Chapter 2
Tectonics and Exhumation of Romanian Carpathians: Inferences from Kinematic and Thermochronological Studies

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Abstract The ultimate topographic expression of intra-continental mountain chains is established during continental collision. The Romanian Carpathians provide a key location for understanding the mechanics of collision during slab retreat because the nappe stacking was not overprinted by back-arc extension, as commonly observed elsewhere. A review of existing kinematic and low-temperature thermochronological data infers that the collisional mechanics is significantly different when compared with high-convergence orogens. The shortening of the orogen at exterior was entirely accommodated by back-arc extension and the area in between simply rotated and moved into the Carpathians embayment. The roll-back collision is driven by foreland-coupling, a process that gradually accretes and exhumes continental material towards the foreland. The topographic expression of the Romanian Carpathians is both inherited from latest Cretaceous—Paleogene times, such as in the Apuseni Mountains or South Carpathians, and overprinted by the Miocene exhumation associated with the roll-back collision, as in the East or the SE Carpathians. The migration of exhumation towards the foreland continued during Pliocene—Quaternary times and is still active modifying the present-day topography in the SE Carpathians. The Transylvanian Basin is one of the best examples available of vertical movements induced by deep mantle processes in what is commonly referred as dynamic topography.

Keywords Kinematics · Thermochronology · Exhumation · Carpathians · Orogenic mechanics

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Introduction

The kinematics and exhumation of collisional orogens have been a constant topic of tectonic studies since the definition of steady-state wedges and associated large-scale hinterland exhumation during the buoyant subduction of continental plates (e.g. Platt 1986; Beaumont et al. 1994; Ring et al. 1999). Continental collision is the moment when out-of-sequence contractional deformation becomes rather the rule than the exception. Understanding its importance is fundamental for a fairy large number of orogenic processes, such as emplacement of metamorphic nappes, exhumation of high pressure rocks, deformation of thin-skinned thrust belts, the interplay between shortening and surface processes, syn-orogenic extension, the geometry of foredeep basins or accretion of continental material (e.g. Marotta et al. 1998; Burov 2007; Doglioni et al. 2007; Burov and Yamato 2008; Haq and Davis 2008; Naylor and Sinclair 2008; Faccenda et al. 2009). When compared with the strain partitioning observed during oceanic subduction stages, the out-of-sequence collision deforms a much wider zone, compressional stresses being transmitted much farther in the orogenic foreland and hinterland (e.g. Ziegler et al. 1998; Roure 2008).

The kinematics, geometry and exhumation of European orogens can be simply divided in two main categories (Fig. 2.1).

High-convergence collisional orogens, such as the Pyrenees or the Alps are characterized by large amounts of contractional exhumation. This exhumation is enhanced in orogenic hinterlands along retro-wedges that display a complex poly-phase deformation with an opposite polarity when compared to the one of the subduction zone (such as the Insubric line, Roure et al. 1989; Schmid et al. 1996; Beaumont et al. 2000). On the contrary, the "Mediterranean"-type of collisional orogens is dominated by subduction processes, resulting in the formation of highly arcuated mountain belts, such as the Apennines, Carpathians, Hellenides and the Betics–Rif system (Fig. 2.1). These orogens evolved rapidly during the retreat (or roll-back) of genetically associated slabs (i.e. Calabrian, Vrancea, Aegean and Gibraltar, respectively) that peaked in almost all situations during Miocene times (e.g. Jolivet and Faccenna 2000; Faccenna et al. 2004; van Hinsbergen et al. 2005; Ismail-Zadeh et al. 2012; Vergés and Fernàndez 2012). The slab retreat is accommodated by coeval extension affecting the hinterland of the upper orogenic plate, which formed large basins floored by either continental or oceanic lithosphere (such as the Pannonian and Aegean Basins, Black Sea or Western Mediterranean). These basins are extensional back-arcs in terms of geodynamic evolution (e.g. Royden 1993; Okay et al. 1994; Jolivet and Faccenna 2000; Horváth et al. 2006; Doglioni et al. 2007) although their relative position behind a magmatic or island-arc (Uyeda and Kanamori 1979; Dewey 1980; Mathisen and Vondra 1983) is not always very clear. In almost all situations, the back-arc extension overprinted and hid the earlier continental accretion, in particular by exhumation along extensional detachments, such as the widely documented core complexes of the

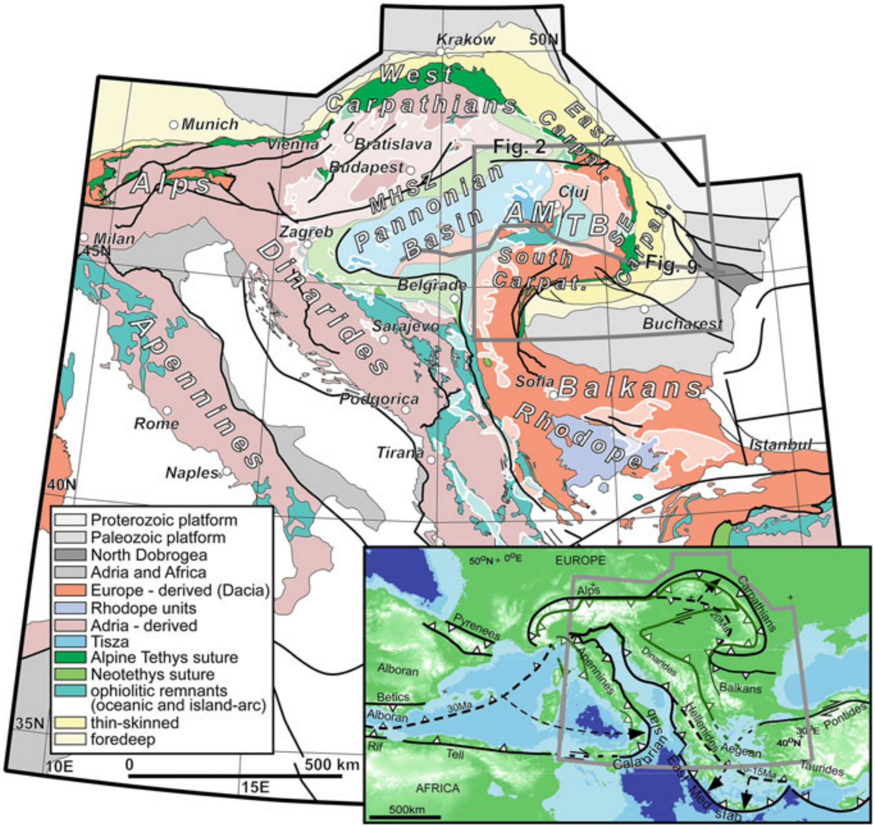

Fig. 2.1 Tectonic map of the Alps–Carpathians–Dinaridic–Hellenidic system (simplified from Schmid et al. 2011) with the extent of the Pannonian and Transylvanian back-arc basins (white transparent background). The *grey rectangle* is the location of Fig. 2.2 *AM*—Apuseni Mountains; *TB*—Transylvanian Basin; *MHSZ*—Mid-Hungarian Shear Zone. The lower inset is the location of the map in the system of European Mesozoic—Cenozoic orogens. *Dashed black line* is the position of the orogenic front prior to the onset of extension associated with the roll-back of the Calabrian, Aegean and Carpathian slabs

Rhodope–Aegean or Betics (Brun and Faccenna 2008; Brun and Sokoutis 2010; Vissers 2012).

An exception is the Romanian segment of the Carpathian Mountains, where the back-arc extension associated with the retreat of a slab kinematically connected with the stable European foreland (e.g. Schmid et al. 2008) took place during Miocene times in the Pannonian Basin, i.e. at far distances from the active subduction. This Miocene extension is rather minor in the areas situated in between, i.e. the eastern Apuseni Mountains, Transylvanian Basin or the East, SE and South Carpathians (e.g. Tiliță et al. 2013). The clockwise rotation and E-ward translation of these Carpathian units accompanied the W-ward subduction of the Carpathian embayment (sensu Balla 1986; Ustaszewski et al. 2008). Both processes cannot be accommodated by the dominantly N-S oriented absolute plate motion of Africa relative to Europe (Kreemer et al. 2003; van der Meer et al. 2010; van Hinsbergen and Schmid 2012). The reduced amount of Miocene extension provides a rare opportunity to quantify the crustal accretion processes and associated exhumation during roll-back and continental collision.

The north- to east-ward translations and coeval clockwise rotations during Cretaceous—Tertiary times have resulted in the formation of the characteristic double loop of the Carpathians around their SE corner and the transition from the South Carpathians to the Balkanides (Fig. 2.1, e.g. Csontos and Vörös 2004; Fügenschuh and Schmid 2005). This provides an opportunity to understand kinematic and exhumation processes associated with the formation of such highly arcuated orogens. The absence of studies connecting tectonic exhumation with landform formation in the Carpathians has resulted in long-lasting controversies concerning the age of topography and its links with deformation still taking place at present (e.g. Wohlfarth et al. 2001; Rădoane et al. 2003; Fielitz and Seghedi 2005; Dombrádi et al. 2007; Necea et al. 2013).

In order to derive the quantitative tectonic background required to understand the evolution of landforms in the highly arcuated Romanian Carpathians, this study is reviewing available kinematic and thermochronological data. The study provides critical constraints for understanding the mechanics of crustal accretion of collision in orogens dominated by roll-back. Furthermore, we analyse the link between deep- and near-surface processes acting in the aftermath of continental collision in this tectonically still active segment of the Carpathians.

Main Tectonic Units of the Romanian Carpathians

The Romanian Carpathian Mountains formed in response to a Triassic to Tertiary evolution of tectonic blocks and intervening oceans. In a fairly simplified terminology, four continental blocks [European foreland, Dacia, Tisza and ALCAPA (i.e. ALps–CArpathians–PAnnonia)] were separated by two intervening oceans (Neotethys/Transylvanides and Alpine Tethys/Ceahlău–Severin) closed by subduction and collision (e.g. Săndulescu 1984, 1988; Csontos and Vörös 2004; Schmid et al. 2008).

The European foreland is made up of a collage of units floored by Precambrian, Caledonian or Variscan basement, overlain by Paleozoic, Mesozoic or Cenozoic sediments with variable thicknesses and degrees of deformation. These were observed at depth in wells or defined by various geophysical techniques beneath the thrusting of the Carpathian units (Figs. 2.2 and 2.3, e.g. Săndulescu and Visarion 1988; Visarion et al. 1988). The East European and Scythian platforms are essentially Precambrian units, the latter representing the margin of the former affected by significant deformation during the latest Precambrian—early Paleozoic tectonic events and lower amounts of subsequent reactivations (Săndulescu and

