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Orogen-parallel extension in the Southern Carpathians

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Abstract

A structural study, including a kinematic analysis based on sense of shear criteria recorded in fault rocks, is combined with fission-track dating. A two-stage Alpine tectonic evolution is proposed for the major tectonic units which constitute the Southern Carpathians of the Parang mountains area. Upper Cretaceous top-to-the-SSE nappe stacking was followed by WSW–ENE orogen-parallel extension in the Eocene. Top-to-the-ENE shearing in lower greenschist-facies mylonites from the eastern part of the Danubian window (Parang mountains) is associated with a low-angle detachment at the base of the brittlely deformed Getic nappe (the Getic detachment). Below the dome-shaped Getic detachment, east-dipping at the eastern termination of the Danubian window, the Danubian units were rapidly exhumed. Hence, the eastern part of the Danubian window represents a greenschist-facies core complex. As a working hypothesis, it is proposed that this orogen-parallel stretch was originally N–S-oriented and that it formed when the Rhodopean fragment (which includes the Getic and Supragetic nappes) moved northward into an oceanic embayment, past the western margin of Moesia. This Eocene extension was part of a process of oroclinal bending in the area of the Southern Carpathians and their continuation into the Balkan mountains. Extension was followed by about 50º clockwise rotation of the Southern Carpathians, associated with dextral wrenching along the northern boundary of Moesia and compression in the Moldavides of the Eastern Carpathians.

Keywords: Carpathians; oroclinal bending; core complex; fission track dating; mylonites

1. Introduction

In eastern Europe, the Alpine mountain system bifurcates into the southeast-trending Dinaric–Hellenic branch and the great double-looped Carpathian branch. The latter rejoins the Dinarides–Hellenides and the Balkan mountains west and south of the Moesian platform (Fig. 1). The structuring of the Carpathian double loop is a long-lasting process. In the Romanian part of the Carpathians orogenic activity peaked during three major periods of convergence and/or collision: (1) the Meso–Cretaceous ('Austrian') phase, pre-dating post-tectonic cover sequences of Albian to Cenomanian age; (2) the Late Cretaceous ('Laramide') phase, pre-dating Maastrichtian to Palaeogene cover sequences; and (3) Miocene shortening, in particular during Burdigalian to Sarmatian (Serravallian) times (e.g. Burchfiel, 1980; Sandulescu, 1984, 1994). Only the Miocene

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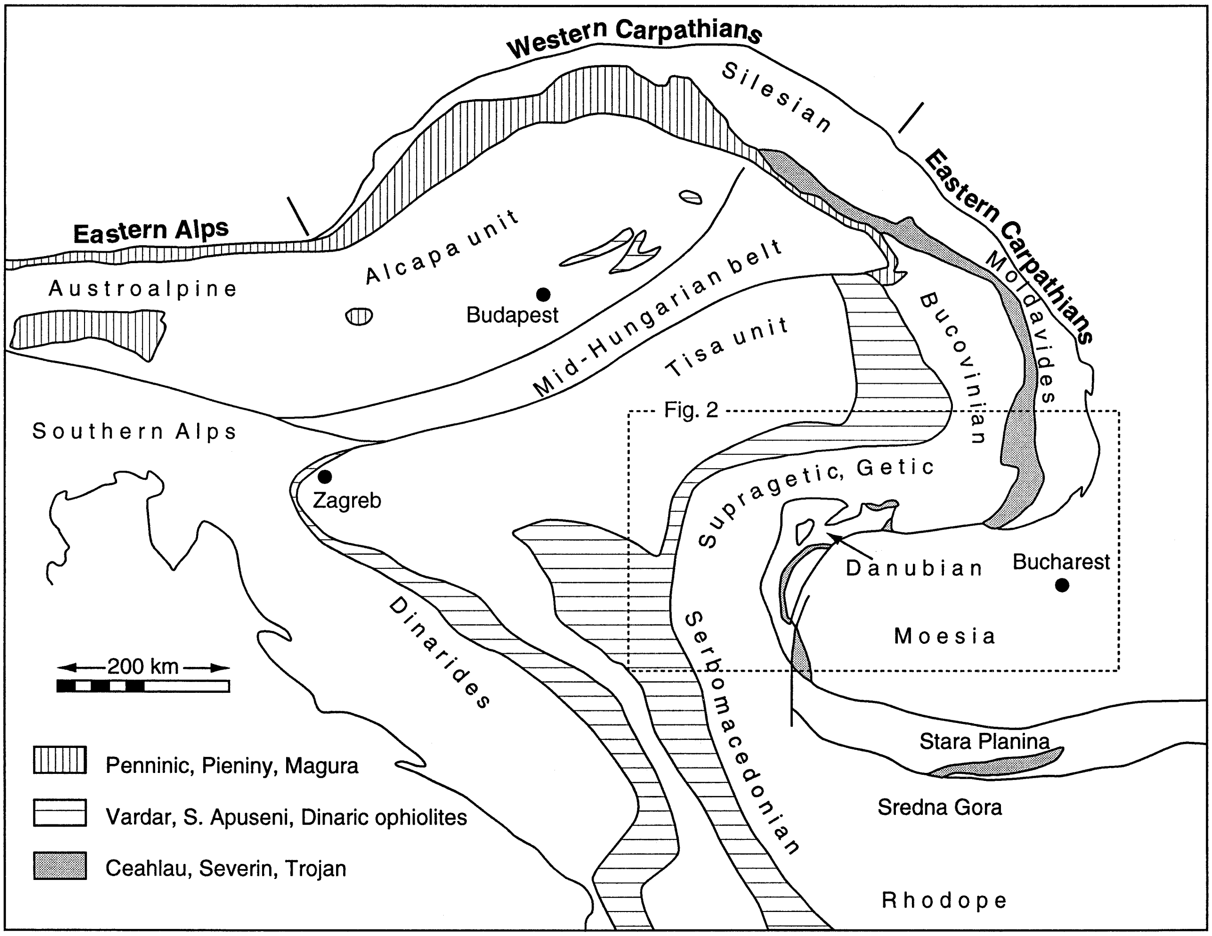


Fig. 1. Tectonic sketch map of the Carpathians, after Sandulescu (1984). Inset: outlines of Fig. 2.

thrust belt of the Silesian units and the Moldavides can be traced in continuity around the entire Carpathian loop. The more internal parts of the Carpathians, deformed during the Cretaceous, are of very heterogeneous origin and composition. The areas with Austroalpine affinities are commonly separated into two different blocks: the Alcapa unit and the Tisa unit. These two blocks underwent rotations of opposite sign during the Neogene and are separated from each other by the Miocene Mid-Hungarian belt, situated between the Balaton and Mid-Hungarian lines (Fig. 1; Balla, 1987; Csontos et al., 1991, 1992). It appears that most of the very significant amount of Palaeogene convergence between Apulia and Europe in the Central Alps (some 540 km according to Schmid et al., 1996) has been taken up by dextrally transpressive movements along the Dinaric–Hellenic branch since only minor amounts of internal shortening are reported for the Eastern and Southern Carpathians, respectively, during that time (Burchfiel, 1980; Morley, 1996).

It is now widely accepted that arc formation in the Western Carpathians and the northern part of the Eastern Carpathians, associated with shortening in the Silesian and Moldavian flysch belt, is primarily related to subduction roll-back of a remnant of oceanic lithosphere adjacent to the western margin of the East European continent, and contemporaneous back-arc extension in the Pannonian basin during the Miocene (e.g. Royden and Burchfiel, 1989). This roll-back propagated from southwest to northeast. It is presently at its terminal stage in the Vrancea area (e.g. Linzer, 1996) and possibly associated with the propagation of slab break-off towards the southeast (Sperner et al., 1996).

In contrast, both the process of arc formation and its timing are still poorly known for the loop of the Southern Carpathians, linking Carpathians and Balkan mountains around the western edge of the Moesian platform (Fig. 1). The Moesian platform was firmly welded to the East European craton since Early Cretaceous time across the Dobrogea mountains. Therefore it undoubtedly provoked an important corner effect during arc formation (Ratschbacher et al., 1993). Oroclinal bending (Figs. 1 and 2) of (1) the Danubian units, (2) various tectonic units which, according to Burchfiel (1980) constitute the Rhodopean fragment (Serbomacedonian massif, Supragetic, Getic and Bucovinian nappes), and (3) of the eastern branch of the Vardar ocean (South Apuseni mountains), is very pronounced. This oroclinal bending of the Southern Carpathians around Moesia asks for substantial dextral movements of these units relative to the Moesian platform during arc formation, as already proposed by Pavelescu and Nitu (1977). This wrenching was combined with clockwise rotations of the Tisa unit, evidenced by palaeomagnetic data (e.g. Patrascu et al., 1994).

It is not clear yet where the large dextral movements are located. The amount of Miocene dextral transpression in the Getic depression (the subsurface continuation of the Moldavides west of Ploiesti and south of the Southern Carpathian mountains, see Fig. 2) does not exceed a few tens of kilometres according to Matenco et al. (1997). Within the Southern Carpathians further to the north substantial dextral wrenching is only reported for Oligocene time (Cerna-Jiu fault, Fig. 2; Berza and Draganescu, 1988). The dissection of Cretaceous nappe boundaries and post-tectonic intramontane basins (Fig. 2), due to Miocene dextral wrenching, is very modest (some 10 km according to Ratschbacher et al. (1993) south of the Hateg basin; see Fig. 2).

This contribution will provide evidence for a hitherto undetected important orogen-parallel stretch which affects a previously formed stack of Danubian nappes, including the sole of the Getic nappe, near the eastern termination of the Danubian window. This presently WSW–ENE-oriented stretch is coeval with updoming of the Danubian window due to a minor amount of contemporaneous shortening perpendicular to the orogen. Based on new fission-track data this post-nappe straining is interpreted to be of Palaeogene age. This extension will be discussed in the context of oroclinal bending of the Southern Carpathians. It will be argued that arc formation in the Southern Carpathians, associated with dextral wrenching between the Southern Carpathians and Moesia, at least partly pre-dates Miocene bending in the Eastern and Western Carpathians.

