Multi-phase exhumation history of a low-topography orogen: A compilation of low-temperature thermochronological data for the Romanian Carpathians

6.1 Introduction

The exhumation of low-topography orogens such as the Carpathians is generally difficult to constrain because of the relatively low amounts of shortening-related uplift. Because bulk exhumation of the upper plate is in the order of <5–6 km, metamorphism in the core of low-topography orogens pre-dates the main orogenic phase (generally pre-Alpine). Such small amounts of bulk exhumation recorded by thermochronological studies are often observed in highly arcuate thrust belts, which are associated with back-arc extension in the overriding plate and rates of subduction that are higher than plate convergence rates. Collision, if present, is generally defined to be early or incomplete compared to other orogens considered to have reached steady-state equilibrium [e.g. Royden, 1993]. This is in contrast with high-convergence orogens like the Alps, for which observation and modelling studies suggest that orogenic steady-state largely involves exhumation in the orogenic core. For such types of orogens, hinterland exhumation is associated with crustal-scale backthrusting (i.e. retro-shears) exposing large metamorphic complexes at the surface [e.g. Beaumont et al., 1994; Schmid et al., 1996; Willett and Brandon, 2002]. In low-topography orogens, however, retro-shears are generally less obvious and the geodynamic processes are usually driven by mechanisms in the lower plate, such as slab retreat or roll-back [e.g. Royden and Burchfiel, 1989], slab detachment [e.g. Wortel and Spakman, 2000] or delamination [Sacks and Secor, 1990]. Exhumation of the orogenic core in the absence of retro-shears,
the formation of a back-arc basin and/or exhumation resulting from syn-orogenic extension are common in Mediterranean-type mountain belts, which are difficult to explain by the simple retro-shear model (e.g. Apennines, Dinarides, Calabria, Betics, Hellenides) [Jolivet and Faccenna, 2000; Brun and Faccenna, 2008].

The Carpathians-Dinarides-Balkans system (Figure 6.1a) is a typical example of a low-topography orogen. This $800 \times 600$ km orogenic belt formed in response to Late Jurassic–Miocene compression linked to the evolution of two oceanic realms, the Alpine Tethys and the Neotethys [e.g. Boccoletti et al., 1974; Schmid et al., 2008; Robertson et al., 2009]. The present-day appearance of the Carpathians, Dinarides and Balkans as separate mountain chains is the result of a superposed Miocene extensional phase [e.g. Ustaszewski et al., 2008]. The Miocene extensional phase created the intra-montane Pannonian and Transylvania basins, starting as back-arc basins at $\sim 20$ Ma due to the rapid roll-back of the external Carpathian slab [Cloetingh et al., 2006; Horváth et al., 2006] (Figure 6.1). Studies dealing with the build-up of individual segments of the Carpathians-Dinarides-Balkans system commonly discuss these mountain belts individually and, therefore, lack integration across basins.

One example is the interpretation of the Late Cretaceous–Paleogene contractional episodes in the Apuseni Mountains, which are commonly linked to the evolution of the external Carpathians or areas in between [e.g. Bleahu et al., 1981; Balintoni, 1994; De Broucker et al., 1998; Krézek and Bally, 2006], even though no significant tectonic events are defined for those areas during that period. However, a retro-deformation of the large-scale extension and collapse of the Pannonian Basin to a situation at $\sim 20$ Ma ago (Figure 6.1b) [Ustaszewski et al., 2008] shows that the Sava subduction zone in the Dinarides [Pamić et al., 2002a; Schmid et al., 2008] is at a similar distance from the Apuseni Mountains as the Ceahlău-Severin subduction zone at the exterior of the Carpathians. The former is the result of the Cretaceous–Eocene convergence between the already sutured Tisza-Dacia block in the upper plate and the Adria-derived thrust sheets in a lower plate position (Figure 6.1a) [Schmid et al., 2008]. Recent kinematic and absolute age datings of deformation structures associated with the Sava Zone have demonstrated substantial shortening and a significant metamorphic event during the latest Cretaceous–Eocene (peak metamorphism at $\sim 40$ Ma), associated with top WSW-ward directions of thrusting, recorded in the distal Adriatic thrust sheets at the contact with the Sava Zone [Schefer, 2010]. This event is coeval with the rotation of the South Carpathians around the Moesian indentor (Figure 6.1b) [Ratschbacher et al., 1993], which took place during latest Cretaceous–Eocene orogen parallel extension of the Danubian units [Bojar et al., 1998; Willingshofer et al., 2001; Fügenschuh and Schmid, 2005] and the subsequent 35 km early Oligocene dextral offset of the Cerna fault [Berza and Draganescu, 1988]. These events are coeval with exhumation recorded at the local scale of the Apuseni Mountains, where the previously assumed Miocene exhumation [Sanders, 1998] is proven to be of latest Cretaceous–Paleogene age (see Chapter 3).

Furthermore, the influence of lower plate rheology on the Miocene subduction and soft-collision in the East Carpathians (strong East European / Scythian Platform) and SE Carpathians (weaker Moesian Plate; Figure 6.1a) on the mechanics of hinterland deformation is still unclear in terms of quantifying uplift and erosion
Figure 6.1: a) Major tectonic units of the Alps, Carpathians and Dinarides (ALCADI) [Simplified after Schmid et al., 2008]. b) Restoration of tectonic units in the ALCADI domain for the early Miocene. Colours and patterns of tectonic units correspond to those in Figure 6.1a [Simplified after Ustaszewski et al., 2008].
through time. This includes the speculated presence of the intermediate Danubian block in the SE Carpathians, as a subsurface continuation of the same unit exposed in the South Carpathians [e.g. Visarion et al., 1978; Iancu et al., 2005]. The presence of an intermediate Danubian block in the SE Carpathians implies a shift from the Ceahlău-Severin subduction zone to mainly continental subduction of the distal parts of the foreland platform, grouped under the generic term of “Carpathian embayment” (Figure 6.1b).

In more detail, one important objective is deriving spatial and temporal constraints on post-collisional deformations (post 11-9 Ma) in the SE Carpathians. These are particularly relevant when discussing the interplay between tectonics and a large sea-level drop at around 6 Ma linked to the Paratethys Messinian Salinity Crisis [e.g. Gillet et al., 2007]. In the SE Carpathians, there is an apparent discrepancy between the ∼6–5 Ma onset of exhumation (latest Miocene–Pliocene) as recorded by thermochronological and provenance studies [Sanders, 1998; Seghedi et al., 2004; Panaiotu et al., 2007] (also see Chapter 4), and the observed ∼3–2 Ma onset of tectonic inversion (latest Pliocene–Quaternary) as indicated by tectonic structures in the field and seismic sections [e.g. Hippolyte and Sândulescu, 1996; Tărăpoancă et al., 2003; Leever et al., 2006].

The small thermal overprint in low-topography orogens prevents an optimal discrimination of individual exhumation events from the long-term patterns of mountain decay using thermochronology [Reiners and Brandon, 2006]. Only the combination of several low-temperature thermochronometers, such as apatite fission track (AFT) and apatite (U-Th)/He thermochronology (AHe), can provide enough details on individual (tectonic) exhumation events. With the combined use of several low-temperature thermochronometers and structural analyses, many of the uncertainties on the evolution of the Romanian Carpathians and geodynamic processes driving it were discussed and partly resolved in this study (Chapters 3 to 5).

This chapter integrates low-temperature thermochronological results of the Apuseni Mountains (Chapter 3), SE Carpathians (Chapter 4) and East and South Carpathians (Chapter 5) in order to reconstruct the large-scale exhumation history of the Romanian Carpathians. For this reconstruction, a large regional thermochronological database is compiled, integrating the results from this thesis with previously published thermochronological data [Sanders, 1998; Bojar et al., 1998; Schmid et al., 1998; Sanders et al., 1999; Willingshofer, 2000; Willingshofer et al., 2001; Schuller, 2004; Fügenschuh and Schmid, 2005; Moser et al., 2005; Gröger, 2006; Gröger et al., 2008; Necea, 2010].

The compiled database contains 229 apatite fission track (AFT) analyses, 112 zircon fission track (ZFT) analyses and 54 apatite (U-Th)/He (AHe) analyses (Figure 6.2). Integration of all ZFT, AFT and AHe data allows reconstructing the link between exhumation patterns in the Carpathian orogen and tectonic deformation episodes from the Cretaceous to present-day at a higher resolution. It allows separation of different tectonic episodes, in contrast to previous bulk exhumation studies [e.g. Sanders, 1998]. In particular, the addition of the AHe thermochronometer provides new constraints on the recent post-collisional exhumation (post 11-9 Ma) and allows the separation between syn- and post-collisional exhumation events.
EXHUMATION HISTORY OF THE ROMANIAN CARPATHIANS

6.2 Brief overview of the main tectonic units and deformation events at regional scale

The Carpathians are the result of a Triassic to Tertiary evolution of four continental blocks and intervening oceans [e.g. Schmid et al., 2008]. At the interior of the Romanian Carpathians, these blocks are the Europe-derived Tisza and Dacia blocks to the S and SW, and the Adria-derived ALCAPA block presently situated N and NW-wards (Figure 6.1a). The Tisza and Dacia blocks were separated by the East Vardar Ocean, which closed gradually during Late Jurassic–Cretaceous times followed by continental collision during the late Early to Late Cretaceous (Figures 6.1a and 6.3) [e.g. Schmid et al., 2008]. The welded Tisza-Dacia block was subsequently sutured with the ALCAPA block during the early Miocene (Figures 6.1a and 6.3). At the exterior, the Carpathians were ultimately thrust over the (Paleozoic) European/Scythian/Moesian continental foreland during Cretaceous–Miocene contractional episodes, closing the Ceahlău-Severin Ocean and the Carpathian embayment (Figures 6.1 and 6.3) [e.g. Oszczypko, 2006].

