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# POST-COLLISIONAL TERTIARY-QUATERNARY MAFIC ALKALIC MAGMATISM IN THE CARPATHIAN-PANNONIAN REGION: A REVIEW

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## ***Abstract***

Mafic alkalic volcanism was widespread in the Carpathian Pannonian region (CPR) between 11 and 0.2 Ma. It followed the Miocene continental collision of the Alcapa and Tisia blocks with the European plate, as subduction-related calc-alkaline magmatism was waning. Several groups of mafic alkalic rocks from different regions within the CPR have been distinguished on the basis of ages and/or trace-element compositions. Their trace element and Sr-Nd-Pb isotope systematics are consistent with derivation from complex mantle source regions, which included both depleted asthenosphere and metasomatized lithosphere. The mixing of DMM-HIMU-EMII mantle components within asthenosphere-derived magmas indicates variable contamination of the shallow asthenosphere and/or thermal boundary layer of the lithosphere by a HIMU-like component prior to and following the introduction of subduction components.

Various mantle sources have been identified: Lower lithospheric mantle modified by several ancient asthenospheric enrichments (source **A**); Young asthenospheric plumes with OIB-like trace element signatures that are either isotopically enriched (source **B**) or variably depleted (source **C**); Old upper asthenosphere heterogeneously contaminated by DM-HIMU-EMII-EMI components and slightly influenced by Miocene subduction-related enrichment (source **D**); Old upper asthenosphere heterogeneously contaminated by DM-HIMU-EMII components and significantly influenced by Miocene subduction-related enrichment (source **E**). Melt generation was initiated either by: (i) finger-like young asthenospheric plumes rising to and heating up the base of the lithosphere (below the Alcapa block), or (ii) decompressional melting of old asthenosphere upwelling to replace any lower lithosphere or heating and melting former subducted slabs (the Tisia block).

***Key words:*** Alkalic mafic magmatism, Miocene-Quaternary, Carpathian-Pannonian region, geochemistry, isotopes, melting model.

## ***Introduction***

Our understanding of Miocene-Quaternary alkalic mafic magmatism in the Carpathian-Pannonian region (CPR) has been greatly increased during the last decade by the acquisition of high quality geochemical, isotopic and radiometric data. In spite of this progress, a comprehensive understanding of CPR magmatism is still lacking. Accordingly, to help answer some important questions about the evolution and distribution of alkalic magmatic activity within the CPR and its relation to regional

geodynamics, we synthesize all available published data (Salters et al., 1988; Embey-Isztin et al., 1993; Dobosi et al., 1995; Downes et al., 1995a; Harangi et al., 1995a; Dobosi & Jenner, 1999; Szabó et al., 1992; Embey-Isztin et al., 1993; Embey-Isztin & Dobosi, 1995; Rosenbaum et al., 1997; Harangi, 2001a,b) and new geochemical and Sr and Nd isotopic data from the Styrian Basin (south-eastern Austria) (Appendix).

Generation of alkalic magmas is usually considered to be a response to extension, which allows the asthenospheric mantle to rise up (as thermally anomalous mantle diapirs) and melt (Wilson & Downes, 1991). However, in the CPR, the alkalic magmatism generally followed post-collisional calc-alkaline magmatism that was related to earlier subduction. Therefore, we also include data for the most primitive subduction-related basic rocks from the CPR (Downes et al., 1995b; Mason et al., 1996; Seghedi et al., 2001a) in order to evaluate the role of subduction-affected sources in the asthenosphere. These will be referred to as the calc-alkaline basalts (CAB).

Here we characterize the alkalic magmatism of the CPR in terms of age, location, and geochemical and isotopic compositions. We then define the melting processes that generated the magmas and investigate the possible effects of other petrogenetic processes (e.g. fractional crystallisation, crustal contamination), that could have affected compositions of primary magmas. Finally, we discuss the source characteristics (i.e., depleted, enriched, asthenospheric, lithospheric, plume) in order to assess compositions of the mantle where the alkalic basalts probably formed and determine whether there is any relationship between the mantle sources for the alkalic mafic magmas and the subduction-modified asthenosphere that generated the earlier calc-alkaline magmas.

### ***Regional occurrences and chronology of the mafic alkalic magmatism***

Neogene-Quaternary mafic alkalic volcanism in the CPR generally postdated the calc-alkaline volcanism that characterizes the inner Carpathian arc and Pannonian Basin, except that in the Perşani mountains, which was contemporaneous. Other types of alkalic volcanism (shoshonitic, adakitic, potassic) also occurred, interpreted as being related to specific geodynamic processes (Pécskay et al., 1995a; Harangi, 2001a; Seghedi et al., submitted). The mafic magmatism is distributed across several volcanic fields, detailed below (Fig. 1), and erupted during several different time periods (Pécskay et al., 1995a) (Fig. 2):

a) In the western part of the CPR, alkalic basalts occurred as small individual volcanic centres in several areas. They are Burgenland, Little Hungarian Plain (LHP) and Danube Basin, Styrian Basin (known also as the Graz Basin) and Balaton (Figs. 1, 2). It has a spatial relationship with an older intermediate alkalic to acidic volcanism in the Styrian Basin (Harangi et al., 1995a; Harangi, 2001b). Two main ages have been identified, one at 11.7 - 10.8 Ma mainly in Burgenland and the second at ~6 - 2 Ma (Balogh et al., 1992; Embey-Isztin et al., 1993; Pécskay et al., 1995a; Harangi, 2001b), following the calc-alkaline magmatism in the West Carpathians. Potassic volcanic rocks, ~14 Ma old, were described at Balatonmaria (Harangi et al., 1995b), but are beyond the scope of this paper;

b) In the central part of the Pannonian basin, between the Danube and Tisza rivers (location 5 in Fig. 1) alkalic basalts with ages of 11-8 Ma (Balogh et al., 1986) were identified from boreholes, but no petrological work is available for these rocks. Potassic volcanic rocks of 2.1 Ma were also recovered from boreholes at Bar (Balogh et al., 1986) and studied by Harangi et al. (1995b), but not included in our review;

c) In the Central Slovakian Volcanic (CSV) area (Fig. 1) alkalic magmas were generated after termination of calc-alkaline activity (ca. 9 Ma) ~ 8 Ma to 6.6 Ma, and

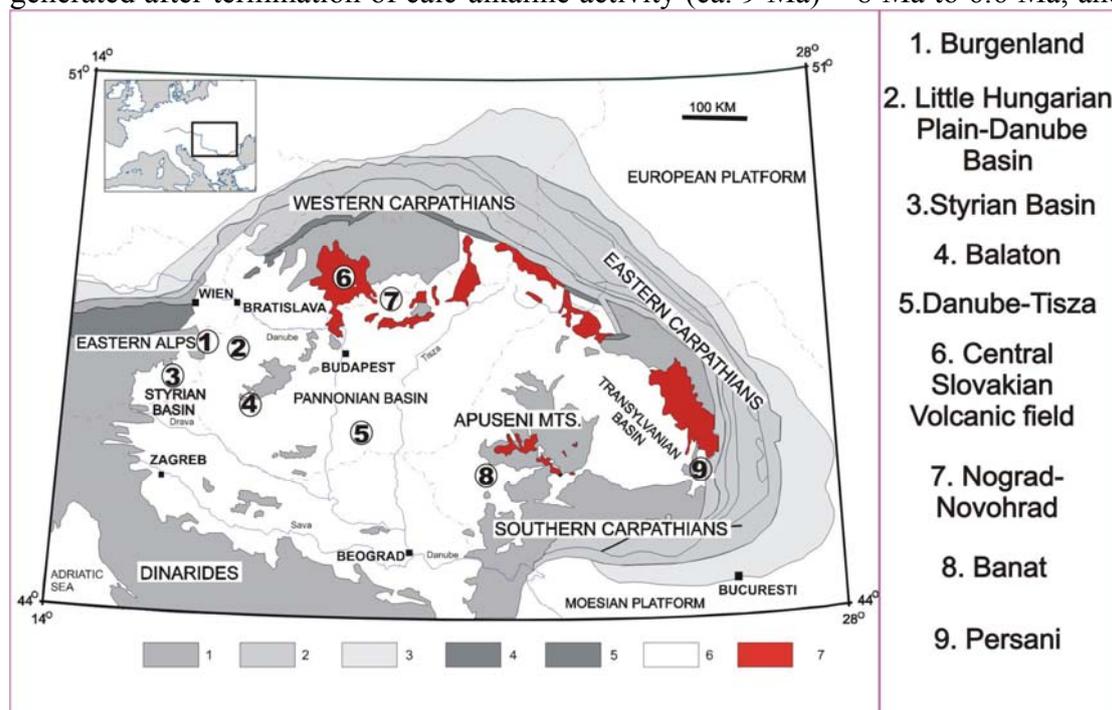


Fig.1. Geological sketch map of Carpathian-Pannonian region showing distribution of the mafic alkalic volcanic areas: 1. Burgenland; 2. Little Hungarian Plain and Danube Basin; 3. Styrian Basin; 4. Balaton; 5. Danube-Tisza; 6. Central Slovakian Volcanic field; 7. Nograd-Novohrad; 8. Banat; 9. Perşani.

Legend: 1. Inner Alpine Carpathian Mountain belt and Dinarides; 2 Alpine-Carpathian Flysch belt; 3. Carpathian Molasse belt; 4. Calcareous Alps; 5. Pieniny Klippen belt; 6. Neogene-Quaternary sedimentary deposits; 7. Outcropping calc-alkaline volcanic rock

a second short pulse occurred much later at ca. 0.5 Ma (Balogh et al., 1981) or 0.13-0.14 Ma (Šimon & Halouska, 1996);

d) In the northern part of the CPR, the Nograd-Novohrad area (Fig.1) (as defined by Pécskay et al., 1995a) is situated ca. 25 km east of the CSV area and the same distance north of the Cserhat-Matra calc-alkaline volcanic area. Here alkalic basaltic volcanism formed small volcanic edifices between ca. 7.2 and 1 Ma, with the largest volume erupted between 2.25 and 1.6 Ma (Konečný et al., 1995, 1999);

e) In the central part of the CPR, a short episode of alkali basaltic volcanism occurred in the Banat area (Fig.1) at 2.5 Ma. Brief shoshonitic volcanism at 1.6 Ma (Pécskay et al., 1995b) occurred close to the Apuseni calc-alkaline adakite-like volcanism (Roşu et al., 2001);

f) In the southern part of the Transylvanian Basin (Perşani mountains) (Fig. 1), alkalic basaltic volcanism occurred between 1.5 and 1.2 Ma, with a second pulse at ~0.6 Ma, coeval with the calc-alkaline adakite-like and shoshonitic magmatism at 1.5-0.5 Ma in the nearby South Harghita Mts. (Seghedi et al., 1987, 2001b; Downes et al., 1995b; Mason et al., 1996).