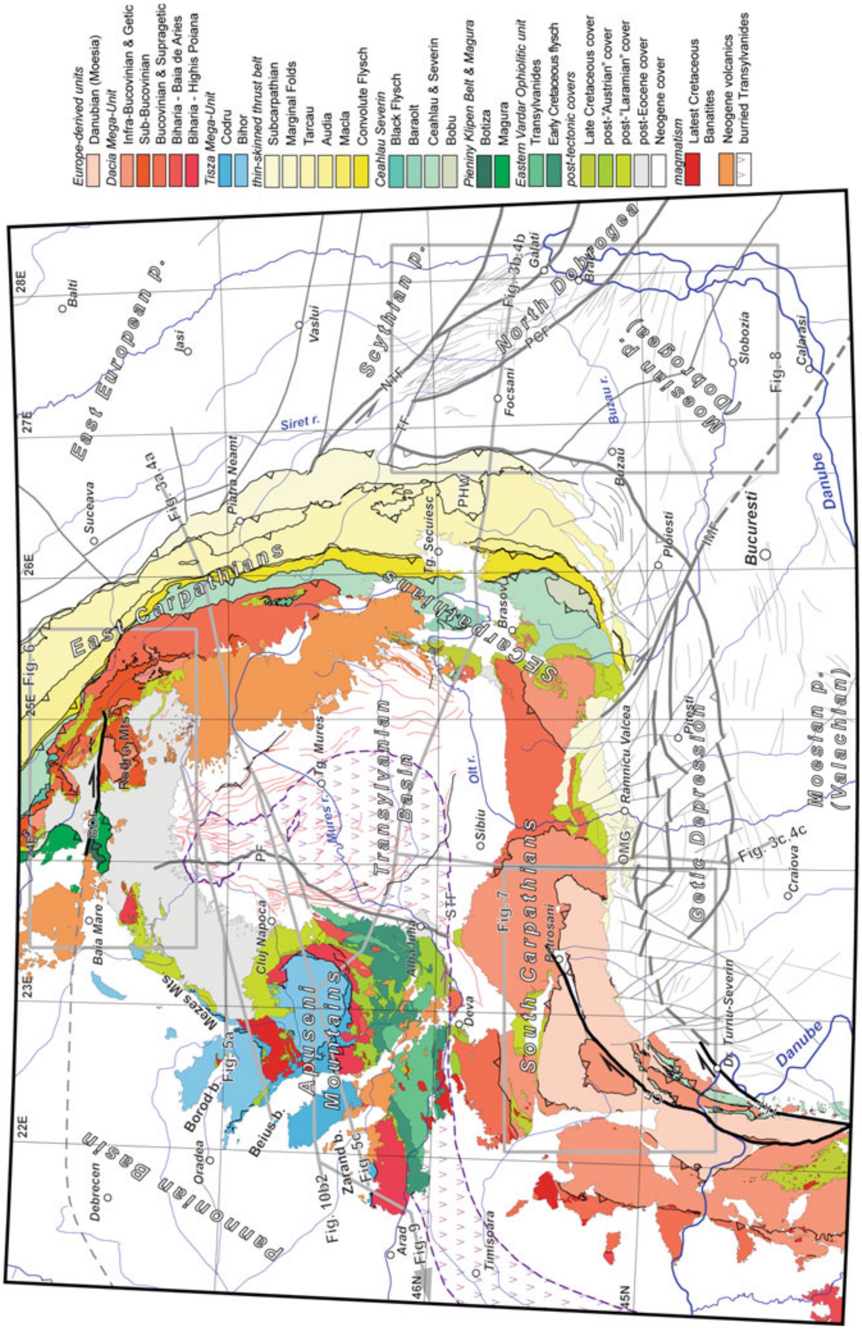

Fig. 2.2 Tectonic map of the Romanian Carpathians (derived by the compilation of 1:50,000 and 1:200,000 maps of the Geological Institute of Romania and Mațenco et al. 2010). *Thick grey lines* are the locations of geological cross sections in Figs. 2.3, 2.4, 2.5 and 2.9. *CF*—Cerna Fault; *BDF*—Bogdan Vodă—Dragoș Vodă fault system; *IMF*—Intramoesian Fault; *TF*—Trotuș Fault; *NTF*—New Trotuș Fault; *OMG*—Ocnele Mari—Govora antiform; *PHW*—Putna half-window

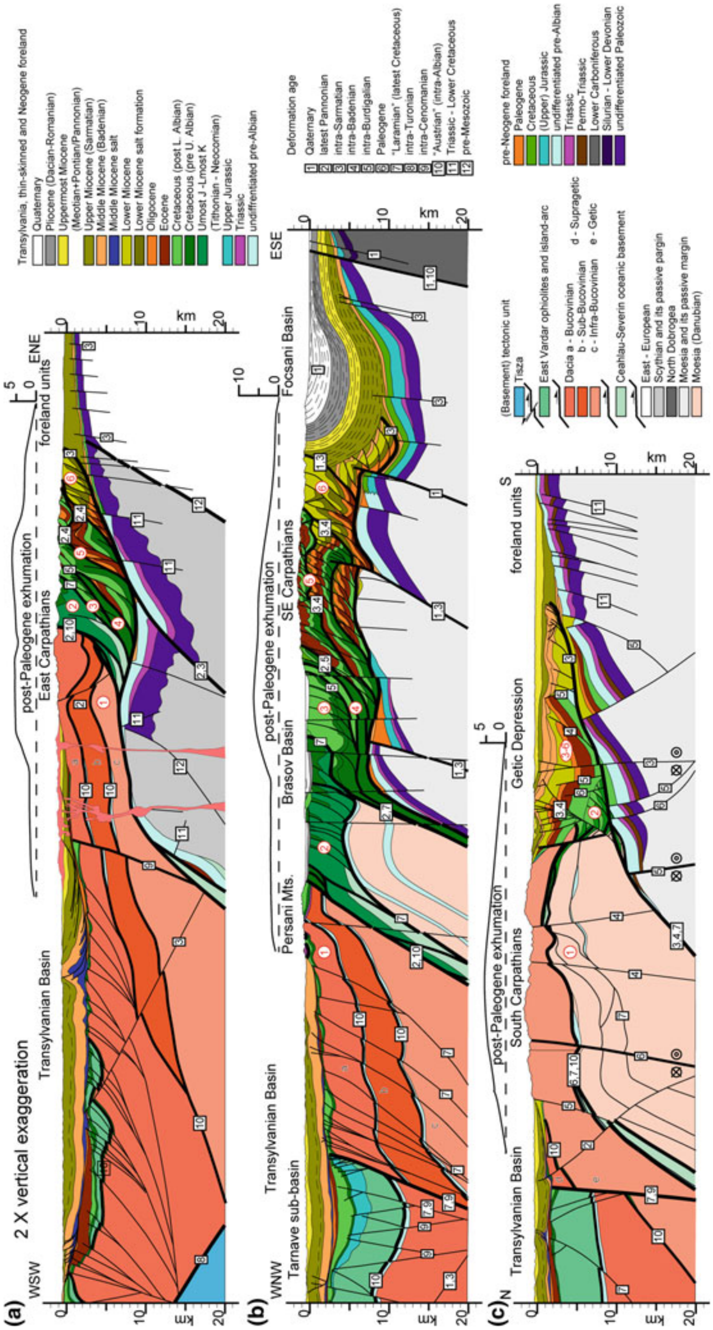

Fig. 2.3 Representative cross sections and corresponding amounts of post-Paleogene exhumation along three transects crossing the (**a**) East, (**b**) SE and (**c**) South Carpathians (modified from Mațenco et al. 2010). The location of cross sections is displayed in Fig. 2.2. *Black numbers* indicate deformation ages, *red numbers* indicate tectonic units. Note the 2× vertical exaggeration, for further details see the text

Visarion 1988; Stephenson et al. 2004; Saintot et al. 2006). The SW margin of these units is often described as the prolongation of the NW–SE striking Teisseyre–Tornquist zone, essentially the limit between Proterozoic and Paleozoic European units that was affected by significant deformations during Paleozoic–Cenozoic times (Ziegler Ziegler 1990; Pharaoh 1999; Stănică et al. 1999). The Cretaceous—Paleogene shortening and along-strike displacements have duplicated locally this lineament in the structure of the SE Carpathians (Fig. 2.3, Bocin et al. 2013). To the SW, the Moesian platform (Fig. 2.2) is a Gondwana-derived terrane accreted to the East European/Scythian margin during the Late Carboniferous, subsequently overprinted by significant deformations, such as Permo–Triassic rifting, large-scale Cimmerian (Triassic–Jurassic) contraction and the Miocene Carpathians collision (Visarion et al. 1988; Tari et al. 1997; Seghedi et al. 2005; Vaida et al. 2005).

The Dacia unit makes the bulk of Romanian Carpathians thick-skinned nappes (Figs. 2.1 and 2.2). This unit is a piece of the European continent that split off during the Jurassic opening of the Ceahlău–Severin Ocean (Ștefănescu 1983; Săndulescu 1988; Schmid et al. 2008). The Dacia unit was sutured back to Europe during the Cretaceous closure of this ocean and the subsequent Paleogene—Miocene subduction of its eastern thinned continental passive margin remnant, i.e. the Carpathians embayment. Ultimately, the Miocene collision established the present-day nappe geometry of the East and South Carpathian Mountains (Săndulescu 1984, 1988; Balla 1986; Ustaszewski et al. 2008; Mațenco et al. 2010; Merten et al. 2010). The Dacia unit is made up of a thick-skinned nappe stack with an overall antiformal geometry that is well exposed in the East and South Carpathians (Figs. 2.2 and 2.3). This nappe stack formed during successive events of contraction taking place in late Early to latest Cretaceous times during the closure of the Ceahlău–Severin Ocean and was subsequently deformed during the Paleogene formation of the Danubian extensional dome in the core of the South Carpathians (Figs. 2.2 and 2.3, Schmid et al. 1998; Kräutner and Bindea 2002; Fügenschuh and Schmid 2005; Iancu et al. 2005a). In the East Carpathians, the E-ward facing Bucovinian nappe stack truncates an earlier Variscan thrust system with an opposite vergence during the peak contractional moments of late Early Cretaceous shortening (Fig. 2.3a, Kräutner and Bindea 2002). In the South Carpathians, the Getic/Supragetic nappes contain a medium to high-grade metamorphic Neoproterozoic to early Paleozoic basement and low metamorphic degree Paleozoic, overlain by a latest Paleozoic—Mesozoic sedimentary cover (Fig. 2.3c, Iancu et al. 2005a, b; Balintoni et al. 2010). These basement nappes are largely covered in the connecting area of the SE Carpathians (Fig. 2.3b), but their lateral (Sub-)Bucovinian—Supragetic and Infra-Bucovinian—Getic correlation has been inferred by a combination of surface observations and deep geophysical studies (Visarion et al. 1978; Săndulescu 1984; Bocin et al. 2009, 2013). At the opposite side of the East and South Carpathians, the Biharia nappe of the Apuseni Mountains (Fig. 2.2) has been recently ascribed to the Dacia unit based on its medium to high-grade Middle–Late Jurassic and Early Cretaceous metamorphic overprint (Dallmeyer et al. 1999; Schmid et al. 2008). Its lateral prolongation is inferred from seismic studies combined with well data beneath the Paleogene–Quaternary cover of Transylvanian and Pannonian Basins (de Broucker et al. 1998; Maţenco and Radivojević 2012; Tiliţă et al. 2013).

The Tisza unit outcrops in the centre and NW part of the Apuseni Mountains, being covered almost elsewhere by the above-mentioned Paleogene–Miocene sediments of the Pannonian–Transylvania Basins (Fig. 2.2). Tisza is a continental unit with mixed affinities that separated from Europe during Middle Jurassic times. It moved southwards in a position adjacent to the Adriatic continental unit (i.e. that is genetically part of the African plate) and was realigned to European blocks (i.e. Dacia) during the late Early Cretaceous moments of closure and emplacement of the Transylvanides nappes, interpreted as a branch of the Neotethys Ocean (or East Vardar, Săndulescu 1975; Vörös 1977; Csontos and Vörös 2004; Haas and Péró 2004; Schmid et al. 2008). The Tisza continental unit is made up by a medium to high-grade Variscan metamorphic basement that is overlain by a dominant Permian–Triassic Germanic facies with significant lateral variations, including Middle Triassic shallow-water carbonates and Upper Triassic deeper water, i.e. Halstatt facies (Burchfiel and Bleahu 1976; Balintoni 1996; Haas and Péró 2004). A high-temperature Early Cretaceous metamorphic overprint with lower degree retromorphism during enhanced late Cretaceous exhumation has been recently detected in the NE-most corner of the Bihor nappe in the Apuseni Mountains (Fig. 2.2), inferring a tremendous metamorphic gradient across this dome (Kounov and Schmid 2013; Reiser et al. 2014).

The basement and sedimentary cover of ALCAPA continental unit are covered by Miocene sediments in the Romanian Carpathians (Fig. 2.1). The ALCAPA unit is made up of far-travelled Adriatic-derived continental nappes, trusted northwards during the Cretaceous–Paleogene closure of the Alpine Tethys and thick-skinned nappes emplacement of the Alps and Western Carpathians. These units were subsequently affected by counterclockwise rotations and ENE-wards translations during the Miocene extrusion of the Eastern Alps and the coeval extension of the Pannonian Basin. These movements were coeval with the Miocene shortening taking place at the exterior of the Western and Polish–Ukrainian East Carpathians (Ratschbacher et al. 1991; Tari et al. 1992; Csontos 1995; Fodor et al. 1999; Krzywiec 2001). In intra-Carpathians units, the thrusting of the ALCAPA over Dacia took place in Early Miocene times, as inferred by the final emplacement of the Piennides–Măgura nappe system in Maramureș area, and was subsequently followed by large-scale late Miocene sinistral strike-slip motions along the Bogdan– and Dragoș–Vodă faults system situated at their contact (Săndulescu et al. 1993; Tischler et al. 2007, 2008).

