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GEOCHEMISTRY AND TECTONIC DEVELOPMENT OF CENOZOIC MAGMATISM IN THE CARPATHIAN-PANNONIAN REGION

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ABSTRACT

This updated review considers the magmatic processes in the Carpathian-Pannonian Region (CPR) during Early Miocene to Recent times, as well as the contemporaneous magmatism at its southern boundary in the Dinaride and Balkans regions. This geodynamic system was controlled by the Cretaceous to Neogene subduction and collision of Africa with Eurasia, especially by Adria that generated the Alps to the north, the Dinaride-Hellenide belt to the east and caused extrusion, collision and inversion tectonics in the CPR. This long-lived subduction system supplied the mantle lithosphere with various subduction components. The CPR contains Neogene to Quaternary magmatic rocks of highly diverse compositions (calc19 alkaline, K-alkalic, ultrapotassic and Na-alkalic) that were generated in response to complex post-collisional tectonic processes. These processes formed extensional basins in response to an interplay of compression and extension within two microplates: ALCAPA and Tisza-Dacia. Competition between the different tectonic processes at both local and regional scales caused variations in the associated magmatism, mainly a result of extension and differences in the rheological properties and composition of the lithosphere. Extension led to disintegration of the microplates that finally developed into two basin systems: the Pannonian and Transylvanian basins. The southern border of the CPR is edged by the Dinaride and Balkans (Sava and Vardar zones) that acted as a regional extensional tectonic setting during Miocene times. This extension was associated with small volume volcanism in narrow extensional sedimentary basins or granitoids in core-complex detachment systems.

Major, trace element and isotopic data of magmatic rocks from the CPR suggest that subduction components were preserved in the lithospheric mantle after the Cretaceous-Miocene subduction and were reactivated especially by asthenosphere uprise via extension. Pre-collisional subduction-related volcanic activity is absent from the CPR area. Changes in the composition of the mantle through time support geodynamic scenarios of collision and extension that are linked to the evolution of the main blocks and their boundary relations. Weak lithospheric blocks (i.e. ALCAPA and western Tisza) generated the Pannonian basin and the adjacent Styrian, Transdanubian and Zărand basins which show high rates of vertical movement accompanied by a range of magmatic compositions. Strong lithospheric blocks (i.e. Dacia) were only marginally deformed, as in the northern and eastern part of the Transylvanian basin, where strike-slip faulting was associated with magmatism and extension. Strike slip tectonic and core complex extension was associated with small volume volcanism along older suture zones (Sava zone and Vardar zone) accommodating the extension in the Pannonian basin. Various magmas acted as lubricants in a range of tectonic processes.

Key words: magmatism, lithosphere, asthenosphere, extension, core-complex-related magmatism, transtension-related magmatism

1. Introduction

This review interprets previous and recent data on magmatic and geodynamic processes in the Carpathian-Pannonian Region (CPR) during Early Miocene to Recent times, as part of the long-lived Alpine-Carpathian-Dinaridic orogenic system (see Schmid et al. 2008 and references therein). We discuss the geodynamic implications of magmatism in an area where several microplates invaded the Carpathian embayment. The microplates (ALCAPA and Tisza-Dacia) represent the northernmost fragments caught between the European mega-plate (including the European plate, East European platform and Moesian microplate) and the northward propagation of the Afro-Arabian mega-plate including its Adria promontory (Fig. 1) (Rosenbaum et al., 2004).