According to the age distributions (Fig. 2), the older volcanism was restricted to the westernmost part (i.e. Burgenland) at ~12-10 Ma, and younger activity was largely of Pliocene to Quaternary age in the Central Slovakia, Nograd-Novohrad and Perşani areas. The longest continuous interval of volcanic activity occurred in the Nograd-Novohrad, Balaton and LHP areas (Pécskay et al., 1995a).

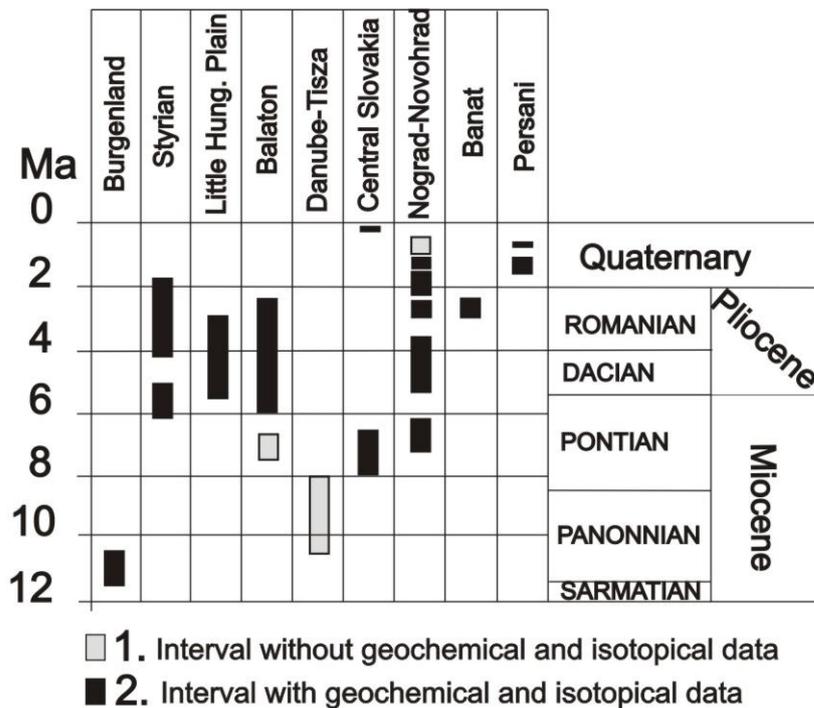


Fig. 2. Diagram of time distribution of mafic alkalic volcanism in Carpathian-Pannonian region. Data from Balogh et al., 1981, 1986, 1994; Pécskay et al., 1995 and Konečný et al., 1999.

### *Geotectonic background*

At the time of generation of the oldest alkalic basaltic magmas in Burgenland, the main tectonic events in CPR area were the collisional kinematics of two microplates, Alcapa and Tisia (or Tisia-Dacia) (Royden, 1993, Horvath, 1993; Csontos, 1995, Fodor et al., 1999). Back-arc extension occurred during Early Miocene block collision due to uneven subduction rollback (Royden & Burchfiel, 1989). Formation of the Styrian Basin was related to Early-Middle Miocene rifting (Tari, 1994). The Little Hungarian Plain and Danube Basin started to develop later, during Middle to late Miocene due to counterclockwise rotations of the Alcapa block (Fodor et al., 1999). In the Lower to Middle Miocene, andesitic to shoshonitic volcanoes erupted in the southeastern part of the Styrian basin (Ebner & Sachsenhofer, 1995; Harangi et al., 1995b). In Late Miocene to Pleistocene times, sporadic alkali mafic volcanism was related to the reactivation of NE-SW or N-S high-angle fault systems during thermal subsidence of the Pannonian basins (Embey-Isztin et al., 1993; Csontos, 1995; Harangi, 2001b).

In the central part of the Pannonian Basin, alkalic magmatism postdated the Middle Miocene crustal thinning, which occurred mostly along the mid-Hungarian belt where the Alcapa and Tisia-Dacia microplates were in contact since the Early Miocene. Opposite rotations of the microplates may have also contributed to crustal thinning (Csontos & Nagymarosy, 1998). In the CSV, as well as in Nograd-Novohrad, generation of alkalic basaltic volcanism was accompanied by limited extensional tectonics (Fodor et al., 1999).

The Pliocene alkali basalts from Banat were generated almost simultaneously with the shoshonites of the South Apuseni area along the reactivated E-W-trending south-Transylvanian fault (Szakács & Seghedi, 1996, Seghedi et al., 1998; Roşu et al., 2001). During this time, important inversion tectonics affected the Pannonian Basin,

followed by renewal of transtensional strike-slip faulting (Csontos, 1995; Fodor et al., 1999).

In the southern part of the East Carpathians, the stress field changed in Pliocene (~5 Ma) times to a NNE-SSW to N-S contraction, coeval with large-scale slab-breakoff and tearing in the southern part of the East Carpathians (Mason et al., 1998; Mañenco & Bertotti, 2000). Breakoff and tearing of the slab at shallow levels (<50 km) in the extreme south of the arc and strike-slip tectonics, accompanied by an extensional stress regime at the surface of the upper plate (Gîrbacea & Frisch, 1998; Ciulavu, 1998; Mañenco & Bertotti, 2000), caused contemporaneous eruption of adakite-like calc-alkaline, shoshonitic and alkalic basaltic magmas (Mason et al., 1998; Seghedi et al., 2004).

### *Geochemical and isotopic evidence for the mantle-source components*

#### *Summary of major, trace element and REE data*

Most CPR compositions center on the basanite and trachy-basalt fields of the TAS diagram (Fig. 3), and some plot in the foidite (nephelinite), basalt, trachy-basalt and basaltic trachy-andesite fields (Embey-Isztin & Dobosi, 1995; Harangi, 2001b). Figure 3 also includes four Neogene subduction-related calc-alkaline basalts (CAB) from the CPR.

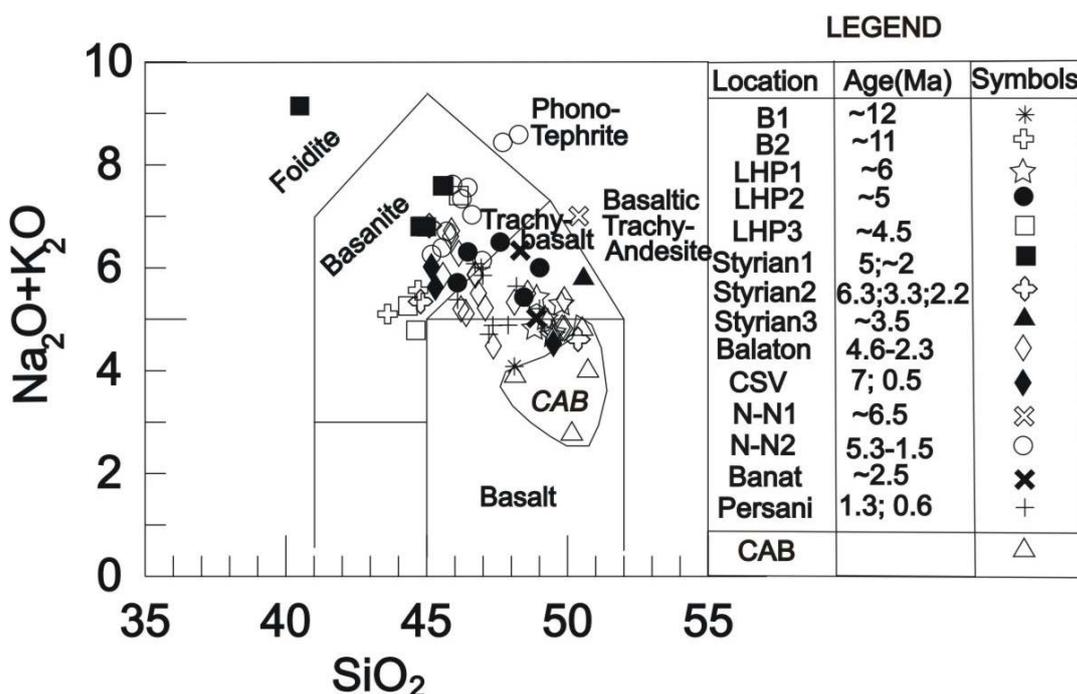


Fig. 3. TAS diagram for alkali mafic volcanic rocks and selected calc-alkaline basalts from the Carpathian-Pannonian region (CPR), showing the legend of symbols. Data from Salters et al., 1988; Embey-Isztin et al., 1993; Dobosi et al., 1995; Downes et al., 1995a; Harangi et al., 1995; Dobosi & Jenner, 1999; Downes et al., 1995b; Mason et al., 1996; Seghedi et al., 2001a. Abbreviations: B = Burgenland; LHP = Little Hungarian Plain; CSV = Central Slovakian Volcanic area; N-N = Nograd-Novohrad.

Calc-alkaline basalts (CAB) from the CPR.: 134t (9.4 Ma) from north-east Hungary (Downes et al., 1995b), UA34 (~13 Ma) from the Ukrainian Carpathians (Beregovo area) (Seghedi et al. 2001a; Pécskay et al., 2000), C64 (~8 Ma) from the East Carpathians (Călimani mountains) (Mason et al., 1996; Pécskay et al., 1995), Lgf (2.14 Ma) from the Perșani mountains (Downes et al., 1995a).

We identify several groups that correspond to the various volcanic fields (Fig. 3). Samples from the LHP volcanic field are shown with different symbols (LHP1, 2, 3) representing three succeeding age intervals (~6; ~5; ~4.5 Ma) as they also show different compositions. Based on their different compositions or ages, we have separated several groups of basalts, basanites and nephelinite from Burgenland1, 2, Styrian Basin (Styrian1, 2, 3) and Nograd-Novohrad 1, 2.

The incompatible trace-element compositions of the different groups are shown normalized to primitive mantle values (Fig. 4), with average OIB values (acc. Sun & McDonough, 1989) shown for reference. Several groups are distinguished according to trace element patterns and sample locations:

(A) Significant and sometimes variable enrichment in incompatible elements (large ion lithophile elements-LILE), light rare earth elements (LREE), high Nb along with a significant relative depletion of K. This pattern is found in Burgenland 2, Styrian 1 and LHP3 samples.

(B) Similar to the average OIB pattern, with minor troughs or spikes, although sometimes with greater depletion of heavy rare earth elements (HREE). They are Burgenland1, LHP1 basalts and some CSV samples.