On a more regional scale, the evolution of two separate groups of oceanic basins that opened during the break-up of Pangea is responsible for the orogenic systems observed in the Alps-Carpathians-Dinarides (ALCADI) realm (Figure 6.1a). The “Alpine Tethys” is a collective name for all oceanic realms that started to open during the Middle Jurassic, in a direct kinematic link to the sea-floor spreading in the Central Atlantic [Favre and Stampfli, 1992]. The term “Neotethys” is used for all oceanic realms located in an area southeast of the Alpine Tethys and the future Western Alps that opened during Triassic–Middle Jurassic times associated with the closure of the Paleotethys, which still separated Gondwana and Laurussia at the end of the Variscan orogeny [e.g. Stampfli and Borel, 2004]. Parts of the Neotethys (the Eastern and Western Vardar Ophiolitic units; Figure 6.1a) opened during Triassic–Jurassic times and were obducted during Late Jurassic times due to the opening of the Central Atlantic [Schmid et al., 2008]. Other parts of the Neotethys (e.g. the Sava Zone, Figure 6.1a) opened later during the Cretaceous and were directly linked with the Alpine Tethys [Schmid et al., 2008]. Among the Alpine Tethys branches, the Ceahlău-Severin Ocean (Figure 6.1a) opened in Middle–Late Jurassic times at the exterior of the Carpathians [Schmid et al., 2008].

Following the opening of these oceanic domains in the ALCADI realm, subduction/obduction and subsequent continental collision is recorded from the Late Jurassic onwards, when large ophiolitic sheets were emplaced in the Dinarides (Figure 6.1a) [e.g. Dimitrijević, 1997]. Starting with the late Early Cretaceous, subduction of these oceanic domains and subsequent continental collision is recorded in all ALCADI domains. The late Early Cretaceous event is observed by substantial shortening and metamorphism in the composite units of the Dinarides,

**Figure 6.2 (facing page):** Sample locations for low-temperature thermochronological data of the Romanian Carpathians with colour codes indicating references used for this overview (see legend). a) ZFT. b) AFT. c) AHe. The geodetic projection of this figure as well as Figures 6.4, 6.5 and 6.7 is Romanian Stereo 70.
Figure 6.3: Ages of main activity along major tectonic contacts in the Alps, Carpathians and Dinarides colour-coded in six time slices [redrawn after Schmid et al., 2008]. Although some contacts were repeatedly active, only the age of the main deformation is shown.

top-E facing obduction of the East Vardar Ophiolites over the Transylvanian- and East Carpathian basement, and internal shortening and thrusting of the latter over the Ceahlău-Severin sediments (Figure 6.3). This was subsequently followed by an isolated phase of shortening in the Apuseni Mountains during Turonian times, and by post-“orogenic” subsidence associated with a partly extensional phase in the Apuseni Mountains and South Carpathians during latest Turonian–middle Senonian times, known as the “Gosau-phase” [e.g. Sândulescu, 1988; Willingshofer et al., 1999; Csontos and Vörös, 2004; Schmid et al., 2008].

Starting with the latest Cretaceous, renewed contraction is observed in all ALCADI units (Figure 6.3). In the East and South Carpathians, this shortening phase is generally restricted to late Campanian–early Maastrichtian times, indicated by an upper Maastrichtian post-tectonic cover [Sândulescu, 1984; Berza, 1994]. In the Dinarides, this is the main phase of shortening (i.e. the “Dinaridic” phase) related to subduction and closure of the Sava Zone (Alpine Tethys, Figure 6.1a), which is generally interpreted to be of latest Cretaceous–Eocene age [e.g. Dimitrijević, 1997]. In the area situated in between the East/South Carpathians and the Dinarides, i.e. the Apuseni Mountains, the timing of this shortening event is generally unclear and is suggested to be of latest Cretaceous or latest Cretaceous–Eocene age [Lupu, 1984; Sândulescu, 1984; Pană and Erdmer, 1994; Balintoni, 1994]. No significant post-Eocene contractional deformation is described in the Dinarides in relationship to the evolution of the Sava Zone. From the early Miocene onward, the Dinarides were affected by a significant extensional collapse coeval with back-arc extension leading to the formation of the
Pannonian Basin [Schefer, 2010]. This collapse is partly coeval with the large-scale rotations of the Tisza-Dacia block around Moesia during the Oligocene–lower Miocene [Ratschbacher et al., 1993; Fügenschuh and Schmid, 2005] and was terminated by the Carpathian collision during the middle–late Miocene [e.g. Horváth et al., 2006]. In Pliocene–Quaternary times, the evolution of the ALCADI domain is dominated by the Adriatic push. This is observed by significant out-of-sequence deformation in the Dinarides and present-day strain measurements [e.g. Pinter et al., 2005; Bennett et al., 2008], and by the localised deformation in the SE Carpathians, generally related to the post-collisional evolution of the Vrancea slab [e.g. Matenco et al., 2007].

### 6.3 Data sources and methodological constraints

An overview of low-temperature thermochronological datasets for the Romanian Carpathians is presented in Figure 6.2. The ZFT, AFT and AHe thermochronometers have relatively low closure temperatures (∼230°C [Brandon and Vance, 1992; Brandon et al., 1998], ∼110 ± 10°C [Gleadow and Duddy, 1981] and ∼75 ± 5°C [Wolf et al., 1996], respectively), thus allowing to study the exhumation of the upper few kilometres of the crust. 28 AFT and 29 AHe ages from this study were combined with 112 ZFT ages available in 7 other publications, 203 AFT ages derived from 11 other publications, and 25 AHe ages from 3 other publications (Figure 6.2). In general, datasets that overlap geographically are compatible with each other. A small number of age versus elevation profiles have also been published (see Figure 6.2 for locations). The data quality varies within the datasets (see section 6.3.1) and samples cover a large range of elevations (see section 6.3.2). These different parameters should be taken into account before any conclusions on age patterns can be drawn.

#### 6.3.1 Quality and homogeneity of the data

All fission track ages included in the database were obtained with the external detector method [Hurford and Green, 1982]. Additional information collected for each sample (where reported): 1) geographic coordinates, 2) elevation, 3) depositional/intrusion age (for sediments/volcanics), 4) lithology and structural unit, 5) homogeneity of single grain ages for FT samples (Chi-squared probability and/or dispersion) and AHe samples (spread in single grain measurements), 6) whether FT ages are pooled or central ages, 7) mean track lengths (MTLs) for FT samples, and 8) thermal history models.

Some publications contained more than one age at a given geographic location in the same tectonic position. These points were discarded from the study to prevent interpolation problems. Sample locations from local publications without geographic coordinates were plotted carefully by reading them off accompanying maps. Note that the errors on the exact sample locations in general are below the resolution of interpolated maps (Figures 6.4 and 6.5). All sample locations were double-checked so that they respect their position relative to major tectonic structures.

Only reset ages (i.e. thermochronological ages younger than the depositional
For the ZFT and AFT data, basement samples in general show homogeneous FT ages ($P(\chi^2)>5\%$ and dispersion$<30\%$). For these samples, the total reset central or pooled age has been taken into account for data interpolation. Many clastic sediment samples, especially from the East and SE Carpathians, do not pass the $\chi^2$-test at 5% and have dispersions higher than 30%, indicating heterogeneous FT ages [Galbraith, 1981; Brandon, 1992]. These samples are generally characterised by a combination of non-reset and reset age populations (e.g. see Chapter 4), indicating partial annealing after sedimentation. For these samples, the minimum reset age population, which represents a minimum estimate for the onset of cooling [Galbraith and Laslett, 1993; Brandon et al., 1998], is taken into account for the age distribution maps.

AHe ages that are more than $2\sigma$ error older than the AFT age were discarded in the age distribution map. AHe ages of samples that did not replicate within uncertainty were discarded, or the youngest single grain ages were taken into account (following the original authors’ interpretation).

### 6.3.2 Elevation

The samples included in this overview represent a wide range of elevations varying from 140 to 2600 m with an average overall elevation of 727 ± 425 m. Because of the general systematic increase of temperature with depth in the crust, sampling over a range of elevations within a small area can provide important information on the geothermal gradient. Generally, samples at higher elevations within a vertical profile will have cooled earlier through the partial retention zone (PRZ) and will therefore have older ages than samples at lower elevations [e.g. Braun et al., 2006]. However, age-elevation relationships for the Apuseni Mountains and the South Carpathians show that ages largely overlap within error regardless of sample elevation (see Figure 3.5 in Chapter 3 and Figure 5.5 in Chapter 5). This was also reported for the South and East Carpathians by Moser et al. [2005] and Gröger [2006], respectively. These vertical age distributions suggest that regional sets of samples experienced rapid cooling events and apparent ages can be interpreted as cooling ages.