The CPR has long been an excellent laboratory in which to study the interaction between tectonics, deep mantle processes and igneous activity (Chalot-Prat and Gîrbacea, 2000; Downes et al., 1995a,b; Harangi, 2001; Harangi et al., 2001, 2006, 2007; Harangi and Lenkey, 2007; Konečný et al., 2002; Kovács et al., 2007; Kovács and Szabó, 2008; Mason et al.,1996, 1998; Pécskay et al., 1995, 2006; Póka et al., 1988, 2004; Salters et al., 1988; Seghedi et al., 2004a,b, 2005, 2010; Szabó et al., 1992). These studies revealed that the most voluminous igneous rocks, the calc-alkaline ones, are derived from a subduction-enriched source, with a typical subduction-related geochemical signature (i.e., high LILE/HFSE ratios, negative Nb, Ta and Ti anomalies; lower $^{87}Sr/^{86}Sr$ and higher $^{143}Nd/^{144}Nd$ radiogenic isotopic ratios), which is commonly suggested to reside within the mantle lithosphere. However genuine subduction-related calc-alkaline magmatic arcs show linear arrays of point magma sources, parallel to the mantle wedge above subduction zones (e.g. Stern, 2002; Kogiso et al., 2009 and references therein) with a melt zone resulting from the release of hydrous fluids from subducted materials lowering the melting temperature of the overlying asthenospheric mantle. Such a linear array cannot be found in the CPR. However, similar situations where lithospheric mantle enriched with subducted material was involved in late melt-production are not uncommon (e.g. McPherson and Hall, 1999; Kovács and Szabó, 2008; Karaoğlu et al., 2010). In the CPR the calc-alkaline magmatism is considered to be post-collisional and entirely of lithospheric origin. It is dominantly a result of extensional processes dependent on the rheological properties and specific lithosphere composition of the microplates and their boundaries (Fig. 2). In addition, the transition toward the asthenosphere-derived magmas, represented mainly by the Na-alkalic suite, is of crucial importance in understanding the tectonic processes.

Within the CPR literature, there is not yet a consensus whether or not the calc-alkaline magmas were generated during direct subduction processes. Disagreements also occur over the relationships between magmatism and geodynamics in the CPR, since the overall geodynamic post-Miocene reconstructions are not very different from the early models of Balla (1984), Csontos et al. (1992) and Ratschbacher et al. (1991). Additionally, there are still many geodynamic reconstructions that ignore or minimize the importance of magmatism in the evolution of the CPR (e.g. Ustaszewski et al., 2008; Lorinczi and Houseman, 2010).

Here we give a new overview and propose a working hypothesis with the aim of giving the most plausible interpretation of the most credible geochemical data-base. The large amount of geological, structural, petrologic and geochemical data (most at the same quality level) has broadened the understanding of the importance of magmatism in the larger scale geodynamic evolution. Our interpretation will follow the changes in the magma compositions through time in connection to the evolution of the main sedimentary basins (Pannonian and

Transylvanian basins) that resulted from the disintegration and rearrangement of ALCAPA and Tisza-Dacia blocks that filled the Carpathian embayment, in order to discuss various geodynamic scenarios. This approach differs from our previous one (Seghedi et al., 2004a) when we discussed the magmatism in relation to the place of main eruption areas in four ―segments‖, without including the magmatism in the Styrian basin or the Sava and Vardar zones because of the scarcity of data for these areas at that time.

We consider the close link between the magmatism and the extensional evolution of the main lithospheric blocks and their boundaries. The rheological properties of the lithosphere (i.e. whether it is rigid or weak) cause specific magmatic evolution during orogenic processes (Cloething et al., 2006, Tesauro et al., 2009). Melt production and migration followed various kinds of extension and we suggest that both large and small volume magmatic activity influenced the dynamics of extension.

