole dataset contains fault-striae readings related to several stress regimes. It was assumed that the first two categories of data were controlled by one tectonic event, the first category carrying just approximate value. Using fault categories described by Bott (1959), these data from monophase locations document that in 14% of the outcrops, the rocks were deformed predominantly by reverse faults, 19% were deformed by a combination of reverse and strike-slip faults, 14% were deformed by normal faults and the remaining 53% were deformed by strike-slip faults.

The polyphase locations were predominantly deformed by oblique- and strike-slip faulting. Figure 3f shows the fault-striae data from this dataset from selected outcrops, which were deformed by several deformational events.

The majority of minor faults in the Magura Unit (Fig. 1) and some faults in other units are lined by calcite fibres. These structures are formed by the steady growth of calcite during small increments of fault slip by a crack-and-seal mechanism (Ramsay 1980), which is slow enough to allow a diffusive mass transfer of these shear fibres by repeated fracturing, an infinitesimal amount of movement and calcite growth, infilling the rupture in the optical continuity with the fractured crystal. These fibres are generally subparallel to the existing striation, although we have found these fibres to be grown at an acute angle to the fault wall-rock in several instances.

The data from creek and road cuts display arrays of meso-scale faults with displacement ranging from a few centimetres to a few metres and show cross-cutting relationships, both among themselves and with striations on each fault plane, allowing the determination of the deformation history. Shear sense was determined from minor structures, including the lower-order shears (e.g. Petit 1987), shear fibres (e.g. Hancock 1985) and frictional structures (e.g. Dżulyski & Kotlarczyk 1965). Among the observed thrusts, both foreland-vergent and hinterland-vergent thrusts are present at outcrop. Older thrusts have formed in subhorizontally dipping sediments, thereby accommodating initial horizontal shortening. They have been subsequently folded and sometimes rotated to steep and locked positions. There are slickensides that have developed along the bedding planes. Some shale horizons contain a weak pressure solution cleavage. The largest proportion of internal slickensides of thrust sheets is formed by meso-scale strike-slip faults.

Gently dipping fault planes are included in the younger generation of thrusts from outcrops. These faults were not affected by any significant rotation. Part of this meso-scale fault population located inside the thrust sheets is subparallel to the thrust sheet defining large-scale thrusts and has the same vergence. The thrust sheets include large-scale folds and meso-scale thrusts developed inside the damage zones of thrust-sheet defining thrusts. Thrust-sheet bounding thrusts have displacements of the order of hundreds or thousands of metres.
Fig. 3. Great circle diagrams of all fault-striae populations (a–e) and selected polyphase populations (f) at outcrops located in Figure 1. The stresses controlling these populations are listed online at http://www.geolsoc.org.uk/sup18257. A hard copy can be obtained from the Society Library. Code above each diagram contains location number, stratigraphic age of deformed rocks, indication ‘poly’ in the case of polyphase fault-striae dataset, and type of the controlling stress regime (R, reverse faulting; N, normal faulting; S, strike-slip faulting; T, transpressional stress). The separated monophase datasets in (f) are numbered, including the location number and the number of the monophase dataset.
Fig. 3. Continued
Fig. 3. Continued
Fig. 3. Continued
Although décollements and major ramps are represented as discrete planes in balanced cross-sections through the study area (e.g. Roure et al. 1993; Roca et al. 1995; Nemčok et al. 2001), they contain zones of fault breccia or gouge that are several tens to hundreds of metres thick in studied outcrops. Each zone shows evidence of repeated cataclasis and calcite cementation. Some of these zones contain just a fault gouge, or a fault gouge with rounded or angular clasts. The Magura sole thrust zone is usually several hundred metres thick. Its breccia zone usually contains clasts representing footwall lithologies, whereas hanging-wall lithologies are present only in the uppermost part of the zone. Clasts and fault gouge contain multiple generations of veining with random orientations. These observations are consistent with a history of relatively large slip increments followed by stress build-up and cycles of fluid overpressure, prior to rupture (e.g. Sibson 1989, 1994; Knipe et al. 1991; Knipe 1993). Blocks of various sandstones, siltstones, cherts and marlstones are contained within a clay matrix, which is frequently intensely sheared and microfolded. The described large-displacement faults are usually accompanied by deformation zones on both sides, which mostly contain meso-scale faults and fractures. Outcrops indicate that more competent layers have accommodated folding through a flexural slip along their bedding planes. These surfaces are identified by striations, minor fault gouge seams or shear fibre development.

