The Neogene–Quaternary volcanism of the Carpathian–Pannonian region: from initial plate tectonic models to quantitative petrogenetic explanations

Szabolcs Harangi[1,2]*, Ioan Seghedi[3] and Réka Lukács[2,4]

[1]Department of Petrology and Geochemistry, Eötvös Loránd University, Institute of Geography and Earth Sciences, Pázmány Péter sétány 1/C, 1117 Budapest, Hungary

[2]HUN-REN-ELTE Volcanology Research Group, Pázmány Péter sétány 1/C, 1117 Budapest, Hungary

[3]Institute of Geodynamics 'Sabba S. Ştefănescu', Romanian Academy, 19–23 Jean-Louis Calderon St., Bucharest-37, Romania

[4]Institute for Geological and Geochemical Research, HUN-REN Research Centre for Astronomy and Earth Sciences (MTA Centre of Excellence), Budaörsi út 45, H-1112 Budapest, Hungary

SH, 0000-0003-2372-4581; IS, 0000-0001-7381-7802; RL, 0000-0002-2338-4209
*Correspondence: harangi.szabolcs@ttk.elte.hu

Abstract: A wide range of volcanic rocks formed in the Carpathian–Pannonian region during the last 20 Myr, closely associated with the tectonic evolution of the area: (1) high-SiO_2 rhyolites and rhyodacites; (2) calc-alkaline-type volcanic suites (basalt–andesite–dacite–rhyolite and adakitic rocks); (3) potassic–ultrapotassic volcanic rocks (shoshonites, latites, leucitites and lamproites); and (4) alkali sodic volcanic rocks (basalts, trachyandesites and trachytes). During the 18.1–14.4 Ma silicic ignimbrite flare-up volcanism, explosive eruptions produced large volume of tephras, which covered extended areas of Europe. Calc-alkaline-type volcanism took place in syn-extensional and post-collisional settings. There is no classic volcanic arc in the Carpathian–Pannonian Region, and so the term 'back-arc basin' should be used for the Pannonian Basin only in a structural geology sense. Alkali basalt volcanism occurred mostly in the peripheral areas of the Pannonian Basin primarily owing to edge-driven convective upwelling of heterogeneous asthenospheric mantle. The youngest eruptions took place 100–30 ka ago, mostly in the geodynamically still active southeast Carpathians. After 50 years of intense research, there are still many open questions to be solved. Interdisciplinary understanding of the results of different disciplines is crucial to better constrain the evolution of this tectonically complex area, to reveal its geoenergy and to assess potential natural hazards.

The Carpathian–Pannonian Region (CPR) is one of the key areas of the Mediterranean in addition to the Aegean, Alboran and Tyrrhenian basins (Fig. 1), where significant lithosphere thinning occurred behind an orogenic area, characterized by an arcuate thrust-fold belt (Horváth *et al.* 1981; Horváth and Berckhemer 1982; Tari *et al.* 1999, 2024; Harangi *et al.* 2006; Horváth *et al.* 2015; Royden and Faccenna 2018). The Pannonian Basin (Fig. 2) has been classified as a typical back-arc basin with highly extended continental crust (Horváth *et al.* 1981; Horváth and Berckhemer 1982; Royden *et al.* 1982, 1983*a*, *b*; Roberts and Bally 2012; Balázs *et al.* 2016, 2022; Tari *et al.* 2024), although it is questionable whether a volcanic arc was indeed developed in the area (Seghedi *et al.* 2013). Special attention has focused on the hydrocarbon and geothermal potential of this region, fostering intense geophysical studies and exploration drilling investigations (e.g. Royden and Horváth 1988; Horváth *et al.* 2006*a*, 2015, 2018; Badics and Vető 2012; Boote *et al.* 2018; Lemberkovics *et al.* 2018). These studies have provided unprecedented views into the basement structure and development of geological and tectonic interpretations, which help to understand the dynamic evolution of the tectonic system.

The formation and evolution of the CPR were accompanied by a wide range of volcanic activities, which have played a significant role in the plate tectonic and geodynamic models developed over the last 50 years (Lexa and Konečny 1974; Póka 1988; Szabó *et al.* 1992; Seghedi *et al.* 1998, 2004*a*, *b*, 2005*a*, 2023; Harangi *et al.* 2001, 2007; Harangi 2001*b*; Konečný *et al.* 2002; Harangi and Lenkey 2007; Kovács *et al.* 2007; Lexa *et al.* 2010; Seghedi and Downes 2011). There is a great diversity of

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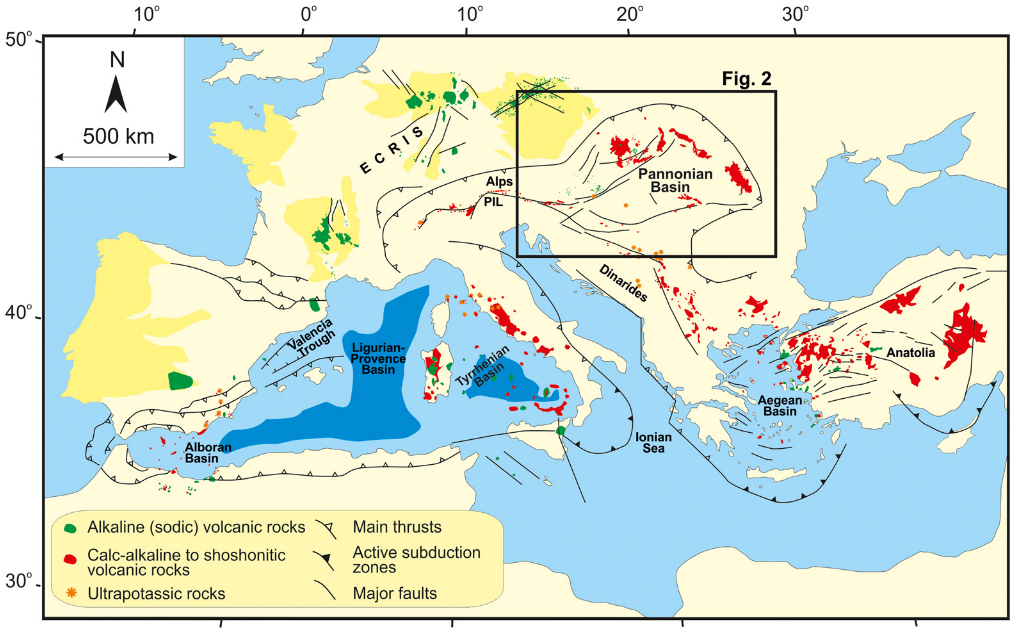


Fig. 1. Neogene–Quaternary volcanic rocks in the Mediterranean region and the surroundings. Note that the tectonic setting of the Pannonian Basin, i.e. development behind the arcuate orogenic belt, shows similarities to back-arc basin areas (Alboran, Tyrrhenian, Aegean) in the Mediterranean. ECRIS, European Cenozoic Rift System; PIL, Periadriatic–Insubric Line. Source: after Harangi *et al.* (2006).

volcanic rocks (from nephelinites to rhyolites, sodic alkaline, calc-alkaline-type and ultrapotassic rock types; Figs 2 & 3) in an area of approximately 600 000 km^2. Thus, the CPR is a natural laboratory not only because of its structural evolution, but also for the better understanding of the origin of these various magmas. Magma generation is a response of the deep mantle as well as plate tectonic processes, therefore knowledge of magmagenesis is vital in any geodynamic models. Igneous petrology and geochronology have made major progress during the last decades, thanks to rapid advances of technology. In this review, we outline the initial thoughts about the volcanism in the CPR proposed during the advent of the plate tectonic concept and show how interpretations became quantitative with time. K/Ar radiometric dating (Balogh *et al.* 1981, 1986; Pécskay *et al.* 2006) played an important role in establishing the temporal relations between the volcanism and the main geological events. More recently, zircon geochronology has become a viable tool to better constrain the eruption ages and to add new information about the lifetime of magma reservoirs and the thermal consequences of magmatism (Lukács *et al.* 2015, 2018, 2021*a*, *b*, 2024). These results have opened a new perspective on how this complex orogenic system involving a basin area underlain by thin crust and lithosphere has evolved. Furthermore, this has enabled a more comprehensive understanding of the resources and geological hazards associated with it.

From the birth of plate tectonic models to the first modern geochemical studies

From the late 1960s, the theory of plate tectonics was increasingly accepted by scientists in eastern-central Europe, and many of them used this new model to explain the regional volcanic activity. One of the most striking elements of the model was subduction of oceanic lithosphere beneath an oceanic or continental plate (Hess 1962; Dickinson and Hatherton 1967; Dickinson 1968; Jakeš and White 1970). The occurrence of volcanoes along these converging plate boundaries was explained by hydrous melting above descending lithospheric slabs (McBirney 1969; Wyllie 1971, 1973; Boettcher 1973). In the CPR, andesitic and basaltic magmas were previously explained by melting of granite bodies (e.g. Kuthan 1968; Danilovich 1972), but in the early 1970s, plate tectonic interpretations were included in published models. Roman (1970) explained the deep earthquakes under the Vrancea area (Fig. 2a), SE Carpathians by a sinking lithospheric slab. Bleahu *et al.* (1973) and Boccaletti *et al.* (1973) connected

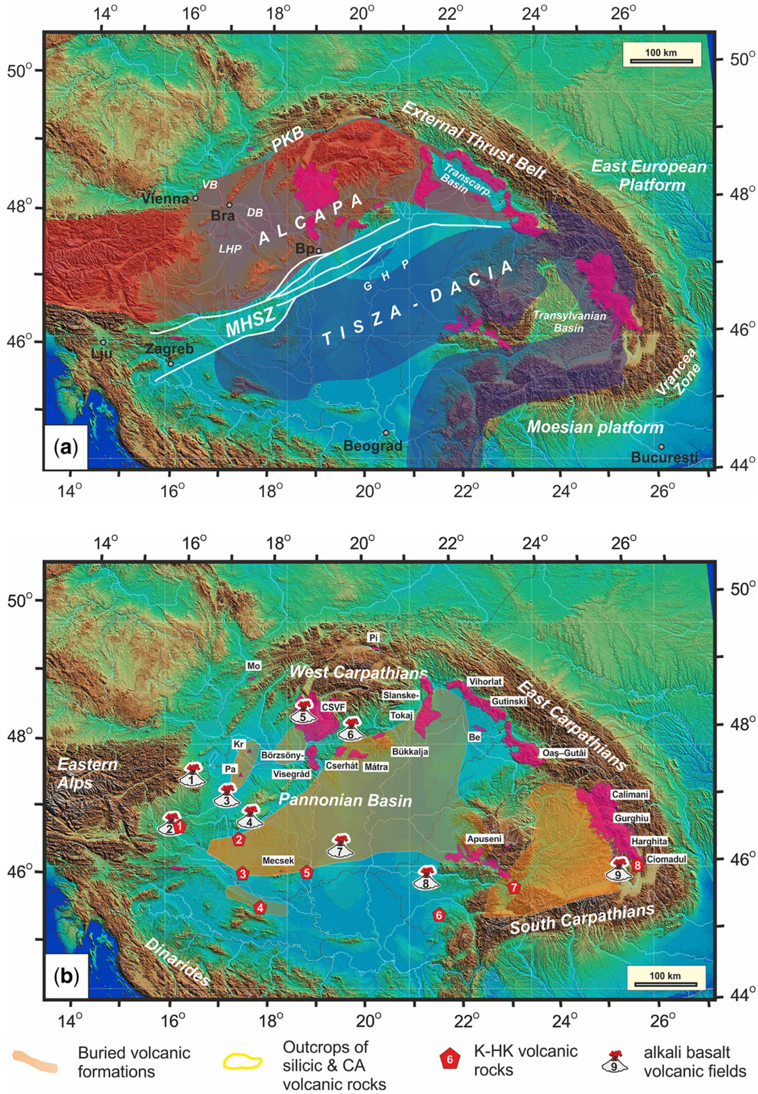


Fig. 2. (**a**) Main structural elements of the Carpathian–Pannonian region. The Pannonian Basin is underlain by two main microplates having different origin and locations during the pre-Cenozoic (Csontos and Vörös 2004). They are separated by the Mid-Hungarian Shear Zone (Csontos and Nagymarosy 1998). The East European Platform and the Moesian Platform are two major crustal blocks, which have distinct thermomechanical properties and contributed to the collision along the East and South Carpathians (Cloetingh *et al.* 2004). (**b**) Spatial distribution of the main groups (CA volcanic rocks; K–HK volcanic rocks and alkali basalt volcanic fields) of Neogene–Quaternary volcanic rocks in the Carpathian–Pannonian region. A large volume of volcanic formations is buried by younger sediments owing to the subsidence of the area. Locations of the potassic–ultrapotassic volcanic rocks: 1, Styrian basin; 2, Balatonmária-1 borehole; 3, Szenta boreholes; 4, Krndija; 5, Bár; 6, Gătaia; 7, Uroi; 8, Malnas-Bixad, South Harghita, Locations of the alkaline basalt volcanic fields: 1, Burgenland; 2, Styrian Basin; 3, Little Hungarian Plain; 4, Bakony-Balaton Upland; 5, Štiavnica; 6, Novohrad/Nógrád-Gemer; 7, Kecel and the surroundings (buried); 8, Lucareţ–Şanoviţa; 9, Perşani. ALCAPA, Alpine–Carpathian–Pannonian and Tisza–Dacia; Be, Beregovo; Bp, Budapest; Bra, Bratislava; CA, silicic and calc-alkaline type; DB, Danube Basin; GHP, Great Hungarian Plain; K–HK, potassic and ultrapotassic rocks; Kr, the buried Kráľová volcano in the Danube basin (Rybár *et al.* 2024); LHP, Little Hungarian Plain; Lju, Ljubljana (in addition to Vienna, Zagreb; Beograd and Bucureşti); Mo, dykes in Eastern Moravia; Pa, the buried Pásztori volcano in the Little Hungarian Plain (Harangi *et al.* 1995*b*; Harangi 2001*a*; Pánisová *et al.* 2018); Pi, dykes at the Pieniny Klippen Belt; PKB, Pieniny Klippen Belt; Transcarp. Basin, Transcarpathian Basin; VB, Vienna Basin. Source: topographic map from Horváth *et al.* (2006*b*).

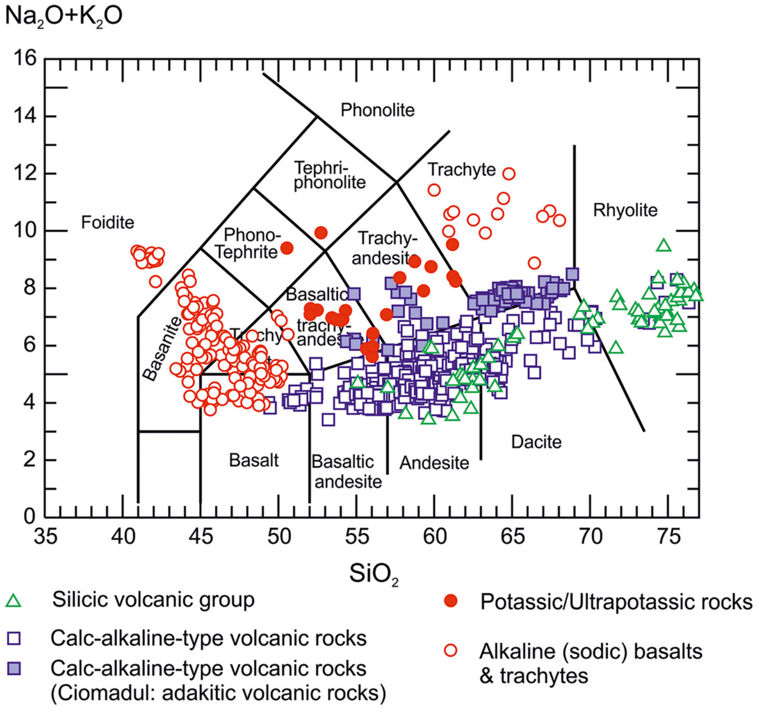


Fig. 3. Compositional diversity and the main groups of Neogene–Quaternary volcanic rocks in the Carpathian–Pannonian Region based on the total alkali–silica diagram (Le Bas *et al.* 1986). SiO_2, Na_2O and K_2O concentrations are in wt%. Source: Data are from Salters *et al.* (1988), Embey-Isztin *et al.* (1993), Dobosi *et al.* (1995), Downes *et al.* (1995*a*, *b*), Harangi *et al.* (1995*a*, 1995*b*, 2001, 2007), Mason *et al.* (1996), Seghedi *et al.* (2001, 2004*a*, 2007, 2008, 2022), Roşu *et al.* (2004), Harangi and Lenkey (2007), Kiss *et al.* (2010) and Lukács *et al.* (2018).

the andesitic volcanism along the East Carpathians to subduction and suggested that an island arc had formed there. Boccaletti *et al.* (1973) put forward that the magmas were formed by melting of the downgoing lithospheric plate at 140–150 km depth, which was at that time a widely accepted explanation (Green and Ringwood 1968). They invoked a westward-dipping subducted slab beneath the East Carpathians. A relationship between the andesitic volcanism and subduction was assumed also by Rădulescu and Săndulescu (1973), although in the case of the Neogene andesitic volcanism in the Apuseni, they had doubts about the connection to subduction because the Miocene deformation was absent. Another proponent of subduction-related andesite volcanism was Szádeczky-Kardoss (1971, 1973, 1974), who identified numerous volcanic arcs in the Pannonian Basin.