2. Geodynamic features

In the eastern Mediterranean area, tomographic models have shown the ―Aegean Tethys‖ to be a subduction zone whose features have been studied extensively and correlated to geologic events. The still-subducting ―Aegean Tethys‖ is inferred to have started at approximate 171±5 Myr and its slab remnant was shown to reach a depth of ~ 1500 km (e.g. Facenna et al., 2003; van der Meer, 2009). According to this scenario the whole post-Miocene evolution of the region between Africa, with its Adria block promontory, is a still-active subduction zone. Inside the Carpathian-Moesian realm, lithospheric block motion was considered to be driven by retreat of a west-dipping European lithospheric slab (e.g. Royden, 1988) and the push of Adria (e.g. Ratschbacher 1991; Rosenbaum et al., 2004) and resulted in various translation movements, and diachronous extensional rotation that broke up the upper crust and squeezed it into the available space (Csontos, 1995; Fodor et al. 1999; Horváth et al. 2006; Márton 2000; Márton and Fodor, 2003; Seghedi et al., 1998). The emplacement of these microplates, recognized as belonging either to Adria (Africa) or Europe, or both (Figs. 2, 3) (Schmid et al. 2008) caused a ‗soft‘‘ continental collision at ~11 Ma with only minor crustal thickening. Continental collision was accompanied by substantial strike-slip faulting, extension and opposed block rotations (e.g. Csontos and Nagymarosi, 1998; Maţenco et al., 2007) involving corner effects at the Bohemian (Sperner et al. 2002) and Moesian (Ratschbacher et al. 1993; Schmid et al. 1998; Fügenschuh and Schmid 2005) promontories. In Lower Miocene times the inferred subduction was at a much smaller scale than classic subduction systems (which can be thousands of km wide) being a land-locked basin several hundred km in width between Moesian and European plate (e.g. Csontos, 1995 and the references therein, Ustaszewski et al., 2008). The Moesian plate was stable during the Miocene, being overthrust and deformed until mid-Eocene times to the south, due to subduction/collision in the Aegean arc, and transpressionally deformed by eastward propagation of the Dacia microplate during Miocene times along its northern margin (Fügenschuh and Schmid 2005; van Hinsbergen et al., 2008). Substantial Miocene dislocations took place along the Mid-Hungarian Fault Zone that disrupted the former connection between ophiolitic units in the West Carpathians and the Dinarides (Schmid et al. 2008). The Adriatic indentor was responsible for Pliocene-Quaternary inversion in the Pannonian basin and all around the Carpathians by fault reactivations, large scale folding and present-day seismicity (Pinter et al., 2005; Jarosinski et al., 2010).

The simultaneous occurrence of compression and extension is considered typical for slab retreat and low-stress subduction zones (Royden, 1993). Slab break-off (Mason et al., 1998; Nemčok et al., 1998; Seghedi et al, 1998; Wortel and Spakman, 2000) or slab delamination (Sperner et al., 2001; Gîrbacea and Frisch, 1998; Knapp et al, 2005, Fillerup et al., 2010) have been invoked to explain magma generation along plate boundaries. The Transylvania basin, consisting of the Dacia block and north-east Tisza block, displayed minor Miocene upper crustal extension, which was replaced during the late Miocene by small scale contraction features and shallow salt diapirs (Krézsek and Bally, 2006; Mațenco et al, 2010a). Its crust and lithosphere have normal thicknesses (Dérerová et al., 2006), the surface heat-flux is low (30-60mW/m2) and shows higher values (120mW/m2) only where overlapping the narrow East Carpathian volcanic range (Demetrescu and Andreescu, 1994; Demetrescu et al., 2001).

Major processes accommodating the extensional deformation in the Pannonian basin were of wide-rift type and core-complex type, characterized by crustal flow sometimes associated with doming and lateral escape of the lower-middle crust at the expense of low154 buoyancy lithosphere (Tari et al., 1999; Csontos and Vörös, 2004). The unusual hot and weak Pannonian lithosphere presently observed (e.g., Cloetingh et al., 2006) with a present-day high heat flow (~ 90-100mW/m2) evident also in the digital elevation models (e.g. Dunkl and Frisch, 2002) and thermal and rheological modeling (Tesauro et al, 2009), implies an overthickened crust and thin lithosphere at the onset of extension (Tari et al., 1999).

A synthesis of Miocene-Quaternary geodynamic developments in the CPR showing time intervals of geodynamic processes leading to final collision in different parts of the system, major rotation during extension and recent inversion due to the push of Adria is given in Fig. 4.

