Cretaceous to Cenozoic thermal evolution of the southwestern South Carpathians: evidence from fission-track thermochronology

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Abstract

The southwestern South Carpathian orogen is composed of various nappe complexes which were assembled during the Cretaceous–Cenozoic orogeny. These are from footwall to hangingwall: (1) the Danubian nappe complex including a Cadomian/Variscan basement; (2) the Arjana and Severin units with Jurassic to Early Cretaceous rift and oceanic sequences; and (3) the Getic nappe complex with Variscan continental basement. Fission track (FT) thermochronology on apatite, zircon and sphene from samples collected from various units of the South Carpathians, in conjunction with field constraints and previous geochronology is used to characterise the Alpine tectonic events and to restore the pattern and amount of exhumation since the Cretaceous. Zircon from the Ilsch unit and the Danubian Liassic cover sequence yields FT ages around 200 Ma suggesting cooling of the rift flanks prior to the opening of the Severin rift. Zircon and sphene from the Getic and Danubian basement units yield FT ages averaging 110 Ma and indicating cooling under 240°C of the basement contemporaneous with, or postdating thrusting. Apatite FT ages display a decreasing age trend from the hangingwall (65 Ma) to the footwall units (30 Ma). The age data and corresponding horizontal confined track length distributions suggest that exhumation of the nappe pile occurred in two stages: the first is related to the Late Cretaceous nappe stacking and the second one to the final thrusting of the South Carpathians onto the top of the Moesian platform. Apatite FT ages along major brittle wrench faults indicate reheating above ca. 120°C during fluid flow associated with fault (re)activation during Oligocene and Neogene times. Thus, shear zone rocks experienced a higher temperature overprint during Cenozoic time than rocks of the unaffected nappe pile. Temperatures of hydrothermal flow along these zones decreased below 100°C progressively starting with the Late Oligocene–Early Miocene when the area began to override the Moesian platform. © 1998 Elsevier Science B.V. All rights reserved.

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1. Introduction

The southwestern South Carpathians (Fig. 1) represent a nappe pile which is mainly composed of pre-Alpine basement nappes (e.g.: Berza et al., 1994; Kühn, 1996), separated by the ophiolitic Severin unit. Tectonostratigraphically upwards, these units include: (1) the Moesian platform with thick early Palaeozoic to Neogene sediments; (2) the Danubian nappe complex (exposed within the Danubian window) with a Cadomian/Variscan basement and a late
Palaeozoic to Mesozoic cover; (3) the Severin flysch-and-ophiolite nappe and the Arjana nappe with Mesozoic syn-rift deposits; and (4) the Getic and the Supragetic nappes (both with Variscan basement and Mesozoic cover). The nappe assembly was completed during the Early to Late Cretaceous, and was mainly overprinted by Palaeogene and Neogene wrenching along steep dextral strike-slip faults (e.g.: Ratschbacher et al., 1993; Matenco et al., 1997).

In the study region the metamorphic overprint related to collision remained largely within very low-grade metamorphic conditions. A reconnaissance study involving fission track analysis was carried out in order to constrain the low-temperature thermal history and the amount of exhumation since the Late Cretaceous.
Apatite, zircon and sphene fission track (FT) studies are commonly used to restore late stage vertical motions within orogens (e.g.: Dumitru, 1989; Wagner et al., 1989; Rohrman et al., 1994) as well as cooling events related to source areas of sedimentary rocks (e.g.: Hurford et al., 1984; Brandon and Vance, 1992; Rohrman et al., 1996). These are due to the fact that FT ages are sensitive to cooling below ca. 300°C in zircon (e.g.: Tagami et al., 1996; Tagami and Shimada, 1996) through 120°C in apatite (Naeser and Faul, 1969; Gleadow and Duddy, 1981; Gleadow et al., 1986a; Green et al., 1989).

Based on sedimentological and geochronological data compiled from the literature, on own field work and on 40Ar/39Ar analyses we present a more complete picture regarding the evolution of the South Carpathians. Zircon ages from the Severin nappe and Liassic sedimentary cover provide information about the Jurassic extension. Sphene and zircon ages from Getic and Danubian basement units record Early Cretaceous collision and Late Cretaceous post-metamorphic cooling in the northern Danubian window. Apatite FT thermochronology on samples from Getic basement, Severin nappe and Danubian cover and basement units quantified by modelling of track length distributions and ages records progressive cooling under 120°C starting with the Late Cretaceous. Final cooling under 60°C was associated with Late Oligocene–Early Miocene folding and faulting of the area, when the South Carpathians started to override the Moesian platform. The data provide a number of new constraints for the tectonic evolution of the South Carpathians and demonstrate how the FT thermochronology can be applied to unravel the role of various tectonic events which affected the region.