The mantle diapir model released by Stegena *et al.* (1975) was an important milestone in the plate tectonic interpretation of the CPR (Fig. 4). Their 'active mantle diapir' explanation was adopted in many subsequent works. However, it is important to note that their interpretation of an 'active mantle diapir' was not the same as hot mantle upwelling, i.e. a mantle plume introduced by Morgan (1971). The authors clearly stated that 'the upwelling, partially melted mantle material – the mantle diapir – is generated at the upper surface of the subductive lithospheric slab (Scholz *et al.* 1971); thus, it is a different term from the "mantle plume" or "hot spot"'. They also considered that subduction along the Carpathians took place during the Early to Middle Miocene and suggested that partial melting occurred in the subducted lithosphere by shear heating. The diapiric mantle upwelling was thought to have consisted of hot partially melted material derived from the subducted slab. When it reached the bottom of the overlying continental lithosphere, it spread laterally and initiated crustal melting ('subcrustal erosion'), which, according to Stegena *et al.* (1975), explained why the lower crust is thinner than the upper crust beneath the Pannonian Basin. Furthermore, they stated that the andesitic and rhyolitic volcanism took place when compressive stresses prevailed in the area because of subduction, whereas it changed

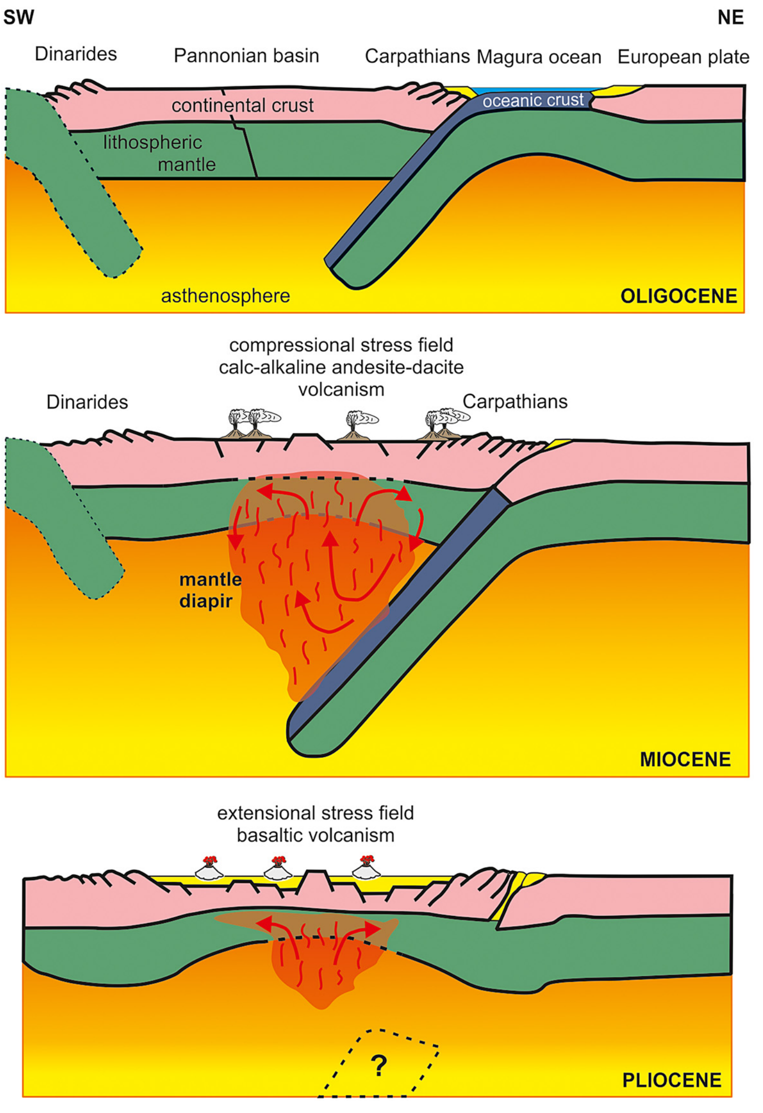


Fig. 4. First and widely referred plate tectonic model for the relationship of geodynamic and volcanic processes of the Pannonian Basin. See text for explanation. Source: model provided by Stegena *et al.* (1975).

to alkali basalt volcanic activity under a subsequent tensional stress environment. This is a very similar scenario to the tectonic evolution and volcanism shown in the Basin and Range area of western USA (Glazner and Bartley 1984; Ormerod *et al.* 1988; Gans *et al.* 1989; Wernicke *et al.* 1996; Putirka and Platt 2012).

The model of Stegena *et al.* (1975) was already presented in an international conference about the new plate tectonic model held in Budapest in 1974 (and published in a special issue of the Acta Geologica Hungarica), and this had a major influence on the thinking of many geologists. Among them, Lexa and Konečny (1974) linked the calc-alkaline andesitic volcanism along the Western Carpathians to diapiric upwelling of the mantle rather than to active subduction. In the 1960s and 1970s, the origin of andesitic magmas in subduction zones was a fundamental question. The most widely accepted model explained magma formation by partial melting of the subducting plate, following Green and Ringwood (1968). However, several authors such as McBirney (1969), Boettcher (1973) and Wyllie (1971, 1973) excluded slab melting and proposed that melting occurred in the mantle wedge above the subducted slab owing to the decrease in melting point by hydrous fluids released by the downgoing lithosphere. Although Lexa and Konečny (1974) accepted the latter model, they rejected a close connection between the andesitic volcanism in the CPR and ongoing subduction. They pointed out that the volcanism was younger by 5–20 Ma than the presumed subduction along the Outer Carpathians, and that no systematic change in major-element composition could be detected in the volcanic rocks perpendicular to the subduction zone. They denoted these volcanic rocks as ‘areal-type’, distinguishing them from subduction-related ‘arc-type’ associations. Within the areal-type volcanic suite, Lexa and Konečny (1974) considered that the rhyolitic volcanism in the interior of the Pannonian Basin was independent of the andesitic volcanic activity and that the silicic magmas were generated by partial melting of the continental crust, in agreement with Pantó (1969) and Konečný and Slávik (1974). In summary, Lexa and Konečny (1974) adopted the mantle diapir model of Stegena *et al.* (1975), the difference being that they considered that mantle upwelling was driven by friction heat from subduction and not by partially melted material. Thus, they considered only an indirect link between subduction and andesite magma generation.

The model proposed by Lexa and Konečny (1974) was sharply criticized by Balla (1981), who followed the view of Szádeczky-Kardoss (1974) and assumed several subduction zones and associated volcanic arcs in the CPR. Balla (1980, 1981) divided the calc-alkaline volcanic areas into two main parts: (1) the Inner Carpathian Volcanic Area, which included the volcanic complexes in the Northern Pannonian Basin and the Apuseni; (2) the Outer Carpathian Volcanic Area, which comprised the volcanic complexes running parallel with the East Carpathians orogen belt. He recognized the differences between these two regions but connected both to active subduction zones.

The original idea, as well as the classification of volcanic rocks in the CPR by Lexa and Konečny (1974), was retained throughout their later publications (Lexa *et al.* 1993, 2010; Konečný *et al.* 1995*a*, 2002; Lexa and Konečný 1998). They distinguished the Neogene–Quaternary volcanic rocks in the CPR based on their spatial distribution and their tectonic and genetic relationships: (1) areal-type, silicic volcanic rocks associated with the initial phase of back-arc basin extension; (2) areal-type andesitic volcanism associated with the more advanced stage of back-arc basin extension; (3) arc-type andesitic volcanism associated directly with subduction; and (4) alkaline basalt volcanism related to post-convergence extension. Nemčok and Lexa (1990) further refined the model of the areal-type volcanism in the Central Slovak Volcanic Field (Fig. 2b). Here they identified a Basin and Range-type volcanism associated with extensional tectonic structures and emphasized that calc-alkaline volcanism occurred under an east–west tensional stress field, because of the diapirically updoming mantle, with only a subordinate role of subduction. The first sign of extensional tectonics is seen in the Early Badenian formations, which corresponds to the onset of the andesitic volcanism, and later, during the Late Badenian and Early Sarmatian, extension became more intense, associated with the culmination of volcanic activity. Later, Harangi (1999, 2001*b*), Harangi *et al.* (2001, 2007) and Harangi and Lenkey (2007) presented petrological and geochemical observations and model calculations about the origin of the magmas in the Northern Pannonian Basin, which were consistent with the early idea of Lexa and Konečny (1974) that the Mid-Miocene andesitic and rhyolitic volcanism was related not to ongoing subduction but to lithospheric and crustal thinning. On the other hand, Seghedi *et al.* (1998, 2004*a*, 2005*a*, 2011, 2019) and Seghedi and Downes (2011) pointed out that the calc-alkaline volcanic rocks along the Eastern Carpathians, classified previously as arc-type, are actually post-collisional, since the volcanic activity did not occur simultaneously with the presumed active subduction, but took place several million years after it, following continent–continent collision. Detailed field mapping of the volcanic areas in the East Carpathians resulted in several papers on the volcanology, petrology and geochemistry of these volcanic rocks (e.g. Seghedi *et al.* 1986, 1987; Szakács and Seghedi 1989; Szakács *et al.* 1993).

A major development occurred in the 1980s, when new analytical techniques allowed determination of major- and trace-element compositions as well as radiogenic isotope ratios with high precision. The first modern geochemical analytical results on the Neogene–Quaternary volcanic rocks were published by Salters *et al.* (1988) in the AAPG Memoir, which contained state-of-the-art papers about the characteristics and evolution of the Pannonian Basin (Royden and Horváth 1988). Based on Sr–Nd–Pb isotope ratios, they suggested a common mantle source for the calc-alkaline and alkaline magmas. This mantle source region was variously metasomatized by components derived from Cretaceous subduction. The calc-alkaline magmas underwent crustal contamination, whereas the alkaline basaltic magmas reached the surface without involvement of crustal material. In this Memoir, Póka (1988) statistically evaluated the available major-element geochemical data on the Neogene volcanic rocks of the CPR and interpreted their genesis based on the Stegena *et al.* (1975) mantle diapir model. Another overview paper was given by Szabó *et al.* (1992), who directly linked the calc-alkaline volcanism to subduction. Using multivariate mathematical techniques, they compared the compositional characteristics of the volcanic rocks in the CPR with an extended database and suggested that the Western Carpathians had an active continental margin affinity, whereas the East Carpathians had an island arc nature.

Novel petrological and geochemical tools were used to investigate the mafic alkaline magmas and the lithospheric mantle using ultramafic xenoliths in the alkaline basalts (Embey-Isztin 1976, 1981; Embey-Isztin and Scharbert 1981; Embey-Isztin *et al.* 1985, 1989, 1990, 1993; Dobosi 1989; Dobosi *et al.* 1991; Kurat *et al.* 1991; Dobosi and Fodor 1992; Downes *et al.* 1992; Konečný *et al.* 1995*c*). The International Association of Volcanology conference held in Mainz in 1990 was a milestone in research on the volcanic rocks of the CPR, with initiation of several new international collaborations. These provided an opportunity to produce modern analytical data and present the results to international audiences. These collaborations and associated results led to a thematic volume of *Acta Vulcanologica* (Downes and Vaselli 1995) about the Neogene–Quaternary volcanism. This volume covered every volcanic area of the CPR, described the first modern volcanological results (e.g. Capaccioni *et al.* 1995; Harangi and Harangi 1995; Karátson 1995; Konečný *et al.* 1995*a*, *b*; Seghedi *et al.* 1995; Mason *et al.* 1995), published a comprehensive geochronological summary about the eruption ages (Pécskay *et al.* 1995*a*, *b*), recognized the group of ultrapotassic volcanic rocks (Harangi *et al.* 1995*a*) and, particularly, brought together many geoscientists working on these topics.

In the late 1990s, four significant papers were published (Mason *et al.* 1996, 1998; Nemčok *et al.* 1998; Seghedi *et al.* 1998), which used geological observations but also modern geochemical data as well as theoretical considerations to explain the complex interplay of volcanism and plate tectonic processes. They included a new element, *slab break-off* (von Blanckenburg and Davies 1995; Wortel and Spakman 2000), which occurs when the subducted dense oceanic slab exerts a significant downward force, while the less dense continental slab cannot follow it, and therefore detachment occurs. A gradual slab break-off was put forward from north to south along the East Carpathians (Mason *et al.* 1998; Nemčok *et al.* 1998; Seghedi *et al.* 1998; Wortel and Spakman 2000; Sperner *et al.* 2001, 2002), which could explain the southward younging of the volcanic activity and parallel alignment of the volcanic complexes close to the suture zone. Seghedi *et al.* (1998) emphasized the complexity of the connection between volcanism and plate tectonics. They distinguished four segments of andesitic volcanic areas within the CPR. The Mid-Miocene andesitic volcanic activity in the northern part of the Pannonian Basin was interpreted similarly to the model of Lexa and Konečny (1974), i.e. magmas originated in a back-arc setting during slab roll-back. In the northern part of the East Carpathians, andesitic volcanism could have been related to slab break-off, whereas in the southern segment, magmas were formed by partial melting of the detached oceanic slab. Andesitic volcanism in Apuseni was due to lithospheric extension triggered by block rotation and translation. These theories were later tested with the advent of high-precision geochemical compositional data on volcanic rocks.

Towards the quantitative petrogenetic models

The early 1990s was a turning point in the interpretation of magmatic processes, since high-precision determination of major- and trace-element and radiogenic isotope composition of volcanic rocks became available, and the political changes allowed scientists from the region to perform analytical work in well-equipped institutes.

The rapidly growing published geochemical composition data both for bulk rocks and various mineral and glass phases (e.g. Embey-Isztin *et al.* 1993; Dobosi *et al.* 1995; Mason *et al.* 1995, 1996; Downes *et al.* 1995*a*, *b*; Harangi *et al.* 1995*a*, *b*, 2001, 2005, 2007, 2015*a*; Harangi 1999; Karátson *et al.* 2000, 2001, 2007; Seghedi *et al.* 2001, 2004*a*, *b*, 2005*a*, *b*, 2011, 2023; Roşu *et al.* 2004; Harangi and Lenkey 2007; Harris *et al.* 2013; Kovacs *et al.* 2017; Lukács *et al.* 2018, 2024; Molnár *et al.* 2018, 2019; Bracco-Gartner *et al.* 2020; Hencz *et al.* 2024) quantitatively constrained the magma source regions and magma evolution. This produced a better understanding of the relationships between volcanism and plate tectonics and enabled testing and refinement of the older models.

Embey-Isztin *et al.* (1993) used trace-element and radiogenic isotope data to constrain the mantle source regions of the Late Miocene to Quaternary sodic alkaline basaltic magmas, whereas Mason *et al.* (1996) published the first comprehensive paper on the petrogenesis of andesitic to dacitic volcanism along the East Carpathians. These studies showed the major differences between these volcanic activities. While the compositional variation in the basaltic rocks was explained by derivation from slightly distinct, mostly asthenospheric mantle sources, the andesitic to dacitic magmas showed more complex evolution with variable contamination by crustal material. Embey-Isztin and Dobosi (1995) and Embey-Isztin *et al.* (1993, 2001) pointed out that the alkali basaltic magmas in the peripheral areas (e.g. Styrian Basin and Novohrad/Nógrád-Gemer) were derived from an isotopically depleted, but trace-element-enriched mantle source similar to that identified beneath the basaltic volcanic fields in other parts of Europe (Cebriá and Wilson 1995; Hoernle *et al.* 1995). In contrast, the isotopic composition of the alkali basalts in the interior of the Pannonian Basin was more diverse, and it was assumed to reflect a subduction component residing in the metasomatized lithospheric mantle and incorporated into the uprising asthenosphere-derived magmas. Embey-Isztin and Dobosi (1995), Šefara *et al.* (1996) and Embey-Isztin *et al.* (2001) proposed involvement of a mantle plume in the genesis of the alkali basaltic magmas based on the isotopic signatures. Seghedi *et al.* (2004*b*) also invoked the role of a limited extent of hot mantle upwelling ('mantle plume finger') similar to that proposed for many alkaline basaltic areas of Europe (Granet *et al.* 1995; Wilson and Patterson 2001). In contrast, Harangi and Lenkey (2007) pointed out that the available data did not support the existence of a mantle plume beneath the Pannonian Basin. Instead, they emphasized the role of lithospheric extension, which could have controlled the formation of the Middle Miocene silicic and andesitic magmas. Formation of the alkali basaltic magmas can be related to the upwelling of enriched asthenospheric material beneath the thinned lithosphere, mainly in the peripheral, western and northern regions (Ali and Ntaflos 2011; Ali *et al.* 2013; Harangi *et al.* 2015*a*). Nevertheless, the mantle plume explanation was employed, even in recent papers (e.g. Neubauer and Cao 2021).

The presence of a geochemically enriched metasomatic component in the lithospheric mantle caused by earlier subduction event(s) was also detected in the ultrapotassic magmas (Harangi *et al.* 1995*a*), and as Harangi (1999, 2001*b*) and Harangi *et al.* (2001, 2006, 2007) pointed out, this was also the source of the primary magmas of the Mid-Miocene calc-alkaline andesitic-to-dacitic volcanism in the Northern Pannonian Basin. Seghedi *et al.* (2008) described a unique Pleistocene lamproite in the southwestern part of the Pannonian Basin (Gătaia, Fig. 2b), assumed to be derived from a highly metasomatized mantle. Thus, the subduction-related compositional features of these volcanic rocks do not necessarily indicate a close relationship to ongoing subduction but are inherited characteristics of the lithospheric mantle. As Harangi (1999, 2001*b*) and Harangi and Lenkey (2007) emphasized, the volcanism was closely associated with the syn-rift phase of the Pannonian Basin when significant lithospheric extension occurred. Thinning of the lithosphere caused decompression-driven partial melting of the most fusible, metasomatized lithospheric mantle, which has a lower solidus temperature, while after progressive thinning, magmas were derived from the asthenospheric mantle (Harangi *et al.* 2007). This is particularly well detected in the Central Slovakian Volcanic Area (Fig. 2b). Occurrence and preservation of primary almandine garnet in the earliest volcanic rocks indicate the changing geodynamic environment from compression to extension (Harangi *et al.* 2001). These new petrological and geochemical results were consistent with the main geodynamic conclusion of Lexa and Konečny (1974) and added a quantitative petrogenetic model for magma generation in this continental extensional setting (Harangi *et al.* 2001, 2006, 2007).