(C) Nograd-Novohrad basalts with a parallel pattern to that for average OIB, but with variable enrichment in K, Nb and La. However, a slight negative K spike along a higher Nb characterizes the ~6.5 and 2.6 Ma age groups of Dobosi et al. (1995) (Fig. 4).

(D) An enriched pattern of LILE, including LREE, with variable enrichment in LILE. (i.e. Ba and Pb and sometimes Sr) and variable depletion of HREE, Zr, P and Ti. This pattern is characteristic for samples from Styrian2 (however, two samples do not show Pb enrichment), LHP2, Perşani and Balaton. Basalts from Banat and CSV show the smallest enrichment. The most extreme enrichment is in Perşani samples (Sărata and Hoghiz) which show high LREE and K depletion, as well as the highest Ba and Pb. These patterns are similar to those of contemporaneous calc-alkaline adakite-like rocks from the nearby South Harghita volcanic area (Mason et al., 1996; Seghedi et al., submitted) (Fig. 4).

(E) Calc-alkaline basalts (CAB) show characteristic variable negative Nb, REE, and P-depletion along with positive spikes of Ba, and strong positive K and Pb anomalies. A similar pattern is shown by Styrian3 basalt (Nh5-Neuhaus).

Chondrite-normalized REE data show parallel patterns, but different levels of REE enrichment in each group (Fig. 5). There are different patterns for Burgenland1 and Burgenland2, and a continuous increase in almost parallel patterns from LHP1 to LHP3. The pattern for the Styrian1 nephelinite is almost parallel to that of Burgenland2 basanite. Exception is found in samples from the Perşani mountains, where the HREE show a parallel pattern for all the samples and LREE display a gradual increase, highest for the Sărata sample.

#### *Isotopic characteristics of the mantle source*

Figure 6 (a,b,c,d) shows the isotopic characteristics of CPR mafic alkalic rocks relative to the mantle components defined by Zindler & Hart (1986) and the EAR (European Asthenospheric Reservoir) defined in Cebria & Wilson, (1995) and Granet et al. (1995). EAR is considered to be an isotopically distinct mantle plume component beneath Europe (Wilson & Patterson, 2001) and falls on a mixing line between DMM and HIMU. The isotopic data for the CPR mafic alkalic rocks separate into several groups some of which lie relatively close to the EAR composition.

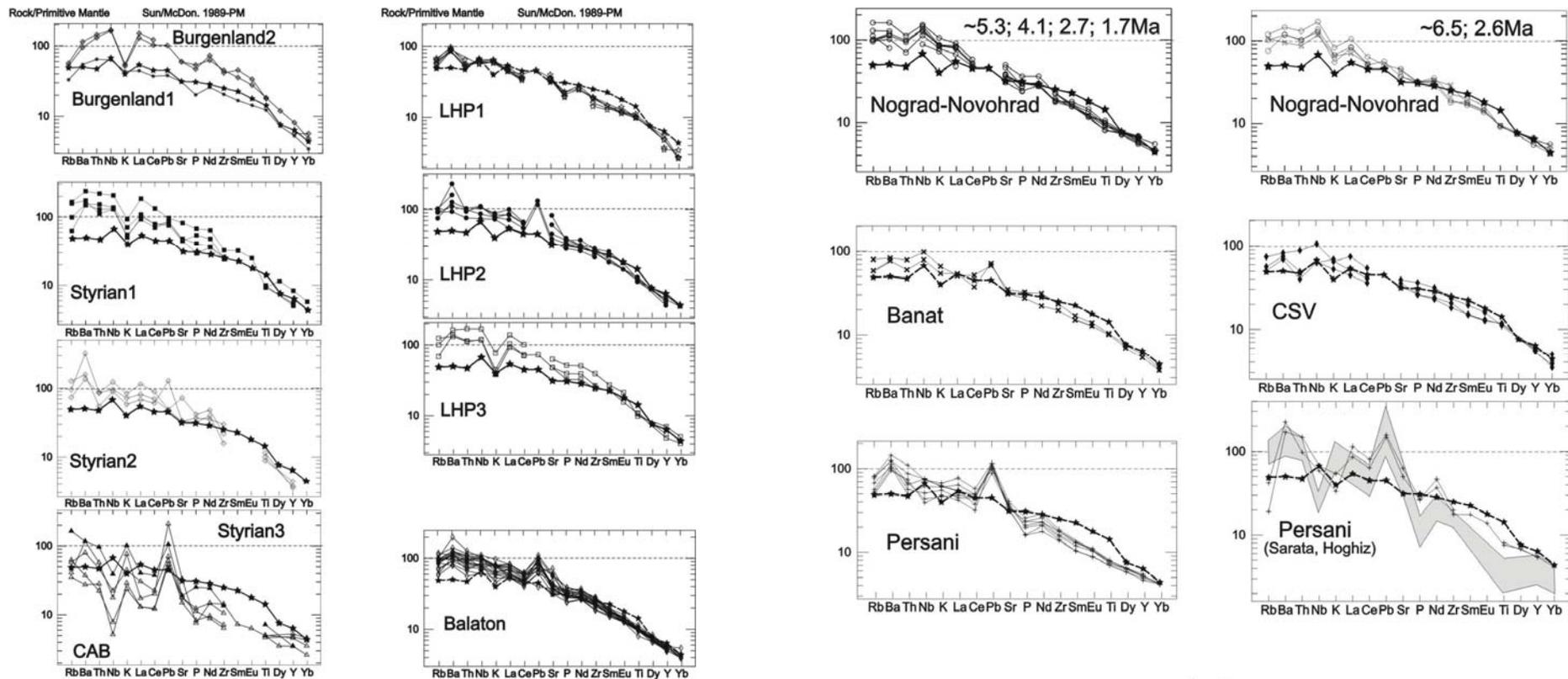


Fig. 4a

Fig. 4b

Fig. 4a,b. Primitive mantle-normalized element diagrams for all available mafic alkalic rocks and selected calc-alkaline basalts in Carpathian-Pannonian region based on volcanic distribution and composition, using the normalizing coefficient of Sun & McDonough (1989); Symbols and source of the data as in Fig. 2. Average OIB values (acc. Sun & McDonough, 1989) plotted with black stars symbol and are shown for reference. Shadow area, in the diagram showing Sărata and Hoghiz samples from Perșani, represents contemporaneous calc-alkaline adakite-like rocks from the nearby South Harghita volcanic area (Mason et al., 1996; Seghedi et al., submitted).

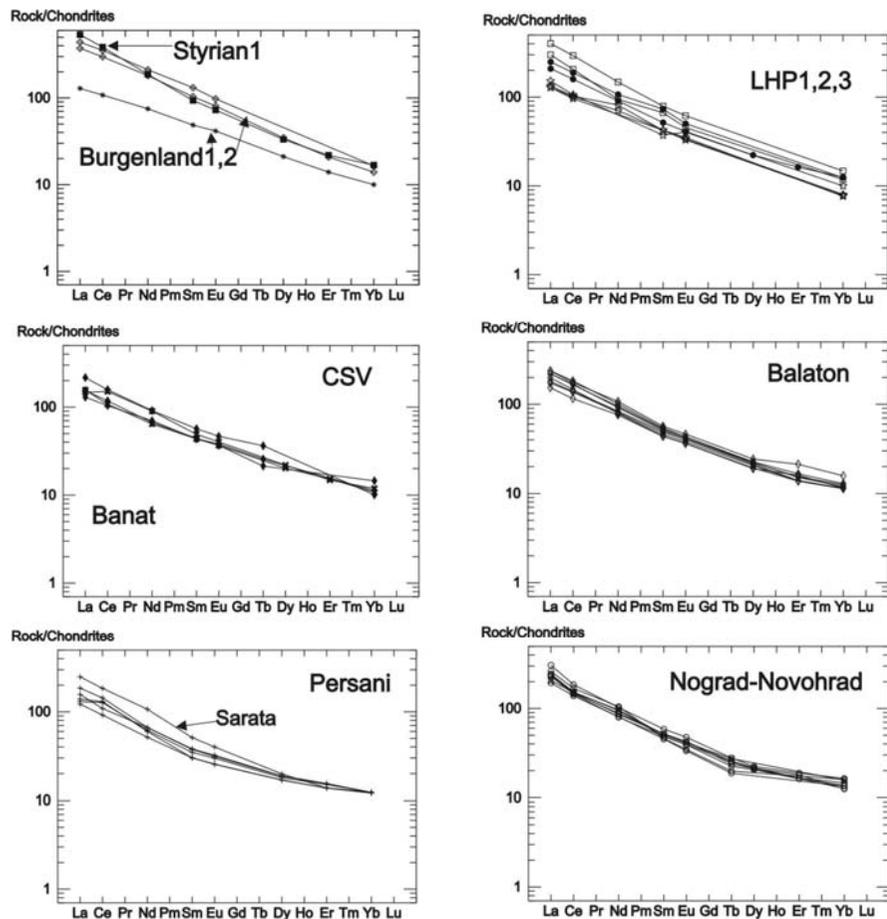


Fig. 5.(left) Chondrite-normalized rare earth element diagram for available mafic alkalic rocks in Carpathian-Pannonian region. Normalizing coefficients from Sun & McDonough, 1989; Symbols and source of the data as in Fig. 2.

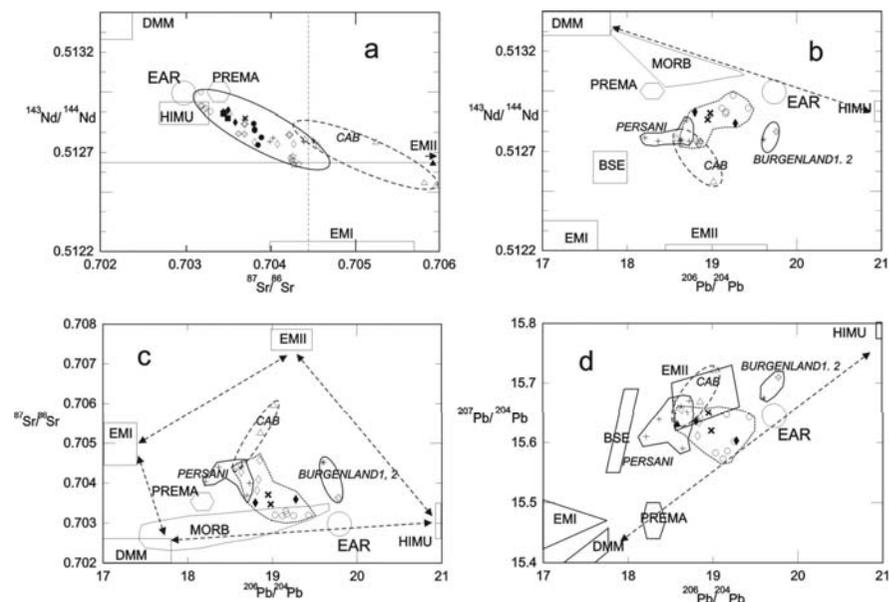


Fig. 6a, b, c, d.(above) Isotopic characteristics of Carpathian-Pannonian mafic alkalic rocks and selected calc-alkaline basalts relative to the mantle components defined by Zindler & Hart, (1986) and the EAR (European Asthenospheric Reservoir) defined in Cebria & Wilson, (1995) and Granet et al., (1995). Symbols and source of the data as in Fig. 2.