The earlier studies of the main Carpathian décollement recognized either subhorizontally oriented tensile fractures, filled by vertical mineral fibres grown against the weight of the whole thrust wedge (Nemčok et al. 1999b, 2000c), or randomly oriented and cross-cutting veins (Nemčok et al. 2000c), both indicating pore-fluid overpressure along the décollement. Another indicator of the pore-fluid overpressure in the accretionary wedge is the injection of sandstone dykes. The observed dykes are probably confined

Fig. 3. Continued
to the northeastern, thickest part of the accretionary wedge, as indicated by their presence at the outcrop of the Krosno beds of the Skole Unit at location 205, the outcrop of Zlin beds of the Magura Unit at location 404 and the outcrop of Menilitic beds of the Skole Unit at location 463 (see Fig. 1 for location). There are several vertical dykes of coarse sandstone within the massive sandstone at location 205. One of these dykes is about 30 cm thick and its top is preserved at this outcrop. Its fill contains carbonaceous plant fragments, which allow for the determination of the upward direction of flow inside the column. In the uppermost parts of the dyke, these pieces made a set of concave surfaces parallel to the dyke top. In the remaining part of the dyke, these surfaces are parallel to the dyke walls. The flow pattern inferred from these contours indicates that the largest flow rate was in the centre of the sandstone dyke.

**Principal stress orientations**

*σ3 orientations from extensional veins*

The distribution of calcite vein measurements indicates that 87% of the veins occur in well-cemented Magura facies (Fig. 4), whereas the remaining units did not undergo much veining. Most of the veins are of the stretched variety (Ramsay & Huber 1983) and were interpreted as having been opened roughly perpendicularly to their walls. Few of them contain fibres, allowing for the exact determination of the opening vectors. Veins in the Magura unit are located either along the main thrusts, where they were...
Fig. 4. Map of the Outer West Carpathian accretionary wedge from Figure 1, with σ directions determined from extensional veins. Further explanation is given in the text.
opened in all directions, or inside thrust sheets, where they have a more regionally consistent opening direction pattern. Nineteen percent of the latter veins were opened in the direction parallel to shortening: NNW–SSE to north–south in the west and NNE–SSW to NE–SW in the east, which is related either to flexure associated with folding or to the initial gravitational collapse of structures developed by shortening. Thirty-eight percent of the data indicate the orogen strike-parallel stretching in front of the Magura unit, ENE–WSW in the west and NW–SE in the east. This pattern seems to be rotated in the vicinity of the Przemyśl Sigmoid (Fig. 1). The extension directions, determined from the veins in regions of tectonic windows, always indicate that the extension was oblique to the direction of the local shortening, which indicates a strike-slip component of the deformation.

\[ \sigma_1 \text{ orientations from folds} \]

Measured meso-scale folds were formed by both fault-propagation and fault-bend folding. The frequent slickensides, formed along the bedding planes, indicate that the shortening vector was roughly perpendicular to the fold axes. This observation was used as justification for the rough determination of the maximum compressional stress \( \sigma_1 \) orientation as perpendicular to fold axes (Fig. 5). This rough determination needs to be used cautiously, because the assumption does not hold for transpressional folds or folds associated with strike-slip faults.

