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Review article

Characteristics of collisional orogens with low topographic build-up: an example from the Carpathians

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ABSTRACT

Sequence stratigraphy in the hinterland, kinematic analysis of thin-skinned thrusting in the foreland and thermochronological tracking of exhumation in the orogenic core are combined to quantify the mechanics of an orogen with low topographic build-up. The Carpathian system demonstrates that collisional deformation can couple and thicken the lower orogenic plate along reverse faults that dip more steeply than the subduction zone, defining a 'foreland-coupling' type of collision. Near the surface, this is expressed by wide antiforms in the upper plate and the thin-skinned orogenic wedge. A sequence stratigraphic analysis of the back-arc Transylvanian Basin demonstrates that the sedimentary architecture records orogenic uplift pulses with both short and long wavelengths. These correspond to the activation of individual thrust sheets in the thin-skinned wedge and to lower-plate coupling events respectively.

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Introduction

Combined field and numerical modelling studies have demonstrated that continental collision coincides with the onset of large-scale out-of-sequence deformation. This is most commonly expressed by concentrated uplift along step-up hinterland vergent retro-shears (i.e. crustal-scale backthrusts), where the upper-plate exhumation is enhanced by denudation (e.g. Beaumont *et al.*, 1994). One example described as typical for such retro-shears is the Insubric line in the European Alps (Schmid *et al.*, 1996). In this type of mountain chain, the steady-state exhumation of an orogen can be defined by reset thermochronological ages nested against the retro-deformation front (e.g. Willett and Brandon, 2002). However, collisional coupling (i.e. far-field transmission of orogenic stresses) has been documented to be responsible for intra-plate compressional deformation in orogenic forelands (Ziegler *et al.*, 1995). This observed feature of strain partitioning in the continental lower (subducting) plate during collision is generally ignored by existing geodynamic models.

Retro-shears are less obvious in orogens that are characterized by 'soft' collision, i.e. where small amounts of bulk exhumation are recorded by thermochronological studies. This is 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 as being early or incomplete when compared with other orogens considered to have reached steady-state equilibrium (e.g. Royden, 1993). The formation of this type of mountain chain has been explained in terms of age of the subducted oceanic lithosphere, slab pull, rates of convergence between plates (Royden and Burchfiel, 1989), or in terms of the general direction of subduction (Doglioni *et al.*, 1999). Independent of their genesis, monitoring the amount of uplift during collision in such orogens is generally hampered by low rates of exhumation. Temperatures reached by exhumed rocks are likely to be near or even below the closure temperatures of low-temperature thermochronometers. In such cases, exhumation pathways may be affected by lateral transport inside the orogen and longer residence in partial retention zones, and thermochronometers tend to reflect the cumulative exhumation of multiple tectonic episodes (Reiners and Brandon, 2006). 'Soft' collision, exhumation of the orogenic cores in the absence of retro-shears, formation of back-arc basins 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 low-topography Carpathian orogen (Fig. 1) is a case where crustal-scale back-thrusting in the orogenic core, although postulated on indirect grounds at the contact with the back-arc Transylvania Basin (Sanders, 1998), has never been confirmed (Krézsek and Bally, 2006). This uncertainty raises the question as to whether collision is accompanied by back-thrusting or not. Because of low rates of exhumation during and after collision, the post-tectonic cover is still present, allowing biostratigraphic dating of nappe-stacking events and discrimination from stages of post-orogenic deformation (Matenco *et al.*, 2007). Improving the resolution of thermochronometers with alternative methods aiming to determine tectonic

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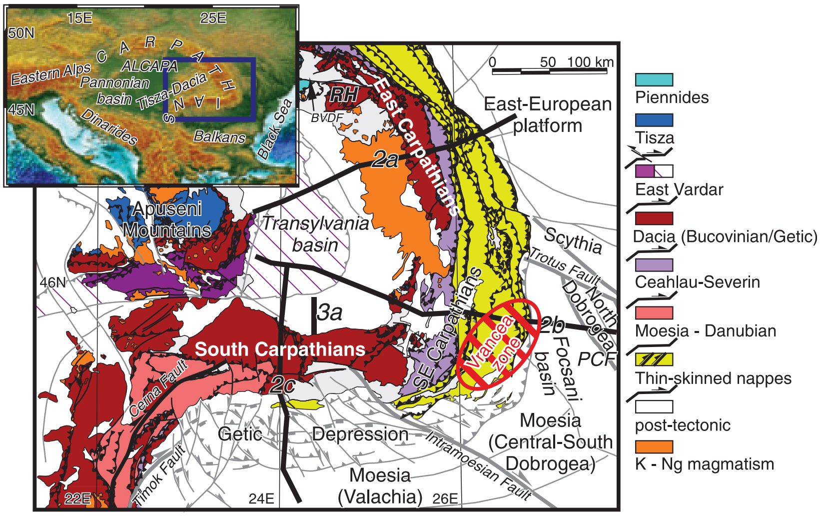


Fig. 1 Tectonic sketch of the Carpathian Mountains bordering the Transylvanian Basin with the location of cross-sections in Figs 2 and 3a (simplified from Matenco *et al.*, 2007 and Schmid *et al.*, 2008). The upper left map shows the location in the Eastern Alps–Carpathians–Dinaridic system. The East-European and Scythian platforms are grouped under the generic term of Europe. The Vrancea zone marks the rough extent of the concentration of intermediate mantle earthquake epicentres in the SE Carpathians. BVDF, Bogdan-Voda Dragos-Voda faults system; RH, Rodna horst.

uplift is important for understanding collisional mechanics of low-topography orogens such as the Carpathians. One pertinent example is the analysis of third-order sedimentological base-level variations (e.g. Van Wagoner *et al.*, 1990) that create regional unconformities in the low-angle dipping sediments of a back-arc basin (Nystuen, 1998). Here, discrimination between tectonic uplift pulses and exhumation effects driven by other factors, such as climatic or eustatic changes, is possible given the localized influence of orogenic growth bordering the back-arc basin (e.g. Catuneanu *et al.*, 2009).