2. Geological setting

2.1. Danubian units

The Danubian basement mainly comprises amphibolite-grade metamorphic and plutonic rocks. The amphibolite-grade metamorphism and associated ductile deformation have been generally regarded as Middle Proterozoic (Sandulescu, 1984; Kräutner et al., 1988). K–Ar data from both upper and lower Danubian basement units show that the Alpine rejuvenation decreases from top to bottom. Only in the southern lower Danubian units, ages within the interval of 500–650 Ma were reported (Grünenfelder et al., 1983). However, recent extensive 40Ar/39Ar mineral data from all the units (Dallmeyer et al., 1996) and structural profiles (Neubauer et al., 1997) show that remnants of a Precambrian orogen are preserved only in the southern lower Danubian nappes. Muscovite and hornblende concentrates from all other high-grade metamorphic samples consistently yield Variscan ages. Within the Danubian window, the Cretaceous overprint is penetrative and reached epidote-amphibolite facies conditions only in the northern part (Retezat, Tarcu Mountains) (Neubauer et al., 1997). To the south, Cretaceous reactivation of the Danubian basement occurs only along major shear zones (Bojar et al., 1993; Dallmeyer et al., 1996).

To constrain the Late Cretaceous evolution as well as the nappe geometry, the stratigraphy and the distribution of the Arjana and Severin Nappes and of the Late Cretaceous deposits belonging to the Danubian cover will be discussed in more detail.

The Danubian cover sequences are preserved in three large synforms: Svinita-Svinecea, Presacina and Cerna-Jiu (Fig. 1) (Fig. 2). Lateral variations of stratigraphy and lithofacies are shown in Fig. 3. The Late Carboniferous and Permian successions are interpreted in terms of molasse deposits following the Variscan orogenesis.

A new cycle of sedimentation started with crustal extension conditions in the Liassic. A second transgression took place at the end of the Early Jurassic, probably as a result of a new extension phase, and Doggerian to Malmian sedimentation shifted to carbonate deposition. In the region westward from the Godeanu klippe, the clastic sedimentation was accompanied by intrusion of mafic rocks (dykes) and deposition of Liassic to Malmian tuffs (Fig. 3).

The dominantly calcareous sedimentation of the Early Cretaceous is followed by a sedimentary gap of late Aptian–Albian age. Thus the Late Cretaceous sequences located in the Svinita-Svinecea, Presacina and Cerna zones are in unconformable contact with the Early Cretaceous calcareous deposits.

1. In the Svinita-Svinecea zone, the Late Cretaceous without palaeontological record, overlies dis-
cordantly the Tithonian–Hauterivian (Pop, 1996) and are considered of Senonian age. (2) In the Presacina zone, located south-southeastward to the Godeanu klippe, Nastaseanu (1980) separates the Presacina facies with Liassic clastic deposits, Doggerian to Early Cretaceous carbonatic deposits and the Fenes facies with Liassic clastic deposits and Doggerian–Malmian volcanoclastics. Both facies are covered by the ‘wildflysch’ formation of early Senonian age (Nastaseanu, 1980) (Fig. 3) which is followed by clastic deposits assumed to be of late Senonian age. Northeastward to the Godeanu klippe, the Late Cretaceous deposits overlie directly Barremian–Aptian limestones. The Late Cretaceous succession described by Pop (1986) (with Nadanova, and ‘wildflysch’ formations) is similar to those described to the east, in the Cerna-Jiu region. (3) In the Cerna zone, the Early Cretaceous limestones are followed by the Nadanova formation of Cenomanian–Turonian age (e.g.: Bercia et al., 1987; Hartopanu et al., 1987) covered by the ‘wildflysch’ which contains mafic rocks, tuffs and various olistoliths (including Late Jurassic–Early Cretaceous limestone, sandstones from the Severin flysch nappe, mafic volcanic and metamorphic rocks). Stanoiu (1978a, 1982a) described sandstones (e.g., the Gardaneasa sandstone) of Campanian–Maastrichtian age as the youngest formation of Late Cretaceous ‘wildflysch’ deposits. Thus, the age of the ‘wildflysch’ formation although with a poor palaeontological record, is considered as late Turonian–Maastrichtian (e.g.: Stanoiu, 1982b).

2.2. Severin and Arjana nappes

The Severin flysch, actually underlying the Bahna and Portile de Fier Getic klippen, is of Jurassic–Neocomian age (e.g.: Stanoiu, 1978a,b; Bercia et al., 1987) or late Jurassic–Aptian age (e.g.: Mutihac, 1990; Pop, 1996). The biostratigraphic age of the formation is not well constrained.