The overall system of interpreted \( \sigma_1 \) vectors indicates a fan pattern inside the accretionary wedge as well as in front of the thick Inner Carpathians, which were developed during pre-Tertiary time. On average, \( \sigma_1 \) vectors have NNW–SSE orientations along profile 1 from west to east in Figure 1, north–south orientations along profile 2, and NNE–SSW to NE–SW orientations along profiles 3–5. A relatively simple \( \sigma_1 \) pattern is present in front of the Magura Unit along profiles 3–5. Disturbances are present only in areas next to geometric complexities along the main thrusts. A more complex \( \sigma_1 \) pattern is determined from the Magura Unit, where complexities are located along major thrusts, bounding either the Magura Unit or its internal thrust sheets.

\[ \sigma_1 \text{ and } \sigma_3 \text{ orientations from faults} \]

The principal stress directions \( \sigma_1 \) and \( \sigma_3 \) calculated from reverse, strike-slip, oblique-slip and normal faulting data are shown in Figure 6. The complete stress directions with the ratio of their magnitudes, in cases where they can be calculated, are available online at http://www.geolsoc.org.uk/sup18257. A hard copy can be obtained from the Society Library. The age of deformed rocks at each outcrop is listed in Figure 3. The diagrams in Figure 3 indicate a 26% success rate in finding data at outcrops, which is 136 out of 525. A total of 110 outcrops were deformed by a monophase fault-striae dataset. To map the reliability of the data from ‘monophase outcrops’, it needs to be said that 18% include fewer than four readings and 30% include 4–6 readings. Fortunately, the distribution of locations with less numerous datasets (Fig. 3) is scattered, allowing locations with robust datasets to make up for locations with minor datasets in local areas. What is surprising is the small amount of outcrops deformed by polyphase fault-striae datasets, only 25 out of 136, and a mere five of them (Fig. 3f) have robust datasets. The small amount of polyphase outcrops is not a search artefact, because the amount of 525 visited outcrops where we could find some structural data is rather robust and already represents a selection from the total, which is several times larger than that. The rare occurrence of robust polyphase locations is further highlighted by the fact that out of five polyphase locations shown in Figure 3f, two of them contain separated monophase subsets, which contain fewer than four fault-striae readings.

The calculated stress states from outcrops can be roughly divided into four basic stress regimes: thrusting (12%), transpressional (14%), normal faulting (18%) and strike-slip faulting (56%) regimes (Fig. 3).

Both thrusting and transpressional stress regimes have been calculated from the outcrops, which are located in the close vicinity of largescale faults such as the contact between the Inner Carpathians with the Tertiary accretionary wedge, the Magura Unit sole thrust, the Silesian unit sole thrust, the Carpathian frontal thrust, faults bounding tectonic windows and faults at the contacts of major sub-units (compare Figs 1 and 3). These outcrops are homogeneously distributed over the study area.

Strike-slip stress regimes have been calculated from a majority of the outcrops. Apart from being located in the vicinity of several large-scale strike-slip faults, strike-slip faults represent the internal fault-striae data from thrust sheets (compare Figs 1 and 3). The number of outcrops with strike-slip stress regimes is largest in the west. It progressively decreases to the east, in accordance with the decreasing proportion of the well-cemented Magura unit length to the total wedge length in the cross-section through the Tertiary accretionary wedge.
Fig. 5. Map of the Outer West Carpathian accretionary wedge from Figure 1, with σi directions determined from folds. Further explanation is given in the text.
Normal faulting stress regimes have been calculated from the outcrops located in the most mountainous rear portion of the wedge, as well as at the frontal anticlines of main units and at the wedge front (compare Figs 1 and 3).

The outcrops with data relating to several successive stress regimes are located inside the Pieniny Klippen Belt, which is located just behind the Tertiary accretionary wedge (compare Figs 1 and 3) in the very rear of the accretionary wedge, at the geometrical complexities of several major unit contacts and in front of the wedge. Most of the stress regimes are located either in the Pieniny Klippen Belt or in the most western part of the accretionary wedge, where the amount of partitioning between thrusting and strike-slip faulting is the largest in the study area, as known from pre-existing studies (e.g. Nemčok et al. 1998b).