The calc-alkaline volcanic suites in the East Carpathians and Apuseni have a distinct origin, as reflected in their geochemical compositions (Mason *et al.* 1996; Seghedi *et al.* 2001, 2004*a*, *b*, 2005*a*, *b*, 2011, 2022, 2023; Roşu *et al.* 2004; Harris *et al.* 2013; Fedele *et al.* 2016). Petrogenetic modelling suggested that these magmas originated in a heterogeneous mantle and underwent various degrees of crustal contamination. In addition, andesites with high Sr/Y ratios, i.e. an adakitic character, in the Southern Harghita and the Apuseni Mts require a different origin, either slab melting or melting of the mafic lower crust. The volcanic activity postdates the presumed subduction and peak continent–continent collision, and therefore has been interpreted as post-collisional (Seghedi *et al.* 2011, 2019). Thus, it appears that, although vast amounts of andesitic to dacitic volcanic rocks were formed in the CPR, no direct connection can be inferred to active subduction.

A similar distinct and controversial origin has been proposed for the small alkaline-calc-alkaline intrusions in the Carpathian Măgura nappes (Pin *et al.* 2004; Trua *et al.* 2006; Nejbert *et al.* 2012; Anczkiewicz and Anczkiewicz 2016; Krmíček *et al.* 2020) from Eastern Moravia and Pieniny (Fig. 2b). They were formed at 11–13 Ma as shown by K/Ar (Pécskay *et al*. 2006, 2015) and U–Pb zircon dating results (Anczkiewicz and Anczkiewicz 2016). Melting of a variably metasomatized lithospheric mantle source and subsequent fractional crystallization of the primary basaltic magma or/and melting of an ancient crustal material have been suggested to explain their origin.

Constraints on the eruption ages

A vast amount of geochronological data constrains the eruption ages of the Neogene–Quaternary volcanism of the CPR (Pécskay *et al.* 1995*a*, 2006), primarily owing to the establishment of a laboratory for K/Ar dating in Debrecen, Hungary (Balogh 1985; Balogh and Simonits 1998). K/Ar dating was performed partly on determination of eruption ages of the sodic alkali basalt volcanism (Balogh *et al.* 1981, 1986, 1994*a*, *b*, 2005; Konečný *et al.*

1999), whereas other projects focused on the more evolved rocks (e.g. Hámor *et al.* 1985; Peltz *et al.* 1985; Pécskay *et al.* 1986, 1994, 1995*a*, *b*, 2000, 2006; Széky-Fux *et al.* 1987; Edelstein *et al.* 1992; Roşu *et al.* 1997; Márton and Pécskay 1998; Birkenmajer and Pécskay 2000; Seghedi *et al.* 2005*b*, 2022, 2023; Szakács *et al.* 2015). Broad cooperation was established with scientists who worked on different volcanic areas of the CPR, resulting in a self-consistent geochronological data set for most volcanic rocks of the CPR. More recently, additional K/Ar and Ar/Ar geochronological data for various volcanic rocks were published by Wijbrans *et al.* (2007), Szakács *et al.* (2012), de Leeuw *et al.* (2010, 2012, 2013, 2018), Chernyshev *et al.* (2013), Panaiotu *et al.* (2013), Lahitte *et al.* (2019), Rybár *et al.* (2019), Dibacto *et al.* (2020), Sant *et al.* (2020), Kohut *et al.* (2021), Šarinová *et al.* (2021) and Karátson *et al.* (2022), among others.

For the last decade, zircon geochronology has been applied to constrain eruption ages for the silicic volcanic rocks (Lukács *et al.* 2015, 2018, 2021*a*, 2024; Brlek *et al.* 2020, 2023; Bercea *et al.* 2023; Šegvić *et al.* 2023; Trinajstić *et al.* 2023), although there were applications also for the andesitic–dacitic and basaltic rocks (Harangi *et al.* 2015*b*, 2020; Molnár *et al.* 2018, 2019; Hurai *et al.* 2013; Bouloton and Paquette 2014; Anczkiewicz and Anczkiewicz 2016; Lukács *et al.* 2021*b*; Seghedi *et al.* 2023). This method provided more accurate ages for eruption timing than K/Ar data and considerably refined the eruption chronology of the Miocene silicic volcanism. Furthermore, these data provided a robust correlation between proximal and distal volcanic deposits. Zircon geochronology also played a role in establishment of the eruption history of Ciomadul, the youngest volcano of the CPR, where a combination of zircon U/Th, (U–Th)/He and U/Pb dating techniques was applied (Harangi *et al.* 2015*b*, 2020; Molnár *et al.* 2018, 2019). In addition to eruption ages, the time-scale of magma reservoirs was also constrained (Harangi *et al.* 2015*b*; Lukács *et al.* 2021*b*). The youngest eruptions were dated also by radiocarbon techniques on charcoal fragments (Moriya *et al.* 1996; Vinkler *et al.* 2007; Harangi *et al.* 2010, 2015*b*; Karátson *et al.* 2016). Indirect dating by luminescence method was also applied to Ciomadul (Harangi *et al.* 2015*b*, 2020; Karátson *et al.* 2016, 2019) and the youngest basalt eruption in the CPR (Šimon and Maglay 2005).

Volcanism in the Carpathian–Pannonian region: from source to surface

By the end of the 1990s, a robust general geodynamic model had been developed for the formation and evolution of the Pannonian Basin (Horváth *et al.* 1981; Royden *et al.* 1982, 1983*a*, *b*; Csontos *et al.* 1992; Tari *et al.* 1992; Horváth 1993, 1995; Fodor *et al.* 1999). The main element of this model was retreating subduction with slab roll-back along the East Carpathians and associated lithospheric thinning behind it. As a result, the Pannonian Basin has been considered as a typical back-arc basin, similarly to the Aegean, Tyrrhenian and Alboran basins in the Mediterranean (Fig. 1). In detail, the evolution of the CPR has involved the following tectonic processes (Fig. 5; Horváth *et al.* 1981, 2006*a*, 2015; Csontos 1995; Schmid *et al.* 2008, 2020; Ustaszewski *et al.* 2008; Handy *et al.* 2015):

(1) Lateral extrusion and eastward movement of the Alpine–Carpathian–Pannonian ALCAPA microplate (Fig. 2a) from the compressional Alpine collision zone (Kázmér and Kovács 1985; Ratschbacher *et al.* 1991*a*, *b*; Csontos *et al.* 1992). Later it was juxtaposed with the Tisza–Dacia microplate (Fig. 2a) and moved further east-northeastward.

(2) Gradual roll-back of subducted oceanic slab, retreating subduction presumably along the foreland of the Eastern Carpathians (Royden *et al.* 1982, 1983*a*). The vertical, slowly subsiding lithospheric fragment beneath the Vrancea zone (Fig. 2a; Roman 1970) may be the last remnant of this subduction after gradual slab detachment (Sperner *et al.* 2004), although other explanations such as delamination of overthickened lithospheric roots haven been also proposed (Ismail-Zadeh *et al.* 2012).

(3) Significant thinning of the overriding continental lithosphere from about 21 to 13 Ma (central and western regions) and to 8–9 Ma (eastern regions; Balázs *et al.* 2016). Core-complex-type extension was followed by formation of several extensional basins in both microplates (Tari *et al.* 1992, 1999, 2024). Thinning affected mostly the mantle and lower crustal parts of the lithosphere (Horváth *et al.* 2006*a*) and was accompanied by passive upwellling of the asthenospheric mantle.

(4) Lateral asthenospheric mantle flow beneath the Pannonian Basin, which may have occurred from the Alps towards the east (Royden *et al.* 1982, 1983*a*; Horváth and Faccenna 2011; Kovács *et al.* 2011, 2012*a*) or from the northwest to the southeast (Qorbani *et al.* 2016; Song *et al.* 2019) due to slab roll-back beneath the Eastern Carpathians or may have taken place actively extruding from the Alpine collision zone. Horváth and Faccenna (2011) and Faccenna *et al.* (2014) suggested that formation of a slab window owing to slab break-off

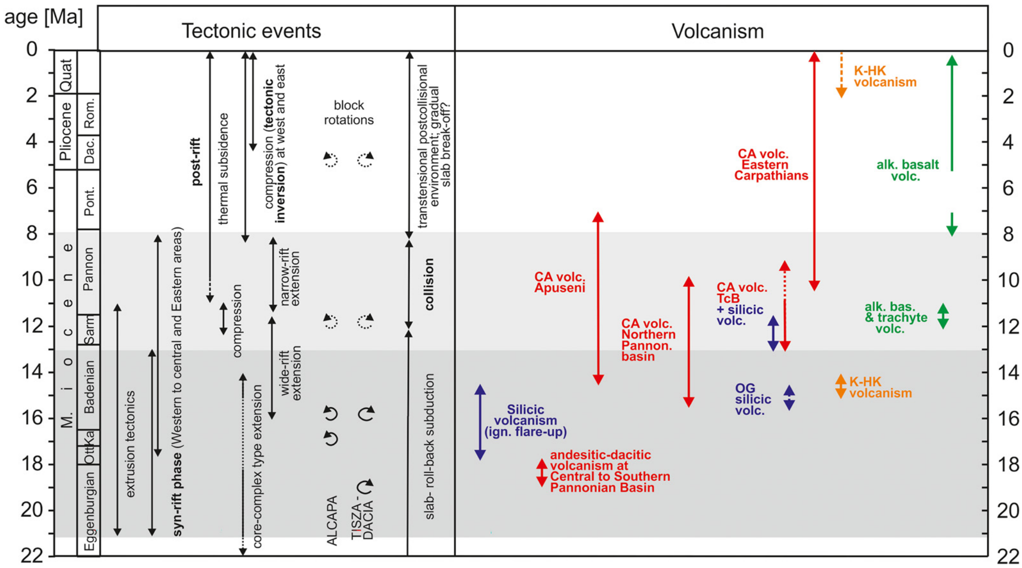


Fig. 5. Summary of temporal range of principal geodynamic events and the volcanism in the Carpathian–Pannonian Region for the last 22 Myr. alk. basalt, alkaline basalt; CA, calc-alkaline-type volcanic suites; ign. flare-up, ignimbrite flare-up; K–HK, potassic and ultrapotassic rocks; OG, Oaş–Gutâi; TcB, Transcarpathian Basin; volc., volcanism. Source: modified from the figure of Harangi (2001*b*) and Harangi and Lenkey (2007) using the results of Balogh *et al.* (1986, 1994*a*, *b*), Horváth (1993), Roşu *et al.* (1997, 2004); Fodor *et al.* (1999), Tari *et al.* (1999), Horváth *et al.* (2006*a*, 2015), Pécskay *et al.* (1995*a*, *b*, 2006), Márton *et al.* (2007), Seghedi and Downes (2011), Balázs *et al.* (2016), Lukács *et al.* (2018, 2021*a*, 2024), Seghedi *et al.* (2022) and Brlek *et al.* (2023) and further references found in the text.

beneath Dinarides could facilitate asthenospheric mantle flow towards the Pannonian Basin.

(5) Continental collision ('soft-collision'; Sperner *et al.* 2002; Bielik *et al.* 2004; Maţenco *et al.* 2007, 2010), which presumably occurred along the Eastern Carpathians around 11 Ma and between the ALCAPA–Tisza–Dacia microplate and the thicker continental blocks of the East European and Moesian platforms (Cloetingh *et al.* 2004).

(6) Thermal subsidence within the Pannonian Basin following the syn-rift phase resulting in deposition of sediments some several kilometres thick (Royden *et al.* 1983*a*, *b*).

(7) Tectonic inversion (Horváth and Cloetingh 1996; Bada *et al.* 2007; Dombrádi *et al.* 2010), starting around 8–9 Ma in the western regions and continuing in other areas over the last 5 million years. Considering this tectonic evolution scenario, it was becoming clear that the volcanism had apparently occurred in the wrong place and at the wrong time (Fig. 5), and this led to unconventional plate-tectonic explanations (e.g. Lexa and Konečny 1974; Harangi 2001*b*; Harangi and Lenkey 2007; Seghedi *et al.* 2011).

The plate-tectonic concept has provided a solid framework for geological interpretation over half a century. However, natural processes are often very complex and cannot be understood by applying only generalized models. In classical plate-tectonic theory, specific magma types are generally connected to particular tectonic settings, so that calc-alkaline volcanic suites are often interpreted to imply active, coeval subduction, and alkaline basalts are interpreted to reflect mantle plumes. The routine association between volcanic rocks and their geodynamic environment often leads to the use of terms such as 'orogenic' and 'anorogenic' volcanic rocks, which are mixture of genetic and descriptive terms. Cañón-Tapia and Walker (2004) noted that 'The theory of plate tectonics … adequately explains the distribution of volcanoes on Earth, but there are also many aspects of volcanism at the global scale that are difficult to understand using the plate tectonics model alone'. Thus, a different approach may be necessary to interpret volcanic activities in each tectonic setting. The composition of erupted magmas is determined by several factors, from the magma generation conditions (e.g. depth and degree of melting, nature of the magma source) to the wide range of magmatic differentiation processes in trans-lithospheric magma reservoirs and the final magma ascent and eruption mechanisms. Mantle source characteristics are of primary importance. Partial melting of lithospheric mantle that was metasomatized by subduction-related fluids in the past could produce very similar magmas to those formed in active subduction zones. Therefore, an apparent subduction-related geochemical fingerprint of volcanic rocks alone cannot be used to simply connect them to a plate tectonic setting; neither can the alkaline character of basalts automatically be connected to a mantle plume (Kempton *et al.* 1991; Bradshaw *et al.* 1993; Hawkesworth *et al.* 1995; Harangi *et al.* 2006; Harangi and Lenkey 2007; Seghedi and Downes 2011). Instead, origin of volcanism involving the reason of magma generation can be interpreted by a ‘from source to surface’ philosophy going down to crystal-scale information (Cañón-Tapia and Walker 2004; Davidson *et al.* 2007; Smith and Németh 2017) and involving further knowledge from geological and geophysical observations. The complexity of the Neogene–Quaternary volcanism and the geodynamic evolution of the CPR offers a challenging opportunity to apply this philosophy. Although, at first glance, the volcanic suites appear to be located in the wrong place and to have erupted at the wrong time (Fig. 5), this is only a matter of interpretation. In fact, they are in the right place and at the right time, and our task is to find the solution, which might depart from the seemingly simple classical plate tectonic interpretation.

Silicic volcanism

The Neogene–Quaternary volcanism of the CPR started with repeated violent explosive eruptions fed by SiO_2-rich magmas (Fig. 5 & 6). These caldera-forming events occurred in the interior of the Pannonian Basin and were coeval with the peak extension of the continental lithosphere. Large-volume pumiceous pyroclastic flow (ignimbrite) and

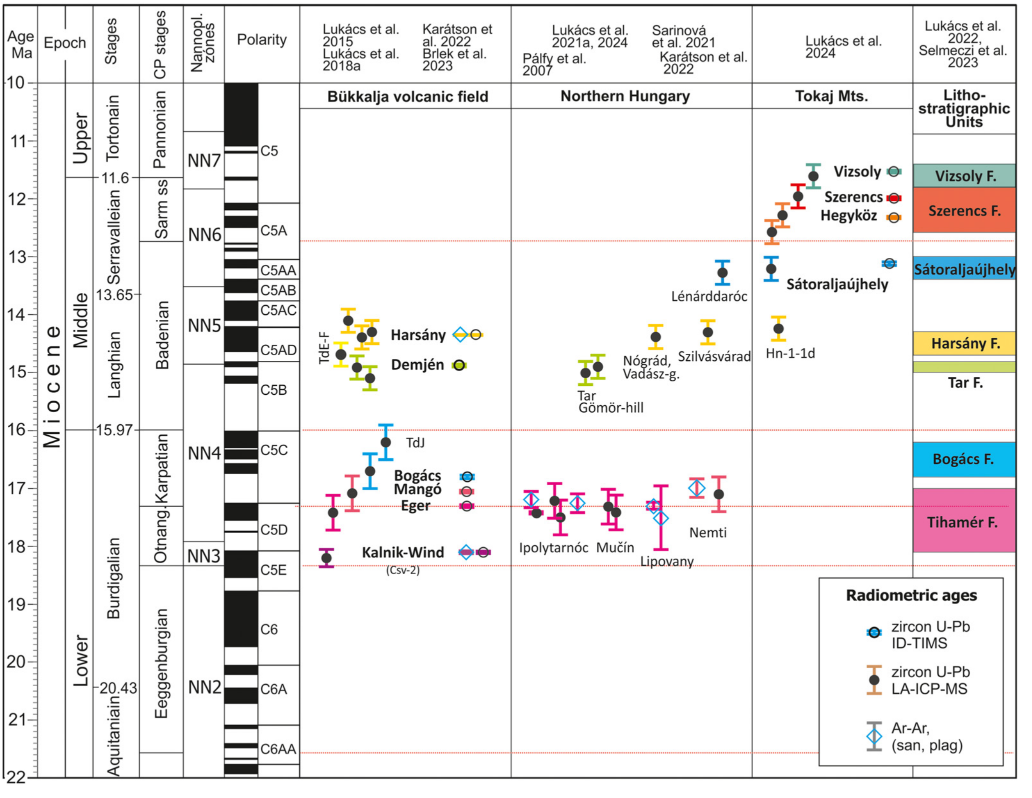


Fig. 6. Eruption chronology of the silicic volcanism in the Northern part of the Pannonian Basin (Fig. 2) based on high-precision zircon U–Pb dating and feldspar Ar–Ar dating (Pálfy *et al.* 2007; Lukács *et al.* 2015, 2018, 2021*a*, 2024; Šarinová *et al.* 2021; Karátson *et al.* 2022) and the defined lithostratigraphic units (Lukács *et al.* 2022; Selmeczi *et al.* 2024). CP, Central Paratethys; ID-TIMS, isotope dilution–thermal ionization mass spectrometry; LA-ICP-MS, laser ablation inductively-coupled plasma mass spectrometry. Source: magnetic polarity epochs are after Gee and Kent (2007); CP stages are after Piller *et al.* (2007) and Kováč *et al.* (2018).

accompanied pyroclastic fall deposits covered extensive areas in Europe. These pyroclastic horizons have great stratigraphic significance, noticed already a century ago (Noszky 1936; Ravasz 1987, Hámor *et al.* 1980), when they were classified into three main regionally widespread units, named Lower Rhyolite Tuff, Middle Rhyolite/Dacite Tuff and Upper Rhyolite Tuff. Their ages were determined by the K/Ar method (Hámor *et al.* 1980; Márton and Pécskay 1998) to be 19.6 ± 1.4 Ma, 16.4 ± 0.8 Ma and 13.7 ± 0.8 Ma, respectively, although the individual dates overlap each other within uncertainties and therefore cover a continuous age range between 21 and 13 Ma. Within this period, two major counterclockwise block rotation events were recognized in the ALCAPA domain shown by palaeomagnetic results (Márton *et al.* 2007) that helped to distinguish at least three main eruptive periods. In the last decade, new zircon U/Pb age data significantly revised the eruption chronology of the silicic volcanism as well as the timing of block rotations (Fig. 6; Lukács *et al.* 2015, 2018, 2021*a*). Lukács *et al.* (2018) identified at least eight major eruption events (eruption units, which involve closely spaced volcanic eruptions) from 18.1 to 14.4 Ma in the Bükkalja area, northern Hungary, which could have been in a proximal position to the eruption centres. This is a shorter period with more eruptive episodes than previously defined.

