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Partial delamination of continental mantle lithosphere, uplift-related crust–mantle decoupling, volcanism and basin formation: a new model for the Pliocene–Quaternary evolution of the southern East-Carpathians, Romania

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Abstract

A geodynamic model is proposed for the Mid-Miocene to Quaternary evolution of the southern East-Carpathians in order to explain the relationships between shallow and deep geological phenomena that occurred synchronously during late-collision tectonics.

In this area, an active volcanic zone cross-cuts since 2 My the suture between the overriding Tisza–Dacia and subducting European continental plates. Mafic calc-alkaline and alkaline magmas (south Harghita and Persani volcanoes) erupted contemporaneously. These magmas were supplied by partial melting of the mantle lithosphere of the subducting, and not of the overriding, plate. In an effort to decipher this geodynamically a-typical setting of magma generation, the spatial and temporal distribution of shallow and deep phenomena was successively examined in order to establish the degree of their interdependence. Our model indicates that intra-mantle delamination of the subducting European plate is the principal cause of a succession of events. It caused upwelling of the hot asthenosphere below a thinned continental lithosphere of the Carpathians, inducing the uplift of the lithosphere and its internal decoupling at the Moho level by isostatic and mostly thermal effects. During this uplift, the crust deformed flexurally whilst the mantle deformed in a ductile way. This triggered decompressional partial melting of the uppermost mantle lithosphere. Flexural deformation of the crust induced its fracturing, allowing for the rapid ascent of magmas to the surface, as well as reactivation of an older detachment horizon at the base of the Carpathian nappe stack above which the Brasov, Ciuc and Gheorghieni hinterland basins formed by extension and gravity spreading. The rapid subsidence of the Focsani foreland basin is controlled by the load exerted on the lithosphere by the delaminated mantle slab that is still attached to it. In this model, crust–mantle decoupling, magma genesis and volcanism, local near-surface hinterland extension are consequences of uplift, induced by asthenospheric upwelling triggered by intra-mantle delamination.

This model enables to conceive that delamination-induced decoupling began to be efficient 9.4 My ago at the northern end of the East-Carpathians, when the mantle slab dipped westwards. Since then, intra-mantle delamination migrated laterally, normal to the slab strike, and follows the arcuate shape (NW → SE → SW) of the Carpathians. Nowadays, whereas the mantle slab is still actively foundering below the Vrancea seismic zone to the SE of the most recent volcanic area (South Harghita–Persani), a significant southwestward shift of the delamination process can be discerned.

Keywords: post-orogenic; mantle delamination; uplift; alkaline/calc-alkaline volcanism; basin formation; East-Carpathians; Romania

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1. Introduction

One of the most puzzling features of the southern East-Carpathians of Romania is the simultaneous extrusion of Plio–Quaternary mafic alkaline and calc-alkaline volcanics in closely spaced areas (Fig. 1). In the Persani Mountains, a mafic alkaline series consists of alkali- and trachy-basalts (Peltz et al., 1971; Peltz and Bratosin, 1986; Downes et al., 1995), yielding K–Ar ages in the 2.25–0.35 My range (Casta, 1980; Mihaila and Kreuzter, 1981). On the other hand, a calc-alkaline association occurs in the South Harghita Mts, 40 km to the northeast of the Persani Mts. This suite includes shoshonitic and high-K andesitic basalts, high-K andesites and dacites (Peccerillo and Taylor, 1976; Seghedi et al., 1986, 1987; Mason et al., 1996) that yielded K–Ar ages in the 2.1–0.2 My range (Pécskay et al., 1995a; Szakacs and Seghedi, 1995). These volcanics are the most recent products of an eruptive cycle that began 9.4 My earlier and lead to the development of the Calimani–Harghita calc-alkaline volcanic chain.

Whereas the Plio–Quaternary alkaline rocks were commonly interpreted as a "within-plate" type association derived by partial melting of the asthenosphere during a phase of lithospheric extension (Downes et al., 1995), the calc-alkaline volcanics were interpreted as being related to the subduction of the European plate beneath the intra-Carpathian Tisza–Dacia microplate. However, volcanic activity in the southern East-Carpathians commenced only after large-scale crustal shortening had ceased in the outer Carpathians, around 10 My, and well after oceanic realms had been closed during late Oligocene–Early Miocene times (Ellouz and Roca, 1994; Linzer et al., 1998). To explain this time-lag between the end of major crustal shortening and lithospheric subduction and the onset of calc-alkaline volcanism, earlier geodynamic models invoke detachment and roll-back of the subducted oceanic lithospheric slab (Csontos, 1995; Mason et al., 1998; Seghedi et al., 1998; Linzer et al., 1998), or delamination of the lower part of the lithospheric mantle from the lower plate (Gîrbacea, 1997; Gîrbacea and Frisch, 1998). As already stressed by Csontos (1995), these models fail to explain the space and time distribution of alkaline *and* calc-alkaline volcanism.

These models, apart from the one proposed by Gîrbacea and Frisch (1998), do not integrate the following three major constraints. (1) Regionally, the southern East-Carpathians were affected by compressional stresses during Pliocene–Quaternary times. This is indicated by reverse faults and folds (Bandrabur et al., 1971) and by palaeostress analysis (Linzer et al., 1998). On the other hand, at a smaller scale, the internal parts of the southern East-Carpathians underwent local extension, as indicated by the subsidence of small Pliocene–Quaternary grabens (Gîrbacea, 1997; Gîrbacea et al., 1998; Linzer et al., 1998). (2) The strong seismicity of the Vrancea area, located 60 km to the southeast of the South Harghita volcanoes, is related to active "subduction" of high density mantle slab, defined by frequent earthquakes ranging in focal depths between 70 and 200 km. This nearly vertical slab is attached to the European foreland lithosphere, some 130 km to the southeast of its collisional suture with the overriding Tisza–Dacia unit (Fan et al., 1998; Wenzel et al., 1998, 1999). (3) The Pliocene–Quaternary volcanic zone crosscuts the suture between the overriding Tisza–Dacia and the subducting Europe plates that is visualised on the seismic profiles (Fig. 2; Radulescu et al., 1976; Mocanu and Radulescu, 1994; Artyushkov et al., 1996).

Therefore, mafic magmas with alkaline and calc-alkaline affinities necessarily originate from the mantle below the lower and not the upper plate, i.e. a mantle that has never had any relationship with the assumed subducted oceanic crust. A fourth and new significant constraint concerns the fact that alkaline and calc-alkaline mafic magmas are likely products of partial melting of the upper part of the mantle lithosphere of the subducting plate. This has been documented by a petrogenetic study of mantle xenoliths coming from the mantle lithosphere of the lower plate just below the most recent volcanoes (Chalot-Prat and Boullier, 1997). In view of this new constraint, the model of Gîrbacea and Frisch (1998) needs reappraisal because it is not necessary to assume contamination of partial melts derived from the asthenosphere to explain the development of calc-alkaline magmas during the delamination of the lower mantle lithosphere. The revised model integrates the physical processes that affected the mantle lithosphere as well as the crust during mantle delamination, as proposed by Gîrbacea and Frisch (1998), but not the chemical processes and the related partial melting effects. In our

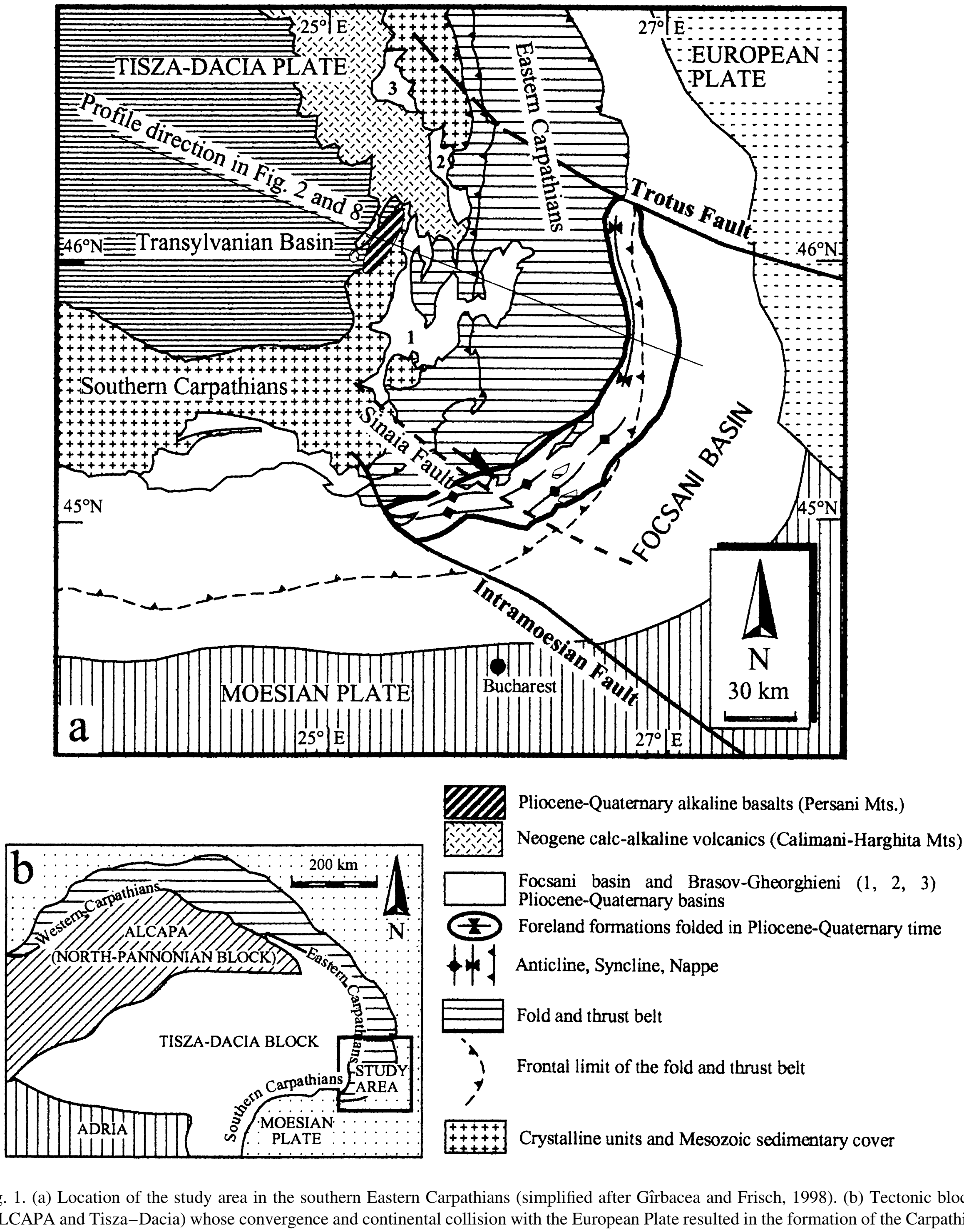


Fig. 1. (a) Location of the study area in the southern Eastern Carpathians (simplified after Gîrbacea and Frisch, 1998). (b) Tectonic blocks (ALCAPA and Tisza–Dacia) whose convergence and continental collision with the European Plate resulted in the formation of the Carpathian arc during Tertiary times (after Csontos, 1995).

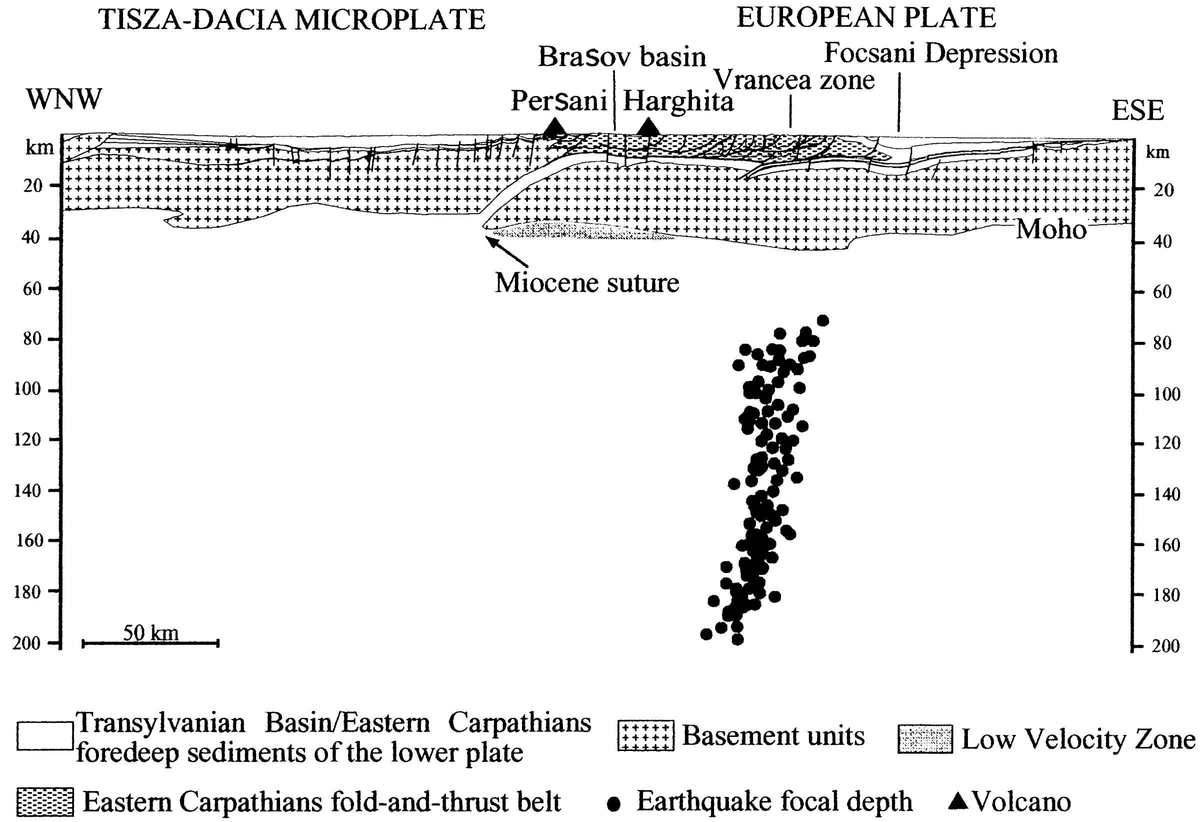


Fig. 2. Crustal profile through the Eastern Carpathians, showing the position of the studied Pliocene–Quaternary basins, volcanoes, and Vrancea seismic zone (simplified after Gîrbacea and Frisch, 1998).

study, we postulate that there is a genetic link between deep and shallow events and that their timing is essential. Major problems that had to be solved were not only the partial melting conditions of the subcrustal mantle, but also the rapid ascent of magmas throughout the crust to the surface. Indeed, some of calc-alkaline and alkaline mafic volcanics display primitive mantle features, suggesting that magma generation had been accompanied by the opening of deep crustal fractures. Our model explains how and why penetration of a subduction slab into the mantle, disturbing asthenospheric convection, is able to induce magma genesis, triggering crust–mantle decoupling, and that ascent of magma to the surface. The latter depends on the ability of the entire crust to behave in a brittle way, which appears to be often possible for only very short periods. Inasmuch as this model allows to decipher the geodynamic evolution of the southeastern Carpathians during the last 17 My, it may contribute to forecast the future behaviour of this seismically and volcanically active area.