\( \sigma_1 \) vectors associated with thrust, strike-slip and transpressional stress regimes show a fan-like pattern inside the accretionary wedge (Fig. 6), as in the case of the rough estimation from folds (Fig. 5). \( \sigma_1 \) directions progressively change from NWW–SSE in the west to NE–SW in the east. However, there are local exceptions to this pattern. For example, locations 354, 359 and 459 indicate about 40° of local clockwise stress rotation in the NW vicinity of the Przemyśl Sigmoid (Fig. 1). Further exceptions are locations 193, 365, 366, 367 and 416, which reside in the Pieniny Klippen Belt, a contact zone between the Inner Carpathians and the Tertiary accretionary wedge. These locations indicate multiple reactivations of this highly deformed zone by various kinds of stress events (Fig. 3f). Exceptions are also present in the far west, where calculated stress events indicate local stress perturbations along sinistral strike-slip faults zones, calculated at outcrops such as 20, 88, 116, 120, 133, 165 and 180 (compare Figs 1 and 3). Some of these zones display small areas of pure thrusting, such as locations 5, 81 and 106, or have a local sinistral transpressional character, demonstrated by locations 17, 114, 315, 317 and 323 (compare Figs 1 and 3). As a result of the increased rotational deformation along these zones, 73% of the polyphase outcrops from the accretionary wedge reside in the west. Their stress regimes include multiple thrust, transpressional and strike-slip events, or a combination of strike-slip and transpressional events with extensional events (compare Figs 1 and 3). The sinistral body rotation in the west is also documented by older and younger stress regimes at location 108, characterized by NW–SE- and NNE–SSW-trending \( \sigma_1 \) vectors, respectively. The NW–SE vectors were calculated from rotated fault-striae data.

\( \sigma_1 \) vectors associated with normal-faulting stress regimes have been determined from 26 outcrops. With the exception of the NNE–SSW extension, \( \sigma_1 \) vectors are calculated from normal faults that occasionally post-dated the shortening, as indicated by the relative event succession from locations 38 and 106. The NNE–SSW extension is different in that it was calculated from outcrops 100, 147 and 148, which were located in the foreland deformed by Early Cretaceous rifting that pre-dates the accretionary wedge development. All observed normal faults were meso-scale. No evidence of large-scale normal fault propagation or reactivation of the synorogenic faults has been found.

The majority of the normal-faulting stress regimes are confined to the more proximal and mountainous parts of the accretionary wedge (Fig. 6), which formed large morphological gradients after the shortening. There are, however, eight exceptions. One such exception is location 106, which is formed by a large recumbent fold located above the detachment formed in the incompetent upper Badenian shale and evaporites. This location may represent a case where the syncontractional extension reacts to the friction reduction along the detachment and lack of lateral constraints, as known from Algeria and the Northern Apennines (Meyer et al. 1990; Bonini et al. 2002).

To summarize this section, the Cenozoic dynamics of the Tertiary accretionary wedge was characterized by a long-lasting orogenic stress regime, which underwent gradual changes in the western portion of the study area. It was subsequently replaced by a weak extension.

**Palaeomagnetic data**

The successful sampling of massive shale or marl from thrust sheets was a result of frequent disintegration of samples along the scaly fabrics. The sites at which the collection of samples was possible are shown in Figure 7 and listed in Table 1. Four to 12 samples were initially taken from 14 locations, but samples from five locations failed in thermal demagnetization, as they had too strong a present magnetic signal and the matching test had too large a scatter of results (E. Márton, pers. com.). The most reliable results came from locations PI 47–57, PI 64–71 and PI 72–83 (where the numbers indicate the number of samples taken at each location). Locations PI 8–19, PI 20–27, PI 40–46, PI 58–63, PI 84–90 and PI 100–108 would require more samples for precisely determined declination and inclination. These results can only be regarded as indications.
The most reliable results show that the mass rotation in the western part of the Magura Unit was up to 90° counterclockwise since the Late Eocene, which is in accordance with the results from the Central Carpathian Palaeogene (Podhale) Basin and earlier studies in the orogenic hinterland (e.g. Túnyi & Gross 1995; Márton & Márton 1996; Márton et al. 1996, 1999; Túnyi & Márton 1996; Grabowski & Nemčok 1999, and references therein), but is larger than the results (60°) from the Palaeocene–Middle Eocene facies of the Magura Unit, further to the east (Krs et al. 1991).