The ophiolite-bearing units of the South Apuseni Mountains are commonly grouped with the ones buried beneath the Miocene sediments of the Transylvanian Basin and exposed as the highest tectonic unit of the East Carpathians under the generic name of Transylvanides (Fig. 2.2, Săndulescu and Visarion 1977; Săndulescu 1984). These are thought to be derived from the Neotethys Ocean that was located SE of the Alpine Tethys. After the closure of the Paleotethys Ocean, the Neotethys Ocean started to open during Triassic times. Its northern branch was affected by large-scale Late Jurassic—Earliest Cretaceous obduction (Stampfli and

Borel 2002). The Transylvanides and the genetically associated island-arc volcanics together with their prolongation in eastern Serbia, Macedonia and Greece are interpreted as the part of the Neotethys (or Vardar in the Dinarides–Hellenides) Ocean that became thrusted or obducted over the European margin in late Jurassic times (the Eastern Vardar of Schmid et al. 2008; Maţenco and Radivojević 2012). The initial Late Jurassic emplacement of the Transylvanides was subsequently followed by their large-scale thrusting over Dacia during late Early Cretaceous times over a distance that spans from the Southern Apuseni Mountains to the East Carpathians (Figs. 2.2 and 2.3, e.g. Săndulescu and Visarion 1977; Nicolae and Saccani 2003; Săsăran 2005; Ionescu et al. 2009).

The Ceahlău–Severin unit (Fig. 2.2) contains the relicts of an oceanic embayment that opened between the European foreland and the Dacia unit, interpreted as the eastern prolongation of the Alpine Tethys, kinematically linked with the opening of the Central Atlantic Ocean (Favre and Stampfli 1992; Schmid et al. 2008). The exact moment of Jurassic opening is not fully clear in the Romanian Carpathians, owing to the almost complete subsequent subduction of the underlying oceanic basement (e.g. Săndulescu 1984; Ştefănescu 1995; Iancu et al. 2005a). However, the relationship of the Late Jurassic–Early Cretaceous Sinaia deep-water sediments with their underlying basement is rather obvious in the lateral prolongation of the Carpatho–Balkanides in Serbia and Bulgaria (i.e. the equivalent Kostel and Trojan deep-water formations, Kounov et al. 2010). It indicates that the initial opening of the Ceahlău–Severin Ocean is Middle Jurassic, coeval with the opening of the Alpine Tethys. Deformation of oceanic basement, deep-water sediments and genetically associated contractional trench turbidites combined with the overlying post-tectonic covers (such as Ceahlău conglomerates) indicate that these oceanic relics were emplaced during the late Early Cretaceous ($\sim$105 Ma) and latest Cretaceous ($\sim$70 Ma) moments of deformation in a system of nappes (Black Flysch, Baraolt, Ceahlău, Severin, Bobu) whose geometry is variable along the strike of the orogen (Fig. 2.2, Săndulescu 1975; Ştefănescu 1976; Săndulescu et al. 1981 and references therein). At the exterior of the Ceahlău–Severin unit, a wide system of nappes form the external Carpathians thin-skinned belt, well exposed in the East and SE Carpathians and buried beneath the latest Miocene–Pliocene cover of the Dacian Basin in the frontal part of the South Carpathians (Figs. 2.2 and 2.3). The various composing nappes (Convolute Flysch, Macla, Audia, Tarcău, Marginal Folds, Fig. 2.2) were emplaced successively in a foreland-breaking sequence during the Miocene subduction of the Carpathians embayment, a thinned continental to possibly oceanic domain formerly connected with the Ceahlău–Severin Ocean (Balla 1986; Săndulescu 1988; Maţenco and Bertotti 2000). The deformation culminated during the final moments of Carpathians collision and emplacement of the frontal Subcarpathian nappe at around 11–8 Ma and was followed by subsidence and differential vertical motions during latest Miocene—Quaternary times, a process that is still active today (e.g. Leever et al. 2006b; Maţenco et al. 2007, 2010; Ismail-Zadeh et al. 2012 and references therein).

The Relationship Between Tectonics and Exhumation in the Romanian Carpathian Mountains

The tectonic units presently exposed in the Romanian Carpathians or buried beneath the Neogene–Quaternary cover of the foreland and back-arc basins have suffered multiple episodes of burial and exhumation during their long poly-phase kinematic history. For instance, the metamorphic basement of the East and South Carpathians retains a dominant latest Proterozoic—Earliest Paleozoic age of metamorphism, but these rocks have been overprinted by numerous other episodes of (re-)burial and exhumation at/from various depths and rates during early or late Paleozoic, Cretaceous, Paleogene or Miocene—Quaternary times, as recorded by both high- and low- temperature thermochonlogical markers (Pană and Erdmer 1994; Medaris et al. 2003; Ciulavu et al. 2008; Balintoni et al. 2009; Merten 2011 and references therein).

We analyse the link between Carpathians exhumation and its present-day topographic expression by describing the effects of last significant tectonic exhumation event. This is obviously a fairly qualitative definition as orogenic burial and exhumation at steady state is a complex concept that assumes continuous exhumation driven by surface processes keeping pace with accretion of continental material in subduction/collision zones (e.g. Willett and Brandon 2002; Braun et al. 2014). In this study, the age of the last significant exhumation event is defined as the cooling age beneath the lowest closure temperature of available thermochronological markers.

The Romanian part of the Carpathians has a high density of low-temperature thermochronological markers that include zircon and apatite fission track (ZFT and AFT, closure temperatures $\sim 230°$ and $\sim 110 \pm 10°$, respectively, Gleadow and Duddy 1981; Brandon and Vance 1992) and apatite U–Th/He (AHe, closure temperature $\sim 75 \pm 5°$, Wolf et al. 1996). These methodologies involve rather complex procedures in orogenic exhumation that are described elsewhere (e.g. Reiners and Brandon 2006 and references therein), including the Romanian Carpathians studies cited below. The individual ages are cooling ages beneath their specific threshold temperatures, which are influences by a multitude of factors, such as the distribution of thermal gradient, surface and tectonic processes, groundwater flow, variations in diffusion controlled by the mineralogy of specific crystals (e.g. Braun et al. 2012). Thermal modelling by using multiple low-temperature thermochronometers and geological markers is constrained by AFT track length analysis (see Ketcham et al. 1999; Farley 2000; Ketcham 2005 for further details). Converting the resulting time—temperature histories into exhumation/denudation estimates has a higher degree of uncertainty than specific ages, because it involves estimating the geothermal gradient at the time of exhumation. Exhumation/burial does not translate directly into vertical movements, as thermochronological markers reflect temperatures (e.g. England and Molnar 1990). In orogenic context, exhumation may be the result of erosion resulting from tectonic uplift, in particular when the rates of exhumation are very high. A direct conclusion can be obtained only by correlation with all other available geological constraints, including kinematics across the mountain chain. If only low-temperature thermochronological markers are used to derive exhumation, the potential error bar is usually in the order of 1–2 km whenever AHe is available, depending on the geothermal gradient at the time of exhumation (e.g. Ketcham 2005).

Collision of the East Carpathians

A representative East Carpathians profile (Figs. 2.3a and 2.4a) shows large amounts of Miocene exhumation (~6 km) widely distributed over the entire orogenic area

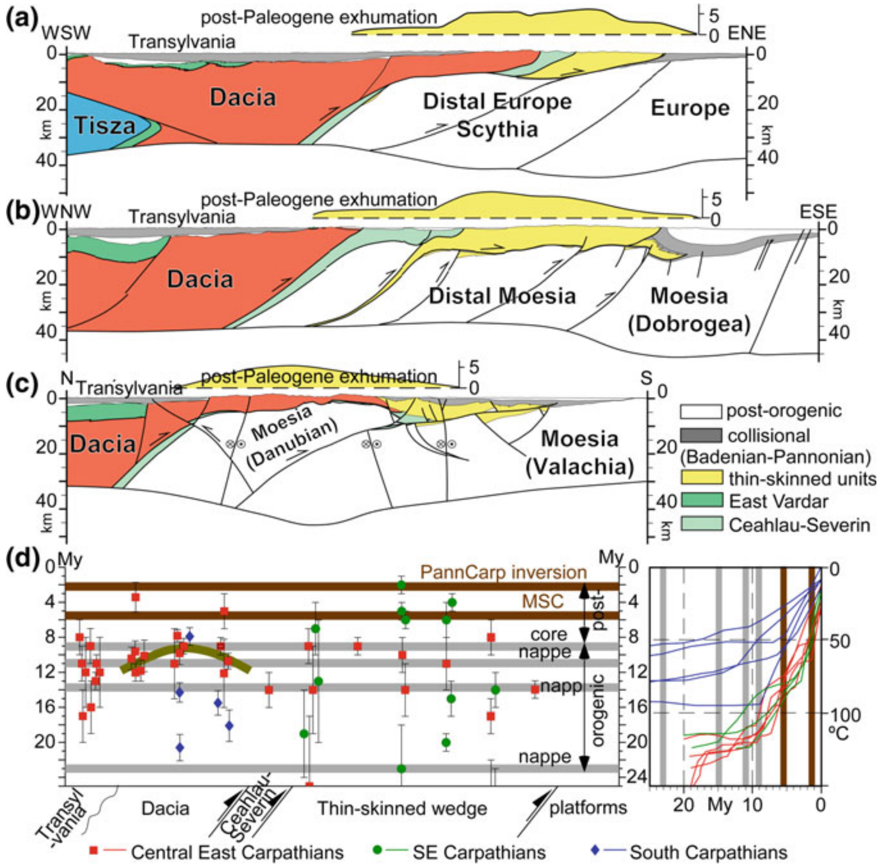

Fig. 2.4 Simplified crustal-scale version of cross sections in Fig. 2.3 by underlying the main tectonic units of the (**a**) East, (**b**) SE and (**c**) South Carpathians and synthetic representation of the apatite fission track data across the three areas an their thermal modelling (**d**) (after Maţenco et al. 2010). Note that only Miocene ages are plotted. For further details, see the text

from the eastern margin of the Transylvanian Basin to the centre of the thin-skinned contractional wedge, which gradually decreases towards the foreland (Maţenco et al. 2010).

AFT ages indicate that most of this exhumation took place during the last stages of collision, ~15–11 Ma (Sanders et al. 1999). This overall Middle Miocene exhumation corresponds roughly to the age of emplacement of the frontal Carpathians nappes (Tarcău, Marginal Folds, Subcarpathian), as demonstrated by the study of post-tectonic covers (e.g. Săndulescu et al. 1981; Maţenco and Bertotti 2000). The overall distribution of exhumation (Fig. 2.4a) infers that these tectonic events have affected not only the frontal nappes, but was distributed with similar cumulative amplitudes over the entire orogen. In more details (Figs. 2.3a and 2.4a, d) exhumation ages correlated with the main moments of uplift recorded by tectonic sequence stratigraphy studies in the Transylvanian Basin (Fig. 2.5) suggested that