In the following we review all data available on surface and deep events in an effort to explore their "cause and effect" relationship and propose a new interpretation for crust–mantle relationships in a late-collisional setting.

2. Geological framework of the southern East-Carpathians

The regional tectonic framework of the southern East-Carpathians is well documented (Csontos et al., 1992; Roure et al., 1993; Csontos, 1995; Artyushkov et al., 1996; Gîrbacea, 1997; Matenco, 1997; Gîrbacea and Frisch, 1998; Gîrbacea et al., 1998; Nemcock et

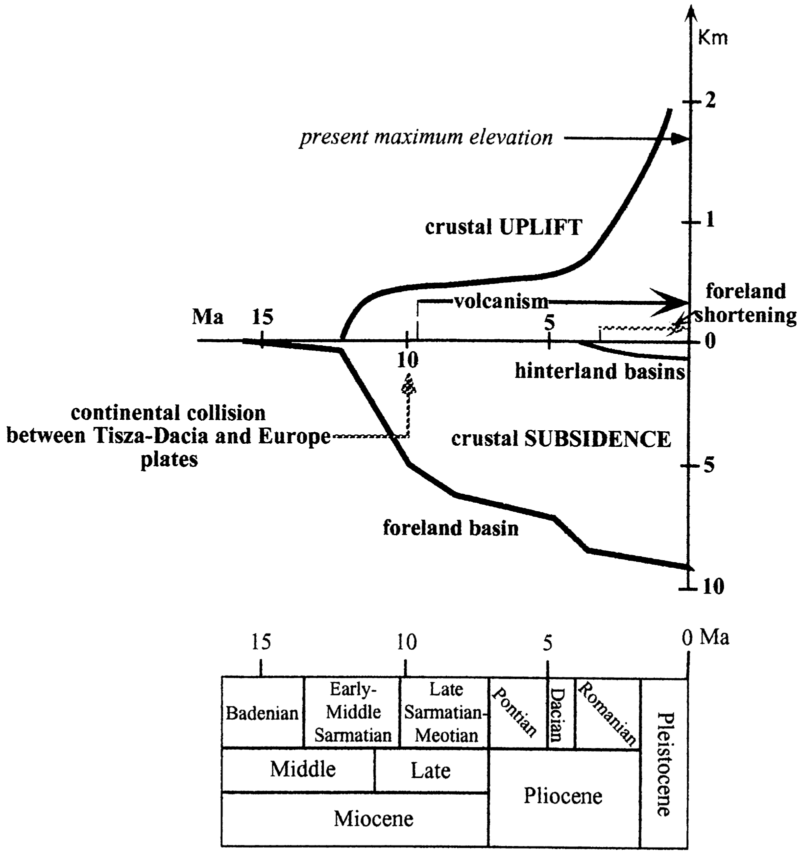


Fig. 3. Timing of successive and coeval surface events in the southern Eastern Carpathians. The different curves represent the subsidence of Focsani foreland, the uplift and the hinterland subsidence leading to the Brasov, Ciuc and Gheorghieni basin formation, while straight lines represent the age intervals of the Late Miocene–Quaternary volcanic activity and of shortening of the internal part of the foreland.

al., 1998; Matenco et al., submit.). The East-Carpathians consist of imbricated thrust sheets of different ages, forming an internal and an external part. The internal East-Carpathians consist of thick-skinned nappes, involving Late Proterozoic (Sandulescu, 1975) and possibly Variscan basement units (Dallmeyer et al., 1998) that, together with their Mesozoic sedimentary cover, were deformed and incorporated into the orogenic wedge during Cretaceous times (Sandulescu, 1975). The external, Outer East-Carpathians, consists of thin-skinned Cretaceous–Miocene turbidite nappes arranged in a southeast-verging fold-and-thrust belt that underwent maximum deformation during Cenozoic times (Royden and Baldi, 1988). The present tectonic position of the East-Carpathians is the result of the Tertiary collision of the Tisza–Dacia microcontinent (upper plate) with the European Plate (lower plate). It is believed that this collision was preceded by roll-back of the subduction zone (Royden, 1988, 1993) and closure of a basin floored by a subductable oceanic lithosphere (Csontos, 1995; Linzer, 1996). Plate convergence stopped essentially during the Sarmatian, around 10 My (Fig. 3; Roure et al., 1993). Refraction seismic profiles throughout the Intra-Carpathian area (Fig. 2; Radulescu et al., 1976; Mocanu and Radulescu, 1994) image the position of the suture zone between the Tisza–Dacia and European plates in the Harghita-Persani area, the location of the youngest volcanic activity. Elsewhere, the Internal Carpathian domain is characterised by extensive calc-alkaline acidic, and alkaline and

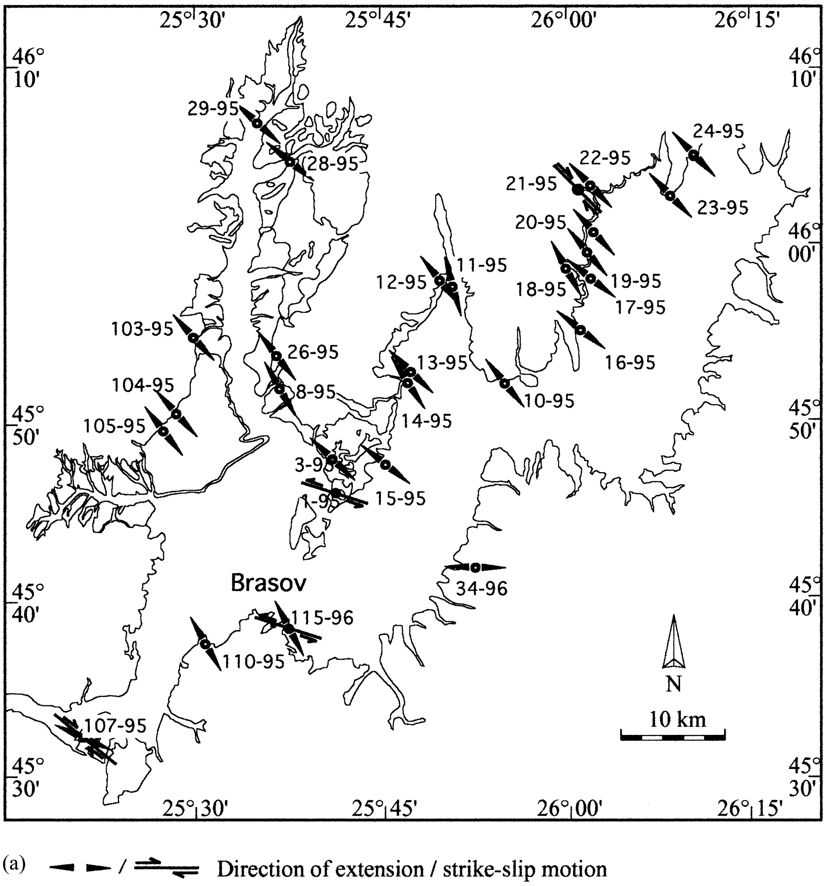
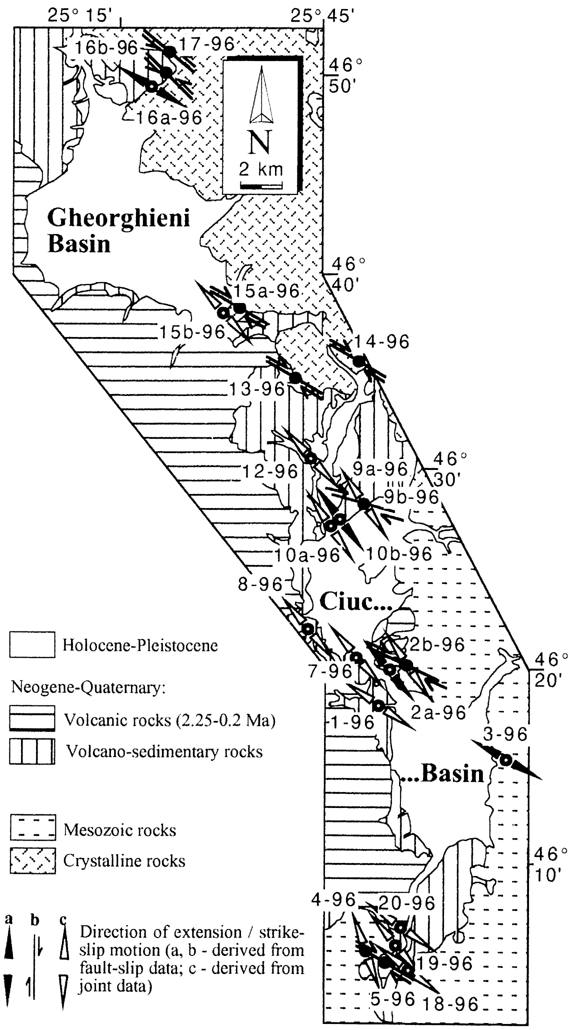


Fig. 4. Map of the Brasov Basin (a) and Ciuc–Gheorghieni basins (b) indicating the Pliocene–Quaternary kinematic directions. Outcrop numbers concern the measuring sites listed in Table 1.

calc-alkaline mafic volcanism and the emplacement of intrusive bodies during the time span of 20 My (early Miocene) and 0.15 My (Pécskay et al., 1995b). Most of these volcanics are scattered throughout the inner East-Carpathians, whereas a few were emplaced at the northern and southern extremities of the outer East-Carpathians (Csontos, 1995). The space and time distribution of mafic volcanism of both affinities is rather complex. Indeed, whatever their affinity, the oldest eruptions occur rather in the centre of the Carpathian–Pannonian area and the youngest ones are located at its periphery, partly far away from the Outer Carpathian front (Pécskay et al., 1995a,b). In general, but not systematically, alkaline eruptions followed the calc-alkaline ones in rather close sites. Both types of eruptions can also occur simultaneously at same site as in Central Slovakia (between 16–14 My) and in South Harghita–Persani Mts (since 2 My; Csontos, 1995; Pécskay et al., 1995b).

The studied area covers those parts of the inner and outer East-Carpathians that are bounded to the South and North by the NW–SE-striking Trotus and Intramoesian faults, respectively (Figs. 1 and 2). It contains three different types of basins. (1) The Focsani foreland basin that is located adjacent to the Carpathian thrust front and contains up to 9000 m of Neogene–Quaternary clastics with marine lithologies until 10.5 My followed by continental lithologies, and gently folded in the westernmost parts of the basin (Motas and Tomescu, 1983; Jipa, 1983). (2) The Brasov–Gheorghieni system of extensional hinterland basins that contains up to 1000 m of Plio–Quaternary



(b)

Fig. 4. (*continued*)

continental deposits (Gîrbacea, 1997; Gîrbacea et al, 1998) and is bounded to the west and northwest by the Harghita and Persani zone of Plio–Quaternary mafic eruptive centres. (3) The Late Cretaceous–Pliocene Transylvanian basin that contains up to 4000 m of predominantly marine sediments deposited in a foreland basin associated with the Apuseny Mountains thrust system (De Broucker et al., 1998; Ciulavu, 1998 and refs. therein). The volcanic province of the Persani Mts, with alkaline affinities, and the Harghita Mts, with calc-alkaline affinities, also includes the isolated 6.8 My old calc-alkaline Rupea volcano, located 15 km northwestwards of the Persani Mts. These calc-alkaline volcanoes form the southeastern termination of the calc-alkaline Calimani–Harghita volcanic chain. The seismically very active Vrancea zone underlies the external parts of the thin-skinned Outer Carpathian (Fig. 2). Seismic refraction data indicate that the mantle lithosphere thins from about 100 km beneath the European foreland across the Vrancea Zone to about 40 km beneath the inner Carpathians and the Harghita and Persani volcanic field where a low-velocity zone is recognised directly below the Moho (Lazarescu et al., 1983; Nemcock et al., 1998).