2. The tectonic units of the Southern Carpathians

2.1. The Danubian nappes

The Danubian nappes represent the most external unit of the Southern Carpathians, apart from the subsurface continuation of the Moldavides in the Getic foredeep (Fig. 2). This pile of basement-cover nappes (see Berza et al., 1994; Berza and Draganescu, 1997, for a general overview) was probably peeled off the western part of the Moesian platform during the Late Cretaceous ('Laramide') orogeny. It is presently exposed in a window (the Danubian window), surrounded by higher tectonic units, or, onlapped by the Late Miocene post-tectonic cover of the Getic depression. The Danubian nappes axially plunge underneath the Getic nappes at the eastern termination of the window. The fact that they do not reappear in front of the Getic nappe in the Eastern Carpathians (Fig. 2) shows that the Danubian nappes are unique to the western part of the Southern Carpathians. However, they can be traced into the northernmost part of the Balkan mountains (Stara Planina; see Krautner, 1996), dissected from the Danubian window by the right-lateral Timok fault (Figs. 1 and 2).

The basement of the Danubian nappes (Berza and Draganescu, 1997) consists of Precambrian high-grade, partly migmatitic series, intruded by granitoids of Panafrican age. According to radiometric age determinations this basement was partly affected by Variscan deformation and metamorphism (Dallmeyer et al., 1996). A steep Variscan fault zone marks the limit between two distinct Precambrian terranes: the Lainici-Paius Group and the Dragsan Group. On the other hand, Variscan metamorphic

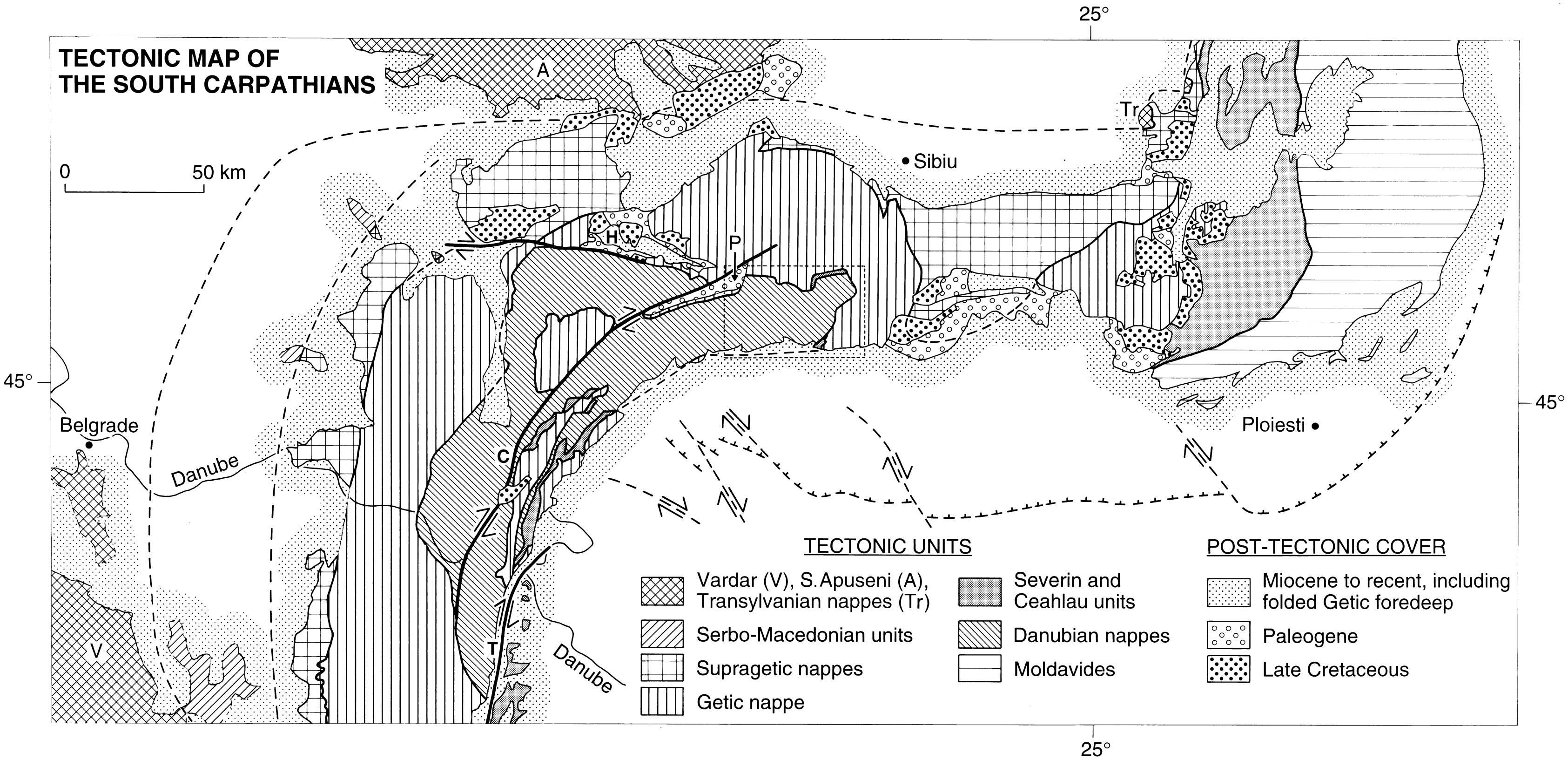


Fig. 2. Tectonic map of the Southern Carpathians, based on the Geological Map of Romania, 1 : 1,000,000 (Sandulescu et al., 1978) and on a tectonic map by Balintoni et al., in Berza et al. (1994). The structures in the Getic foredeep are after Matenco et al. (1997). *V* = Vardar zone; *A* = South Apuseni mountains; *Tr* = Transylvanian nappes; *T* = Timok fault; *C* = Cerna fault; *H* = Hateg basin; *P* = Petrosani basin.

overprint of Palaeozoic series preserved north of the Petrosani basin is low and does not exceed Alpine metamorphic grade. The Mesozoic cover, transgressive on basement or Permian redbeds, starts with Lower Jurassic Gresten facies (Schela formation and white arkoses), followed by Upper Jurassic to Lower Cretaceous platform carbonates (locally referred to as Oslea or Lupeni limestone, erroneously mapped as Palaeozoic, together with parts of the Schela formation, on some of the Romanian map sheets). Deeper marine deposits set in with pelagic limestones and marls of Albian to Turonian age (Nadanova formation), followed by poorly dated terrigenous flysch of Turonian to Senonian age. Continental margin sediments in a different facies are only preserved in the northwestern part of the Danubian window (Arjana nappe). Alpine lower greenschist-facies metamorphic overprint (chloritoid and pyrophyllite commonly occurring in the Schela formation) is pronounced in the northeastern part of the Danubian window, sub-greenschist-facies conditions prevailing in its southwestern part. In many places of the eastern part of the Danubian window Alpine mylonitization under lower greenschist-facies conditions is pervasive, as will be discussed later.

2.2. The Severin unit

The Severin unit tectonically overlies the Danubian nappes. It marks the most external oceanic suture zone of the Romanian Carpathians, suturing the Danubian nappes (and the Moesian platform) with the most external unit of the Rhodopean fragment, i.e. the Getic nappe. Ultramafic rocks, gabbros and basalts are locally preserved in melange zones (Savu et al., 1991). In most places the Severin unit is represented by sediments only, predominantly by Sinaia beds, an upper Tithonian to Hauterivian carbonaceous turbidite sequence with impure radiolarian cherts (Azuga beds) at its base.

In the Danubian window, the Severin unit is either very thin or, more commonly, completely missing at the sole of the Getic nappe. Yet, its lateral equivalents in the Eastern Carpathians, in particular the Ceahlau nappe, reach a considerable thickness. In the Eastern Carpathians, the tectonic contact between the Ceahlau nappe and the Getic nappe (or its equivalents) is sealed by Albian molasse and therefore considered to be of Meso–Cretaceous age. In analogy to the situation in the Eastern Carpathians, and because sedimentation in the Severin unit stopped during the Early Cretaceous, the contact between the Rhodopean fragment and the Severin unit in the Southern Carpathians is also considered to have formed during the 'Austrian' phase. However, the Severin unit cannot have overriden the Danubian nappe pile before the latest Cretaceous because Turonian to late Senonian flysch is present in the Danubian units. Sandulescu (1984) and Krautner (1996) parallelize the Sinaia beds with the Trojan flysch of the Balkan mountains.

2.3. Getic and Supragetic nappes

The Getic nappe (to be correlated with the Infrabucovinian nappe of the Eastern Carpathians according to Sandulescu, 1984) constitutes the lowest structural unit of the Rhodopean continental fragment. The base of the Getic nappe, as well as most of the remainder of this nappe, is made up of high-grade Proterozoic basement (Iancu and Maruntiu, 1994). Variscan reworking, including low-grade metamorphism, is more pronounced when compared to the basement of the Danubian nappes (Dallmeyer et al., 1996). A striking contrast with the Danubian nappes concerns the Alpine evolution and was already described in the pioneering work of Mrazec (1904) and Murgoci (1905). In general, the Getic basement shows no retrogression due to mylonitization of the pre-Alpine mineral assembly (in contrast to the predominantly mylonitic Danubian basement). Furthermore, the cover of the Getic nappes (Late Carboniferous to Cretaceous) completely lacks Alpine metamorphic overprint. The Getic nappe is overridden by various Supragetic nappes (e.g. Streckeisen, 1934), referred to as Sub-Bucovinian and Bucovinian in the Eastern Carpathians (Sandulescu, 1994). This nappe pile formed during the Meso–Cretaceous orogeny because sedimentation stopped in the Aptian. Post-tectonic basins, with Albian to Cenomanian deposits at their base, seal the nappe contacts (see 'Late Cretaceous post-tectonic cover' in Fig. 2). Structurally, Getic and Supragetic units define a pronounced structural depression south and southeast of Sibiu (Fig. 2), adjacent to the east-plunging structural dome of the eastern part of the Danubian window. In this area (east of the Olt valley) the Getic nappes are entirely covered by Supragetic nappes (Hann, 1995), the Getic nappes and the eastern continuation of the Severin unit (the Ceahlau nappe) reappearing further to the east, at the southern termination of the Eastern Carpathians.