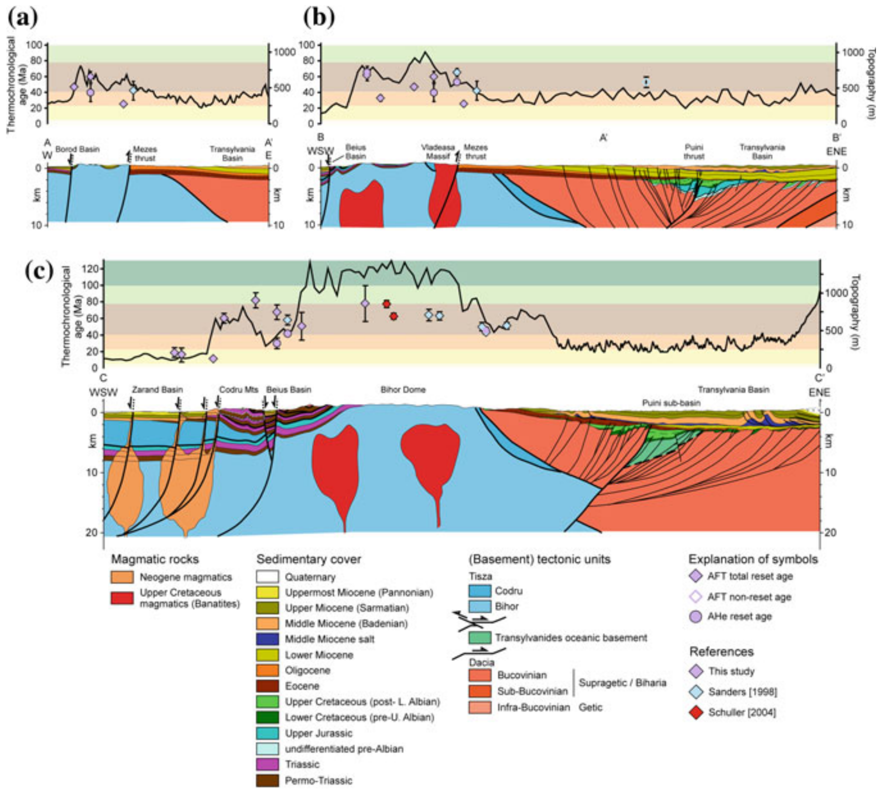

Fig. 2.5 Cross sections over the Apuseni Mountains and neighbouring parts of the Transylvanian and Pannonian Basins, and ages of exhumation derived from apatite fission track and apatite U–Th/He data (modified from Merten et al. 2011). Note the 2× vertical exaggeration. Locations of cross sections are displayed in Fig. 2.2. **a** Cross section over the Mezeş Mountains. **b** Cross section over the Vlădeasa Mountains. **c** Cross section over the Bihor dome

enhanced exhumation took place during 13–8 Ma in two main moments (Maţenco and Bertotti 2000; Krézsek et al. 2010). A first moment of in sequence nappe stacking (~14–10 Ma) deformed the above-mentioned frontal Carpathians nappes and is correlated with three pulses of uplift observed by sequence stratigraphy along the Transylvanian Basin margins (Late Badenian, Middle and Late Sarmatian), which infer a cumulated uplift in the order of 2.7 km of the adjacent East Carpathians. The second moment of out-of-sequence contractional deformation (~10–8 Ma) has resulted in enhanced exhumation of the hinterland and exaggeration of the Bucovinian antiformal nappe stack. This is correlated with two pulses of uplift observed by sequence stratigraphy along the Transylvanian Basin margins (Early and Late Pannonian) that infer a cumulated uplift in the order of 2.3 km in the core of the Bucovinian nappes. This late stage of enhanced exhumation in the orogenic core is genetically linked with a migration of contractional deformation from the former subduction zone to a crustal-scale thrust fault truncating the Scythian lower plate, creating an overall anticlinal ramp structure in the overlying Bucovinian units (Fig. 2.4a).

More recent studies have combined AHe data with existing AFT into a higher resolution exhumation history of the East Carpathians (Merten 2011). This study inferred a more symmetrical distributed Miocene exhumation, estimates ranging from 1.5 to 2.7 km for the Transylvanian Basin to 4.3 km in the centre of the Audia nappe at a rate of ~0.7 ± 0.1 mm/year, decreasing gradually to 2 km in more external thin-skinned nappes.

In this overall East Carpathians evolution, localized uplift occurred during Miocene times in the region of Rodna Mountains (Figs. 2.2 and 2.6). Zircon fission track data have demonstrated that the Rodna Mountains was affected by significant amount of cooling during Late Cretaceous times (~85–70 Ma, Santonian–Campanian) coupled with slightly earlier exhumation in the Bucovinian basement located northwards (~99–95 Ma, Cenomanian). This has been interpreted as a large-scale extensional event not yet quantified (Gröger et al. 2013), but most likely a detachment uplifting the core of the Rodna Mountains in its footwall. This exhumation event has been subsequently followed by reburial followed by a Miocene exhumation depicted by AFT and AHe data attributed to deformation near the major Bogdan Vodă–Dragoș Vodă fault system (Fig. 2.6, Gröger et al. 2008). Because the ZFT data were not reset by the reburial and exhumation, the Miocene uplift must have amplitudes in between the AFT and ZFT cooling ages, which translate somewhere between 3–5 and 7–10 km at normal geothermal gradients (Fig. 2.6). This Miocene exhumation spans in the interval 15–10 Ma and is divided into a two-phase deformation history, a first 16–12 Ma thrusting event (exhumation ages 15–13 Ma) being subsequently followed by the exhumation of Rodna Mountains during the sinistral strike-slip transpression taking place at 12–10 Ma (exhumation ages 13–8 Ma). The latter localized 4–5 km exhumation in the Rodna Mountains added to the overall pattern of collisional exhumation, which in this area has values in the order of 1–2 km (Tischler et al. 2007; Gröger et al. 2008; Merten 2011).

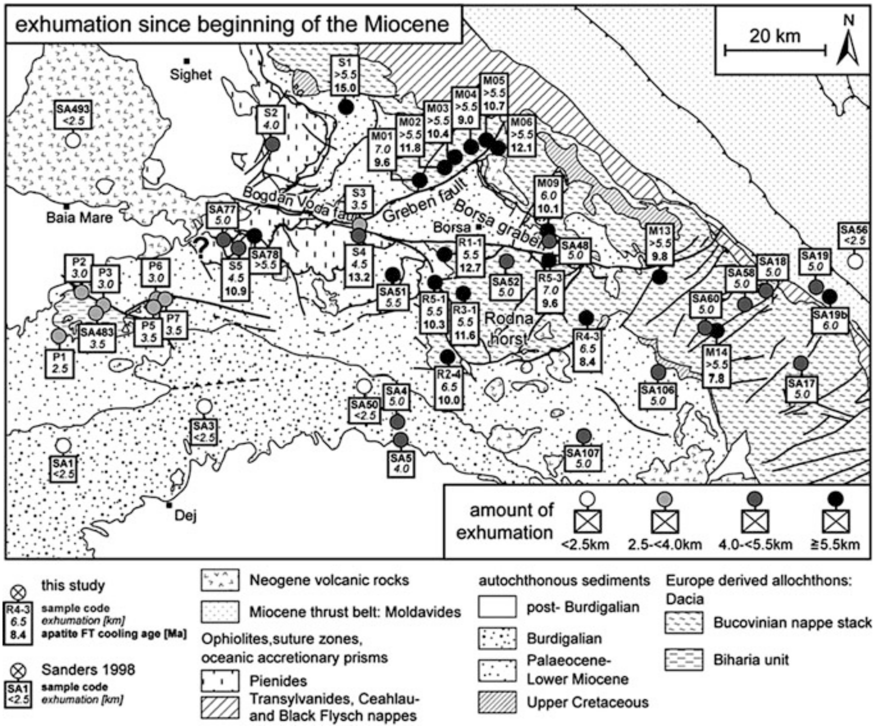

Fig. 2.6 Map showing ages and amounts of exhumation as derived from apatite fission track data in the northern part of the East Carpathians, Rodna Mountains and adjacent areas (Gröger et al. 2008). Location of the map is displayed in Fig. 2.2

Out-of-Sequence Thin- Versus Thick-Skinned Accretion in the SE Carpathians

A representative profile based on AFT data indicates an unusual situation in the SE Carpathians (Fig. 2.4b). The amounts of Miocene exhumation in the Bucovinian nappes and the ophiolitic-bearing Ceahlău unit are fairly low and contrast with the much higher, up to 6 km cumulative exhumation values in more external thin-skinned nappes. This exhumation geometry is at odds with normal critical wedge tapper geometries in collisional orogens, where the maximum amounts of exhumation are recorded in the upper plate hinterland. A number of thick-skinned thrusts truncate the tectonic lower plate (i.e. Moesia) and post-date the emplacement of overlying thin-skinned nappes (Fig. 2.4b). These out-of-sequence thrusts have been inferred by the near-surface geometry of nappes and were confirmed by deep geophysical studies, including reflection/refraction seismics, tomographic inversion, ray-trace analysis and gravity/magnetic modelling (e.g. Hauser et al. 2007; Bocin et al. 2005, 2009, 2013). The thick-skinned thrusting is associated with large amounts of Quaternary differential subsidence and uplift recorded in the Focșani Basin and the neighbouring external nappes of the SE Carpathians, respectively (Leever et al. 2006b; Mațenco et al. 2007). The continuous subsidence creating the exceptionally thick (up-to-13 km) Miocene–Quaternary sediments in Focșani Basin (Fig. 2.4b, Tărăpoancă et al. 2003) is related to the pull exerted by the Vrancea slab of the SE Carpathians (Fig. 2.4b, Tărăpoancă 2004; Mațenco et al. 2007, 2010). This pull was explained by a number of different geodynamic models explaining the deep mantle structure of the SE Carpathians (see the review of Ismail-Zadeh et al. 2012 and references therein).

More recent AFT and AHe data have inferred a higher resolution exhumation history of the SE Carpathians (Merten et al. 2010; Necea 2010). The results indicate that exhumation migrated towards the foreland, ages degreasing from Cretaceous in internal basement nappes to Miocene–Quaternary in the external part of the thin-skinned wedge (Merten et al. 2010). The Miocene nappe thrusting exhumation occurred at rates of around 0.8 mm/year, which is the same order of magnitude when compared to the East Carpathians. However, in the SE Carpathians this deformation has been overprinted by two subsequent events of out-of-sequence thick-skinned thrusting, possibly enhanced by the sea-level drop that took place near the Miocene—Pliocene limit in the Paratethys during the Messinian Salinity Crisis. These latest Miocene—Early Pliocene and Pleistocene exhumation events had much higher rates in the order of 1.6–1.7 mm/year (Merten et al. 2010). These rates have been confirmed by a high-resolution profile located in the northern part of the SE Carpathians (Putna Valley, Necea 2010). In this area, similar migration of exhumation from Cretaceous to Quaternary from the internal to the external part of the orogen has been observed, the Quaternary exhumation being overly exaggerated in the area of the Putna half-window (Fig. 2.2). Here, the youngest 1 Ma AHe exhumation age from the entire Carpathian Mountains has been recorded, inferring peak Quaternary exhumation in the order of 5 km (Necea 2010).

The total amount of Miocene shortening recorded by the thin-skinned thrust wedge of the East and SE Carpathians is laterally variable along the strike of the orogen. It increases northwards from 140 to 160 km in the northern part of the SE Carpathians to around 220–240 km in the Polish segment of the East Carpathians (Roure et al. 1993; Ellouz and Roca 1994; Ellouz et al. 1994; Morley 1996).

Nappe Stacking, Orogen Parallel Extension and Transcurrent Movements in the South Carpathians

Available exhumation data in the South Carpathians indicate that the bulk of their last event of tectonic exhumation took place during (Late) Cretaceous—Paleogene times (Bojar et al. 1998; Sanders 1998; Schmid et al. 1998; Sanders et al. 2002; Fügenschuh and Schmid 2005; Merten 2011), significantly predating the main Miocene–Quaternary exhumation recorded by the East and SE Carpathians.

The main tectonic processes was the closure of the Ceahlău–Severin Ocean that took place coeval and was subsequently followed by large-scale ~90° clockwise rotations accompanying the N- to E-ward translations of the Tisza–Dacia unit around the Moesian promontory (e.g. Ratschbacher et al. 1993; Csontos 1995). This process started with the late Early Cretaceous (intra-Albian, ~105 Ma) onset of closure of the Ceahlău–Severin Ocean accompanied by the thrusting of the Supragetic over the Getic units. This was subsequently followed by coupling of the Moesian margin, over-thrusting of the Getic–Supragetic system and the formation of the Danubian nappes (Fig. 2.2) during late Cretaceous (late Campanian—Early Maastrichtian, ~80–68 Ma) times (e.g. Berza et al. 1983; Săndulescu 1984; Iancu et al. 2005a and references therein). The latter led to burial of the Danubian nappes that were affected by an overall sub-greenschists facies metamorphism (Ciulavu et al. 2008). These events are in agreement with the distribution of ZFT and AFT data that indicate dominantly a Cretaceous background in terms of exhumation ages in the Getic–Supragetic system (Fig. 2.7).