The Severin rift formed during the Jurassic, separated the Danubian and the Getic continental plates (e.g.: Sandulescu, 1984) and widened until the Early Cretaceous.

Rift deposits are partially preserved in different facies of all three Danubian synforms. (1) In the Svinita-Svinecea region, the Severin nappe with Tithonian–Aptian deposits overthrust late Senonian deposits (Pop, 1996). (2) In the Presacina region, northwestern of the Godeanu klippe, Gherasi and Hann (1990) separated the Arjana nappe consisting of a ‘volcano-sedimentary formation’ of Liassic to early Malmian (?) age, covered by Malmian limestones and by Late Cretaceous black shales, coarse sandstones and conglomerates. We suggest that the Fenes formation (Nastaseanu, 1980) belongs not (at least partially) to the Danubian cover, but represents the southward extension of the Arjana nappe (Fig. 1). Thus, the Arjana nappe and the cover of the Danubian realm are both covered by the Late Cretaceous wildflysch formation. (3) In the Cerna region the Severin flysch nappe with Tithonian to Valanginian deposits (e.g.: Stanoiu, 1978b) overlies Late Creta-
Fig. 3. Stratigraphy and lateral variations of cover sequences within the Danubian and of Severin/Arjana nappes. **Danubian cover**: NF = Nadanova Formation, WF = wildflysch Formation, GB = Gardaneasa beds. **Severin=Arjana nappes**: SN = Severin flysch (Tithonian–Neocomian = Aptian?), AN = Arjana nappe (Lias–Upper Cretaceous). Data were compiled from Stanoiu (1978a, 1982a), Nastaseanu (1980), Pop (1986, 1996), Bercia et al. (1987), Hartopanu et al. (1987), Gherasi and Hann (1990), Mutihac (1990). The age of the first Neogene overstep deposits on the Getic nappe basement is Aquitanian.

2.3. **Getic nappe**

In the Getic and Supragetic basement, the penetrative deformation is of Variscan age (Dallmeyer et al., 1996). Alpine $^{40}\text{Ar}/^{39}\text{Ar}$ mineral and whole rock phyllite ages have been recorded only along ductile shear zones. A muscovite concentrate from the western border of the Bahna klippe records a $194\pm0.1\text{ Ma}$ plateau age suggesting an Early Jurassic event (Bojar et al., 1996). A similar Jurassic tectonothermal overprint in the Supragetic nappe was described by Udubasa et al. (1997) for the Fagaras Mountains.
Within the Getic klippen, the sedimentary cover is largely missing. Senonian deposits occur sporadically under the Oligocene deposits of the Petrosani basin (Moisescu, 1981). Westward, in the Resita synform, the sedimentation stopped in the Albian (Mutihac, 1990). In the Hateg basin thick Vraconian to late Maastrichtian deposits are also known (Stilla, 1986) to overlie discordantly the Albian continental deposits.

2.4. Tertiary basins

The first evidence of Palaeogene deformation is the activation of the dextral Cerna fault (Berza, 1988) which separates the Godeanu and Bahna–Portile de Fier Getic klippen. The Petrosani basin is interpreted as a negative flower structure related to the strike-slip zone. The sedimentation started with shales, coals and fine limestones of Chattian age (Moisescu, 1981). The basin is mainly filled (80%) by conglomerates and coarse sandstones with a thickness of 700–800 m, also of Chattian age. The Miocene sedimentation continued with Aquitanian marls, coals, marine sandstones and ended with the ‘gravels formation’ of Middle (?) to Late Miocene age. Actually the maximum thickness of the Tertiary sediments reaches 1500 m. The oldest sediments are still underlain by a continuous thin slice (ca. 1–2 km) of the Getic nappe and by discontinuous Senonian deposits. The age of the first overstep sequences for both Danubian and Getic domains seems to be Early Miocene (Dessila-Codarcea et al., 1968). To the north, in the Hateg basin, Palaeogene deposits are also underlain by the Getic basement.

Other important Tertiary basins in the region are the Caransebes-Mehadia graben (eastward from the Godeanu klippe) and the Orsova pull-apart basin. In the Orsova basin, the sedimentation started in the late Aquitanian (Hartopanu et al., 1987) with clastic deposits and stopped at the level of the Sarmatian. The maximal preserved thickness of the sedimentary pile is ca. 1200 m. The first Tertiary deposits of the Caransebes basin are of late Badenian age (e.g.: Marinescu and Popescu, 1987; Marunteanu et al., 1995), the age of the last sediments being also Sarmatian. Thus the oldest deposits from the Orsova and Caransebes basins are of late Aquitanian respectively late Badenian age. These deposits overstep the assembled Getic and Danubian nappes.