The Carpathians: framework of kinematics and collision

The Carpathian mountains are the result of a Triassic to Tertiary evolution of continental blocks and intervening oceans. The continental blocks comprise the interior Tisza–Dacia and ALCAPA, and the exterior European/Scythian/Moesian continental foreland with respect to the arcuate Romanian Carpathians (e.g. Csontos and Vörös, 2004; Schmid *et al.*, 2008). Two oceanic domains separated Tisza, Dacia and the foreland, namely the East Vardar ocean to the west and the Ceahlău-Severin ocean to the east (Figs 1 and 2). The East Vardar Ocean was situated between the Tisza and Dacia continental blocks (Fig. 1) and closed gradually during the Late Jurassic–Cretaceous. This closure was followed by continental collision during the late Early to Late Cretaceous (e.g. Schmid *et al.*, 2008). The Ceahlău-Severin ocean, important for the Miocene Carpathian collision discussed in this study, opened between the Dacia block and the European/Scythian/Moesian foreland (Fig. 1) during the Late Jurassic. It started to close during the late Early to Late Cretaceous. The Carpathian embayment, situated in a more eastward position in front of the Carpathians, did not close until the late Middle Miocene. This concave-shaped embayment follows the present-day curved configuration of the Carpathian mountains (Fig. 1) and it was invaded by the Tisza–Dacia block in an upper-plate position during the Neogene retreat of a slab associated with the Alpine Tethys, i.e. by subduction roll-back (e.g. Ustaszewski *et al.*, 2008). All the relics of this slab were entirely subducted at the time of termination of nappe emplacement over the European and Moesian foreland at *c.* 11 Ma (Fig. 2; Matenco and Bertotti, 2000).

The Miocene, outward-vergent thrusting in the highly arcuate Carpathian orogen was coeval with extension and subsidence observed in the back-arc basins (Pannonian/Transylvanian, Fig. 1), similar to what is found in other orogens of the Africa/Europe collision zone (Faccenna *et al.*, 2004). The subsidence led to the deposition of thick Middle–Upper Miocene sediments in the Transylvanian Basin, which unconformably overlie the suture zone between Tisza and Dacia (Fig. 2; Schmid *et al.*, 2008). Post-11 Ma crustal deformation peaked near the limit between the Pliocene and Quaternary and was restricted to the area of the SE Carpathians (Fig. 2b). A pronounced sea-level drop corresponding to the Messinian Salinity Crisis is recorded in the Carpathian foreland, and potentially enhanced denudation around 5–6 Ma by increasing the exposure of source areas (Leever *et al.*, 2010). Presently, strong mantle anisotropy characterizes the Vrancea seismogenic zone of the SE Carpathians, with a high-velocity body observed in seismic tomography at depths of up to 370 km (Martin *et al.*, 2006). This has been explained by a large number of often contradicting geodynamic models involving slabs or evolution of continental blocks (see Matenco *et al.*, 2007 for a review).

Geometry and exhumation of the Carpathian Mountains

The structure of the East, SE and South Carpathians is dominated by the asymmetry inherited from the late Early Cretaceous to Miocene period of subduction and continental collision between the Tisza–Dacia and the European/Scythian/Moesian foreland.

The present-day East Carpathian subduction zone forms a large antiform beneath the Bucovinian basement (Fig. 2a). This is the result of deformation along higher-angle thrusts located in the foreland of the Ceahlau-Severin subduction zone, which truncate the European/Scythian lower plate (see also Stefanescu *et al.*, 1988). This late stage deformation ramps up into to the frontal sole thrust, where it is recognized as a