At the end of these contractional deformations, the clockwise rotations have brought the South Carpathians strike in a position that was likely sub-parallel with the Moesian margin (Fügenschuh and Schmid 2005), which is also parallel with the direction of subsequent translations into the Carpathian embayment. The continuation of deformation with such an orogenic geometry resulted in the elongation of

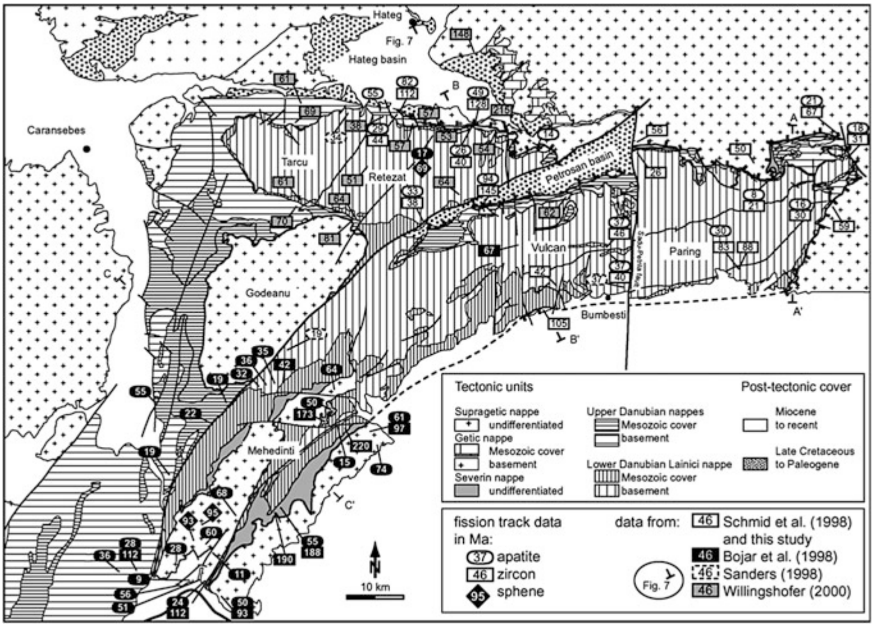

Fig. 2.7 Map compilation showing ages of exhumation in the Danubian nappes and adjacent Getic units of the South Carpathians, as derived from thermochronological studies (Fügenschuh and Schmid 2005)

the South Carpathians (i.e. orogen—parallel extension) and the initiation of large-scale transcurrent motions in respect to the Moesian margin. In a first stage, the orogen parallel extension formed an elongated extensional dome that exhumed the previously buried Danubian nappes (Schmid et al. 1998; Mațenco and Schmid 1999). This overall process culminated during Late Eocene times with the formation of the presently top to E–NE Getic detachment that reactivated the earlier late Cretaceous Getic–Danubian thrust. The gradual erosion of the Cretaceous orogen coupled with the mechanical variability of the orogen—parallel extensional in relationship with segmenting strike-slip and normal faults created a pattern of exhumation where the ZFT and AFT ages are widely distributed throughout the latest Cretaceous—Eocene times in the Danubian nappes, but are generally younger E-wards (Fig. 2.7, Willingshofer et al. 2001; Fügenschuh and Schmid 2005).

The N to NE translations and rotations of the South Carpathians accommodated initially by the orogen parallel extension continued during Oligocene—Lower Miocene times by the activation of the Cerna–Timok system of large-offset curved strike-slip faults cumulating ~ 100 km total dextral offset (Fig. 2.2). The Cerna Fault accommodate ~ 35 km of Early Oligocene dextral offset (Berza and Drăgănescu 1988), its rotational kinematic resulting in transtension, most relevant by the formation of the Petroșani Basin (Fig. 2.2, Ratschbacher et al. 1993). This was subsequently followed by the activation of the curved dextral Timok Fault that cumulates one of the largest offsets (~ 65 km) observed in continental Europe (e.g. Krautner and Krstic 2003). The age of this fault is still uncertain, but is likely late Oligocene—Early Miocene. To the S, the Timok offset is gradually transferred to the thrusting of the Srednogorie and Western Balkans units (Fig. 2.1). To the ENE and E, the Timok offset was significantly overprinted by the subsequent Miocene deformation of the Getic Depression characterized by strike-slip (mostly dextral) and thrusting of the South Carpathians over the Moesian platform (Răbăgia and Mațenco 1999; Tărăpoancă et al. 2007; Răbăgia et al. 2011; Krézsek et al. 2013). The Timok Fault is the likely precursor of the modern contact between the Carpathian units and the Moesian platform (the S Călimănești–Tg. Jiu Fault of Visarion et al. 1988). It was connected during Oligocene—Early Miocene times with the Carpathians embayment by transferring its dextral offsets to the thrusting of internal Moldavides (Audia, Macla and Curbicortical Flysch nappes, Fig. 2.2) in the East and SE Carpathians (Schmid et al. 2008; Krézsek et al. 2013).

These events were followed by the Middle–Late Miocene transpressional docking of the South Carpathians against the Moesian platform and thrusting emplacement of thin-skinned sediments of the Getic Depression. A regional profile crossing roughly the central part of the South Carpathians (Fig. 2.3c, see also a more recent interpretation based on modern wide angle reflection data of Krézsek et al. 2013) indicates a roughly symmetric geometry of the Miocene exhumation that reached maximum values of ~ 5 km in the centre of the orogen, decreasing rapidly elsewhere. This is in agreement with the much lower amounts of thrusting of the Getic Depression (up to ~ 35 km) when compared with East and SE Carpathians, and with exhumattion data, only few AFT Miocene ages as young as ~ 8 Ma being obtained in the centre of exhumation (Fig. 2.4c, d). A larger number of AFT exhumation ages distributed throughout the Miocene have been obtained in the western part of the Danubian nappes (Bojar et al. 1998), near the connecting area with the Serbian Carpathians and the Balkanides (Figs. 2.1 and 2.2). These are likely the result of higher amounts of tectonic uplift driven by the combination between Miocene thrusting and transpression.

Higher resolution exhumation studies based on AFT and AHe data suggest that the exhumation of the South Carpathians was a more continuous process that started already in Paleogene times with the formation of the core complex and gradually decreased with lower values into the Miocene as a combination between tectonic induced uplift with low values and the gradual erosional breakdown of an earlier built orogen (Merten 2011). This uplift gradually increases E-wards towards the connection with the SE Carpathians, where what was thought to be Miocene tectonic exhumation started already during late Oligocene times (Merten et al. 2010; Necea 2010), confirming the connection with the Timok fault and the continuity of deformation.

The Orogenic Evolution and Subsequent Stability of the Apuseni Mountains

Owing to their high structural complexity, the Apuseni Mountains have the most distributed exhumation history in the entire Romanian Carpathians. The multi-phase deformation at the contact between Tisza and Dacia combined with poor outcrops exposures and widespread covering by post-tectonic Cretaceous, Paleogene and Miocene–Quaternary sediments or volcanics makes them also the least understood mountains in the entire chain. On the overall, the results of low-temperature thermochronology studies (Sanders 1998; Sanders et al. 2002; Schuller and Frisch 2006; Schuller et al. 2009; Merten et al. 2011; Kounov and Schmid 2013; Reiser et al. 2014) agree that the exhumation of the Apuseni Mountains took place during two main periods.

The first late Jurassic—late Cretaceous (Early Campanian) exhumation period was driven by multi-phase large-offset structures, often with opposite polarities and mechanics (contraction vs. extension). Following their Middle Jurassic separation, the onset of the contractional deformation at the contact between Tisza and Dacia started during the late Jurassic ($\sim$Tithonian) emplacement of the East Vardar ophiolites and island-arc volcanics, presently exposed in the Southern Apuseni Mountains, over the Dacia margin (Schmid et al. 2008). This was possibly driven by obduction (emplacement of oceanic lithosphere over a continent) or island-arc collision, as the coeval emplacement of similar ophiolites is documented all along the European-derived margin, from Romania to Greece (Fig. 2.1). This event is associated with the deposition of a shallow-water Late Jurassic—Early Cretaceous carbonatic sequence directly over the Transylvanides ophiolitic sequence in the Southern Apuseni Mountains and their continuation beneath the Transylvanian

Basin and East Carpathians (e.g. Săsăran 2005). However, differently from elsewhere, the final emplacement by thrusting of the Transylvanides over the top of the Bucovinian nappe pile took place much later, during late Early Cretaceous times (Săndulescu and Visarion 1977; Săndulescu 1984). In the Apuseni Mountains, these events are recorded by burial in high-T thermochronology and by exhumation recorded by ZFT data in the Biharia nappes, Transylvanides and their sedimentary cover (Schuller 2004; Kounov and Schmid 2013; Reiser et al. 2014). Given the widespread ages from latest Jurassic to late Early Cretaceous ($\sim$140–100 Ma), it is rather unclear if these were two separate events or just one continuous process leading to the $\sim$150 km inferred thrusting of the Transylvanides up to the East Carpathians. This complex deformation was followed by the main intra-Turonian deformation that created the Biharia/Codru/Bihor nappe stack of the Apuseni Mountains with a presently top-NW transport direction. This event has no reasonable attached mechanism in Romanian literature, as is not observed in the East, SE and South Carpathians. Interestingly, the restoration of latest Cretaceous—Miocene $\sim$90° clockwise rotations and Miocene extension of the Pannonian Basin (e.g. Ustaszewski et al. 2008) indicate that this deformation has the required SW vergence, timing and proximity to be driven by a Dinarides event during the closure of the Neotethys (Dimitrijević 1997).

The second exhumation period was driven by a long-wavelength contractional uplift of the Bihor dome during latest Cretaceous—Eocene times and by lower offset Oligocene (possibly also Earliest Miocene) thrusting that took place at the contact with the Transylvanian Basin sediments in the area of the Mezeș Mountains, Puini thrust and its southern prolongation (Figs. 2.2 and 2.5). These events are observed in AFT and AHe thermochronology combined with field kinematics (Fig. 2.5, Merten et al. 2011).

In contrast with the East, SE and South Carpathians, the Miocene tectonics did not created any significant exhumation in the Apuseni Mountains. The few low-temperature thermochronological ages obtained are related to the magmatic cooling of Neogene volcanics (Fig. 2.5, Merten et al. 2011). The relative Miocene stability at AHe resolution of the Apuseni Mountains ($\sim$1 km given the high geothermal gradients) is in sharp contrast with the coeval large amounts of subsidence recorded by the Pannonian and Transylvanian Basins, and with the large uplift of the East, SE and South Carpathians. In fact, the Apuseni Mountains are a relic of a much wider and continuous mountain chain that occupied the space between the Dinarides and the Carpathians at the end of Paleogene times.

Dynamic Topography in the Transylvanian Basin

One of the most striking European examples of dynamic topography is the Miocene formation and evolution of the Transylvanian Basin (Fig. 2.2). The up to 3.5 km thick Middle–Upper Miocene sedimentary cover has an apparent symmetric geometry both in cross sections and in map views (Figs. 2.2 and 2.3). It overlies an earlier basement and cover affected by large-scale tectonic movements that include the late Early Cretaceous emplacement of the ophiolite-bearing Transylvanides nappes drilled by exploration wells and inferred from the magnetic and gravity anomalies in the central and western part of the basin (Figs. 2.2 and 2.3, e.g. Ciupagea et al. 1970; Săndulescu and Visarion 1977; de Broucker et al. 1998; Ionescu et al. 2009; Tiliţă et al. 2013). This was followed by the extensional formation of two main late Cretaceous sub-basins (Târnave and Puini, Figs. 2.3b and 2.5c) that was likely connected with the coeval extensional exhumation of the Rodna Mountains (Gröger et al. 2008). These two sub-basins were inverted in two subsequent contractional episodes that took place during the latest Cretaceous and latest Eocene, which affected the entire Transylvanian Basin (de Broucker et al. 1998; Krézsek and Bally 2006). This contraction resulted in the formation of an orogenic area that was subsequently affected by large-scale erosion (the pre-Paratethys buried denudational surface of Paraschiv 1997). The deposition of a foredeep wedge in the northern part of the basin took place in response to the thrusting of ALCAPA over Dacia during late Oligocene—Early Miocene times (Krézsek and Bally 2006; Tischler et al. 2008). The onset of regional subsidence in the Transylvanian Basin took place during Middle Miocene times (Badenian), continuing with accelerated pulses throughout the Middle–Late Miocene (Filipescu and Gîrbacea 1997; Krézsek and Filipescu 2005; Tiliţă et al. 2013). The subsidence started near the East and SE Carpathians and gradually extended over the entire basin (Tiliţă et al. 2013). In parallel, the exhumation of the neighbouring Carpathians during Middle–Late Miocene uplifted the margins of the Transylvanian Basin in several pulses of movement and created forced regressive sequences with coarse deltaic deposition (Krézsek and Filipescu 2005; Krézsek et al. 2010; Mațenco et al. 2010). These vertical tectonic movements were interrupted by an eustatic sea-level drop during Middle Badenian times, which lead to the deposition of thick salt and other evaporitic sequences throughout the basin (Peryt 2006; de Leeuw et al. 2010). These evaporites migrated during the latest Middle–Late Miocene and created locally large exaggerated diapirs due to the overburden, contractional stresses and volcanic sagging (Krézsek and Bally 2006; Szakacs and Krezsek 2006; Tiliţă et al. 2013). Towards the end of the late Miocene times ($\sim$8 Ma) the entire Transylvanian Basin was uplifted to the $\sim$600 m maximum topographic elevations of the sedimentary fill, being subsequently affected by significant erosion and local deposition of Pliocene–Quaternary continental sediments (Mațenco et al. 2010).