3. Classification of magmatic rocks and their spatial and temporal distribution

Analyses of approximately 650 magmatic rocks have been divided into four main groups based on their variation in the SiO_2 versus $Na_2O + K_2O$ diagram (Fig. 5): (1) Calc-alkaline, (2) Na-alkalic, (3) K-alkalic and (4) Ultra-K. The reason for using this traditional classification (Le Bas et al., 1986) is to avoid any genetic connotation, since this is mostly debated.

The most complex and abundant group is the calc-alkaline group (**Group 1**), that, based on dissimilar geochemical and petrographic features, discussed later, has been divided into several subgroups:

(1a) a felsic subgroup, dominantly acid (>70 wt% SiO_2) pyroclastic rocks and rare minor domes (Lexa et al., 2010) generated during the Miocene (21–10 Ma) in the Pannonian and Transcarpathian basins (Fig. 3). This subgroup is the same as the ―silicic suite‖ of Harangi and Lenkey (2007);

(1b) a normal calc-alkaline subgroup spread across the CPR during middle Miocene to Quaternary times (17–0.1 Ma) is dominated by andesites and dacites with minor basalts, basaltic andesites and rhyolites. Garnet-bearing andesites, dacites and rhyolites are part of this group and were erupted at different times in most areas (e.g. Harangi et al., 2001; Nițoi et al., 2002; author‘s unpublished data) (Fig. 3);

(1c) an adakite-like subgroup characterized by rocks dominated by amphibole and biotite as mafic phenocrysts with high Sr/Y ratio occurred in the Apuseni Mts. at 12.5-7.5 Ma and in South Harghita at 3.5-0.03 Ma (Roşu et al, 2004; Seghedi et al., 2004a, 2007, 2010);

(1d) a transitional subgroup, formed of basaltic andesites to rhyolites characterized by higher Nb (35ppm) and Nb/Y, shifting to a more enriched composition similar to asthenosphere-derived magmas, present between 11 and 8 Ma in the Pannonian basin (Central Slovakian volcanic field), or as rare occurrences at 7-8 Ma in Apuseni Mts. or at 2 Ma in the Perşani Mts. (Harangi et al., 2007; Mason et al., 1996, 1998; Roşu et al., 2004; Seghedi et al., 2005).

K-alkalic volcanism (**Group 2**) includes large volume shield volcanoes in the Styrian basin (15.5-16.5 Ma), which form part of a complex and voluminous buried volcanic area (e.g. Ebner and Sachsenhofer, 1995; Ntaflos et al., 2007). A similar buried trachybasalt volcano, 200m thick, was found in southern Transdanubia and is dated at 14.5-15.0 Ma (Harangi, 2001). The younger 1.5-1.8 Ma K-alkalic volcanism situated at the southern margin of Transylvanian basin is of small volume (e.g. Pécskay et al., 1995b).

Ultra-K volcanism (**Group 3**) is rare, existing as a buried 2-2.2 Ma leucitite in the interfluve region between the Danube and Tisza rivers (Pécskay et al., 1995; Harangi et al., 1995b) and as a unique lamproite volcano at 1.3 Ma at the south-eastern border of Pannonian basin (Seghedi et al., 2008).

Na-alkalic volcanism (**Group 4**) was generated during the late Miocene to Quaternary (~11–0.2 Ma), generally following the calc-alkaline magmatism in different areas. It comprises monogenetic volcanic fields of maars, diatremes, tuff cones, cinder/ spatter cones and lava flows (Lexa et al., 2010). However, even though it may be important, we are not able to discuss the 11–12 Ma large trachyandesite to alkaline trachyte volcano associated with minor alkali basalts situated beneath 2000m thick sediments in the Little Hungarian plain, eastern Pannonian Basin, attributed by Harangi et al. (1995a) and Harangi (2001b) to the Na208 alkalic volcanic series, because of a lack of published geochemical data. For the same reason, (missing geochemical data) we cannot discuss the 10-8 Ma Na-alkali basalts buried in the interfluve region between the Danube and Tisza rivers (Pécskay et al., 1995a, 2006) (Fig. 2).