The pyroclastic products were classified into four main lithostratigraphic units (Fig. 6) following the ISG recommendations and decisions of the Hungarian Stratigraphic Commission (Lukács *et al.* 2022), whereas later, Hencz *et al.* (2024) proposed another informal lithostratigraphic scheme with more units. Zircon ages suggest that the first major 40–50° counterclockwise rotation occurred between 17.1 and 16.8 Ma, significantly later than previously thought (18.5–17.5 Ma; Márton *et al.* 2007), whereas the second, smaller (30°) rotation took place between 16.2 and 14.88 Ma. Lukács *et al.* (2021*a*) provided zircon fingerprints (U/Pb ages, trace-element and Hf isotope data) for each main eruptive unit, and these were successfully used to correlate the distal deposits with the proximal ones. Trace-element compositions of fresh glass shards and mineral compositional data were also used for tephra correlation (Harangi *et al.* 2005; Lukács *et al.* 2021*a*; Karátson *et al.* 2022; Hencz *et al.* 2024). The zircon-based methodology was extended to pyroclastic formations around and far from the CPR, and these studies showed that the 18.1, 17.3, 17.1, 14.9 and 14.4 Ma eruptions (eruption events with closely spaced eruptions) were extremely large and covered major parts of central and southern Europe with volcanic ash (Lukács *et al.* 2018, 2022; Rocholl *et al.* 2018; Brlek *et al.* 2020, 2023; Arp *et al.* 2021; Bercea *et al.* 2023; Šegvić *et al.* 2023; Trinajstić *et al.* 2023). Ongoing correlation studies using zircon and glass fingerprinting are helping to achieve a high-resolution chronostratigraphic framework for sedimentary sequences in the isolated Paratethys basins in the Mediterranean area and in the CPR.

The minimum volume of erupted material was estimated based on the evaluation of descriptions of several hundred boreholes in the Pannonian Basin, and as a result, Lukács *et al.* (2018) concluded that the cumulative tephra volume of the Early to Mid-Miocene silicic volcanism could be >4400 km^3. These large silicic eruptions produced even far-reaching pyroclastic flows (>100 km runout; Brlek *et al.* 2023), and some of them could yield several hundred cubic kilometres of tephra (Lukács *et al.* 2018; Karátson *et al.* 2022). Most of the volcanic deposits are buried by young sediments in the subsided sub-basins (e.g. Great Hungarian Plain; Fig. 2a; Pantó 1963; Széky-Fux and Kozák 1984), where drill cores revealed pyroclastic deposits >1000 m thick. Similarly, thick pyroclastic suites were interpreted along the Mid-Hungarian Shear Zone (Fig. 2a) by seismic sections and drill-core data (Csontos and Nagymarosy 1998; Horváth *et al.* 2015). In summary, this volcanism was an ignimbrite flare-up, the largest one in Europe in the last 20 Ma (Lukács *et al.* 2018). It is comparable with the secondary-to-tertiary magmatic pulses of ignimbrite flare-up episodes (De Silva *et al.* 2015) or Category 2 flare-up events (Gravley *et al.* 2016). Eruptions of silicic magmas occurred in two main periods: from 18.1 to 16.2 Ma and from 14.9 to 14.4 Ma, separated by a 1.3 ± 0.3 Myr gap (Fig. 6). This quiescence coincided with the onset of calc-alkaline intermediate volcanism in the northern part of the Pannonian Basin (Harangi *et al.* 2007), associated with graben formation along normal faults (Nemčok and Lexa 1990; Harangi *et al.* 2001). The peak extension stage in this area could have enabled fast evacuation of magmas, even from deep crustal reservoirs. Renewal of the silicic volcanism at 14.9 Ma took place in a changing tectonic environment, when the lithosphere was already thinned considerably.

Following the silicic ignimbrite flare-up, further significant silicic volcanism took place in the northern and northeast CPR. Rhyolite magmas extruded in the Ziar graben along with north–south-striking fault zones within the andesitic volcanic field of Central Slovakia (Fig. 2b) at approximately 12.2–11.4 Ma (Chernyshev *et al.* 2013). Large silicic explosive eruptions, accompanied by rhyolitic lava dome-forming and andesitic-to-dacitic volcanic periods, were also identified in the Tokaj Mts (Fig. 6; Szepesi *et al.* 2019, 2023; Lukács *et al.* 2024). The silicic explosive volcanism was characterized by at least four explosive eruption episodes between 13.1 and 11.6 Ma. The 13.1 and 12 Ma eruptions (called

Sátoraljaújhely and Szerencs eruptions, respectively) were remarkably large and resulted in ignimbrites and ash-fall deposits over extended areas (Lukács *et al.* 2024). Additional ignimbrite-forming rhyolitic eruptions occurred in the Oaş–Gutâi Volcanic Zone in the East Carpathians (Fig. 2b) at 15.4–14.8 Ma (K/Ar ages; Fülöp 2002; Kovacs *et al.* 2017). Szakács *et al.* (2012) described the Dej tuff complex in the Transylvanian Basin and proposed an eruption age of 14.8–15.1 Ma based on multiple dating techniques. On the other hand, de Leeuw *et al.* (2013) suggested an age of 14.38 $\pm$ 0.06 Ma by $^{40}Ar/^{39}Ar$ dating, which fits remarkably with the age of the Harsány ignimbrite (Lukács *et al.* 2018). Nevertheless, the eruption centre is still not yet well constrained, and further accurate dating is still needed.

In situ zircon U/Pb dates also allowed estimations of residence times of crustal magma reservoirs. They spread over several hundreds of thousands of years (300–700 kyr; Lukács *et al.* 2015, 2018, 2024), suggesting that large volumes of silicic magmas persisted in the upper crust for prolonged timescales. This is consistent with models of silicic magma reservoirs with long-term incremental growth by repeated recharge from below (Hildreth 2004; Lipman 2007). Remarkably, silicic magmas with distinct composition existed and evolved coevally (e.g. those of the Eger and Mangó eruptions and the Demjén and Harsány eruptions; Fig. 6) and presumably separately in the inner Pannonian Basin (Fig. 7). However, thermal maturation and shallowing of the brittle–ductile transition are prerequisites for emplacement of such voluminous magmas. This can be achieved by basaltic underplating and/or volcanism prior to the silicic eruptions (De Silva *et al.* 2015; Karakas *et al.* 2017). An extended andesite–dacite volcanism in the central-southern part of the Pannonian Basin during the onset of the syn-rift phase at 18–20 Ma (Fig. 5) could have been the forerunner of the silicic volcanic activity. Furthermore, significant thinning of the lithosphere could have been associated with melting of the metasomatized lower lithosphere owing to decompression or heating by passively upwelling asthenospheric mantle and production of basaltic magmas, which became stuck at the crust–mantle boundary (Harangi *et al.* 2001, 2007). These processes could all have contributed to the thermal maturation of the continental crust. The thermal impact of the silicic magma emplacement events for $>$ 4 Myr resulted in an anomalously high temperature in the crust and maturation of organic material (Horváth *et al.* 1988, 2015; Sachsenhofer *et al.* 1998). This is supported also by the partial or complete resetting shown by the apatite fission track and (U–Th)/He thermochronometer data in the Pannonian Basin (Dunkl and Frisch 2002; Danišík *et al.* 2015).

Petrological observations, trace-element composition of pumices and glass shards, and trace-element and isotopic composition of zircon all indicate that basaltic magmas played a significant role in magma evolution. Crystal fractionation with various amounts of crustal assimilation, mixing of mantle- and crustal-derived magmas and intermittent magma mixing/mingling in the upper crust were the dominant magmatic processes (Harangi *et al.* 2005, 2007; Harangi and Lenkey 2007; Lukács *et al.* 2009, 2018, 2024; Czuppon *et al.* 2012), whereas pure crustal anatexis (Póka 1988; Lexa and Konečný 1998; Póka *et al.* 1998; Chernyshev *et al.* 2013) seems to be implausible. Bulk rock geochemical data of pumices and the cognate lithoclasts suggest that a bimodal basalt/andesite–rhyolite volcanism (Fig. 8) characterized the Early to Mid-Miocene silicic volcanic episode (Lukács *et al.* 2005; Harangi and Lenkey 2007). The dominance of crystal-poor rhyolitic ignimbrites implies effective extraction of melt-dominated bodies from longstanding crystal mush reservoirs followed by rapid evacuation and eruption (Lukács *et al.* 2009, 2018, 2024). The Sr–Nd isotope data of the bulk pumices and glasses, and the zircon and glass Hf-isotope values (Figs 8 & 9; Lukács *et al.* 2018, 2024; Brlek *et al.* 2023) indicate that the initial rhyolitic magmas contained significant crustal components, but a major change has been detected at around 16.5 Ma, when a decrease in crustal magma involvement and/or an increase in asthenospheric mantle-derived magmas suggest that the peak extension was achieved, and the lithosphere became considerably thinned. Interestingly, the same Hf isotopic trend was found in the Tokaj Mts pyroclastic rocks (Lukács *et al.* 2024), where the change occurred after 12.3 Ma. This is consistent with a petrogenetic model in which mafic magmas were formed by melting in the metasomatized lithospheric mantle and later possibly even in the passively upwelling asthenospheric mantle.

Evolution of the Early to Mid-Miocene silicic volcanism of the Pannonian Basin can be regarded as a response to significant thinning of the lithosphere (Fig. 10) and shows many similarities with other rift-related silicic volcanic regions (Hildreth 1981) such as the Taupo Volcanic Zone (New Zealand) and the Basin and Range Province (USA). The silicic ignimbrite flare-up event in the interior of the Pannonian Basin could have been related to back-arc basin-type regional extension.

Andesite–dacite rhyolite volcanism

A prominent volcanic feature of the CPR is the chain of complex, dominantly andesitic to dacitic volcanoes along the Carpathian orogenic belt (Fig. 2). They are usually referred to as belonging to 'calc-alkaline volcanism', and as this type of

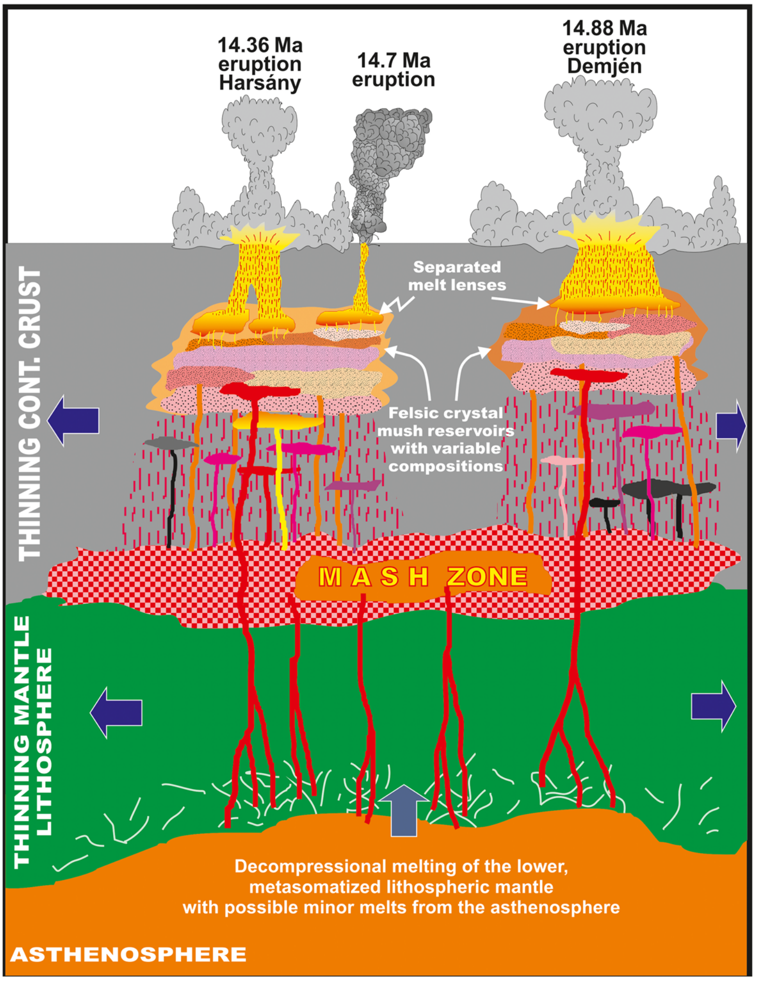


Fig. 7. Conceptual model for the architecture of the magma storage system during the early to mid-Miocene silicic volcanism as illustrated by the case of the 14.33–14.88 Ma eruption stages (Harsány and Demjén volcanic eruptions; Fig. 6). The silicic magma reservoirs could have been existing over several hundred thousand years before the eruptions and considerably affected the thermomechanical properties of the continental crust. Based on the extensive zircon *in situ* U–Pb dates and trace-element data (Lukács *et al.* 2015, 2018), the Demjén and Harsány silicic plutonic systems evolved partially coeval, although they had a distinct magma composition, so they were presumably spatially separated from one another. MASH zone: extensive, open-system magma reservoir developed at the crust–mantle boundary (Hildreth and Moorbath 1988). MASH, Melting, Assimilation, Storage and Homogenization.

volcanic rocks is typical of subduction zones, their origin is often connected to subduction (Bleahu *et al.* 1973; Boccaletti *et al.* 1973; Szabó *et al.* 1992; and further papers dealing with the tectonic evolution of the CPR). However, this is not entirely correct for several reasons. First, the 'calc-alkaline' term is generally used for a specific magma differentiation trend leading to silica-oversaturated rocks with gradually increasing SiO_2 and alkalis (Na_2O and K_2O), i.e. forming basalts through andesites and dacites to rhyolites (Sisson and Grove 1993; Grove *et al.* 2003). Nevertheless, this term goes back to the classification of volcanic rocks by Peacock (1931), who distinguished them based on the 'alkali-lime' index (i.e. based on the relative variation of $Na_2O + K_2O$ and CaO along with SiO_2). Later, Miyashiro (1974) divided the subalkaline rocks into tholeiitic and calc-alkaline suites based on the variation of FeO^{tot}/MgO ratio as a function of SiO_2, whereas Peccerillo and Taylor (1976) constructed a classification scheme based on the K_2O v. SiO_2 content. In summary, the meaning of the term

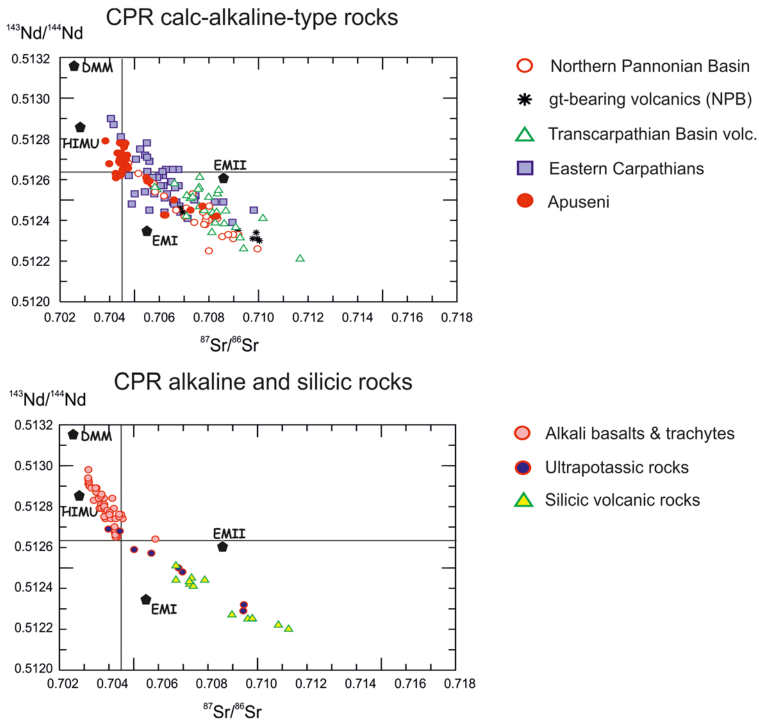


Fig. 8. Sr–Nd isotopic composition of the various volcanic rocks in the CPR. The wide isotopic variation can be explained by involvement of crustal components both from fluids metasomatized the lithospheric mantle source and from the lower crustal melts. CPR, Carpathian–Pannonian Region; DMM, depleted mid-ocean ridge basalt mantle; HIMU, high-µ mantle; EMI, enriched mantle I; EMII, enriched mantle II; NPB, Northern Pannonian Basin. Source: DMM, HIMU, EMI, and EMII components are from Zindler and Hart (1986). Carpathian–Pannonian Region data are from Salters *et al.* (1988), Embey-Isztin *et al.* (1993), Dobosi *et al.* (1995), Downes *et al.* (1995*a*), Harangi *et al.* (1995*a*, 1995*b*, 2001, 2007), Mason *et al.* (1996), Seghedi *et al.* (2001, 2007, 2008), Roşu *et al.* (2004), Harangi and Lenkey (2007) and Lukács *et al.* (2018).