3. Supracrustal deformation and crustal fracturing during volcanic eruptions

3.1. Timing of supracrustal deformation and volcanism

The timing of supracrustal deformations and volcanism is summarised in Fig. 3. The Focsani foreland basin began to subside around 16 My (Matenco, 1997), before the end of major crustal shortening around 10 My. Uplift of the East-Carpathians began around 12 My (Sanders et al., 1997; Sanders, 1998), was paralleled by a dramatic subsidence acceleration of the Focsani basin, and was shortly followed by the onset of the volcanic activity. Subsidence of the tensional hinterland basins began at about 4 My, and was followed at 3 My by the shortening of the western parts of the Focsani basin. The latter two phenomena were associated with a dramatic increase in uplift rates, leading to elevations of up to 2000 m. Since 2.25 My, the active volcanic area as well as the zone of maximum subsidence in the Focsani depression (Matenco et al., 1997) are shifting southwestwards.

3.2. Formation of the hinterland basin system

Since Pliocene times, the internal area of the southern East-Carpathians was affected by SE–NW-directed extension, controlling the subsidence of three graben-like sedimentary basins, namely the Brasov, Ciuc and Gheorghieni basins (Fig. 4a and b; Gîrbacea, 1997; Gîrbacea et al., 1998). These basins are superimposed on Carpathian nappes and contain up to

1000 m thick lacustrine–fluvial–volcaniclastic (Bandrabur et al., 1971; Bandrabur and Codarcea, 1974). Based on faunal assemblages (Liteanu et al., 1962; Samson and Radulescu, 1963; Radulescu et al., 1965) and magnetostratigraphic data (Ghenea et al, 1979), sedimentation in these basins began around 3.8–3.6 My. A lava flow underlying the oldest formations in the Ciuc Basin yielded a K–Ar age of 4 My (Pécskay et al., 1995a).

Extension is documented by well-developed NE–SW-oriented normal faults and vertical joint sets (Fig. 5a and b; Table 1). The estimated amount of extension is between 14 and 22%, based on outcrop-scale structures and an area-balanced kinematic model for basin formation. In addition to kinematic data indicating NW–SE extension, the individual segments of the Brasov, Ciuc, and Gheorghieni basins seem to be linked by strike–slip faults.

Basin formation was synchronous with accelerated uplift rates of the entire area, amounting to a total of 1500–2000 m since 4 My (Fig. 3). The age and amount of uplift was deduced from fission track studies (Sanders et al., 1997; Sanders, 1998) and deposition rates of coarse sediments supplied into the foreland basin from the eroding orogen (Artyushkov et al., 1996; Sanders, 1998). As a consequence of uplift and associated exhumation, the vertical stress became the maximum compressive stress in rock formations approaching the surface, whilst the horizontal stress became the minimum compressive stress, thus promoting extension (Price, 1966). Therefore, uplift of the area is thought to be the driving mechanism for the observed extension at upper crustal levels, such that basins formed by "gravity spreading", a well-known and documented process during rifting phases (Neugebauer, 1978; Houseman and England, 1986).

Coeval with hinterland extension, the sedimentary fill of the northwestern parts of the Focsani basin was folded between 3 and 1 My (Sandulescu, 1975). Taking into account seismic activity indicating strike–slip motion along the NW–SE-oriented Trotus and Sinaia faults (Gîrbacea et al., 1998), a kinematic model was proposed to explain the observed deformation patterns (Fig. 6). In this model, uplift-induced extension controls basin formation in the internal part of the East-Carpathians, as well as shortening in the foreland basin by southeastward gravity-driven motion of the Carpathian nappes between the Trotus and Sinaia faults (Fig. 6a); this involves the reactivation of a pre-existing detachment horizon at the base of the fold-and-thrust stack (Fig. 6b). Until now, this late compressional phase did not substantially affect the regional subsidence of the Focsani basin (cf. Fig. 9 in Artyushkov et al. (1996).

3.3. Synchronous development of the Persani and South Harghita volcanoes

For the *Persani basalts*, which yield K–Ar ages between 2.25 and 0.35 My (Casta, 1980; Mihaila and Kreuzter, 1981; Downes et al., 1995), palaeomagnetic data and faunal assemblages from lacustrine sediments interbedded with lava flows indicate a late Pliocene–late Pleistocene age (Peltz et al., 1971; Ghenea et al., 1979). Eruptions were initially phreatomagmatic, then pyroclastic with formation of strombolian scoria cones (Seghedi and Szakacs, 1994). The basalts consist to 20% of olivine and augite phenocrysts in a groundmass of plagioclase, olivine, augite and magnetite microlites in a glassy matrix (Peltz and Bratosin, 1986). Geochemically, these rocks are alkali basalts and trachybasalts (classification of Le Maitre, 1989), rather enriched relative to the MORB in titanium, alkalis, and also in the most incompatible elements, such as Th, U, Ta, Nb, LREE and also of Pb, Sr and Zr (Fig. 10; Peltz and Bratosin, 1986; Downes et al., 1995). Their Nd and Sr initial isotopic ratios (Fig. 9) indicate an enriched mantle source (Downes et al., 1995). This mantle source is believed to be not a deep primitive mantle (Downes et al., 1995) but the continental lithospheric mantle as indicated by xenoliths brought to the surface by these alkaline lavas (Chalot-Prat and Boullier, 1997; for details see next section).

The *South Harghita* volcanoes were active between 2.1 and 0.2 My (Pécskay et al., 1995a; Szakacs and Seghedi, 1995). Eruptions were mainly effusive (flows and domes); explosive phreatomagmatic and pyroclastic activity, with numerous pumice- and ash-flows, played a secondary role (Szakacs and Seghedi, 1995). The South Harghita volcanics are, according to their K_2O/SiO_2 ratio and homogeneous trace element patterns (Fig. 10; Mason et al., 1996; Chalot-Prat and Boullier, 1997), calc-alkaline high-K basalts, andesites and dacites, with a gradual increase

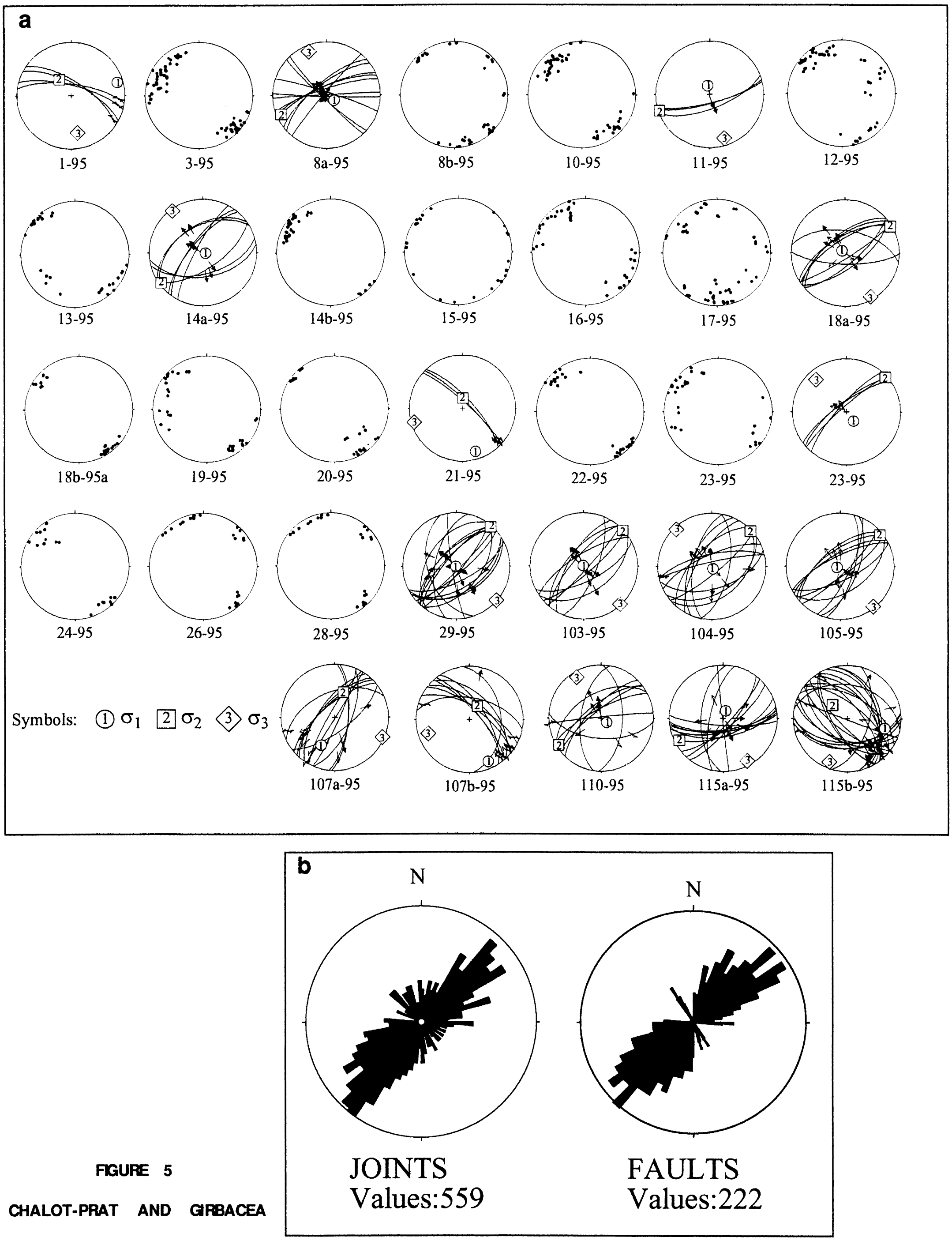


Fig. 5. (a) Joint and data sets interpreted to reflect the Pliocene–Quaternary extensional event from the Eastern Carpathians hinterland (after Gîrbacea and Frisch, 1998, modified). The fault planes are represented in equal area, lower hemisphere stereonets as great circles, with arrows showing the direction of slip of the hangingwall. The joint planes are plotted in equal area, lower hemisphere stereonets as poles. Maximum pole density is assumed to be parallel to the maximum direction of extension. The geographic position and kinematic parameters of these data sets are listed in Table 1. (b) Orientation of Pliocene-Recent vertical joints (left) and normal faults (right).

Table 1
The results of the kinematic analysis of data collected in the Braçov–Gheorghieni basins, Persani, and Harghita mountains (fault-slip and joint data). Outcrop numbers (No.) are located in Figs. 4a, b and 7 (Ages: Pz — Paleozoic; J — Jurassic; Cr — Cretaceous; Mc — Miocene; Pl — Pliocene; P — Pleistocene; H — Holocene; E — Early; L — Late)

No.	Age	Location Lat. N/Long. E	Bedding	No. of data	s_1	s_2	s_3 [a]	R [b]	F [c] (°)
1–95	Cr	45°46′38″/25°40′30″	054/12	4	074/11	325/59	170/29	0.546	10
3–95	EP	45°48′00″/25°40′27″	220/12	61			132/9		
8a–95	EP	45°52′27″/25°36′32″	124/07	11	124/76	247/08	338/02	0.307	16
8b–95	EP	45°51′03″/25°37′42″	128/10	36			316/2		
10–95	ECr	45°52′24″/25°54′31″	284/37	49			320/20		
11–95	EP	45°57′53″/25°50′46″		3	347/76	252/1	162/14	0.499	1
12–95	ECr	45°57′39″/24°48′56″	125/46	46			320/8		
13–95	EP	45°55′00″/25°47′28″		29			313/2		
14a–95	EP	45°52′30″/25°46′38″	119/05	8	100/86	234/3	324/3	0.434	12
14b–95	EP	45°52′30″/25°46′38″	119/05	32			302/10		
15–95	EP	45°46′23″/25°45′18″		23			307/5		
16–95	ECr	45°55′22″/25°59′00″	021/16	35			127/3		
17–95	EP	45°55′49″/26°00′08″		52			307/27		
18a–95	Cr	45°58′46″/26°00′52″	274/70	9	307/84	59/2	150/5	0.444	10
18b–95	Cr	45°58′46″/26°00′52″	274/70	22			140/5		
19–95	ECr	46°00′04″/26°01′10″	199/24	33			139/6		
20–95	ECr	46°01′52″/26°02′02″	215/61	24			137/3		
21–95	EP	46°03′01″/26°02′09″		3	163/15	3/74	254/5	0.501	2
22–95	ECr	46°03′01″/26°01′48″	234/55	28			315/2		
23a–95	Ec	46°03′01″/26°07′54″	291/24	4	144/77	48/2	317/18	0.504	2
23b–95	Ec	46°02′56″/26°07′54″	291/24	28			308/9		
24–95	Ec	46°04′59″/26°09′55″	283/19	18			136/2		
26–95	ECr	45°53′32″/25°36′02″	293/60	23			321/3		
28–95	EP	46°04′42″/25°36′49″	102/15	25			308/7		
29–95	EP	46°06′50″/25°34′31″	128/16	16	226/89	41/1	131/0	0.401	12
103–95	ECr	45°54′18″/25°28′32″	140/65	9	308/89	47/0	137/1	0.474	8
103o–95	ECr	45°54′18″/25°28′32″	140/65	10	129/22	284/63	34/10	0.866	13
104–95	ECr	45°51′22″/25°29′00″	230/17	12	150/86	47/1	317/4	0.510	18
104o–95	ECr	45°51′22″/25°29′00″	230/17	12	295/2	29/86	295/4	0.525	5
105–95	ECr	45°50′02″/25°27′08″	135/19	11	253/85	51/5	141/2	0.431	15
105o–95	ECr	45°50′02″/25°27′08″	135/19	8	300/6	44/67	208/20	0.261	13
107a–95	Ec	45°33′21″/25°18′10″	160/47	12	206/65	19/48	113/3	0.591	8
107b–95	Ec	45°33′21″/25°18′10″	160/47	9	156/14	31/66	251/19	0.556	9
110–95	J	45°37′40″/25°29′58″		9	127/79	238/4	239/11	0.377	10
115a–95	J	45°38′31″/25°37′00″		13	19/78	243/98	151/8	0.336	21
115b–95	J	45°38′31″/25°37′00″		23	105/28	313/59	202/12	0.358	21
1–96	Pl	46°18′02″/25°47′24″		22			123/24		
2a–96	Pl	46°19′41″/25°49′16″		11	20/84	218/5	128/2	0.486	10
2b–96	Pl	46°19′41″/25°49′16″		8	138/14	337/75	229/5	0.485	7
2c–96	Pl	46°19′41″/25°49′16″		19			144/7		
3–96	Cr	46°15′53″/25°58′39″	144/43	10	106/87	208/1	298/3	0.443	14
4–96	Cr	46°05′42″/25°04′22″	111/16	20			158/8		
5–96	Pl	46°04′58″/25°50′04″		12	304/1	232/86	70/4	0.556	10
7–96	LMc–Pl	46°20′18″/25°48′10″		21			138/3		
8–96	LMc–Pl	46°21′52″/25°42′02″		22			139/4		
9a–96	LMc–Pl	46°27′53″/25°46′57″		7	327/8	210/74	59/14	0.534	14
9b–96	LMc–Pl	46°27′53″/25°46′57″		28			325/24		
10a–96	LMc–Pl	46°27′39″/25°42′21″		14	46/83	217/7	307/1	0.453	12
10b–96	LMc–Pl	46°27′39″/25°42′21″		29			322/3		
12–96	LMc–Pl	46°30′45″/25°44′52″		21			136/4		