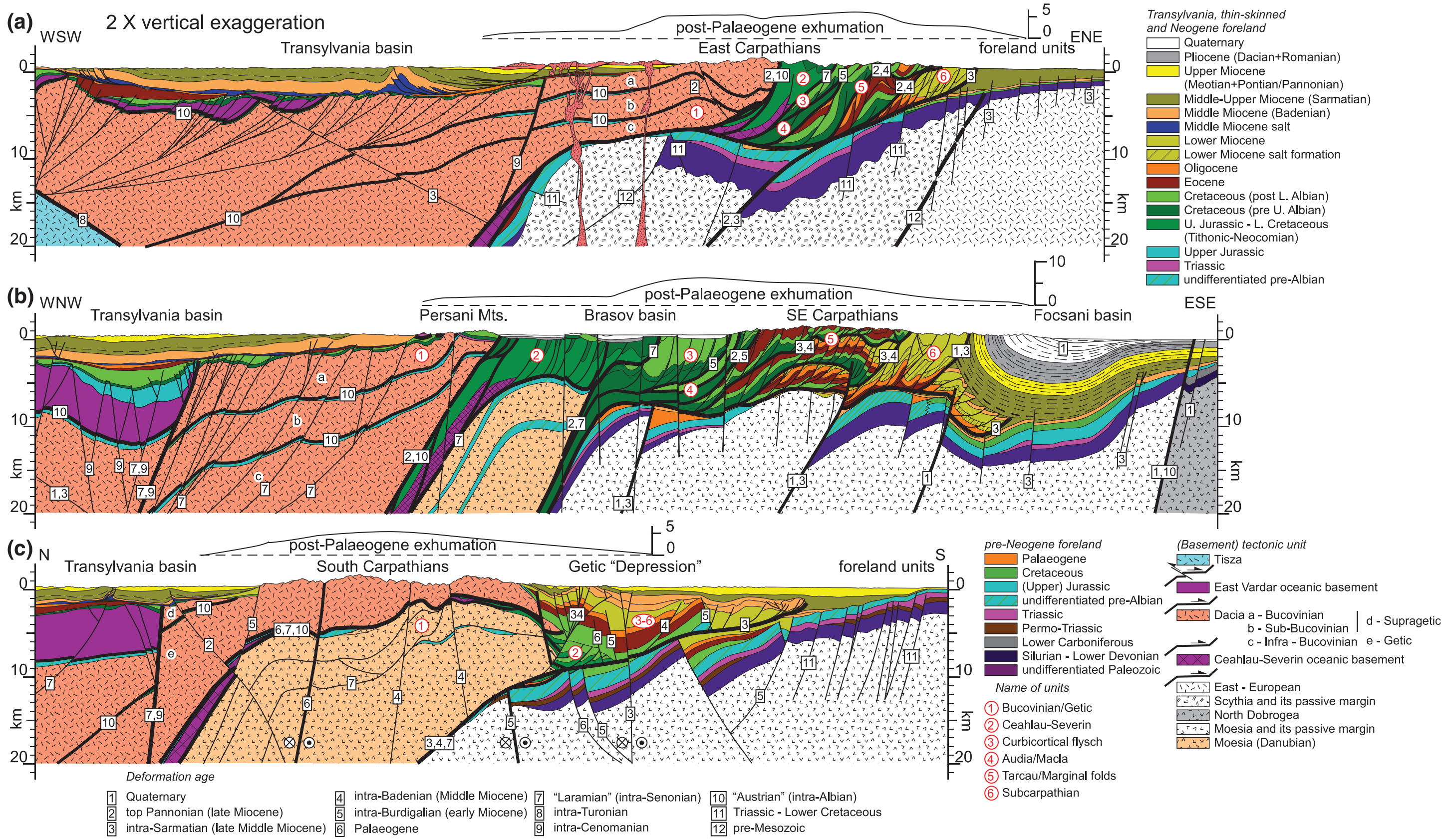


Fig. 2 Simplified cross-sections and bulk post-Palaeogene exhumation amounts along three representative cross-sections in the Romanian Carpathians (location of the sections is displayed in Fig. 1). Note the 2X vertical exaggeration. The amounts of post-Palaeogene exhumation are taken from Sanders *et al.* (1999). Numbers indicating deformation ages are derived from the overlying post-tectonic covers and are described in Matenco and Bertotti (2000), Matenco *et al.* (2007) and references therein. This excludes the previously undefined deformation 2 (top Pannonian), with its effects derived from the exhumation geometry at 9 Ma. Additional information on these ages was taken from Răbăgia *et al.* (2010) for the Getic Depression and the underlying Moesia, Fügenschuh and Schmid (2005) and Iancu *et al.* (2005) for the South Carpathians, and Leever *et al.* (2006) for the Focsani basin. The interpretation inside the Transylvanian Basin comes from high density data in the form of interpreted seismic lines calibrated by wells published by Krézsek and Bally (2006) and Krézsek *et al.* (2010). In this area, the presented sections are slightly modified, depth-converted versions of this information. (a) Regional cross-section across Transylvanian Basin and East Carpathians. The central and external parts of the thin-skinned units are controlled by seismic lines and wells (Matenco and Bertotti, 2000; Tărăpoancă *et al.*, 2003 and references therein), while the Bucovinian nappes and the underlying crustal structure are taken from surface geology and potential field geophysical interpretations controlled by deep seismic information published by Krautner and Bindea (2002) and Stefanescu and working-group (1988). (b) Regional cross-section across Transylvanian Basin and SE Carpathians after Matenco *et al.* (2007) and Schmid *et al.* (2008). Detailed interpretation based on seismic lines is available here in the external thin-skinned nappes and the foreland (Tărăpoancă *et al.*, 2003; Leever *et al.*, 2006). (c) Regional cross-section across Transylvanian Basin and South Carpathians. The Getic Depression and the foreland units are well controlled by the interpretation of seismic lines calibrated by wells (Răbăgia and Matenco, 1999; Leever *et al.*, 2010; Răbăgia *et al.*, 2010). The interpretation of South Carpathian basement and its depth geometry is taken from lateral projection of surface studies (e.g. Iancu *et al.*, 2005 and references therein) calibrated by thermochronology (Fügenschuh and Schmid, 2005). For a similarly interpreted cross-section, but located farther west, see Schmid *et al.* (2008).

system of antiformal stacks and out-of-sequence thrusts (Fig. 2a). Apatite fission track (AFT) studies have demonstrated that overall Miocene subduction and collision induced up to 6 km of bulk exhumation in the East Carpathians between 17 and 8 Ma (Sanders *et al.*, 1999). Significant exhumation during the Late Miocene (15–10 Ma) has been recorded by combined AFT and structural studies near the sinistral transcurrent system of the Bodgan-Voda Dragos-Voda faults, located inside the Early Miocene suture zone between the internal ALCAPA and Tisza–Dacia blocks (Fig. 1; Sanders *et al.*, 1999; Tischler *et al.*, 2007; Gröger *et al.*, 2008). This is a local effect of strain partitioning along inherited crustal boundaries in the upper-plate basement and is therefore less relevant for the overall collision mechanics of the East Carpathians. Outside this restricted area, Miocene AFT ages are widely distributed from the Transylvanian Basin to the thin-skinned wedge, suggesting similar amounts of bulk exhumation across the orogen and increased exhumation rates during the late stages of plate convergence (15–11 Ma, Fig. 2a; Sanders *et al.*, 1999).

The geometry of the SE Carpathians shows a much steeper contact between the Bucovinian upper-plate basement and the Ceahlau-Severin subduction zone (Fig. 2b). Furthermore, the Moesian lower plate and the sole thrust at the base of thin-skinned units are truncated by a number of younger, high-angle thrusts. This geometry at depth is substantiated by seismic reflection/refraction studies, tomographic inversion analysis and ray-trace modelling (e.g. Bocin *et al.*, 2005, 2009; Hauser *et al.*, 2007). The SE Carpathian foreland is dominated by the large-scale syncline geometry of the Focsani Basin (Fig. 2b), visible in a large array of interpreted seismic lines calibrated by wells (e.g. Tărăpoancă *et al.*, 2003). This geometry was interpreted as an effect of Quaternary inversion of a foreland domain previously affected by increased subsidence rates throughout the Miocene to Pliocene (Bertotti *et al.*, 2003; Leever *et al.*, 2006). This subsidence was possibly driven by the Vrancea slab pull in the SE Carpathians (Matenco *et al.*, 2007). AFT studies have suggested that the overall Miocene–Quaternary deformations induced 4–5 km of bulk exhumation, distributed asymmetrically across the orogen (Fig. 2b; Sanders *et al.*, 1999). The highest amounts of exhumation are observed in the thin-skinned wedge, while the most internal (Ceahlau) nappe and the upper-plate (Bucovinian) basement record almost no exhumation. Although post-Cretaceous AFT ages are distributed in the 27–5 Ma interval, thermal modelling of AFT samples suggests an acceleration of denudation during the Pliocene (*c.* 5 Ma), when 2 km of exhumation took place (Sanders *et al.*, 1999).

The geometry of the South Carpathians is largely inherited from the patterns of thick-skinned nappes emplaced during the Cretaceous, core-complex formation during the latest Cretaceous–Eocene orogen-parallel extension (i.e. the dome shape of the Danubian nappes) and dextral transtensional faulting during the Oligocene to Early Miocene (Fig. 2c; Răbăgia and Matenco, 1999; Fügenschuh and Schmid, 2005; Răbăgia *et al.*, 2010). These deformation phases correspond to the main pulses of exhumation recorded by a high density of zircon and apatite fission-track ages, which span from late the Early Cretaceous to Palaeogene (see the review of Fügenschuh and Schmid, 2005 and references therein). Deformation associated with the Miocene collision is recorded by numerous faults in the Getic Depression, but the overall offset along the sole thrust is rather small when compared with that seen for the area of the East or SE Carpathians (~15–20 km in Fig. 2c). A small number of Miocene AFT ages have been obtained for the centre of the South Carpathians (Schmid *et al.*, 1998; Fig. 2c) and the area to the SW, near the connection with the Dinarides (Fig. 1; Bojar *et al.*, 1998). These ages have been interpreted to be the result of the final stage of thrusting and transpression recorded by the thin-skinned units (Fügenschuh and Schmid, 2005).