Fig. 3 is a simplified and updated spatial and temporal distribution (Pécskay et al., 2006) in which we have distinguished the development of magmatism as related to the main sedimentary basins (Pannonian and Transylvanian) generated during the main tectonic events in the CPR. These events include tectonic escape and extension of the main ALCAPA and Tisza-Dacia blocks, resulting in a simultaneous breakup; and collision in the East Carpathians and Pliocene-Quaternary inversion tectonics derived from the push of Adria. We have added the contemporaneous magmatism situated at the southern limit of the CPR that includes the Sava and Vardar zones ophiolitic accretionary prisms. Calc-alkaline intrusions in Fig. 3 are of two kinds: (1) those induced by extensional exhumation located in the footwall of core complexes in the Styrian Basin (18.2-16.5 Ma) and Vardar zone (20-16 Ma) (Cvetković et al., 2007; Fodor et al, 2008; Schefer et al., 2010; Koroneos et al., 2010), that will be not commented on from a geochemical point of view, and (2) hypabyssal intrusive complexes, which are frequently associated with volcanic rocks (not shown), or piercing flysch-type sedimentary rocks along the internal margin of the Carpathian accretion prism (Pieniny at 13.5-11 Ma, Trua et al, 2006) or following a transtensional fault system in the northern part of Transylvanian basin (12-8 Ma), where they are known as the ―subvolcanic zone‖ (Pécskay et al., 2009) (see also Fig. 2).

4. Geochemical features

In previous summaries of the geochemical features of the Neogene to Quaternary volcanic rocks of the CPR (Harangi, 2001a; Harangi et al., 2006; Harangi and Lenkey, 2007; Seghedi et al. 2004a, 2004b, 2005), there are detailed discussions of the petrological features of magmatic rocks. Here we will use only a limited number of diagrams: Nb/Y and $^{87}Sr/^{86}Sr$ as source indicators; SiO_2 as an indicator of magma chamber processes, and Th/Y as an indicator of subduction or crustal input, in order to outline the most important characteristics connected to the geodynamic processes (Figs 6a, b, 7a, b, 8a, b). High Nb/Y, low SiO_2 and low $^{87}Sr/^{86}Sr$ are indicators of mantle (asthenosphere) whereas low Nb/Y, high SiO_2 and high $^{87}Sr/^{86}Sr$ are indicative of crust or metasomatized mantle. We use Y instead of Yb in the same way as Pearce (2008), since Yb and Y behave similarly and Yb is sometimes missing from our data base. Besides being a source indicator, Nb/Y is also sensitive to fractional crystallization, as Nb is more strongly partitioned into melt during magma chamber processes than Y, although its behaviour is somewhat different at higher pressure when Y partitions strongly into garnet (e.g. Natland, 2007). Brief interpretations of geochemical features will be discussed along with the regional magmatic and geodynamic development (Figs. 3-10).

5. Regional geodynamic development-connection with magmatic processes

The post-Miocene evolution of the Alcapa and Tisza-Dacia blocks attests to an important disintegration and rearrangement, so the accompanying magmatic activity was connected temporally and spatially to the main basin system development: a Pannonian Basin system that splintered into several sub-basins each with specific associated magmatism: Styrian, main Pannonian, Transcarpathian and Zarand basins, and the Transylvanian basin system (Figs. 2, 3).

5.1 Pannonian Basin system

5.1a Styrian basin and surrounding areas - the south westernmost Pannonian sub-basin

The volcanic events (Fig. 3) occurred after the low-angle extensional core-complex deformation that controlled lower crust exhumation along the western margin of the Styrian basin and generation of the Pohorije pluton along the transpressional Periadriatic fault system (Dunkl et al., 2003; Fodor et al., 2008).

