Middle Miocene volcanism in the vicinity of the Middle Hungarian zone: evidence for an inherited enriched mantle source – a review

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Abstract

Middle Miocene igneous rocks in the vicinity of the Middle Hungarian zone (MHZ) show a number of subduction-related geochemical characteristics. Many of these characteristics appear to be time-integrated, showing a decreasing subduction signature with time. In contrast to previous models, which suggest southward-dipping subduction of European lithosphere beneath the Alcapa microplate (along the Western Carpathians) is responsible for the chemical characteristics seen in middle Miocene volcanics, we propose that source enrichment occurred via the subduction of either the Budva-Pindos or Vardar Oceans. Recent seismic studies have revealed that the proposed southward-dipping subduction was not developed beneath the entire Western Carpathians or, even if it had, was overprinted by the collision of the European plate and the Alcapa unit at 16 Ma. This subduction is thought to have started 30 Ma ago, therefore the time between the onset of subduction and collision cannot account for extensive source enrichment in the overlying mantle wedge. It is also pertinent to note that the middle Miocene igneous rocks of the MHZ in their reconstructed positions are not parallel to the supposed suture expected for subduction-related arc volcanoes.

Our review suggests an alternative hypothesis, whereby source enrichment is related to the subduction of either the Budva-Pindos or Vardar Ocean during the Mesozoic-Paleogene. In this model the Alcapa microplate was transferred to its present tectonic position via extrusion and rotations. Geophysical modeling and mantle xenoliths provide evidence that this process occurred at the scale of the lithospheric mantle, indicating that the subduction-modified lithospheric mantle was coupled to the crust.
Melting in the lithospheric mantle of the Alcapa unit was triggered by the extension during the formation of the Pannonian Basin. The preserved subduction-related geochemical character of volcanics in intra-plate settings that are otherwise directly unaffected by subduction, can be attributed to tectonic transport of metasomatised mantle from a previous subduction-affected setting. This model provides an alternative approach to understanding the geochemical complexity seen among intra-plate calc-alkaline volcanics, where chemical characteristics can be explained without the involvement of plumes.

Key words: Carpathian-Pannonian region, middle Miocene, volcanics, mantle xenoliths, mantle tomography, geodynamics
The Tertiary and Quaternary volcanism of the Carpathian-Pannonian region (CPR) has been the focus of extensive recent research because the area provides an excellent opportunity to understand the interplay between tectonics and volcanism (Chalot-Prat and Boullier, 1997; Chalot-Prat and Girbacea, 2000; Downes et al., 1995; Harangi, 2001b; Harangi et al., 2001; Harangi et al., 2006; Harangi and Lenkey, 2007; Konečny et al., 2002; Konečny and Lexa, 1999; Pécskay et al., 2006; Pécskay, 1995; Póka, 1988; Póka et al., 2004; Salters et al., 1988; Seghedi et al., 1998; Seghedi et al., 2004a; Seghedi et al., 2005; Szabó et al., 1992; Zelenka et al., 2004; Vaselli et al., 1995). These studies revealed that most of the Tertiary igneous rocks in the CPR were derived from a subduction-enriched source, with both the degree of crustal contamination and strength of the subduction-related geochemical signature (i.e., high LILE/HFSE ratios, negative Nb, Ta and Ti anomalies; lower $^{87}\text{Sr}/^{86}\text{Sr}$ and higher $^{143}\text{Nd}/^{144}\text{Nd}$ radiogenic isotopic ratios) decreasing with time. Most younger Plio-Pleistocene alkaline basaltic rocks of the region, which mainly postdated the main phase of calc-alkaline volcanism (with the exception of the Eastern Transylvanian basin where it was contemporaneous), show very little evidence for subduction. These Plio-Pleistocene basalts show geochemical affinities similar to that of an OIB-like asthenospheric source (Dobosi et al., 1998; Embey-Isztin et al., 1993b; Seghedi et al., 2004b). This being the case the relationship between the onset of proposed subduction along the Carpathians and the climax of magmatism still remains a matter of debate. Most previous studies have dealt with the whole CPR and distinguished different volcanic provinces based on
geochemistry and spatial distribution. In this paper we focus only on the middle Miocene volcanics of the “Western segment” (see Fig. 1) which were formed during the first phase of extension in the CPR (Husimans et al., 2001). This enables us to gain better spatial and temporal resolution and put forward an alternative explanation for the subduction-related source enrichment of these magmas. In addition to temporal and spatial considerations, this paper incorporates recent results and seismic tomography from the region. Although integration of such information can be used to help decipher the geodynamic evolution of the region, such attempts have been seen limited considerations in previous studies.

The most recent reviews on the Tertiary volcanism of the region (Harangi, 2001b; Harangi and Lenkey, 2007; Seghedi et al., 2004a; Seghedi et al., 2005) provide no alternative to the present geodynamic paradigm of the Carpathian-Pannonian region. This paradigm suggests that magmas were mainly produced by contemporaneous subduction-related arc volcanism along the Carpathians. Existence of a classic subduction margin for the entire Western Carpathians is now questioned by several authors (i.e., Grad et al., 2006; Szafián and Horváth, 2006; Szafián et al., 1997; Szafián et al., 1999, Tomek, 1993). There is evidence to suggest the existence of subduction along eastern sections of the Western Carpathians - the roll-back effects of which can be used to account for the observed block-rotations and fault patterns in the Northern Pannonian Basin (i.e., Bielik et al., 2004; Márton and Fodor, 2003; Sprener et al., 2002). The reconstructed position of middle Miocene volcanics of the “Western segment” in their reconstructed position are not parallel to the hypothesized southward subduction zone along the Carpathian mountain belt, but in fact, were generated in the vicinity of the
Middle Hungarian zone (MHZ, Fig. 1). Here the term “vicinity” refers to a zone 300-350 km wide and is centered on the MHZ. This width is not an arbitrary number, but is twice the distance at which typical arc-volcanoes occur relative to the trench (i.e., Ellam and Hawkesworth, 1988; Hawkesworth et al., 1994). Paleogene-early Miocene igneous rocks show a very similar spatial distribution close to the MHZ, which suggests there may be a link between the middle Miocene and this magmatic episode (Fig. 1, Kovács et al., 2007). This observation raises the possibility that subduction-related geochemistry is not related to any contemporaneous subduction, but it is an “inherited” feature from a previous geodynamic setting. This “inherited” model is a known model and its overview is given before the discussion. Our main aim in this paper is, therefore, to address the subduction-related source enrichment at the origin of the middle Miocene igneous rocks along the MHZ. We consider all potential subduction zones which may account for this source enrichment and examine how each subduction zone matches the temporal, and geodynamic characteristics of volcanism seen in the CPR.