### 6.3.3 Maps of interpolated thermochronological ages

Three interpolated maps were created of ZFT ($n = 82$), AFT ($n = 184$) and AHe ages ($n = 41$) derived from the thermochronological database following the criteria outlined in section 6.3.1 (Figures 6.4b-d). Major faults characteristic for the evolution of the Romanian Carpathians [e.g. Schmid et al., 2008] were included as tectonic boundaries, acting as discontinuities during interpolation (Figures 6.3 and 6.4). Contouring was done using a Minimum Curvature interpolation method [Briggs, 1974; Lam, 1983], generating the smoothest possible surface while maintaining data accuracy.

Age data are colour coded following the main tectonic episodes modified after Schmid et al. [2008], identifying seven main age groups (Figures 6.4b-d). These
are 1) Early Cretaceous (>100 Ma), 2) Late Cretaceous (100–78 Ma), 3) latest Cretaceous–middle Eocene (78–41 Ma), 4) middle Eocene–Oligocene (41–23 Ma), 5) Miocene (23–5 Ma), 6) early–middle Pliocene (5–3 Ma) and 7) latest Pliocene–Quaternary (3–0 Ma).

6.3.4 Reconstruction of exhumation and burial in the Romanian Carpathians

Available thermochronological ages, time-temperature models, structural and sedimentological data were integrated to reconstruct a series of maps with estimated amounts of exhumation and burial in the Romanian Carpathians from the latest Cretaceous to present-day (Figure 6.5). Miocene to present-day syn- and post-collisional exhumation histories were contradictory for some areas in the East and SE Carpathians, i.e. time-temperature histories modelled for AFT indicate rapid cooling for the last few Ma whereas age-elevation relations and AFT/AHe age pairs do not confirm this recent cooling [e.g. see Gröger, 2006; Necea, 2010]. These studies used the Laslett et al. [1987] annealing model for thermal modelling, which is known to introduce overestimations of late-stage cooling as a modelling artefact [Ketcham et al., 1999; 2000]. Thus, for Miocene to present-day times, the age-elevation relations and AFT/AHe age pairs were preferred over modelled time-temperature histories in assessing and separating syn- and post-collisional cooling. For each tectonic episode, active structures that play a main role in the system’s evolution were included as tectonic boundaries and are highlighted in Figures 6.5a-f. Exhumation/burial estimates were calculated assuming an average surface temperature of 10°C and a constant (paleo-) geothermal gradient of 25 ± 10°C/km for the Apuseni Mountains and South Carpathians [e.g. Bojar et al., 1998; Schmid et al., 1998] and 20 ± 5°C/km for the East and SE Carpathians [e.g. Sanders, 1998; Gröger, 2006; Necea, 2010]. These values are in accordance with present-day heat flow data [e.g. Veliciu and Visarion, 1984; Dövenyi and Horváth, 1988; Demetreacu and Andreescu, 1994; Demetreacu et al., 2007]. However, a constant paleo-geothermal gradient should be regarded as estimation rather than derived or modelled values. For example, given the significant Miocene re-heating of the Pannonian lithosphere during back-arc extension, the paleo-geothermal gradient in the Apuseni Mountains might have been lower during Cretaceous–Paleogene times (also see Chapter 2). The uncertainties on cooling estimates and paleo-geothermal gradients generally result in propagated errors on exhumation/burial estimates in the order of ~25% (e.g. SE Carpathians; also see Chapter 4), reaching up to ~45% (e.g. Apuseni Mountains; also see Chapter 3).

Overall, the series of maps aims to give both an overview and a quantification of the main areas subjected to exhumation and burial, and their evolution in time. Three regional cross-sections were built and used to further illustrate the main mechanisms acting during the evolution of the Romanian Carpathians (Figure 6.6). Burial estimates for the Transylvania Basin and Focşani Basin were derived from and are compared with sedimentation and subsidence estimates obtained from seismic and sedimentological studies [Tărăpoancă et al., 2003; Krészek and Bally, 2006; Leever et al., 2006].
6.4 Main features of thermochronological age patterns

On the overall, the patterns derived from thermochronological ages are representative for the Cretaceous and Miocene exhumation episodes that took place along two subduction/obduction and continental collision zones in the Romanian Carpathians (i.e. the East Vardar and Cealhău-Severin zones, Figure 6.4). In more detail, the maps demonstrate a separation into different age groups, which deserve a more detailed explanation.

6.4.1 ZFT age patterns

The ZFT age distribution map shows mainly Late Cretaceous to Paleocene ZFT ages for the Apuseni Mountains (Figure 6.4b). Oldest ages are recorded for basement samples from the Bihor Dome (95–86 Ma). The latest Cretaceous–Paleocene ZFT ages (70–64 Ma) are recorded for magmatic Banatites and are indistinguishable from Re-Os and $^{40}\text{Ar}/^{39}\text{Ar}$ amphibole and biotite ages [Wiesinger et al., 2005; Zimmerman et al., 2008]. These ages can be interpreted as magmatic cooling (also see Chapter 3).

In the northern part of the East Carpathians, Early to Late Cretaceous ZFT ages are recorded for the Preluca Massif and the Bucovinian and Sub-Bucovinian nappes, whereas younger latest Cretaceous to Paleocene ages are recorded for the Infra-Bucovinian nappe (Figure 6.4b). The ZFT age pattern suggests a possible NNW-SSE striking fold-like geometry following the orientation of the Bucovinian thrust, with the youngest ZFT ages in the core of the antiform and ages increasing towards the flanks. The youngest and structurally lowest nappe of the Bucovinian nappe stack (i.e. the Infra-Bucovinian nappe) records the youngest ZFT ages. No ZFT data are available for the central and southern parts of the East Carpathians, and the SE Carpathians.

In the South Carpathians, ZFT ages in the Getic nappes north of the Getic detachment [senso Schmid et al., 1998] range from Early Cretaceous to Paleogene (Figure 6.4b). Directly south of the Getic detachment, the Danubian nappes record younger latest Cretaceous to Oligocene ZFT ages decreasing from west (latest Cretaceous–middle Eocene) to east (middle Eocene–Oligocene). Within the Danubian window, ZFT ages increase to Early Cretaceous towards the south. The Getic nappes in this southern part and in the Godeanu Klippen record Late Cretaceous ZFT ages.

Figure 6.4 (facing page): a) Tectonic sketch of the Romanian Carpathians. Solid black lines indicate the locations of cross-sections in Figure 6.6. b-d) Patterns of thermochronological data in the Romanian Carpathians. Major faults characteristic for the evolution of the Romanian Carpathians [e.g. Schmid et al., 2008] were used as tectonic boundaries. Areas for which data are absent are blanked (the Neogene basins, i.e. the Pannonian and Transylvania basins and the foredeep). b) ZFT pattern. Out of 112 available data, 82 total reset ages are selected (see text for explanation). c) AFT data pattern. Out of 229 available data, 184 total reset ages or age populations are selected (see text for explanation). d) AHe data pattern. Out of 54 available data, 41 reset ages are selected (see text for explanation).
Figure 6.4: Continued
6.4.2 AFT age patterns

The AFT age distribution map (Figure 6.4c) shows Cretaceous to middle Eocene AFT ages for the Apuseni Mountains and Preluca Massif. Late Cretaceous AFT ages are restricted to the centre of the Bihor Dome, whereas AFT ages decrease outwards to latest Cretaceous–middle Eocene. Thus, on the overall, the AFT age pattern reflects the domal structure of the Bihor Autochthon and the structurally overlying Codru nappes with the oldest ages located in the centre of the dome. Younger middle Eocene–Oligocene AFT ages seem to be limited by two basement faults, the Puini and Mezeş thrusts. Miocene ages are recorded for the Neogene volcanics in the WNW-ESE striking Zărand and Beius basins and overlap their volcanic emplacement age.

The Getic and Supragetic nappes of the South Carpathians generally show latest Cretaceous–middle Eocene AFT ages, with some younger Oligocene ages in the northern part of the Făgăraș Mountains (Figure 6.4c). South of the Getic detachment, the Dambian nappes record latest Cretaceous to Miocene AFT ages decreasing from west (latest Cretaceous–middle Eocene) to east (Miocene), which is similar to the age pattern observed in the ZFT data. Towards the south, Oligocene to Miocene AFT ages in the Danubian window appear to be limited by the Cerna-Jiu fault.

The AFT age distribution in the East and SE Carpathians clearly reveals a major difference between the two areas (Figure 6.4c). In the East Carpathians, Miocene AFT ages are widespread, both in the upper Dacia block (Bucovinian nappes), the Ceahlău unit, and in the external sedimentary wedge of the subducting East European foreland. In the SE Carpathians, however, Miocene AFT ages seem to be limited to the Ceahlău unit and the thin-skinned nappes overlying the Moesian platform. The upper Dacia block does not show a Miocene overprint and records older Early Cretaceous–Eocene AFT ages. In the SE Carpathians, much younger early Pliocene to Quaternary AFT ages occur in the external nappes NW of Buzău.

6.4.3 AHe age patterns

The AHe age distribution map (Figure 6.4d) shows latest Cretaceous to Oligocene AHe ages for the Apuseni Mountains and South Carpathians. In the Apuseni Mountains, latest Cretaceous–middle Eocene AHe ages appear in the Bihor Dome, whereas slightly younger Middle Eocene–Oligocene ages are observed along the Mezeş thrust and in the area surrounding the Neogene volcanics of the Beius Basin. One latest Cretaceous–Paleocene age is recorded for the Făgăraș Mountains in the South Carpathians.