This basin contains allochthonous felsic pyroclastic deposits (not discussed) and intermediate calc-alkaline and K-alkalic volcanic rocks generated at 17.5-14 Ma. The geochemical diagrams suggest a common source and isotopic variability suggests a variably enriched lithospheric source (perhaps enriched during earlier subduction events) without significant fractionation (Figs. 6a, 7a). The magmatism may be associated with extension (see also Harangi et al., 1995) following extrusion tectonics of low-angle core-complex type during counterclockwise rotation (Márton and Fodor, 2003) that triggered melt generation at different levels in the lithosphere.

Small volumes of Na-alkali basalts occurred at ~11Ma in Burgenland and at 4-1.8 Ma in the Styrian basin. Burgenland basalts were generated during small scale lithospheric extension at the western edge of the Pannonian basin through passive upwelling and adiabatic decompression melting of asthenospheric mantle (Ali and Ntaflos, 2010). The Styrian basin basalts show heterogeneous isotopic features that suggest asthenospheric decompression melting from various depths (Embey-Isztin et al., 1993; Ali et al., submitted). We associate this with the push of Adria and tectonic inversion which caused a north-east directed asthenospheric mantle flow coupled to small scale lithospheric extension (Márton and Fodor, 2003).

5.1b. Western Carpathians and Pannonian Basin

Here the magmatism shows the most complex compositional variation and longest duration (Fig. 3). It started as large volume felsic pyroclastic eruptions at 21-18 Ma, followed at 18-8 Ma by large volume felsic pyroclastic and intermediate calc-alkaline lavas and pyroclastics, and ended with small volume Na-alkalic basaltic volcanism (10-0.1 Ma) (Lexa et al., 2010). Magmatism became younger towards the north, where it intruded the Outer Carpathian nappes (Pécskay et al., 2006). Geochemical data imply a change in source from a crustal one (showing highest SiO_2, $^{87}Sr/^{86}Sr$, Th/Y and lowest Nb/Y), through a mixed crustal/lithospheric mantle source, to a lithospheric mantle source with decreasing subduction component through time (Figs. 6a, 7a). The younger intrusions in the Carpathian nappes have the lowest Th/Y, suggesting a less enriched lithosphere source (Trua et al., 2006). Garnet289 bearing varieties occurred mainly at 16.4-15 Ma and suggest high pressure partial melting, as well as contamination via mixing of lithosphere mantle-derived magmas with variable amounts of lower crustal metasedimentary material, explaining the large Nb/Y and $^{87}Sr/^{86}Sr$ isotope variation (Harangi et al., 2001). The garnet-bearing rocks were interpreted as a sign of change in the regional stress field from compressional to tensional, since garnet is not stable at shallow depths and its preservation requires a rapid ascent of the host magma to the surface (Harangi and Lenkey, 2007). The high Nb/Y ―transitional‖ andesites generated at 11-8 Ma show a different Nb/Y trend and fractionation. The data demonstrate the diminution of subduction-components in the lithospheric mantle with time (Fig. 7a, 1b, Th/Y vs. Nb/Y) in favor of less affected asthenospheric mantle that had already started to generate melts at that time (Harangi et al., 2007; Seghedi et al., 2005). Mixing of magmas derived from the lithosphere and asthenosphere probably caused the rather sudden increase of Nb/Y (Fig.7a). Wide-rift extension and block rotation contemporaneous with extrusion tectonics towards the south (e.g. Marton and Fodor, 2003), triggered partial melting initially at the crustal level and then at the upper lithosphere mantle level favored by asthenosphere upwelling (Figs. 4, 9). A Miocene (17-18 Ma) thermal event in the basement outcrops suggests that a regional distribution around volcanic areas in the Western Carpathians and Pannonian basin was associated with the high heat flow during crustal extension (Danišík et al., 2008). It is possible that extension and rotation triggered the crustal flow that may have also generated a small volume of magma (e.g., Teyssier et al., 2005), however we suggest that the heat from the asthenosphere produced melting in the hydrated part of lithosphere (with probable underplating) and favored crustal melting. In this scenario, the addition of both volume and heat from crustal and lithospheric magmas led to melt-induced weakening at the crust-mantle boundary that initiated the detachments (Tari et al., 1999).