Sequence stratigraphy in the Miocene back-arc Transylvanian Basin

Miocene subsidence in the Transylvanian Basin led to the deposition of up to 4 km of sediments between *c.* 16 and 9 Ma, and was post-dated by dominantly continental conditions (Figs 3 and 4). The Miocene sediments are not associated with large-scale normal faulting and massive lithosphere attenuation, as is the case in the Pannonian Basin (Fig. 2; Tari *et al.*, 1992; Dérerová *et al.*, 2006). The only significant structures at regional scale are shallow salt-decollements, which are accentuated near the East Carpathians as a result of loading and thermal sagging induced by the emplacement of overlying volcanic edifices (Fig. 2a; Szakacs and Krézsek, 2006).

The evolution of the Miocene depositional system in the Transylvanian Basin, situated near the emerging Carpathians during the Middle and Late Miocene, was analysed by Krézsek *et al.* (2010), in a sedimentological study of outcrops correlated with wells and reflection seismic data (Fig. 3a). These data were calibrated by detailed biostratigraphic analysis and were used as a primary input for a novel sequence stratigraphic interpretation (Fig. 4a).

Miocene pulses of orogenic shortening created tectonic uplift episodes in areas of the Transylvanian Basin situated in the proximity of the East, SE and South Carpathians (Fig. 3). These uplift episodes are recorded in the basin by local base-level (or 'relative sea-level') drops, which can be detected by means of a sequence stratigraphy study. However, this type of third-order base level variation is potentially influenced not only by tectonics but also by eustatic variations, climate or local hydrological balance (e.g. Abreu and Haddad, 1998; Catuneanu, 2002). In the larger Middle–Late Miocene, Pannonian/Transylvanian basin system, the base level variations are rather well known and discussed in terms of eustasy vs. tectonics of a Paratethys realm, which evolved as an isolated (large) lake (e.g. Vakarcs *et al.*, 1994; Horváth, 1995; Horváth *et al.*, 2006; Csato *et al.*, 2007; Juhász *et al.*, 2007 and references therein).

For example, the uplift of the Carpathians towards the end of the Middle Miocene closed the gateways between the Central and Eastern Paratethys and created a large transgression in the entire intra-Carpathians area at the beginning of the Pannonian (e.g. Harzhauser *et al.*, 2003;

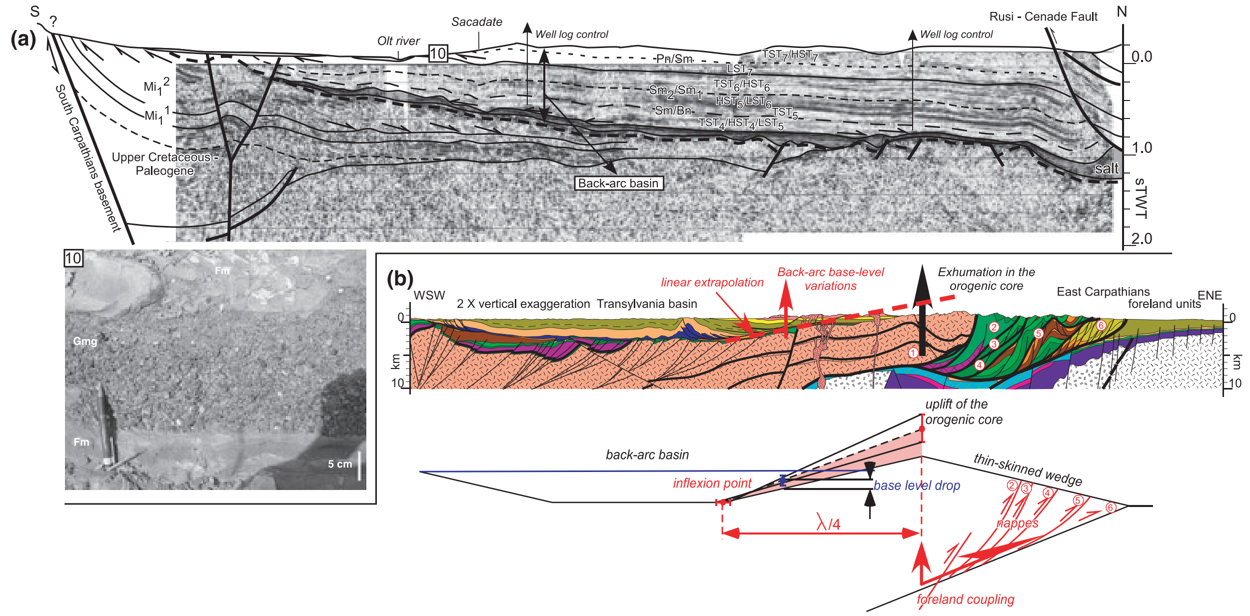


Fig. 3 (a) Example of data used to derive the sedimentological evolution of the Transylvanian Basin near the East, SE and South Carpathians. Interpreted seismic line near the South Carpathians correlated with well logs, outcrop information and detailed biostratigraphy. $Mi_1^{1,2}$, lower and upper parts of Lower Miocene; Bn, Badenian; Sm, H_Sarmatian; Pn, Pannonian (see Fig. 4a for local Paratethys stages); LST, TST, HST – low-stand, transgressive, high-stand system tracts, 10 – is the location of the outcrop photograph below. The outcrop photo shows sandy tempestites, often highly bioclastic in alternation with marls. Gmg, matrix-supported fining upwards gravels; Fm, massive fines (shales) (see Miall, 1996). The full dataset and its sedimentological interpretation are available in Krézsek *et al.* (2010). (b) Methodology of projecting the base-level variations recorded in the back-arc Transylvanian Basin towards the orogenic core. The inflexion point is the place where the measured uplift cannot be observed in the basin. It is detected by seismic interpretations through dip changes in distal sediments deposited horizontally by aggradation (see Krézsek *et al.*, 2010). Section in Fig. 2(a) is used as an illustrative to calibrate the conceptual sketch of projection in the orogenic core using a simple linear function. Note that the error bars in locating inflexion points cumulated with the ones in base-level changes, lead to rather large errors in calculating the uplift in the core of the orogen.