Consistent Miocene AHe ages are shown for the Preluca Massif and East Carpathians, whereas AHe ages in the SE Carpathians range from Eocene to Quaternary (Figure 6.4d). The AHe age distribution in the East and SE Carpathians reveals a similar pattern as the AFT data. Miocene AHe ages appear in the East Carpathians, both in the upper Dacia plate, the Ceahlău unit, and in the sedimentary wedge. Furthermore, Miocene ages are observed in the Preluca Massif. In the SE Carpathians, however, the Bucovinian nappes record older Eocene–Oligocene AHe ages, suggesting that the presumed upper plate with respect to
the Miocene subduction has not been exhumed from >60°C since the Eocene–Oligocene. Miocene AHe ages are found in the Ceahlău unit, and the sedimentary wedge of the subducting Moesian plate. The external nappes of the SE Carpathians consistently record very young Pliocene to Quaternary AHe ages along a NNE-SSW strike, hence crosscutting the early–middle Miocene thrusts.

6.5 Spatial and temporal distribution of exhumation patterns and differential vertical movements

The Early Cretaceous (140–100 Ma) and Intra-Turonian (100–78 Ma) contractional deformation events cannot be properly quantified at regional scale with the available data set due to later thermal overprint. Thus, the main exhumation episodes and related tectonic events that will be discussed are (Figures 6.5 and 6.6): 1) latest Cretaceous–middle Eocene, 2) middle Eocene–Oligocene, 3) early–middle Miocene, and 4) late Miocene–present-day. For the SE Carpathians, the last episode can be further subdivided into 4a) late Miocene, 4b) latest Miocene–early Pliocene and 4c) latest Pliocene–Quaternary.

6.5.1 Latest Cretaceous–middle Eocene (78–41 Ma)

The latest Cretaceous–middle Eocene map shows overall exhumation of Tisza (Bihor and Codru nappes) and Dacia (Bucovinian/Getic/Biharia nappes), but also of the Danubian nappes in the South Carpathians and of the Ceahlău units in the SE Carpathians (Figure 6.5a).

Results suggest ~2–4 km of exhumation for the Apuseni Mountains (Figure 6.5a). The largest amount of exhumation is localised in the area of the Bihor Dome, suggesting that doming and subsequent folding of the Codru nappes might be a latest Cretaceous–Paleocene feature (Figures 6.5a and 6.6b). Timing is coeval with the latest Cretaceous emplacement of the Banatites in the Bihor Dome area [Wiesinger et al., 2005; Zimmerman et al., 2008], and deposition of the Gosau deposits and the Bozes flysch at the flanks of the Bihor Dome [Schuller, 2004]. An influx of detrital material with ZFT age populations characteristic for the Bihor and Codru nappes recorded for uppermost Cretaceous–Paleocene sediments supports latest Cretaceous–Paleocene exposure and erosion of these units as a result of compressional tectonic activity [Schuller et al., 2009]. The data suggest that the Mezeș Mountains and the adjacent northwestern part of the Transylvania Basin were buried at ~2 km (Figures 6.5a and 6.6a), which is corroborated by the latest Cretaceous–middle Eocene deposition of red continental deposits of the Jibou Formation [e.g. Proust and Hosu, 1996].

Large exhumation estimates (~5–10 km) are indicated for the Preluca Massif and the Bucovinian nappes in the northern part of the East Carpathians (Figures 6.5a and 6.6a). In the northern part of the East Carpathians, exhumation is localised in the core of the antiformal shaped Bucovinian nappe stack (Figures 6.4b, 6.5a and 6.6a), thus exposing its lowest structural level, i.e. the Infra-Bucovinian nappe (Figure 6.4a). In the SE Carpathians, ~3–5 km of exhumation is suggested for the frontal part of the Bucovinian units and the Ceahlău nappes (Figures 6.5a and 6.6b).
In the South Carpathians, 2–5 km of exhumation is suggested for both the Getic and the Danubian nappes following the antiformal-shaped nappe-stack (Figures 6.5a and 6.6c).

6.5.2 Middle Eocene–Oligocene (41–23 Ma)

The middle Eocene–Oligocene map (Figure 6.5b) suggests overall exhumation of the Apuseni Mountains and South Carpathians, whereas it indicates burial for the East and SE Carpathians and the northern and eastern parts of the Transylvania Basin.

In the Apuseni Mountains, exhumation in the order of ∼2 km is limited by the Mezeş thrust (Figures 6.5b and 6.6a). The rest of the Apuseni Mountains appears to be a stable area during this time with no significant exhumation or burial recorded (Figures 6.5b and 6.6b). Directly NE of the Mezeş and Puini thrusts, burial is indicated for the NW part of the Transylvania Basin (including the Preluca Massif). For both faults, the exhumation data suggest relative upward movement of the western block with respect to the eastern block, which suggests reverse faulting during this period (Figures 6.5b and 6.6a, b). The pre-Oligocene and pre-Miocene subcrop maps for the Transylvania Basin of Krézsek and Bally [2006] suggest a similar subsidence pattern for the NW part of the basin influenced by fault activity of the Puini thrust. Burial estimates of <3 km are comparable with the subsurface thicknesses for the Paleo–Eocene and Oligocene of Krészek and Bally [2006], which are ∼800–1500 m for the area between the Mezeş and Puini thrusts and ∼2500 m east of the Puini thrust.

In the South Carpathians-Dinarides connection, middle Eocene–Oligocene exhumation is suggested for the Danubian window along the Getic detachment (Figure 6.5b). A significant jump along the Getic detachment is suggested by high exhumation estimates (<8 km) for the Danubian nappes compared to <1 km for the Getic nappes to the north (Figure 6.6c). In the northern part of the Danubian window, exhumation estimates are highest for the eastern part (3–8 km) and show an overall decrease towards the west (<2 km) across the Cerna-Jiu fault. Towards the south, exhumation values range between 1 and 3 km, with the highest values localised along the Cerna-Jiu fault (Figure 6.5b). The degree of exposed metamorphism in the Danubian window shows a minor general increase from ∼250°C at ∼2.5 kbar in the SW to ∼300–350°C at 3–4 kbar in the NE [Ciulavu et al., 2008]. This provides a critical constraint on the maximum exhumation of the Danubian window, which cannot exceed about 14 km, using a constant geothermal gradient of 25°C/km (E. Willingshofer, personal communication).

The data suggest that the eastern part of the Transylvania Basin, the East
a) Latest Cretaceous - middle Eocene

b) Middle Eocene - Oligocene

EXHUMATION HISTORY OF THE ROMANIAN CARPATHIANS
c) Early-middle Miocene

Figure 6.5: Continued
Figure 6.5: Continued
Carpathians, the external part of the SE Carpathians and its European foreland experienced 1–7 km of burial (Figures 6.5b and 6.6a, b). In contrast, the southeastern part of the Transylvania Basin was characterised by a period of non-deposition or erosion (Figure 6.5b). The pre-Oligocene and pre-Miocene subcrop maps for the Transylvania Basin of Krézsek and Bally [2006] corroborate these observations, indicating erosion for the SE part of the Transylvania Basin and accelerated sedimentation in the domain of the East and SE Carpathian external nappes [e.g. Fusaru and Kliwa sandstones, see Sândulescu et al., 1981].

6.5.3 Early–middle Miocene (23–11/9 Ma)

The early–middle Miocene map suggests major exhumation in the East Carpathians (2–5 km) and external SE Carpathians (2–4 km), whereas the South Carpathians and Apuseni Mountains generally show only minor exhumation during this period (Figure 6.5c).

The distribution of exhumation estimates in the East and SE Carpathians (Figure 6.5c) reveals a similar trend as observed in the AFT and AHe age patterns (Figures 6.4c and d). In the northern East Carpathians, early–middle Miocene exhumation takes place both in the upper Tisza-Dacia block and in the sedimentary wedge shortened above the subducting East European foreland with values reaching up to 5.7 km (Figure 6.6a). However, early–middle Miocene exhumation in the SE Carpathians is limited to the Ceahlău nappes and the sedimentary wedge of the subducting Moesian plate, reaching values of maximum 4.5 km (Figure 6.6b). In the SE Carpathians, the upper Dacia block (Bucovinian units) forms a stable area, i.e. no significant exhumation is recorded here for this period (Figure 6.6b).

For the South Carpathians, data suggest no exhumation to minor burial (0–1 km, Figure 6.5c). Only in the area NW of Rimnicu Vilcea, 1–3 km of exhumation is suggested for both the Danubian and Getic nappes (Figure 6.6c, Parang Mts.). Minor exhumation (0–1 km) is also indicated for the core of the Apuseni Mountains (Figures 6.5c and 6.6a, b). Minor early–middle Miocene exhumation of the Apuseni Mountains and South Carpathians contradicts previous studies, which postulated that Miocene tectonic events induced large-scale exhumation over the entire Romanian Carpathians [e.g. Sanders, 1998].