The origin of the small intrusions in the Pieniny and Eastern Moravia (Western Carpathians) has been a matter of long debate. Since at around 14 Ma the western part of ALCAPA had already collided with Europe (e.g. Fodor et al., 1999) (see Fig. 4), we suggest their generation was related to regional transtensional faulting at the block boundary during and after major counterclockwise rotation and collision of the westernmost ALCAPA block at around 14-12 Ma that generated the Transcarpathian basin (Márton et al., 2007).

Generation of Na-alkalic magmatism at 4-3 Ma along the mid-Hungarian line in the Balaton area, included in the 1b group (Pécskay et al., 2006, Wijibrans et al., 2007) suggests (as in the case of Styrian basin basalts) a north-east-directed asthenosphere mantle flow and small volume partial melting that produced volcanism along NW-SE strike-slip lithospheric faults. The process was most likely controlled by mantle perturbations resulting from the counterclockwise rotation of the Adriatic microplate and tectonic inversion in the Pannonian basin (Márton and Fodor, 2005; Bada et al, 2007). Continuous generation of small volume Na-alkalic basalts between 8 and 0.13 Ma, following normal and transitional calc-alkaline magmatism (16.5-11; 11-8 Ma) suggests a long period of small volume asthenospheric melt production via decompression melting (Harangi and Lenkey, 2007). The region of Na-alkali magma generation that is superposed on the previous calc-alkaline volcanism (Lexa et al., 2010) encircles the place of asthenosphere uprise and could potentially be the site of a mantle plume (Konečný et al, 2005). This is possibly suggested by recent tomography data (Koulakov et al., 2009), but not yet conclusively demonstrated.

5.1c Transcarpathian basin

In the north-westernmost part of the Pannonian Basin, felsic and normal calc-alkaline volcanism erupted in the Transcarpathian basin at a triple junction between ALCAPA, Tisia-Dacia and the European foreland. This volcanism occurred all around the margins of the Transcarpathian basin at 15-9 Ma (Pécskay et al, 2006), following a crustal fault system that allowed magma extrusion along N-S tectonic depressions and an E-W transtensional fault system (e.g. Baráth et al, 1997). It formed an aligned chain of composite volcanoes, but also erupted inside the basin that attains a thickness up to 6-7 km, with more than 1.5 km of volcanic products (Lexa et al, 2010). Geochemical studies indicate a heterogeneous lithospheric mantle source associated with assimilation-fractional crystallization (AFC) processes in crustal magma chambers (Seghedi et al., 2001). The calc-alkaline magmas show small scale fractionation with higher Th/Y for the volcanic rocks in the interior of the basin, where younger felsic magmas are present, suggesting a derivation and possible mixing of upper lithosphere mantle and lower crust sources (Fig. 6a, 7a,-1c). Little is known about the garnet-bearing rocks in this area; recently Konečný et al. (2010) showed the presence of garnet in rhyolites, attributing them to a crustal origin, based on Sr and Nd isotopes. Magmatism in the Transcarpathian basin was generated at the mantle lithosphere/crust level as a result of major extension via counter-clock rotation of the easternmost part of ALCAPA that caused core-complex exhumation of lower-middle crustal units (Soták et al, 2000, Bárath et al., 1997). The counter-clockwise rotations of ALCAPA and the clockwise rotation of Tisza-Dacia started at ~18.5 Ma. However after 14.5 Ma the Transcarpathian Basin area disintegrated from ALCAPA along the Hernad fault (Márton et al, 2007) and up to 11 Ma collided with the European plate. The Transcarpathian basin appears to have resulted from typical core-complex extension that allowed decoupling and flow of lower-middle crust and magma generation within the decoupling area. Back-arc rotational extension and asthenosphere uprise was invoked for magma generation (Seghedi et al, 2004a), however although the back-arc concept has been invoked by most specialists in the Pannonian area, it is an ambiguous concept for any basins within the CPR, since no true ―arc-type‖ magmatism exists, as required (e.g. Stern, 2002).