2.4. The eastern branch of the Vardar zone

According to correlations by Sandulescu (1984, 1994) a part of the Vardar zone is also affected by oroclinal bending. According to this author, the South Apuseni mountains represent the eastern continuation of the eastern branch of the Vardar zone (Fig. 2). The Transylvanian nappes are obduction-type nappes overlying the Rhodopean fragment further to the east (Fig. 2). These nappes are derived from the eastern part of the South Apuseni ophiolitic zone (but issued from a separate, more easterly 'root zone'), presently situated in the subsurface of the Transylvanian basin according to Sandulescu (1994). Hence, the arc of the Southern Carpathians also affects these very internal units of the Carpathians, structured during the Meso–Cretaceous. However, in contrast to Sandulescu (1994), we do not consider the Vardar–South Apuseni ophiolitic branch to continue into the Pieniny ophiolitic belt. Both these ophiolite-bearing belts have different palaeogeographical and tectonic positions with respect to the Apulian microplate.

3. Kinematic analysis of fault rocks in the Parang mountains

Our preliminary study focusses on the kinematic analysis of fault rocks, in particular lower greenschist-facies mylonites of Alpine age, in the easternmost part of the Danubian window (Fig. 3): the Parang mountains east of the Jiu valley. Kinematic indicators (Simpson and Schmid, 1983), predominantly shear bands and asymmetric sigma clasts, are widespread and can often be detected already in the field. The number of stations with known sense of shear was increased by microscopic inspection of oriented specimens. Hence, other shear sense criteria, such as flattening of recrystallized quartz grains oblique to the mylonitic foliation, could be used. The sense of shear criteria, where available, generally were unambiguous, indicating strongly non-coaxial flow.

3.1. Top-SE-directed nappe stacking

In the Parang mountains (Fig. 3), from bottom to top, the four Danubian basement nappes are: the lower Danubian Schela and Lainici nappes, followed by the upper Danubian Urdele and Vidruta nappes (see Berza and Draganescu, 1997). A generally thin veneer of Mesozoic cover is often preserved near nappe contacts. Viewed in a N–S section this nappe stack defines a domal structure (Fig. 4, top), also affecting the base of the Getic nappe, including occasionally preserved remnants of the Severin unit. Occurrences of upper Danubian nappe units, separated from the lower Danubian nappe units by the main Intradanubian thrust (Berza et al., 1994), are restricted to the northern half of the dome. In the south, the Getic nappe, including the Severin unit, directly overlies the lower Danubian Lainici nappe. This geometry, although dissected by numerous post-nappe faults (in particular the Cerna-Jiu strike-slip fault, Berza and Draganescu, 1988) and overprinted by orogen-parallel extension described below, is valid for the entire area of the Danubian window north of the Danube. Upper Danubian units are systematically preserved near the northern and western margin of the Danubian window and invariably missing below the numerous klippen of the Getic nappe near the Getic depression. This geometry was interpreted in terms of a duplex structure between a roof thrust at the base of the Getic nappe and an unexposed floor thrust at the base of the Danubian nappe stack, indicating S- to SE-directed transport for Late Cretaceous nappe stacking in the Southern Carpathians (Seghedi and Berza, 1994).

Interestingly, our systematic study of lineation orientation and sense of displacement in mylonitized basement and cover of the Danubian nappes revealed top SE- to SSE-directed movement in very few places only (Fig. 5a). These occurrences are restricted to the central and southern parts of the outcropping area of the Schela nappe (localities with top-SE to SSE movements indicated in Fig. 3), i.e. in an area which is structurally farthest away from the base of the Getic nappe.

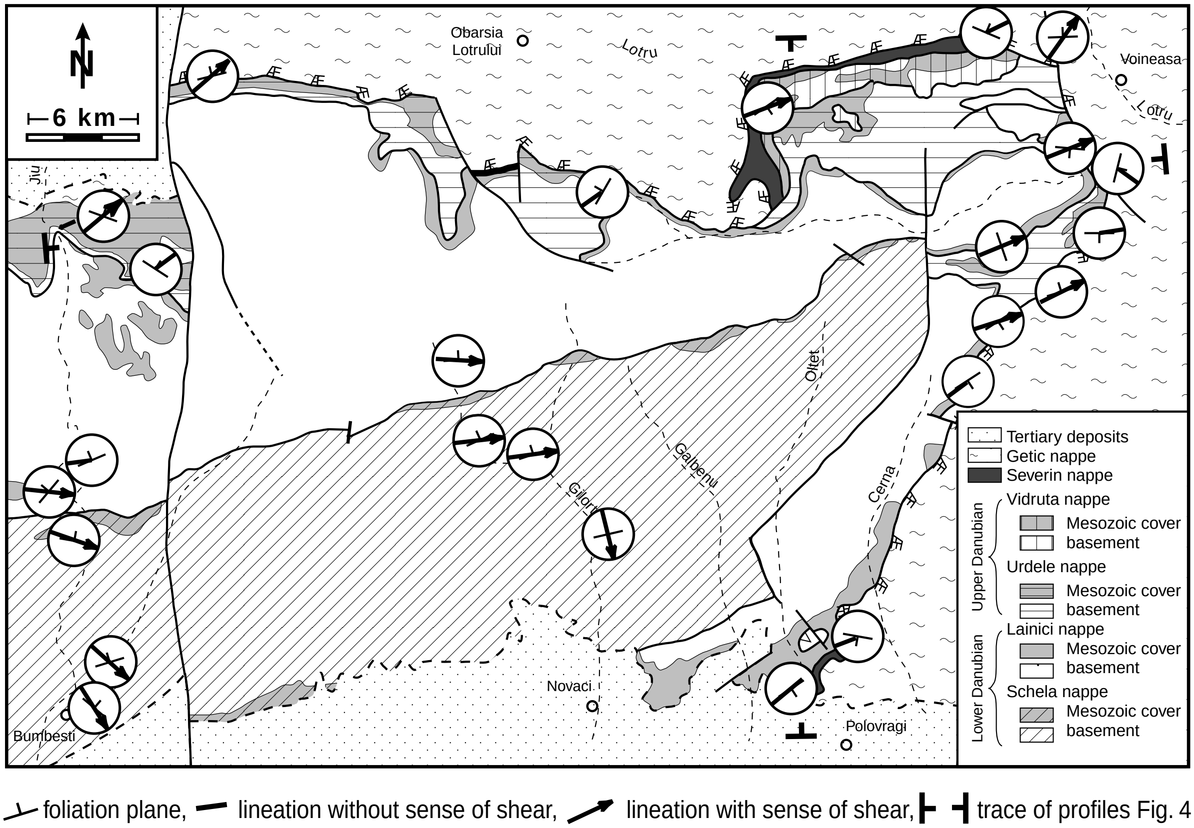


Fig. 3. Tectonic map of the Parang mountains, modified after Berza et al., in Berza and Iancu (1994). The map also indicates the average of the orientation of mylonite foliation and lineation at a number of stations. Lineations associated with a known sense of displacement are marked with an arrow indicating relative sense of movement of the hangingwall in respect to the footwall.

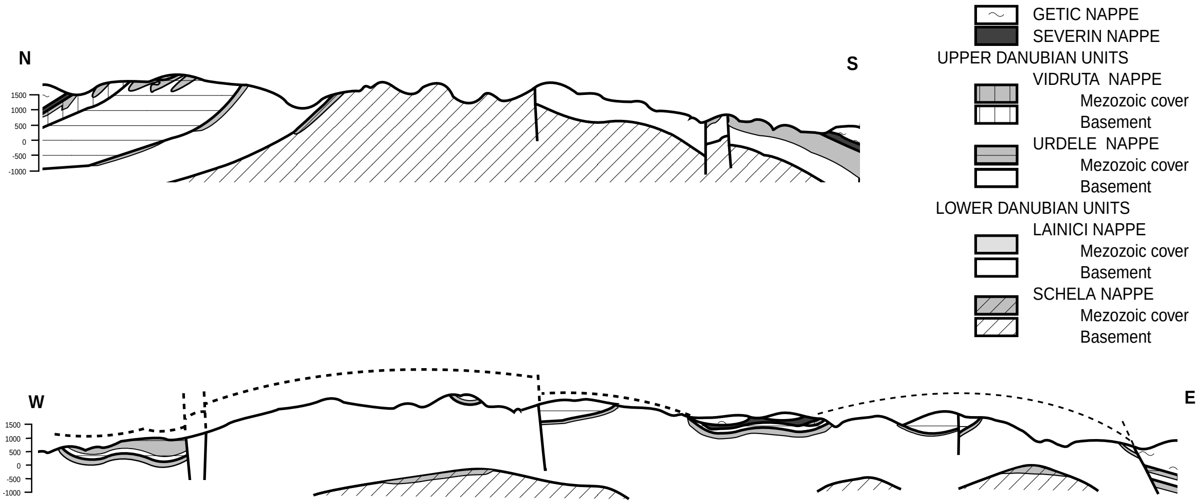


Fig. 4. Schematic N–S (top) and E–W (bottom) sections through the Parang mountains, traces of profiles indicated in Fig. 3.