(i) The first group, two samples from the ~11Ma Burgenland area (Burgenland1, 2), is closest to the EAR and trends toward EMII (Fig. 6 b, c, d). (ii) The second group is of mafic lavas of the Pannonian Basin, all of which are <8 Ma old. These also trend toward EMII, but with a larger proportion of the DMM component than the EAR. (iii) The third group is 1.5-0.6 Ma basalts from the Perşani mountains, which show a multi-component composition, implying the presence of EMI and EMII components as well as DMM. (iiii) The fourth group is the CAB, which intersects the 3<sup>rd</sup> mafic alkalic group and trends toward EMII. It appears that the first group (comprising older magmas) originated from an EAR component toward EMII, whereas the second group originated from a different initial component similar to the EAR in that it is a mixture between HIMU and DMM, but it has a greater contribution from the DMM component (Fig.6 b,c,d). The third group is distinctive by its larger involvement of an EMI component.

In comparison with Western and Central European magmatism (Fig. 7a,b), the first group of CPR mafic alkalic magmas resembles mafic magmas from Poland and France, whereas the second group of CPR mafic magmas shows some similarity with the main group of alkaline magmas from France and Germany (Woerner et al., 1986; Blusztajn & Hart, 1989; Wilson & Downes, 1991; Granet et al., 1995; Wedepohl & Bauman, 1999; Wilson & Patterson, 2001). Suggestion for the involvement of an EMI component (Perşani mountains basalts) is a peculiarity of CPR.

## ***Discussion***

### *Partial melting*

Embey-Isztin & Dobosi (1995) and Harangi (2001b) determined that the genesis of CPR alkalic basalts resulted from 0.5-2% degree of partial melting of a garnet-lherzolite source in the upper asthenosphere beneath the region.

We use REE abundances and ratios for modeling partial melting processes and to constrain source characteristics. We adopted the non-modal batch melting equation of Shaw (1970), using the method of Albarede (1995) and REE partition coefficients from McKenzie & O'Nions (1991). For example the La/Sm ratio is not affected by source mineralogy and therefore provides information on the bulk geochemical composition of the source. Also, Sm/Yb, an incompatible element related to an element compatible with garnet, was used to constrain the source mineralogy and degree of partial melting. Figure 8 displays reference compositions representing possible asthenospheric mantle sources (N-MORB and PM, Sun & McDonough, 1989). Our modelling attempt shows that only PM may partially account for the source of magmas in the CPR. In fact a PM source for the CPR alkalic basaltic magmas can only be accepted considering a significant enrichment in highly incompatible elements. As a consequence, by extrapolation of melting trajectories and applying the inversion method of Zou (1998) we designate a new source, here named Enriched Asthenospheric Mantle (EAM) (La = 1.5 ppm, Sm = 0.7 ppm, Yb = 0.8 ppm) which is very similar to the Western Anatolian Mantle (WAM) composition proposed by Aldanmaz et al. (2000).

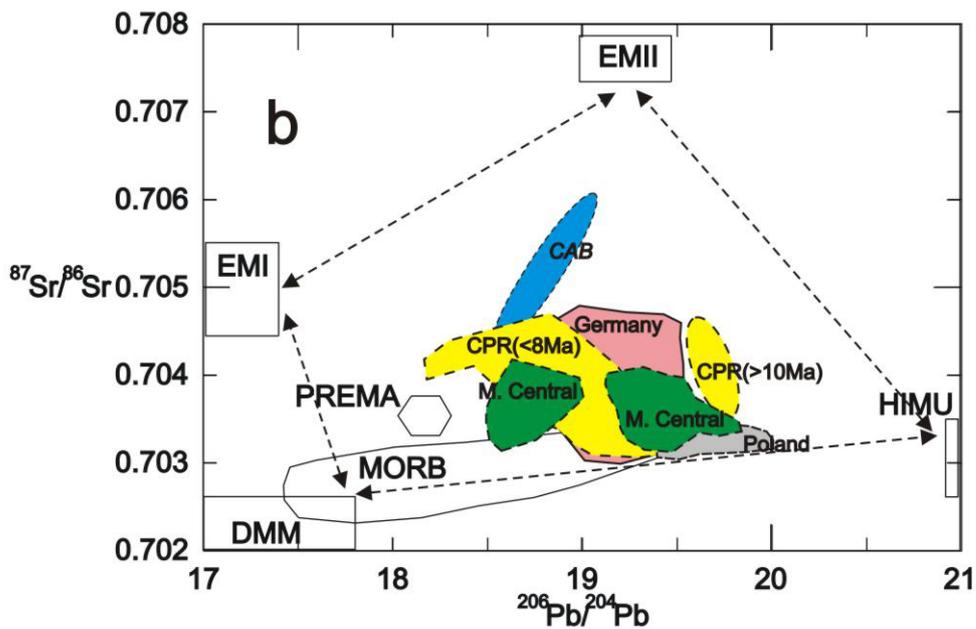
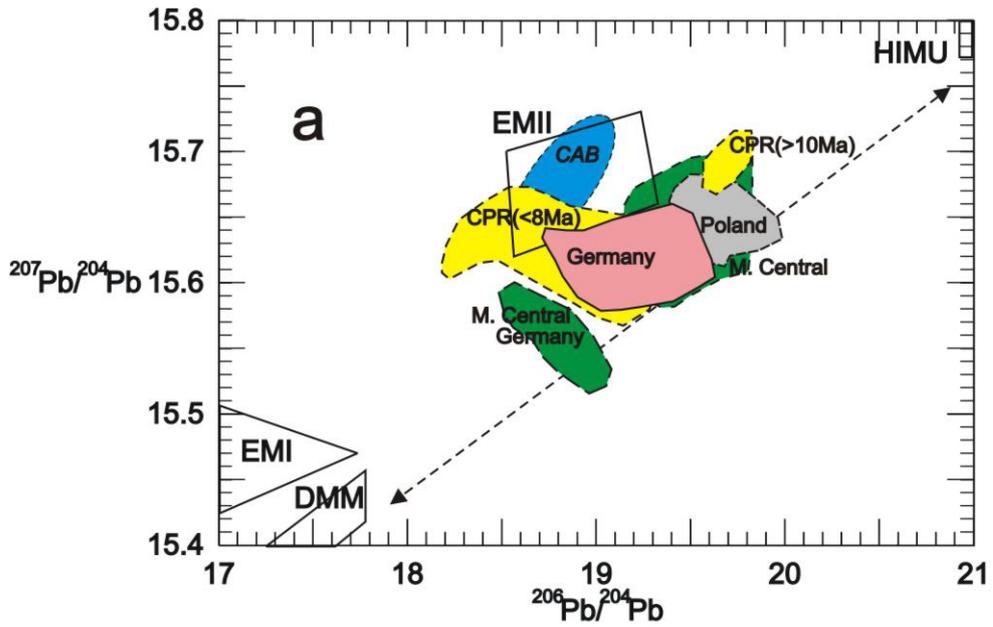


Fig. 7a, b.  $^{206}\text{Pb}/^{204}\text{Pb}$  -  $^{207}\text{Pb}/^{204}\text{Pb}$  and  $^{206}\text{Pb}/^{204}\text{Pb}$  -  $^{87}\text{Sr}/^{86}\text{Sr}$  diagrams of available data from mafic alkalic rocks and selected calc-alkaline basalts in Carpathian-Pannonian region in comparison with western and central European alkali mafic magmatism (France -Masif Central, Poland,, Germany). Symbols and source of the data as in Fig. 2. Additional data from Wörner et al., 1986; Blusztajn & Hart, 1989; Wilson & Downes, 1991; Wedepohl & Bauman, 1999.