Table 1 (*continued*)

No.	Age	Location Lat. N/Long. E	Bedding	No. of data	s$_1$	s$_2$	s$_3$[a]	R[b]	F[c] (°)
13–96	Pz	46°36′02″/25°48′04″		5	323/3	203/84	53/5	0.455	13
14–96	Pz	47°37′30″/25°44′14″		8	136/13	331/76	227/3	0.481	9
15a–96	Pz	47°37′53″/25°37′27″		12	336/3	239/69	67/21	0.534	7
15b–96	Pz	47°37′53″/25°37′27″		35			137/12		
16a–96	Pz	46°50′22″/25°29′03″		4	163/87	40/2	310/3	0.362	16
16b–96	Pz	46°50′22″/25°29′03″		9	90/15	242/73	358/8	0.493	5
17–96	LMc–Pl	46°51′22″/25°25′54″		9	111/20	262/67	17/10	0.474	3
18–96	P–H	46°04′44″/25°50′33″		36			306/17		
19–96	Pl	46°06′30″/25°50′44″		47			319/20		
20–96	Pl	46°07′22″/25°51′13″		29			304/4		
22–96	Pl–P	46°01′51″/25°25′52″		25	245/86	48/4	138/1	0.452	12
23–96	Pl–P	46°01′45″/25°24′32″		11	215/86	59/4	329/2	0.524	6
24–96	Pl–P	46°01′14″/25°24′31″		6	359/85	220/4	130/3	0.456	10
25–95	Pl–P	45°57′22″/25°21′17″		17	23/87	221/3	131/1	0.476	10
26–96	Pl–P	45°59′26″/25°19′35″		12	123/85	232/2	322/4	0.478	6
27–96	Pl–P	45°53′30″/25°53′10″		18	9/88	226/2	136/1	0.427	11
34–96	Cr	45°42′02″/25°18′14″		11	188/40	347/48	88/11	0.623	21

[a] Data sets consisting exclusively of joints are those where only σ_3 is given, as the direction of maximum density of joint poles.
[b] R is the ratio between the stress magnitudes [$R=(\sigma_2-\sigma_3)/(\sigma_1-\sigma_3)$]. R defines the regime of deformation, as: plane strain, with $R=0.5$ (deformation occurs only parallel to σ_1 and σ_3, and $\sigma_1=\sigma_3$, $\sigma_2=0$); axial extension (constriction), with $R=1$ (shortening occurs parallel to both σ_1 and σ_2); axial shortening (flattening), with $R=0$ (extension occurs parallel to both σ_2 and σ_3).
[c] F indicates the average value of the differences between the measured striae on fault planes and the orientation of the calculated maximum compressive stress (σ_1).

in K_2O and decrease in SiO_2 for the youngest rocks (Peccerillo and Taylor, 1976). This increase in K_2O, associated with an H_2O enrichment, is reflected in a change of the phenocryst composition from pyroxene-bearing ones, through hornblende-, to biotite-bearing ones (Seghedi et al., 1987). Hybrid magmatic products, resulting from mixing of distinct magmas during their ascent to the surface, are abundant (Mason et al., 1996). In the southernmost part of South Harghita Mts., two lava domes are described as shoshonitic basalts due to their high K_2O and low SiO_2 contents, with a mineralogical assemblage of olivine, clino- and ortho-pyroxene, and hornblende. The Nd and Sr initial isotopic ratios of the most mafic, and thus the most primitive products (Fig. 9; Mason et al., 1996), are rather scattered and typical of crust-contaminated mantle products. This heterogeneity of isotopic ratios, combined with a great homogeneity of trace element patterns, is however not believed to be a crustal contamination effect during the rise of the magmas to the surface, as assumed by Mason et al. (1996), but rather a source effect (Chalot-Prat and Boullier, 1997). Therefore, we propose that their source is likely in the subcontinental mantle that was previously metasomatised by crustal melts (Chalot-Prat and Boullier, 1997; for details see next section).

The analysis of fault-slip data collected from the south Harghita (Fig. 4b) and Persani (Fig. 7) scoria cones indicates a general NW–SE extension direction, resulting in the development of normal faults, domino, horst and graben structures (Gîrbacea, 1997).

In map view, the eruptive centres of the Persani basalts appear to be aligned in two directions, NE–SW and NW–SE (Fig. 7). The craters of the Calimani–Harghita volcanic chain, that becomes younger to the SE, are aligned along a NW–SE trend only (Szakacs and Seghedi, 1995). Both directions are of regional importance in the southern East-Carpathians. Nevertheless, only the NW–SE trend, which is also the direction of local extension, is thought to underline the orientation of major deep supply conduits. Indeed it is assumed that SW–NE deep fractures do not exist as they would have

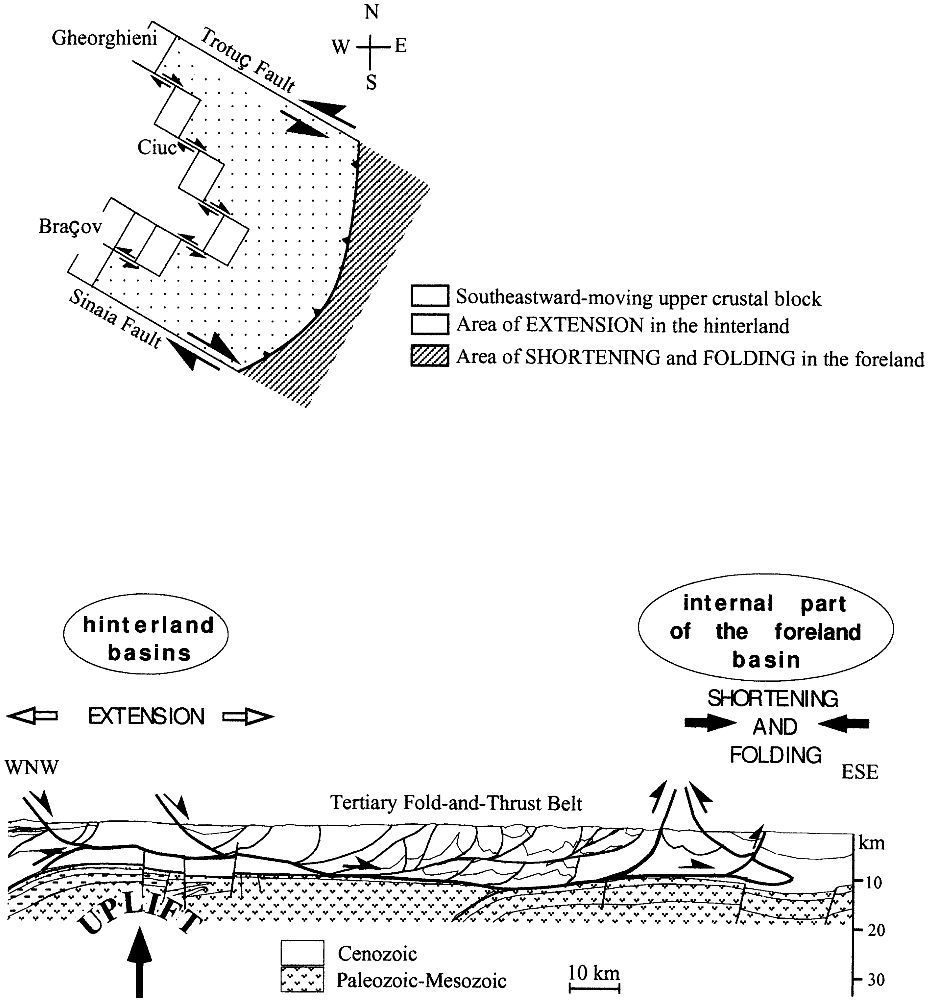


Fig. 6. (a) Kinematic model of the Brasov–Gheorgieni basin formation and foreland folding. The uplift-induced southeastward motion of a crustal block between two strike–slip faults (Trotus and Sinaia) resulted in extension and basin formation in the hinterland, accommodated by coeval shortening in the foreland (see Fig. 1 in Gîrbacea et al., 1998). (b) The crustal motion has taken place above a detachment horizon within the fold-and-thrust belt (after Gîrbacea and Frisch, 1998; Gîrbacea et al., 1998, modified).

prevented horizontal detachment faulting at upper crustal levels and basin formation by gravity spreading. In the Calimani–Harghita chain, opening of a “deep magmatic fracture” has progressively propagated southeastwards since 9.4 My. However within this chain, local N–S and NNE–SSW alignments of volcanoes (Szakacs and Seghedi, 1995) are apparent and indicate secondary fractures. This suggests that opening of supply conduits at upper crustal levels resulted from the interplay between a set of deep and shallow fractures, the distribution of which needs further mapping and structural analysis. In any case, whether they played a major or minor role, supply conduits are not randomly distributed but are closely related to the regional tectonic constraints. Moreover, it is noteworthy that the abundance of differentiated magmatic products among the Calimani–Harghita calc-alkaline series probably reflects that crustal fracturing was not readily achieved, restraining magma ascent.

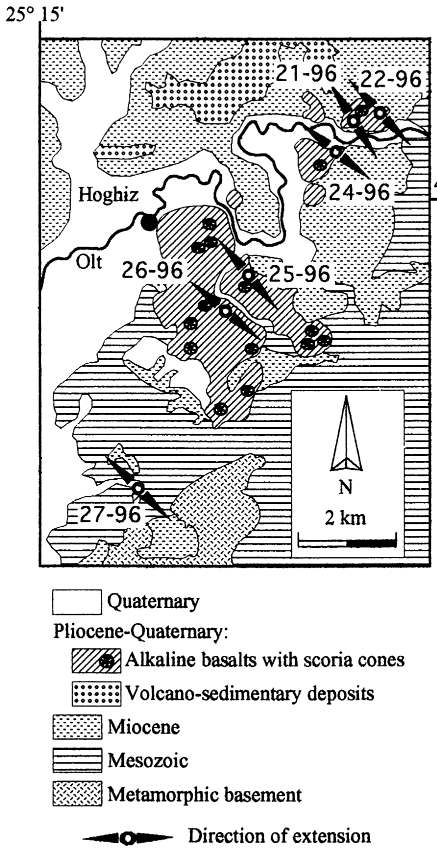


Fig. 7. Geological map of the Persani Mts. and calculated Pliocene–Quaternary kinematic directions. Outcrop numbers concern the measuring sites listed in Table 1.

4. Delamination of the lower mantle lithosphere and involvement of its upper parts in magma genesis

4.1. Intra-mantle delamination model

An intra-mantle delamination model was proposed to explain the Plio–Quaternary tectono-sedimentary evolution of the East-Carpathians (Fig. 8). Physical processes involved in this model are constrained by the following observations. (1) Present seismicity in the East-Carpathians is largely confined to the Vrancea zone where epicentres of earthquakes (up to magnitude 7.4) define a 20/60 km wide and 130 km long, sub-vertical high-density slab at depths between 70 and 200 km (Fuchs et al., 1979; Oncescu, 1984; Fan et al., 1998; Wenzel et al., 1998, 1999); this slab is thought to sink gravitationally into the asthenosphere as suggested by tensional focal mechanism solutions (Oncescu, 1984; Oncescu and Trifu, 1987). (2) This 130 km long slab is located 130 km southeastward of the Neogene suture between the Tisza–Dacia and European plates (Fig. 2). (3) Thermal and seismic models give evidence for thinning of the lithosphere across the Vrancea zone from about 100 km under the European foreland to 40 km beneath the internal East-Carpathians (Fig. 8; Radulescu et al., 1976; Demetrescu and Andreescu, 1994; Mocanu and Radulescu, 1994). (4) A high heat flow of up to 100 mW/m^2 below the volcanic area compares to 45–55 mW/m^2 elsewhere (Geothermal Atlas of Europe, 1991; Mocanu and Radulescu, 1994). (5) The isotopic signature of natural fluids exsolving presently at surface ($\delta^{18}O$ and δD in waters, $^3He/^4He$, $\delta^{13}C$ and CH_4 in gas samples) attests to a dominant asthenospheric mantle contribution in the inner part of the Carpathian arc (volcanic arc and adjacent Transylvanian Basin), and a dominant crustal contribution in its outer Focsani basin part (Minissale et al., 1999).