After Late Cretaceous nappe stacking, the sedimentation continued eastward of the study area in the Getic depression (e.g.: Motas and Tomescu, 1983). The Palaeocene sediments overstepped the Getic and Severin nappe fronts. The first Tertiary deposits covering Mesozoic sequences of the Moesian platform are of Early Miocene age (e.g.: Paraschiv, 1983; Stefanescu, 1984; Vinogradov, 1988). The regional distribution and geometry of these deposits suggest that they are related to the loading of the South Carpathians orogen (together with the Getic depression) onto the Moesian platform, the age of the first overstep sediments of the two units being Sarmatian. Thus thrusting started in the Oligocene=Early Miocene and ceased in the Sarmatian. A thick pile of Sarmatian to Quaternary deposits is known for the Getic depression and Moesian platform. To the west, the thickness of these deposits progressively decreases so that in the klippen region the Sarmatian occurs discontinuously (e.g.: Caransebes and Orsova basins).

3. Analytical procedures

Mineral concentration has been done using standard heavy liquid and magnetic separation techniques. Zircon crystal were mounted in FEP Teflon and later in PFA Teflon (Danhara et al., 1993) and etched at 200°C in NaOH–KOH–LiOH eutectic melt (Zaun and Wagner, 1985). Sphene and apatite crystals were mounted in epoxy resin. Sphene crystals were etched in HNO₃±HCl±HF±H₂O solution between 20 and 30 min (Gleadow, 1978). Apatite samples were etched for 35–40 s at 20°C in 7% HNO₃ solution.

Mineral samples together with an external detector (low uranium muscovite) and mineral standards were irradiated in the TRIGA reactor of the ‘Atominstut der Österreichischen Universitäteten’. During the irradiation, neutron flux was monitored using SRM 612 and CN standard glasses (Tables 1 and 2). The spontaneous track densities were determined using a ×100 oil objective mounted on an Olympus BH-2 microscope. Systematic checks of the grains were made, and all suitable grains were counted.
Only properly etched grains were counted, where the section was parallel to the crystallographic c-axis.

For apatite tracks, length measurements were done on horizontal confined tracks (Gleadow et al., 1986b) as this method seems to be the best one in estimating track length distributions.

The ages were calculated using the standard fission track age equation (Hurford and Green, 1982), for a geometry factor of 0.5 (Gleadow and Lovering, 1977) and the zeta calibration recommendation (e.g.: Hurford and Green, 1983; Green, 1985; Hurford, 1990). Errors were calculated assuming a Poisson distribution for both spontaneous and induced tracks (Green, 1981). When \( \chi^2 \) tests failed (\(<5\%\) (Galbraith and Laslett, 1985), mean ages and standard errors were calculated following procedures proposed by Green (1981).

The stability of spontaneous tracks in apatite is controlled mainly by the temperature history between 120°C and 60°C (e.g.: Laslett et al., 1987; Duddy et al., 1988), and subordinately by the chemical composition (e.g.: Gleadow and Duddy, 1981; Green et al., 1989). Although tracks in apatite significantly anneal between 120° and 60°C (PAZ, partial annealing zone), annealing at ambient temperatures is also occurring. Rapid cooling in volcanic rocks produces narrow distributions of tracks (standard deviations around 1.3 μm), with mean track lengths between 14 and 15 μm. If samples did not transit the PAZ quickly, the track length distributions reflect more annealing, and therefore reduction in age (Green, 1988). Assuming a geothermal gradient of 20–40°C/km the thermal stability of apatite will extend to a depth of 4–6 km.

For zircon, the temperature intervals recovered by FT are controlled by the temperature history between 175°C (e.g.: Harrison et al., 1979) and 230°C (e.g.: Hurford, 1983; Zaun and Wagner, 1985). Some more recent studies (e.g. Foster et al., 1996; Tagami et al., 1996; Tagami and Shimada, 1996) argue for higher limits of zircon PAZ, between 230° and 310°C. Sphene PAZ is also weakly constrained, its higher temperature stability being estimated at around 300°C (Gleadow and Lovering, 1978).

Microprobe analyses on selected apatite concentrates were carried out on a Jeol microprobe at the University of Salzburg (conditions during measurement: 15 kV, 20 mA, standard ZAF corrections).

Time correlations are based on calibrations of Gradstein et al. (1995) and Gradstein and Ogg (1996) for the Mesozoic and Cenozoic.