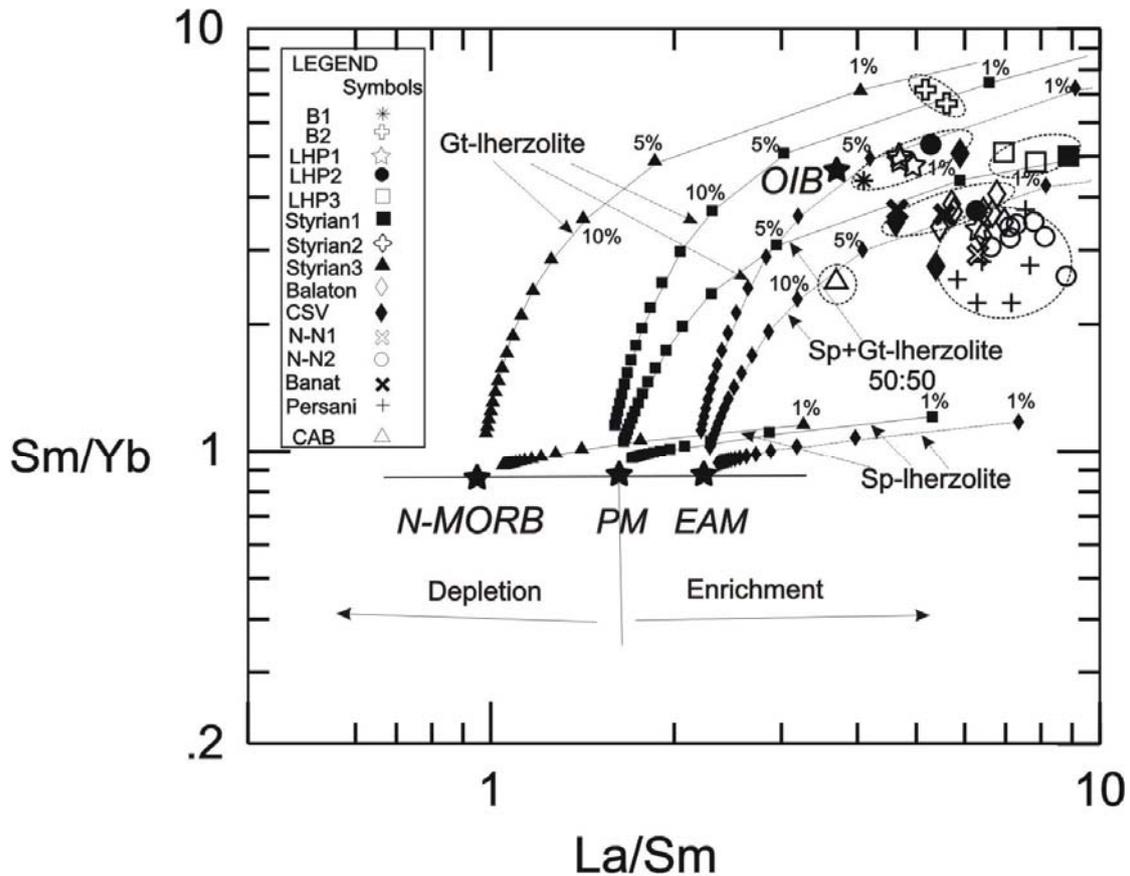


Fig. 8. Log Sm/Yb vs. log La/Sm diagram showing melting curves obtained using the non-modal batch melting equation of Shaw, (1970) and the method proposed by Albarede, (1992). Melt curves are drawn for spinel Iherzolite and garnet Iherzolite with melt mode of  $ol_{0.53}+opx_{0.27}+cpx_{0.17}+sp_{0.03}$  and  $ol_{0.60}+opx_{0.20}+cpx_{0.10}+gt_{0.10}$ . Mineral-matrix partition coefficients are from McKenzie & O'Nions, 1991. EAM represents an Enriched Asthenospheric Source defined by using melt inversion methods (Zou, 1998); PM = primitive mantle; N-MORB = average value of mid oceanic ridge basalts. N-MORB, PM and OIB after Sun & McDonough, (1989). Symbols and source of the data as in Fig. 2. Abbreviations: B = Burgenland; LHP = Little Hungarian Plain; CSV = Central Slovakian Volcanic area; N-N = Nograd-Novohrad. Symbols and source of the data as in Fig. 2.

The results of our modeling confirm significant variation in source mineralogy and point mostly to the garnet and garnet + spinel stability fields. Partial melting trajectories from PM and EAM sources yield a degree of partial melting between 6 and 0.6 %. The larger degree of partial melting is suggested for Burgenland1 (~6%) and LHP1 (~4%) along the garnet Iherzolite trajectory of EAM, whereas the lower degrees of melting are suggested for the Burgenland2 samples (~2%) along the garnet Iherzolite trajectory of PM, or LHP3 (0.8-0.9%; 1-2 %) and the Styrian1 nephelinite (0.6-0.7%; 0.9-1%) along the garnet-spinel Iherzolite trajectory of PM or EAM, respectively. Most of the other basalts (Banat, Balaton, CSV, Nograd-Novohrad, Perşani) follow the garnet-spinel Iherzolite trajectory of EAM with 1-4% partial melting. However, most of the Perşani mountains samples and some of N-N show scatter, implying processes other than partial melting (fractional crystallization, mixing). The compositions of CAB (e.g. sample Lgf) suggest larger proportions of partial melting (7-8%) along the garnet-spinel Iherzolite trajectory of EAM, compared to the mafic alkalic magmas.

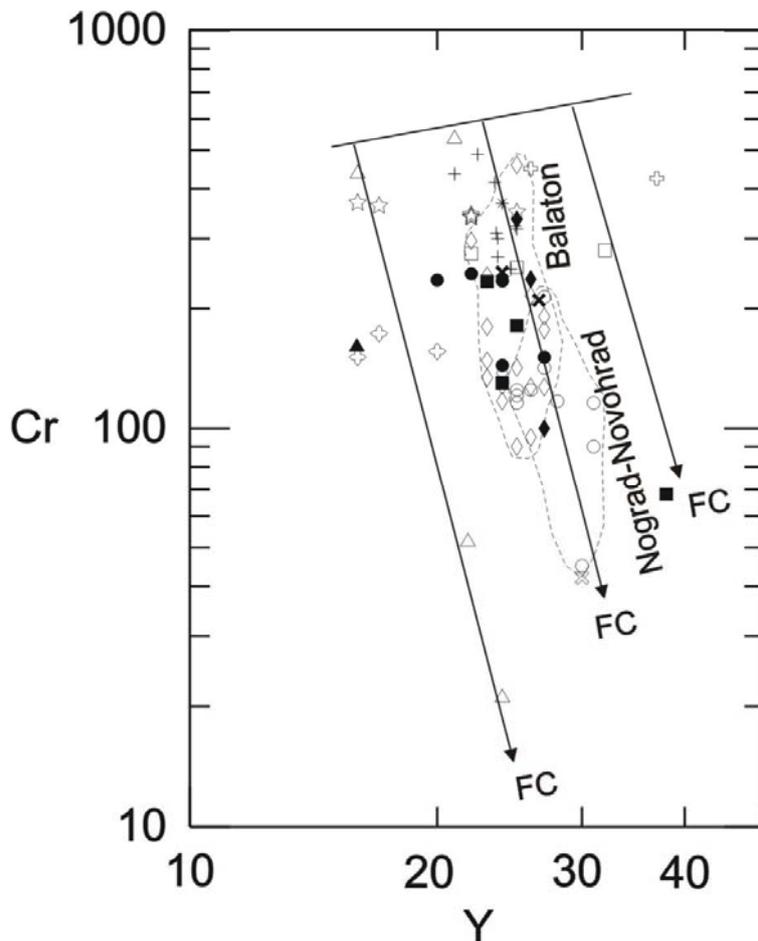


Fig. 9. Log Y vs. log Cr diagram for mafic alkalic rocks and selected calc-alkaline basalts in Carpathian-Pannonian region. Symbols and source of the data as in Fig. 2. FC = fractional crystallization

#### *Fractional crystallization processes*

Fractionation in CPR alkali basaltic magmas was demonstrated in Nograd-Novohrad basalts by Dobosi et al. (1995) and Dobosi & Jenner (1999). Fig. 9 suggests that the Nograd-Novohrad magmas are the most affected by fractionation. Variable Y might reflect partial melting, whereas variable Cr indicates fractional crystallisation. As the highest Cr contents of Nograd-Novohrad basalts do not exceed 200 ppm, it is clear that most of these magmas fractionated before rising to the surface, at about 30 km, at the crust-mantle boundary (Dobosi & Jenner, 1999). Differentiation processes are evident in basalts from other areas such as Balaton, but there the trend starts from high Cr contents (400-500 ppm) characteristic for primitive magmas. Fractionation also affected Styrian Basin and CSV magmas. Compositions of two of the CAB also suggest differentiation process.

#### *Crustal contamination*

According to the Nd-Sr isotopic systematics, LHP1, Burgenland basalt, Styrian3 basalt and some of the Balaton and Perşani basalts have enriched signatures, which could suggest a crustal contaminant (Figs. 6, 12). However,  $\delta^{18}\text{O}$  isotope ratios in clinopyroxene and olivine phenocrysts (5.03-5.2 ‰) in most of these alkali basalts, are close to the value of the N-MORB mantle and do not support crustal assimilation processes (Dobosi et al., 1998). Also, the “primitive” major and trace element distribution of these “enriched” magmas are contradictory to a crustal contamination

history (Downes et al., 1995; Embey-Isztin & Dobosi, 1995; Harangi et al., 1995; Harangi, 2001b).

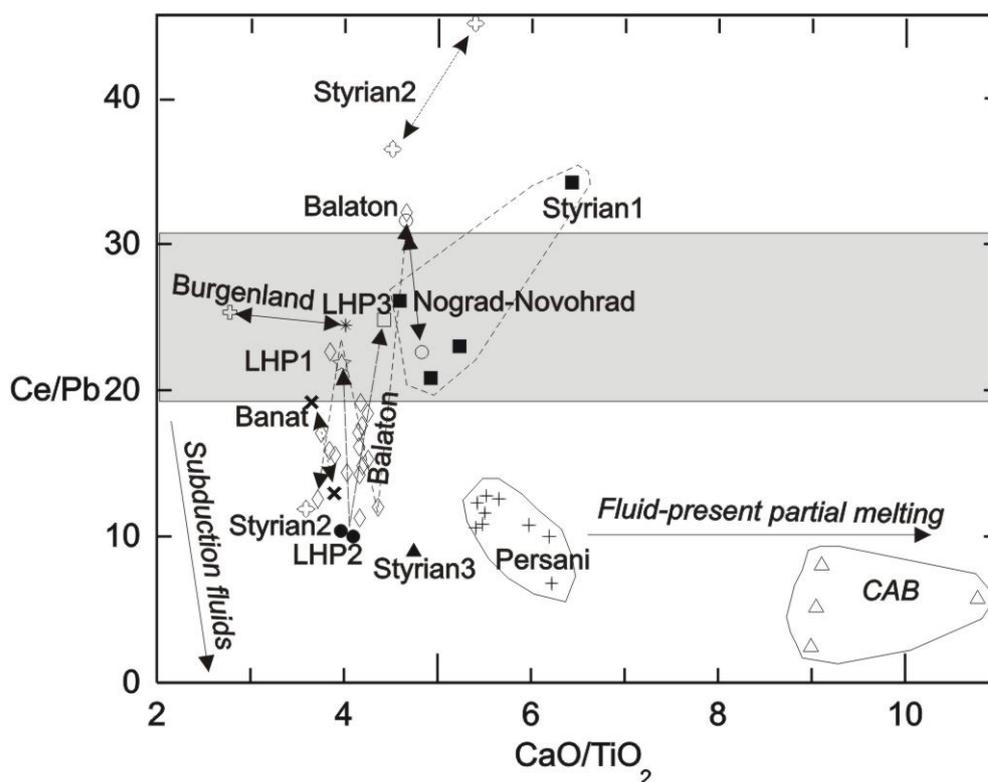


Fig. 10.  $\text{CaO}/\text{TiO}_2$  vs.  $\text{Ce}/\text{Pb}$  diagram for mafic alkalic rocks and selected calc-alkaline basalts in Carpatho-Pannonian region. Shadow area represents the field of mantle-derived magmas ( $\text{Ce}/\text{Pb} \sim 20\text{-}30$ ) according to Hofman et al., (1986). Symbols and source of the data as in Fig. 2.

#### *Subduction-modified mantle*

Besides differences related to different degrees of partial melting and fractional crystallization, there is evidence for chemically diverse mantle sources involved in the generation of CPR alkali basaltic magmas. The diversity mainly resulted from the previous introduction of subduction components, mainly Ba and Pb, which are soluble in aqueous slab-derived fluids (Szabo et al., 1992; Downes et al., 1995b; Mason et al., 1996; Seghedi et al., 2001a) (Fig. 4).