None of these substantial Miocene vertical movements can be explained by the few low-offset faults observed in the Transylvanian Basin (Fig. 2.3). Both the initial subsidence and the subsequent regional exhumation are dynamic topography processes, i.e. related to the deep mantle evolution during the roll-back of the Carpathians slab. It is rather unclear which dynamic topography process may be responsible from the wide variety inferred, although some are more likely, such as mantle thinning during extension and its subsequent response (for a review see Tiliţă et al. 2013). Furthermore, the evolution of the adjacent intra-mountain Pliocene–Quaternary Brașov and Tg-Secuiesc Basins (Figs. 2.2 and 2.3b) and the associated alkaline and adakitic volcanism is also driven by dynamic topography processes. The formation of these extensional grabens with sediments averaging few hundreds to few tens of metres is related to the rise of the astenospheric mantle in the hinterland of the rapidly sinking Vrancea slab, as inferred by a wide array of deep geophysical observations and numerical modelling (Seghedi et al. 2011; Ismail-Zadeh et al. 2012 and references therein).

In fact, the postulation of Transylvania as a back-arc basin behind the volcanic arc of the East Carpathians is highly questionable. In mechanical terms, the Pannonian Basin is the typical back-arc driven by the Miocene roll-back of the Carpathians slab (e.g. Horváth et al. 2006). Given the coeval stability of the Apuseni Mountains and the pre-Miocene continuity of the orogen in the entire intra-Carpathians region, the Transylvanian Basin can be alternatively interpreted as an unusual large fore-arc basin overlying the frontal part of a wide orogen, as commonly discussed elsewhere (e.g. Fuller et al. 2006). In other words, the current dilemma is: can the Transylvanian Basin be considered a small back-arc or a large fore-arc basin? A response would be eventually important for deciphering the underlying mechanisms driving its evolution.

Mechanisms Driving the Miocene–Quaternary Evolution of the Carpathians Foredeep and Foreland Platforms

Typical models of orogenic mechanics (e.g. Beaumont 1981; Platt 1986; Naylor and Sinclair 2008; Cawood et al. 2009 and references therein) assume that foredeeps form in response to flexure of the continental lower plate involved in subduction and collision under load exerted by the coeval orogenic nappe stacking. This mechanism creates syn-kinematic foredeep wedges that record the moments of thrust loading and are locally involved in thrusting. In the Carpathians, this mechanism should have controlled the deposition of the undeformed foredeep and the thrusting of the Subcarpathian molasse sequence. The same mechanics assumes that the flexural isostatic rebound associated with the ceasure of contractional stresses should have created a limited amount of uplift in post-orogenic times. Along the strike of the Romanian Carpathians, only the East Carpathians segment underlain by the East European/Schythian platform units respects these general rules of foredeep evolution and post-orogenic rebound. The SE and South Carpathians segments underlined by the various units of the Moesian Platform show atypical foredeep evolution and post-collisional geometries (Mațenco et al. 1997; Bertotti et al. 2003; Leever et al. 2006a).

The external East Carpathians displays typical wedge-shaped foredeep geometries of the Miocene sediments overlying the East European/Scythian platform units (Fig. 2.3a). These deposits show a clear syn-kinematic character with lateral transitions of sedimentological facies, such as from shallow-water clastic and carbonates at far distances from the orogen to coarse clastic deep-water molasse overlying the earlier turbiditic/flysch deposition in its proximity. The latter was involved in thrusting by the final emplacement of the Subcarpathian nappe (Ionesi 1994; Miclăuș et al. 2009, 2011). This foredeep displays a characteristic transgressive–regressive cyclicity driven by the Miocene moments of thrust loading. Post-dating these moments of deformation, the entire East Carpathians—foredeep—East European/Scythian transect underwent exhumation and uplift to continental conditions. The rather high present-day topographic elevations of the foreland East European/Scythian units (Fig. 2.2) of around 300–400 m are associated with ongoing uplift (van der Hoeven et al. 2005; Schmitt et al. 2007) and suggest that the flexural rebound was associated with a process such as slab detachment in the East Carpathians after the Miocene collision (Wortel and Spakman 2000; Sperner et al. 2001).

In contrast, the external parts of the SE and South Carpathians display atypical foredeep geometries (Fig. 2.3b, c). Foredeep wedges are observed in Upper Miocene (Middle–Upper Sarmatian) sediments in the frontal part of the Getic Depression and at high depths beneath the Focșani Basin (e.g. Fig. 2.3c, see also Tărăpoancă et al. 2003; Krézsek et al. 2013 and references therein). These wedges are overlain by a thick latest Miocene–Quaternary mostly post-tectonic sedimentation. This sedimentation is part of a large basin that overlies the Moesian Platform and segments of the thin-skinned nappes of the SE Carpathians and Getic Depression (Fig. 2.2, the Dacian Basin of Jipa and Olariu 2009) and was driven by the continuous subsidence of the Moesian platform in post-collisional times (Mațenco et al. 2003, 2007). This subsidence is responsible for the low and flat topography that presently dominates the relief in most of the Dacian Basin.

The driving mechanisms of this subsidence cannot be, yet again, quantified by the study of the upper crustal structural geometries of the Moesian platform, but it is certainly driven by the evolution of the Vrancea slab. Modern high-resolution local tomography and modelling studies (Martin and Wenzel 2006) have demonstrated that the high-velocity anomaly often associated with the Vrancea slab is much larger, extending from the SE Carpathians in a SW-ward direction well beneath the Moesian platform, and can be divided in two segments. While the Vrancea segment is (bearly) still attached, the anomaly beneath the Moesian platform is apparently detached from the Moesian lithosphere (Heidbach et al. 2007). This is a typical geometry that in other similar collisional settings has been interpreted as a STEP (subduction-transform edge propagator) fault. Such lithospheric-scale structures have been often invoked in highly arcuated orogenic systems, such as in the formation of the Apennines (Govers and Wortel 2005). In the case of the South and SE Carpathians, this mechanism implies that a lithospheric tear in the Carpathians subducting slab propagated along the South Carpathians during Miocene—Quaternary times, eventually arriving in the present-day configuration. Such a hypothesis is still speculative and requires a better understanding of the coupling between mantle dynamics and near-surface deformation (Ismail-Zadeh et al. 2012).

Limited to the SE Carpathians area located between the Intramoesian and Trotuș faults and their WNW-ward prolongation (Fig. 2.2), one other mechanism is juxtaposed over the longer term pattern of subsidence, the latter reaching extreme values in the centre of the Focşani Basin (~6 km of Upper Miocene–Quaternary sediments, Fig. 2.3b). Large-scale differential vertical motions were recorded during late Pliocene—Quaternary times by ~5 km of uplift of the external Carpathians nappes and up to 2 km of subsidence in the neighbouring Focşani Basin, which accommodated a total amount of shortening in the order of 5 km (Leever et al. 2006b; Maţenco et al. 2007; Necea 2010). This process is still active today: outside the well-known intermediate mantle Vrancea earthquakes (Ismail-Zadeh et al. 2012 and references therein), there is a direct correlation between crustal seismicity and activity of recent faults derived from geophysical or neotectonic studies. Crustal seismicity correlates with thick-skinned thrusting beneath the external nappes of the SE Carpathians (Bocin et al. 2009) and with the activation of the large array of normal faults farther in the foreland (Maţenco et al. 2007), most likely genetically related with the 2013 seismicity of the Galaţi area.

These studies are important for prediction of natural hazards, such as seismicity or flooding. For instance, the correlation of active faulting with the vertical movements derived from GPS studies resulted in predicting the evolution of areas threatened by flooding in the foreland of the SE Carpathians (Fig. 2.8). This prediction infers a clear pattern of acceleration of flooding and rivers instability, which is in agreement with existing geomorphological studies (Rădoane et al. 2003).

The Extension and Inversion of the Pannonian Basin

The Pannonian Basin is made up of a large number of Miocene (half-)grabens distributed in a wide area, from the Apuseni Mountains to the Alps in the west, Western Carpathians in the north and Dinarides in the south. A transect connecting the Apuseni Mountains with the SE part of the basin (Fig. 2.9) indicates that the extensional mechanics was asymmetric and the deformation migrated in space and time across the basin, from Early Miocene to early Pontian (20 Ma or older to ~8.5 Ma, Maţenco and Radivojević 2012). The asymmetry is reflected by the formation of large-scale crustal detachments that accommodated the local formation of core complexes with significant footwall exhumation These structures are well documented near the Dinarides (Ustaszewski et al. 2010; Stojadinovic et al. 2013; Toljić et al. 2013) or speculated near the South Carpathians and Apuseni Mountains (such as the Makó–Tomnatec trough, Fig. 2.9, or the Békés Basin, Tari et al. 1999; Magyar et al. 2006; Balázs et al. 2016). In this area, the extension started near the Dinarides during Early Miocene times, continued everywhere in the basin during the Middle Miocene and finished in the area close to the Apuseni Mountains and South Carpathians during the late Miocene (Fig. 2.9).

The first and last stages of extension were associated with the formation of detachments and half-grabens, while the second Middle Miocene stage of extension was more symmetric, resulting in the widespread formation of grabens across the

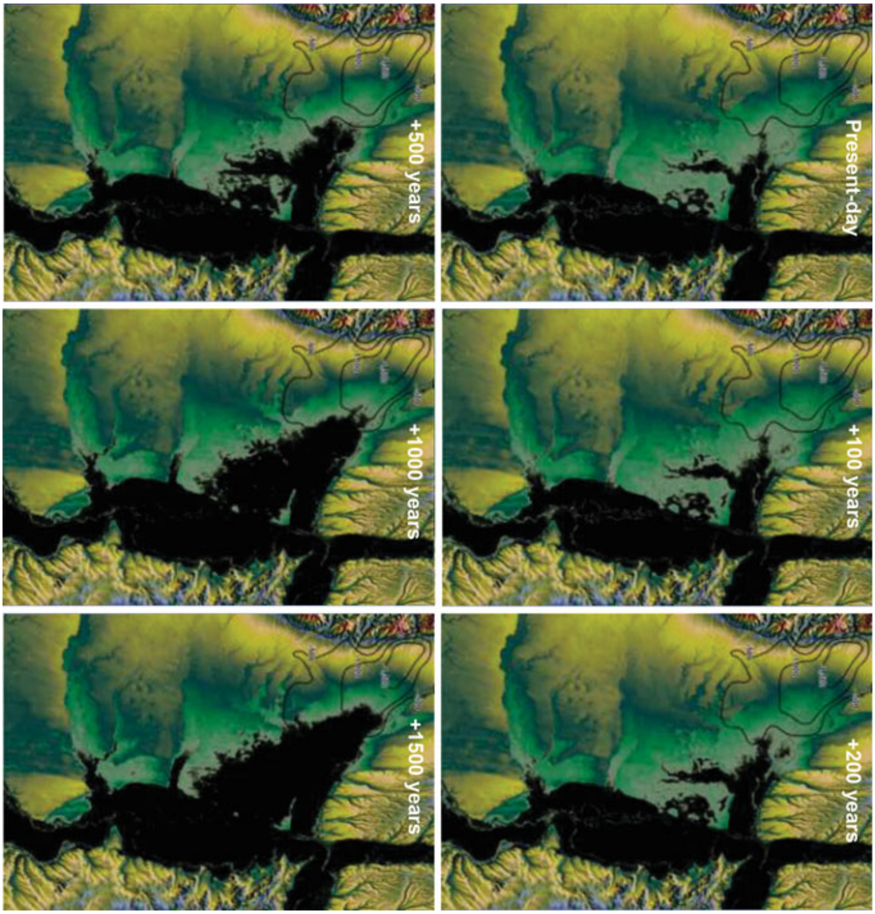

Fig. 2.8 Model prediction of the evolution of areas threatened by flooding in the SE Carpathians by correlating vertical movements derived from GPS studies in the Carpathians foreland (van der Hoeven et al. 2005) with active faulting patterns (see Maţenco et al. 2007 for further details)

basin (Maţenco and Radivojević 2012). The last, late Miocene (Pannonian–Early Pontian) stage of extension was associated with the large-scale mantle lithospheric thinning and the formation of the astenospheric upraise that is presently still observed beneath the Pannonian Basin (Horváth et al. 2006, 2015). The variability of the Miocene extensional mechanics is observed in the Miocene extensional basin cross-cutting the western margin of the Apuseni Basin and their W-ward prolongation (Fig. 2.2): the Middle Miocene Beiuş Basin is a roughly symmetric graben, while the Borod, Zarand Basins and Tomnatec Depression are highly asymmetric and are associated with footwall uplift during extension (Dinu et al. 1991; Răbăgia 2009; Maţenco and Radivojević 2012).