The rock mass in front of the Magura Unit rotated counterclockwise in the western part, with the declination value becoming progressively smaller from the Cretaceous (43–64°) to the Neogene (20°), as indicated by the present data in conjunction with the data of Krs et al. (1982), which came from the west area of our region. The combination of these data with the data of Koráb et al. (1981), located in the easternmost Silesian Unit of our region, further indicates a progressive decrease in the counterclockwise rotation of the declination along the orogen strike, from west (64°) to east (25°), for the same stratigraphy for the units in front of the Magura Unit.

It needs to be emphasized that the interpreted rotations have only preliminary character, because the quality of the early determinations inside the wedge (Koráb et al. 1981; Krs et al. 1982, 1991) may be questionable (E. Márton, pers. com.) and the new data require further precision and completion.

Discussion

The main goal of this discussion is to select the development model of the Carpathian arc from the three possibilities outlined in the introduction, as follows.

1) If the Carpathian arc was developed by orocinal bending, it should have recorded a stress field characterized by outward-fanning σ₁ trajectories, which is a typical result for a push from behind. This stress record should be present in both the orogen and its foreland.

Although the determined stress field in the accretionary wedge is characterized by a system of outward-fanning subhorizontal σ₂ trajectories (Figs 5 and 6), this stress type is not present in the orogenic hinterland (Fig. 2) and therefore characterizes a rather narrow frontal zone of the orogen. This stress field type is also absent in the orogenic foreland (see Bergerat 1987), which lacks the expected stress record typical for the region around the outer arc of the fold. Therefore, the orocinal bending model can be ruled out.

2) If the Carpathian arc was developed by a spreading of the overthickened lithosphere, it should have recorded a gravity-dominated stress field that controls the polydirectional material flow into low-energy areas. This stress record should be present in the orogen but not in the foreland.

The new stress results from the accretionary wedge do not indicate any distinct gravity collapse (Fig. 6), apart from a tenuous initiation in the most proximal part, indicating that the accretionary wedge formed a narrow compressional zone in front of the intensively stretched hinterland (Fig. 2). The studies of extension timing in the hinterland have demonstrated that the extension was progressively expanding to a more northeasterly region of the ALCAPA megablock during the Miocene (e.g. Nemčok et al. 1998a), which is in accordance with subduction roll-back in a similar direction. Furthermore, the roll-back is indicated by subduction-related volcanism that becomes progressively younger in the northeasterly direction inside the ALCAPA megablock (see Nemčok et al. 1998c, and references therein). Therefore, the gravity spreading model can be ruled out as well.

3) If the Carpathian arc was developed by back-arc spreading, it should have recorded a stress field characterized by a narrow zone of frontal compression and a wide zone of internal extension. This stress record should be present in the orogen but not in the foreland.

When the data for the hinterland and the accretionary wedge are compared (Figs 2 and 4–6) it is shown that the orogen was characterized by a relatively narrow frontal zone of compression and a broad internal zone of extension, especially during the time period after there was some northeastward expansion of the extensional conditions in the hinterland during the Miocene. The ALCAPA megablock was affected by subduction-related volcanism, which migrated towards the NE with time, along with the subduction roll-back. Perhaps the best indication that the subduction roll-back ran obliquely to the

Fig. 6. Map of the Outer West Carpathian accretionary wedge from Figure 1, with σ₁ and σ₂ directions determined from fault-striate data. Method S is the computer program of Sperner et al. (1993), method HH is from Hardcastle & Hills (1991); method NKL is from Nemčok & Lisle (1995) or Nemčok et al. (1999a). Further explanation is given in the text.
Fig. 7. Map of the Outer West Carpathian accretionary wedge from Figure 1, with determined palaeomagnetic declinations. The numeric codes at the locations indicate the number of analysed samples.
orogen strike and controlled the stress field in the Carpathians is stress changes in the regions behind the ceased subduction segment. These stress changes were reacting to neighbouring segments with subduction that was still active, and that was ceasing in an easterly direction along the Carpathian arc (see Nemčok 1993). All of the aforementioned points indicate that the back-arc spreading model is the best candidate for the Carpathian arc development.