4. Results

4.1. Apatite

The apatite fission track data are summarised in Table 1, and the regional distribution of age data is shown in Fig. 4. Apatite was concentrated from basement and Mesozoic cover rocks of all tectonic units. Apatite ages range from 74 ± 4 to 7 ± 1 Ma. Single grain age distributions and the radial plot (Galbraith, 1990) were used for the graphical analysis of ages. In the radial plot, the x-axis represents the inverse value of the single grain age standard deviation. Thus, the ‘more precise’ ages plot on the right of this diagram. The y-axis represents the single ages normalised to a mean value. The diagram is designed in such a way that drawing a line trough the origin and a point, the single grain age can be read on the right side.

Plotting the apatite mean ages against elevation (Fig. 5a) no clear relationship is observed. Plotting the fission track age against mean track length (Fig. 5b), two age groups are observed. ‘Age group I’ shows older ages and a progressive decrease of the mean track length with age. If the ages from this group are plotted against the tectonostratigraphic position of the sample (Fig. 2), a decreasing age and mean track length trend from the hangingwall to the footwall is observed. The single grain age histograms from the samples form Getic nappe display maximum ages between 60 and 80 Ma (Fig. 6). Mean track lengths are 12.5–13.5 μm, with standard deviations around 1.5 μm. Length distributions are narrow and unimodal. Some samples from the Getic basement show younger ages, around 50 Ma. For example, sample AB35 shows a strong single grain age peak between 20 and 40 Ma (Fig. 6), mean track length around 10.5 μm and a much broader distribution (standard deviation around 2 μm).

Microprobe measurements of apatite for samples AB6, AB36, AB137 show 2.5–3 wt% F and 0.3 wt% Cl (Fig. 9). In order to model the data, the Laslett
### Table 1
Apatite fission track ages and horizontal confined length measurements

<table>
<thead>
<tr>
<th>Sample</th>
<th>Rock type and tectonic unit</th>
<th>Elev. (m)</th>
<th>Nr. of Spontaneous Induced Dosimeter Age</th>
<th>Spontaneous track density ( \rho_s (N_s) ) ( (10^6 \text{ cm}^2 \text{N}_s) )</th>
<th>Induced track density ( \rho_i (N_i) ) ( (10^6 \text{ cm}^2 \text{N}_i) )</th>
<th>Dosimeter ( \rho_d ) ( (10^6 \text{ cm}^2 \text{N}_d) )</th>
<th>Age ± 1σ (Ma)</th>
<th>( P (\chi^2) ) probability (%)</th>
<th>U (ppm)</th>
<th>Mean track length (µm)</th>
<th>Standard deviation (µm)</th>
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<td>1.470</td>
<td>0.7212</td>
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<td>&lt;5</td>
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<td>Liassic sandstone D</td>
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<td>0.6763</td>
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Table 1 (continued)

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<th>Sample</th>
<th>Rock type and tectonic unit</th>
<th>Elev. (m)</th>
<th>Nr. of crystals</th>
<th>Spontaneous track density $\rho_s \times 10^6 \text{cm}^{-2}$ (N_s)</th>
<th>Induced track density $\rho_i \times 10^6 \text{cm}^{-2}$ (N_i)</th>
<th>Dosimeter $\rho_d \times 10^6 \text{cm}^{-2}$ (N_d)</th>
<th>Age ± 1σ (Ma)</th>
<th>$P \ (\chi^2)$ probability (%)</th>
<th>U (ppm)</th>
<th>Mean track length (μm)</th>
<th>Standard deviation (μm)</th>
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<td>AB140</td>
<td>granite D</td>
<td>350</td>
<td>21</td>
<td>0.140 (88)</td>
<td>1.21 (762)</td>
<td>0.7212 (3534)</td>
<td>15 ± 2</td>
<td>42</td>
<td>13</td>
<td>13.82 ± 0.39</td>
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<tr>
<td>AB144</td>
<td>granite D</td>
<td>910</td>
<td>19</td>
<td>0.186 (106)</td>
<td>0.705 (401)</td>
<td>0.7212 (3534)</td>
<td>35 ± 4</td>
<td>38</td>
<td>8</td>
<td>13.70 ± 0.28</td>
<td>1.22</td>
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<tr>
<td>AB145</td>
<td>granite D</td>
<td>930</td>
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<td>0.172 (124)</td>
<td>0.616 (445)</td>
<td>0.7212 (3534)</td>
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<td>7</td>
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<td>770</td>
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<td>0.328 (248)</td>
<td>1.250 (945)</td>
<td>0.7212 (3534)</td>
<td>32 ± 2</td>
<td>51</td>
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<td>0.917 (424)</td>
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<td>0.437 (248)</td>
<td>2.060 (1171)</td>
<td>0.7212 (3534)</td>
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<tr>
<td>AB154</td>
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<td>260</td>
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<td>0.359 (276)</td>
<td>1.33 (1025)</td>
<td>0.7212 (3534)</td>
<td>36 ± 2</td>
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<td>15</td>
<td>12.90 ± 0.18</td>
<td>1.60</td>
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<tr>
<td>ABR3</td>
<td>chlorite schist</td>
<td>300</td>
<td>20</td>
<td>0.081 (127)</td>
<td>0.609 (919)</td>
<td>0.6763 (2865)</td>
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<td>12.09 ± 0.51</td>
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<td>AB171</td>
<td>gneiss D</td>
<td>200</td>
<td>10</td>
<td>0.209 (93)</td>
<td>1.350 (599)</td>
<td>0.6763 (2864)</td>
<td>19 ± 2</td>
<td>&lt;5</td>
<td>18</td>
<td>13.25 ± 0.51</td>
<td>2.05</td>
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</table>