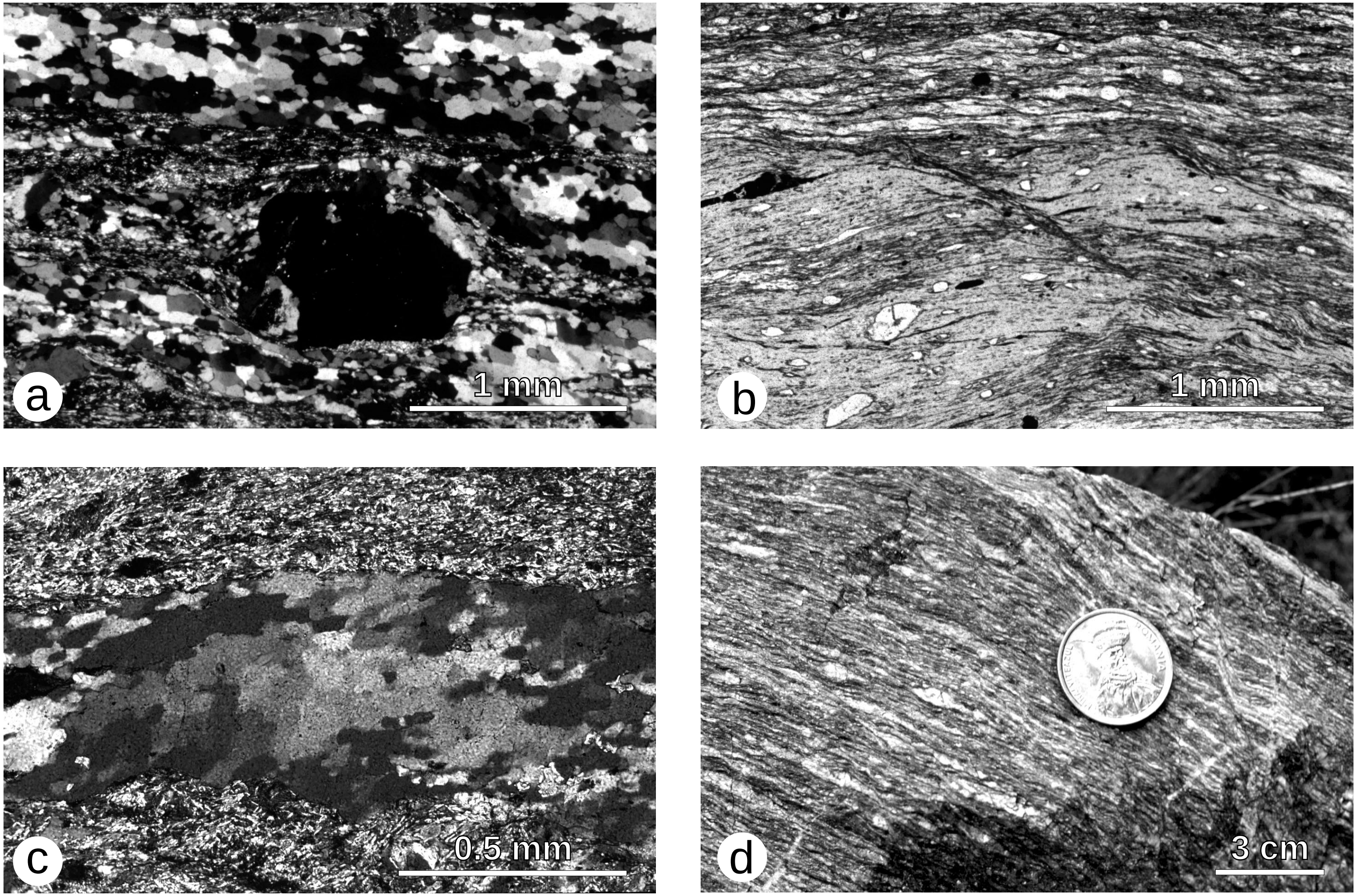


Fig. 5. Sense of shear indicators in lower greenschist-facies mylonites from the Danubian nappes. (a) Sigma clast indicating sinistral (top SE) shearing in a basement mylonite of the Schela nappe, Gilort valley. (b) Shear bands indicating dextral (top ENE) shearing in a mylonite of the Lainici nappe, immediately underneath mylonitized Lupeni limestone and about 50 m below the cataclastically deformed base of the Getic nappe (compare with Fig. 6a), uppermost Cerna valley at the eastern margin of the Danubian window. (c) Dynamic recrystallization of quartz by subgrain rotation, associated with grain flattening oblique to the macroscopically visible foliation, indicating dextral (top ENE) shearing in a basement mylonite of the Lainici nappe near Voineasa at the NE corner of the Danubian window, at a few hundred metres from the base of the Getic nappe. (d) Outcrop of dextrally (top ENE) sheared basement of the Urdele nappe with sigma clasts. Note that ductile shearing was followed by cataclastic faulting, indicating continued top-ENE sense of movement (Riedel shear immediately adjacent to the coin) at significantly lower temperatures. Northern rim of the Danubian window, 7 km east of Petrosani, Jiet valley.

3.2. Pervasive mylonitization associated with a top-NE sense of movement

Apart from the few occurrences mentioned above, all mylonitic stretching lineations found in the area of Fig. 3 are E–W to NE–SW oriented, with a mean orientation around azimuth 64 (Fig. 6). Sense-of-shear criteria invariably indicate top ENE-directed movement (Fig. 5 b–d). Direct overprinting relationships between the two sets of lineations are only locally found. In general, one set of lineations prevails at any given locality, hence no relics of the SE- to SSE-oriented lineation, taken to be older for reasons outlined below, are preserved over most of the area investigated. Interestingly, mylonitic horizons associated with ENE-oriented lineations are pervasive throughout the entire exposed part of the nappe stack, except for the deepest structural levels.

Fig. 3 documents that the orientation of the lineations does not vary systematically around the eastern termination of the Danubian window, although the mylonitic foliation containing the stretching lineations considerably varies in orientation. The foliation is south- to southeast-dipping in the southeast and generally north-dipping in the north, in accordance with the dome structure of the window. In Fig. 6 the foliation poles of all the mylonitic rocks are seen to scatter along a great circle whose best-fit axis almost coincides with the orientation of the stretching lineations, the senses of shear (Figs. 3 and 5b–d) remaining top-to-the-ENE all around the window. In map view, this implies a component of dextral and sinistral movement between the core of the Danubian window and the overlying Getic nappe at the northern and southern margin of the Danubian window, respectively. At the eastern edge of the window, the mylonitic foliation often exhibits an easterly dip direction (Fig. 3), senses of shear indicating downfaulting of the Getic nappe. There, a late east-dipping cataclastic normal fault cuts through the more gently east-dipping mylonites (E–W profile of Fig. 4). However, except for the eastern edge of the window, the nappe stack appears sub-horizontal over the entire length in the E–W section of Fig. 4.

Mylonites associated with ENE-oriented lineations are not restricted to nappe contacts, as would be expected if this mylonitization, associated with a strong retrogressive metamorphic overprint (in respect to pre-Alpine metamorphism), would only be associated with nappe stacking. Except in a few

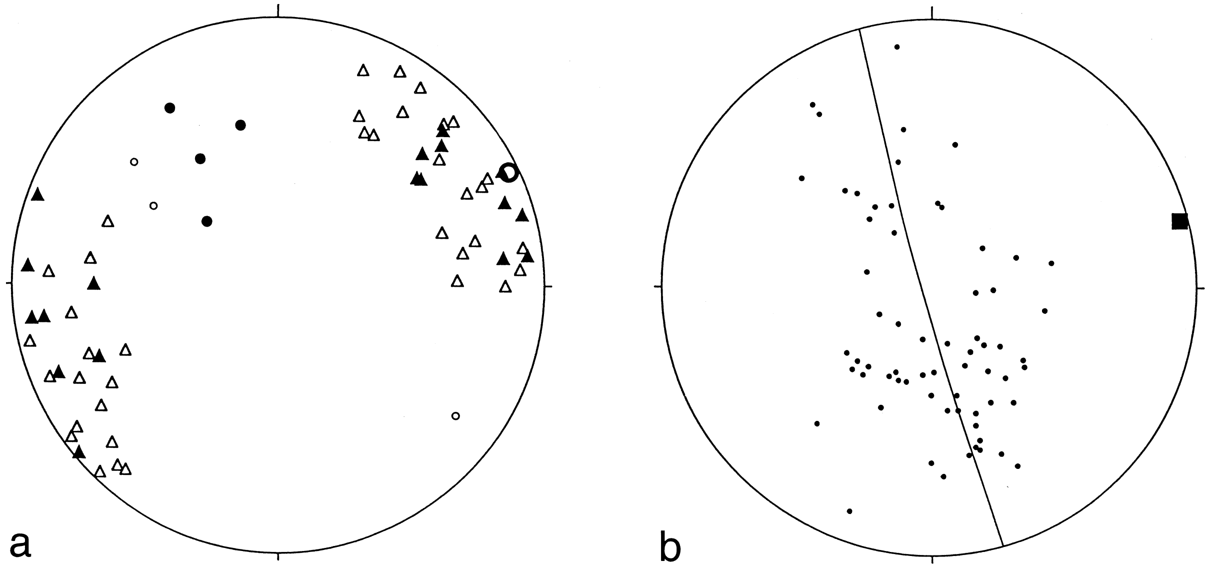


Fig. 6. (a) Lower-hemisphere equal-area plot of mylonitic lineations; data collected in the area depicted in Fig. 3. Closed circles: top-SE lineations with known sense of shear; open circles: lineations presumably related to top-SE nappe stacking; closed triangles: top-ENE lineations with known sense of shear; open triangles: lineations presumably related to normal faulting; large open circle: best fit to all ENE-oriented lineations associated with normal faulting (azimuth 64°). (b) Foliation poles of all mylonites within the Danubian nappes with best-fit great circle, suggesting doming around an axis dipping 3° towards azimuth 75°.

places, Mesozoic cover rocks are invariably mylonitic, the typical aspect of the Oslea and Lupeni limestone being that of a platy fine-grained marble mylonite. The basement of the lower Danubian nappes exhibits a dense network of anastomosing mylonite horizons, each of them tens or hundreds of metres wide. Within our working area all basement rocks of the upper Danubian nappes have been mylonitized, including the mafic rocks of the Dragsan Group. In the lower Danubian nappes, however, the amphibolites of the Dragsan Group generally escaped mylonitization, intense deformation concentrating in quartzo-feldspatic lithologies. Hence, strain intensity within the basement lithologies increases towards higher tectonic units, indicating a strain gradient towards the base of the Getic nappe. In the southeastern part of the Danubian window of the Parang mountains, where the upper Danubian units are missing, the strain related to this second mylonitization event seems to be more concentrated within a narrow area following the base of the Getic nappe.