Minor burial of ~0.8 km occurred in the NW-most part of the Transylvania Basin (Figures 6.5c and 6.6a, b), which is corroborated by <1000 m of lower Miocene sedimentation [Krézsek and Bally, 2006]. The central parts of the Transylvania Basin record higher amounts of burial in the order of 2–3 km (Figures 6.6a and b). The western margin of the Apuseni Mountains is subsiding simultaneously with normal faulting taking place in the western basins (Borod, Beius, Zaran, the Mureş corridor) [e.g. Csontos, 1995]. This is confirmed by the pre-middle Miocene and early Sarmatian subcrop and outcrop maps of Krézsek and Bally [2006], which indicate continental to shallow-marine sedimentation in these basins for this period.

6.5.4 Latest Miocene–present-day (11/9–0 Ma)

The latest Miocene to present-day map (11–0 Ma) suggests generalised exhumation throughout the entire Romanian Carpathians. Only in the SE Carpathian
external nappes and foreland, burial is recorded (Figures 6.5d and 6.6). Exhumation estimates are uniform across the East, internal SE and South Carpathians and the Apuseni Mountains (0.5–2.5 km). These values are corroborated by the onset of inversion of the Transylvania Basin at the end of Pannonian times (~9 Ma) inducing 0.5–1.5 km of erosion [Krézsek and Bally, 2006].

Latest Miocene–present-day exhumation of the SE Carpathians in the area limited by the Peceneaga-Camena and Intra-Moesian faults is significantly different from the rest of the Romanian Carpathians (Figure 6.5d). The data suggest that the SE Carpathian external nappes were subsiding from the latest Miocene until ~6 Ma (Audia thrust contact) to ~3 Ma (between Audia thrust and Pericarpathian line), suggesting 1–3 km of burial by sedimentation (Figures 6.5d and 6.6b). This is corroborated by remnants of uppermost Miocene sediments still unconformably covering parts of the external nappes north and southwards (marked as post-tectonic cover in Figure 6.4a). Values of 1–3 km are in agreement with regional tectonic reconstructions on the western flank of the Focșani Basin, which have postulated ~2 km of restored sediment thickness over the external nappes [Leever et al., 2006]. Towards the SE Carpathian foreland, post-11 Ma shallow marine, lacustrine and alluvial deltaic sediments of the Focșani basin reach thicknesses of ~6 km [Jipa, 1997; Tărăpoancă et al., 2003].

For the latest Miocene–early Pliocene (between 6 to 3 Ma), rapid exhumation is suggested for the SE Carpathians at the Audia thrust contact (Figures 6.5e and 6.6b). Data suggest 1–3 km of uplift and erosion with the axis of maximum exhumation striking NNE-SSW to NE-SW, thus crosscutting the nappe contacts. Timing coincides with a change in sediment source area around 5 Ma recorded in sandstone petrography data, from a volcanic arc province derived from the hinterland volcanoclastic edifices [Seghedi et al., 2004], to a recycled orogen province [Panaiotu et al., 2007]. At the same time, the foredeep basin was still subsiding [Leever et al., 2006].

In the latest Pliocene–Quaternary (3 Ma to present-day), the data suggest 2–4 km of uplift and erosion in the external SE Carpathians, localised in the area between the Audia thrust and the Pericarpathian line (Figures 6.5f and 6.6b). The NE-SW oriented axis of maximum exhumation, which is the similar to the orientation of open folds and thrusts in the SW areas (between Buzău and Ploiești; Figure 6.4a), is crosscutting early to middle Miocene nappe contacts (Figure 6.5f). Timing coincides with syn-kinematic thickness variations recorded from the uppermost Pliocene onward. Lower Quaternary coarse deposits have been subsequently tilted (~5–10° eastward dip) and uplifted to present-day elevations of up to 1 km [e.g. Leever et al., 2006]. Synchronous with accelerated exhumation in the external SE Carpathians, the Brașov, Ciuc and Gheorghieni hinterland basins formed by NW-SE extension and gravity spreading during the Pliocene–Quaternary (from ~4 Ma onward), accumulating up to 1000 m of Plio-Quaternary continental deposits [Gîrbacea et al., 1998; Chalot-Prat and Gîrbacea, 2000].

6.6 Kinematic evolution of the Romanian Carpathians

Several major processes were operating in the Carpathian realm during Cretaceous to present-day times (Figures 6.1 and 6.3): 1) the Jurassic to Cretaceous
Figure 6.6
EXHUMATION HISTORY OF THE ROMANIAN CARPATHIANS

Figure 6.6: Continued
Figure 6.6: Simplified crustal-scale cross-sections illustrating the mechanisms playing a major role in the evolution of the Romanian Carpathians (bottom). Plotted on top of each cross-section are (from bottom to top): thermochronological ages (see legend for references and explanation of symbols), and estimated exhumation patterns through time for the latest Cretaceous–Oligocene and the Miocene–Quaternary respectively (bars indicate input data). The Early Cretaceous (140–100 Ma) and Intra-Turonian (100–78 Ma) deformation events cannot be properly quantified with the data set, due to later thermal overprint. Note that the exhumation lines do not represent a paleotopography because the zero-lines are in fact paleomorphologies inherited from previous events. a) SW-NE cross-section through the Mezeș Mountains (simplified after Chapter 3), the NW part of the Transylvania Basin [simplified after Krézsek and Bally, 2006] and the northern part of the East Carpathians (simplified after Chapter 5). b) W-E to NW-SE cross-section through the Bihor Dome, the Transylvania Basin [simplified after Schmid et al., 2008] and the SE Carpathians (simplified after Chapter 4). c) N-S cross-section through the South Carpathians [Matenco et al., 2010].
subduction/obduction of the East Vardar Ocean and the collision of Tisza and Dacia, 2) the Cretaceous subduction of the Ceahlău-Severin Ocean and the subsequent Neogene slab roll-back and Miocene collision of Tisza-Dacia with the European foreland, and 3) the Cretaceous–Eocene subduction/collision recorded in the Sava Zone of the Dinarides.

Intra-Albian exhumation ages in the Southern Apuseni Mountains, the South Carpathians and in the Bucovinian nappes of the East and SE Carpathians probably record the subduction and obduction of the East Vardar Ocean, and the onset of subduction in the Ceahlău-Severin Ocean (Figure 6.4). ZFT data from the northern part of the East Carpathians (Figure 6.4b) show at least partial annealing related to burial by Alpine age nappe-stacking [Gröger, 2006] and the subsequent exhumation of the nappe pile. The effect of the onset of shortening in the Adriatic lower plate of the Dinaridic Sava zone during the late Early Cretaceous [Schefer, 2010] on the exhumation of the Romanian Carpathians is unknown and difficult to quantify in the absence of exhumation data for the Dinarides.

By restoring the Paleogene–Miocene ∼90 degrees clockwise rotation of the Tisza-Dacia Block [e.g. Pătraşcu et al., 1990; 1994], the presently NW-facing Intra-Turonian nappe stack of the Apuseni Mountains would indicate top SW-ward shortening direction, which is compatible with the Late Cretaceous–Eocene Dinaridic kinematics [e.g. Schmid et al., 2008]. It is difficult to differentiate between the retro-shear type of orogenic thickening in the East Vardar domain and the effects of the Dinaridic shortening due to the subduction recorded in the Sava zone [see plate 3 in Schmid et al., 2008].

### 6.6.1 Latest Cretaceous–middle Eocene tectonics

One key result of this study is the observation that the latest Cretaceous tectonic event [the “Laramian” phase of Sândulescu, 1988] is related to continuous exhumation which is prolonged all the way into the Paleocene–middle Eocene. This is the moment when the Ceahlău-Severin units were coupled [foreland-coupling collision senso Ziegler et al., 1995] together with the underlying Danubian block and incorporated into the upper plate in the Moesian domain (Figures 6.6b and c). In the northern part of the East Carpathians where this block is absent, the Ceahlău units were docked against more external thin-skinned nappes (Figure 6.6a). Thrusting was accompanied by exhumation and erosion of the Bucovinian nappes at ZFT resolution, indicating the highest amounts of exhumation in the core of the Bucovinian antiform (Figure 6.4b). This antiformal geometry presently observed in the northern part of the East Carpathians is likely to be a feature inherited from the Late Cretaceous–Eocene deformation phase.

The southern border of the South Carpathians is unconformably covered by a late Senonian post-tectonic cover. The stacking of the Getic and Danubian nappes over the Moesian platform is demonstrated by the exhumation recorded by the thermochronological data (Figures 6.5a and 6.6c) and by the heavy mineral provenance spectra (Bojar et al. 2010). Note that this exhumation is later overprinted by the formation of the Danubian core complex [see next section and Fügenschuh and Schmid, 2005]. However, it is likely that both processes form part of a more continuous evolution with a gradual change from shortening to elongation along
CHAPTER 6

the orogen, due to the onset of Tisza-Dacia rotations in the same overall stress regime.

The data suggest that doming of the Bihor Autochthon and subsequent folding of the Codru nappes in the Apuseni Mountains are of latest Cretaceous–Eocene age (Figures 6.5a and 6.6b). This is likely to be an effect of the coeval deformation episodes in the Sava zone of the Dinarides, which recorded continental collision and significant exhumation near the contact between Getic/Serbo-Macedonian upper plate and Jadar-Kopaonik lower composite unit [the Dinaridic phase; Schmid et al., 2008]. This collision is responsible for latest Cretaceous–Paleogene Banatitic magmatism that took place in a long curved belt stretching from the Northern Apuseni Mountains, western South Carpathians, Dinarides and further east into the central Balkans [see Schmid et al., 2008].