Kovac *et al.*, 2004; Sztanó *et al.*, 2005). It is unlikely that this transgression is related to global eustasy because it occurred in an already endemic Pannonian/Transylvanian lake (Magyar *et al.*, 1999). It is more likely that the gradual rise in the lake level was controlled by local climatic factors, in particular, runoff precipitation, where basin widening is controlled by the height of the gateways towards adjacent basins (Garcia-Castellanos, 2006), in this case, in the Carpathian foreland. Such an overspill over a gateway, for example, recorded during the Upper Pontian, waters from the Pannonian Basin invaded across the South Carpathians during the Messinian Salinity Crisis (Figs 1 and 4a; Leever *et al.*, 2010). This example demonstrates that the balance among eustasy, climate, local hydrological balance and tectonics is complex, despite the advantage of a closed sedimentary system. Therefore, our interpretation is based on the assumption that tectonics is the only parameter that focuses base level variations in one restricted part of a large basin, which do not have any coeval and comparable regional equivalents.

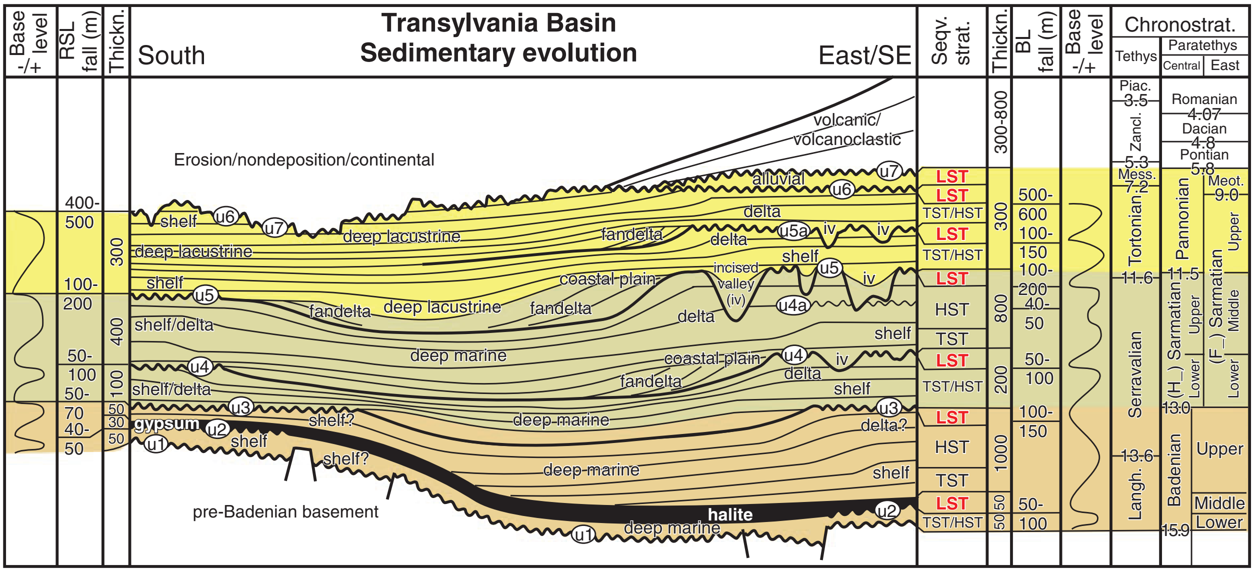
The balance between shortening and orogenic uplift depends on a variety of parameters, such as the backstop, subduction angles, plate rheology, erosion and subduction velocities (e.g. Hoth *et al.*, 2007). In a back-arc basin, base-level drops record absolute uplift values during moments of orogenic shortening, which can be used to derive the uplift in the core of the orogen once an extrapolation function is defined (Figs 3b and 4b). A simple linear function was used to project the local base-level variations measured near the basin margins in the core of the orogen, starting from inflexion points observed in the distal sediments of the Transylvania Basin (Fig. 3b).

Orogenic pulses recorded by the sediments of the back-arc Transylvanian Basin

The tectonic uplift pulses detected by sequence stratigraphy are generally coeval with individual nappe shortening events defined by structural mapping in the Carpathian foreland (Fig. 2). Two types of uplift are calculated as a function of their effect in the Transylvanian Basin: short wavelengths, only influencing proximal areas (observed $\lambda/4$ = 50–90 km) and large wavelengths, influencing the entire basin (observed $\lambda/4$ = 100–150 km, Fig. 4b).

The Miocene back-arc evolution started with a Lower–Middle Badenian transgressive–regressive cycle, which took place at reduced sedimentation

(a)



(b)

Time (see Fig. 3)	Recorded base level variations (metres)			Uplift (kilometres)		
	East	SE	South	East	SE	South
Upper Badenian (u3)	125 ± 25	125 ± 25	60 ± 10	1.1 ± 0.5	0.9 ± 0.4	0.8 ± 0.4
Middle H_Sarmatian (u4a)	45 ± 5	45 ± 5	-	0.4 ± 0.2	0.3 ± 0.1	-
Upper H_Sarmatian (u5)	150 ± 50	150 ± 50	150 ± 50	1.2 ± 0.6	0.9 ± 0.4	1.2 ± 0.6
Lower Pannonian (u5a)	125 ± 25	125 ± 25	-	0.8 ± 0.4	0.8 ± 0.4	-
Uppermost Pannonian (u6)	550 ± 50	550 ± 50	450 ± 50	1.5 ± 0.3	1.1 ± 0.3	1.5 ± 0.4
*Top Pliocene	-	-	-	-	2 0 ± 1.0	-
* Matenco *et al.* (2007)		***Cumulative total***		5 0 ± 2.0	6.0 ± 2.6	3.5 ± 1.4

Fig. 4 (a) Sequence stratigraphic summary of the Middle–Upper Miocene evolution of the Transylvanian Basin in the vicinity of the East, SE and South Carpathians (Fig. 1). The sequence stratigraphy is mainly based on transgressive surfaces and maximum flooding zones (e.g. Galloway, 1989; Sharland *et al.*, 2001). Note the endemic biostratigraphy of the Central and Eastern Paratethys, used in the Transylvania Basin and Carpathians foreland, is the result of the Miocene uplift of the Carpathians (see Rögl, 1996). To separate the same biostratigraphic stage with different durations, the Sarmatian has the prefix H_ and F_ for the Transylvania Basin and Carpathians foreland respectively. BL – base level; LST, TST, HST – low-stand, transgressive and high-stand system tracts. (b) Recorded base level variations vs. their extrapolation in the area recording maximum uplift, following the methodology of Fig. 3(b).