3.3. Interpretation of top-ENE shearing in terms of orogen-parallel extension

Top-ENE shearing in lower greenschist-facies mylonites is not restricted to the Parang mountains. It has previously been reported from the northern part of the Danubian window, north and south of the Oligocene Petrosani basin by Ratschbacher et al. (1993). These authors, however, attributed this deformation to Cretaceous nappe stacking which, according to them, took place parallel to the strike of the chain. In the following we will argue that this top-ENE shearing is caused by an orogen-parallel stretch which post-dates nappe stacking.

The principal argument to attribute top-ENE shearing to an extensional event is based on the observation that the mylonites of the Danubian window (Fig. 5) are directly overlain by cataclastic fault rocks, invariably found at the base of the Getic nappe (Fig. 7). According to preliminary observations, cataclastic faulting at the base of the Getic nappe also indicates ENE-directed extension (see Fig. 7a). Moreover, lower greenschist-facies mylonites at lower structural levels are often found to be overprinted by east-dipping cataclastic Riedel shears (Fig. 5d), indicating that top-E shearing is continuing under decreasing temperatures. This temperature drop is interpreted to be caused by exhumation during ongoing extension. Cataclasites at the base of the Getic nappe, however, directly overprint pre-Alpine high-grade fabrics of the basement. Alpine retrograde mylonites or foliated cataclasites are totally absent within the base of the Getic nappe within our working area (Fig. 3). Hence, there is an extremely rapid transition from mylonites of the footwall to exclusively cataclastic fault rocks at the base of the Getic nappe in the hangingwall. This change in fault rock fabric coincides with a considerable jump in Alpine metamorphic grade recorded by the Mesozoic cover rocks across the base of the Getic nappe. Unmetamorphic cover rocks of the Getic nappe are found in the immediate vicinity of the base of the Getic nappe north and south of the Petrosani basin, immediately west of our working area. In summary, the base of the Getic nappe is interpreted to represent a low-angle extensional detachment (referred to as ‘Getic detachment’) above the pervasively deformed lower greenschist-facies core complex of the Danubian window.

Additional arguments support this interpretation (see also the fission-track data described in the next section). Firstly, lineations related to the Getic detachment are top-to-the-ENE, while nappe stacking was top to-the-SE, as inferred from the large-scale geometry of the Danubian nappes (Berza et al., 1994), and from the SE-oriented lineations (Figs. 3 and 5a) preserved in the Schela nappe. Secondly, in many places the Severin unit, which formerly represented an important oceanic suture, is omitted at the base of the Getic nappe, as are the passive continental margin units (i.e. the Arjana unit), preserved in the southeastern parts of the Danubian window only. Finally, a considerable thickness of basement at the base of the Getic nappe must have been omitted in many places, in particular along the northern rim of the Danubian window, south of the Petrosani and Hateg basins. There, the cover of the Getic nappe is preserved extremely close to the Getic detachment.

The Danubian units appear subhorizontally oriented in the E–W section (Fig. 4), constructed subparallel to the movement vector during extensional overprint. Hence, the Getic detachment is likely to be inclined to the east-northeast only near the east-

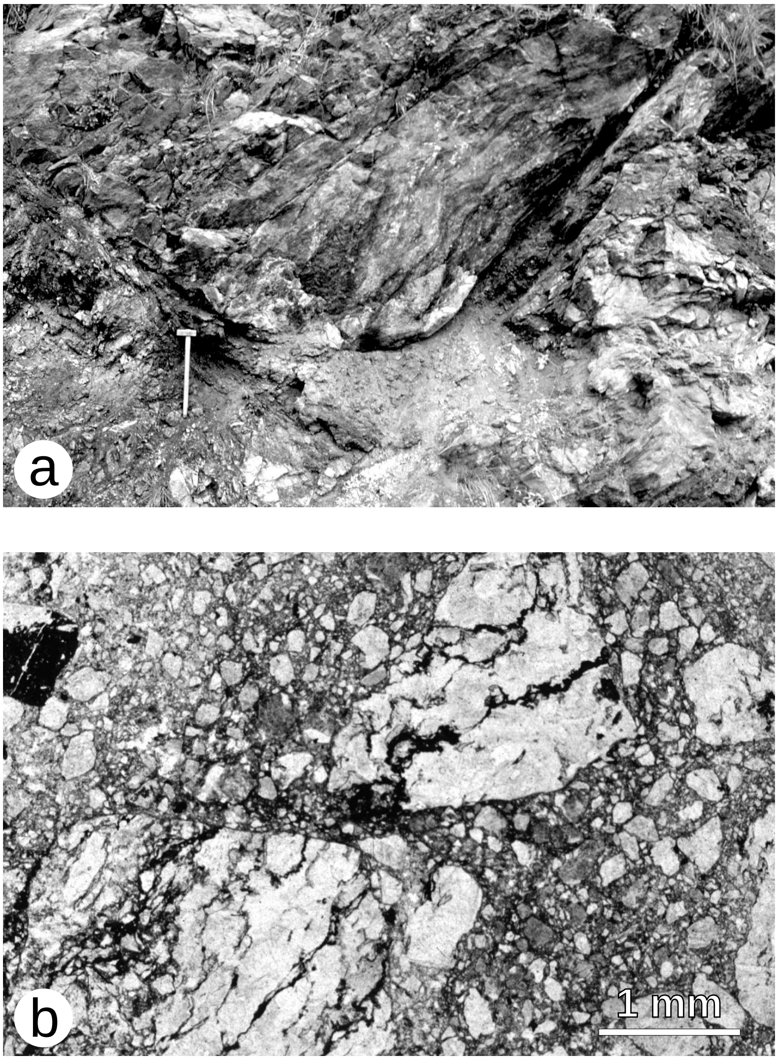


Fig. 7. Brittle deformation at the base of the Getic nappe. (a) Conjugate normal faults at the base of the Getic nappe, cutting through the pre-Alpine high-grade foliation; immediately above greenschist mylonites of the Danubian units and 10 km east of Petrosani (Jiet valley), at the northern margin of the Danubian window. East is to the right. (b) Unfoliated cataclasite, characterized by a wide spectrum in the size of abraded rock and mineral fragments; from the base of the Getic nappe near Polovragi monastery.

ern termination of the Danubian window. It probably had a subhorizontal orientation above the axial culmination of the dome. This, together with the fission-track results discussed below, suggests rotation of the originally east-dipping detachment west of the eastern margin of the Danubian window during ongoing normal faulting, as predicted according to the rolling hinge model (Buck, 1988; Axen et al., 1995).

Doming of the Getic detachment and parts of the Danubian nappe stack in the N–S profile (Fig. 4) also affects the mylonites related to E–W extension. Hence, at first sight, doming due to N–S compression appears to post-date extension. However, no mesoscopic N–S-directed compressional overprint of the mylonites could be observed on the outcrops. An additional argument against substantial N–S shortening of an originally planar, N–S-striking, east-dipping detachment is provided by the observation that the Getic detachment can be followed over a distance of about 100 km to the west along the northern margin of the Danubian window. Mylonites that we relate to top-ENE extensional shearing are found all the way to the west as far as south of the Hateg basin (Ratschbacher et al., 1993, fig. 5). The modest amount of shortening associated with doming would not be sufficient for reorienting a former east-dipping detachment over this distance. Hence, doming may be regarded as contemporaneous with exhumation and E–W extension. The moderate amount of N–S compression during exhumation could be taken to be contemporaneous with orogen-parallel extension, a scenario also known from other regional settings (Mancktelow and Pavlis, 1994). On the other hand, doming during Miocene N–S compression cannot be completely ruled out.

4. Evidence for exhumation by orogen-parallel extension and dating of this event on the basis of fission-track data

Fission-track dating has been carried out on six samples, five of which contained zircon and apatite, one only zircon. Mineral separation, mounting, polishing and etching were carried out according to the techniques described by Seward (1989). All samples were treated using the external detector method with a ζ-value (Hurford and Green, 1983) of 348 (SRM 612 and FCT zircon) and 357 (CN5 and Durango apatite). All ages are 'Central ages' (Galbraith and Laslett, 1993) and errors are quoted as 1σ. Annealing temperatures for fission tracks are 90 ± 30°C for apatite (Green et al., 1989) and 240 ± 60°C for zircon (Yamada et al., 1995). Track lengths analysis in apatite was made on horizontal confined tracks (tints and tincles, Bhandari et al., 1971) and measured at 10 × 100 × 1.25 magnification. Fission-track data are listed in Table 1 and ages are shown in Fig. 8, together with their radial plots (Galbraith, 1990) and track lengths distributions.