2. Geological background

The wider Carpathian-Pannonian region (CPR) comprises several mountain chains, i.e., the Eastern Alps, the Carpathian arc and the Dinarides, which surround the Intra-Carpathian Basin System. The Intra-Carpathian Basin System, the internal part of which is commonly referred to as the Pannonian Basin, has a significantly extended continental crust. This extended crust outcrops in several smaller internal mountains, and is filled by middle to late Miocene sediments (Fig. 1).
Based on the Mesozoic tectonostratigraphy and structural analysis, the internal area is subdivided into two major units (Csontos, 1995; Csontos and Vörös, 2004; Haas et al., 1995; Haas et al., 2000; Kovács et al., 2000): Alcapa (ALps-CArpathians-PAnnonian, i.e. northern Intra-Carpathian Basin System and Western Carpathians) and Tisza-Dacia (southern Intra-Carpathian Basin System; East and South Carpathians) (Fig. 1). In this paper the Alcapa unit is defined as the unit bounded to the south by the Periadriatic and Balaton faults (Balla, 1984; Csontos, 1995; Fodor et al., 1999; Fodor et al., 1998). The Balaton fault separates Alcapa from the Mid-Hungarian zone. The major fault separating the Mid-Hungarian unit from the Tisza unit to the south is termed the Mid-Hungarian (or Zagreb-Zemplin) fault (Balla, 1984; Csontos and Nagymarosy, 1998; Haas et al., 2000; Kovács et al., 2000; Wein, 1969) (Fig. 1). The Mid-Hungarian zone (MHZ, Fig. 1) is composed of the Bükk Mts. (N Hungary) and a narrow structural belt between Lake Balaton and the Mecsek Mts. (SW Hungary). Exposures seen within this zone consist of low- to high-pressure metamorphic Paleozoic-Mesozoic continental margin sediments, a mélange, and dispersed remains of a Jurassic ophiolite nappe (Csontos and Vörös, 2004; Haas et al., 2000; Harangi et al., 1996; Wein, 1969). The Mid-Hungarian zone was strongly deformed during Tertiary times (Balla, 1987; Csontos and Nagymarosy, 1998).

Based on numerous structural studies and tectonic reconstructions (Balla, 1984; Csontos, 1995; Csontos et al., 2002; Fodor et al., 1999), the study area was formed in three major steps during Tertiary times. The earliest tectonic phase took place in the Paleogene, with a major right lateral shear event along the Periadriatic zone and the Balaton fault. This shear event was initiated in the Late Eocene (Fodor et al., 1992), but the bulk of shearing appears to have occurred during the Oligocene. As a consequence,
Alcapa was subjected to major right lateral displacement (Kázmér and Kovács, 1985). This event was contemporaneous with the proposed onset of subduction along the Western Carpathians ~30 Ma ago (Oszczypko, 1992; Pécskay et al., 2006; Tomek, 1993; Tomek and Hall, 1993). The existence of subduction along the entire Western Carpathians is now doubted by many reconstructions (i.e., Mantovani et al., 2000), which proposed that the subduction had a NW-SE strike parallel to the eastern portion of the Western Carpathians and the Eastern Carpathians. The remaining western part of the Western Carpathians was characterized by transtensional movements with a strike subparallel to ENE-WSW (Fig. 1). It is unclear then, whether well-developed subduction occurred along the entire Western Carpathians (see discussion for more detail).

The next tectonic phase was dominated by opposite rotations of the Alcapa and Tisza-Dacia microplates within the CPR (Csontos et al., 2002; Márton, 1987). Based on paleomagnetic data (Márton, 1987), the two microplates were detached from their southern neighbor, the Dinarides, and were extruded or rotated into the Carpathian embayment during the late Oligocene and early Miocene (ca. 20-18 Ma).

The last major tectonic phase occurred from the Middle-Miocene to subrecent, and still shapes the area today (Bada and Horváth, 2001; Csontos, 1995; Csontos et al., 2002; Horváth, 1993; Horváth et al., 2006; Horváth and Cloetingh, 1996). This final tectonic stage includes the formation of the Carpathians and opening of the Pannonian basin, which occurred in two distinct phases. The first extension phase took place in the early-middle Miocene (17.5-14 Ma) which affected both the internal and external part of the Pannonian basin with a β factor of 1.4-1.6 (Huismans et al., 2001). This first stage of basin formation was also contemporaneous with major counter-clockwise rotation of the
Alcapa block (Márton and Fodor, 2003). This tectonic phase is considered to be driven by the southward and westward directed subduction of the European margin beneath the internal Carpathian area (Balla, 1984; Horváth, 1993; Jolivet and Faccenna, 2000; Márton and Fodor, 2003). Since the “classical” development\(^1\) of subduction along the entire Carpathians is now in question, there may be other driving forces for basin formation. One possible driving force is eastward directed asthenospheric flow from the Alpine collision belt analogous to that seen in the Himalaya collision belt (Flower et al., 2001).

The collision between Alcapa and the European platform is thought to have started 16 Ma ago, with collision becoming younger towards the southeast. During this time the MHZ was subjected to extensive deformation between the rotating Alcapa and Tisza-Dacia blocks (Csontos and Nagymarosy, 1998) and was almost in perpendicular position to the remaining subduction front along the Carpathians. This phase was followed by a second extension stage in the late Miocene (11-8 Ma) which is related to the gravitational collapse of the asthenospheric dome. This phase affected only the central part of the CPR and the mantle lithosphere with a δ factor of 6-8 (Husimans et al., 2001). At 6 Ma ago there was a smaller rotation of the Alcapa block due to the change in the rotation of the Adriatic indenter (Márton and Fodor, 2003).

### 3. Temporal and spatial distribution of middle Miocene igneous rocks

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\(^1\) By the term “classical subduction” we refer to subduction of an oceanic slab, where the subducted slab reaches considerable depth (>200 km), the Benioff-zone is well-developed and the slab is capable of causing extensive metasomatism in the overlying mantle wedge (i.e., Peacock, 1991; van Keken, 2003; Schmidt and Poli, 1998)
The Tertiary igneous rocks that cover a considerable part of the CPR have been classified in different ways by different authors (see Seghedi et al., 2004a; Seghedi et al., 2005; Harangi, 2001; Harangi and Lenkey, 2007 for more detail). In this review we try to adopt a spatial classification that combines these previous attempts, but does not discuss volcanites of the Apuseni Mts (“Central segment”). Here we distinguish a “Western” and “Eastern segment”. The “Eastern segment” covers igneous rocks parallel to the Eastern Carpathians with NNW-SSE strike, from the Vihorlat Mts. to the East Transylvanian basin (Fig. 1). These are mainly middle-Miocene-subrecent calc-alkaline rocks which have been interpreted as subduction related arc volcanites (Seghedi et al., 2004a; Seghedi et al., 2005). In the Persani Mts. (East Transylvanian basin, ETB) there are contemporaneous Plio-Pleistocene calc-alkaline and mafic alkaline rocks whose genesis is still matter of debate (cf., Chalot-Prat and Boullier, 1997; Chalot-Prat and Girbachea, 2000; Mason et al., 1998; Seghedi et al., 2004).

The “Western segment” is much more complex than the “Eastern”, showing wider geochemical, spatial and temporal variability. In this study the “Western segment” includes igneous rocks of the Alcapa, MHZ and northern portion of the Tisza units, from the Tokaj Mts. to the Styrian basin which line up along a NE-SW directed strike that is almost perpendicular to the “Eastern segment” (Fig. 1). The genesis of middle Miocene igneous rocks in the vicinity of the MHZ is mainly attributed to the first phase of extension (induced by subduction roll-back), that caused melting in the previously metasomatised lithosphere (i.e., Harangi, 2001b; Harangi and Lenkey, 2007).

The middle Miocene mainly calc-alkaline, silicic and very rarely ultrapotassic volcanites in the “Western segment” either built up mountain ranges as a series of
volcanic edifices (that have subsequently been strongly disintegrated and eroded by now) or are buried under the thick sedimentary sequence of the Great and Little Hungarian plain nearly parallel to the MHZ (Fig. 1). The most important parts of the mountain ranges (from west to east) are as follows: Visegrád Mts., Börzsöny Mts., Central Slovakian Volcanic Field, Cserhát-Mátra Mts. Bükk Mts and Tokaj-Slanac Mts (Fig 1). In addition to volcanites observed in these ranges, Zelenka et al. (2004) reported large andesitic and dacitic strato-volcanoes buried below southern Transdanubia, Danube-Tisza Interfluve and Great Hungarian plain (Fig. 1). Similarly, Hajnal et al. (2004) identified a thick middle Miocene igneous body buried beneath the Nyírség district via a low-frequency 3-D seismic survey. This area is on the boundary between the Alcapa and Tisia microplates and sits on the MHZ. Partially buried volcanic rocks have been also reported in the Styrian basin (SB) and Little Hungarian Plain (LHP), and are dominantly potassic, intermediate to acidic middle and late Miocene volcanic rocks (Harangi et al., 1995; Harangi, 2001a). Pamić et al. (1995) identified a thick middle Miocene volcanic succession buried in the South Pannonian Basin (including the Drava Basin) at the southern side of the MHZ. Among the observed igneous rock types erupted along the MHZ, there is a general younging tendency towards NE (Pécskay et al., 1995; 2006).