6.6.2 Middle Eocene–Oligocene rotations and shortening

The Eocene–Oligocene was generally considered as a tectonically quiescent period [Sanders, 1998] except for the South Carpathians [e.g. Fügenschuh and Schmid, 2005 and references therein], where Paleogene exhumation is known to accommodate the Paleogene–early Miocene rotation of Tisza-Dacia around Moesia [senso Ratschbacher et al., 1993]. However, the new compilation of thermochronological data clearly indicates that middle Eocene–Oligocene exhumation also occurs in parts of the Apuseni Mountains and SE Carpathians (Figures 6.5b and 6.6b and c).

In the Apuseni Mountains latest Eocene–earliest Oligocene ENE-WSW shortening has been recorded near the Mezeş Thrust (Chapter 3), inducing local exhumation in the order of 1–3 km (Figure 6.6a). In the footwall of the Mezeş thrust, pulses of subsidence and uplift are recorded in the Transylvania basin during the Paleogene, associated with local erosion [Paraschiv, 1979; Proust and Hosu, 1996]. Timing is coeval with collision in the Sava Zone (Figures 6.1a and 6.3) and thus the rather minor middle Eocene–Oligocene shortening episode and associated exhumation localised along the Mezeş and Puini thrusts might represent a retro-shear of the Dinarides (Figure 6.2b). The earlier mentioned collision between the Dinarides and the Tisza block [e.g. Pamić et al., 2002b] is associated with widespread thrusting in the external Dinarides and metamorphism/exhumation near the contact with the Tisza-Dacia upper plate [Schefter, 2010]. Shortening along steep basement faults (Mezeş and Puini thrusts) does correspond with retro-shear exhumation [senso Beaumont et al., 1996] and erosion, but at a much smaller scale.

In the East and SE Carpathians no clear Paleogene tectonic structures have been observed in kinematic studies. However, thermochronological data indicate that parts of the Dacia block (Bucovinian nappes) in the SE Carpathians were exhumed during the middle Eocene–Oligocene (Figures 6.5b and 6.6b), associated with thick turbiditic sedimentation in more distal parts of the flysch basin. Since the renewed and large-scale exhumation largely post-dates latest Cretaceous contractional events it is unlikely that middle Eocene–Oligocene exhumation represents the breakdown or collapse of the Cretaceous orogen. This suggests that the exhumation represents renewed contraction/subduction at the contact between
Tisza-Dacia and the distal parts of the foreland platforms, especially in the SE Carpathians. This would imply a potential earlier onset of the Miocene tectonic events affecting the external Carpathian nappes [the “Moldavides” of Săndulescu, 1988], and specifically an earlier onset of the early Miocene tectonics affecting the internal parts of this nappe pile. The post-tectonic covers, which classically define this event, probably record only the temporary cessation of thrusting during the early Miocene. The upper Eocene–Oligocene wedge-type geometries of the turbiditic sedimentation defined to represent trench asymmetries or even extensional events [e.g. Săndulescu et al., 1981; Săndulescu, 1994] can in fact be related to a foredeep-wedge type of sedimentation during active thrusting. This suggested subduction-related thrusting would be in a retro-wedge position with respect to the major collision recorded coeval in the Sava zone, but in a far distance (Figure 6.1a). However, the SE Carpathians cannot be a typical retro-shear, as deformation takes place along a subduction zone.

Latest Eocene core-complex formation resulted in exhumation of Danubian unit in the footwall of the Getic detachment (Figures 6.4c, 6.5b and 6.6c). Significant differences across the Cerna-Jiu fault system detected by FT studies, and limited by the degree of metamorphism to up to 14 km of exhumation, can be explained by the differential mechanics of the Danubian core-complex formation. The part of the Danubian core-complex situated SE of the Cerna-Jiu fault system reflects distinct middle Eocene–Oligocene tectonic denudation, whereas the Danubian part NW of this fault recorded slow cooling since the latest Cretaceous. The change from extension to strike-slip dominated tectonics along the curved Cerna-Jiu fault allowed for further latest Eocene–Oligocene exhumation on the concave side of this strike-slip fault, while exhumation ceased on the convex side [Fügenschuh and Schmid, 2005].

This South Carpathian along-strike core-complex deformation can be correlated with differential deformation between the Tisza-Dacia block and Moesia during the invasion of the Carpathian embayment and the coeval collision recorded in the Sava Zone of the Dinarides. The orogen parallel extension of the Danubian units was subsequently followed in the South Carpathians by the early Oligocene 35 km dextral offset of the Cerna fault [Berza and Draganescu, 1988] and early Miocene 65 km dextral offset of the Timok fault [Kräutner and Krstić, 2003]. This dextral rotation and invasion of the Tisza-Dacia block into the Carpathian embayment (Figure 6.1b) might explain the localisation of middle Eocene–Oligocene exhumation in the area of the SE Carpathians in the proximity of the South Carpathians (Figure 6.5b). This effect is absent in the rest of the SE and East Carpathians as shown by the suggested burial for this time period (Figures 6.5b and 6.6a).

6.6.3 Miocene orogenic events

The Miocene contraction between Tisza-Dacia and the European foreland was generally considered to have induced exhumation and uplift in the entire Romanian Carpathians [e.g. Sanders, 1998]. However, the compilation of thermochronological data clearly illustrates that only the East Carpathians and parts of the SE and South Carpathians were exhumed during this period (e.g. Figures 6.5c
and 6.6). This was the result of thin-skinned thrusting developing in a typical foreland propagating sequence, scraping off sediments deposited over the European/Moesian passive continental margin under subduction.

Towards the end of the middle Miocene (∼11-9 Ma), collision was locked against the non-thinned continental parts of the foreland platforms. Generalised exhumation occurred in the entire East Carpathians, as indicated by the 20–11 Ma (locally as young as 8 Ma) AFT and AHe ages (Figures 6.4c and d). In the northern part of the East Carpathians, uplift and exhumation in the order of 3–5 km took place over the internal parts of the orogen (Figure 6.5c), in combination with local transpression along major strike-slip faults (e.g. Bogdan-Voda and Dragos-Voda system). This study indicates that Miocene exhumation at AHe resolution also occurred in the SE Carpathians in the order of 3–4 km (Figure 6.5c). Here, Miocene exhumation was restricted to the external Ceahlău nappes and the thin-skinned nappes eastwards (Figure 6.6b). Since an upper plate must be exhumed during collision, the logical conclusion is that Miocene subduction in the SE Carpathians took place between the Danubian block and overlying Ceahlău nappes in an upper plate position and the Moesian foreland in a lower plate position (Figure 6.6b). This suggests that the Danubian block in the SE Carpathians was already accreted to the upper plate during earlier Cretaceous (to Paleogene?) phases of subduction. This type of accretion of a continental block during subduction and the shift of the major plate contact is a well known process at active margins [e.g. Brun and Faccenna, 2008; Ricketts, 2008]. The localisation of this process in the SE Carpathians can be related to the presence of the rather weak Moesian plate in that part of the European foreland [e.g. Cloetingh et al., 2004], which can localise foreland-coupling deformation [senso Ziegler et al., 1995]. North of the Trotuş fault, the rheologically stronger East European Platform blocked a forward propagation of the deformation front, thus causing generalised exhumation of both the subducting and overlying plates (Figure 6.6a).

This compilation suggests that widespread Miocene exhumation did not occur in the South Carpathians (except for the area NW of Râmnicu Vâlcea) as suggested by earlier thermochronological studies [e.g. Sanders, 1998] or speculated from sedimentological studies on source areas [Krészsek and Filipescu, 2005]. This is demonstrated by the higher resolution of the combined use of the AFT and AHe methods, which corrects for any track length artefacts related to AFT thermal modelling.

Similarly, there is no evidence for significant Miocene exhumation in the Apuseni Mountains as suggested by earlier thermochronological studies [e.g. Sanders, 1998; Schuller, 2004]. However, upper Miocene deposits (Pannonian–Pon- tian) are the youngest deposits, indicating non-deposition or erosion at least since then. Furthermore, strike-slip faulting with NW-SE shortening and NE-SW extension has been recorded for middle–late Miocene times, associated with opening of NW-SE trending graben structures. This was associated with volcanism peaking during Sarmatian times, which is the result of magma generation by lower crustal decompression during extensional opening of the Pannonian basin [Seghedi, 2004].
6.6.4 Post-collisional evolution

After the middle–late Miocene “soft”-collision (∼9 Ma), the entire system (Tisza-Dacia including the Apuseni Mountains, South Carpathians, Transylvania Basin and East Carpathians) was exhumed 1–2 km (Figure 6.5d). Timing coincides with the onset of inversion of the Transylvania Basin at the end of the Pannonian (∼9 Ma) [Krézsek and Bally, 2006]. Thermal models for the Northern Apuseni Mountains, South Carpathians and East Carpathians indicate a similar timing for the onset of generalised exhumation (Chapters 3 and 5). Low-temperature thermochronological results indicate coeval subsidence for the external part of the Miocene fold-and-thrust belt in the SE Carpathians (∼2 km of burial, Figure 6.5d) and the Focsani foredeep basin (∼6 km; Leever et al., 2006), which is most probably related to lower plate loading as a result of foredeep sedimentation during shortening. This mechanism has a large wavelength geometry, due to the antiformal updoming of the upper plate (Tisza-Dacia), already locked against the European foreland (Figures 6.6a and b). The overall exhumation was rather minor and decreases at larger distances from the former subduction zone, corroborated by the observation that no major exhumation was recorded for the Apuseni Mountains. Here, uplift and exhumation are only suggested by the hydrographical river network incision along the western flank of the Apuseni Mountains, where middle–upper Miocene (Pannonian–Pontian) sediments are incised up to 300–400 m topographic difference with respect to river base level.