Within the footwall of the Getic detachment (Danubian nappes) fission-track ages for both zircon and apatite decrease from west to east. In the west zircon FT-ages of 46 and 40 Ma and two apatite FT-ages of 37 Ma could be determined (samples 9624 and 9627, Table 1). This indicates rather fast cooling (about 25°C/Ma) of the Danubian in the western Parang mountains during Late Eocene time. At the easternmost end of the Danubian window, immediately adjacent to the Getic detachment, two Danubian samples (9607 and 9640) yielded zircon ages of around 30 Ma, and substantially younger apatite ages of 16 and 18 Ma, respectively. Within the Getic nappe two samples, also collected near the Getic detachment at the eastern end of the Danubian window and in the immediate vicinity of samples 9607 and 9640 (samples 9611 and 9641), yielded zircon FT-ages of 64 and 59 Ma, while an apatite age of 21 Ma could be determined for one of them. The cooling rate for this sample is <4°C/Ma for the time span covered.

Our results indicate very distinct cooling (and exhumation) histories of the Getic and Danubian units, and additionally, differences in the cooling rates within the Danubian window, going from east to west. The offset in zircon fission-track ages from 30 Ma at the eastern margin of the Danubian window to 64 and 59 Ma in the Getic nappe is spectacular in the sense that it occurs over a few tens of metres across the east-dipping Getic detachment. In combination with the structural evidence, this indicates that the zircon fission-track data are related to differential cooling and exhumation in the context of normal faulting across the Getic detachment at around 30 Ma. Thus, this detachment brings cold, early exhumed parts of the Getic nappe in direct contact with Danubian lower greenschist-facies mylonites, which did not cool below about 200°C before the Tertiary. On the other hand, the apatite ages are offset by 3 to 5 Ma only. This small difference in apatite ages between Getic and Danubian in the east indicates almost coeval final cooling of both these units below 60°C after about 20 Ma (lower limit of the partial annealing zone for apatite). Substantial normal faulting therefore pre-dates the Early Miocene. The bimodal track lengths distribution in sample 9641 from the hangingwall (cf. Fig. 8) indicates reheating of an earlier cooled sample to around 100°C, possibly related to the advection of heat from the exhumed warm footwall. The two samples from the footwall further west (9627 and 9624), recording rather fast cooling, suggest that normal faulting started to be active during the Late Eocene to Early Oligocene. Since the apatite ages in the western samples are even older than the zircon ages obtained at the eastern margin of the Danubian window, rapid cooling related to

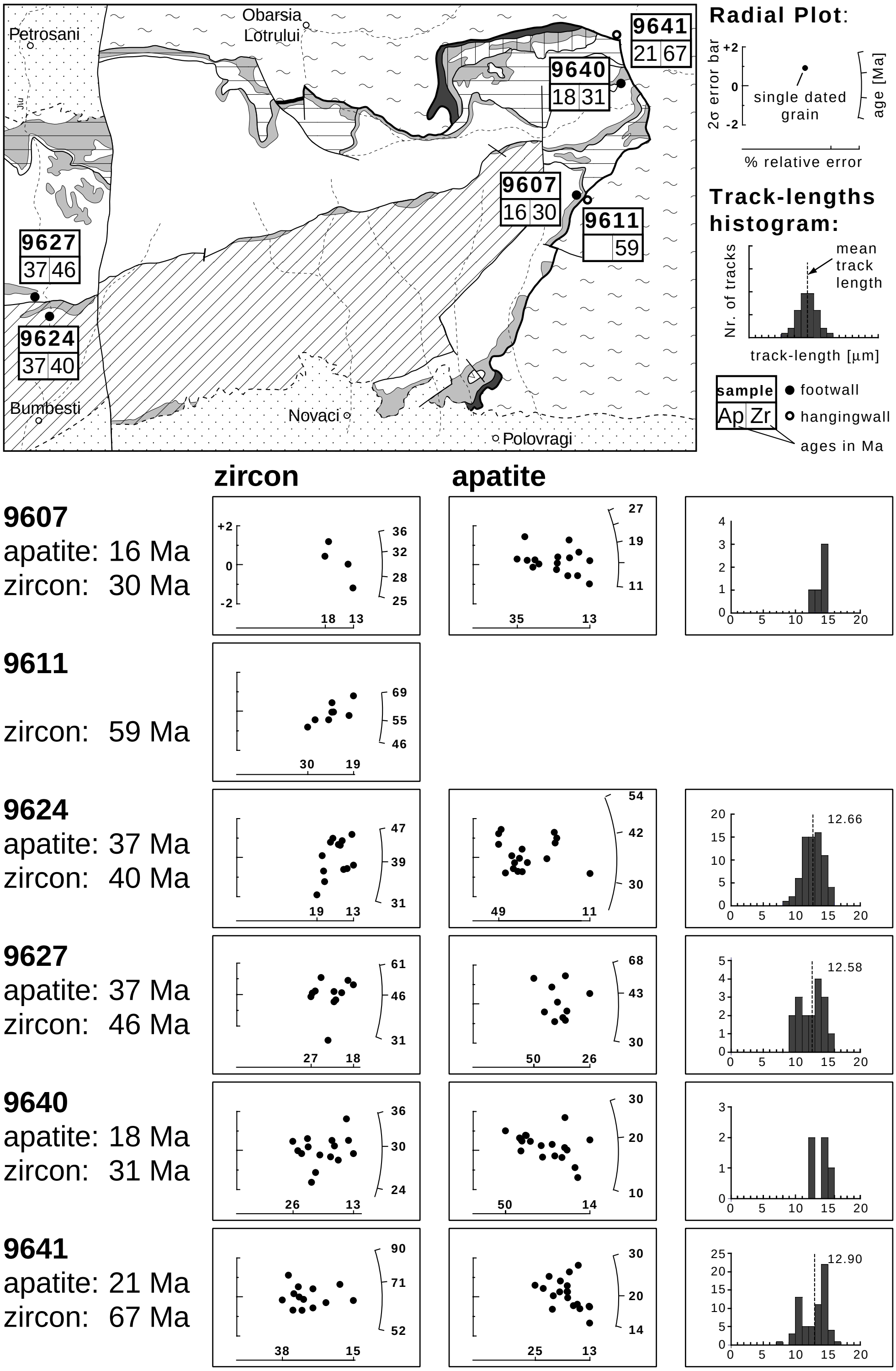


Fig. 8. Zircon and apatite fission-track ages, together with radial plots (Galbraith, 1990) for single grain age distributions and track lengths histograms for confined horizontal tracks in apatite. Sample localities are indicated in the tectonic map of the Parang mountains (cf. Fig. 3).

Table 1
Fission-track data

Sample	Alt. (m)	Nr. of grains counted	Track density ($\times 10^4$ cm^{-2}) (counted)			P (χ^2) (%)	Age $\pm$ 1σ (Ma)
			Standard	ρ_s	ρ_i		
96-07 A	1900	17	138 (1857)	14.61 (132)	231.9 (2095)	98	15.5 $\pm$ 1.4
96-07 Z	1900	4	17.36 (1292)	526.3 (352)	530.8 (355)	41	29.9 $\pm$ 2.4
96-11 Z	1900	9	17.22 (1292)	1574 (466)	794 (235)	98	59.1 $\pm$ 5.0
96-24 A	350	18	131.1 (1857)	34.96 (209)	223.2 (1334)	84	36.6 $\pm$ 2.8
96-24 Z	350	13	16.94 (1292)	486.7 (1231)	358.2 (906)	50	39.9 $\pm$ 2.1
96-27 A	700	10	136.5 (1857)	2.68 (44)	134.9 (287)	52	37.2 $\pm$ 6.1
96-27 Z	700	12	16.80 (1292)	738.3 (613)	466.1 (387)	35	46.1 $\pm$ 3.3
96-40 A	1100	19	135.7 (1857)	19.72 (117)	262.9 (1560)	96	18.1 $\pm$ 1.8
96-40 Z	1100	15	15.39 (1292)	1055 (1111)	958 (1009)	88	29.4 $\pm$ 1.5
96-41 A	750	19	135 (1857)	27.63 (220)	322 (2564)	91	20.6 $\pm$ 1.5
96-41 Z	750	13	16.52 (1292)	816.4 (659)	363.0 (293)	99	64.3 $\pm$ 4.9

Apatite (A) and zircon (Z) fission-track data. All samples have been treated using the external detector method. First and second column refer to sample number and altitude, respectively. Number of grains counted is given in column 3. Standard, spontaneous (ρ_s) and induced (ρ_i) track densities are shown in columns 4, 5 and 6, respectively. Number of counted tracks is given in parenthesis. (χ^2) test (Galbraith, 1981). Zircon ages were calculated with a zeta value of 348 for dosimeter glass SRM 612. Apatite ages were calculated with a zeta value of 357 for dosimeter glass CN5. All ages are 'Central ages' (Galbraith and Laslett, 1993), errors are quoted as 1σ.

normal faulting must have been earlier (Eocene) in the west.

According to the available data, and especially by taking into account the differences in cooling history between the eastern and western sample localities within the Danubian window, we propose that rather fast cooling and exhumation started in the Eocene along a low-angle normal fault, dipping with about 20º to the east. This normal fault was located at the present-day eastern end of the Danubian window. The samples, now situated in the western part of the Parang mountains (i.e. samples 9624 and 9627), are inferred to have been subsequently carried westward below a flat-lying detachment, according to the rolling hinge model (Buck, 1988; Axen et al., 1995). This westward transport of samples 9624 and 9627 had to occur under still elevated temperatures above 60ºC, within the partial annealing zone of apatite (Green et al., 1989), as demonstrated by the rather short mean track lengths determined. Computer-based thermal modelling (Gallagher, 1995) suggests a period of about 20 Ma within the partial annealing zone, allowing for the observed mean track lengths. Later stages of cooling are thought to have been erosion-controlled, as cooling and exhumation slowed down during Oligocene time and finally ceased in the Miocene.