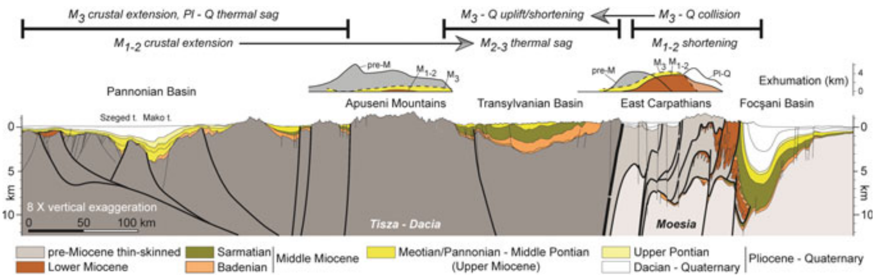

Fig. 2.9 Simplified geological cross section across the SE part of the Pannonian Basin, Apuseni Mountains, Transylvanian Basin and SE Carpathians and amounts of exhumation over the Apuseni Mountains and SE Carpathians derived from low-temperature thermochronology (modified from Maţenco and Radivojević 2012). The geological cross section displays only Miocene–Quaternary sediments geometries and faults patterns. All pre-Miocene structures were ignored. The location of the cross section is displayed in Fig. 2.1. pre-M = pre-Miocene; M_1 = Early Miocene; M_2 = Middle Miocene; M_3 = Late Miocene; Pl = Pliocene; Q = Quaternary

The latest Miocene–Quaternary inversion of the Pannonian Basin driven by the indentation/subduction of the Adriatic promontory has created thrusting near the Dinaridic margin, associated with transcurrent (both transpressional and transtensional) deformation observed in the basin centre and to the NE (Fodor et al. 2005; Pinter et al. 2005; Bada et al. 2007). The effects of this inversion are fairly reduced in the eastern Dinarides near the South Carpathians and Apuseni Mountains, being limited to the formation of either positive flower structures with uplift in the order of few hundred metres (e.g. Fruska Gora on the flank of the city of Novi Sad in Serbia, Toljić et al. 2013), or small-scale transtensional structures with similar offsets (such as the Derecske Trough west of the Apuseni Mountains, Windhoffer and Bada 2005). Quaternary uplift of the Apuseni Mountains is suggested by deep river incisions with highly elevated terrace systems (such as on the NE flank of the Borod Basin) and non-equilibrated river network (Dombrádi et al. 2007), but this uplift must be below the resolution of AHe thermochronometer ($\sim$1 km given the high geothermal gradient).

Mechanics of Continental Collision and Large-Scale Evolution of Topography

All available studies suggest that the Miocene slab retreat/roll-back of the Carpathians slab was accommodated by extension in the Pannonian Basin. The extension was accompanied by different translational and rotational kinematics in the two intra-Carpathians ALCAPA and Tisza–Dacia blocks during Paleogene—Miocene times (counterclockwise and clockwise, respectively, Csontos 1995; Ustaszewski et al. 2008). The Miocene roll-back of the Romanian Carpathians was accommodated almost entirely in the Tisza–Dacia sector of the chain, the contact with ALCAPA being somewhere in the Maramureș Piennides and the Bogdan Vodă—Dragoș Vodă fault system, Figs. 2.1 and 2.2). In this Tisza–Dacia sector of the chain, balanced cross sections and various reconstructions have inferred a total amount of shortening in the external Carpathians in the order of 120–160 km, increasing from north to south (Roure et al. 1993; Ellouz et al. 1994; Morley 1996). The amount of back-arc extension in the Tisza–Dacia sector of the Pannonian Basin is variable due to coeval clockwise rotational kinematics, decreasing from the contact with ALCAPA along the Mid-Hungarian Shear Zone towards the SE junction between the South Carpathians and Dinarides (Fig. 2.1). Available estimates indicate total amount of extension in the order of 140–180 km near the Mid-Hungarian Shear Zone (Lenkey 1999; Ustaszewski et al. 2008). These estimates of coeval Miocene extension and contraction are similar, which means that there was no large-scale absolute plate motion involved in the Carpathians shortening. In other words, the continental unit composed by the Apuseni Mountains, Transylvanian Basin, East, SE and South Carpathians simply rotated clockwise and moved E-wards into the Carpathian embayment, driven solely by pull and sink of the slab roll-back, collapsing the Pannonian Basin in the back and shortening the Carpathians in front (Fig. 2.9). There was no other driving force pushing or otherwise moving this continental unit E-wards, in agreement with the overall N-wards movement of Africa relative to Europe during Miocene times. Or, otherwise said, the mantle lithosphere of the Carpathians embayment simply sank into the asthenosphere and pulled the Romanian Carpathians E-wards.

Such a simple translation process of continental units, collapsing by extension the back-arc and shortening the foreland, is not unique. It is also interpreted in most of other highly curved Mediterranean orogens. For instance, the W-wards movement of the Betics–Rif system was accommodated by Miocene shortening at the exterior and coeval extensional collapse of the Alboran Domain, driven by the roll-back of the Gibraltar slab, while Africa–Iberia convergence was N–S to NE–SW oriented (Vergés and Fernàndez 2012). The W-wards movement of the Calabrian slab was accommodated by Oligocene—Quaternary shortening at the exterior of the Apennines and coeval back-arc extension and opening of the oceanic Liguro–Provencal and Tyrrhenian domains of the Western Mediterranean. In both situations, the velocity of roll-back was almost one order of magnitude higher than the slow N–S approach of Africa relative to Europe/Iberia (Faccenna et al. 2005, 2014). One other much larger example is the $\sim$900 km of S-ward translation of Aegean units that was accompanied by $\sim$500 km of N–S Europe—Africa convergence, while the amount of subducted material in the mantle is in the order of 1500 km (van Hinsbergen et al. 2005; Jolivet and Brun 2010).

Mechanics of Collision in the Romanian Carpathians

What is unique in the Romanian Carpathians is the location of the Miocene extensional back-arc, situated at far distances from the contractional plate margin.

Therefore, the successive amounts of shortening and associated exhumation are preserved and can be quantified to derive the mechanics of collision during roll-back. The link between exhumation and kinematics can be analysed along a transect from the Pannonian Basin to the SE Carpathians, i.e. towards the place where the slab is still preserved beneath the orogen in the Vrancea area (Fig. 2.9). Note that this transect is based only on the latest Cretaceous—Quaternary AFT and AHe data (Merten et al. 2010, 2011) and, therefore, indicate only low-temperature exhumation. This transect excludes strain partitioning at transcurrent structures, such as in the South Carpathians or Bogdan Vodă—Dragoș Vodă fault system of the East Carpathians. It also excludes the exhumation associated with intra-Carpathians deformation, such as the late Oligocene—Early Miocene thrusting of ALCAPA over Tisza–Dacia or the late Early Cretaceous collision between Tisza and Dacia and the emplacement of the Transylvanides nappes. However, it does not exclude reactivations of the latter contacts.

The distribution of pre-Miocene (latest Cretaceous–Paleogene) values (Fig. 2.9) indicates large-scale exhumation of the Apuseni Mountains and more reduced values at the western termination of the SE Carpathians, in the Bucovinian nappes and neighbouring Ceahlău–Severin units. Although such data are missing in the Transylvanian Basin, the overall pattern indicates a gradual increase of exhumation values towards the hinterland. All available kinematic data in the Apuseni Mountains, Transylvania Basin and East/SE Carpathians generally agree that the successive latest Cretaceous–Paleogene deformation events have the same kinematics and therefore are driven by the subduction that took place at the exterior of the Carpathians. This points towards large-scale exhumation increasing towards the Carpathians hinterland during the pre-Miocene, i.e. pre-roll-back times.

The pattern of exhumation changed once the roll-back started during the Early Miocene by the onset of simple translations and rotations of the Romanian Carpathians (Fig. 2.9). The Apuseni Mountains stopped their enhanced tectonic uplift and remained relatively stable and the low amounts of exhumation recorded are likely related to the erosional breakdown of an elevated topography. The entire Miocene exhumation took place in the external Carpathians thin-skinned nappes (or Moldavides, sensu Săndulescu 1988). The combination between low-temperature thermochronology and kinematics infers that individual moments of nappe accretion created exhumation in their immediate hinterland, but the overall Miocene exhumation migrated foreland-wards (Fig. 2.10b). The involvement of buoyant non-thinned platform units in the subduction zone during the late Miocene, possibly accompanied by slab detachment, stopped this process in the East Carpathians, exhumation being shifted backwards in the orogenic core between 11 and 8 Ma. By contrast, the foreland-ward exhumation migration continued in the SE Carpathians, driven by the thick-skinned accretion of blocks from the lower Moesian plate. The transition from thin- to thick-skinned enhanced the exhumation as the ~5 km recorded during the Pliocene–Quaternary has the highest rate in the post-Paleogene history.

A comparison of Carpathians kinematics with a “standard” double-vergent orogen helps to understand the significance of this orogenic mechanics (Fig. 2.10).

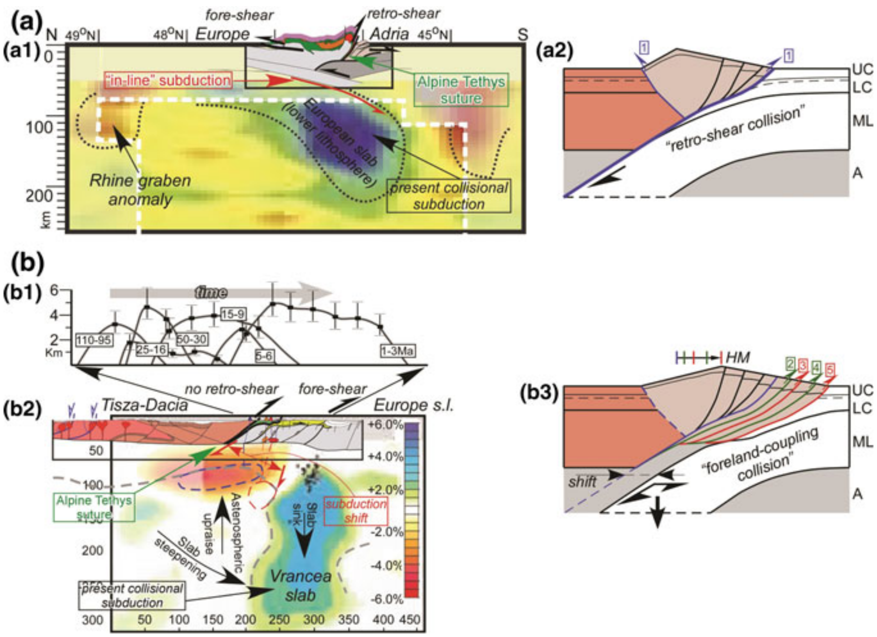

Fig. 2.10 Illustration of the concept of orogenic collision in two types of end-members orogens. **a** Tectonic cross section overlain by teleseismic mantle tomography (*a1*) along a transect crossing the Central Alps (simplified from Lippitsch et al. 2003; Schmid et al. 2004) and cartoon (*a2*) of the envisaged mechanical model of double-vergent orogenic collision (adapted from Willett and Brandon 2002). **b** Tectonic cross section overlain by teleseismic mantle tomography (*a2*) along a transect crossing the Apuseni Mountains and SE Carpathians (modified from Martin and Wenzel 2006; Maţenco et al. 2010) and cartoon (*a3*) of the envisaged mechanical model of foreland-coupling collision (adapted from Maţenco et al. 2010). The graph above the orogenic transect (*a1*) represents exhumation values and ages derived from thermochronology (compiled and simplified from Merten 2011). The exhumation data infer that exhumation migrates in time towards the foreland in the foreland-coupling collision of the Carpathians. This is in contrast with nested exhumation ages that characterize double-vergent orogens at steady state

The N–S Europe—Africa convergences have created one of the most studied orogens in the world, the Alps (Fig. 2.10a). The Middle Jurassic opening of the Alpine Tethys was followed by a prolonged history of oceanic subduction during Cretaceous—Eocene times, including the accretion of intervening Brianconnais domain (Schmid et al. 2004 and references therein). The beginning of collision with the European lower plate starting with Oligocene times is marked by the onset of deformation in the Southern Alps retro-wedge (such as illustrated in the NFP20 East profile, simplified in Fig. 2.10a1, Schmid et al. 1996). The onset of collision enhanced the exhumation along the Periadriatic lineament (the Insubric segment in Fig. 2.10a1), while the exhumation is reduced N-wards (e.g. Rosenberg and Berger 2009). The subsequent Miocene–Quaternary deformation takes place dominantly in the Southern Alps, including the lower crustal indentation of Adria (Fig. 2.10a1).