$\rho = \text{track density} \times 10^6 \text{tracks cm}^{-2}$; (N) = number of tracks counted; s, i and d denote, respectively, spontaneous, induced tracks and tracks in the fluence monitor glass. Samples were dated using the external detector method (Hurford and Green, 1983). Ages were calculated using $\zeta = 346 ± 10$ for SRM612 (indicated with #) and $\zeta = 364.4 ± 10$ for CN5. Details on zeta calibration are available from the first author. When a sample failed the $\chi^2$ test ($P(\chi^2) < 5\%$), the mean age is given. For comparison the pooled age is also given in parentheses (Green, 1981). GN = Getic nappe, SN = Severin nappe, DB = Danubian basement.
model for Durango apatite was chosen (Laslett et al., 1987).

The samples from the sedimentary sequences, below the Getic nappe, show younger ages (Fig. 7). In the same time, confined track length distributions are positively skewed or bimodal. Taking into account that the stratigraphic age of samples from flysch (samples AB37, AB142) are Jurassic to Aptian, the data support total reset after deposition and, subsequent, cooling. Sample AB172 from the Liassic cover shows also total resetting (see also the radial plot, Fig. 7). The samples from the Danubian basement are almost from granitic rock (Fig. 7). Microprobe analyses of apatite from the granitic rocks...
(samples AB144, 149, 154) indicates F concentration around 3–3.5 wt% (Fig. 9). For modelling the time temperature path for these samples, the annealing model for fluor apatite (Crowley et al., 1991) is more appropriate.

The other age group (Fig. 5) shows younger ages with track lengths between 13 and 14 μm. Samples from this group are from rocks adjacent to brittle fault zones which show extensive hydrothermal alteration within cataclastic rocks and fault gouges. The ages scatter between 22 Ma and 7 Ma and are not correlated with the tectonostratigraphic position of the sample. Due to the low spontaneous track densities and the bad quality of apatite crystals (these are mostly altered or contain fluid inclusions) only few confined tracks could be measured. Nevertheless, the mean track lengths seem to decrease from the older ages (AB16, 22 ± 2 Ma) to the younger one (AB150, 7 ± 1) (Fig. 5, ‘age group II’ and Fig. 8). Standard deviations are larger for younger ages and lower for older ages (Table 1).

For both age groups, the mean track lengths are below 15 μm (Fig. 5b). This shows that the rocks underwent partial annealing and therefore age reduction. Thus, the ages (Fig. 5b) are not dating a geological event, they must be interpreted in terms of a multiple cooling–reheating history within a temperature interval between 120°C and 60°C.

4.2. Zircon and sphene

Zircon fission track ages were determined for eleven samples and sphene for three sample. The results are summarised in Table 2 and the regional distribution is given in Fig. 4.

The zircons from the Cretaceous flysch sandstones (ZAB37, ZAB40, ZAB141) and from Liassic cover (ZAB139) give the oldest ages of 188 ± 19, 190±23, 173±20 and 220±27 Ma. Single grain ages for both Liassic sandstones and flysch (Tithonian–Neocomian stratigraphic age of 150–130 Ma) are in proportion of 80% older than the depositional age of the sediments suggesting that the samples have experienced minor partial annealing since deposition (see also Fig. 10). Taking into account that the thermal overprint of the flysch and Liassic deposits remained within very low-grade conditions, the data support cooling of the area adjacent to the rift zone, cooling
Table 2
Zircon and sphene fission track ages

<table>
<thead>
<tr>
<th>Sample</th>
<th>Rock type</th>
<th>Elev. (m)</th>
<th>Number of grains</th>
<th>Spontaneous track density $\rho_s \times 10^6$ cm$^{-2}$</th>
<th>Induced track density $\rho_i \times 10^6$ cm$^{-2}$</th>
<th>Dosimeter Age (Ma)</th>
<th>$P$ ($\chi^2$) (%)</th>
<th>$U$ (ppm)</th>
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<td>112 ± 12</td>
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<td>(356)</td>
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<td>4.126</td>
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<td>0.323</td>
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<td>97 ± 11</td>
<td>90</td>
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<td>(115)</td>
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Notation as in Table 1. Age determinations: for zircon using $\xi = 113 ± 8$ for CN2; for sphene using $\xi = 354 ± 10$ for CN5.
which mainly pre-dated the opening of the Severin rift and the formation of the flysch deposits.