Alkaline mafic volcanism in the “Western segment” (Styrian basin (SB), Little Hungarian Plain (LHP), Bakony-Balaton Highland (BBH) and Nógrád-Gömör (NG)) post-dated the middle Miocene, predominantly calc-alkaline magmatism. Alkaline mafic volcanism started at 11 Ma ago and was active till the Plio-Pleistocene (Pécskay et al. 1995, 2006).
4. Petrology and geochemistry

The middle Miocene volcanic rocks of the “Western segment” which are considered in this study are summarized in Table 1. They show wide petrologic variety, ranging from basalts to rhyolite. Projected in the TAS diagram (Le Bas et al., 1986) most compositions correspond to basaltic andesite and andesite fields (Fig. 2). Rare Earth Elements (REE) display an enriched, convex-upward pattern relative to chondritic values (Fig. 3a). General trace element patterns show a positive anomaly in Large Ion Lithophile Elements (LILE) and Pb, with a negative anomaly in the High Field Strength Elements (HFSE). The enrichment of LILE and LREE over HFSE (especially Nb) is a typical feature of subduction-related volcanic rocks (Ellam and Hawkesworth, 1988) (Fig. 3b). Paleogene-Early Miocene igneous rocks along the MHZ are similar to Middle-Miocene rocks in terms of trace element patterns, displaying the same anomalies and comparable concentrations (Fig. 3c and 3d).

Trace element discrimination diagrams also show that the geochemical characteristics of more primitive middle Miocene volcanics are similar to subduction-related magmas. The Ba/Zr vs. Nb/Zr diagram of Pereplov et al. (2005) distinguishes IAB, MORB and WPB sources (Fig. 4a). Ba/Zr ratio is indicative of fluid contribution from a subducted slab, whereas Nb/Zr ratio is sensitive to the effect of slab-derived melts. Middle Miocene volcanics show a transitional character, where the most primitive magmas with SiO$_2$<56 wt%$^2$ lie between the IAB and MORB and WPB fields - displaying an almost constant Nb/Zr ratio but a large variation in Ba/Zr ratio. More

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$^2$ These rocks represent the most primitive calc-alkaline rock that are likely to be unaffected by any contamination and fractionation.
silica-rich samples, however, are closer to the IAB field. Paleogene-early Miocene igneous rocks along the MHZ (Kovács et al., 2007) are very similar to their middle Miocene counterparts, with a somewhat higher Ba/Zr ratio. Typical fields for different OIB compositions (HIMU, EMI and EMII) are also displayed, showing that middle Miocene and Paleogene-early Miocene rock of the MHZ fall close to the EMI component.

A plot of Th/Y vs. Nb/Y gives constraints on different mantle source components (Fig. 4b). The Nb/Y ratio can be used to differentiate among mantle sources, whereas Th/Y can be used to monitor the transfer of slab-derived components to the mantle wedge (Elliot et al., 1997; Peate et al., 1997; Seghedi et al., 2004a). Only middle Miocene magmas with <56 wt% SiO$_2$ were used in this plot. Their relatively high Th/Y ratio is indicative of considerable sediment influx, whereas variation in Nb/Y refers to heterogeneous mantle source. From the OIB components only EMI and EMII show similar Th/Y values, while their Nb/Y ratios are higher. Paleogene-early Miocene rocks of the MHZ display larger variation in both ratios implying significant source variation and even larger sediment input.

Nd-Sr isotopic systematics show an enriched character, where $^{143}$Nd/$^{144}$Nd and $^{87}$Sr/$^{86}$Sr ratios are in the range of 0.5122-0.5125 and 0.706-0.714, respectively (Fig. 4c). More mafic rocks (with <56 wt% SiO$_2$) are very similar to the EMI component. Paleogene-early Miocene rocks again show a very similar pattern to middle Miocene volcanites (Fig. 4c).

Younger, mainly post-Miocene alkaline basalts in the “Western segment” showing common OIB character, still have slight subduction overprint on their
geochemistry (Dobosi et al., 1998, Seghedi et al., 2004a). Boron concentrations are anomalously high in alkaline mafic magmas of the Bakony-Balaton Highland (BBH, Fig. 1), which is indicative of subduction contamination (Gméling et al., 2007).

5. Xenoliths

Over the last decades our knowledge of the lithosphere beneath the CPR has been greatly improved by studies of xenoliths hosted in the Plio-Pleistocene alkaline basalts. Here we attempt to demonstrate that xenoliths provide very useful information in geodynamic exploration. Xenolith-bearing host basalts occur at the edge of the CPR (Styrian basin (SB), Nógrád-Gömör (NG) and Eastern Transylvanian basin (ETB)) and in its central part (Little Hungarian Plain (LHP), Bakony-Balaton Highland (BBH)) (Fig. 1). These xenoliths in alkali basalts are of lithospheric and in some cases asthenospheric origin (Szabó et al., 2004; Falus, 2004). Because eruption of the host alkali basalts postdate the major tectonic events of the CPR, their xenoliths record effects preceding tectonic events - which include deformation (Falus et al., 2000; Falus, 2004; Falus et al., 2007; Hidas et al., 2007), metasomatism and melt extraction (i.e., Bali et al., 2006; Chalot-Prat and Boullier, 1997). The upper mantle xenoliths are mostly spinel peridotites (i.e., Downes et al., 1992; Embey-Isztin et al., 2001; Szabó et al., 2004; Vaselli et al., 1995), accompanied by subordinate pyroxenites and lower crustal granulites in the CPR (Dobosi et al., 2003; Embey-Isztin et al., 2003; Embey-Isztin et al., 1990; Kovács and Szabó, 2005; Kovács et al., 2004; Török et al., 2005; Zajacz et al., 2007). The peridotites represent residual mantle material showing textural and geochemical evidence for a
complex history of melting and recrystallization, irrespective of location within the region (e.g., Falus, 2004; Szabó et al., 2004; Szabó et al., 1995). This deformation has been attributed to a combination of extension and upwelling of asthenosphere in the middle Miocene (Szabó et al., 2004 and references therein). The mantle xenoliths contain hydrous phases (pargasitic and kaersutitic amphiboles and rarely phlogopite) as evidence for modal metasomatism, in all occurrences of the region.