In order to discuss the subduction component in the mantle we use the  $\text{Ce}/\text{Pb}$  vs  $\text{CaO}/\text{TiO}_2$  diagram (Fig. 10). Ce and Pb both show similar bulk distribution coefficients that are not affected by fractional crystallisation at most slightly affected by the degree of partial melting (Sims & De Paolo, 1997). Since the degree of partial melting in CPR alkalic mafic magmas is low, this effect is negligible (Fig. 8). For mantle-derived magmas, the  $\text{Ce}/\text{Pb}$  ratio is 20-30 (Hofman et al., 1986), which is characteristic for most of the mafic alkaline rocks of Western Europe (e.g. Wilson & Downes, 1991; Jung & Hoernes, 2000). However, low  $\text{Ce}/\text{Pb}$  ratios may be due to incorporation of Pb into the asthenosphere via subduction fluids, and  $\text{Ce}/\text{Pb}$  values  $<20$  therefore suggest the influence of subduction-related fluids within the asthenosphere.  $\text{CaO}/\text{TiO}_2$  ratio increases as a consequence of increasing degree of partial melting and/or the presence of fluids (Tatsumi & Eggins, 1995). Figure 10 shows that the region of the CPR mantle most affected by the subduction component is the Styrian3 basalt and the Perşani area, whereas the Balaton, Banat and LHP2

magmas are less affected. The mantle beneath Burgenland and Nograd-Novohrad seems to have been unaffected by the Miocene subduction fluids. High Ce/Pb values of two samples of Styrian2 suggest enrichment processes that probably arose in the lithosphere. The subduction component in the mantle can also be highlighted using the La/Nb vs. La diagram (Fig. 11).

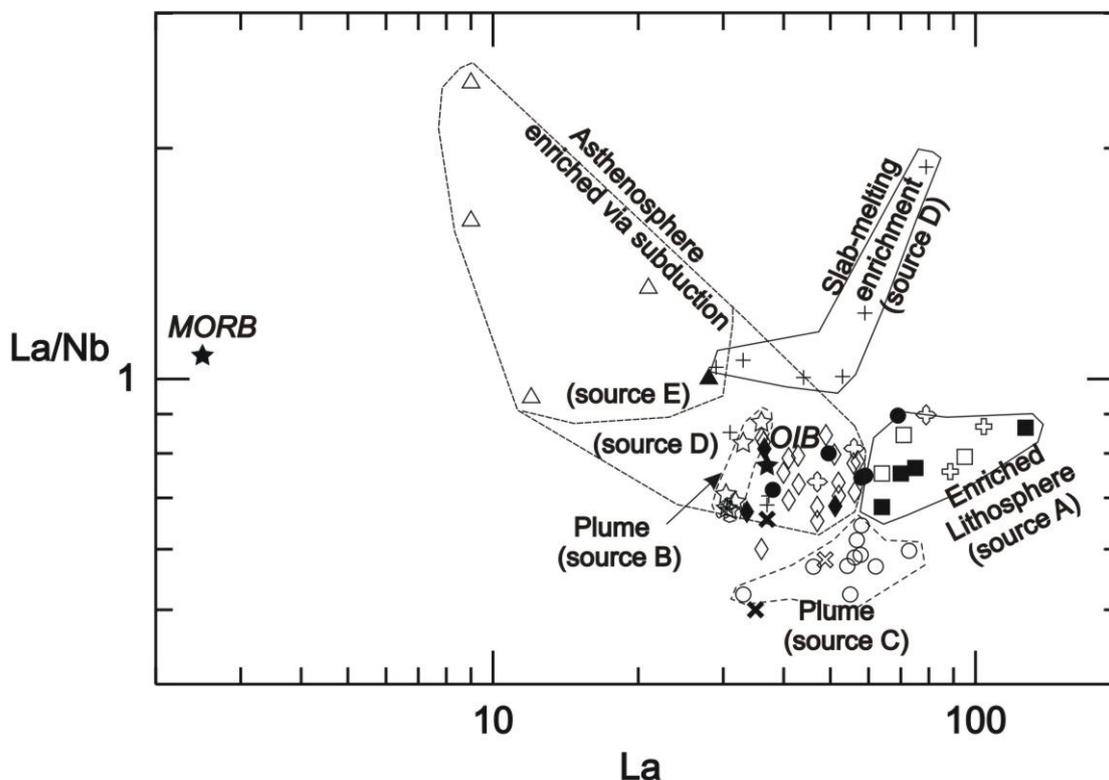


Fig. 11. Log La vs. log La/Nb diagram for mafic alkalic rocks and selected calc-alkaline basalts in Carpatho-Pannonian region showing the main mantle source regions. MORB and OIB after Sun & McDonough, (1989). Symbols and source of the data as in Fig. 2.

#### *Asthenospheric vs. lithospheric reservoirs*

A striking feature of the compositions of alkali basalts from Burgenland and LHP, when normalized to primitive mantle values, is the existence of two types of magmatic reservoirs: **(A)** enriched, but showing a significant relative depletion of K and **(B)** rather similar to average OIB (Fig. 4a). The enriched reservoir can be modeled as an amphibole-bearing source (Embey-Isztin & Dobosi 1995; Harangi (2001b); Späth et al., 1996, 2001; Class & Goldstein, 1997). However, amphibole is unstable in the upper asthenospheric mantle and in thermal mantle plumes, but is stable in the lithospheric mantle, where it is thought to be generated by asthenosphere-derived metasomatic melts (Class & Goldstein, 1997). Lavas which show the enriched pattern (Burgenland2 basanites, LHP3 basalts and Styrian1 basanites and nephelinite) (Fig. 4a) are therefore interpreted to have their origins in the lithosphere, which was metasomatized by asthenosphere-derived melts in earlier events (Rosenbaum et al., 1997).

La/Nb ratio will separate subduction-modified asthenospheric sources since Nb is variably depleted in subduction-modified mantle. La is an indicator of enrichment processes, either is fluid-dependent via subduction (asthenosphere) or via asthenospheric metasomatism (lithosphere) (Fig. 11). CAB are clearly separated

(source **E**), as well as the transition to relatively volatile-poor conditions of the mantle source (source **D**). Perşani mountains (source **D**) are distinct in having high La/Nb ratios, but also higher La contents than any other CAB. Lower lithospheric mantle affected by ancient asthenospheric enrichments (source **A**) is also well constrained by its high La (>60 ppm) and low La/Nb ratio. Source **C**, defining the Nograd-Novohrad basalts, plot also distinctly at the lowest La/Nb ratio.

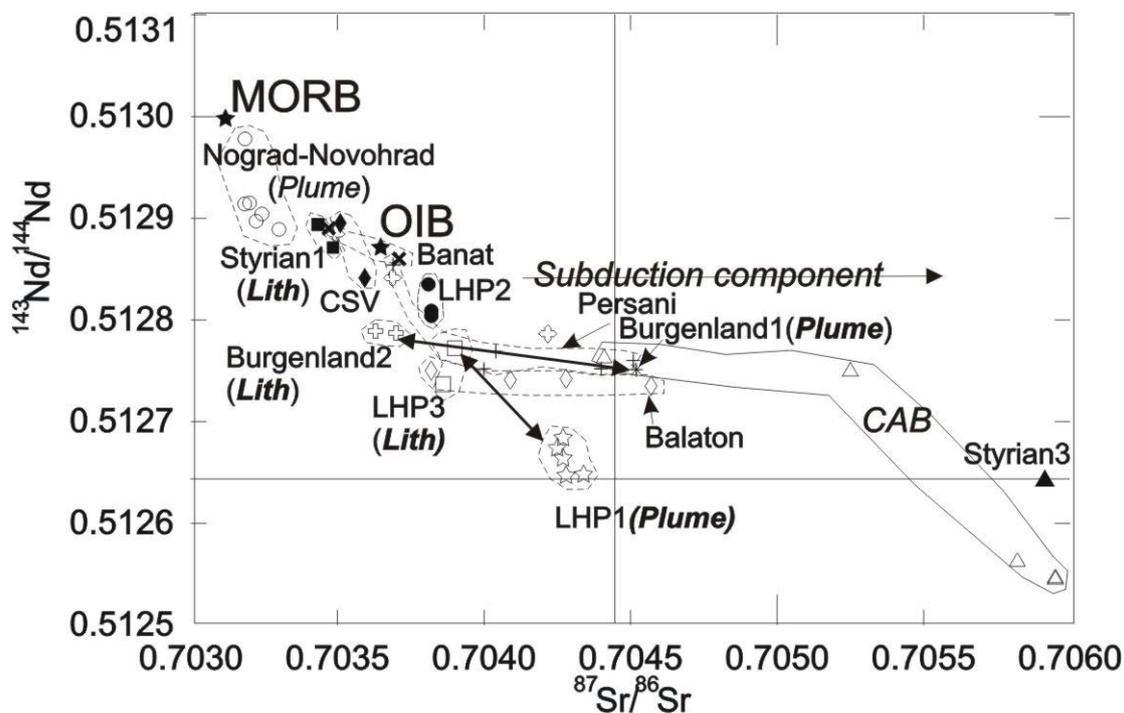


Fig. 12.  $^{87}\text{Sr}/^{86}\text{Sr}$  vs.  $^{143}\text{Nd}/^{144}\text{Nd}$  variation of mafic alkalic volcanic rocks and selected calc-alkaline basalts from the Carpathian-Pannonian region, showing the source region of the magmas. Symbols and source of the data as in Fig.2.

The enriched lithospheric sources (**A**) are rather isotopically depleted compared with the asthenospheric OIB-like ones (**B**), as Nd and Sr isotopic characteristics show (Fig. 12). We explain the enriched isotopic characteristics of **B** sources as reflecting the composition of new asthenospheric mantle plume material (Fig. 12). Class & Goldstein (1997) and Class et al. (1998) documented a similar situation for the Hawaiian Islands and Comores. Class & Goldstein (1997) suggested that small degree melts from the metasomatized lithospheric mantle (in our case Burgenland2, LHP3 and Styrian1 samples) are formed by conductive heating by the plume and these may interact with larger degree plume-derived melts (e.g., Burgenland1 basalts; LHP1). This hypothesis is in accordance with our previous suggestions of a larger degree of melting for Burgenland1 basalts (~6 %) and LHP1 (~4 %) (**B**) and a smaller degree for Burgenland2 basanitic and LHP3 magmas (< 1 %) (**A**) (Fig. 8). This plume model is also viable in the Nograd-Novohrad area (source **C**), the most isotopically depleted, which was not affected by a subduction component, but was instead influenced by deep-seated fractionation processes. The remaining CPR basaltic alkaline rocks indicate little participation of the lithospheric mantle in their genesis (Rosenbaum et al., 1997), but show some subduction components (source **D**), which suggest that they were derived from the same source that previously generated the calc-alkaline suite (Styrian3 basalt is closest to the CAB) (source **E**).