In the klippen region, zircon samples from the Getic and Danubian basement units yield ages between 115 and 110 Ma, indicating cooling of the basement under 200°C during that time. Single grain ages show a large spread and suggest that, prior to Aptian cooling, the maximum temperature of these samples never significantly exceeded 200–250°C.

Two sphene samples (SAB 166, SAB167) show slightly younger ages (95 ± 6, 93 ± 12 Ma). For the reason that the sphene grains present strong and irregular uranium zoning patterns, these ages should be regarded as relative values.

A zircon sample from the Jiu Valley and a sphene sample from the Retezat Mountains, both from mylonitic rocks, show ages of 67 ± 7, 69 ± 5 Ma, respectively. Single grain age distributions show a relatively narrow pattern (Fig. 10). The data suggest complete annealing at temperature in excess of 240–300°C, supporting the fact that in the northern Danubian window the Late Cretaceous metamorphism reached greenschist to epidote-amphibolite facies conditions.

One zircon sample (ZAB 145) from a fault zone east of the Cerna fault zone has an age of 42 ± 5 Ma. This age could be due to a Paleocene activation of the shear zone.

5. Discussion and conclusions

Based on the stratigraphic, geochronologic and metamorphic arguments already presented, we will
Fig. 7. Apatite fission track data from sedimentary sequences and Danubian basement. For explanation of graphics, see Fig. 6.

Fig. 8. Apatite fission track data from the fault zones. For explanation of graphics, see Fig. 6.
discuss a possible model for the geological evolution of the South Carpathians.

5.1. Jurassic extension

The zircon fission track ages from flysch and Lias-
sic cover indicate that regional exhumation mainly
pre-dated the opening of the Severin rift. According
to the Buck model (Buck, 1991), these data support
an active origin for the Severin rift and the involve-
ment of a mantle plume in generating extension. Consid-
ering a pre- to syn-rift geothermal gradient of
40°C/km, to exhum the zircon PAZ from the area
bordering the rift zone the total amount of erosion
should be in the order of 5–6 km.

The regional distribution of the syn-rift sediments
indicates an asymmetric lithospheric extension, with
the Danubian realm acting as hangingwall and the
Getic realm as footwall (Fig. 12a). An Early Jurassic
ductile event is also supported by an 40Ar/39Ar
plateau-like age pattern on mica concentrate from a
basement gneiss within the Getic nappe (Fig. 4).

5.2. Early Cretaceous collision

During the Early Cretaceous (Neocomian–
Aptian), the oceanic crust exposed within the Severin
unit, was totally subducted. Synchronously the rift
deposits were deformed in an accretionary wedge
between Danubian and Getic domains and partially
overridden by the Getic crystalline (Fig. 12b). Cool-
ing of both basement units under 200°C during
continent–continent collision is recorded by the zir-
con and sphene ages.

During the Cenomanian, the sedimentation area
shifted eastward, the sedimentary record showing
progressive subsidence of the basin. The Arjana
nappe was partially covered by Late Cretaceous de-
posits (Figs. 2 and 3). This is in contrast with the
Severin nappe which lies on the top of Late Cre-
taeous sequences. Meanwhile the wildflysch contains
olistoliths from the Severin nappe which played the
role of a source area at that time.

5.3. Late Cretaceous nappe stacking

During the late Campanian to possibly early
Maastrichtian, footwall propagation of the thrust
front was responsible for the emplacement of Sev-
erin nappe on the top of the wildflysch (Fig. 12c).
Progressive involvement of the Danubian basement
produced stronger crustal thickening in the presently
northern Danubian window, consequently Late Cre-
taceous tectonic exhumation (e.g.: England and Mol-
nar, 1990; Stüwe and Barr, 1998) were higher north-
ward to the klippen area. This fact is also supported
by the formation of the Hateg basin adjacent to the
Retezat Mountains. This basin is filled mainly with
thick early to late Maastrichtian deposits (Fig. 12c).

As a result of different amounts of denudation
between the two areas, zircon and sphene fission track data from the klippen region record Early Cretaceous ages, in contrast with the data from the Jiu Valley or Barbat Valley (Retezat Mountains) which depict Maastrichtian ages (67 ± 7 Ma and 69 ± 5 Ma, respectively) thus indicating exhumation of a deeper crustal level at that time.