This illustrates a typical retro-wedge collision where continental material entering collision zone is exhumed by retro-shears in the hinterland by an opposite polarity deformation (Fig. 2.10b, Schmid et al. 1996; Willett et al. 2006). The resulting position of the subducted slab shows almost perfect alignment with the crustal geometry of subducted Europe (Fig. 2.10a1, Lippitsch et al. 2003).

At the other extreme, there is no retro-wedge deformation in the Romanian Carpathians during the Miocene collision (Fig. 2.10b2). Exhumation is localized over the thin-skinned wedge and shows a general pattern of migration in time towards the foreland (Fig. 2.10b1). The high-velocity anomaly of the Vrancea slab is shifted with 70–100 km towards the foreland from the correct position of the oceanic suture zone between Dacia and Moesia. This shift is the result of the thick-skinned accretion of continental blocks during the Miocene–Quaternary collision. In mechanical terms, this means that material is gradually accreted from the lower plate during collision (Fig. 2.10b3), but the rapid roll-back does not allow large-scale exhumation in upper crustal retro-shears and remains at depth. Such a process creates mechanical decoupling and migration of the slab towards the foreland, which may be associated with processes such as lithospheric delamination or lithospheric instability (Cloetingh et al. 2004; Duretz and Gerya 2013). The continental accretion of material in the lowed plate during roll-back collision is called foreland-coupling collision, as defined in the Romanian Carpathians (Mațenco et al. 2010).

Inferences for the Tectonic Age of Romanian Carpathians Topography

The mode of collision has direct implication for the present-day distribution of topography in orogens such as the Romanian Carpathians. As explained above, defining tectonic topographic ages is highly qualitative, but helps understanding the last significant tectonic event responsible for the creation of present-day relief. Such a map was defined for the Romanian Carpathians in a landmark study (Fig. 2.11, Merten 2011). This map ignores exhumation processes at higher resolution than AHe closure temperature, which is in the order of 1 km for the Miocene evolution of the Apuseni Mountains and $\sim$2 km elsewhere.

The overall topography of the Apuseni Mountains is the result of latest Cretaceous–Paleogene shortening moments that locally reactivated the inherited Tisza–Dacia contact (Fig. 2.11). This is expressed by the formation of the large Bihor dome and deformation along the Puini Fault system during latest Cretaceous—Middle Eocene times in the Apuseni Mountains and the NW part of the Transylvanian Basin. Renewed Middle Eocene–Oligocene contraction reactivated the Puini thrust and exhumed its hanging-wall at the eastern margin of the Apuseni Mountains. This is coeval with the thrusting of the Mezeș Mountains and their southern prolongation over sediments of the Transylvanian Basin (Fig. 2.11).

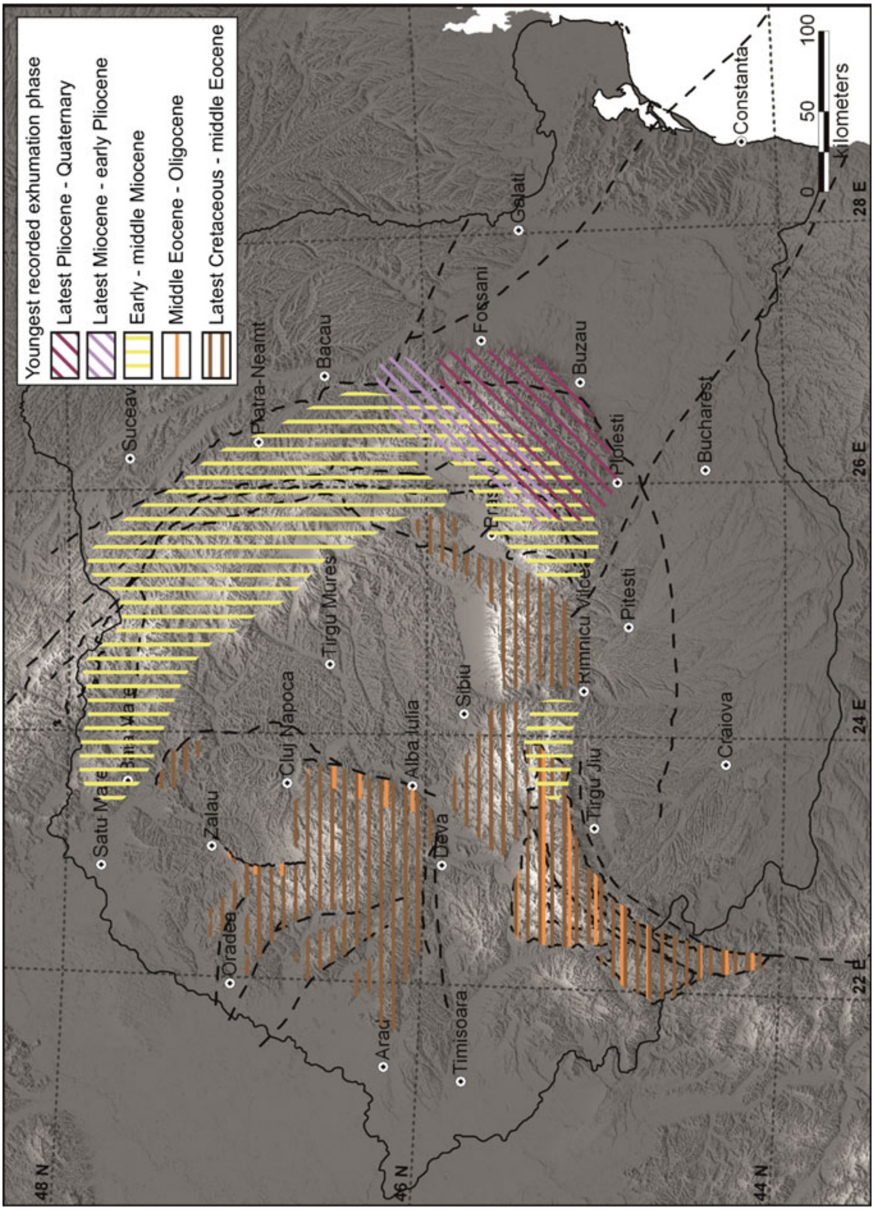

Fig. 2.11 Topographic map of Romanian Carpathians showing the age of the youngest tectonic exhumation phase (after Merten 2011), inferring the tectonic age of topographic relief. The bulk of the present-day topography in the areas of the Apuseni Mountains and South Carpathians were formed during latest Cretaceous—Paleogene times. Only a limited area in the centre of the South Carpathians indicates enhanced Miocene exhumation. The present-day topography of the entire East and SE Carpathians is the result of post-Paleogene exhumation events. The bulk of this exhumation is Miocene, being overprinted by a younger latest Miocene–Quaternary exhumation event that was restricted to the external areas of the SE Carpathians

The South Carpathians topography reflects dominantly the Latest Cretaceous—Paleogene moments of (Supra-)Getic/Danubian nappe stacking and the subsequent orogen parallel extension and transcurrent deformation along the Cerna fault (Figs. 2.2 and 2.11). Most of the Getic–Supragetic nappes in the eastern hanging-wall of the extensional detachment have a topography that is inherited from the latest Cretaceous nappe stacking and subsequent erosional breakdown. The western part of the South Carpathians displays mixed latest Cretaceous—Paleogene topography ages. Note that the overprinting Miocene exhumation in this western area is too localized to be visible in the map in Fig. 2.11. However, the Miocene exhumation has overprinted the central—southern part of the South Carpathians, most likely coeval with the formation of the Ocnele Mari–Govora anticline (Figs. 2.2 and 2.3c, Răbăgia et al. 2011). The present-day topography of the Transylvanian Basin was influenced by the Miocene moments of uplift in the neighbouring mountain chains and was ultimately established towards the end of the Miocene times by the exhumation of the basin at the $\sim$600 m maximum present elevation of Pannonian sediments. It is one of the most spectacular examples of dynamic topography in Europe.

The topography in the East and SE Carpathians is certainly inherited from Miocene times, where 4–6 km of total exhumation has overprinted all earlier events (Fig. 2.11). The topography in the external part of the SE Carpathians has been subsequently established during the two latest Miocene–Early Pliocene and latest Pliocene–Quaternary exhumation events. The overall migration of this exhumation in a foreland direction is visible in topography by the migration of the water divides, deflection of rivers and high topographic non-equilibrium (Rădoane et al. 2003; Leever 2007; ter Borgh 2013).

Conclusions

Deriving the mechanics of continental collision is of fundamental importance to understand processes driving the topographic expression of orogens. Romanian Carpathians provide a key location for understanding mechanics of collision during slab retreat because the genetically associated back-arc extension is situated at far distances and did not overprinted the nappe stacking, which is often observed elsewhere. Our review of available kinematic and low-temperature thermochronological data infers that collisional mechanics during roll-back is significantly different when compared with high-convergence orogens, such as the Alps or the Pyrenees. In most roll-back orogens, the large amount of subducted material observed by high-velocity anomalies in teleseismic tomography studies is derived from the passive translation of the upper plate accommodated by back-arc extension. In our studied case, there were no absolute plate motions during Miocene—Quaternary times, the entire subduction and accretion at the exterior of the orogen was accommodated by back-arc extension and the continental block of the Romanian Carpathians simply rotated clockwise and translated E-wards into the

Carpathians embayment. The exhumation pattern changed with time, from enhanced in hinterland during pre-roll-back times to focused near or in the thin-skinned wedge during roll-back. The latter is characterized by gradual accretion of sediments or continental basement derived from the lower subducting plate. This material was not exhumed in retro-wedges as is the common case of double-vergent orogens, but remained at high depths and, therefore, shifted the location of the main subduction zone towards the foreland. This process is called foreland-coupling collision and near the topographic surface induces exhumation that migrates towards the foreland with gradually increased values due to the transition from thin-skinned to thick-skinned thrusting.

This type of collision has significant inferences for the topographic evolution of the Romanian Carpathians. The present topographic expression of Apuseni Mountains is inherited from pre-roll-back times, when the exhumation was higher at farther distances in the hinterland of the subduction zone. The topography of the East and SE Carpathians is inherited from the moments of roll-back collision that shifted exhumation to the vicinity of the main subduction zone. The migration of exhumation towards the foreland continued during Pliocene—Quaternary times and is still active, modifying the present-day topography in the SE Carpathians. The Transylvanian Basin is one of the best examples of dynamic topography, i.e. near-surface movements induced by deep mantle processes, both during its initial Middle–Late Miocene subsidence and during the exhumation of the basin towards the end of Miocene times. The kinematics and exhumation of the South Carpathians were driven by the coupled rotational–translational movements of the Tisza–Dacia unit into the Carpathians embayment.

All these findings demonstrate the strong coupling between deep-Earth and near-surface processes. Their understanding is of fundamental importance for the evolution of continental topography and mitigation of geo-hazards.

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