For the purpose of this study only xenoliths from the BBH and NG will be considered in detail. This is because these localities are in close proximity to the middle Miocene volcanics of the “Western segment” with which we are concerned. The LHP is also close, but the available xenolith suite is limited (Falus et al., 2007). In the BBH, xenolith lithologies include extremely rare orthopyroxene-rich olivine websterites and a quartz-bearing orthopyroxene-rich websterite (Bali, 2004; Bali et al., 2006). From petrographic and geochemical evidence, these orthopyroxene-rich mantle rocks are probably the products of interaction between subduction-related melts (i.e., adakite, boninite) and mantle lithologies, as pointed out by McInnes et al. (2001) and Santos et al. (2002) for xenoliths from Papua New Guinea and tectonic peridotites from Spain, respectively. The original (i.e., slab-derived) melts are highly reactive in the mantle. The websterites, are therefore, though to be produced in fore-arc or arc setting - because melts formed in this environment cannot migrate far from their source without being completely consumed (Rapp et al., 1999). Downes et al. (1992) and Rosenbaum et al. (1997) have already proposed subduction-related characteristics of the BBH mantle xenoliths based on their radiogenic isotope signatures.
Granulite xenoliths have also been found at this locality, with Dobosi et al. (2003) and Embey-Isztin et al. (2003) hypothesizing a subduction-accretion origin for these lower crustal rocks. These authors went on to suggest that the protolith from which these xenoliths derived was a subducted oceanic slab which got stuck at lower crustal depth.

Previously, Embey-Isztin et al. (1990) suggested that a considerable volume of basaltic magmas were trapped and crystallized close to the MOHO beneath the BBH, causing massive underplating in the region.

Szabó et al. (1996) presented volatile-rich silicate melt inclusions in peridotite xenoliths as evidence for the existence of a metasomatic agent with subduction-related origin. Granulite and pyroxenite xenoliths are also present in great abundance, with lower-crustal granulites interpreted as being either the remnant of an older underplated magmatic body or a subducted oceanic slab (Kovács and Szabó, 2005). These rocks show indication of incipient melting around former garnets. It was also proposed that the incipient melting of these granulites may have played an important role in the formation of the Tertiary calc-alkaline volcanic rocks of the NG representing the “lower-crustal component” (Harangi, 2001a). Kovács et al. (2004) and Zajacz et al. (2007) reported clinopyroxenite xenoliths with silicate melt inclusions and concluded that they are the fragments of underplated basaltic rocks trapped at the MOHO or in the uppermost mantle beneath the NG.

6. Geophysical constraints
Seismic tomographic studies in the CPR show that the remnants of subducted slab(s) cannot be adequately identified. Piromallo and Morelli (2003), Spakman (1990) and Wortel and Spakman (2000) defined a cool slab lying subhorizontally beneath the CPR at a depth of 600 km. In the east, the Vrancea zone was thoroughly examined by Oncescu et al. (1984), where a cool and steeply dipping body is found. This seismogenic body was thought to be a remnant of the westward-subducted European slab. Detailed seismic tomography of Wortmann et al. (2001), however, showed a very steep (but slightly eastward-dipping) localized body at the same location. Geophysical constraints are even more controversial for the Western Carpathians, where the existence of a southward dipping subduction is ambiguous. Bielik et al. (2004) and Sroda et al. (2006) found features as southward dipping reflectors, thickened crust for the central and western part of the Western Carpathians deploying geophysical modeling (i.e., seismic and gravitational measurements) which is compatible with the idea of a southward dipping subduction of European plate beneath the Alcapa unit. In comparison to Bielik et al. (2004) and Sroda et al. (2006), Tomek (1993) and Tomek and Hall (1993) suggested that either the European slab did not subduct beneath the western CPR, or that slab is now completely detached, disintegrated and sunk into a considerable depth. Grad et al (2006) came to a very similar conclusion by studying the Carpathian transect (CEL05) of the Celebration 2000 project. The CEL05 transect is NNE-SWS directed and runs through the point where the W-E strike of the Western Carpathians turns into the NNW-SSE strike of the Eastern Carpathians. Grad et al. (2006) found northward-dipping reflectors beneath the Western Carpathians at mantle depth, which may exclude the existence of a well-developed southward-directed subduction of the European lithosphere beneath
Alcapa. In their interpretation, even if southward-dipping subduction existed, it would have to be a relatively old feature overprinted by a younger most probably northward directed thrusting of the Alcapa unit. Furthermore, 2D and 3D gravitational modeling showed that a deep crustal root did not develop beneath the Western Carpathians which would be expected for a “classical” convergent margin (Szafián et al., 1997; Szafián et al., 1999; Szafián and Horváth, 2006). These later authors suggested alternatively that this is due to the fact that the Western Carpathians are characterized by transtensional-transpressional plate motions with little or no collision at least since the middle Miocene.

The nearest unambiguous trace of a subducted slab outside the CPR has been identified along the eastern shoreline of the Adriatic Sea and at its northern continuation beneath the Po Plain (Koulakov et al., 2002; Lippitsch et al., 2003; Piromallo and Morelli, 2003). It is a north-northeastward-dipping slab that can be detected down to 600 km. This slab is most probably the remnant of the former Budva-Pindos Ocean (Csontos and Vörös, 2004).

7. An overview of the “inherited” enriched source theory

It is a common geological situation that magmas show subduction-related geochemical signatures in places otherwise unaffected by any recent or contemporaneous subduction (i.e., Basin and Range, (USA), [Fitton et al. (1991); Hawkesworth et al., (1995); Harry and Leeman (1995)]; Southern Sonora, (Mexico), [Till et al. (2007)]; Anatolia, (Turkey), [Pearce et al. (1990); Keskin et al. (1998); Aldanmaz et al. (2000)]; Eastern Rift (Morocco), [El Bakkali et al. (1998)]; This is because the subduction-related
signature from a previous geodynamic setting can be preserved in the mantle as metasomatic alteration(s), which can be reactivated via melting under various circumstances.

The subduction-related signature commonly refers to rocks which show relative enrichment in LILEs (Ba, Sr), alkalis and LREEs and depletion in HFSEs (Nb, Ta, Ti, Zr and Hf) (Saunders et al., 1991). This geochemical signature can originate from two main sources: the subducted slab flux and the hybridized mantle flux. The former one is due to the dehydration and melting of the subducted oceanic slab and interaction of these melts/fluids with the overlying mantle wedge. The melting of the slab occurs in the range of 90-120 km (Drumont and Defant, 1990), whereas the dehydration is a more or less continuous process from the compaction at relatively low pressures until the break-down of the most stable hydrous phases at deep-mantle depth (Ringwood, 1991; Frost, 2006; Schmidt and Poli, 1998). During these processes LILEs and LREEs go preferentially into the fluid phase, whereas HFSEs are retained in the actually stable titanium phase (titanite, rutile) (Saunders et al., 1991 and references therein). This is generally what leads to the formation of the “subduction” signal in mantle wedges. The second geochemical source, the hybridized mantle flux, is the result of melting in the previously metasomatised portion of the mantle wedge (i.e., “hybridized mantle”) above subducting slabs. This remelting produces a similar subduction-related geochemical signature that can differ somewhat depending both on P-T-fO₂ conditions and solid phases present in the residuum. It is for this reason that intra-plate magmas with subduction signature usually show larger variation in geochemistry than arc magmas (Foley et Wheller, 1990). An additional mechanism for producing subduction signature is via delamination and melting
of over-thickened eclogite-granulite facies lower crust. In this case melts/fluids released from this delaminated lower-crust resemble those of subducted oceanic crust and metasomatise the asthenosphere filling up the gap behind the detached lower crust (i.e., Lustrino and Wilson, 2007; Lustrino et al., 2007).

The mantle wedge consists of the original MORB-like subarc mantle with metasomatism-related magmatic veins and hydrous minerals, thought to be due to metasomatism occurring above a subducting slab. These veins are usually of tonalitic-dacitic character, which represents melting of the subducted slabs (Rapp et al., 1999; Ringwood, 1991). At the contact of these dikes one often finds pyroxenite and websterite reaction zones (i.e., Bali et al., 2007a and reference therein). There can also be hydrous phases in both the mantle wedge and reaction zones (pargasite, phlogopite), whose modal proportion can be up to 10% (Niida and Green, 1999; Wallace and Green, 1991). This means that an extensively metasomatised mantle portion can contain up to 0.4 wt % of H₂O (Hawkesworth and Gallagher, 1993). The presence of hydrous phases suggests that melting temperature can be significantly lowered relative to that of MORB mantle (i.e., Green and Lieberman, 1976; Green and Fallon, 2005).