rates associated with normal faults with small offsets (Figs 3a and 4a, between u1 and u2). The amount of local stretching is negligible and is interpreted as a far field effect of the much larger coeval extension in the Pannonian basin (Krézsek *et al.*, 2010). In the Transylvanian Basin, a regional, up to 100 m, Middle Badenian base-level fall (u2, Fig. 4a) is correlated with a climatic event described as the Middle Miocene salinity crisis, which restricted marine connections and triggered deposition of thick salt and gypsum (Peryt, 2006). The Upper Badenian was deposited during a complete third-order transgressive–regressive cycle. Deep-marine sediments gradually onlapped onto all the basin margins starting from the areas near the East and SE Carpathians. The base-level fall towards the end of the Badenian exposed earlier shelf areas, while fan delta systems, dominated by coarse-grained sediments, supplied short submarine fans (u3, Fig. 4a). This base-level fall was local and asymmetric, reaching 100–150 m in the E/SE and 50–70 m in the south, and resulted from tectonic uplift in the Carpathians. This interpretation can be correlated with the Middle Miocene nappe stacking observed in the Carpathian foreland (e.g. Matenco and Bertotti, 2000).

During H_Sarmatian–Pannonian times (Fig. 4a), pulses of tectonic uplifts with short wavelengths are recorded by coarse-grained fan delta systems, with depositional rates of ~1 mm yr^{-1} in the basin centre. One local and two regional sequence boundaries can be observed during the H_Sarmatian. The first regional base-level drop of 50–100 m (u4,

Fig. 4a) was coeval with a global sea-level fall and was recorded regionally, while the second one, of up to 200 m (u5, Fig. 4a), was more localized and coincided with the last tectonic movement recorded at the exterior of Carpathian orogen along the frontal sole thrust near 11 Ma. The coeval exhumation of the Eastern and SE Carpathians led to increased depositional energy with widely distributed, thick deltaic and fluvial conglomerates, and shelf deposits near the South Carpathians. This exhumation event post-dates a base-level drop of 40–50 m, defined only near the East and SE Carpathians (u4a, Fig. 4a), which is the result of local tectonic uplift.

Sedimentary deposits of Pannonian age are made up of distal sands and marls. Proximal facies are observed near the East and SE Carpathians. An initial base-level rise was followed by deposition of sediments of shallow-lacustrine to fluvial facies, which were subsequently eroded during basin exhumation. These sediments were preserved only in places where they were covered by volcanoclastic successions (Fig. 4a). Near the East Carpathians, a 100–150 m base-level fall is added to the overall transgressive–regressive Pannonian pattern (u5a, Fig. 4a), recording another localized tectonic uplift. The overall progradation is interrupted by an erosional surface, locally overlain by fluvial gravels (Fig. 4a). This erosional surface records the moment when the entire Transylvanian Basin was exhumed, which led to more than 400 m erosion (u6, Fig. 4a). Deep lacustrine Pannonian deposits are only recorded in the centre of the basin. Although regression took place near the basin margins, a regressive stage of complete basin fill is not observed; the infilling process was interrupted by a regional tectonic event, which completely exhumed the basin.