'calc-alkaline' is not obvious (Sheth *et al.* 2002; Arculus 2003). There are different geochemical definitions, and the term is often used as a synonym of subduction-related volcanism. However, there are many examples of calc-alkaline-type volcanic rocks that were formed in a continental extension setting (e.g. Gans *et al.* 1989; Hawkesworth *et al.* 1995; Hooper *et al.* 1995; Morris and Hooper 1997; Morris *et al.* 2000; and also, many parts of the Mediterranean; Harangi *et al.* 2006). The specific 'subduction-type' geochemical features, such as enrichment of large ion lithophile elements (LILEs) and depletion in high-field strength elements (HFSEs), reflect the magma source characteristics, which can be inherited from an earlier subduction-related fluid metasomatism (Johnson *et al.* 1978; Cameron *et al.* 2003; Harangi *et al.* 2006). Therefore, it is important to distinguish between compositional or magma evolution character and the tectonic setting of the volcanism. Although there is confusion about the meaning of the term 'calc-alkaline', we retain it in this paper to describe the volcanism that produced andesitic and dacitic, up to rhyolitic rocks, since it is widely used in the literature, but with a strong emphasis that this is only a descriptive term.

Although the surficial occurrence of Neogene–Quaternary volcanic rocks with intermediate composition may suggest that there is a volcanic arc along the northern to eastern parts of the CPR, borehole data shade this apparent simple spatial distribution (Fig. 2). A vast amount of andesitic volcanic rocks, even complete volcanic structures (e.g. the 14.1 Ma Kráľová volcano in the Danube basin; Fig. 2b; Rybár *et al.* 2024), is present in the inner Pannonian Basin, buried by several hundred metres of young sediments deposited during the intense subsidence period of the Pannonian Basin (Pantó 1963; Széky-Fux and Kozák 1984; Zelenka *et al.* 2004; Petrik

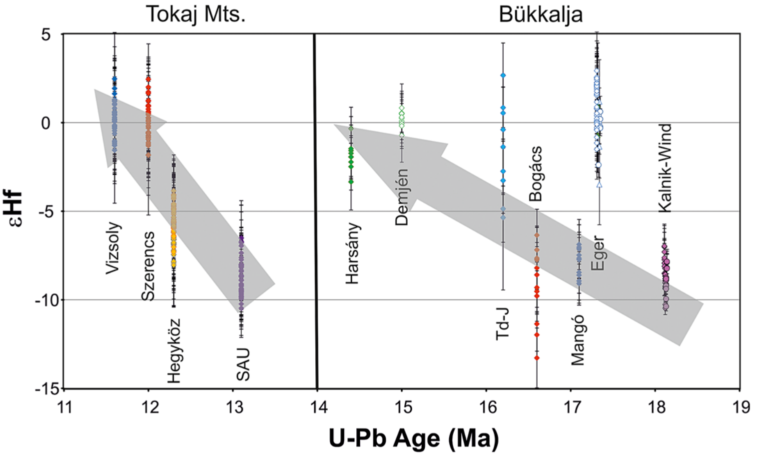


Fig. 9. εHf isotope (^{176}Hf/^{177}Hf isotope values relative to the bulk Earth Hf isotope value at the time of the volcanism; Bouvier *et al.* 2008) values of zircon from distinct eruptive units (Fig. 6) of the silicic volcanism in the Bükkalja and the Tokaj Mts (Fig. 2). In general, zircon from the older eruptive units has more negative εHf values, suggesting significant crustal components (either from lithospheric mantle metasomatized by fluids with crustal signature or from lower crustal melts mixed with mantle-derived magmas). However, the εHf value is higher in zircon from the younger eruption units in both areas. This can be explained by contemporaneous lithosphere thinning when a progressively larger fraction of magmas could come from the asthenospheric mantle, which is characterized by a higher εHf value, and the incorporated crustal melt fraction with lower εHf is decreasing. Source: data are from Lukács *et al.* (2018, 2024) and Brlek *et al.* (2023).

et al. 2019). Furthermore, Neogene calc-alkaline volcanic rocks in significant volume are found also in the Apuseni Mts (Roşu *et al.* 2004; Seghedi *et al.* 2007; Harris *et al.* 2013) and in a smaller amount in the Mecsek Mts (Árváné-Sós and Ravasz 1978). Following the pioneering work by Salters *et al.* (1988) and Downes *et al.* (1995*b*), detailed petrological and geochemical characterization of this volcanism for the last three decades has increased our understanding of the origin of the magmas and their relationships to tectonic processes (e.g. Mason *et al.* 1996; Harangi *et al.* 2001, 2006, 2007; Seghedi *et al.* 2001, 2004*a*, *b*, 2005*a*, 2023; Roşu *et al.* 2004; Kovacs *et al.* 2017, 2021; Rottier *et al.* 2020*a*, *b*). There are significant differences between the four segments of the calc-alkaline volcanic suites, i.e. the Northern Pannonian Basin, the volcanic chains around the Transcarpathian Basin, the Eastern Carpathians and the Apuseni Mts, so we describe them separately.

In the Northern Pannonian Basin (NPB; comprising the Central Slovakian Volcanic Area, the Visegrád–Börzsöny, Cserhát and Mátra; Figs 2 & 5), volcanism lasted from 16 to 9 Ma, followed by alkaline basaltic volcanism from 8 to 0.1 Ma (Balogh *et al.* 1981; Konečný *et al.* 1995*a*, 2002; Pécskay *et al.* 2006; Chernyshev *et al.* 2013; Rybár *et al.* 2024). Volcanoes developed well within the Pannonian Basin, where they are underlain by thin crust and lithosphere. They do not form a well-defined volcanic chain, but rather form an amalgamated cluster of volcanic complexes in an area of approximately 150 × 150 km. Structural and palaeomagnetic studies suggest that volcanism occurred in an extensional tectonic environment (Nemčok and Lexa 1990; Nemčok *et al.* 1998) with major block rotations (Karátson *et al.* 2000, 2007). A wide range of volcanic rocks were formed from basalt and basaltic andesite through the dominant andesite and dacite to rhyolites (Konečný *et al.* 1995*b*; Harangi *et al.* 2001, 2007; Karátson *et al.* 2001; Rottier *et al.* 2020*a*, *b*).

One of the peculiarities of the NPB volcanism is that almandine garnet is relatively common in the early (14–16 Ma) eruptive products (Embey-Isztin *et al.* 1985; Lantai 1991; Harangi *et al.* 2001; Rottier *et al.* 2020*b*; Bouloton 2021), whereas it is usually absent in calc-alkaline volcanic rocks worldwide. Most of the almandine crystallized from magma, although xenocrysts presumably from the granulitic lower crust can also be found (Harangi *et al.* 2001). Almandine is stable only at high pressure (>700 MPa; Green 1977, 1992; Alonso-Perez *et al.* 2009; Blatter *et al.* 2023), and so its presence in volcanic rocks implies rapid magma ascent from the lower crust without stopping in a shallower magma reservoir. Onset of the volcanism in this area coincides with peak extension of the Pannonian Basin, suggesting that the tensional stress field and rifting allowed the rapid ascent of the viscous, garnet-bearing magmas. Petrogenetic models based on Sr–Nd–Pb-O

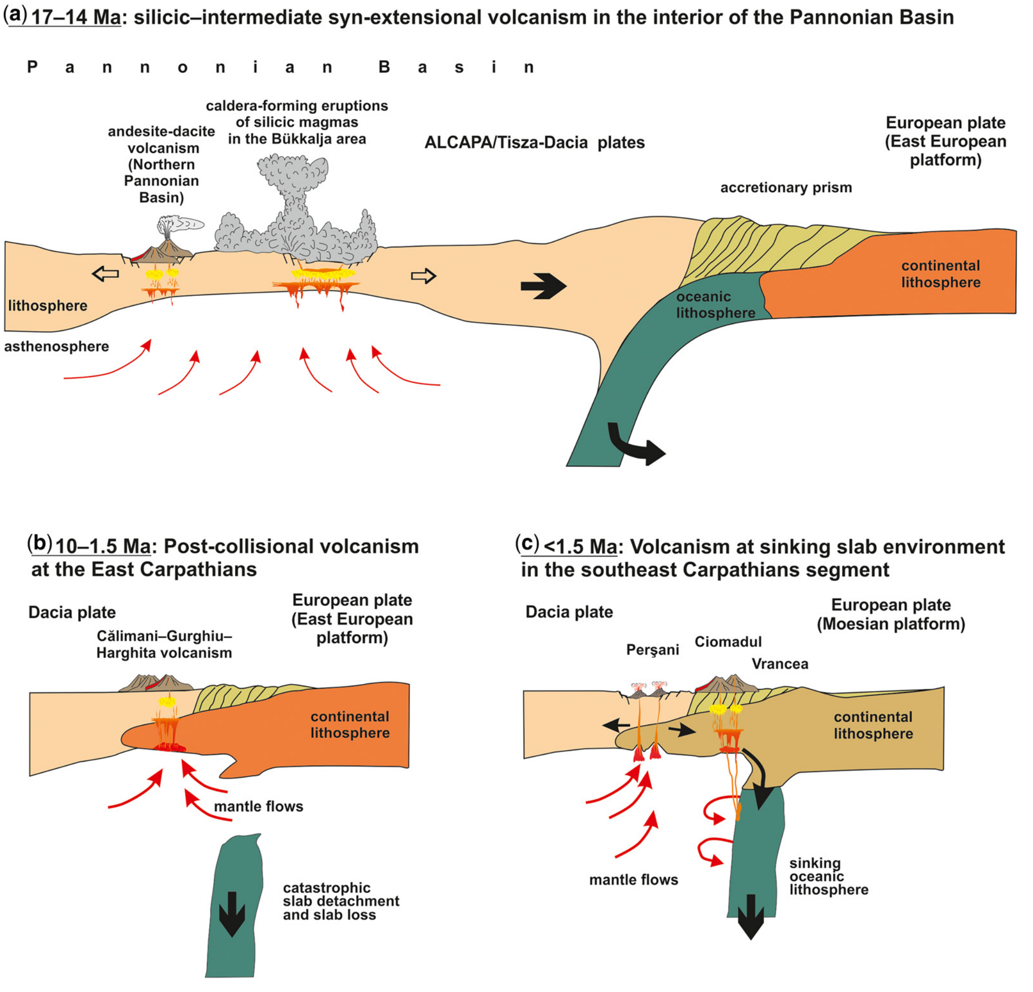


Fig. 10. Suggested models for the origin of the intermediate-to-silicic volcanism at various segments of the Carpathian–Pannonian region (idealized west–east cross-sections). **(a)** Large explosive eruptions of silicic magmas (Bükkalja; Fig. 2), later accompanied by andesite–dacite volcanism (Northern Pannonian Basin: Fig. 2) in the interior of the Pannonian Basin at 17–14 Ma. Magmas are generated by decompression melting of metasomatized lithospheric mantle owing to extension and heating by passively uprising asthenospheric mantle, and these mafic magmas were mixed with various amounts of crustal melts. (**b**) East Carpathians. A post-collisional volcanism took place from 10 to 1.5 Ma resulting in the Călimani–Gurghiu–Harghita volcanic chain (Fig. 2). This could have been controlled by gradual slab detachment and catastrophic slab loss causing asthenospheric mantle upwelling. This resulted in melting of the lower lithosphere metasomatized previously by subduction-related fluids. (**c**) Youngest volcanism of the Carpathian–Pannonian Region that has taken place at the southeast Carpathians for the last 1.5 Ma. Magma generation occurs both in the asthenosphere and in the metasomatized lower lithosphere owing to mantle flows initiated by the slab roll-back and sinking slab beneath the Vrancea zone. This resulted in the development of the Perşani alkali basalt volcanic field and the shoshonitic to adakitic volcanism of Ciomadul. ALCAPA: Alpine–Carpathian–Pannonian microplate. Source: the models were constructed using ideas of Lexa and Konečny (1974, 1998), Lexa *et al.* (1993), Mason *et al.* (1998), Harangi *et al.* (2001, 2006, 2007), Harangi (2001*b*), Seghedi *et al.* (2004*a*, 2011, 2023), Seghedi and Downes (2011) and Lukács *et al.* (2024).

isotopes suggest that mixing of mantle-derived magmas formed by melting of enriched, metasomatized lithosphere and magmas from the metasedimentary lower crust resulted in peraluminous magmas, where almandine could crystallize (Harangi *et al.* 2001, 2007). Ponding of large amounts of mafic melt beneath the crust could have caused lower crustal melting. Later, magma reservoirs stabilized in the heated upper part of the continental crust, where almandine was no longer stable. Although crystallization followed either a dry OPAM (olivine–plagioclase–augite–magnetite) trend or a wet evolution with amphibole and biotite, the variation of Dy/Yb ratio with SiO_2 content clearly suggests a significant role for amphibole crystallization in the lower crust (Fig. 11; 'amphibole-sponge'; Davidson *et al.* 2007). The presence of an amphibole-sponge is corroborated by the occurrence of amphibole megacrysts and xenoliths containing high-Al and high-Mg pargasitic amphibole in the andesites (Demény *et al.* 2012). The Štiavnica volcano represents one of the largest composite volcanic complexes in the Central Slovakian Volcanic Area (Fig. 2), where longstanding (>3 Myr) magma evolution occurred at shallow crustal depth with a presence of sulfide-saturated dacitic-to-rhyolitic melt (Rottier *et al.* 2020*a*). The magmatic reservoir was sulfide-saturated during its entire evolution, and this led to thorough porphyry mineralization with a contrasting Cu/Au ratio similarly to the nearby Javorie volcanic complex (Rottier *et al.* 2020*b*). A striking change in the magma source occurred at around 13 Ma (Harangi *et al.* 2007), when mafic magmas were generated mostly by melting of a heterogeneous asthenospheric mantle.

The subduction-related compositional character of the magmas can be explained by the nature of the lithospheric mantle source, metasomatized during previous subduction (Fig. 10; Harangi *et al.* 2001, 2006, 2007; Kovács *et al.* 2007; Kovács and Szabó 2008). This inherited LILE-enriched mantle region was reactivated and underwent partial melting as the lithosphere thinned at around 14–16 Ma. Composition of magmas changed in time, showing progressively less crustal contamination and more magma fractions from the upwelling asthenospheric mantle. Basalts and basaltic andesites formed around 9–10 Ma show a transitional chemical composition towards the later alkali basalts (Harangi *et al.* 2007), i.e. volcanism became fed almost entirely by asthenosphere-derived magmas. This scenario is consistent with a progressively thinning lithosphere and crust, and, as a result, passively uprising asthenospheric mantle, allowing the magma generation and volcanic eruptions for a protracted time, from *c.* 16 Ma to present in the same area (Harangi and Lenkey 2007; Harangi *et al.* 2007). Thus, the calc-alkaline volcanism in the NPB is considered as extension-related (Fig. 10) similarly to the Neogene syn-rift volcanism in the Basin and Range area, western USA (Gans *et al.* 1989; Fitton *et al.* 1991; Hawkesworth *et al.* 1995). Noteworthy, even at the advent of the plate tectonic models, Lexa and Konečny (1974) excluded the direct link between active subduction and the calc-alkaline volcanism in the NPB based on geological considerations and they suggested that this volcanic activity could be related to the back-arc extension of the Pannonian Basin (so-called 'areal-type' volcanism).

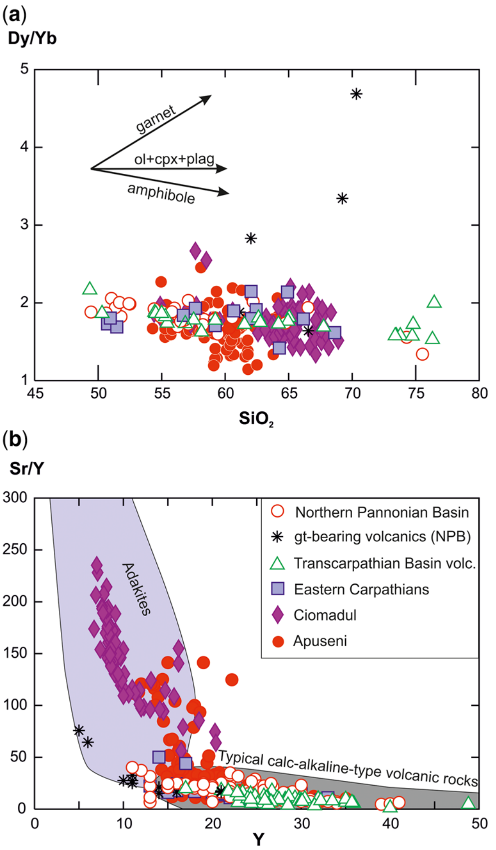


Fig. 11. Compositional features of the intermediate volcanic rocks of the Carpathian–Pannonian Region. (**a**) Initial garnet fractional control of the bulk composition of the magmas, followed by amphibole (wet magma condition; e.g. some of the volcanic suites in the Northern Pannonian Basin and Transcarpathian Basin; Fig. 2) and olivine + clinopyroxene + plagioclase (dry magma condition; e.g. East Carpathians; Fig. 2) fractionation trends. Mineral vectors are based on Rayleigh crystal fractionation model calculations. (**b**) Discrimination of volcanic rocks with adakitic composition from the typical calc-alkaline-type suites (Defant and Drummond 1990). The Ciomadul and Apuseni (Fig. 2) volcanic rocks have adakitic compositional characteristics. cpx, clinopyroxene; ol, olivine; plag, plagioclase; NPB, Northern Pannonian Basin. Source: Carpathian–Pannonian Region data are from Salters *et al.* (1988), Downes *et al.* (1995*b*), Harangi *et al.* (2001, 2007), Mason *et al.* (1996), Seghedi *et al.* (2001, 2007, 2008, 2022), Roşu *et al.* (2004), Harangi and Lenkey (2007), Kiss *et al.* (2010).