The new interpretation of the deep lithospheric configuration of the southern East-Carpathians by Gîrbacea and Frisch (1998) implies that slab roll-back and break-off induced delamination of the European lower mantle lithosphere and upwelling of the asthenosphere into the newly created space (Sperner, 1996). Partial delamination of the European continental mantle lithosphere, possibly along a pre-existing rheological boundary of unknown origin, was caused by slab-pull exerted by the subducting heavy oceanic lithosphere still attached to the European plate. Following activation of this intra-mantle detachment level, the delaminating segment of the European lower mantle lithosphere, together with the attached oceanic slab, rolled-back towards the foreland. By now, the delaminated segment stands nearly vertical beneath the Vrancea zone, 130 km to the southeast of the suture zone between the Tisza–Dacia and the European plates. According to this model, the strong Vrancea zone earthquakes, confined to the 130 km long delaminated continental lower lithospheric slab, are caused by gravitational forces exerted on it by the attached oceanic lithosphere.

At larger scale, tomographic studies indicate that a mantle slab extends laterally below the northern

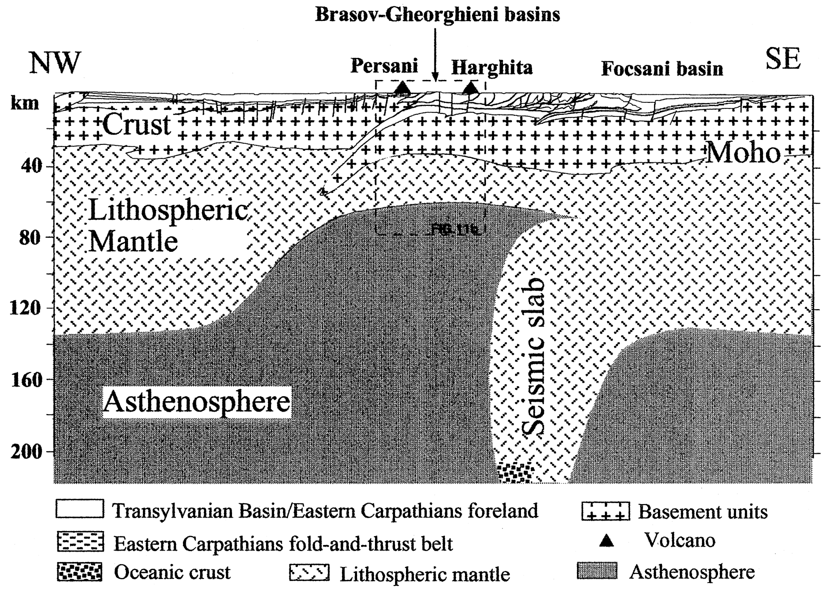


Fig. 8. Delamination model proposed for the Pliocene-recent evolution of the Eastern Carpathians (after Gîrbacea and Frisch, 1998, modified). This model suggests that, after the continental collision in Miocene times, break-off of the west-dipping subducting slab occurred at a depth of 70 km. Slab break-off propagated horizontally towards the east, inducing lithospheric delamination, counterclock-wise rotation of the delaminated lithospheric segment and movement of the Vrancea slab (seismically active due to ongoing pull of the oceanic lithosphere) into its present position. Location of Fig. 11b is shown for a better understanding of the model.

branch of the East-Carpathian arc (Fan et al., 1998). As observed by Wenzel et al. (1999), the uppermost high-velocity material of this slab strikes roughly SW–NE, suggesting lower mantle subduction to the NW, while northwards the deeper portions of the high velocity material are oriented more N–S, close to the strike of the East-Carpathians. Presumably, this was also the strike orientation of subduction before the Vrancea segment was delaminated. Thus the directional change of mantle subduction, from N–S to NE–SW, is likely preserved in the tomographic image. Furthermore, seismicity, although mainly confined to the Vrancea zone, becomes increasingly more active southwestwards and westwards (Fan et al., 1998). This possibly reflects lateral propagation of continental mantle delamination below the southern branch of the East-Carpathian arc.

The hypothesis of mantle delamination is reinforced by the presence of the deep Focsani basin just above the area where the detached lower mantle slab is still attached to the lithosphere. Indeed the Neogene and recent high subsidence rates of this basin are probably caused by the load exerted by this heavy and vertically hanging slab on the overlying continental lithosphere. In other words, the subsidence of the Focsani basin reflects the isostatic response of the lithosphere to a dramatic, though local thickening of the lithospheric mantle in the area of the Vrancea slab (Fig. 8), and not a thrust-loaded downward flexing of the entire lithosphere, as proposed by Royden and Karner (1984). The south-westwards shifting of the actively subsiding zone in the Focsani basin (Matenco et al., 1997) would reflect in turn a shift of the delamination process in the same direction.

4.2. *Upper part of the subcontinental mantle, a likely source of calc-alkaline and alkaline mafic magmas*

As already underlined, the youngest volcanic area cross-cuts the suture between the overriding Tisza–Dacia plate and the subducting European plate. Thus, the mafic volcanoes are necessarily fed by partial melts from the mantle below the lower plate. The upper part of the subcontinental mantle of this lower plate, found as xenoliths in the Persani alkaline

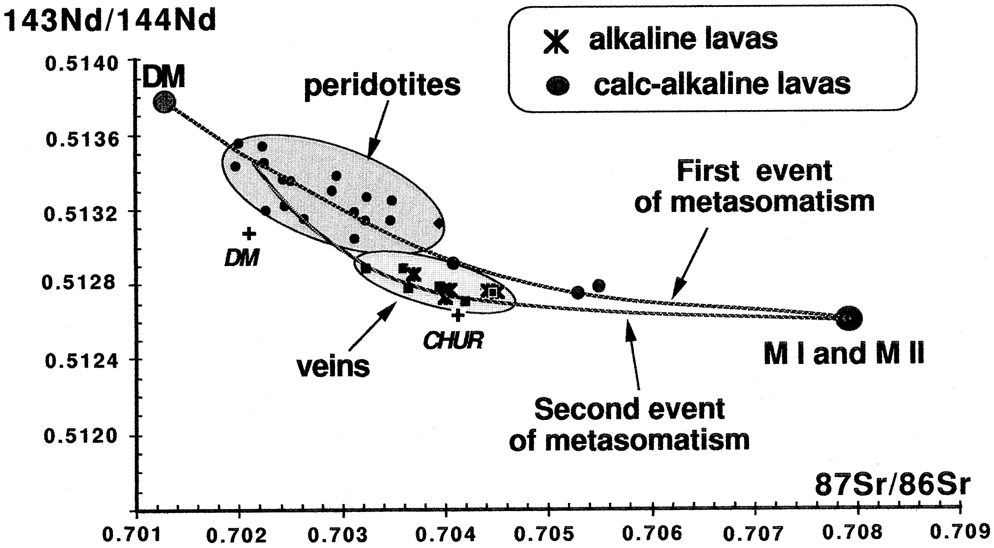


Fig. 9. Relationships, in terms of Nd and Sr isotopic ratios, between alkaline (Persani) and calc-alkaline (south Harghita) volcanics and the subcrustal mantle (peridotites crosscut by veins) below the southern Eastern Carpathians (after Chalot-Prat and Boullier, 1997; modified); see text for explanations. DM: initial depleted mantle, harzburgitic in composition; M I and M II: melts which were successively produced by partial melting of a same source-rock (most likely an eclogite from a polymetamorphic lower crust) and which percolated and metasomatised the subcrustal mantle during a orogenic event well before the occurrence of volcanic eruptions.

basalts, is thought to be the source of calc-alkaline and alkaline mafic volcanics. This hypothesis is strongly constrained by a petrological study of the subcrustal mantle xenoliths (Chalot-Prat and Boullier, 1997). The results of this study, the aim of which was to decipher the nature of the relationships between the subcrustal mantle and the mafic volcanics found at surface, can be summarised as follows: the xenoliths are samples of a subcontinental mantle, mainly represented by peridotites similar to those of the lithospheric mantle elsewhere below the European continent. These peridotites are mainly secondary lherzolites (metasomatic clinopyroxene (cpx)- and/or amphibole (amph)-bearing harzburgites) cross-cut by cpx or amph veins of metasomatic origin. This means that the subcontinental mantle was initially a residual mantle that successively underwent metasomatism by percolation of melts, first in small quantities along grain boundaries leading to formation of interstitial metasomatic minerals and thus to secondary lherzolites, and later in large quantities along fracture systems leading to formation of monomineral (clinopyroxene; amphibole) veins cross-cutting the secondary lherzolites. From the major and trace element compositions of metasomatic minerals (= products of the reaction between host-rock minerals and a percolating melt), these metasomatic events correspond to the percolation of two different ultramafic melts (M I and M II). The Nd and Sr initial isotopic ratios of metasomatic minerals, interpreted in terms of mixing in various proportions between a residual mantle and the contaminant melts, suggest that these melts originated from successive partial melting of a source including a lower crust component (Fig. 9). This source would correspond either to pyrope-bearing eclogites from a deep and polymetamorphic continental crust, or to a carbonated and already crustally metasomatised mantle, or even to a multilayer of both, as often described at crust–mantle interfaces, for example in the Alps (Aurisicchio et al., 1985; Godard et al., 1996; Hermann et al., 1997; Martin et al., 1998; Muntener and Hermann, 1996). The subcontinental mantle of the East-Carpathians, below the European lower plate, shows thus a long geological history. This is a residual mantle which has been modified by several crust recycling events during underthrusting

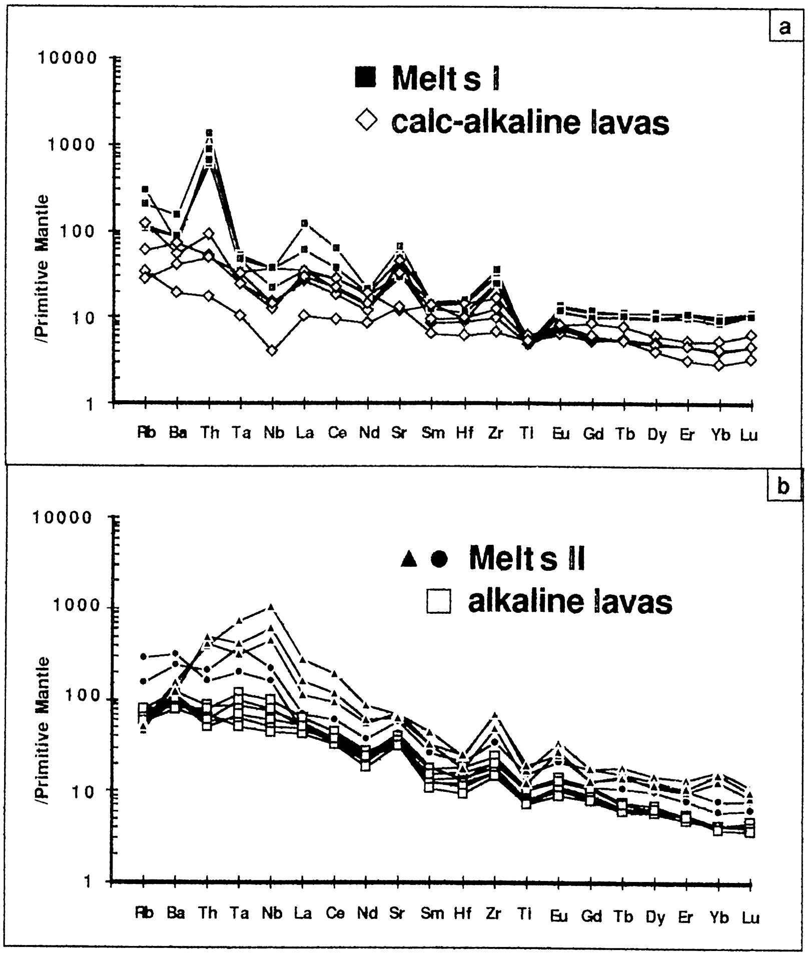


Fig. 10. Relationships, in terms of trace element spectra, between: (a) the contaminant melt M I and the calc-alkaline basic lavas (from the south Harghita Mts); and (b) the contaminant melt M II and the alkaline lavas (from the Persani Mts) (after Chalot-Prat and Boullier, 1997; modified); see text for explanations.

of lower crust into the mantle either at time of Tisza–Dacia — Europe plate convergence and collision, or more likely before during the Variscan and Proterozoic orogenies.

From this, the following partial melting relationships between mantle and erupted basalts are deduced. The inferred bulk compositions of contaminant melts are clearly different from those of alkaline and calc-alkaline basic lavas. Indeed the relative proportion of major elements is not the same and their trace element contents relative to Primitive Mantle are much higher (Fig. 10). Nevertheless, both contaminant melts show strong positive anomalies in Th, Sr and Zr, as well as a negative Ti anomaly. Both types of anomalies are present, although attenuated, in the patterns of the two types of lavas. In the same way, melts I and calc-alkaline lavas on one hand, and melts II and alkaline lavas on the other hand differ from each other by displaying opposed Ta and Nb contents and different REE fractionation patterns. Thus contaminant melts and mafic eruptive magmas can be assumed to be only indirectly related. Therefore, it is suggested that the lavas originated by melting of peridotites already metasomatised by melts I and II. This is supported by: (1) the similarity of isotopic ratios of alkaline basalts and monomineral veins (Fig. 9), which means that the alkaline basic magmas originate from partial melting of secondary lherzolites rich in veins; (2) the fact that the isotopic ratios of the most primitive calc-alkaline lavas (Mason et al., 1996), although they differ from those of the studied secondary lherzolites, lie on the mixing trend DM–MI. This simply means that the mantle source of the calc-alkaline lavas was secondary lherzolites but more metasomatised by melt I than those brought up as xenoliths.