The ongoing subsidence recorded in the foreland of the SE Carpathians is the result of the pull exerted by the high velocity anomaly detected in tomography studies beneath the Vrancea area [e.g. Martin et al., 2006], generally interpreted as a slab remnant from the subduction history of the Carpathians [e.g. Wenzel et al., 1999]. In the area of the SE Carpathians situated between the Trotuş, Peceneaga-Camena and Intra-Moesian faults, the overall latest Miocene to present-day evolution is strongly overprinted by a younger exhumation event (Figure 6.6b). In the internal parts of the orogen, exhumation starts around 5 Ma with ∼3 km of exhumation (at rates of 1.7 mm/yr) over the core of the orogen (Chapter 4). This event peaks around 3–2 Ma, when significant exhumation is recorded more to the foreland in the external nappes of the SE Carpathians and the western foredeep margin. This induced gradual E-wards tilting of foreland strata in northern areas (NW of Focşani, Figure 6.4a), and short-wavelength faulting/folding in the southern areas where the structural grain changes from N-S to E-W (West of Buzău, Figure 6.4a). Most pronounced exhumation occurs in the Tarcau nappe along Buzau Valley, where ∼3–4 km of rock has been eroded during latest Pliocene–Quaternary times at a rate of ∼1.6 mm/yr (Chapter 4). The overall Pliocene–Quaternary configuration indicates that post-collisional exhumation in the SE Carpathians migrated towards the foreland, which is partly coeval with a forward-migration of the foredeep depocenter during the Quaternary [Leever et al., 2006]. The latter mechanism may also be valid for the Pliocene; however associated sediments were subsequently eroded and thus prevent a clear conclusion. The overall migration in uplift and subsidence patterns implies a change in the positioning of the associated mechanism. As the inversion is explained by a combination of slab-pull and coeval crustal shortening, it is likely that the Trotuş and Intramoesian faults acted as strike-slip faults accommodating
foreland coupling and slab steepening. The ongoing shortening is accommodated along steep basement faults in the lower plate, inducing major uplift and erosion (Figure 6.6b).

Whether the late-stage exhumation of the SE Carpathians is the result of far-field stresses transmitted by the Adriatic push along the active external Dinaridic boundary as previously suggested [e.g. Matenco et al., 2007] is still unclear. The observation that the amplitudes of exhumation decrease from the Vrancea area westwards can be combined with GPS-derived strain patterns, which indicate a decrease of deformation from the Dinarides to the Apuseni Mountains [e.g. Grcerczy et al., 2005]. Therefore, these two mechanisms (Dinaridic push and Vrancea slab-pull) seem to be independent. However, numerical modelling of the Pannonian lithosphere indicates that strain tends to concentrate at non-stretched continental margins during the onset of inversion, while the thinned Pannonian lithosphere would respond later and with smaller amplitudes in the order of tens to hundreds of metres [Jarosinski et al., 2009]. To answer this question, a detailed analysis of exhumation patterns at the transition between the Apuseni Mountains and the Pannonian Basin would be required, which is beyond the resolution of this low-temperature thermochronological study.

6.7 Present-day topography and its link with exhumation and tectonics

This study has demonstrated that the exhumation history of a low-topography orogen such as the Romanian Carpathians can be studied in detail by using a combination of thermochronological and structural data. Previous regional-scale thermochronological studies suggested that the present-day topography of the Romanian Carpathians is a Miocene–Quaternary feature [e.g. Sanders, 1998; Sanders et al., 2002]. However, the present compilation of data from three different low-temperature thermochronometers shows that only the East and part of the South and SE Carpathians were significantly exhumed in early-middle Miocene syn-collisional times (Figure 6.5c). In latest Miocene–Quaternary post-collisional times, the entire hinterland was slowly exhumed (Figure 6.5d). Only the most frontal part SE Carpathian thin-skinned nappes show major exhumation (i.e. more than ~1–2 km) in Pliocene–Quaternary times (Figure 6.5e and f) following latest Miocene–early Pliocene subsidence (Figure 6.5d).

The final topographic expression of the Romanian Carpathians can be discussed on the basis of the youngest recorded exhumation phase (Figure 6.7). Note that the relatively minor overall post-collisional uplift (<2 km) of the entire Carpathian orogen (except the SE Carpathians) is herewith omitted.

The data suggest that the main part of the topography of the Apuseni Mountains and South Carpathians is inherited from Cretaceous–Paleogene tectonic events. In the Northern Apuseni Mountains, the main topography is formed by the Bihor Dome (Figure 6.7). The area was subsequently affected by late Eocene–Oligocene thrusting, creating the Mezeş Mountains and the exhumation of Paleogene deposits of the western part of the Transylvania Basin. Subsequently, Miocene normal faulting created the WNW-ESE to NW-SE oriented basins at the western Apuseni margin. However, this was not associated with
significant footwall exhumation (i.e. below AHe resolution). The South Carpathians display mainly a Paleogene topography (Figure 6.7) [see also Fügenschuh and Schmid, 2005], with younger Oligocene–Miocene exhumation observed only along the dextral faults that accommodated the rotation around the Moesian promontory (i.e. Cerna fault system) [e.g. Ratschbacher et al., 1993] and the subsequent dextral transpression against the E-W oriented Moesian margin [e.g. Răbăgia et al., 2009].

The East Carpathians display a topography which is mostly of early–middle Miocene age (Figure 6.7), a result of the Miocene orogenic processes, middle–late Miocene continental collision and the subsequent uplift and erosion. Interestingly, in the northern segment, both the upper (Bucovinian nappes) and lower (European foreland and sedimentary wedge) plates were exhumed during this time interval and therefore represent a Miocene topography. Differential amounts of exhumation are related to (sinistral) transpression along the Bogdan-Voda and Dragos-Voda fault systems, kinematically linked to the Mid-Hungarian fault zone accommodating the dextral translations and rotations of Tisza-Dacia with respect to ALCAPA [Tischler et al., 2007].

The topography of the SE Carpathians seems to be a combination of Cretaceous, Miocene and Pliocene–Quaternary exhumation episodes. Only the Ceahlău and more foreland-ward thin-skinned nappes were exhumed during the Miocene orogenic processes, whereas the Perşani Mountains still represent an older Cretaceous–Paleogene palaeorelief. The topography of the most frontal part of the
SE Carpathians is the youngest topography in the Romanian Carpathians. The area that records the youngest AHe ages and largest amounts of exhumation corresponds to the highest present-day topography of the SE Carpathians (Figure 6.7). This indicates that active topography building processes in the SE Carpathians started during the rapid late Pliocene–Quaternary deformation episode. The morphology of the East and South Carpathians clearly reveals that the highest topography in the SE Carpathian bend zone is shifted towards a more external position to the SE (Figure 6.7). This can be related to the effect of the Pliocene–Quaternary foreland-ward migrating exhumation and inversion induced by a combination of Vrancea slab-pull and coeval crustal shortening.

### 6.8 Accelerated Pliocene–Quaternary exhumation of the SE Carpathians in a regional perspective – the effects of tectonics and climate

The rapid Pliocene–Quaternary exhumation of the SE Carpathians (~4–5 km at a rate of ~1.6 mm/yr, see Chapter 4) is a striking feature that is not observed in the rest of the Romanian Carpathians (Figures 6.5 and 6.6). Such an increase of Pliocene–Quaternary exhumation rates has also been reported for other areas in Europe, in particular for the Swiss and Western Alps [e.g. Schlunegger and Hinderer, 2003; Cederbom et al., 2004; Willett et al., 2006; Champagnac et al., 2007; Vernon et al., 2008], but also for the Betics [Johnson, 1997; Reinhardt et al., 2007], the Appennines [Balestrieri et al., 2003] and the Caucasus [Morton et al., 2003; Saintot et al., 2006]. A Pliocene–Quaternary increase in erosion rates is in line with indications for a global increase of sedimentation rates at 2–4 Ma, in a variety of settings including active and inactive mountain belts [e.g. Zhang et al., 2001; Molnar, 2004]. Possible causes that have been proposed include global cooling and onset of widespread glaciation [Molnar and England, 1990; Hinderer, 2001; Ehlers et al., 2006], an increase in climatic variability [Zhang et al., 2001; Molnar, 2004], and an increase in uplift rates of major orogens [Raymo and Ruddiman, 1992]. However, much is debated on the relative importance of tectonics and climate on increased Pliocene–Quaternary erosion and sedimentation rates. Studies in favour of climatic forcing argue that plate motions do not show a globally synchronized change in rates at any time in the past 5 to 10 Ma to support a tectonically forced global increase in uplift rates and the subsequent increase of erosion rates [e.g. Molnar, 2004]. However, other studies have shown that climate-induced erosion and the subsequent relief production and isostatic compensation is often not sufficient to explain the observed rock uplift [e.g. Champagnac et al., 2007; 2008]. Recent studies even question whether the increase in global weathering and erosion actually occurred, and showed that on a global scale the chemical weathering flux was essentially constant over the past 10 Ma [Willenbring and Von Blanckenburg, 2010].