However, several complexities in the dynamics and kinematics of the study area must still be discussed, as they do not precisely fit the dynamics and kinematics that are typical for the back-arc stretching development model. The first complexity is that the West Carpathians lack a prominent retrowedge. Analogue material modelling (Malavieille 1984; Storti et al. 2000) and numerical modelling (Willett et al. 1993) have shown that basal pull has a tendency to develop doubly vergent wedges, which are characterized by an initial high-velocity thrusting in the retrowedge and a high-frequency–low-displacement thrusting in the prowedge, followed by a low-frequency– high-displacement thrusting in the prowedge and a low-velocity thrusting in the retrowedge. Although the retrowedge has yet to be described as a significant structural element in the West Carpathians, recent studies have described a retrowedge located to the south of the Pieniny Klippen Belt in Slovakia (e.g. Sperner et al. 2002, and references therein). It is possible that the insignificant development of the retrowedge in the West Carpathians is caused by a slightly oblique convergence.

The strike-slip component of the displacement in the western region of the West Carpathian accretionary wedge (Fig. 1) is indicated by kinematic data for locations in the rear portion of the accretionary wedge (compare Figs 1 and 3). If paleomagnetic declination data are to indicate similar results, there must be a progressively lower amount of counterclockwise rotation in the wedge from west to east, as well as towards younger time periods. The available data for the western portion of the wedge indicate a post Late Eocene counterclockwise rotation of about 90° for the Magura Unit and about 64° for the units located in front of the Magura Unit. Over time, the 64° value decreases to 20°, which is a decrease characteristic of the

| Location | No./No. | D° | P | k | α95° | Dc° | Ic° | Cleaning | Dip | Rock age |
|----------|---------|----|---|---|------|------|-----|---------|-----|---------|
| Pl 1–7   | Large scatter | 030–60 | 22 | 12 | 008 | 35 | AF 220–350 | TH 350–420 | 167/32 | Early Eocene |
| Pl 8–19  | 08/12 | 014–26 | 22 | 14 | 014 | 26 | TH 250–400 | 162/60 | Early Eocene |
| Pl 20–27 | 06/08 | 315 | 70 | 18 | 18 | 68 | TH 250 | 064/18 | Oligocene |
| Pl 28–32 | Large scatter | 303–17 | 06 | 32 | 302 | 30 | TH 400–450 | 168/30 | Late Eocene |
| Pl 33–39 | Strong present field | 304 | 46 | 18 | 12 | 267 | TH 250–375 | 244/37 | Late Eocene |
| Pl 40–46 | 05/07 | 125 | 50 | 06 | 42 | 102 | TH 350–375 | 249/04 | Late Eoc–Olig |
| Pl 47–57 | 10/11 | 286 | 58 | 24 | 13 | 282 | TH 300–375 | 074/15 | Badenian |
| Pl 58–63 | 04/06 | 318 | 54 | 15 | 14 | 340 | TH 300 | 268/18 | Oligocene |
| Pl 64–71 | 07/08 | 318 | 54 | 15 | 14 | 340 | TH 300 | 268/18 | Oligocene |
| Pl 72–83 | 09/12 | 304 | 46 | 18 | 12 | 267 | TH 250–375 | 244/37 | Late Eocene |
| Pl 84–90 | 02/07 | 315 | 70 | 18 | 18 | 68 | TH 250 | 064/18 | Oligocene |
| Pl 91–95 | 09/12 | 154 | 30 | 41 | 12 | 164 | TH 300–400 | 352/25 | Early Eocene |
| Pl 96–99 | Failed | 153 | 32 | 15 | 15 | | |