These metasomatic alterations can be preserved both in the subcontinental lithospheric mantle and asthenosphere. The SCLM is defined as the uppermost layer of the mantle which is not involved in convection; it is therefore capable of preserving its distinct geochemical signature for a long time. The SCLM also has a better potential of transferring its geochemical heterogeneities during plate tectonic processes (i.e., Harry and Leeman, 1995). There is evidence for the existence of even 1 Ga old subduction enriched mantle source of kimberlites under western Australia (McCulloch et al., 1983).
and similar Proterozoic-Mesozoic source of calc-alkaline rocks in the Basin and Range province (USA) (Fitton et al., 1991; Hawkesworth et al., 1995). In contrast, the asthenosphere is convective which makes it less likely to preserve its geochemical heterogeneities for a long period of time. This is due to its continuous deformation (convection) and higher temperature relative to SCLM.

There are basically three ways to trigger melting in the subduction modified mantle after the cessation of subduction:

1) Continental extension arguably represents the simplest way in which the SCLM may cross the solidus. Melting in this scenario is caused by decompression associated with extension. Hawkesworth and Gallhager (1993) studied the relationship among initial lithospheric thickness, stretching factor and temperature, in order to find out if melting can occur within the SCLM or asthenosphere or in both (assuming that 0.4 wt% of H2O is present). The authors found that melting preferentially occurs in the SCLM if the degree of extension is small, while the SCLM is thick.

2) Plumes can trigger melting too because they disturb the thermal structure and raise the temperature above the solidus.

3) Slab delamination can place the lower part of the SCLM beneath thick lithosphere, due to its negative buoyancy. As the delaminated slab sinks, its temperature exceeds the solidus (as it is demonstrated in Anatolia, Turkey) by Keskin et al. (1998) and Pearce et al. (1990))

It has also been realized that subduction-related signal usually vanishes with time after the cessation of subduction or commencement of volcanism in intra plate settings. This is clearly reflected in decreasing LILE/HFSE ratios and increasing HFSE
concentrations with time (i.e., Almandaz et al., 2000; El Bakkali et al. 1998; Fitton et al. 1991; Hawkesworth et al., 1995). This implies that the original enrichment is used up by multiple melting events until the state where the mantle becomes depleted and, henceforth, loses its potential to yield further melt batches. Such phenomena was demonstrated by Till et al. (2007) for the southern Sonora volcanites (Mexico), where volcanism following the cessation of subduction (after a short quiescence) displayed decreasing subduction signal and eventually terminated after a 4 Ma of activity. Another noteworthy observation is that OIB-like mafic alkaline volcanism often follows the paroxysm of volcanism in the mature phase of extension. This can be explained by decreasing degree of melting in the deeper enriched asthenosphere, rather than in the lithosphere (SCLM) (El Bakkali et al., 1998; Fitton et al., 1991). It was suggested by Hawkesworth and Gallagher (1993) that plume and extension related volcanism of enriched lithosphere can be distinguished by their volume and temporal evolution. This is because extension-related volcanism is less voluminous and subduction signature turns into OIB character with time. By comparison, plume volcanism is quiet extensive and generally starts with OIB character.

7. Discussion

7.1. Subduction-related geochemistry, crustal contamination and underplating

Studies of the petrology and geochemistry of the middle Miocene volcanic rocks of the “Western segment” have revealed a clear subduction character (Fig 3 and 4).
Their mantle source appears to be heterogeneous, showing evidence for predominantly enriched mantle components, with EMI dominating over EMII component. The trace element and isotopic compositions of EMI have been interpreted in terms of mixing of various components such as: pelagic and/or terrigenous sediments, sub-continental lithospheric mantle, ancient oceanic plateau, oceanic crust and lower continental crust (Lustrino and Dallai, 2003). The geochemical character of the middle Miocene volcanites in the “Western segment” was interpreted as being due to subduction-related diapiric uprise of a large-volume of partial melts from the metasomatised lithospheric mantle and asthenosphere. These partial melts were thought to have caused underplating and crustal anatexis (i.e., Harangi, 2001b; Harangi and Lenkey, 2007; Seghedi et al., 2005). Crustal contamination of the volcanites is indicated by $^{87}\text{Sr}/^{86}\text{Sr}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ isotope systematics (Harangi, 2001b) and the clear correlation between $^{87}\text{Sr}/^{86}\text{Sr}$ and Th/La ratios. Such isotopic and geochemical characteristics imply that crustal contamination was more significant than aqueous fluids from a subducted slab during the genesis of middle Miocene volcanites (Harangi and Lenkey, 2007). The mixing between crustal and mantle-derived melts led to the formation of middle Miocene volcanites (Harangi, 2001b; Harangi and Lenkey, 2007; Seghedi et al., 2004b; Seghedi et al., 2005). Oxygen isotopic compositions of the middle Miocene volcanites also refer to either source contamination or mixing of mantle-derived magmas with lower crustal rocks (Harangi et al., 2001; Harangi et al., 2006; Seghedi et al., 2004a). Melting in the subduction-enriched mantle, the melting temperature of which was lowered by the presence of volatiles (Green, 1973a; Green, 1973b), was triggered by the first phase of extension in the middle Miocene (17-14 Ma ago). Although the tectonic processes (i.e.,
a roll-back) driving the extension are well constrained (i.e., Horváth, 1993), the subduction which may have caused the enrichment in the mantle still remains matter of intense debate. For the Central Slovakian Volcanic Field (Fig. 1) it is thought that delamination of part of the lithospheric mantle played a further role in shaping the temporal distribution and geochemistry of volcanism (Seghedi et al., 1998; Seghedi et al., 2004a).

As discussed earlier, underplating is well-documented beneath both BBH and NG. Such underplating may be responsible for the granulite-facies metamorphism in the lower crust (both localities are in the vicinity of the middle Miocene volcanites of the “Western segment”, Fig. 1). Furthermore, underplating could have supplied heat for lower crustal melting and played a role in the generation of middle Miocene intermediate and silicic magmas. It is plausible that the large underplated bodies may represent basaltic melts trapped at MOHO levels during the first extension phase in the middle Miocene. The exact timing of these processes is, however, not yet known. The relationship between melt underplating and tectonics is, therefore, somewhat unclear.