5. Additional constraints on the timing and amount of orogen-parallel extension in the Southern Carpathians

The onset of orogen-parallel extension in the Eocene significantly post-dates Cretaceous nappe stacking and a probable Late Cretaceous extension event. Many of the available radiometric age determinations of Alpine tectonometamorphic events in the Danubian nappes (mostly K–Ar ages on muscovite and biotite, and a few Rb–Sr muscovite ages) span the 120–70 Ma time interval (see compilation of data by Ratschbacher et al., 1993, and data by Grünenfelder et al., 1983, and Dallmeyer et al., 1996). Rb–Sr muscovite ages (Ratschbacher et al., 1993), give 76 and 72 Ma, respectively, and represent formation ages related to 'Laramide' nappe stacking according to these authors. Hence, in contrast to the fission-track ages, the isotope data record an earlier (Cretaceous) event of deformation and/or cooling.

The post-tectonic cover of the Getic nappe, preserved at the northern border of the Danubian nappes and near the Getic detachment (Fig. 2), starts with Upper Cretaceous sediments and indicates that Cretaceous nappe stacking was immediately followed by exhumation and erosion of structurally higher parts of the Getic nappe. However, the tectonic scenario leading to Late Cretaceous cooling is not clear yet. In our opinion it is related to an earlier (Late Cretaceous) extension event. Neubauer et al. (1993) interpreted Late Cretaceous normal faulting, associated with the Late Cretaceous post-tectonic basins, to have occurred in a transpressional regime. Later, Neubauer et al. (1997) postulated this normal faulting to be linked with Late Cretaceous core complex formation exhuming the Danubian window.

The fission-track data presented in this study clearly indicate that the above-discussed Late Cretaceous phase of exhumation (probably related to extension) did not lead to substantial cooling of the underlying Danubian units in the Parang mountains. According to the fission-track data, the eastern Danubian units must have stayed at temperatures above 240ºC until about 37 Ma ago, i.e. for some additional 35 Ma since Maastrichtian time. At least for the area of the eastern Danubian window, the fission-track data exclude core complex formation during the Late Cretaceous, as postulated by Neubauer et al. (1997). It is not clear yet, why mylonitization related to Neogene extension did not reset some of the radiometric ages. Possibly, the temperature conditions prevailing during Neogene mylonitization (estimated to be around 300ºC to 350ºC) were insufficient for resetting the Cretaceous ages in the specimens analysed so far. On the other hand, it has to be emphasized that most of the Cretaceous radiometric ages were obtained within the Danubian window west of our working area.

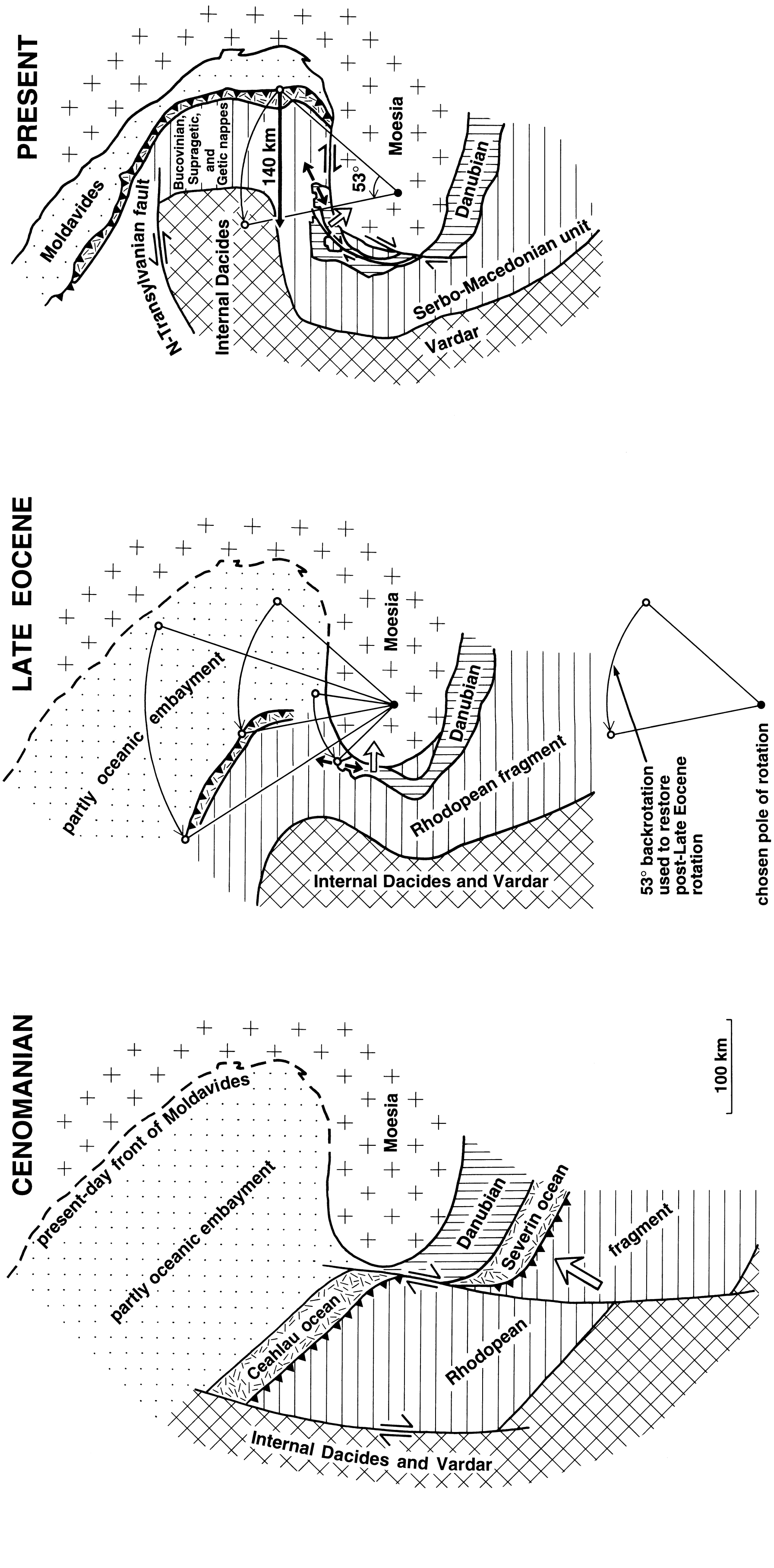
Two conclusions may be drawn from this discussion. (1) Orogen-parallel extension related to the Getic detachment occurred during the Eocene. Because it substantially post-dates Cretaceous nappe stacking (and probable Cretaceous normal faulting), it represents a separate tectonic event. It cannot result from the immediate extensional collapse of an overthickened crust formed by Cretaceous nappe stacking. (2) Judging from the very different temperature regime prevailing in the Danubian and Getic units during the time interval 70 to 35 Ma (240–300ºC and about 100ºC, respectively), Eocene extension must have led to the omission of a very thick slice of crustal material at the base of the Getic nappe: on the order of 5–6 km, assuming a temperature difference of 170ºC and a normal geothermal gradient. Consequently the E–W stretch during Eocene time, related to mylonitization with top-ENE movement, must be considerable. It is likely to substantially exceed the minimum estimate of 20 km (given by the present-day distance between the western and eastern fission-track sampling localities within the footwall and assuming a rolling hinge model). However, the total amount of extension remains to be quantified.

The Cerna-Jiu fault and the Oligocene deposits in the Petrosani basin (Fig. 2) provide important constraints regarding the time when orogen-parallel extension was completed. The Cerna-Jiu fault clearly offsets the northern margin of the Danubian window, and hence the Getic detachment, by some 35 km (Berza and Draganescu, 1988). Ratschbacher et al. (1993) interpret the Petrosani basin (an Oligocene half-graben, Fig. 2) to have formed during dextrally transtensive movements near the NE termination of the curved Cerna-Jiu fault, Chattian sediments also recording some of the deformation related to this strike-slip fault. This means that the Getic detachment at least pre-dates the Chattian infill of the Petrosani basin (i.e. about 30 Ma). This is compatible with our fission-track ages obtained at the eastern end of the Danubian window (also 30 Ma for zircon FT), because rapid cooling related to extension is likely to post-date movements along the Getic detachment by a few Ma.

In summary, orogen-parallel extension within the eastern part of the Danubian window appears to represent an Eocene event which substantially post-dates Laramide nappe stacking. This event is therefore unrelated to Cretaceous orogeny. It terminated no later than 30 Ma ago, i.e. during the Early Oligocene, being immediately followed by a new stress regime, i.e. E–W compression related to dextral shearing along the Cerna-Jiu fault (Ratschbacher et al., 1993).