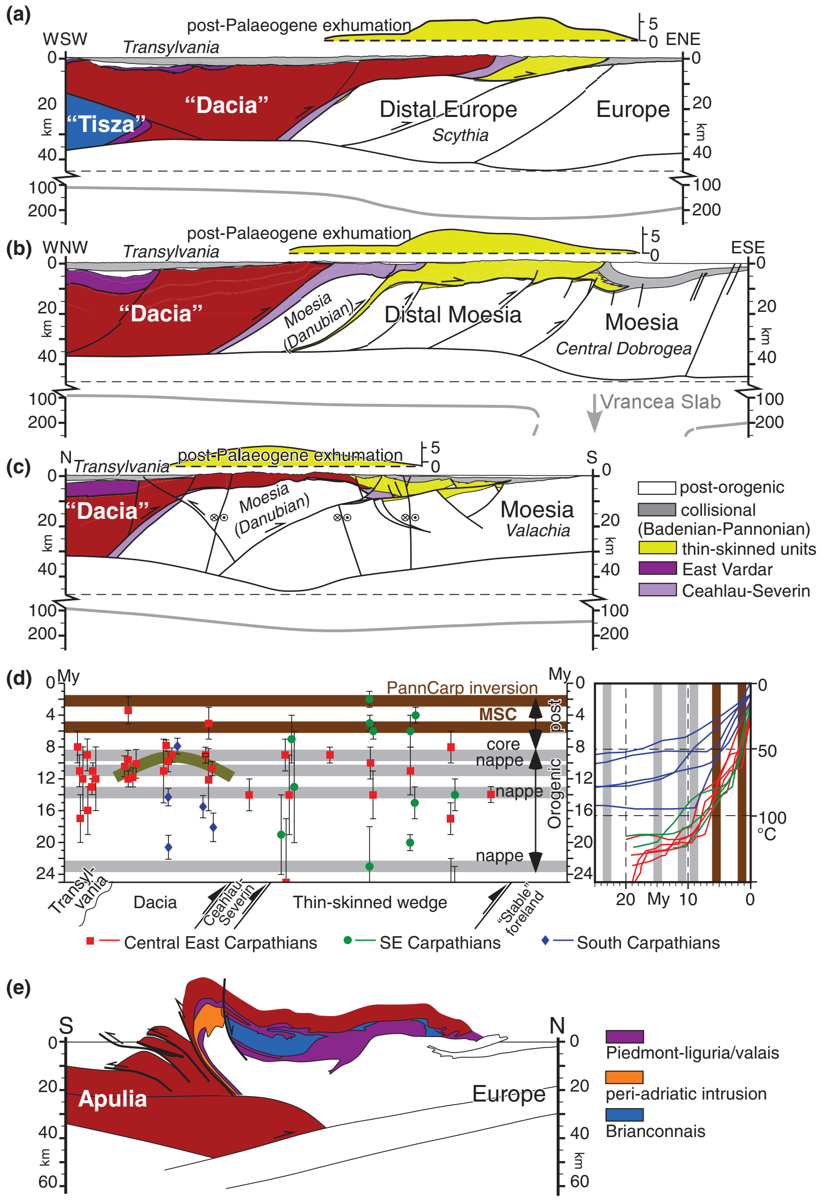


Fig. 5 (a–c) Simplified crustal-scale versions of the cross-sections displayed in Fig. 2. Grey lines in (a–c) are the corresponding base of lithosphere (after Dérerová *et al.*, 2006 and Horváth *et al.*, 2006). Note that crustal sections have the same horizontal and vertical scales, while the sub-crustal lithosphere is vertically exaggerated. For further lithospheric details and the geometry of the Vrancea slab, see Martin *et al.* (2006) and Matenco *et al.* (2007). (d) Synthetic Miocene AFT data across the Romanian Carpathians (left) and thermal modelling (right). Exhumation data in (a–d) are from Sanders (1998), Fügenschuh and Schmid (2005) and Gröger *et al.* (2008). Strain partitioning near major strike-slips was excluded (e.g. BVDF, Fig. 1). Note that only Miocene exhumation ages are plotted. Grey lines in (d) are deformation events derived from post-tectonic covers (Schmid *et al.*, 2008) and exhumation of the Transylvanian Basin at 9 Ma; brown lines are the Messinian Salinity Crisis (MSC) and Quaternary inversion (e.g. Matenco *et al.*, 2007). (e) Simplified cross-section in the Central Alps along profile NFP20 East (Schmid *et al.*, 1996), illustrative for the retro-shear collision (Fig. 6b).

Integration of structure, kinematics, exhumation and base-level variations

The complete exhumation of the Transylvanian Basin is coeval with a clustering of 9 Ma AFT ages in the orogen (Fig. 5). Error bars and tectonic events with high frequency (0.5–2 Ma) previously precluded more detailed interpretations (Fig. 5d). However, several patterns are still detected by thermochronometers when comparing the upper plate (Dacia basement and its Mesozoic cover) with the sedimentary wedge (Ceahlau nappe and more eastward thin-skinned units) located at the contact with the lower plate (Europe/Moesia, Fig. 5). The exhumation patterns are amplified by observed base-level variations in the Transylvanian Basin interpreted as tectonic-driven.

The major exhumation of the East Carpathians upper plate (Fig. 5a,d) is restricted to the 13–8 Ma time interval. The centre of the outcropping basement indicates slightly younger exhumation ages (10–8 Ma) when compared with its flanks (Fig. 5d). The overall antiformal shape of exhumation can be correlated with a ramp anticline in the lower plate beneath the orogen (distal Europe in Fig. 5a). After the major, latest Cretaceous–Palaeogene exhumation episode, the South Carpathians upper plate records Miocene uplift in the centre of the exposed basement (Fig. 5c). One 8 Ma AFT age was measured in its centre, while the flanks were partly exhumed during the Middle Miocene shortening, with samples having long residences in the partial annealing zone (Fig. 5c,d). Miocene thrusting took place beneath the Danubian nappes, i.e. far towards the foreland, indicating that the Ceahlau-Severin subduction zone was already abandoned (Figs 2c and 5c; Iancu *et al.*, 2005). In contrast with the East and South Carpathians, the Dacia basement and its Mesozoic cover in the SE Carpathians were almost not exhumed (i.e. by amounts below AFT resolution) since the accretion of the oceanic Ceahlău-Severin nappe during Cretaceous times. This cannot characterize an upper-plate indentor during collision and indicates that continental units underlying the Ceahlău-Severin nappe (i.e. the same Danubian nappes) are above the Miocene subduction zone of the Carpathian embayment (Fig. 5b,d; Schmid *et al.*, 2008).

The exhumation ages in the thin-skinned nappes of the East and SE Carpathians are generally Early–Late Miocene, coinciding with the moment of overthrusting over the thick continental European margin (Figs 2a and 5a,d). Older Early–Middle Miocene ages are associated with individual nappe emplacements in the East and SE Carpathians, which are difficult to distinguish as individual events using thermochronology alone (Fig. 5d). In the East Carpathians, the high-angle thrust in the lower-plate basement ramps up into the thin-skinned sole thrust, which is responsible for exhumation dated to 10–8 Ma in the external nappes (Fig. 5a).

The back-arc Transylvanian Basin recorded Miocene tectonic events at much higher resolution than thermochronology through base-level variations. Uplift pulses with short wavelengths are observed at *c.* 13, 12.5, 12 and 11 Ma, the second and latter only in the East and SE Carpathians. A simple lateral projection into the core of the orogen (Fig. 4b) indicates individual pulses of exhumation in the order of 0.4–1.2 km, cumulating 3.5, 2.9 and 2.0 km in the East, SE and South Carpathians respectively. Exhumation of the Transylvania Basin at *c.* 9 Ma yields calculated uplifts in the order of 1.5, 1.1 and 1.5 km in the same areas (Fig. 4b). However, this latter event has a much larger exhumation wavelength; a 200–300 km wide area was uplifted, distributing young exhumation ages across the entire Carpathians (Figs 5 a,d and 6a).

After 9 Ma, thermochronology results and deformation structures in the SE Carpathians indicate exhumation clusters around two age groups: at 5–6 Ma and at 2–3 Ma. The first one is coeval with the Messinian Salinity Crisis and the desiccation of the Carpathian foreland basin resulting from the coeval Black Sea sea-level drop (Gillet *et al.*, 2007). As syn-tectonic patterns are absent in the nappes and foredeep sediments, this might be interpreted as climate-driven exhumation similar to the coeval event defined in the Alps (Willett *et al.*, 2006). The second exhumation group is the result of the SE Carpathian inversion during the Quaternary, when high-angle reverse faults duplicated the lower plate and truncated the nappe pile with fold-like geometries near the surface, with the overall exhumation estimated at ~2 km (Figs 2b and 5b,d; Matenco *et al.*, 2007). Hence, the Transylvanian Basin records only continental sedimentation during these time intervals. Deriving tectonic signals with continental sediments is a daunting task (Catuneanu *et al.*, 2009) and is much less reliable than deriving this type of signals from lacustrine or marine deposits.

Given their error bars (Figs 3b and 4b), the uplift estimates derived from tectonic-driven base-level variations and extrapolated in the core of the orogen are surprisingly comparable with the bulk exhumation calculated using thermochronological data. The AFT estimates in the order of 5.5, 5.5 and 5 km for the East, SE and South Carpathians, respectively (Sanders, 1998), are similar to base-level estimates, although the value in the South Carpathians is higher.