In summary, the upper part of the mantle lithosphere of the East-Carpathians, which is a residual mantle, the composition of which was significantly modified and diversified during previous orogenesis by successive percolation of melts of mainly continental crustal origin, can be considered as the likely source of calc-alkaline and alkaline basalts.

If the composition of the asthenosphere is that of a residual mantle, as commonly accepted (Wilson, 1989), the lithospheric mantle clearly appears here as a mixing interface between asthenosphere and crust. We also note that transfer of continental crustal material into the mantle has been repeatedly invoked on the base of deep reflection- and refraction-seismic surveys (Ziegler et al., 1995, 1998, 2000) to explain an apparent layering and the occurrence of seismic reflectors in the mantle lithosphere. These reflectors were interpreted "as being related to shear zones and relict subduction zones along which lower crustal material was inserted into the mantle" (Ziegler et al., 1995, 1998).

Another relevant observation in these mantle xenoliths is the occurrence of intergranular basaltic pockets (less than 1 mm in diameter) filled by basaltic glass, including a few euhedral spinel, olivine and clinopyroxene crystals (Chalot-Prat and Boullier, 1997; Chalot-Prat and Arnold, 1999). These randomly scattered basaltic pockets occur at triple points and grain boundaries throughout the mantle fragment, and are *everywhere* isolated from each other without any connection with the host-lava of the mantle fragment. They are believed to represent partial melting effects during decompression of the subcrustal mantle at depth, just prior to eruption. This process, already invoked by many authors (Chazot et al., 1996 and refs. therein; Yaxley et al., 1997; Lustrino et al., 1999), requires a dramatic and sudden pressure decrease associated with a volume increase of the material and thus an increase of the available space. Moreover, the periphery of these mantle xenoliths is eventually microfissurated (max. outermost 2–3 mm) and even in process of disagregation in the host-lava that often infiltrated the xenolith along microfissures. These cracking and infiltration processes are heterogeneous at every scale, inducing damaging on the periphery of the mantle fragment. The observed effects of this fissuring, triggered by decompression of the mantle fragment within magma ascending to the surface are different from, and cannot be confused, with those of melting pockets.

Thus, the mantle xenoliths brought to the surface can be regarded as pieces of the subcontinental mantle that were dislocated from just below the mantle–crust interface during the melting process. This hypothesis is fully supported by the seismic refraction data that indicate the presence of a low-velocity zone (LVZ) below the Moho, interpreted as a partially molten mantle zone (Lazarescu et al., 1983). As this LVZ is observed beneath the Plio–Quaternary volcanoes, it is likely that it supplied the partial melts.

Nevertheless, eruption of primitive magmas suggests that basaltic melts, expelled from the host–mantle, migrated rapidly towards the surface throughout a rather thick continental crust. Although density gradients associated with gravity effects easily explain the expulsion of basaltic melts from the host–mantle, these must be injected under pressure into and throughout the crust to prevent their storage at intracrustal levels. This requires a brittle behaviour of the crust at the time of extrusion of such melts. However, at the time of magma genesis, the response of the crust depends not only on its rheological properties but also on regional tectonic constraints (see previous chapter). Thus, in the case where fusion occurs by decompression, the rather localised and partially molten zone then becomes an area of weakness that is subjected to the surrounding lithostatic pressure. Correspondingly, magma is forced out of this partially molten zone and injected upward in the crust with fractures paving the way for their ascent to the surface. Obviously, magma eruption takes only place when magma generation is synchronous with fracturing of the entire crust, a timing taken into account in the new geodynamic model presented below.

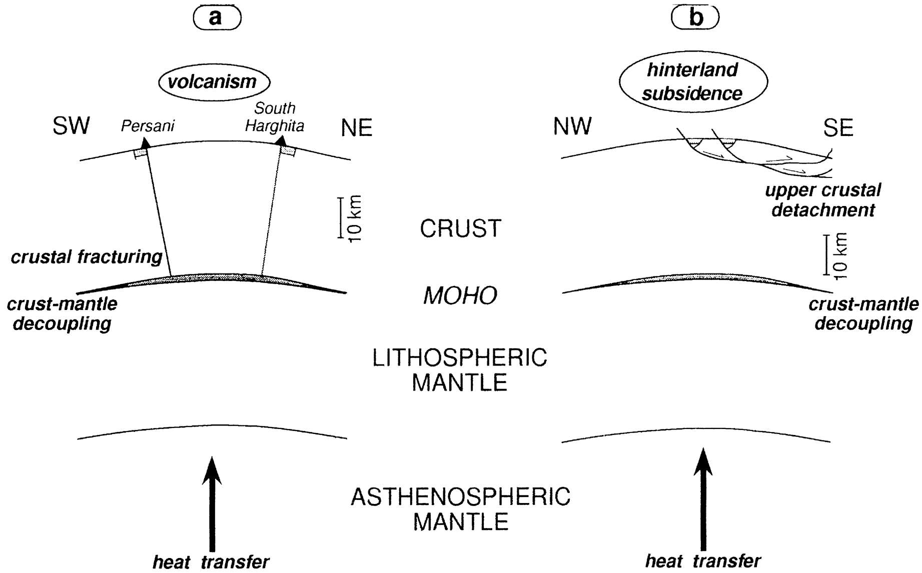


Fig. 11. Geodynamic model explaining together lithospheric uplift, volcanism and basin formation in the southern East Carpathians during the Late Miocene–Quaternary times. (a) SW–NE profile showing fracturation of the crust leading to volcanic eruptions; (b) NW–SE profile showing upper crustal detachment leading to hinterland basin formation by gravity spreading. The succession of events is as follows. (1) Asthenospheric upwelling, itself induced by delamination of the lower part of the lithospheric mantle (see Fig. 8). (2) Temperature increase and density decrease of the overlying lithosphere leading to its uplift. (3) Mantle–crust mechanical decoupling at MOHO level due to the density gap existing between mantle and crust and the different rheological behaviour of each material; crust bends and fractures, while mantle deforms in a ductile way, undergoes decompression and partial melting. (4a) Injection under pressure of magmas throughout the crustal fractures and eruption at surface. (4b) Crustal extension and remobilisation, in the upper crust, of an old detachment horizon above which hinterland basins form by gravity spreading towards SE (see Fig. 6).

5. "Cause and effect" of deep and shallow processes: a new geodynamic model for the southern East-Carpathians

Based on the data presented above, we propose the following model summarised in Figs. 8 and 11. (1) Delamination of the lower part of the European continental mantle lithosphere induces upwelling of the asthenosphere and rapid thermal uplift of the remnant lithosphere. (2) Under a tectonic setting of fast thermal uplift of the lithosphere, partial melting of the upper part of the mantle lithosphere is linked to crust–mantle decoupling. (3) Uplift-induced fracturing of the crust allows for forceful injection of magmas into the crust and their eruption at the surface. (4) Regional uplift of the crust is coupled with upper crustal extension, controlling the development of hinterland basins by gravity spreading above a major supra-crustal detachment level.

In the southern East-Carpathians, the shallow and deep processes, as well as their relative timing, are best explained by the delamination of the lower part of the European mantle lithosphere, giving rise to an asthenospheric upwelling beneath the thinned lithosphere. The isostatic and thermal effects of the upwelling hot and buoyant asthenosphere are responsible for the uplift of the overlying lithosphere (Allen and Allen, 1990; Lachenbruch and Morgan, 1990).

However, as crust and mantle have not only different density, thermal expansion coefficient, textures and compositions, a mantle–crust rheological decoupling occurred along the Moho. While the crust was uplifted relatively faster than the continental mantle, the crust deformed flexurally while the mantle behaved in a ductile way and somewhat expanded such that indeed none voids appeared. This rheological contrast at the Moho level is able to explain: (1) mantle melting without crustal melting; and (2) a brittle failure throughout the crust facilitating volcanic eruptions. Indeed, the temperature increase due to the asthenospheric rise and the pressure drop induced by uplift being equal for both materials at the mantle–crust interface, the decompression experienced by the upper part of the mantle, in response to the crustal bending, is believed to trigger its partial melting. The melting pockets found in mantle xenoliths are small-scale tracers of this decompressional event, while the geophysically defined LVZ is a supporting evidence of this decoupling process at a lithospheric scale. Besides, decompressional partial melting of the upper part of the mantle lithosphere and accumulation of partial melts at the base of the crust enhanced, in return, crust–mantle decoupling and induced further uplift and bending of the crust, its brittle deformation and the always brief opening of crustal-scale fractures through which the mantle-derived magmas ascended rapidly to the surface. Indeed the very short duration of an eruptive period in the same site (Pécskay et al., 1995a; Szakacs and Seghedi, 1995) implies that the brittle behaviour of the lower crust is always a transient process.

During a phase of uplift rate acceleration, 4 My ago, extension reactivated a pre-existing detachment horizon at supra-crustal levels, above which the hinterland basins formed in areas of localised strain along shallow SW–NE extension faults (Fig. 6). Tensional collapse of the internal East-Carpathians, and related south-eastwards mass transfer by gravity spreading, induced shortening and folding of the internal part of the Focsani foreland basin. Tensional hinterland basins developed only at the south-eastern periphery of the youngest volcanic area.

The vertical and very deep main fracture system, controlling the development of the calc-alkaline volcanic chain in a NW–SE direction, propagated from 9.4 My onwards with a rather limited opening rate. In our model, this systematic shift in eruption centres, both in time and space, is related to a south-eastwards propagation of the lower mantle lithosphere delamination. This implies that the delaminating mantle slab initially dipped westwards below the East-Carpathians, which is consistent with tomographic data (Fan et al., 1998; Wenzel et al., 1999). The opening trend of the main magma supply conduits roughly paralleled the strike of the rolling-back mantle slab. Since 2.25 My a new major fracture system, on which the alkaline Persani volcanoes are located, developed to the SW of the earlier one. Note that 6.2 My earlier and 15 km to the NW of the Persani Mts, deep crustal fracturing facilitated the development of the calc-alkaline Rupea volcano. Progressively increasing crustal fracturing and crater development went hand in hand with a dramatic increase in uplift rates, with hinterland basin subsidence beginning at 4 My. That suggests that the area of the crust–mantle decoupling was rather localised until 2.25 My, when it began to enlarge south-westwards. Thus, we assume that: (1) the area, in which the lithosphere is subjected to asthenospheric thermal effects, is currently enlarging south-westwards; (2) the lower mantle delamination is propagating south-westwards as indicated by the SW shift of foreland subsidence and of the zone of deep epicentres.

It cannot be excluded that LVZ was employed as detachment level between the crust and the mantle, facilitating gravitational crustal extension. Although such a mechanism allows for the development of shear zones along which individual blocks slip against each other (Herman and Muntener, 1996; ter Voorde et al., 1998), it is unlikely to provide opening of magma conduits and volcanic eruptions. Therefore, if such a detachment occurs at the Moho level, its vergence should be NW to SE, as in the upper crust. Therefore, the trend of deep and/or shallow detachment horizons depends on the opening directions of supply conduits, themselves controlled by strike of the retreating mantle slab. This may explain, in return, why the strike of major magma supply conduits is NW–SE and normal to the opening direction of hinterland basins (Figs. 6a and 7).

The main subsidence phase of the Focsani basin was controlled by the load of the delaminated lithospheric slab. Its initial slow subsidence during early to middle Miocene times may be related to thrust-loaded deflection of the lithosphere during the onset of underthrusting of the European continental lithosphere below the Tisza–Dacia plate (Demetrescu and Andreescu, 1994). However, only 5 My later, during the Sarmatian, is the onset of the delamination process evident by a dramatic subsidence acceleration of Focsani basin that is coeval with a sharply increased uplift rate of the hinterland that is followed by the onset of the volcanic activity. This suggests that delamination and retreat of the lower mantle lithosphere really began around 12–10 My, triggering upwelling of the asthenosphere beneath the thinned lithosphere.

From the timing and spatial distribution of volcanic eruptions, it is clear that delamination became effective first at the northern end of the East-Carpathians, then propagated southeastwards and is now shifting southwestwards and westwards.

6. Discussions

6.1. Decoupling in an uplift context: primary and secondary effects of a rather common process

In the northern Central Alps, underplated Permian and Early Triassic gabbro sills or laccoliths are observed at the interface between lower crust and the mantle. Their injection occurred in an uplift context and always in very short time frame (Cawthorn, 1975; Rivalenti et al., 1975, 1981; Siena and Coltorti, 1989; Voshage et al., 1990; Herman and Muntener, 1996; Lu et al., 1997a,b; Hermann, 1997; Muntener, 1997; Vavra et al., 1999). They are believed to evidence crust–mantle decoupling which, in this case, must be related to magma injection and not to in-situ partial mantle melting, as invoked in our model. Indeed most of these Permo-Triassic gabbros have not only sharp contacts with the surrounding rocks, but are also representative of MORB-type magmas that were derived by partial melting of the asthenosphere below the host mantle lithosphere. Magma underplating, and thus magma storage at crust–mantle interface, must be due to: (1) the forceful injection of magmas along a lithological and rheological discontinuity which is also a density boundary where in that case magmas find their density equilibrium; (2) the lack of crustal-scale fractures due to the ductile rheological behaviour of the lower crust at the time of magma generation; indeed only few several tens of meter long gabbro dykes cross-cut the base of granulites in the Malenco area. This example of magma underplating in an uplifting context is interesting inasmuch as it gives to us another glance at the *decoupling process during a cycle of post-orogenic uplift* and erosional and tectonic unroofing of the crust (Ziegler and Stampfli, 2000).