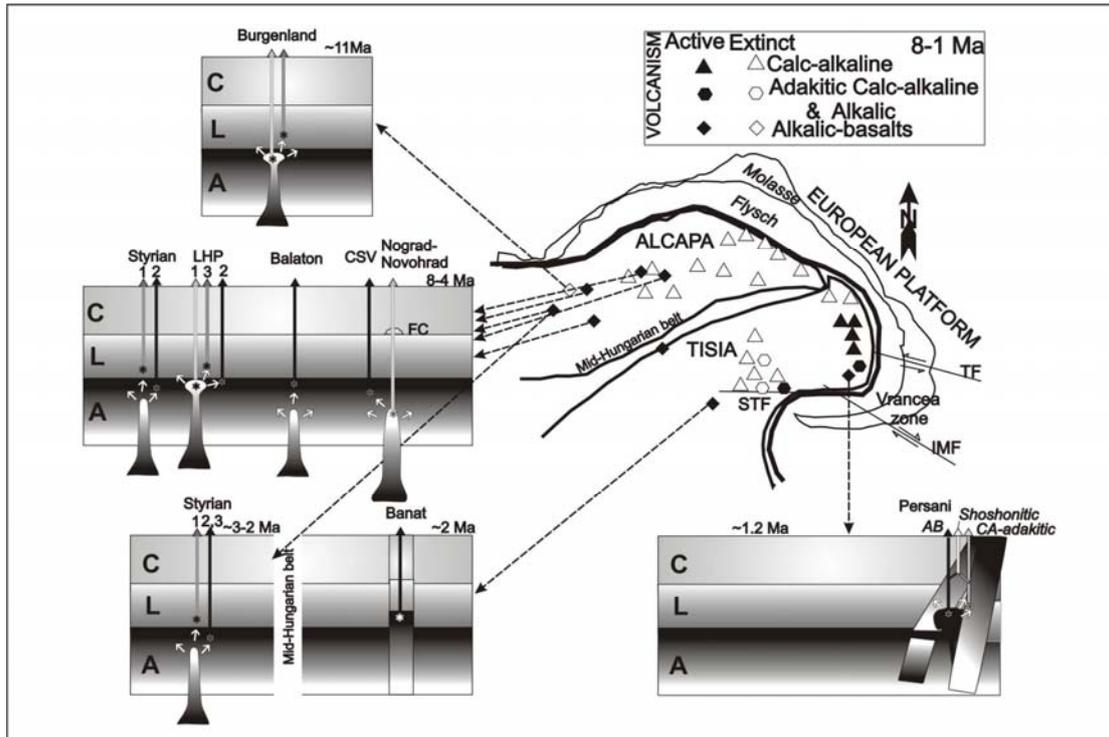


Fig. 13. Geotectonic sketch and profiles showing a tentative model for the relationships between main structural units (Alcapa and Tisia) and evolution of mafic alkalic magmatic activity in the Carpathian-Pannonian region, at 8-1 Ma. DVF = Dragoş Vodă fault; STF = South Transylvanian fault; TF = Trotuş fault; IMF = Intra-Moesian fault. C = crust; L = subcontinental lithospheric mantle; A = asthenosphere; FC = fractional crystallization; arrows = plume-generated heat; asterisks = site of melt generation

### *Finger-like mantle plumes*

Tertiary plume initiation beneath Europe is thought to be related to Alpine post-collisional mantle convection (Wilson & Patterson, 2001). Tertiary-Quaternary alkali basaltic magma generation beneath Central Europe are considered to be related to small-scale mantle plumes with finger-like geometries inducing high thermal anomalies at the base of the lithosphere (Granet et al., 1995; Wilson & Patterson, 2001).

Because of the variable geochemical and isotopic signatures in the western part of the CPR (Burgenland1, LHP1 and Nograd-Novohrad), we suggest the existence of three distinct mantle-source components related to diapiric upwelling of small finger-like plumes. Only one of them (the 11Ma old Burgenland1 source) is similar to EAR, although it is enriched in its Sr and Nd isotopic components. The geochemical data do not provide constraints on the location of isotopic components of each plume, but the enriched Sr and Nd signature of Burgenland1, LHP1 and the generally low volume of the magmatic products (as related to small plume dimensions) suggest a mantle enriched in EMII, corresponding to the upper asthenosphere, rather than lower asthenosphere. In contrast, the depleted character of the Nograd-Novohrad plume could suggest a lower mantle origin. Thus, the mantle plumes below CPR, at least those that evolved contemporaneously (LHP1 and Nograd-Novohrad), did not belong to a common asthenospheric source. The temperature anomaly defining the axial zone of the plume was probably sufficient to promote anhydrous melting of the upwelling material (Campbell & Griffith, 1990).

We also suggest that the Styrian, CSV and Balaton basalts were generated by the ascent of finger-like diapirs that probably did not reach the base of the lithosphere. However, it is not possible to determine their deep origin. According to age data, early stages of volcanism (e.g. Burgenland1 and LHP1) were initiated by plume melting that was facilitated by tectonically weak zones. This melting promoted conductive heating and thinning of the mechanical boundary layer, causing further melting from the base of the lithosphere and/or the upper asthenosphere and generating the later volcanism (e.g. Burgenland2 , LHP2, 3, Styrian2).

#### *Magma generation beneath CPR*

Prior to subduction, the asthenosphere beneath the CPR was already heterogeneous and melts derived from it acted as metasomatizing agents that formed amphibole-(phlogopite) veins in the lower lithosphere (e.g. Vaselli et al., 1995). The asthenosphere was later exposed to subduction metasomatism and the generation of calc-alkaline magmas (see Harangi, 2001a; Seghedi et al. submitted and references therein). Finally, it experienced the influence of new deep asthenospheric plumes, which acted both as source and as initiator of melting in upper asthenosphere and metasomatized lithosphere. Adiabatic decompression in an extensional system was also involved in the upper asthenosphere melting and alkali basaltic magma generation. The DM-HIMU-EMII mixing components of the asthenospheric magmas of the mafic alkalic rocks are in agreement with variable contamination of the shallow asthenospheric and/or thermal boundary layer of the lithosphere by a HIMU- mantle-plume component prior to and after the introduction of subduction components (Rosenbaum et al., 1997). A genetic model is presented in Fig. 13 and shows a number of magmatic events as follows:

(1) Burgenland mafic alkalic lavas were the result of a short-lived event at ~11Ma and are characterized by (1.1) Burgenland1 basalts with high SiO<sub>2</sub> and <sup>87</sup>Sr/<sup>86</sup>Sr ratios and a trace element pattern similar to OIB and (1.2) Burgenland2 basanites which have lower <sup>87</sup>Sr/<sup>86</sup>Sr, but high LILE and variable K depletion (Fig. 4), suggesting derivation by partial melting of a source containing a residual K-bearing phase such as amphibole and phlogopite (Späth et al., 1996). Phlogopite (e.g. Späth et al. 2001) in the source of Burgenland lavas is suggested by these lavas having the highest Ti contents in the CPR (Embey-Isztin & Dobosi, 1995). Amphibole and phlogopite occur in many mantle-derived peridotite xenoliths in alkali basaltic rocks in the CPR (e.g. Embey-Isztin et al., 1989; Downes et al., 1992; Szabo et al., 1995; Vaselli et al., 1995). We suggest that the source region of the Burgenland2 basanitic lavas was located at the base of the lithospheric mantle (as the site of an ancient asthenospheric metasomatic event) rather than in the coexisting asthenosphere, as amphibole is unstable in the convecting asthenosphere (Class & Goldstein, 1997). In contrast, the Burgenland1 basalt was derived from a mantle plume. The isotopic data of the Burgenland2 basanites suggest a rather depleted source, which requires derivation from an enriched HIMU-mantle plume component similar to the EAR from the Western-Central Europe. We agree with Rosenbaum et al. (1997) that this lithospheric metasomatic event is older, prior to the introduction of subduction components in CPR. Consequently, enriched isotopic composition of Burgenland2 basanites is interpreted as being derived from a deep plume. In summary, the ~11 Ma magmas in Burgenland area were derived from the base of a strongly enriched lithosphere (source **A**), which suggests an inherited chemical and isotopical source, implying ~ 1% partial melting and from a rising plume (source **B**), implying ~ 6 %

partial melting of a geochemically depleted and isotopically enriched source (Figs. 8,13).

(2) The LHP lavas, generated between 6 and 4 Ma show a scenario similar to Burgenland, but also have the most complex geochemical pattern of all the basalts from CPR, reflecting heterogeneity within the mantle sources: (a) An enriched geochemical pattern (source **A**- LHP3) with lower  $^{87}\text{Sr}/^{86}\text{Sr}$ , but high LILE and variable K-depletion in mantle-normalized incompatible element pattern, suggesting derivation from a metasomatised lithospheric mantle; (b) A geochemical pattern similar to OIB (source **B**- LHP1) with higher  $^{87}\text{Sr}/^{86}\text{Sr}$  and lower  $^{143}\text{Nd}/^{144}\text{Nd}$ , showing a larger degree of melting (~4 %) and suggesting derivation from an asthenospheric plume; (c) A pattern which shows subduction enrichment (source **D**-LHP2) identifies an upper asthenosphere variably affected by subduction components. We can speculate that a new upwelling plume at the base of the lithosphere gave rise to the LHP1 lavas by ~4% degree melting in an extensional environment. This plume interacted with upper asthenosphere previously penetrated by slab-derived aqueous fluids generating the LHP2 lavas and then heating the amphibole-bearing lithospheric mantle (apparently not affected by subduction metasomatism) gave rise to the LHP3 lavas.