Foreland-coupling collision

The two major units, which compose the Carpathians lower plate, Europe and Moesia (Fig. 1), have a distinct mechanical contrast (i.e. strong vs. weak in terms of lithospheric scale rheology), which strongly influences exhumation during collision (Cloetingh *et al.*, 2004). In the European domain, the Tisza–Dacia basement was stacked and exhumed together with more external thrust-sheets during Miocene emplacement events (Fig. 5a). The European lower plate was deformed, but thrusting did not change the location of the major subduction zone. This situation is different in the South and SE Carpathians, where the Danubian part of Moesia (in the lower plate) was accreted to the upper plate at the end of the Cretaceous. The SE Carpathians recorded further thrusting and duplications in the lower plate during Miocene–Quaternary (Fig. 6a). This process of deforming plate boundaries by gradual accretion of material from the lower plate during subduction and/or collision is defined here as 'foreland-coupling' (Fig. 6b), by analogy with the concept of collisional coupling defined by Ziegler *et al.* (1995).

The Carpathian example demonstrates that the lower plate is not always a 'conveyer belt' that transfers material to the upper plate during subduction and/or collision. This might be the case for orogens with high convergence rates, such as the Alps (Schmid *et al.*, 1996), which show few tens of kilometres of material exhumed in the collision zone along retro-shears (Fig. 5e). This is also relevant for Carpathian-type orogens, which are dominated by subduction-related processes such as

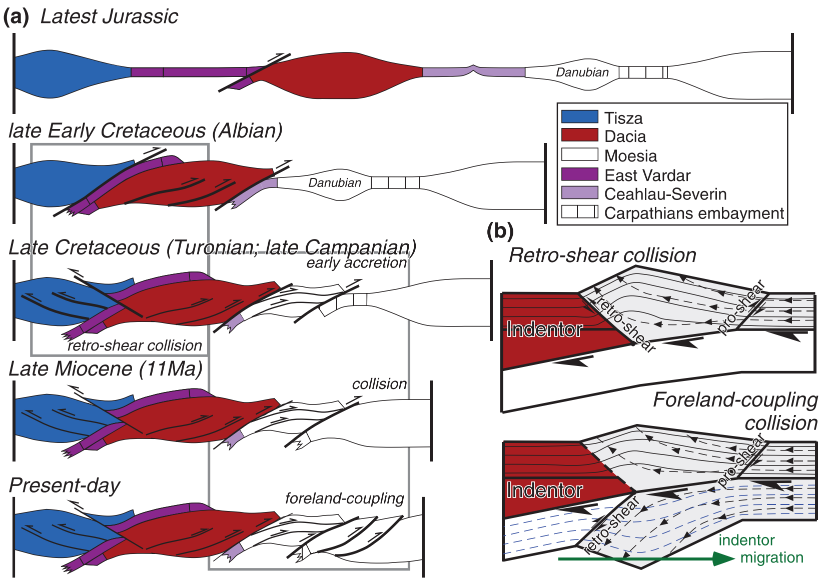


Fig. 6 (a) Sketch of plate-tectonic reconstruction along a transect crossing the SE Carpathians (modified after Schmid *et al.*, 2008). (b) Comparison between steady-state orogenic wedges starting from the mechanism of Willett and Brandon (2002) with particle paths and closure isotherms in the upper panel, as a generic mechanism applicable for various high-convergence orogens. The equivalent foreland-coupling collision in the lower panel is a conceptual cartoon, i.e. not constrained by modelling. Blue dashed lines in the lower plate are strain lines.

slab roll-back, during separate episodes of gradual nappe accretion. Deformation can thicken the lower plate along reverse faults inclined at a higher angle than the sole thrust, particularly when the subduction reaches thicker continental parts of this lower plate (Fig. 6a). These faults can be defined as retro-shears because the material entering the collision zone is moving towards the hinterland, but structurally these are foreland-vergent reverse faults. In terms of exhumation, retro-shear collision will nest zones of reset ages against the retro-deformation front at exhumational steady state (Fig. 5a; Willett and Brandon, 2002). The foreland-coupling collision will distribute exhumation ages across the orogen because of a gradual foreland-ward shift of accretion in the lower plate (Fig. 6b). The latter is a deeper-seated mechanism and, therefore, the timing of activation can be detected by larger wavelengths near the surface and by exhumation ages shifted towards the foreland. The orogens produced by foreland-coupling collision have a characteristic first-order feature: crustal and lithospheric roots are not located beneath the core of the orogen, but shifted towards the foreland (Fig. 5a–c).

The Romanian Carpathians scenario of subduction/collision is certainly not singular among Mediterranean-type orogens. Few examples may include the buried 'thrust-ridges' beneath the thin-skinned nappe pile of the West Carpathians (e.g. Siliesian, Southern Magura, Poprawa and Malata, 2006), which are foreland-coupling events in the lower plate (Palaeozoic Europe) with respect to the Pieniny Klippen Belt and Magura subduction zones (see also Oszczypko, 2006). Cretaceous–Eocene deformations in the Dinarides (pre-dating the Miocene extension of the Pannonian basin) are almost exclusively concentrated in the lower plate, with respect to the Alpine Tethys (Sava) subduction zone. This is indicated by large-scale thrusting and exhumation recorded in the External Dinarides, East Bosnian–Durmitor, Drina–Ivanjica and Jadar–Kopaonik thick-skinned thrust sheets (e.g. Schmid *et al.*, 2008 and references therein). Blind thrust faults with ramp-flat geometries of Quaternary age seem to be rooted at mid-crustal levels in the Adriatic lower plate beneath the Apennines. Here, deformation is observed near the surface through large areas of uplift with an antiformal shape (Picotti and Pazzaglia, 2008). Furthermore, basement thrusting, post-dating nappe emplacement, accounts for strain partitioning in the lower plate in many other orogens (Roure, 2008). For example, the typical retro-shear collision in the European Alps also shows potential foreland-coupling structures, such as a localization of Pliocene deformation in the Palaeozoic basins of Jura Mountains beneath the intra-Triassic thin-skinned decollement (Roure *et al.*, 1994).

The foreland-coupling model of lower-plate thickening is also supported by a simple observation that is confirmed by modelling studies; the orgenic evolution is dependent on horizontal shortening, denudation and, last but not least, the evolution of crustal roots (e.g. Avouac and Burov, 1996). Mechanically, it is easier to build 'mountains' in the mantle than at the topographic surface.

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