7.2. Potential subductions being responsible for source enrichment

The most appropriate candidate responsible for metasomatism beneath the CPR is the former Penninic-Magura Ocean along the Carpathians (Fig. 5, Downes et al., 1992; Downes et al., 1995; Szabó et al., 1992). There may have been intervening, smaller continental blocks within the Penninic-Magura Ocean (e.g. Czorsztyn ridge). This was a Mesozoic ocean of possible Jurassic age, was partially closed in the late Cretaceous (Birkenmajer, 1986) and/or Paleogene-Miocene (Kovać et al., 1994; Meulenkamp et al.,
Evidence for subduction of this ocean can be found in a wide accretionary prism composed of turbidites, i.e., the external Carpathians. The oceanic nature of the basement of these “Flysch nappes” is doubted by several authors (Winkler and Slaczka, 1992). The facts that both this basement disappeared entirely or easily gave place for the overriding Alcapa and Tisza-Dacia microplates during their Paleogene-Miocene movements, suggest that it could have an oceanic basement. The final collision occurred diachronously along the Carpathian arc: early Miocene in the west, middle Miocene in the north-northeast, late Miocene in the east of the arc (Nemcok et al., 1998). Although subduction of the external part of the Magura Ocean might have started in the late Eocene (Oszczypko, 1992), the downgoing slab could barely have reached magma-generating depths by the middle Miocene. This is also supported by Gerya et al. (2002) who stated that at least 24 Ma is necessary for creating extensive metasomatism of a mantle wedge above an active subduction zone. Similarly, Arcay et al. (2006) and Arcay et al. (2005) showed that at least 15 Ma is needed to create substantial hydration and erosion in the mantle wedge (considering various convergent rates). There is also a spatial problem at the time of intense middle Miocene igneous activity, as the subducted slab was too far north (~300 km) from Alcapa (in its reconstructed position) to account for the intense mantle metasomatism (Fig. 5; Kovács et al., 2007). The existence of well-developed subduction beneath the entire Western Carpathians is now also doubted by several geophysical studies as discussed above (i.e., Grad et al., 2006; Szafián and Horváth, 2006; Szafián et al., 1997; Szafián et al., 1999; Wortel and Spakman, 2000). An alternative hypothesis has been proposed, where the lithospheric structure beneath the (the western portion of) Western Carpathians can be explained either by transcurrent
motion of nappes along the orogen without major convergence of the subcrustal
lithosphere (Szafián et al., 1997) or a collisional model implying a “crocodile mouth”-
like structure (Grad et al., 2006). There are multiple models that suggest that subduction
along the Western Carpathian arc explains inadequately the metasomatised nature of the
mantle beneath the “Western segment”.
An alternative possibility for the mantle metasomatism at the origin of these
magmas is that it was caused by an earlier, pre-Miocene subduction outside the present
CPR. Kovács et al. (2007) showed that there is a geochemical and geodynamic link
between the Paleogene and Early Miocene subduction related volcanics of the wider
CPR, including the eastern segment of the Alps and the Dinarides. Kovács et al. (2007)
also proposed that the Paleogene-early Miocene igneous rocks of the Alpine-Carpathian-
Pannonian-Dinaric region formed a uniform volcanic arc mainly along the trench of the
Budva-Pindos Ocean (Fig. 5). Mantle tomography revealed the remnant of this former
subducted slab along the eastern Adriatic shoreline, which is the only well documented
slab remnant in the wider CPR (Koulakov et al., 2002; Lippitsch et al., 2003). Middle
Miocene rocks in the “Western segment” largely overlap with the occurrence of the
Paleogene-early Miocene rocks of the MHZ and share common geochemical characters
(Fig. 3 and 4). Therefore, we conclude that the Budva-Pindos (and Vardar) subductions
are likely the wanted candidates for explaining source enrichment of middle Miocene
volcanites.

7.3. A model for the “inherited” subduction-related character
Kovács et al. (2007) hypothesized that the enriched lithospheric mantle of the Alcapa and Tisza blocks could have been transferred from the former supra-subduction environment to the CPR during extrusion and block rotations in the late Oligocene-early Miocene. This inherited enrichment in the source implies that the uppermost mantle moves together with the crust during major block rotations, and should preserve its metasomatic components from a previous geodynamic setting. Accordingly, Falus (2004) and Hidas et al. (2007) suggested that the Alcapa unit and possibly the Tisza block (Fig. 1) moved together with a portion of their lithospheric mantle on the basis of deformation analysis of mantle xenoliths. They found that shallower ("older") sections of the upper mantle beneath the BBH suffered multiple deformation events, whereas the deeper sections ("younger") show only one strong deformation event most probably related to basin formation. In this scenario the shallower part represents lithospheric mantle which could have been displaced with the crust during the extrusion of Alcapa and Tisza. Although it is still a matter of debate whether the extrusion happened at crustal or lithospheric scale in the CPR, the available data suggest that this process likely happened at lithospheric scale. If so, the question arises at what depth did the decoupling happen in the lithospheric units close to the Budva-Pindos subduction? At this depth the rheology is controlled by amphibole stability (Niida and Green, 1999; Wallace and Green, 1991) because amphibole decomposition leads to a rheological low-point in the mantle where the decoupling is most likely to occur. This depth corresponds to ~ 90 km, which is excellent agreement with the 100 km depth of lithosphere-asthenosphere boundary under Phanerozoic continental areas (Plomerova et al., 2002). This also matches calculation of Falus et al. (2000) who suggested on the basis of mantle xenoliths...
from the Styrian basin (SB), that the maximum thickness of the lithosphere was 90-120 km before the extension events. Rheological considerations and modeling of Willingshofer and Cloetingh (2003) and Willingshofer et al. (1999) also confirm that models show better agreement with field observation in the Eastern Alps if coupling is stronger between the crust and mantle.

Thus, it is likely that the Alcapa block was extruded from the Alpine realm as a lithospheric unit in the late Oligocene-early Miocene. This is also indicated by the preservation of orthopyroxene-rich mantle rocks, subduction-related volatile-rich melts trapped as inclusions in peridotites and the widespread modal metasomatism (i.e., the presence of amphibole and phlogopite) in mantle xenoliths from the BBH and NG. This suggests that the uppermost mantle must have been in or close to a supra-subduction setting. Stable isotopic studies of Demény et al. (2004) and Demény et al. (2005) also support this hypothesis, as some amphibole megacrysts from the mantle show clear results of subduction-related metasomatism beneath the BBH and NG. A recent study of xenoliths from the Styrian basin by Coltorti et al. (2007) (see Fig. 1) also shows evidence for metasomatism by slab-derived melts. In addition, granulite xenoliths beneath the BBH (and NG) may represent remnants of a subducted oceanic lithosphere that were emplaced at lower crustal levels (Dobosi et al., 2003; Embey-Isztin et al., 2003).

Besides the SLCM there may have been contribution from the asthenosphere as well to middle Miocene magmas. It is plausible to assume that not only the lithospheric mantle but also the less viscous, convective asthenosphere may have suffered eastward directed extrusion from the Alpine realm most probably during the late Oligocene-early Miocene. This flow may have happened independently from the lithosphere, and was
parallel to the Alpine orogen (W-E) as it was observed for other orogens with recent convergence (i.e., Himalaya; Flower et al., 2001; Meissner et al., 2002). In this scenario the asthenospheric flow may have played an important role in opening up the Pannonian Basin (and Western Mediterranean basins) and driving subduction roll-back along the Carpathians as well. It is hypothesized; therefore, that the asthenosphere may also have transferred its enriched geochemical signature from the Alpine collision belt to the marginal basins. Unfortunately, seismic anisotropy measurements are not yet available which may give a clearer picture about the potential existence of an eastward directed, orogen parallel anisotropy in the asthenospheric mantle beneath the CPR. However, there may be mantle xenoliths of possible asthenospheric origin that show signs only one deformation event, primitive geochemistry and deeper origin consistent with this model (Falus et al., 2000; Falus, 2004). This model explains OIB character of younger alkaline basalts of the region which are mainly of clear asthenospheric origin.