No./No. indicates the ratio of used and collected samples. D° and P are the mean palaeomagnetic declination and inclination before tilt correction; Dc° and Ic° are the declination and inclination after tilt correction; α95° and k are the statistical parameters. Tilt corrections are made with the use of local bedding. Thermal demagnetization (TH) was used. AF, alternating field. Samples from locations Pl 1–7, 28–32, 33–39, 91–95 and 96–99 failed to provide results, either because of the large scatter of results, or because a strong magnetic field was present. The results with no statistical parameters or α95° > 20 must be regarded as mere indications. The location of samples is shown in Figure 7.
post early Badenian time, and then finally decreases to 0°, which is a decrease characteristic of the late Badenian time. The along-strike decrease of the rotation value in the wedge for the same stratigraphy is indicated by a post Late Eocene counterclockwise rotation of 25° in the east, instead of a rotation of 64° in the west. It must be emphasized that this interpretation of the palaeomagnetic data is far from conclusive. The quality of earlier results (Koráb et al. 1981; Krs et al. 1982; 1991) may be questionable (Márton, pers. com.) and the new data (Fig. 7, Table 1) are not robust enough to prove or disprove this trend. Therefore, for the results to be conclusive, more locations with palaeomagnetic declination data and further precision for locations Pl 8–19, Pl 20–27, Pl 40–46, Pl 58–63, Pl 84–90 and Pl 100–108 are needed.

The second complexity is that the determined stress field for the accretionary wedge (Figs 4–6) is similar to the development model dominated by push from behind. The determined stress field in the accretionary wedge is characterized by a system of outward-fanning subhorizontal σ1 trajectories, which point towards an indentation effect that must be discussed. Large-scale indentation can be ruled out, because the thrust fault traces in the arcuate accretionary wedge do not converge in the apex of the wedge (Fig. 1), as shown by the analogue material modelling for indenter-controlled orogenic salients (see Macedo & Marshak 1999), but rather they converge at salient end points instead. In fact, this type of convergence implies that there was pre-existing basin control in the location of the West Carpathian arc, which would require that the deep basin location coincides with the orogenic apex location. The existence of Early Cretaceous horsts and younger intra-basinal sources in the sedimentary record of the West Carpathian accretionary wedge (e.g. Książkiewicz 1960; Roure et al. 1993, 1994; Oszczypko & Oszczypko-Clowes 2002) and their absence in the sedimentary record of the East Carpathian accretionary wedge (e.g. Stefanescu & Melinte 1996) indicate the progressive deepening of the remnant Carpathian Flysch Basin from the NW to the SE (see also Ryłko & Adam 2005). This deepening is in accordance with the apex location being controlled by the deepest part of the basin.

In contrast, smaller-scale indentation is likely, because rather than being homogeneous, the orogenic hinterland is composed of megablocks (Fig. 1) that underwent separate movements during the development of the accretionary wedge (see Csontos et al. 1992). Their strength is superior to the strength of the wedge, because of the crustal thickness contrast between the hinterland and the accretionary wedge. This makes them good candidates for local indenters. These indenters are capable of affecting the deformation of the surrounding wedge, as suggested by the analogue material studies applied to the Argentinean Andes (Marques & Cobbold 2000).

It should be noted that palaeostress data indicate noticeably different dynamics for the accretionary wedge and the orogenic hinterland, whereas the dynamics of the hinterland can be described as a system of distinct tectonic events, characterized by relatively quick transitions and relatively consistent stress regimes typical for each event (Fig. 2), the accretionary wedge dynamics are characterized as though they were more stable than those of the hinterland and were undergoing slow gradual changes (Fig. 6).

It seems as though hinterland stress has been influenced by a complex stress transfer through a system of Inner Carpathian microplates (Fig. 1). The best example of this complexity is the Sarmatian docking of the ALCAPA microplate, which previously moved toward the NE, and the coeval accelerated eastward movement of the Tisza–Dacia microplate, which caused a roughly east-west-trending extension inside the ALCAPA microplate. The timing of the docking can be inferred from balanced cross-sections that are located in front of ALCAPA (e.g. Nemčok et al. 1999b, and references therein), and the timing of the accelerated Tisza–Dacia movement can be inferred from the timing and magnitude of the magnetic declination rotations (Panaiotu et al. 1998). The ALCAPA stress regime for this time period is shown in Figure 2.