Based on detailed thermochronological analyses, potential climatic controls on the latest Miocene–Quaternary exhumation history of the SE Carpathians were also suggested in this study (see Chapter 4). Although the first- and second post-collisional exhumation phases (at 5–6 Ma and 2–3 Ma ago, respectively) are local SE Carpathian features and most likely represent a tectonic effect of deep
processes related to the evolution of the Vrancea slab (Chapter 4), clear structural and sedimentological evidence associated with tectonics is absent for the first exhumation phase. Timing of the first exhumation phase is however overlapping with a base-level drop and rapid fill observed in the Carpathian foreland after 6 Ma [Leever, 2007] related to the large-scale sea-level drop of the Messinian Salinity Crisis (MSC, 5.96–5.33 Ma) [Krijgsman et al., 1999]. Recent studies of orogenic evolution in the Mediterranean realm, such as for the Alps, indicated a shift of exhumation to orogenic interiors and a change to orogenic destruction coeval with the MSC [e.g. Willett et al., 2006], which enhanced denudation by increasing the exposure of the source areas [e.g. Clauzon et al., 1996; Foeken et al., 2003]. Similar conditions are met for the orogens in the Paratethys realms such as the SE Carpathians, where a coeval large sea-level drop associated with massive sedimentation has been reported for the Black Sea [see Dina et al., 2005; Gillet et al., 2007]. This MSC middle Pontian sea level drop recorded in the Paratethys domain [e.g. Stoica et al., 2007] has potentially induced a shift of depocenters towards the deeper parts of the Black Sea, which is coeval with enhanced erosion of the orogen and exposed parts of the basins by re-equilibration of river profiles. This would partly represent a climatically induced event, since it is generally agreed that the MSC resulted from a complex combination of tectonic and glacio-eustatic processes although the relative importance of both processes is debated [e.g. Clauzon et al., 1996; Krijgsman et al., 1999; Warny et al., 2003; Fauquette et al., 2006].

The second post-collisional exhumation phase has been interpreted as tectonically induced rock uplift followed by subsequent erosion, which is corroborated by structural and sedimentological observations (Chapter 4). The proposed interpretation of a crustal-scale shortening mechanism by reverse faulting along steep basement thrusts and the subsequent erosion of the uplifted areas would also explain the shallow position of the basement beneath the nappe pile [Landes et al., 2004; Bocin et al., 2005; 2009], and the shallow Moho configuration [Hauser et al., 2007] (Figure 6.6b). Combined with the subsidence in the foreland, this would explain the fold-like geometry that is observed at a crustal scale (Figure 6.6b).

The growth and geometry of an orogen are also controlled by erosion, because it directly influences topographic decay and modulates the pattern and rates of surface uplift [e.g. Beaumont et al., 1992; Burbank, 2002; Reiners and Brandon, 2006]. In convergent settings where erosion cannot sufficiently remove the volume of uplifted rock, orogens grow to high, wide orogens, like the Himalayas [e.g. Hodges, 2000; Thiede et al., 2004]. Other areas characterised by erosion rates much higher than or comparable to rock uplift rates, such as the Southern Alps in New Zealand and the European Alps, remain much smaller [e.g. Kamp and Tippett, 1993; Willett et al., 1993; Seward and Mancktelow, 1994; Beaumont et al., 1996; Bernet et al., 2001]. The SE Carpathians seem to be part of the latter category. Although estimated exhumation rates related to rock uplift are in the order of ~1.6 mm/yr (Chapter 4), the elevation of the SE Carpathians is minimal (~average of 1000–1500 m). This can be explained by simultaneous rapid erosion, which is corroborated by rapid Pliocene–Quaternary sedimentation recorded in the foredeep basin (~1.5 mm/yr; Vasiliev et al., 2004) coeval with rapid foreland subsidence (~0.9 mm/yr) and the gradual filling of the Focșani Basin.
CHAPTER 6

[Tărăpoancă et al., 2003]. Because erosion is also related to climate (e.g. precipitation), it provides an important feedback mechanism between climate and tectonics [e.g. Reiners and Brandon, 2006]. This raises questions as to why the SE Carpathians have been eroding so rapidly and have not been able to form a higher topography. This is especially interesting in relation to the climate-related globally observed Pliocene–Quaternary increase in denudation rates [e.g. Vernon et al., 2008] and sedimentation rates at 2–4 Ma ago [e.g. Zhang et al., 2001]. For the Carpathians, possible climatic controls have also been speculated based on low-amplitude measurements of geomorphological features such as terrace levels, and the deposition of coarse Quaternary conglomerates [e.g. Necea et al., 2005; Necea, 2010]. Alternatively, rapid erosion could be related to the rock properties of the rapidly exhuming area, which is made up of the Paleogene sedimentary wedge (Chapter 4). This sedimentary wedge mainly consists of turbidites that can be eroded relatively easily compared to crystalline basement. Such an effect of the exposed rock lithologies on surface erosion has been described for the Alps, where exposure of the crystalline core is likely to have resulted in a 25–40% decrease of sediment discharge [Schlunegger et al., 2001; Schlunegger and Simpson, 2002]. Partly related to this, the migration of tectonic deformation and exhumation towards the foreland (Chapter 4) might represent a factor that restricted the growth of a more elevated SE Carpathian orogen. The overall foreland-ward migration restricted to the SE Carpathians is a direct effect of the rheologically weak characteristic of the Moesian Plate, which localised deformation in a narrow zone gradually migrating towards the exterior of the orogen.

6.9 Conclusions

This study has integrated the large number of available thermochronological data for the Romanian Carpathians with new thermochronological and structural data to study the exhumation patterns and the mechanics of a low-topography orogen. The main features of this orogen are the relatively low amounts of collisional exhumation (<5–6 km), and the localisation of deformation at the plate contact or in the lower plate. These features are a direct result of the slab-dominated tectonic regime that characterises the Tertiary evolution of the Carpathians.

Collisional exhumation of the Carpathian orogen does not seem to respect the typical pattern of upper plate exhumation, which assumes the “rolling-carpet” model of adding material to the upper plate and exhumation in the core of the orogen to generate the build-up of the orogenic wedge (Figure 6.8a). Instead, most of the material is accreted to the frontal part of the orogen, while limited to no exhumation is recorded in the upper plate basement (Figure 6.8b). Deformation during collision has thickened the lower plate, particularly in the areas where the lower plate consists of rheologically weak crust (i.e. the Moesian Platform). The balance between upper plate exhumation and lower plate thickening is roughly equal. This results in large-wavelength fold-like geometries in the exposed surface geology (Figure 6.8b). The exposed basement is generally dominated by metamorphism of pre-Alpine age obviously related to the low amounts of collisional exhumation. This is contrastingly different from high-convergence orogens, which show significant Alpine-age metamorphism in the exposed geology.
Figure 6.8: a) Cartoon illustrating the exhumation pattern associated with retro-shear collision. Maximum exhumation occurs in the retro-shear area, spatially located above the contact zone between the upper and lower plate [e.g. Beaumont et al., 1996]. b) Cartoon illustrating the exhumation patterns associated with foreland-coupling collision. Foreland-coupling deformation occurs along high-angle faults in the lower plate, creating a wide antiformal bulk exhumation geometry (as observed in the East Carpathians). In the case of a weaker lower plate, foreland-coupling deformation events occur successively towards the foreland (t_1, t_2, and t_3), indicated by a migration of relatively small individual exhumation peaks in time (as observed in the SE Carpathians), resulting in a similar bulk exhumation geometry.
The Carpathian example also demonstrates the necessity of incorporating kinematics across syn- to post-orogenic basins, because studies focusing on isolated mountain chains have problems in defining genetic mechanisms. The most striking example is the evolution of the Apuseni Mountains, which cannot be solved as an individual orogen, but must take into account the evolution of other mountainous areas such as the external Carpathians and the Dinarides. Although the impact of the shortening and collision observed in the external Carpathians on the exhumation of the Apuseni Mountains has often been speculated, the amounts of exhumation have never been quantified. This study however showed that these mountains were not exhumed at AHe resolution after the Paleogene (i.e. less than 1–2 km). This contradicts previous studies, which speculated significant exhumation during the Miocene shortening in the Carpathians. Furthermore, the speculative definition of a latest Cretaceous–Eocene phase of Apuseni exhumation driven by the collision in the Dinarides is novel.

The detailed exhumation history of the East and SE Carpathians also demonstrates the significant influence of the mechanics of the lower plate on the exhumation of the upper plate during shortening. Although the upper plate is rather similar in composition and structure, the strong mechanical contrast between the Moesian and European domains results in significantly different exhumation patterns during subduction and collision. While the strong lower plate will distribute young exhumation ages all over the orogen, the mechanical weak lower plate will at first not exhume the upper plate and subsequently will do so, but indirectly, by mechanical coupling along high-angle reverse faults.

This study has demonstrated the applicability of using a combination of several low-temperature thermochronometers to derive a detailed late-stage exhumation history for a low-topography orogen. Especially the use of the AHe thermochronometer has added a significant value in distinguishing Ma-scale tectonic events. This is particularly interesting considering the focus on active tectonic processes and the interaction with surface processes.