No data are available for the palaeo-geothermal gradient, but the estimated present geothermal gradient varies between 15°C and 20°C/km (Demetrescu and Andreescu, 1994). Consequently for the further interpretation of fission track data we adopt a geothermal gradient of 20°C/km.

The thermal histories for apatite age and length distributions were modelled using a program developed by R. Ketcham (Ketcham and Donelick, in prep.).

Apatite fission track data from the Getic basement (Fig. 11) show cooling under 120°C starting in Maastrichtian times, cooling most probably associated with exhumation after the nappe stacking. During the Palaeogene, the whole nappe pile cooled progressively through the apatite partial annealing zone. This is the reason why the apatite ages (Fig. 5b) show a regional decreasing trend from hangingwall to footwall units. Assuming 20°C/km, the modelled data indicate that in the klippen area, 6–7 km of crust was removed since the Late Cretaceous.

5.4. Oligocene to Sarmatian collision with the Moesian platform

The data from the Danubian nappe complex show that the final cooling of the footwall units under 60°C of this unit occurred during an accelerated
phase of erosion at the beginning of Miocene time. These results fit together with the age of the first Neogene sediments from the Orsova and Caransebes basins, sediments which overstep both Danubian and Getic domains. So far we have shown, throughout the region, that Palaeogene deposits are still underlain by Getic basement. For example, the Chattian deposits from the Petrosani basin are underlain by 1–2 km thick Getic crystalline and by ca. 1 km thick Jurassic to Cretaceous Danubian cover. In contrast, the Aquitanian sediments from the Orsova basin are overstepping the Danubian as well as the Getic realm. Therefore we can conclude that ca. 2 to 4 km have been removed from Chattian to Aquitanian times to exhume the Danubian basement. This event was associated with the starting of a transpressive collision (Ratschbacher et al., 1993; Neubauer et al., 1994) between the South Carpathian orogen and the Moesian platform. In the study area, the deformation of the South Carpathian orogen started with large-scale folding and wrench faulting and continued to subsequent Miocene basin formation. During the
large-scale folding the antiforms were subjected to relative faster erosion and cooling than the adjacent synforms. The late large-scale folding brought the Danubian basement with younger apatite ages in a higher topographic position than the overlying Getic nappe with older ages (Fig. 2). Thus, in the klippen region, the apatite FT ages are no more correlated with the elevation (Fig. 5a).

At a regional scale, this event can be related to the Late Oligocene to Early Miocene convergence between the Adria (and Africa) plate and Europe. At that time, the escape tectonics was initiated by the collision of the Adriatic promontory with Europe (e.g.: Royden et al., 1982; Csontos et al., 1992). Deformation and convergence occurred mainly in the West and East Carpathians and, as discussed before, also in the South Carpathians.

The apatite ages and confined track length distribution support the idea that in the study area the Early Miocene–Sarmatian cover had a much larger distribution than it has now. During burial under Early Miocene to Sarmatian deposits, temperature rose ca. 20° to 40°C in the subjacent units (Fig. 11). Moreover, we can relate the cessation of sedimentation at the level of the Sarmatian in the Caransebes, Orsova and Petrosani basins with the final stacking of the South Carpathian on the top of the Moesian platform. To the east, the thrust front was progressively buried under post-tectonic Sarmatian to Quaternary deposits.

A second group of ages comes from brittle fault zones which show strong hydrothermal alteration. The rocks within the fault zones yield much younger ages indicating that the samples were exposed during fluid flow to higher temperatures than the adjacent rocks which remained cooler during the Cenozoic. Apatite ages are totally reset within this zones, (e.g.: 22 ± 2, 7 ± 1 Ma, Fig. 8) and partially reset adjacent to them (e.g.: 50 ± 4 Ma, Fig. 6).

Thus the fission track ages from these zones are interpreted to indicate the existence and decay of a hydrothermal convection system. Progressive overriding of the Moesian platform caused regional cessation of the fluid flow. Hot springs which produced the thermal anomalies can still be found along the Cerna fault (Airinei et al., 1976).

An apatite age from the Retezat area (Barbat Valley) suggests a different history of the Danubian basement during the Cenozoic. The sample comes from a rock which does not show hydrothermal alteration. The young age (17 ± 2 Ma) and the length distribution indicate much stronger Oligocene regional exhumation of the northern part of the Danubian window than in the klippen area. This can be due to the fact that collision with the Moesian platform, shortening and continental thickening were much stronger in the northern part (the subsurface thrust front strikes E–W) than in the eastern part (the subsurface thrust front strikes NE–SW).

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