A weak mechanical continuum seems to represent the accretionary wedge located between the foreland and the Inner Carpathians. As has been discussed above, the combined structural (Figs 2 and 4–6) and palaeomagnetic results (Fig. 7, Table 1) indicate that there is progressively less body rotation from west to east in the study area. The large scatter of the σ1 directions in the west (Fig. 6) can be attributed to the fact that the stresses have been calculated from the data at outcrops, which (1) underwent various amounts of sinistral body rotation and (2) were originally controlled by stress with a certain amount of σ1 deflection, depending on the proportion of strike-slip to thrust components in the western portion of the study area. The eastern portion, which lacks the strike-slip component, is characterized by a relatively consistent synorogenic σ1 pattern (Fig. 6).
The difference between the dynamics in the western and eastern portions of the accretionary wedge accounts for the third complexity, which is the highly asymmetric orogenic salient of the study area (Fig. 1). This complexity is a result of the NE–SW convergence, which is oblique to the strike of the northern West Carpathians. The asymmetry is best expressed by the internal fault patterns of the wedge, which are characterized by thrust faults that are parallel to the leading edge of the eastern limb of the arc that trends at a large angle to the northeasterly transport direction. The asymmetry is also expressed by the sinistral strike-slip fault systems that are parallel to the western limb of the arc that parallels the transport direction. A similar geometric situation is known from the Pennsylvania Salient of the Appalachians, as well as from the Sulaiman Range in Pakistan, both of which were generated by oblique convergence (see Macedo & Marshak 1999).

Conclusions

(1) Meso-structural data indicate an initial horizontal shortening, followed by a large-scale development of detachments, ramps and folds.

(2) The main faults are zones that are several tens to hundreds of metres thick, composed of fault breccia or gouge. Each zone shows evidence of repeated cataclasis and calcite cementation. Both the clasts and the fault gouge contain multiple generations of veining of random orientations, indicating that there were episodes of fluid overpressure prior to rupture. Fluid overpressure is indicated by subhorizontal veins with vertically grown mineral fibres along the frontal part of the main décollement and the presence of sandstone dykes. The dykes are confined to the northeastern, thicker portion of the wedge.

(3) Most of the veining inside the thrust sheets was developed in the now well-cemented sediments of the Magura Unit. A significant portion of the structural data indicates orogen-strike-parallel stretching rather than gravitational collapse.

(4) The arcuate Carpathian orogen is best explained by the back-arc stretching development model. This model best explains the coexistence of the narrow frontal zone of compression with the wide internal zone of extension expanding in a northeasterly direction. It also explains the northeasterly younging of subduction-related volcanism and the stress changes in the segments behind ceased subduction that reacted to northeasterly subduction roll-back in neighbouring segments with active subduction, which was ceasing in an easterly direction along the Carpathian arc.

(5) The stress patterns of the accretionary wedge and the orogenic hinterland differed from each other. The fan-shaped and gradually changing stress patterns in the wedge compared with the preferred-directional and abruptly changing stress patterns in the hinterland probably reflect their rheological differences. The hinterland was characterized by having a mosaic of blocks with weak boundaries, whereas the wedge behaved as a weak continuum.

(6) $\sigma_1$ trajectories that were calculated from fault-striae data in the accretionary wedge have an overall fan-shaped pattern, which indicates local indentation effects caused by the strength contrasts between the accretionary wedge and the blocks of the orogenic hinterland. A comparison of the thrust fault traces converging to the ends of the arcuate orogen with trace convergence patterns in analogue material models indicates that there was a pre-existing basin effect at the location of the Carpathian arc.

(7) The northern West Carpathian arc is asymmetric. The eastern part is characterized by thrust faults parallel to the leading edge of the arc in this region, and the western part is characterized by sinistral strike-slip fault systems that are parallel to the western limb of the arc, which indicates that the arc was formed by oblique NE–SW-trending convergence.

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