In the northeast segment of the Pannonian Basin, distinct volcanic zones surround the Transcarpathian Basin (Fig. 2): the southwest–northeast-oriented Slanske–Tokaj Mts, the northwest–southeast-oriented Vihorlat–Gutinski and the Oaş–Gutâi in the southeast (Kováč *et al.* 1995; Lexa and Kaličiak 2000; Soták *et al.* 2000; Seghedi *et al.* 2001; Seghedi and Downes 2011; Kovacs *et al.* 2017, 2021; Lukács *et al.* 2024). The linear chains of volcanoes on the surface are, however, apparent, since vast volumes of volcanic rocks are buried by young sediments at basins between Slanske–Tokaj and Vihorlat–Gutinski (Fig. 2; Vass 2002; Tschegg *et al.* 2019; Subová *et al.* 2022) and around Nyírség, the northeastern part of the Great Hungarian Plain (Fig. 2; Széky-Fux *et al.* 1987, 2007). Volcanism in each segment was confined mostly to 13.5–9 Ma (Fig. 5), although caldera-forming eruption of silicic magmas at Gutâi occurred slightly earlier (15.4–14.8 Ma; Fülöp 2002; Szakács *et al.* 2012; de Leeuw *et al.* 2013). Noteworthy, the volcanism in the Oaş–Gutâi was longstanding within a restricted area, lasting from 15.4 to 7 Ma (Kovacs *et al.* 2013). Large explosive events in the Tokaj Mts related to high-silicic magmas were described by Lukács *et al.* (2024), who distinguished four major explosive eruption periods. Among them, the 13.1 and 12.0 Ma eruptions (Fig. 6) were very large (>100 km^3 erupted material) and buried an extensive area with volcanic ash. Rhyolitic magmas extruded and caused small-scale explosive events between these large explosive events. They formed lava dome fields and produced rheomorphic ignimbrite in the Tokaj Mts and in the Beregovo area (Fig. 2; Seghedi *et al.* 2001; Kiss *et al.* 2010; Szepesi *et al.* 2019, 2023). Crystal fractionation was particularly extreme, which is rare worldwide (Lukács *et al.* 2024). Trace-element compositional features suggest that both the 'dry–reduced–hot' and the 'cold–wet–oxidized' rhyolitic groups (Bachmann and Bergantz 2008) are found in the Pannonian Basin, similarly to the Taupo zone (Deering *et al.* 2008, 2010) suggesting that melting occurred in mantle regions of various fluid content. The contribution of continental crust to magma evolution is reflected by the wide Sr–Nd–Pb isotopic values of the bulk rocks (Fig. 8; Salters *et al.* 1988) and the Hf isotope data of zircon (Fig. 9; Lukács *et al.* 2024). Some of the rhyolites have the highest $^{87}Sr/^{86}Sr$ isotope ratio (>0.711; Salters *et al.* 1988) in the CPR, corroborated by the strongly negative zircon εHf isotope values in the oldest silicic pyroclastic rocks (−10 to −8; Fig. 9; Lukács *et al.* 2024). This implies significant melting of the lower crust. The variation of isotope ratios indicates that the crustal contribution gradually decreased over time, and mantle-derived magmas became more ubiquitous. The Sr–Nd-O isotope ratios and trace-element compositions of the andesites and dacites also imply variable amounts of a crustal component in the magma genesis (Fig. 8; Seghedi *et al.* 2001; Kovacs *et al.* 2017), although this could also be a subducted sediment signature, as shown by the wide isotopic variation of Quaternary volcanic rocks in central Italy formed during lithospheric extension (Peccerillo 2005; Harangi *et al.* 2006). Detailed crystal-scale studies proved that complex trans-crustal magma storage systems were developed with a deep hot 'MASH' (Melting, Assimilation, Storage and Homogenization; Hildreth and Moorbath 1988) magma accumulation zone at the crust–mantle boundary, which were responsible for the protracted (>4–6 Myr long) volcanism within individual volcanic segments (e.g. Oaş–Gutâi; Kovacs *et al.* 2013, 2021).

The decrease in crustal contribution to the magmas and the increase in mantle melting are in accord with a progressively thinning lithosphere. In fact, this volcanic area is still underlain by an anomalously thin crust and lithosphere (25–27 km and around 80 km, respectively; Horváth 1993; Lenkey *et al.* 2002; Horváth *et al.* 2006*a*, 2015; Kalmár *et al.* 2021, 2023), and it is associated with a relatively high heat flow (>100 mW/m^2; Dövényi and Horváth 1988; Lenkey *et al.* 2002; Horváth *et al.* 2015). However, such a setting seems surprising, since the volcanic areas are situated very close to, and even partly overlapping, the thin-skinned fold-and-thrust belt, which represents the accretionary prism developed at the convergence zone between the East European Platform and the ALCAPA–Tisza–Dacia domain (Fig. 2a). The age of volcanism overlaps the time of cessation of contraction in the external front of the wedge at around 12 Ma (Nakapelyukh *et al.* 2018; Roger *et al.* 2022). Thus, the andesitic-to-rhyolitic volcanism occurred partly at the final stage of subduction, when the subducted slab became almost vertical after continuous roll-back subduction retreat (Royden *et al.* 1982, 1983*a*, *b*). This exerted a pull in the upper lithosphere, which led to thinning and accelerated subsidence of the Transcarpathian Basin (Fig. 2), and induced asthenospheric mantle flow. Lithosphere foundering (e.g. Ducea *et al.* 2013; Murray *et al.* 2015) is another process that may have resulted in the longstanding magmatism, particularly in the Oaş–Gutâi volcanic segment (Kovacs *et al.* 2021). These geodynamic conditions could lead to massive magma generation both in the lower lithosphere and in the upper asthenosphere. The subduction-related compositional character of the magmas was inherited from the formerly metasomatized lithospheric mantle reactivated by decompression melting and/or heating by the upwelling asthenospheric mantle.

The volcanic chain of Călimani–Gurghiu–Harghita (CGH) is approximately 160 km long and runs parallel with the Carpathian orogen zone from northwest to southeast (Fig. 2; Mason *et al.* 1995, 1996; Seghedi *et al.* 1995). A subvolcanic district (Rodna-Bârgău) was developed at the northern margin of the Călimani volcanic complex (Fedele *et al.* 2016), where magmatism slightly preceded and partially overlapped (11.5–8 Ma) the onset of the Călimani volcanism. Here, the presence of garnet-bearing andesites/dacites argues for lower crustal involvement reflected by a transtensional stage (Gröger *et al.* 2008; Tischler *et al.* 2008). Within the CGH chain, each volcanic complex was relatively short-lived, and a gradual age progression in the volcanic activity is observed southeastward from 10 to 0.03 Ma (Fig. 5; Pécskay *et al.* 1995*b*, 2006). The youngest eruptions occurred between 1 and 0.03 Ma in Ciomadul (Szakács *et al.* 1993, 2015; Harangi *et al.* 2010, 2015*b*, 2020; Molnár *et al.* 2018, 2019). This area may still be potentially active, as shown by indications of a magma body beneath the volcano at the present day (Popa *et al.* 2012; Szakács and Seghedi 2013; Harangi *et al.* 2015*c*). Volcanism in Ciomadul was partially contemporaneous with development of the alkali basalt volcanic field at Perşani (Downes *et al.* 1995*a*; Seghedi *et al.* 2016). These volcanic activities took place close to (*c.* 100 km distance) the seismically active Vrancea zone, where a near-vertical lithospheric slab is still descending (Fig. 10; Roman 1970; Wenzel *et al.* 1998; Ismail-Zadeh *et al.* 2012). Thus, the area is geodynamically still active and therefore poses both seismic and potentially volcanic hazards (Radulian *et al.* 2004; Szakács and Seghedi 2013).

Mason *et al.* (1996) evaluated the petrogenesis of the basalt-andesite–dacite–rhyolite volcanic succession of the CGH based on Sr–Nd–Pb-O isotope (Fig. 8) and trace-element data and concluded that crystal fractionation combined with assimilation of variable amounts of crustal material was responsible for the origin of these volcanic rocks. Melt inclusions in olivine and pyroxene revealed that the primary magmas were generated by variable degree (2–12%) of melting of the lithospheric mantle metasomatized by ~2% sediment-derived melts (Bracco-Gartner *et al.* 2020). Mafic K-alkaline melts in Southern Harghita originated from a melt- and fluid-metasomatized lithospheric mantle containing amphibole (±phlogopite) by ~5% melting. The youngest volcanism at Ciomadul has been studied using crystal-scale approaches (Kiss *et al.* 2014; Harangi *et al.* 2015*c*; Laumonier *et al.* 2019; Lukács *et al.* 2021*b*; Cserép *et al.* 2023) demonstrating that a magma reservoir system has been present beneath the volcano for approximately 1.5 Myr. This comprises a deep hot zone of accumulation of mafic magmas at the crust–mantle boundary and a shallower (8–20 km depth) felsic, low-temperature (670–780°C) strongly crystalline magma storage region. Its prolonged existence can be explained by an average magma flux of 1.3×10^{-4} km^3/yr (Lukács *et al.* 2021*b*). Volcanism has been characterized by long-quiescence periods (several tens of thousands of years) and reactivation initiated by high-temperature (>900 °C) more mafic recharge magmas (Kiss *et al.* 2014; Cserép *et al.* 2023).

Most of the volcanism of the CGH postdates the presumed subduction along the East Carpathians and also the continental collision, which occurred at around 11 Ma (Maţenco *et al.* 2007, 2010). Thus, this volcanism is considered post-collisional (Fig. 10; Seghedi *et al.* 2011, 2019), characterized by features such as (a) absence of primitive volcanic rocks; (b) relatively narrow magma evolution from (basaltic) andesite to dacite; (c) variable, small-volume, complex open-system evolution, implying interaction with lower and possibly, upper crustal material; and (d) evidence for large, shallow subcontinental mantle heterogeneities. During the youngest volcanism, two marked changes in magma composition occurred in the southern segment of Harghita at 2 and 1 Ma (Seghedi *et al.* 2023). They correspond with the Trotuş Fault Zone, which is a boundary between two major crustal blocks (the East European Platform at the north and the Moesian platform in the south; Fig. 2), which have distinct thermomechanical properties (Cloetingh *et al.* 2004). After 1 Ma, the erupted products had adakitic compositional features, i.e. elevated Ba and Sr and depleted heavy REE and Y content leading to high Sr/Y and La/Yb ratios (Fig. 11; Szakács *et al.* 1993; Seghedi *et al.* 2004*a*, 2011, 2023; Molnár *et al.* 2018, 2019). Magmas with such compositional characters can originate by various processes, including melting of a subducted oceanic slab, partial melting of basaltic lower crust metamorphosed to eclogite and garnet-amphibolite, fractional crystallization combined with crustal assimilation and magma mixing (Defant and Drummond 1990; Martin *et al.* 2005; Castillo 2006; Wang *et al.* 2020; Zhang *et al.* 2021). The Ciomadul magmas are typically wet and oxidized, and even the recharge magmas are thought to be extremely hydrous (Cserép *et al.* 2023), requiring significant water in the magma source. Slab melting could be a potential explanation, assuming that a remnant oceanic slab is descending beneath Vrancea, and the intermediate depth earthquakes are caused by dehydration in the slab (Ferrand and Manea 2021). North of the Trotuş tectonic zone, no descending slab is detected in the upper mantle (Wortel and Spakman 2000), which can be explained by catastrophic slab loss. In South Harghita, slab detachment at shallow depth could have resulted in upwelling of asthenospheric mantle, which caused partial melting in the metasomatized lithospheric mantle leading to short-lived volcanism (Fig. 10; Seghedi *et al.* 2023). Gradual slab detachment from northwest to southeast was postulated also by Mason *et al.* (1998) and Seghedi *et al.* (1998), who suggested that slab break-off occurred progressively at shallower depth. Hot asthenospheric mantle filled the void left behind by the broken slab, increasing efficiency of dehydration of the slab remnants, one of which might occur now at the Vrancea zone. This model explains well the progressively younger volcanism along the CGH.

The Apuseni Mts (Fig. 2) is composed of nappes of pre-Neogene formations at the contact of the Tisza and Dacia units (Fig. 2; Dallmeyer *et al.* 1999; Pană *et al.* 2002; Schmid *et al.* 2008; Balintoni *et al.* 2010). Here, a persistently long volcanic activity in a relatively restricted area started around 14.5 Ma and lasted until 7 Ma (Fig. 5; Roşu *et al.* 1997, 2004; Seghedi *et al.* 2022). Most of the volcanic activity was confined to northwest–southeast-oriented grabens. Notable mineralization was associated with the volcanism (Neubauer *et al.* 2005; Manske *et al.* 2006; Harris *et al.* 2013), where the youngest basaltic andesite eruption occurred at 7.2–8 Ma at Detunata; Roşu *et al.* 1997, 2004). New zircon U–Pb dating suggests an age of 8.52 Ma for Detunata and a younger age (7.33 Ma) for a nearby andesite (Ene *et al.* 2024). Magmas during the initial volcanic activity (14.5–12 Ma) evolved via a typical calc-alkaline differentiation trend, whereas garnet-bearing dacite magmas could have had a lower crustal origin (Seghedi *et al.* 2022). After 12 Ma, the magma composition changed significantly. The younger volcanic rocks are characterized by lower $^{87}Sr/^{86}Sr_o$ and $\delta^{18}O$ isotope values and have high Sr/Y ratios, typical of adakitic magmas (Fig. 11; Defant and Drummond 1990; Martin *et al.* 2005; Seghedi *et al.* 2007; Castillo 2012; Wang *et al.* 2020; Zhang *et al.* 2021).

The Apuseni volcanic rocks contain high LILE abundance and depletion in HFSE, features generally attributed to flux melting in the mantle wedge above a subducted slab. Linzer (1996) suggested that subduction before slab break-off along the East Carpathians might be responsible for the calc-alkaline-type volcanism in the Apuseni; however, Seghedi *et al.* (2007, 2022) and Harris *et al.* (2013) pointed out that this could not be the case. The volcanism occurred far (*c.* 200 km) from the assumed subduction zone along the East Carpathians. Furthermore, the youngest, <12 Ma volcanic phase took place simultaneously with the post-collisional stage in that segment (Maţenco and Bertotti 2000). During the Miocene, complex extensional graben structures were formed along with considerable thinning of the lithosphere (Tari *et al.* 1999, 2024; Balázs *et al.* 2017). Major block rotations were contemporaneous with the older volcanic phase (Roşu *et al.* 2004). As a result, most authors (Roşu *et al.* 2004; Neubauer *et al.* 2005; Seghedi *et al.* 2007, 2022; Harris *et al.* 2013) consider the volcanic activity in Apuseni as extension-related, where decompression melting of the metasomatized lithospheric mantle and the upwelling asthenospheric mantle led to the formation of primary magmas. An extensive MASH zone (Hildreth and Moorbath 1988) could have been formed at the crust–mantle boundary, and this fed the prolonged volcanic activity. Emplacement of large amount of mafic magma beneath the crust could also initiate melting in the lower crust. The >12 Ma calc-alkaline volcanic suite was formed by crystal fractionation accompanied by variable amounts of crustal contamination. The subsequent, <12 Ma volcanism signifies a major change in magma composition, being adakitic in character. Seghedi *et al.* (2022) attributed this to a larger degree of partial melting of wet basaltic (amphibolitic) lower crust, combined with early crystallization of amphibole.

Potassic to ultrapotassic volcanism

Potassic volcanic rocks (ultrapotassic if $K_2O/Na_2O>2$ and $K_2O>3$ wt%, or potassic if $K_2O/Na_2O = 1–2$ and $K_2O>3$ wt%; Foley *et al.* 1987) are relatively rare. They are most common in cratonic continental regions, but also occur in orogenic areas and even in continental rift zones. A type-area is central Italy (Peccerillo 2005), where Washington (1906) recognized the association of unusual rock types variably rich in potassium. The high K_2O content is usually associated with a strongly enriched trace-element character, attributed to the magma source region, i.e. phlogopite- or amphibole-bearing metasomatized lithospheric mantle and low degrees of melting. The metasomatic agents are usually related to subducted sediments, although McKenzie (1989) suggested that very-low-volume melts from the asthenosphere could react with peridotite in the colder lower lithosphere. The extremely high $^{87}Sr/^{86}Sr$ and low $^{143}Nd/^{144}Nd$ isotope ratios reflect the old ages of the sedimentary component and/or ancient metasomatism. Remobilization of such veined lithospheric mantle can take place by decompression melting or heating by the asthenospheric mantle (Thompson *et al.* 1990). Thus, the occurrence of these rock types is an indication of melting in the lithospheric mantle metasomatized by fluids with crustal components and either extension or mantle plume heating.

In the CPR (Fig. 2), ultrapotassic rocks were first recognized by Harangi *et al.* (1995*a*), who revised the former description of the 2.1 Ma mafic rock at Bár, southern Hungary (previously classified as potassic basalt by Szederkényi 1980) and the volcanic rocks in drill cores of the Balatonmária-1 borehole (described as andesite). The former was termed olivine–leucitite, whereas the latter was termed latite, both having an ultrapotassic character.