(3) The Styrian Basin lavas, erupted between 6 and 2 Ma, also show a complex scenario of magma generation reflecting large heterogeneity within the mantle sources: (a) Styrian1 basanites and nephelinite generated in different periods (between 5-2 Ma) show an enriched geochemical pattern (source **A** similar to Burgenland2 and LHP3) with lower  $^{87}\text{Sr}/^{86}\text{Sr}$ , but high LILE and variable K-depletion in their mantle-normalized incompatible element pattern. They suggest derivation by variable degrees of melting (the lowest degree represented by nephelinite) of a lithospheric mantle (metasomatised in an earlier event). The nephelinites may be due to thermal relaxation after heating by a small plume (Jung & Hoernes, 2000); (b) Styrian2 basalt, which suggests an asthenospheric source enriched by subduction processes (source **D**). Two samples in this group, however, show low Pb and Th content, together with high Ba and LREE, which could result from melting of a local enriched lithospheric source (type **A**), because they also show a slightly negative K spike. Alternatively, they could be a mixture between asthenospheric and lithospheric melts; (c) Styrian3 basalt suggests derivation from an asthenospheric source affected by subduction components (source **E**).

Although Styrian Basin magmas were derived from a lithospheric source (source **A**) or an asthenospheric source variably influenced by subduction components (source **D** and **E**), the melting was likely triggered by a local upwelling of an asthenospheric mantle plume, which probably did not reach the base of the lithosphere.

(4) Balaton basalts were generated from an asthenosphere that still carried subduction components, suggesting that they were derived from the same source that previously generated the calc-alkaline suite (source **D**). However, limited interaction with previously enriched lithosphere cannot be excluded (Embey-Isztin & Dobosi, 1995). Increasing  $^{87}\text{Sr}/^{86}\text{Sr}$  at a constant  $^{143}\text{Nd}/^{144}\text{Nd}$  along with trace element data suggests that the asthenosphere was variably enriched by subduction components rather than mixing with a newly enriched asthenospheric plume source similar to LHP (Figs. 11). It is probable that heat transfer from the upwelling of a finger-like asthenospheric plume triggered melting in Balaton area.

(5) CSV and Banat areas show mantle sources that were least affected by a subduction component (source **D**). They are probably related to an ancient upper

asthenospheric reservoir. Although magma generation in Banat could be related to decompression melting event along the reactivated South-Transylvanian fault (Szakács & Seghedi, 1996; Roşu et al., 2000; Seghedi et al., submitted), it is more difficult to account for a melting process for the CSV basalts. It was probably the long-lived Nograd-Novohrad asthenospheric plume that triggered melting of pre-existing asthenosphere, as magmatism was reactivated after a long period (more than 5 million years gap).

(6) The mantle source (**C**) of Nograd-Novohrad basalts is the most isotopically depleted, showing only some Pb isotope variation, and the magmas are also fractionated. Fractionation occurred at the mantle-crust boundary (Dobosi & Jenner, 1999), which was stable for a long period (Fig. 2) after collision. The plume model is viable in the Nograd-Novohrad area.

(7) The Perşani mountains alkalic basalts were erupted between 1.5 and 0.6 Ma, coeval with adakite-like calc-alkaline and shoshonitic magmas nearby. This situation can be explained by breakoff and tearing of the subducted slab at shallow levels in the extreme southern part of the arc (Maţenco, 1997), which allowed the uprise of the subduction-metasomatized asthenospheric mantle (source **D**). Heat from the asthenospheric plume produced melting of the torn slab, as well as generating melt via asthenospheric mantle decompression. Further mixing may have occurred between the produced melts (Mason et al., 1996; Seghedi et al., submitted).

## **Conclusions**

This review of geochemical and isotopic data of alkalic mafic rocks from the CPR has identified various mantle sources during the Late Miocene-Quaternary post-collisional period. The following geochemically defined mantle reservoirs which contributed to the generation of alkalic mafic magmas in the CPR (see Figs. 4, 11, 12) can be recognized:

(Source **A**) Lower lithospheric mantle affected by several ancient asthenospheric enrichments (Burgenland2 basanite, LHP3, Styrian1 basanites and nephelinite). This enrichment was derived from volatile-bearing melts, which penetrated the lithosphere and formed amphibole (phlogopite)-bearing assemblages, that either have the bulk isotopic composition of the metasomatic agent, or are intermediate between those of the unmetasomatized lithosphere and the metasomatic agent. Spatial differences in the geochemical and isotopic enrichment patterns in the lithosphere below Burgenland, Styrian and LHP areas, argue for multiple ancient asthenosphere-lithosphere interactions. Alternatively, they could be explained by the independent evolution of different parts of the Alcapa terrane-puzzle lithosphere and upper asthenosphere prior to the Miocene collision (Embey-Isztin & Dobosi, 1995).

(Sources **B** and **C**) Younger asthenospheric plumes (Burgenland1 basalt, LHP1, Nograd-Novohrad). Small-scale mantle plumes with finger-like geometry inducing high thermal anomalies at the base of the lithosphere are thought to be the triggering mechanism for Tertiary-Quaternary alkalic basaltic magma generation beneath Western and Central Europe (Granet et al., 1995, Wilson & Patterson, 2001). CPR plumes have an OIB-like geochemical pattern and isotopically are either depleted (Nograd-Novohrad) or variably enriched (Burgenland1 and LHP1 basalt). Since the plumes also show variable geochemical and isotopic composition they argue for heterogeneous composition of the lower asthenospheric mantle below CPR during Miocene-Pleistocene time. The initiation of these plumes could be due to hot mantle

upwelling from a conjunct asthenospheric reservoir beneath Europe (Hoernle et al., 1995).

(Source **D**) Old upper asthenosphere heterogeneously contaminated by DM-HIMU-EMI-EMII components (LHP2, Balaton, Styrian2, CSV, Banat and Perşani) and slightly influenced by subduction enrichment. The compositions of the basalts derived from the asthenosphere, affected by subduction enrichment are consistent with an origin under relatively volatile-poor conditions compared to the earlier calc-alkaline magmas.

(Source **E**) Old upper asthenosphere heterogeneously contaminated by DM-HIMU-EMII components, and significantly influenced by subduction enrichment (Styrian3). The presence of asthenospheric mantle slightly to strongly affected by subduction enrichment is a peculiarity of alkalic basalts in most of the CPR area, except for Burgenland and Nograd-Novohrad. This suggests that after the generation of calc-alkaline magmas, the upper asthenospheric/lower lithosphere mantle source was still accessible for melting under new conditions due to the interaction with deep-generated mantle plumes.

We have also identified the mechanism of melt generation, which was triggered either by (i) young Late Micene and Pliocene-Quaternary asthenospheric mantle plumes rising up and heating the base of lithosphere (below the Alcapa block), or by (ii) tectonic processes which allowed decompressional melting of the old asthenosphere as it rose to replace the convectively removed lithosphere or to heat and melt former subducted slab component (related to Pliocene-Quaternary tectonic processes in the Tisia block) (Fig 13).

This review also reveals the importance of completing the collection of geochemical and isotopic data in the Danube-Tisza area and acquiring Pb isotopes for LHP and the Styrian Basin areas. A more complete data set will offer the opportunity to discriminate the isotopic influences of the asthenosphere during successive mantle plume enrichment events below the CPR.

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Appendix

**Geochemical, Sr and Nd isotopic and K-Ar age data  
from the Styrian Basin**

Sample	Kfs. 1	TS3	GS6	NH5	GR13
Locality	Kapferstein	Tieschen	Gussing	Neuhaus	Grad
Age (Ma)	4.68± 0.26	2.17± 0.13	6.3±0.50	3.11± 0.75	3.37±0.5 7
SiO <sub>2</sub>	44.95	45.04	50.38	50.58	44.73
TiO <sub>2</sub>	2.127	2.17	2.498	1.565	1.929
Al <sub>2</sub> O <sub>3</sub>	14.33	15.09	14.07	15.81	14.06
Fe <sub>2</sub> O <sub>3</sub>	11.63	10.07	10.78	8.43	10.35
MnO	0.32	0.17	0.17	0.09	0.21
MgO	7.26	9.31	4.45	7.33	6.76
CaO	9.78	9.82	8.98	7.43	10.53
Na <sub>2</sub> O	5.26	4.64	2.81	2.71	2.79
K <sub>2</sub> O	1.53	2.17	1.77	3.06	2.53
P <sub>2</sub> O <sub>5</sub>	1.18	0.70	0.80	0.55	0.91
LOI	1.04	0.29	3.66	1.98	4.69
Total	99.41	99.47	100.37	99.54	99.49
Nb	93	69	64	28	88
Zr	273	178	335	152	245
Y	23	17	16	16	20
Sr	949	704	704	401	1525
Rb	100	47	81	105	61
Th	9.5	7.3	4.7	8.2	7.4
Pb	5.4	3.4	9.1	7.5	3.5
Zn	99	60	84	56	85
Cu	26	32	44	31	33
Ni	187	152	75	130	113
Cr	234	173	151	154	156
Ce	141	124	108	67	158
Nd	65	54	48	33	65
La	70	56	47	28	79
V	186	215	234	188	171
Ba	1213	949	1105	821	2236
Sc	8.9	18	18	19	12
<sup>87</sup> Sr/ <sup>86</sup> Sr	0.703438	0.703688	0.704218	0.705914	0.703494
<sup>143</sup> Nd/ <sup>144</sup> Nd	0.512894	0.512843	0.512788	0.512643	0.512886

### *Analytical techniques*

Basaltic rocks were crushed with a jaw-crusher or in an agate mortar with agate pestle. After quartering samples were ground in an agate mortar ball mill. Bulk-rock chemical analyses were performed by X-Ray Fluorescence at the University of Edinburgh (Dept. of Geology) with a Phillips PW 1400 XRF using glass disks and pressed pellets for major and trace elements respectively.

Sr and Nd isotopic analyses were carried out at Royal Holloway, University of London, using a VG354 multicollector mass spectrometer. Analytical procedures are reported in Thirlwall (1991a, b), Downes et al. (1992) and Vaselli et al. (1995). Sr and Nd blanks were 0.95-1.06 ng and 0.2-0.7 ng, respectively. These blank values do not affect the measured  $^{87}\text{Sr}/^{86}\text{Sr}$  and  $^{143}\text{Nd}/^{144}\text{Nd}$  isotopic ratios. During the period of measurement, the standard values were  $^{87}\text{Sr}/^{86}\text{Sr} = 0.710525 \pm 21$  (2s) for SRM987 and  $^{143}\text{Nd}/^{144}\text{Nd} = 0.511420 \pm 11$  (2s) for the laboratory standard, identical to values reported by Thirlwall (1991a, b).

For K/Ar dating the samples were degassed in a conventional extraction line by means of induction heating. After spiking with  $^{38}\text{Ar}$  measurements were made by mass spectrometry in a static mode. K concentrations were determined by flame photometric techniques. Inter-laboratory standards Asia 1/65, HD-B1, LP-6 and GL-O as well as atmospheric Ar were used for controlling and calibration of the determinations. Details of the instruments, the applied methods and results of calibration have been described elsewhere (Balogh, 1985).

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