6. The role of Eocene orogen-parallel extension in the context of arc formation in the Southern and Eastern Carpathians

During orogen-parallel extension in the Southern Carpathians a considerable thickness of Palaeogene flysch deposits accumulated in the Moldavides of the Eastern Carpathians (i.e. the Tarcau sandstone, Sandulescu et al., 1981). In the more internal units of the Moldavides (i.e. the Convolute Flysch nappe)


PRESENT
Moldavides
N-Transylvanian fault
Internal Dacides
Bucovinian, Supragetic, and Getic nappes
140 km
53°
Moesia
Danubian
Serbo-Macedonian unit
Vardar
LATE EOCENE
partly oceanic embayment
Moesia
Danubian
Rhodopean fragment
Internal Dacides and Vardar
53° backrotation used to restore post-Late Eocene rotation
chosen pole of rotation
CENOMANIAN
present-day front of Moldavides
partly oceanic embayment
Moesia
Danubian
Severin ocean
fragment
Ceahlau ocean
Rhodopean
Internal Dacides and Vardar
100 km

the Palaeogene unconformably overlies older flysch, deformed during the 'Laramide' phase. According to Sandulescu et al. (1981), these Palaeogene deposits are not affected by crustal shortening before the Miocene. However, Zweigel (1996) postulates Palaeogene shortening in the internal Moldavides, indicating continuous convergence during the Tertiary. In the more external units (i.e. the Tarcau nappe), not affected by Laramide compression, shortening certainly does not set in before Burdigalian time, the Palaeogene flysch deposits being partly shed from an internal 'cordillera', partly from the foreland (Sandulescu et al., 1981). Many authors (e.g. Linzer, 1996) assume that the Palaeogene flysch of the Moldavides was at least partly deposited on an embayment of oceanic lithosphere, still preserved between the East European craton and the front of the internal East Carpathian nappes (the Rhodopean fragment and the Ceahlau unit). During the Palaeogene a considerable thickness of deposits is also found in the Getic depression, south of the Southern Carpathians. There, Matenco et al. (1997) found evidence for syndepositional extension with NNE–SSW to NW–SE oriented minimum principal stress axes. Hence, the stretching direction inferred for the Getic detachment (WSW–ENE) appears to be rotated in a clockwise sense in respect to contemporaneous extension in the Getic foredeep.

In order to restore the position of the orogenic front of the internal units of the Eastern Carpathians (Ceahlau nappe and Rhodopean fragment) at the time of orogen-parallel extension in the Southern Carpathians during the Eocene (Fig. 9), it is necessary to account for post-Eocene E–W shortening of the Moldavides in the order of some 140 km in a transect south of the Vrancea window in the southernmost Eastern Carpathians (Morley, 1996, see figs. 4 and 6). As pointed out by Morley (1996), palinspastic restoration around the very arcuate bend near the transition from Eastern to Southern Carpathians has to take into account transport directions which change in time, producing divergent displacement trajectories. For our admittedly speculative restoration of the Eocene situation (Fig. 9) we assumed that the estimated 140 km of E–W shortening essentially resulted from a rigid block rotation of the Rhodopean fragment around a rotation pole situated in the centre of Moesia. Also, the dextral movements along the Cerna-Jiu and Timok faults were retrodeformed in Fig. 9. The choice of the position of this rotation pole was governed by finding a best-fit rotation pole, compatible with the curved fault traces of the post-Eocene Cerna-Jiu and Timok dextral strike-slip faults. These faults took up some, but by no means all of the required dextral strike-slip between Moesia and the internal parts of the Rhodopean fragment (an estimated total of 80 km: 35 km along the Cerna-Jiu fault and 45 km along the Timok fault). The amount of rotation chosen in this reconstruction (53°) is a function of the E–W shortening estimate in the Vrancea transect (140 km) and the chosen position of the rotation pole.

The dextral movements along the Cerna-Jiu and Timok faults are not only insufficient in magnitude to accommodate the postulated 53° clockwise rotation. They are also unable to produce the dextral offset between Moesia and the Danubian units, situated south of the Cerna-Jiu fault. Also, a large part of the rotation must have occurred during the Miocene, the major period of E–W shortening in the Moldavides, i.e. after movements along these strike-slip faults ceased. Because the amount of Miocene dextral strike-slip movements within the Southern Carpathians is very modest (Ratschbacher et al.,

Fig. 9. Sketches of the present configuration and palinspastic reconstructions for Eocene and Cenomanian times. The semi-quantitative sketch of the Late Eocene was arrived at by 53° back-rotation of all units behind the front of the Ceahlau unit, around a rotation pole situated in the Moesian platform. It also includes retrodeformation of the Cerna and Timok dextral strike-slip faults (but not the effects of Late Eocene to Oligocene Danubian core complex formation). The sketches for the present and Late Eocene situations also indicate the orientations of the principal stretch related to Palaeogene extension (arrows presently striking ENE) and the directions of Late Cretaceous nappe stacking (arrows presently striking southeast). The sketch for the Late Eocene was drawn in order to illustrate the larger-scale situation during core complex formation in the Danubian window. The sketch for the Cenomanian is arrived at by further back-rotation and by restoring dextral movements west of Moesia in a purely qualitative way. It is meant to illustrate that Late Cretaceous nappe stacking (arrow) possibly took place in a northeasterly direction, compatible with the plate tectonics scenario at this time. Also note that the Rhodopean fragment appears as a continental terrane, caught between the Vardar and Severin–Ceahlau oceans in this reconstruction.

1993), most of the required 53º clockwise rotation had to take place between Moesia and the southern rim of the Danubian units. The amount of dextral strike-slip documented within the sediments of the Getic foredeep is modest according to Matenco et al. (1997) but compatible with some clockwise rotation. Much of the required dextral strike-slip displacement must have occurred between Moesia and the Danubian nappe pile: along a hypothetical fault zone situated within the northern prolongation of the Getic foredeep, presently buried underneath the Danubian nappe pile.

Large-scale clockwise rotation of the Southern Carpathians, together with the South Apuseni mountains and the Tisia unit, is strongly supported by palaeomagnetic data. Patrascu et al. (1994) provide evidence for some 60º clockwise rotation of the Apuseni mountains, affecting and hence post-dating Eocene–Oligocene magmatic rocks formed in the 43 to 30 Ma age interval and prior to the Late Miocene. However, the timing of the rotation is still controversial because of problems in dating the analysed magmatic rocks. According to Balla (1987) the northern boundary of the clockwise-rotated domain coincides with the Mid-Hungarian belt (Fig. 1), including the N Transylvanian fault indicated in Fig. 9.

According to the Eocene reconstruction of Fig. 9, the stretching direction inferred for the Getic detachment is almost N–S-oriented at the time of orogen-parallel extension. Laramide nappe stacking would be east-directed after removal of the Oligocene to Early Miocene clockwise rotation. However, additional clockwise rotations, pre-dating the Eocene, are indicated by clockwise rotation of up to more than 90º of Cretaceous and older formations (Patrascu et al., 1994) in the Tisa unit. Hence Laramide nappe stacking most probably took place in a northeasterly direction.

The situation depicted for the end of the Meso–Cretaceous ('Austrian') phase (see Burchfiel, 1980, for a very similar reconstruction) suggests that the Rhodopean fragment originally overrode the Severin ocean and its northwestern continuation, the Ceahlau ocean towards the north-northeast. This is compatible with plate convergence vectors for Meso–Cretaceous times, obtained on the basis of the reconstructions by Dercourt et al. (1986) (see fig. 2c in Ratschbacher et al., 1993). The corner effect provided by the western edge of Moesia initially provoked dextrally transpressive, followed by dextrally transtensive deformation in the Southern Carpathians, as they (together with the rest of the Rhodopean fragment) moved northward and past the western edge of Moesia (see discussion in Ratschbacher et al., 1993). This differential northward movement, later followed by clockwise rotation of the Rhodopean fragment, is facilitated by the existence of a partly oceanic embayment in the depositional realm of the Moldavian flysch units. This oceanic lithosphere, being progressively subducted underneath the advancing Rhodopean fragment, offers little resistance to the Rhodopean fragment which invades it.

According to this scheme, Laramide nappe stacking, characterized by the offscraping of Danubian nappes from the southwestern edge of Moesia, would have taken place near the SW edge of Moesia. It was associated with top-NE movements during a dextrally transpressive stage, between the Cenomanian and Eocene stages depicted in Fig. 9. After an intermediate stage of purely dextral strike-slip between Moesia and the Rhodopean fragment, the Danubian–Getic nappe stack was affected by tangential stretching, i.e. NNE–SSW extension, as the Rhodopean fragment started to spread into the oceanic embayment. Hence, the situation during orogen-parallel extension in the Eocene would be somewhat similar to that envisaged during the Miocene in respect to extension in the Pannonian basin and contemporaneous compression in the thrust belt of the Western Carpathians (Royden and Burchfiel, 1989). Eocene extension in the Southern Carpathians, leading to core complex formation below the Getic detachment, may be viewed as contemporaneous with plate convergence between the Rhodopean fragment and the still partly oceanic embayment in front of an advancing accretionary wedge (the 'Cordillera' shedding detritus into the Eocene flysch basin of the Tarcau unit). Later, i.e. during the Oligocene and Miocene, the Rhodopean fragment underwent a clockwise rotation, only partly accommodated by dextral wrenching within the Southern Carpathians. Much of this wrenching was taken up in the Getic foredeep, particularly within its projection underneath the Danubian nappes. This same dextral wrenching was instrumental for the formation of the narrow arc situated near the southern tip of the Eastern Carpathians (see Zweigel et al., 1998, fig. 14).

7. Conclusions

The kinematic analysis of fault rocks in the Danubian nappe stack and at the basis of the Getic nappe, together with fission-track data, suggests that the Danubian window was exhumed during orogen-parallel extension in the Eocene. This extension significantly post-dates Cretaceous nappe stacking and pre-dates Oligocene to Miocene dextral strike-slip faulting. It is very likely that this orogen-parallel stretch, oriented WSW–ENE in present-day coordinates, subsequently underwent a significant clockwise rotation.

According to our working hypothesis, proposed in Fig. 9, arc formation in the Southern Carpathians and the transitional area into the Balkans would essentially be due to the corner effect of Moesia. It would largely pre-date Miocene subduction roll-back in front of the Western and Eastern Carpathians. We propose that this mechanism of arc formation was initially associated with a very substantial orogen-parallel N–S stretch in the Southern Carpathians during the Eocene, as the Rhodopean fragment moved northward, past Moesia. This extension was then followed by 50° clockwise rotation of the Southern Carpathians, associated with dextral wrenching, largely taken up along the northern boundary of Moesia, now buried underneath the Danubian units.

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