Furthermore, Harangi *et al.* (1995*a*) pointed out that the mid-Miocene potassic trachyandesite (latite) in the Styrian Basin (Gleichenberg; Scharbert *et al.* 1981; Ebner and Sachsenhofer 1991) also has an ultrapotassic character and shows similarities to the Balatonmária rocks. Another outburst of potassic–ultrapotassic magmas took place at the same time, when shoshonitic rocks were formed at Loncarski Vis and Mt. Krndija in the southern Pannonian Basin (Pamić *et al.* 1992, 1995).

Another ultrapotassic volcanic rock type in the CPR is the 1.32 Ma lamproite, first recognized by Seghedi *et al.* (2008), situated 5 km south from Gătaia (western Romania). Further potassic volcanic formations (often called shoshonite) are found in the south-southeast CPR, such as the 1.6 Ma Mt Uroi (Figs 2b & 5; Roşu *et al.* 2004; Ene *et al.* 2024) in the southern Apuseni Mts, and two 0.9 Ma subvolcanic bodies in the Ciomadul dome field (Fig. 2b; Seghedi *et al.* 1987; Szakács *et al.* 1993, 2015; Mason *et al.* 1996; Molnár *et al.* 2018; Bracco-Gartner *et al.* 2020). The occurrence of isolated Quaternary potassic volcanic centres along a west–east trend in the southern CPR is noteworthy, although no clear explanation has been put forward about it yet.

The ultrapotassic and potassic rocks of the CPR show a wide range of petrological and geochemical characters (Fig. 8). Both the silica-undersaturated Bár olivine–leucitite and the Gătaia lamproite are primitive, having a high Mg number (>0.7) and containing high-Mg olivine and clinopyroxene (Harangi *et al.* 1995*a*; Seghedi *et al.* 2008). They have remarkably similar Sr–Nd isotope ratios. Leucitite is a relatively rare rock type, but it can be found in many parts of Europe (Lustrino *et al.* 2019). Lamproites are even rarer, but are known from central Italy and SE Spain (Venturelli *et al.* 1988; Conticelli and Peccerillo 1992; Peccerillo 2005; Seghedi *et al.* 2007; Prelević *et al.* 2007). The Balatonmária and Gleichenberg latites are silica-saturated, having sanidine instead of leucite (Harangi *et al.* 1995*a*). They have higher $^{87}Sr/^{86}Sr_o$ ratios (Fig. 8) and a more pronounced negative Nb anomaly than the leucitite and lamproite. They contain abundant clinopyroxene, which shows remarkable zoning patterns. The complexity of compositional zoning in clinopyroxene, together with mica phenocrysts, provides evidence of magma mixing and fractional crystallization (Klébesz *et al.* 2009). The source region of the ultrapotassic primary magmas of the CPR is inferred to have been a strongly depleted (garnet–harzburgite) lithospheric mantle veined by clinopyroxene, mica and fluorapatite, metasomatized either by an alkaline melt (leucititic and lamproitic magmas) or by a subduction-related fluid (parental magma of latites) (Harangi *et al.* 1995*a*; Seghedi *et al.* 2008). The strongly metasomatized lithospheric mantle played a crucial role in the origin of other potassic magmas in the CPR (Roşu *et al.* 2004; Bracco-Gartner *et al.* 2020).

The occurrence of ultrapotassic and potassic rocks is a clear sign that the lower lithosphere beneath the Pannonian Basin is thoroughly metasomatized. Furthermore, it provides unambiguous evidence for the presence of mantle lithosphere in contrast to the model presented by Horváth *et al.* (2006*a*), who considered that a thick crustal block without lithospheric mantle came from the southern Alpine compressive region and translated towards the east. Other evidence that the ALCAPA block also comprised a lower lithospheric mantle part is the abundance of various peridotite xenoliths in the alkali basalts (Embey-Isztin *et al.* 1989, 1990, 2001; Downes *et al.* 1992; Konečný *et al.* 1995*c*; Bali *et al.* 2002, 2008; Szabó *et al.* 1995, 2004). Kovács *et al.* (2012*b*) pointed out that most of these peridotites should belong to the ALCAPA lithosphere for a prolonged time. Thus, the escaping ALCAPA was a lithospheric block rather than a crustal block. The metasomatic veins contain various mineral assemblages with amphibole and phlogopite. Phlogopite-bearing xenoliths were found in the western Pannonian Basin (e.g. Szabó *et al.* 1995). Such veined peridotites have a lower solidus owing to the volatile-bearing phase capable of melting in case of decompression (lithosphere thinning as occurred in the Mid-Miocene in the CPR) or heating by an upwelling asthenosphere (as inferred for the potassic–ultrapotassic magma generation in the southern margin of the CPR). Such melting events appear to have been localized and temporally scattered (from 2.1 to 0.9 Ma within the Quaternary), processes that are not entirely understood, although better knowledge is needed because of the unknown potential hazard.

Alkaline basalt volcanism

Alkaline basaltic volcanic activity began 12 Myr ago, culminating at 2–5 Ma (Figs 2 & 5; Balogh *et al.* 1986, 1990, 1994*a*, *b*), and the last eruption of basaltic magma was the Putikov volcano in the NPB, only 100 000 (Šimon and Halouzka 1996; Šimon and Maglay 2005) or 500 000 years ago (Balogh *et al.* 1981; Chernyshev *et al.* 2013). This long-lasting volcanism postdates the calc-alkaline-type volcanic activity in the NPB and was contemporaneous with development of the Călimani–Gurghiu–Harghita volcanic chain in the east (Fig. 5). It produced several monogenetic volcanic fields (Fig. 2; Martin and Németh 2004; Harangi *et al.* 2015*a*), each active for several million years. However, eruption phases were separated by prolonged repose times, often >1 Myr. The youngest basalt volcanic field is in the southeast CPR

(Perşani), close to Ciomadul and the Vrancea zone, i.e. in a geodynamically active area. It was active between 1.3 and 0.6 Ma, with long repose periods (Panaiotu *et al.* 2013; Seghedi *et al.* 2016). All the other basalt volcanic fields are in the western (Styrian, Burgenland–Little Hungarian Plain and Bakony–Balaton Uplands) and northern (Novohrad/Nógrád-Gemer) margins of the Pannonian Basin (except for the buried volcanic field in the western part of the Great Hungarian Plain; Fig. 2). The last basaltic eruption in Putikov occurred *c.* 5–7 Ma after the previous volcanic events in that area (Konečný *et al.* 1999; Chernyshev *et al.* 2013). In the southern CPR, basaltic magma erupted near Lucareţ–Şanoviţa at 2.5 Ma (Pécskay *et al.* 2006) without any precursor volcanism. This isolated volcanic event is similar to the formation of single, short-lived Quaternary volcanic eruptions of potassic and ultrapotassic magmas (Bár, Gătaia, Mt. Uroi; Fig. 2b) along the southern margin of the Pannonian Basin. Although this type of volcanic activity was predominantly basaltic, an 11–11.5 Ma alkaline trachyte composite volcano is present in the sub-surface beneath the Little Hungarian Plain (Pásztori; Fig. 2b; Harangi *et al.* 1995*b*; Harangi 2001*a*; Pánisová *et al.* 2018). Trace-element and isotopic data of the trachytes clearly indicate that they were formed by crystal fractionation from a basaltic parental magma without crustal contamination. This means that large volumes of basaltic cumulate material may reside in the deep crust (Harangi *et al.* 1995*b*; Harangi 2001*a*), as also inferred from 3D geophysical models (Pánisová *et al.* 2018). Small volumes of basaltic magma also reached the surface or formed dykes in the volcano. They show geochemical similarities with the 11.5 Ma basalts in Burgenland. Trachytic volcanoes are inferred to have formed also in the central Pannonian Basin, based on seismic section interpretations and drill core from boreholes in the western part of the Great Hungarian Plain (Balázs and Nusszer 1987).

The major- and trace-element compositions of the alkaline basalts (Embey-Isztin *et al.* 1993, 2001; Dobosi *et al.* 1995; Embey-Isztin and Dobosi 1995; Downes *et al.* 1995*a*; Harangi *et al.* 1995*b*, 2013, 2015*a*; Harangi 2001*a*; Seghedi *et al.* 2004*b*; Ali and Ntaflos 2011; Ali *et al.* 2013) indicate that their parental magmas were silica-undersaturated, ranging from nephelinite, through basanite to trachybasalt and alkali basalt. Most have primitive compositions (Mg-value >0.65), showing only limited olivine fractionation. Magma ascent calculations (Harangi *et al.* 2013; Jankovics *et al.* 2013) suggest that the mafic magmas ascended rapidly through the crust and reached the surface in a few days from the upper mantle. In other cases, the basaltic magmas accumulated at the crust–mantle boundary, where successive magma batches mixed and formed a heterogeneous crystal cargo (Jankovics *et al.* 2013, 2016, 2019; Harangi *et al.* 2015*a*). Crystal-scale studies, particularly the variation in olivine and spinel compositions (Fig. 12), revealed that the erupted basaltic magmas are often composed of distinct magma fractions coming from heterogeneous mantle sources.

Each monogenetic eruption in a volcanic field corresponds to a unique magma generation event in the upper mantle and the compositional diversity of basaltic magmas implies that the magma batches were formed by various degrees of partial melting and from distinct mantle sources (Figs 13 and 14a). Their Sr–Nd–Pb isotope ratios reflect the

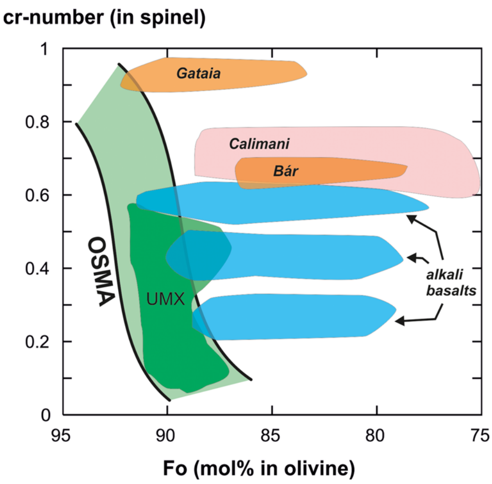


Fig. 12. Relationship between the host olivine and spinel inclusion compositions as expressed by the Fo end-member of olivine (mol%) and cr-number of spinel in various volcanic rocks of the Carpathian–Pannonian Region. The cr-number of spinel reflects the nature (primarily the degree of depletion) of the mantle source region, i.e. a higher cr-number suggests more depleted peridotites. Spinels from the Gătaia lamproite (Seghedi *et al.* 2008) and the Bár leucitites (Harangi *et al.* 1995*a*) as well as the Călimani basaltic andesite (Bracco-Gartner *et al.* 2020) suggest depleted harzburgitic mantle source metasomatized by various fluids. However, the high cr-number of spinel could indicate also melting of pyroxenite lithology (Laukert *et al.* 2014). The remarkably large variation in the spinel composition from the alkali basalts (Harangi *et al.* 2013, 2015*a*; Jankovics *et al.* 2019), even in single samples, indicates a heterogeneous mantle source possibly with involvement of pyroxenitic/eclogitic lithology, as shown also by the trace-element petrogenetic model in Figure 13. Fo, forsterite; OSMA, Olivine-Spinel Mantle Array (Arai 1994); UMX, olivine–spinel compositions in peridotite xenoliths in the Pannonian Basin (Downes *et al.* 1992; Szabó *et al.* 1995; Patkó *et al.* 2024).

compositional heterogeneity of the source (Fig. 8), since these isotope ratios do not change during melting and crystallization, and crustal contamination was limited or absent. Mafic magmas transported various crustal fragments to the surface described mostly in the Novohrad/Nógrád-Gemer Volcanic Field (Huraiová *et al.* 1996, 2005, 2017). They had no detectable effect on the chemical composition of the host magma (Dobosi *et al.* 1995) but provided insights into the magma–crust interaction at depth. Thus, the differences in isotope ratios indicate compositional heterogeneity in the mantle (Lustrino and Wilson 2007). Wilson and Downes (1991, 2006) found similar isotopic variation in the Neogene–Quaternary basalts of Europe and proposed that this reflects variable contributions from the asthenospheric and lithospheric mantle similar to the Basin and Range Province, USA (e.g. Fitton *et al.* 1991; Ormerod *et al.* 1991). Lithospheric mantle is rather refractory, but its melting point is depressed if amphibole-rich veins are present (Pilet *et al.* 2008; Mayer *et al.* 2013, 2014). Melting in the presence of amphibole can be quantitatively modelled. Residual amphibole during melting results in a negative K-anomaly, whereas consumption of amphibole causes high K_2O in the basalt composition. In the Pannonian Basin, distinct groups of basalts can be distinguished, one of which is indeed characterized by a negative K-anomaly in normalized trace-element plots (Harangi and Lenkey 2007; Harangi *et al.* 2015*a*). However, Harangi *et al.* (2015*a*) pointed out that the K-anomaly correlates with several trace-element and isotopic ratios and may be a source feature rather than indicating the presence of amphibole during magma generation.

The large compositional diversity of basalts reflects small, variable degrees of partial melting of upper mantle rocks (Fig. 13), probably entirely in the asthenosphere. The reason of melt generation is, however, somewhat elusive, since it occurred well after the main extensional phase of the Pannonian Basin. A possible explanation for alkaline basalt volcanism is a mantle plume upwelling. Such a scenario, considering a localized mantle plume, 'baby-plume' or 'plume-finger', was used to explain the basalt volcanism in the Pannonian Basin (Granet *et al.* 1995; Embey-Isztin *et al.* 2001; Wilson and Patterson 2001; Seghedi *et al.* 2004*b*; Seghedi and Downes 2011), similarly to other geochemical, thermomechanical and geophysical models (Goes *et al.* 1999; Buikin *et al.* 2005; Burov and Cloetingh 2009; Neubauer and Cao 2021). However, Harangi and Lenkey (2007) and Harangi *et al.* (2015*a*) argued that this is not viable, because of a lack of

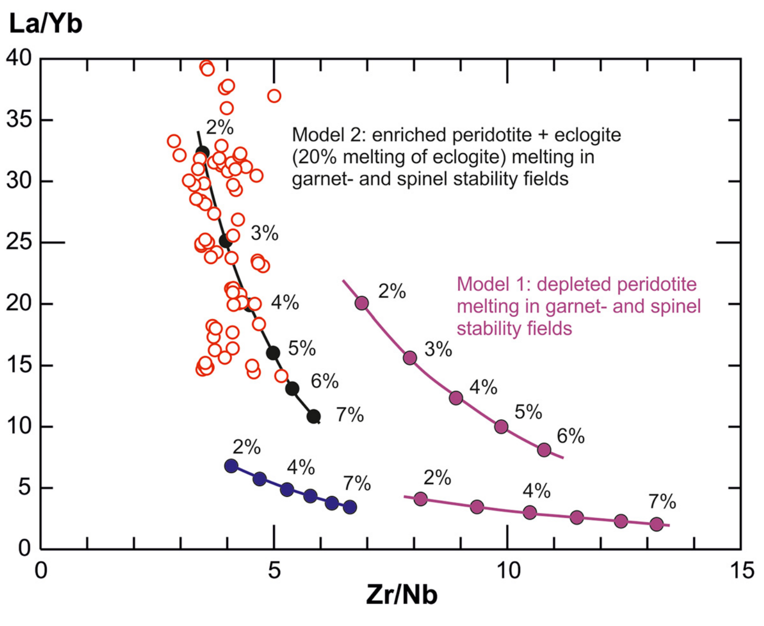
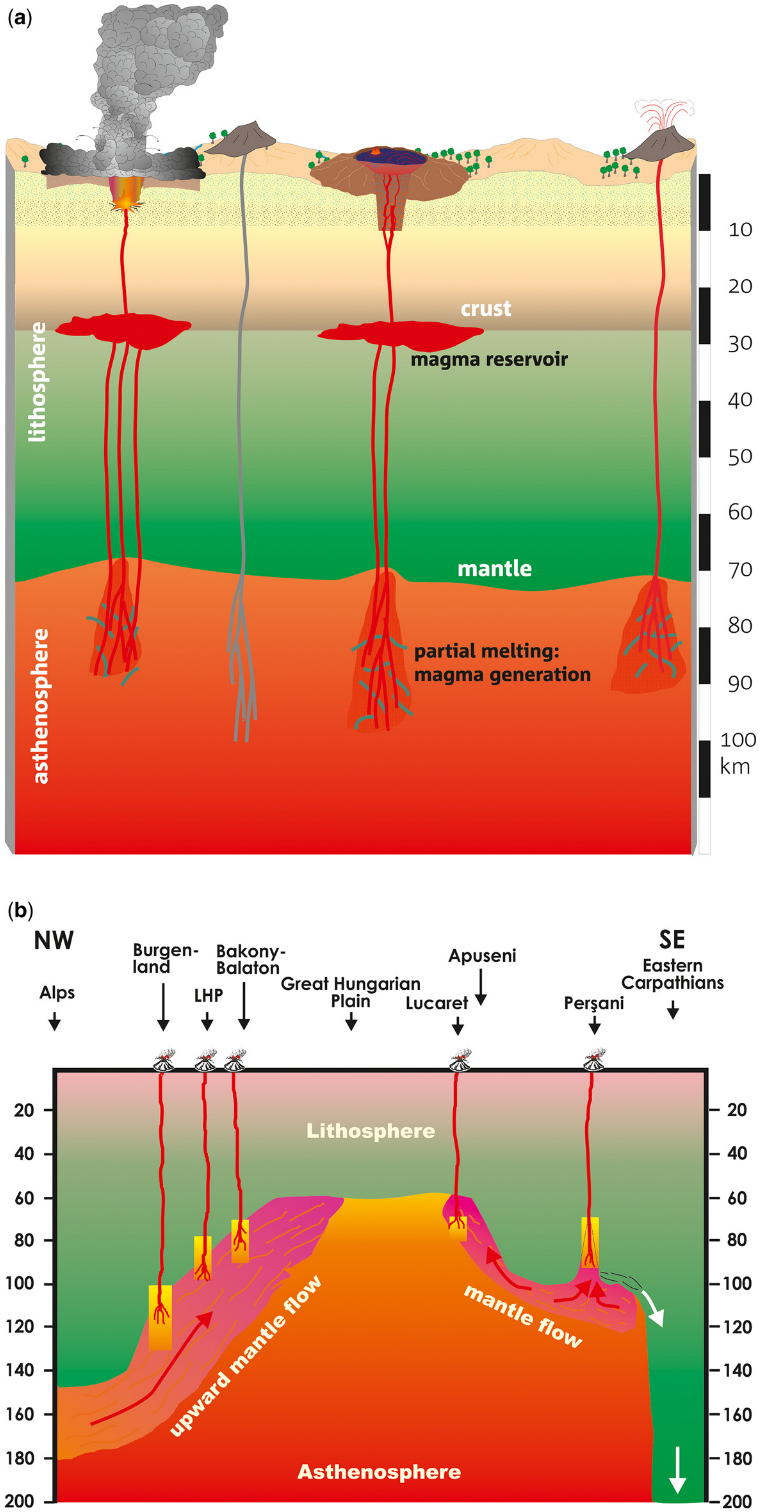


Fig. 13. Petrogenetic model calculations for the origin of the alkali basalt magmas represented by volcanic products with red circle symbols in the Carpathian–Pannonian Region. These suggest that Model 2 explains the compositional variation, i.e. melting of peridotite with some eclogitic lithology. Approximately 20% partial melting of eclogite produces approximately 3% contribution to the final melt. For model parameters and basalt data, see Harangi *et al.* (2015*a*). Source: mid-ocean ridge basalt-type eclogite composition is from Boniface *et al.* (2012). In the model calculation, the dynamic partial melting model of Zou (1998) was used.

evidence for a high-temperature mantle upwelling. Instead, magma generation occurred in melting columns in the uppermost asthenosphere with mantle potential temperatures of 1310–1390°C. In a few cases, higher melting temperatures (1410–1490°C) were calculated, but these were localized eruptions fed by low-volume basaltic magmas, whereas mantle plume upwelling requires excessive melting and protracted volcanism. Furthermore, the basaltic volcanic fields developed at the margins of the CPR (Fig. 2b), not in the interior, as would be expected if a mantle plume was present.