Decoupling of crust and mantle appears to be primarily due to the fundamental role of the asthenospheric heat flux, causing differential thermal expansion of the less dense crust as compared to the mantle, resulting in greater uplift rates of the crust than the mantle lithosphere. This is supported by the preliminary modelling results of Ershov and Stephenson (1999) that demonstrate that asthenospheric heat flow triggers decoupling, the different rheological behaviour of crust and mantle reinforcing decoupling and dissociation effect. In the case of underplated gabbros, the lower crust behaved in a ductile way while the upper part of the underlying mantle lithosphere behaved in a brittle way to enable rise of asthenospheric magmas. Therefore, a crust–mantle decoupling process is not systematically related to partial melting of the uppermost parts of the mantle lithosphere. The reasons for such a turnaround of the rheological behaviour between mantle and crust depends mainly on crust and mantle lithosphere thickness and the prevailing thermal regime (Cloethingh and Banda, 1992; Zeyen et al., 1997; Ziegler et al., 2000 and references therein). Let us note that magma injection is always very brief in time, which means that the brittle behaviour of the subcontinental mantle, if it is synchronous with magma genesis, is of very short duration. This is another point of comparison with our model where we emphasised that accounting of the short duration of the eruptive activity in the same site, the brittle behaviour of the whole crust is always a transient event at geological scale.

6.2. Different asthenospheric components of the mantle lithosphere

Apart from triggering a strong heat flux increase, upwelling of the asthenosphere, as well as of mantle plumes, is associated in the southern East Carpathians with enormous fluid transfers throughout the lithosphere and even to the surface, fluids having a more or less primitive mantle isotopic signature, as documented by Minissale et al. (1999). During their percolation towards the surface, a great part of these fluids are injected and trapped within microfissures, cross-cutting minerals of the lithospheric mantle (e.g. Wilson, 1989). Most often, several microfissure systems cross-cutting each other are observed within mantle rock samples, attesting for different phases of percolation, as seen in Persani mantle xenoliths. Overall, the lithospheric mantle includes two distinct asthenospheric components, a fluid component and a silicate component corresponding to the initial (before metasomatism) harzburgitic residual mantle, each of them having a different history since extraction from the asthenosphere. This remark is important inasmuch as until now these two components have not always been clearly distinguished in any approach to study the origin of mantle-derived magmas. In our model, the fluid component enables to roughly delimit the depth from which the asthenospheric plume is/was coming (as deeper as it is more primitive); in contrast, the silicate component enables to decipher the partial melting history of the asthenospheric mantle before it was accreted to the lithosphere and subjected to metasomatism of various origin.

7. Conclusions

In trying to understand the ins and outs of an a-typical volcanism, now operating since 9.4 My in the southern East-Carpathians, we have come up with a geodynamic model that establishes the "cause *and* effect" relationship between shallow and deep processes affecting this area during the last 17 My, that is since the middle Miocene.

The sequence of events was controlled by delamination of the lower lithospheric mantle of the subducting European plate. This intra-mantle delamination began during the Sarmatian (12 My) after the oceanic slab that was attached to the European plate had been subducted and the collision of the Tisza–Dacia and Europe continental plates was well advanced. Gravity-driven sinking and retreat of this lithospheric slab initiated the subsidence of the Focsani foredeep basin in front of the outer Carpathians, where this slab is still attached to the European lithosphere. Detachment and roll-back of this slab triggered beneath the internal Carpathians asthenospheric upwelling and a huge heat flux. Thermal, rather than isostatic effects of the asthenospheric plume, lowered the density of the remnant lithosphere, thus triggering a dramatic uplift of the lithosphere and its internal decoupling at Moho level beneath which the uppermost mantle was subjected to partial decompressional melting. Decoupling at the crust–mantle boundary was reinforced by the different rheological behaviour of crustal and mantle lithologies. However, their rheological behaviour cannot be determined a priori and needs to be constrained by the magmas that were generated coeval with this uplift. In this study the most primitive mantle-derived magmas came from partial melting of the subcrustal mantle during a decompression phase, then were forcefully injected into the fractured continental crust and erupted at surface. It is inferred, that during the uplift-related crust–mantle decoupling, the mantle behaved in a ductile way and expanded while the crust was flexed and locally fractured, providing conduits for magma ascent and volcanic eruptions. At supra-crustal levels, uplift and flexure of the crust induced local extension and the re-activation of a detachment horizon at the base of the Carpathian nappe stack, resulting in the subsidence of tensional hinterland basins, and shortening and folding of the inner part of the Focsani foredeep. Uplift of the inner Carpathians domain was essentially synchronous with a sharp acceleration of the foreland subsidence, and preceded the onset of hinterland basin subsidence.

One of the key points of our geodynamic model is to attempt to constrain the past and future evolution of the ongoing slab delamination and related volcanic activity in the East-Carpathian arc. Tomographic data and the location of recent earthquake epicentres, the eruption timing and the strike of magma supply conduits (NW → SE → SW) are closely linked to mantle delamination that began 9.4 My ago, first at the northern end of the East-Carpathian chain, while the mantle slab still dipped westwards. Since then, the intra-mantle delamination propagated laterally, normal to the slab strike and followed the curvature of the Carpathian arc. Knowing that volcanic activity was never really interrupted, and that the average length of an eruptive period at the same site is around 1 My (cf. Fig. 3 in Szakacs and Seghedi, 1995), the crust–mantle decoupling process appears to be rather fastly stabilised. Nowadays, as the mantle slab is actively foundering below the Vrancea area, a significant southwestwards enlargement of the zone involved in the mantle delamination process is indicated by the progressive shift of deep earthquakes, foreland subsidence and volcanism. Other interesting information on the possibility of next eruptions will come from the location of the LVZ, its outlines and thickness, below the southeast Carpathian Mountains.

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References

Allen, P.A., Allen, J.R., 1990. Basin Analysis — Principles and Applications. Blackwell Scientific Publications, Oxford, 451 pp.

Artyushkov, E.V., Baer, M.A., Mörner, N.A., 1996. The East Carpathians: indications of phase transitions, lithospheric failure and decoupled evolution of thrust belt and its foreland. Tectonophysics 262, 101–132.

Aurisicchio, C., Bocchio, R., Liborio, G., Mottana, A., 1985. Petrogenesis of the eclogites from Soazza, Switzerland. Chem. Geol. 50, 47–63.

Bandrabur, T., Codarcea, V., 1974. Contributii la cunoaSterea depozitelor plio-cuaternare din regiunea cursului superior al MureSului. Stud. Tehnice Economice, H, 5, 23-60, Bucharest.

Bandrabur, T., Ghenea, C., Sandulescu, M., Stefanescu, M., 1971. Neotectonic Map of Romania, Scale 1:1.000.000, Geological Institute, Bucharest.

Casta, I., 1980. Les formation quaternaires de la dépression de Brasov. Université D'Aix Marseille, 256 pp.

Cawthorn, R.G., 1975. The amphibole peridotite–metagabbro complex, Finero, northern Italy. J. Geol. 83, 437–454.

Chalot-Prat, F., Arnold, M., 1999. Immiscibility between calcio-carbonatitic and silicate melts and related wall rock reactions in the upper mantle: a natural case study from Romanian mantle xenoliths. Lithos 46 (4), 627–659.

Chalot-Prat, F., Boullier, A.M., 1997. Metasomatic events in the subcontinental mantle beneath the Eastern Carpathians (Romania): new evidences from trace elements. Contrib. Mineral. Petrol. 129, 284–307.

Chazot, G., Menzies, M.A., Harte, B., 1996. Determination of partition coefficients between apatite, clinopyroxene, amphibole and melt in natural spinel lherzolites from Yemen: implications for wet melting of the lithospheric mantle. Geochim. Cosmochim. Acta 60, 423–437.

Ciulavu, D., 1998. Tertiary tectonics of the Transylvanian Basin. PhD thesis, Vrije Universiteit Amsterdam, Netherlands Research School of Sedimentary Geology, Publ. 981105: 154 pp.

Cloethingh, S., Banda, E., 1992. Europe's lithosphere — physical properties: mechanical structure. In: Blundell, R.F.S.M.D. (Ed.). A Continent Revealed: The European Geotraverse. Cambridge University Press, Cambridge, pp. 80–91.

Csontos, L., 1995. Tertiary tectonic evolution of the Intra-Carpathian area: a review. Acta Vulcanol. 7, 1–14.

Csontos, L., Nagymarosy, A., Horváth, F., Kovac, M., 1992. Tertiary evolution of the intra-Carpathian area: a model. Tectonophysics 208 (1–3), 221–241.

Dallmeyer, R.D., Neubauer, F., Fritz, H., Mocanu, V., 1998. Variscan v.s. Alpine tectonothermal evolution of the Southern Carpathians orogen: constraints from $^{40}Ar/^{39}Ar$ ages. Tectonophysics 290, 111–135.

De Broucker, G., Mellin, A., Duindam, P., 1998. Tectono-stratigraphic evolution of the Transylvanian basin, pre-salt sequence, Romania. In: Dinu, C., Mocanu, V. (Eds), Geological Structure and Hydrocarbon Potential of the Romanian Areas. BGF Special Volume No. 1, Bucharest, pp. 36–69.

Demetrescu, C., Andreescu, M., 1994. On the thermal regime of some tectonic units in a continental collision environment in Romania. Tectonophysics 230, 265–276.

Downes, H., Seghedi, I., Szakacs, A., Dobosi, G., James, D.E., Vaselli, O., Rigby, I.J., Ingram, G.A., Rex, D., Pecskay, Z., 1995. Petrology and geochemistry of late Tertiary/Quaternary mafic alkaline volcanism in Romania. Lithos 35, 65–82.

Ellouz, N., Roca, E., 1994. Palinspastic reconstructions of the Carpathians and adjacent areas sind the Cretacous: a quantitative approach. In: Roure, F. (Ed.), Peri-Tethyan Platforms. Editions Technip, Paris, pp. 51–78.

Ershov, A., Stephenson, R., 1999. Driving forces, strain and temperature necessary for intracratonic rifting and inversions from the viewpoint of rheological models of lithosphere. Joint Meeting of Europrobe Tesz, Pancardi and Georift projects. Tulcea, Romania, 25 Sept.–6 Oct. 1999. Rom. J. Tect. Reg. Geol. 77 (1), 81.

Fan, G., Wallace, T.C., Zhao, D., 1998. Tomographic imaging of deep velocity structure beneath the Eastern and Southern Carpathians, Romania: implications for continental collision. J. Geophys. Res. 103 (B2), 2705–2723.

Fuchs, K., Bonjer, K., Bock, G., Cornea, I., Radu, C., Enescu, D., Jianu, D., Nourescu, A., Merkler, G., Moldoveanu, T., Tudorache, G., 1979. The Romanian earthquake of March 4, 1977: II, Aftershocks and migration of seismic activity. Tectonophysics 53 (3–4), 225–247.

Geothermal Atlas of Europe, 1991. Atlas International Heat Flow Commission of Central Institute for Physics of the Earth — Europe. In: Hurtig, E., Cermak, V., Haenel, R., Zui, V. (Eds.), Geo International Association for Seismology and Thermal Physics of the Earth's Interior. Herman Haack Verlagsgesellschaft