In this study we propose a similar scenario to that of Kovács et al. (2007) for the formation of the middle Miocene volcanics of the “Western segment” along the MHZ. It is proposed, therefore, that the source of the middle Miocene had been metasomatised along the subduction of either the Budva-Pindos or Vardar Ocean in the Late Mesozoic and Paleogene and then it was transported to the CPR during the eastward extrusion and rotation of Alcapa and Tisza. The available geological record is not yet appropriate to resolve and distinguish the exact contribution of the two proposed oceanic realms to the source enrichment. Furthermore, the exact history of both the Budva-Pindos and Vardar Ocean is still controversial (i.e., Csontos and Vörös, 2004; Pamić et al., 2002; Rosenbaum and Lister, 2005). Formation of middle Miocene igneous rocks was
triggered by the first phase of extension between 17-14 Ma causing extensive melting in
the previously enriched (metasomatised) mantle (Harangi and Lenkey, 2007; Seghedi et
al., 2004a). The middle Miocene volcanics are aligned roughly parallel to the MHZ,
which suggests that this highly fractured and deformed zone may have played a
significant role in the genesis and tapping of the melt formed in the mantle (Csontos and
Nagymarosy, 1998; Kovács et al., 2007). Basaltic volcanism of this period has not yet
been identified on the surface; however large bodies of underplated mafic magmas have
been recognized beneath the BBH and NG in pyroxenite xenoliths (Embey-Isztin et al.,
1990; Kovács et al., 2004; Zajacz et al., 2006) and by geophysical studies (Guterch et al.,
2003). These bodies may represent the trapped basaltic melts, which caused anatexis and
granulite-facies metamorphism of the lower crust, eventually leading to the formation of
middle Miocene mainly calc-alkaline volcanics (i.e., Harangi, 2001b; Harangi and
Lenkey, 2007; Kovács and Szabó, 2005; Seghedi et al., 2005).

Seghedi et al. (2005) pointed out that the geochemical and isotopic data confirm
the diminution of subduction-modified mantle with time, in favor of less affected
asthenospheric OIB-like (EMI) mantle in the “Western segment”. Geochemistry of the
older Paleogene-early Miocene volcanics of the “Western segment” provide evidence for
a stronger enrichment in Th, LIL elements (Fig. 4) which are typical indicators of
sediment and fluid contribution from a subducted slab. On the other hand, the youngest
Plio-Pleistocene alkali basalts of the “Western segment” (Fig. 1, i.e., LHP, BBH, SB,
NG) display only a slight subduction signature with a strong OIB character (Dobosi et al.,
1998; Gméling et al., 2007; Harangi, 2001b; Harangi and Lenkey, 2007; Seghedi et al.,
2004b), whereas middle Miocene volcanites are of transitional character. This declining
subduction signature with time and OIB character of the younger alkaline basalts with their relatively moderate volume exclude plume(s) as potential source of volcanism as it is proposed by Hawkesworth and Gallhager (1993). The OIB character of younger late Miocene to subrecent volcanics, as we demonstrated above, may have been accounted for by eastward-directed, and enriched asthenospheric flow from the Alpine collision belt. In contrast, older middle Miocene volcanites resemble geochemical patterns of magmas formed in intra-plate setting various time after cessation of subduction worldwide (see overview on the inherited enriched mantle source). This implies that extensive metasomatism gained in a subduction setting and, therefore, the potential for melting in the mantle does not last forever, but it vanishes with the increasing degree of melt extraction owing to active plate tectonic processes. In a simple analogy the mantle works like a “rechargeable battery” which is filled up with “melting potential” at subduction zones and than this potential is used up by subsequent episodic melting events. “Melting potential” refers to the depression of the solidus relative to the “dry” mantle, which is dictated by the amount of available volatiles as melt and fluid inclusions, hydrous phases and H in nominally anhydrous minerals (i.e., Gaetani and Grove, 2003; Green, 1971; Green and Liebermann, 1976; Green and Falloon, 2005; Niida and Green, 1999).

This “inherited enrichment” model has the advantage of not raising any space-time problem or requiring contemporaneous subduction. It is also not excluded, bearing in mind the long and adventurous tectonic history of the CPR, that even earlier subduction events with pre-Mesozoic age could have left their fingerprints on the mantle beneath the region (Dobosi et al., 1998; Downes et al., 1992; Embey-Isztin et al., 1993a; Rosenbaum et al., 1997). Given the “inherited enrichment” model, a contemporaneous
(middle Miocene) subduction is not crucial to explain the enrichment in the source region of the middle Miocene volcanic rocks of the “Western segment”. Subduction along Eastern Carpathians could only influence the magma generation by its roll-back effect, triggering significant thinning of the lithosphere.

Some previous studies in the CPR (Harangi et al., 2006; Seghedi et al., 2005) also raised the possibility of “inherited” source enrichment for Tertiary volcanics of the “Western segment”. They also proposed that this source was remelted during athenospheric upwelling, however, they did not attempt to link the proposed inherited enrichment to any former subduction in the CPR. As we showed earlier this “inherited enrichment” model is not unknown, however, its detailed application to the middle Miocene volcanics of the MHZ integrating geophysics, xenoliths and igneous rocks is a new effort.

8. Conclusions

Middle Miocene volcanic rocks in the “Western segment” of the CPR show subduction-related geochemistry, in common with the Paleogene-early Miocene rocks of the same area. Magma generation in the middle Miocene was triggered by contemporaneous extension; however the subduction responsible for the source enrichment has been the subject of intense debate. We propose that the subduction-related geochemical character is an inherited feature, where the metasomatism happened before the Miocene along the subduction of either the Budva-Pindos or Vardar Ocean. The major continental blocks (i.e, Tisza and Alcapa) together with their metasomatised
mantle lithosphere were transported subsequently to the Carpathian embayment during the late Oligocene-early Miocene. Considerable thinning of the lithosphere happened during the first extension phase in the middle Miocene, which induced extensive melting in the already metasomatised mantle. The produced mafic melts were trapped in the lower crust, resulting in massive underplating. The heat flux associated with this underplating process led to granulite-facies metamorphism and anatexis of the lower crust as being represented by partially molten granulite and pyroxenite xenoliths from the “Western segment”. Wide geochemical variations in middle Miocene magmas are due to source heterogeneity and various ratios of mantle and lower crustal melts involved in magma genesis. The inherited enrichment in the source is supported by the diminishing subduction-related geochemistry of the magmas with time. Accordingly, Paleogene arc magmas display very clear subduction overprint on their geochemistry. Subduction character is weaker for the middle Miocene volcanics of the “Western segment” showing only little traces of subduction-induced enrichment in the source. This model also suggests that subduction, which may have been able to supply sediments and volatiles down to the mantle beneath the “Western segment”, may not have developed along the entire Western Carpathians.

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Fig. 1. Major tectonic units and spatial distribution of middle Miocene to subrecent igneous rocks of the wider Carpathian-Pannonian region. The map is modified after Fig. 1 in Kovács et al. (2007). Volcanic provinces as “Western” and “Eastern segments” of the middle Miocene igneous rocks (see text for more details) and localities of some Paleogene-early Miocene rocks are also indicated. Plio-Pleistocene alkali basalts with mantle and lower crustal xenoliths with which the study is concerned are also highlighted (BBH – Bakony-Balaton highland, LHP – Little Hungarian Plain volcanic field, NG – Nógrád-Gömör volcanic field, SB – Styrian basin, ETB – East Transylvanian basin).

Abbreviation for the middle Miocene igneous occurrences of the “Western segment”: B = Börzsöny, Bü = Bükk, C = Central Slovakian Volcanic Field, Cs = Cserhát, M = Mátra, SB = Styrian Basin, SBP = South Pannonian Basin, T = Tokaj, V = Visegrád Mts.

Abbreviations for the major faults are: PAF-Periadriatic fault, BF-Balaton fault, MHF-Mid-Hungarian fault, SF-Sava fault, MBF-Main Balkan fault, VF-Vardar fault.

Fig. 2. Total alkalis vs. silica diagram for middle Miocene volcanic rocks of the “Western segment” (Le Bas et al., 1986). Data are taken from references in Table 1.