Kovács *et al.* (2020) and Koptev *et al.* (2021) proposed a new explanation for the origin of the alkali basalt volcanism in a stage when the Pannonian Basin was in compression rather than in extension. They suggested that basaltic melts were present in the upper asthenosphere for a long time, because of the elevated water content, but they were not extracted during the syn-rift phase. In their model, they proposed that several million years later, these melts were squeezed out because a vertical foliation developed in the upper asthenosphere as a response to compression between the northeastward-moving Adriatic plate and the European platform. Thus, as they stated, neither decompressional melting nor the presence of pyroxenite or eclogite was required to explain the origin of the basaltic melts. This explanation, however, cannot be held for several reasons. The large variation of chemical composition of the basalts, both in trace elements and particularly in isotopic ratios, even in single volcanic fields, does not fit with a suggestion that they are derived from a melt zone, presumably with homogeneous composition if it is in the upper asthenosphere for a protracted time. Furthermore, trace-element data clearly suggest that melting occurred in the presence of garnet, which implies melting at >80 km depth, i.e. melting started deeper than the asthenosphere–lithosphere boundary. Bracco-Gartner and McKenzie (2020) used rare earth element modelling and concluded that partial melting took place at the spinel–garnet transition, although they invoked a depth of 63–72 km, just beneath the lithosphere. Nevertheless, they considered that mantle upwelling caused by thermal buoyancy is necessary to initiate intermittent minor partial melting. In addition to trace elements, major elements such as FeO are also sensitive to the depth of melting (Langmuir *et al.* 1992; Hirose and Kushiro 1993) and can be used to constrain the initial and final pressure of the melting column. Harangi *et al.* (2015*a*) performed model calculations for each volcanic field and determined the depth of partial melting. The initial pressure corresponds to the mantle solidus, whereas the final pressure usually gives the depth of the asthenosphere–lithosphere boundary (LAB). Based on the calculated melting columns, partial melting beneath the Pannonian Basin started about 90–120 km (Fig. 14b), where garnet is stable and was continuous upwards to a depth where spinel–garnet peridotite and spinel peridotite underwent melting (Fig. 13). Nevertheless, trace-element patterns reflect mostly the initial melting stage in the garnet-peridotite field (Niu *et al.* 2011). Overall, the wide range of trace-element composition of the basalts is consistent with small-degree (1–3%) melting predominantly of garnet-peridotite lithology in the asthenospheric mantle (Fig. 13; Harangi *et al.* 1995*b*, 2015*a*; Harangi 2001*a*, 2001*b*; Harangi and Lenkey 2007; Bracco-Gartner and McKenzie 2020). Similar petrogenetic results have been obtained for several other volcanic fields, such as in Germany (Jung *et al.* 2005, 2006, 2012), Auckland (McGee *et al.* 2013) or the Basin and Range (Ormerod *et al.* 1991; Bradshaw *et al.* 1993). Furthermore, the asthenosphere is a convective part of the upper mantle, where vertical foliation as inferred by Kovács *et al*. (2020) cannot be developed owing to the compressional stress such as in the lithosphere, particularly in the crustal parts owing to the convergence of microplates or plates. In contrast, we argue that in a monogenetic volcanic field, each eruption event corresponds to a melting event in the upper mantle (Fig. 14a).

The reason for the melting starting > 80 km depth during the post-rift phase of the Pannonian Basin is, however, still elusive. Melting of the asthenosphere requires upwelling, i.e. decompression, which allows the solidus to decrease below the ambient mantle temperature. However, considering a pure dry peridotite lithology, a high mantle temperature (>1500°C) is necessary at these depths. Even a small amount of water supresses the peridotite solidus (Green *et al.* 2001, 2014; Hirschmann 2006), but this can also be achieved by the presence of pyroxenite or eclogite in the asthenospheric peridotite (Hirschmann and Stolper 1996; Hirschmann *et al.* 2003; McGee *et al.* 2013; Lambart *et al.* 2016) because pyroxenite and eclogite have lower melting points than peridotite. This scenario can be tested using olivine compositions (Sobolev *et al.* 2007; Herzberg 2011; Le Roux *et al.* 2011; Søager *et al.* 2015; Yang *et al.* 2016; Howarth and Harris 2017; Korkmaz Gencoğlu *et al.* 2022), particularly trace elements such as Mn and Zn. Our preliminary olivine compositional data suggest that the primary mafic magmas were generated in a mixed peridotite–pyroxenite/eclogite source. This is also supported by trace-element modelling invoking eclogitic melting followed by partial melting of a heterogeneous (97% peridotite + 3% eclogite) asthenospheric mantle lithology (Fig. 13). However, mantle upwelling is still necessary to induce melting, even in such heterogeneous lithology. Harangi and Lenkey (2007) and Harangi *et al.* (2015*a*) pointed out that the spatial distribution of the basalt volcanic


(a)
crust
magma reservoir
lithosphere
mantle
asthenosphere
partial melting:
magma generation
10
20
30
40
50
60
70
80
90
100
km
(b)
NW
SE
Alps
Burgen-
land
LHP
Bakony-
Balaton
Great Hungarian
Plain
Apuseni
Lucaret
Perşani
Eastern
Carpathians
Lithosphere
upward mantle flow
mantle flow
Asthenosphere
20
40
60
80
100
120
140
160
180
200

fields (Fig. 2) is significant, showing that partial melting took place mostly beneath the margins of the Pannonian Basin, where there is a strong gradient of the LAB from >130 km beneath the Eastern Alps or Western Carpathians to <70 km beneath the Pannonian Basin. Mantle flow towards the thin spot (Fig. 14b) has a strong vertical upwards component inducing decompressional melting of the pyroxenite-rich mantle peridotite with its low solidus temperature. Such heterogeneity can also explain the compositional variation of the basalts both in trace elements and in isotopic ratios, as well as in spinel chemical composition (Harangi *et al.* 2015*a*). Further research can refine this model or the role of water at such depths. Nevertheless, melting of the asthenospheric mantle is a prerequisite to an eruption, and this provides a viable model for basalt magmagenesis against a longstanding basaltic melt layer in the uppermost asthenosphere (cf. Kovács *et al.* 2020; Koptev *et al.* 2021). Melting of the heterogeneous asthenospheric mantle beneath the Pannonian Basin is still capable of producing magma, as the youngest eruption event of Putikov volcano proved 100–500 ka ago.

The model shown in Figure 14 can be applied to most of the Pannonian Basin; however, a different view is necessary for the volcanism in the south-southeast CPR (Fig. 10). The youngest basalt volcanic field here is Perșani, located in the transtensional Brasov basin, close to the Vrancea zone and to the young Ciomadul volcano. Alignment of the volcanic centres is tectonically controlled, since they sit on northeast–southwest-trending normal faults (Ciulavu *et al.* 2000; Seghedi *et al.* 2019). $^{40}Ar/^{39}Ar$ data indicate that volcanism took place between 1.2 and 0.68 Ma (Panaiotu *et al.* 2013) in several stages (Seghedi *et al.* 2016). The alkali basalts have compositions close to the primary magmas, so they ascended rapidly from the upper mantle (Downes *et al.* 1995*a*; Harangi *et al.* 2013). Melt generation occurred in a depth range of 90–60 km, mostly in the spinel-peridotite stability field in the asthenosphere. The asthenospheric mantle is heterogeneous, as shown by spinel compositions (Fig. 12; Harangi *et al.* 2013; Bracco-Gartner *et al.* 2020), and has a minor subduction component indicated by the low Ce/Pb ratio and slightly elevated $^{87}Sr/^{86}Sr$ isotope ratio (Embey-Isztin *et al.* 1993; Downes *et al.* 1995*a*). Nevertheless, this subduction component could be derived from the metasomatized lithospheric mantle, where asthenosphere-derived magmas interacted with parts of the lithosphere metasomatized by LILE-rich fluids (Bracco-Gartner *et al.* 2020; Faccini *et al.* 2020). Variation in Zn/Fe and $^{87}Sr/^{86}Sr$ ratios can be explained by initial pyroxenite/eclogite-dominated melting, whereas later magma generation took place predominantly in peridotite lithology at relatively higher temperatures (1300–1380°C) (Ducea *et al.* 2020). Abundant peridotite xenoliths in the basalts (Vaselli *et al.* 1995; Falus *et al.* 2008; Faccini *et al.* 2020) clearly imply the existence of the mantle part of the lithosphere, which contradicts the model by Fillerup *et al.* (2010), who invoked delamination in the lower crust and direct contact of the continental crust with asthenospheric mantle. Development of the Perșani basaltic volcanic field was related to the geodynamic evolution of the SE Carpathians, where slab roll-back, lower lithosphere rupture or delamination of the lowermost lithosphere and associated asthenospheric flow played a role in magma generation (Fig. 10; Chalot-Prat and Girbacea 2000; Harangi *et al.* 2013; Bracco-Gartner *et al.* 2020; Ducea *et al.* 2020), processes that are still occurring and may cause potential hazards.

Concluding thoughts

This review outlined our present knowledge of the Neogene–Quaternary volcanism in the CPR. It emphasized the role of early plate tectonic explanations and how these ideas have developed through half a century. Volcanism is a surficial phenomenon of magma generation and magma evolution in the upper mantle and crust, and it is important to reveal the reason for these deep processes. Volcanic rocks formed during the eruptions can provide this information, such as the time of eruptions, and the origin and evolution of the magmas. The ‘source-to-the-surface’ studies with high-resolution, crystal-scale investigation of volcanic rocks give us a snapshot of these events, even quantitatively, which can be combined with other geological and geophysical data and observations to formulate a geodynamic model. Classic plate tectonic explanations have to

Fig. 14. (**a**) Conceptual model for the formation of a basalt volcanic field. Each eruption event corresponds to individual melting in the heterogeneous asthenospheric mantle. Thus, the erupted magmas over time have distinct compositions reflecting the mantle source characteristics and the degree of melting. (**b**) Proposed model for the reason of the Late Miocene to Quaternary alkaline basalt volcanism in the Pannonian Basin (Harangi and Lenkey 2007; Harangi *et al.* 2015*a*). Asthenospheric mantle flow along the sharp changes in the lithosphere–asthenosphere boundary results in decompression melting at various depths depending on the source lithology. The depth ranges of melting columns calculated from the major- and trace-element composition of the most primitive basalts (Harangi *et al.* 2015*a*) are consistent with this model. LHP, Little Hungarian Plain.

be refined based on new data. The Carpathian–Pannonian Region, and within it the Pannonian Basin, is a natural laboratory, where development of an extensional basin with thinned continental crust can be better understood, and numerical thermomechanical models can be introduced and tested (Bally and Snelson 1980; Horváth and Berckhemer 1982; Royden *et al.* 1983*a*, *b*; Horváth 1993; Royden 1993; Tari *et al.* 1999, 2024; Horváth *et al.* 2006*a*, 2015; Bada *et al.* 2007; Balázs *et al.* 2016, 2017, 2018).

The new results may even prompt debate on fundamental issues, such as whether the Pannonian Basin can be considered a back-arc basin (Bally and Snelson 1980; Roberts and Bally 2012; Tari *et al.* 2024), as already discussed by Seghedi *et al.* (2013). The concept of the back-arc basin was introduced by Karig (1971) for tectonic setting characterized by active seafloor spreading behind subduction zones. Later, this definition was extended to other areas, even to continental settings where major lithospheric thinning occurred with relation to active subduction. A thorough discussion of the nature and geodynamics of back-arc basins was given by Artemieva (2023), who pointed out that very different processes can lead to the formation of such environments also involving cases with no associated subduction. Within the Mediterranean, the Alboran, the Ligurian–Provence, the Tyrrhenian, the Aegean and the Pannonian Basin are all regarded as back-arc basin systems (Horváth and Berckhemer 1982; Roberts and Bally 2012; Balázs *et al.* 2016, 2022; Tari *et al.* 2024). Nevertheless, there are alternative models for their origin, such as in the case of the Aegean Basin (e.g. Pe-Piper and Piper 1989; Agostini *et al.* 2010), as pointed out by Artemieva (2023). The evolution of the Pannonian Basin is complex; although, there is a general consensus that the significant extension of the continental lithosphere was largely driven by the retreating subduction along the eastern margin of the CPR. Therefore, structurally, the Pannonian Basin can still be regarded as a back-arc basin area. Yet, as many papers have demonstrated, no classic volcanic arc developed during the active subduction. The formation of volcanic rocks with chemical compositions analogous to those observed in modern subduction zones can be attributed to lithosphere thinning by melting of the mantle regions metasomatized by various fluids derived partly from previous subduction events. This is distinct from the classic subduction-related volcanic arcs, where magmas are formed by fluid-flux melting above a downgoing slab. An intriguing question is why calc-alkaline-type volcanism took place in the syn-rift period, basically as a response of lithospheric extension and in other places during a post-collisional period and not contemporaneously and directly related to active subduction. Lexa and Konečny (1974) already argued for extension-related calc-alkaline volcanism, calling them 'areal-type' volcanic rocks and separating them from the classic subduction-related volcanic arcs. Subsequent petrological and geochemical data and model calculations supported this interpretation (Harangi *et al.* 2001, 2007). Although the calc-alkaline-type volcanic rocks are found mostly along a belt parallel to the Carpathians, which apparently looks like a volcanic arc, a significant proportion of the Neogene volcanic material is present within the basement, buried by young sediments in the Pannonian Basin. Thus, the volcanoes cover a much wider area and are not confined to the northern and eastern margins of the Pannonian Basin. Ongoing studies of these volcanic rocks will shed new light on the lateral and temporal extent of the volcanism.

The intense volcanism, and particularly the associated deep magmatic processes, had a major impact on the thermal nature of the crust and caused the brittle–ductile transition zone to move to shallower depths. This allowed emplacement of large volumes of silicic magma in the shallow crust and may have played a role in hydrocarbon maturation (Lukács *et al.* 2018). Thermomechanical numerical models (Balázs *et al.* 2016, 2017, 2018; Fodor *et al.* 2021) will certainly help to understand this process better, as well as the lithosphere–asthenosphere and mantle–crust interactions leading to magma generation.

Extension of the lithosphere and crust could be due to retreating subduction (Csontos *et al.* 1992; Csontos 1995), but this latter process seems to be decoupled from the magmatism and volcanic activity. A crucial question is the role of subduction along the Dinarides (Royden and Faccenna 2018), how it influenced asthenospheric flow and, if so, whether it controlled the formation and evolution of the Pannonian Basin (Kovács *et al.* 2012*a*; Qorbani *et al.* 2016). Another elusive question is the origin of the alkali basalt volcanism. There is no simple scheme to decipher the origin of mafic magmas during post-rift thermal subsidence and tectonic inversion. From a magmatic point of view, a crucial question to be resolved is why there is no basalt volcanism in a location where the present asthenospheric updoming is up to *c.* 30 km depth such as in the Békés basin, the southeastern part of the Great Hungarian Plain (Posgay *et al.* 1995; Tari *et al.* 2024). Under such conditions, partial melting is expected, since mantle material, even dry peridotite, should melt when at depths <40 km (McKenzie and Bickle 1988). In short, after more than 50 years of plate-tectonic concepts and the first modern models for the origin of the CPR, there are still open questions to be solved. Science is evolving by questioning and discussing even basic issues. Mutual understanding of the results of different disciplines is crucial, and we hope that our outline of the

Neogene–Quaternary volcanism and magmatism could help in this effort.

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