mbH, Geographisch-Cartographische Anstalt, Gotha, Germany, 156 pp.
Ghenea, C., Andreescu, I., Bandrabur, T., CepalÎga, A., Mihaila, N., Trubihin, V., 1979. Bio- and magnetostratigraphic correlations on the Pliocene and Lower Pleistocene Formations of the Dacic Basin and BraSov Depression (East Carpathians). Dari Seama Inst. Geol. Geofiz, Geological Institute, Bucharest, LXVI, pp. 139–156.
GÎrbacea, R., 1997. The Pliocene to recent tectonic evolution of the Eastern Carpathians (Romania). Tübinger Geowissenschaftliche Arbeiten: 1997, pp. 136.
GÎrbacea, R., Frisch, W., 1998. Slab in the wrong place: lower lithospheric mantle delamination in the last stage of the Eastern Carpathian subduction retreat. Geology 26 (7), 611–614.
GÎrbacea, R., Frisch, W., Linzer, H.-G., 1998. Post-orogenic uplift-induced extension: a kinematic model for the Pliocene to recent tectonic evolution of the Eastern Carpathians (Romania). Geol. Carpathica 49 (5), 315–327.
Godard, G., Martin, S., Prosser, G., Kienast, J.R., Morten, L., 1996. Variscan migmatites, eclogites and garnet-peridotites of the Ulten zone, Eastern Austroalpine system. Tectonophysics 259, 313–341.
Herman, J., Muntener, O., 1996. Extension-related structures in the Malenco–Margna-system: implications for paleogeography and consequences for rifting and Alpine tectonics. Schweitz. Mineral. Petrogr. Mitt. 76, 501–519.
Hermann, J., 1997. The Braccia gabbro (Malenco, Alps): Permian intrusion at the crust to mantle interface and Jurassic exhumation during rifting. PhD ETH Zurich, 1997, 193 pp.
Hermann, J., Muntener, O., Trommsdorff, V., Hansmann, W., 1997. Fossil crust-to-mantle transition, Val Malenco (Italian Alps). J. Geophys. Res. 102 (20), 123–132.
Houseman, G., England, Ph., 1986. A dynamical model of lithosphere extension and sedimentary basin formation. J. Geophys. Res. 91 (B1), 719–729.
Jipa, D., 1983. Molasse and marginal geosynclinal facies. A discussion. Ann. Inst. Geol. Geofiz. 62, 251–255.
Lachenbruch, A.H., Morgan, P., 1990. Continental extension, magmatism and elevation; formal relation and rules of thumb. Tectonophysics 174, 39–62.
Lazarescu, V., Cornea, I., Radulescu, F., Popescu, M., 1983. Moho surface and recent crustal movements in Romania; geodynamic connections. Anu. Inst. Geol. Geofiz. LXIII, 83–91.
Le Maitre, R.W., 1989. A Classification of Igneous Rocks and Glossary of Terms. Blackwell, London, 193 pp.
Linzer, H.-G., 1996. Kinematics of retreating subduction along the Carpathian Arc, Romania. Geology 24, 167–170.
Linzer, H.-G., Frisch, W., Zweigel, P., Gîrbacea, R., Hann, H.-P., Moser, F., 1998. Kinematic evolution of the Romanian Carpathians. Tectonophysics 297, 133–156.
Liteanu, E., Mihaila, N., Bandrabur, T., 1962. Contributii la studiul stratigrafiei Cuaternarului din bazinul mijlociu al Oltului. Stud. Cerc. Geol. 1 (VII), 485–511.
Lu, M., Hofmann, A.W., Mazzucchelli, M., Rivalenti, G., 1997a. The mafic–ultramafic complex near Finero (Ivrea-Verbano zone). I. Chemistry of MORB-like magmas. Chem. Geol. 140, 207–222.
Lu, M., Hofmann, A.W., Mazzucchelli, M., Rivalenti, G., 1997b. The mafic–ultramafic complex near Finero (Ivrea-Verbano zone). II. Geochronology and isotope geochemistry. Chem. Geol. 140, 223–235.
Lustrino, M., Melluso, L., Morra, V., 1999. Origin of glass and its relationships with phlogopite in mantle xenoliths from central Sardinia (Italy). Per. Mineral. 68 (1), 13–42.
Martin, S., Godard, G., Prosser, G., Schiavo, A., Bernoulli, D., Ranalli, G., 1998. Evolution of the Deep crust at the junction Ausstroalpine/southalpine: the Tonale Nappe: Evolution of the deep crust in the Central and Eastern Alps. Padova, 1, S.C.T. Padova, Universita di Padova, pp. 3–50.
Mason, P.R.D., Downes, H., Thirwall, M.F., Seghedi, I., Szakacs, A., Lowry, D., Mattey, D., 1996. Crustal assimilation as a major petrogenetic process in the East Carpathian Neogene to Quaternary continental margin arc. J. Petrol. 37, 927–959.
Mason, P.R.D., Seghedi, I., Szakacs, A., Downes, H., 1998. Magmatic constraints on geodynamic models of subduction in the East Carpathians, Romania. Tectonophysics 297, 157–176.
Matenco, L., 1997. Tectonic evolution of the Outer romanian Carpathians: constraints for kinematic analysis and flexural modelling. PhD thesis, Vrije Universiteit, Faculty of Earth Sciences, Amsterdam, 160 pp.
Matenco, L., Zoetemeijer, R., Cloetingh, S., Dinu, C., 1997. Lateral variations in mechanical properties of the Romanian external Carpathians: inferences of flexure and gravity modelling. Tectonophysics 282, 147–166.
Mihaila, N., Kreutzer, H., 1981. Contributii la cunosterea cronologiei vulcanitelor bazaltice din Persanii centrali si sudici. Terra 4, 37–43.
Minissale, A., Vaselli, O., Tassi, F., Seghedi, I., Magro, G., Ioane, D., 1999. Fluid sources in orogenic areas, two examples: Northern Apennines and Eastern Carpathians. Rom. J. Tect. Reg. Geol. 77 (1), 33.
Mocanu, V.I., Radulescu, F., 1994. Geophysical features of the Romanian territory. Rom. J. Tect. Reg. Geol. 75, 17–36.
Motas, C., Tomescu, L., 1983. L'avant-fosse carpathique roumaine: evolution et contenu. Anu. Inst. Geol. Geofiz. 60, 147–158.
Muntener, O., Herman, J., 1996. The Val Malenco lower crust — uper mantle complex and its field relations (Italian Alps). Schweitz. Mineral. Petrogr. Mitt. 76, 475–500.
Muntener, O., 1997. The Malenco peridotites (Alps): petrology and geochemistry of the subcontinental mantle and Jurassic exhumation during rifting. Dipl. Natw. ETH, Swiss Fed. Inst. Techn., Zurich, 205pp.
Nemcock, M., Pospisil, L., Lexa, J., Donelick, R.A., 1998. Tertiary subduction and slab break-off model of the Carpathian–Pannonian region. Tectonophysics 295, 307–340.
Neugebauer, H.J., 1978. Crustal doming and the mechanism of rifting. Part 1, Rift formation. Tectonophysics 45, 159–186.
Oncescu, M.C., 1984. Deep structure of the Vrancea region, Romania, inferred from simultaneous inversion for hypocenters and 3-D velocity structure. Ann. Geophys. 2 (1), 23–27.
Oncescu, M.C., Trifu, C.I., 1987. Depth variation of moment tensor principal axes in Vrancea (Romania) seismic region. Ann. Geophys. Series B: Terrestr. Planet. Phys. 5 (2), 149–154.

Peccerillo, A., Taylor, S.R., 1976. Rare earth elements in the East Carpathian volcanic rocks. Earth Planet. Sci. Lett. 32, 121–126.

Pécskay, Z., Edelstein, O., Seghedi, I., Szakács, A., Kovacs, M., Crihan, M., Bernad, M., 1995a. K–Ar datings of Neogene–Quaternary calc-alkaline volcanic rocks in Romania. Acta Vulcanol. 7 (2), 53–62.

Pécskay, Z., Lexa, J., Szakacs, A., Baloh, K., Seghedi, I., Konecn, V., Kovacs, M., Marton, E., Kaliciak, M., Széky-Fux, V., Poka, T., Gyarmati, P., Edelstein, O., Rosu, E., Zec, B., 1995b. Space and time distribution of Neogene–Quaternary volcanism in the Carpatho–Pannonian region. Acta Vulcanol. 7 (2), 15–28.

Peltz, S., Bratosin, I., 1986. New data on the geochemistry of the Quaternary basalts in the Persani Mountains. Dari de Seama ale Sedintelor Institutului de Geologie Si Geofizica 70–71 (1), 389–403.

Peltz, S., Vasiliu, C., Bratosin, I., 1971. Petrologia rocilor bazaltice Plio–Cuaternare din România. Anuarul Institutului de Geologie si Geofizica, vol. XXXIX, pp. 111–175.

Price, N.J., 1966. Fault and Joint Development in Brittle and Semi-brittle Rock. Pergamon Press, Oxford.

Radulescu, C., Kovacs, Al., Mihaila, N., Samson, 1965. Contributions á la connaissance des faunes de mammifères pléistocènes de la Dépression de BraSov (Roumanie). Eiszeitalter und Gegenwart, vol. 16, Öhringen, pp. 132–188.

Radulescu, D., Cornea, I., Sandulescu, M., Constantinescu, P., Radulescu, F., Pompilian, A., 1976. Structure de la croûte terrestre en Roumanie — essai d'interprétation des études seismiques profonds. Anu. Inst. Geol. Geofiz. L, 5–36.

Rivalenti, G., Garuti, G., Rossi, A., 1975. The origin of the Ivrea-Verbano basic formation (western Italian Alps): whole rock geochemistry. Boll. Soc. Geol. Italy 94, 1149–1186.

Rivalenti, G., Garuti, G., Rossi, A., Siena, F., Sinigoi, S., 1981. Existence of different peridotite types and of a layered igneous complex in the Ivrea zone of Western Alps. J. Petrol. 22, 127–153.

Roure, F., Roca, E., Sassi, W., 1993. The Neogene evolution of the outer Carpathian flysch units (Poland, Ukraine and Romania); kinematics of a foreland/fold-and-thrust belt system. Sediment. Geol. 86 (1–2), 177–201.

Royden, L.H., 1988. Late Cenozoic tectonics of the Pannonian Basin system. AAPG Mem. 45, 27–48.

Royden, L.H., 1993. Evolution of retreating subduction boundaries formed during continental collision. Tectonics 12 (3), 629–638.

Royden, L.H., Baldi, T., 1988. Early Cenozoic tectonics and paleogeography of the Pannonian and surrounding regions. In: Royden, L.H., Horvath, F. (Eds.), The Pannonian Basin: A Study in Basin Evolution. Am. Assoc. Petrol. Geol. Mem. 45, 1–16.

Royden, L.H., Karner, G.D., 1984. Flexure of the continental lithosphere beneath Apennine and Carpathian foredeep basins. Nature 309 (5964), 142–144.

Samson, P., Radulescu, C., 1963. Les faunes mammalogiques du Pléistocène inférieur et moyen de Roumanie. C.R.H. Sc. Ac. Sci. 257 (5), 1122–1124.

Sanders, C., 1998. Tectonics and erosion, competitive forces in a compressive orogen: a fission-track study of the Romanian Carpathians. PhD thesis, Vrije Universiteit, Amsterdam, 204 pp.

Sanders, C., Huismans, R., Bertotti, G., 1997. The East Carpathians: deformation in a double vergent orogenic wedge: European Union of Geosciences EUG-9 Conference. Terra Nova 9, 154 (Abstract Supplement No. 1).

Sandulescu, M., 1975. Essai de synthèse structurale des Carpathes. Bull. Soc. Geol. Fr. 17 (3), 299–358.

Seghedi, I., Szakacs, A., 1994. Upper Pliocene to Quaternary basaltic volcanism in the Persani Mountains. Rom. J. Petrol. 76, 101–107.

Seghedi, I., Szakács, A., Udrescu, C., Stoian, M., Grabari, G., 1986. Trace elements geochemistry of the South Harghita volcanics (East Carpathians): calc-alkaline and shoshonitic associations. Dari Seama Inst. Geol. Geofiz. 72–73, 381–397.

Seghedi, I., Grabari, G., Ianc, R., Tanasescu, A., Vâjdea, E., 1987. Rb, Sr, Th, U, K, distribution in the Neogene volcanics of the South Harghita Mountains. Dari Seama Inst. Geol. Geofiz. 70–71, 353–473.

Siena, F., Coltorti, M., 1989. The petrogenesis of a hydrated mafic–ultramafic complex and the role of amphibole fractionation at Finero (Italian Western Alps). Neues Jahrb. Mineral. Mh. 6, 255–274.

Sperner, B., 1996. Computer programs for the kinematic analysis of brittle deformation structures and the Tertiary tectonic evolution of the Western Carpathians (Slovakia) Tuebinger Geowissenschaftliche Arbeiten (TGA), A27, 120 pp.

Szakacs, A., Seghedi, I., 1995. The Calimani–Gurghiu–Harghita volcanic chain, East Carpathians, Romania: volcanological features. Acta Vulcanol. 7, 135–143.

ter Voorde, M., van Balen, R.T., Bertotti, G., Cloething, S.A.P.L., 1998. The influence of a stratified rheology on the flexural response of the lithosphere to (un)loading by extensional faulting. Geophys. J. Int. 134, 721–735.

Vavra, G., Schmid, R., Gebauer, D., 1999. Internal morphology, habit and U–Th–Pb microanalysis of amphibolite-to-granulite facies zircons: geochronology of the Ivrea zone: (Northern Alps). Contrib. Mineral. Petrol. 134, 380–404.

Voshage, H., Hofman, A.W., Mazzucchelli, M., Rivalenti, G., Singoi, S., Raczek, I., Demarchi, G., 1990. Isotopic evidence from the Ivrea zone for a hybrid lower crust formed by magmatic underplating. Nature 347, 731–736.

Wenzel, F., Achauer, U., Enescu, D., Kissling, E., Russo, R., Mocanu, V., Musacchio, G., 1998. Detailed look at final stage of plate break-off is target of study in Romania: American Geophysical Union, 48, T. Eos, vol. 79, pp. 589–594.

Wenzel, F., Lorenz, F., Sperner, B., Oncescu, M., 1999. Seismotectonics of the Romanian Vrancea area. In: Wenzel, F. (Ed.). Vrancea Earthquakes: Tectonics, Hazard and Risk Mitigation. Kluwer, Bucharest.

Wilson, M., 1989. Igneous Petrogenesis: A Global Tectonic Approach. Unwin Hyman, London, p. 466.

Yaxley, G.M., Kamenetsky, V., Green, D.H., Falloon, T.J., 1997. Glasses in mantle xenoliths from western Victoria (Australia), and their relevance to mantle processes. Earth Planet. Sci. Lett. 148, 433–446.

Zeyen, H., Volker, F., Whehrle, V., Fuchs, K., Sobolev, V., Altherr, R., 1997. Styles of continental rifting: crust–mantle detachment and mantle plumes. Tectonophysics 287, 329–352.

Ziegler, P.A., Stampfli, G.M., 2000. Late Palaeozoic-Early Mesozoic plate boundary reorganization: collapse of the Variscan Orogen and opening of Neotethys. In: G. Cassinis (Ed.), The Continental Permian of the Southern Alps and Sardinia (Italy). Mem. Mus. Natl. Hist., Bercia (in press).

Ziegler, P.A., Cloethingh, S., Van Wees, J.-D., 1995. Dynamics of intra-plate compressional deformation: the Alpine foreland and other examples. Tectonophysics 252, 7–59.

Ziegler, P.A., Van Wees, J.-D., Cloethingh, S., 1998. Mechanical controls on collision-related compressional intraplate deformation. Tectonophysics 300, 103–129.

Ziegler, P.A., Cloetingh, S., Guiraud, R. and Stampfli, G.M., 2000. Peri-Tethyan platforms: Constraints on dynamics of rifting and basin inversion. In: P.A. Ziegler, W. Cavazza, A.H.F. Robertson and S. Crasquin-Soleau (Eds.), Peri-Tethyan Rift!Wrench Basins and Passive Margins. Peri-Tethys Mem. 6. Mèm. Mus. natl. Hist Nat., Paris (in press).