Fig. 3. a) Chondrite normalized (Nakamura, 1974) rare earth element pattern of the middle Miocene igneous rocks of the “Western segment”; b) Primitive mantle normalized (McDonough and Sun, 1988) trace element pattern of the middle Miocene igneous rocks of the”Western segment” (data are taken for a) and b) from references in Table 1); c)
Chondrite normalized (Nakamura, 1974) rare earth element pattern of the Paleogene-early Miocene igneous rocks of the Middle Hungarian zone (see Fig. 1 for the location of the Mid-Hungarian zone); d) Primitive mantle normalized (McDonough and Sun, 1988) trace element pattern of the Paleogene-early Miocene igneous rocks of the Mid-Hungarian zone (data are taken for c) and d) from Kovács et al., 2007 and references therein).

**Fig. 4.** a) Ba/Zr vs. Nb/Zr diagram for the middle Miocene igneous rocks of the “Western segment”. Diagram is modified after Pereplov et al. (2005) and data are from references in Table 1.; b) Ba/Zr vs. Nb/Zr diagram for the Paleogene-early Miocene igneous rocks of the Mid-Hungarian zone. Diagram is modified after Pereplov et al. (2005) and data are from Kovács et al. (2007) and references therein; c) Nb/Y vs. Th/Y diagram for the middle Miocene igneous rock of the “Western segment”. Diagram is modified after Seghedi et al. (2005) and data are from references in Table 1.; d) Nb/Y vs. Th/Y diagram for the Paleogene-early Miocene igneous rock of the Mid-Hungarian zone. Diagram is modified after Seghedi et al. (2005) and data are from Kovács et al. (2007) and references therein; e) $^{86}$Sr/$^{87}$Sr vs. $^{143}$Nd/$^{144}$Nd diagram for middle Miocene igneous rocks of the “Western segment”. Data are from references in Table 1.; f) $^{86}$Sr/$^{87}$Sr vs. $^{143}$Nd/$^{144}$Nd diagram for Paleogene-early Miocene igneous rocks of Mid-Hungarian zone. Data are from Kovács et al. (2007) and references therein.

Fields for different kinds of Ocean Island Basalts (OIB) are from Dupuy et al. (1988), Palacz and Saunders (1986) and Weaver (1991) for HIMU, Humphris and Thompson (1983), Storey et al. (1988) and Weaver (1991) for EM1 and Dostal et al.
(1982) and Palacz and Saunders (1986) for EM2, respectively. Compositions of enriched and normal middle ocean ridge basalt (E-MORB and N-MORB) are from McDonough and Sun (1988).

Fig. 5. Shematic summary of major tectonic events in the Carpathian-Pannonian region since the late Eocene from a “mantle perspective”. Main tectonic features as subduction zones, sutures and faults are indicated on the map. Abbreviations are the followings for tectonic units: EA – Eastern Alps, SA – Southern Alps, Ti – Tisza, Da – Dacia, Di – Dinarides, Ap – Apulian indenter; for tectonic lines: PAF – Periadriatic fault, BF – Balaton fault, MHF – Middle-Hungarian fault, ZZF – Zagreb–Zemplén fault, VF – Vardar fault. Please note that ZZF and MHF is basically the same fault, the northern part of which is called as the MHF. Rotations of the Tisza-Dacia block are only schematic; see Patrascu et al. (1994) for more details.
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Table 1. Location, age and petrology of the Middle to Late Miocene igneous rocks in the vicinity of the Middle Hungarian zone

<table>
<thead>
<tr>
<th>Reference</th>
<th>Locality*</th>
<th>Age</th>
<th>Petrology</th>
</tr>
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<tr>
<td>Póka et al., 1988</td>
<td>V, B, Cs, M, C, T</td>
<td>Karpathian-Late Sarmatian</td>
<td>rhyolite, andesite, basalt, dacite</td>
</tr>
<tr>
<td>Salters et al., 1988</td>
<td>T, B, M, Cs</td>
<td>Karpathian-Late Sarmatian</td>
<td>basaltic andesite-rhyolite</td>
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<td>Szabó et al., 1992</td>
<td>B, Bű, C, Cs, M, T, V</td>
<td>Karpathian- Sarmatian</td>
<td>rhyolite, andesite, basalt, dacite</td>
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<td>Pamic et al., 1995</td>
<td>SPB</td>
<td>Karpathian-Late Badenian</td>
<td>trachyandesites, basalts, andesites</td>
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<td>Downes et al., 1995</td>
<td>B, M, T, Cs</td>
<td>Karpathian-Late Sarmatian</td>
<td>andesite, dacite</td>
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<td>Harangi et al., 1995</td>
<td>LHP</td>
<td>Late Miocene</td>
<td>trachybasalt, trachyandesite, trachyte</td>
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<td>Harangi, 2001a</td>
<td>SB</td>
<td>Badenian</td>
<td>trachyandesite, trachyte</td>
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<td>Harangi et al., 2001a</td>
<td>B, C, V</td>
<td>Badenian</td>
<td>andesite, dacite, rhyodacite</td>
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<td>Seghedi et al., 2004</td>
<td>Bű, T</td>
<td>Karpathian-Sarmatian</td>
<td>rhyolite, dacite</td>
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<td>Póka et al., 2004</td>
<td>Cs</td>
<td>Karpathian- Sarmatian</td>
<td>andesite, dacite, rhyodacite, dacite</td>
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<td>Zelenka et al., 2004</td>
<td>buried under Miocene sediments along the MHZ</td>
<td>Karpathian-Sarmatian</td>
<td>andesite, dacite, rhyodacite, dacite, basalt</td>
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<td>Gméling et al., 2005</td>
<td>C, B, V</td>
<td>Badenian</td>
<td>basaltic andesite-andesite</td>
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<td>Harangi et al., 2005</td>
<td>Bű</td>
<td>Karpathian-Sarmatian</td>
<td>rhyolite**</td>
</tr>
</tbody>
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** glass shards and not bulk-rock analyses
Figure

Middle Miocene-subrecent magmatic rocks exposed/subcrop

Plio-Pleistocene alkali basalts

Xenoliths described in the text

Subductions, sutures

Bohemian Promontory

"Western segment"

Penninic subduction

"Eastern segment"

Bohemic Promontory

TISZA-DACIA

ALCAPA units

"Eastern segment"

Western Carpathians

"Western segment"

"Eastern segment"

Magura subduction

Penninic subduction

Budva-Pindos subduction

Vardar suture

Sava-Vardar zone

Subductions, sutures

Etappe Basin

European foreland

Alpine-Carpathian foredeep

Alpine-Carpathian flysch belt

Penninic ophiolites

ALCAPA units

TISZA-DACIA units

Serbo-Macedonian

Southern Alpine, Dinaric, Adriatic platform units

Bosnian flysch, Slovenian trough

Mesozoic mélange and ophiolites

Kovács and Szabó, Fig. 1 (.ai)
Kovács and Szabó, Fig. 2 (.cdr)
Kovács and Szabó, Fig. 3 (.cdr)
Middle Miocene (SiO\textsubscript{2} < 56 wt\%)  
Middle Miocene (SiO\textsubscript{2} > 56 wt\%)  
Paleogene-early Miocene (SiO\textsubscript{2} < 56 wt\%)  
Paleogene-early Miocene (SiO\textsubscript{2} > 56 wt\%)  

Kovács and Szabó, Fig. 4 (.ai)
late Eocene - early Oligocene (~30 Ma)
late Oligocene - early Miocene (~20 Ma)
middle Miocene (~17 Ma)
middle to late Miocene (~12 Ma)

Legend

- Subduction
- Suture zone
- Fault zones
- Proposed fault zones
- Rotations
- Direction of block movements
- Volcanic activity

Kovács and Szabó, Fig. 5 (.cdr)