We suggest that magmas generation, closely correlated to extension, occurred at the same time in both the crust and lithospheric mantle (already affected by subduction components). We correlate extension to an asthenosphere uprise (Konečný et al., 2002, Seghedi et al., 2004a), at a smaller scale than in the Pannonian area, that at the end of collision did not produce asthenosphere-related magmas (Figs. 4, 9). The evidence is the high heat flow of 80-100 mW m^{-2} (Čermák, 1977) and the lithosphere thickness up to ~60 km (Horváth et al., 2006).

4.1d. Zărand basin and Apuseni Mts

Calc-alkaline and adakite-like magmas erupted in the Zărand Basin and Apuseni Mountains at 15-9 Ma and are known to be buried in the Bekes basin toward the west. Garnet-bearing rocks occurred at 13-12 Ma and a small volume of transitional-type andesitic basalts were generated at ca. 8 Ma. After a long time-gap, magmatic activity resumed with small volume Na-alkalic basalts (2.5 Ma), K-alkalic at 1.6 Ma and ultrapotassic magmas at 1.3 Ma. The SiO_2 and $^{87}Sr/^{86}Sr$ show that AFC processes were unimportant in the formation of these magmas (Fig. 6b), typical for a crustal and lithospheric origin that evolved in an extensional setting. Th/Y shows a narrow variation parallel with the mantle array, higher in the adakite-like magmas. Younger adakite-like magmas and transitional basaltic andesites have the lowest Th/Y that suggests mixing of asthenospheric and lithospheric magmas (Fig. 7b). Generation of most adakite-like magmas occurred via delamination and partial melting of a high density garnet-bearing (eclogitic) lower crust (Seghedi et al., 2007). Missing geochemical data makes it difficult to understand the generation of garnet-bearing rocks in this area, however we suppose that they have a similar origin as those from the Pannonian basin (upper lithosphere/crust level) and show evidence of extension and rapid emplacement (see Harangi et al., 2001).

Lithosphere breakup during extreme block rotations (~60°) at 14-12 Ma was responsible for extension with core-complex formation at the easternmost continuation of the Bekes basin. This mainly led to decompression melting of an enriched lithospheric mantle/crust source during major extension and by asthenosphere uprise (Figs. 4, 10).

The magmas generated after 2.5 Ma suggest a close relationship with Pliocene inversion tectonics along the South Transylvanian fault due to the push of Adria, with small volume melt generation (Na-alkalic, K-alkalic and ultrapotassic) from diverse lithospheric and asthenospheric sources (see Seghedi et al., 2010, Fig. 6b).

5.2. Transylvania basin system

5. 2.a Northern part of Transylvanian Basin

Here calc-alkaline magmatism was generated at 12-8 Ma in the northern part of the Tisza-Dacia block following the Dragoş Vodă - Bogdan Vodă transcurrent fault system (Pécskay et al., 2009). It is entirely intrusive, ranging from basalts to rhyolites. Garnet-bearing varieties occurred at 9.5-10.5 Ma. The intrusive bodies may represent magma chambers that fed surface volcanism Their appearance may be related to strong erosion during the uplift of middle crust metamorphic rocks (at least 1 km) as shown by exhumation histories (e.g. Tischler et al. 2006, 2008; Gröger et al., 2008).

The rocks scatter in the SiO_2 vs. $^{87}Sr/^{86}Sr$ or Nb/Y diagrams, and also show a large variation in Th/Y ratios. AFC processes are suggested, but since the source was highly heterogeneous these are difficult to demonstrate for a specific case. Most probably each body evolved independently with specific AFC and/or magma mixing processes; a conclusive petrological study is still lacking.

A sinistral transtensional stress regime at 12-10 Ma along the Dragoş Vodă transcurrent fault system at lithosphere scale (following a transpressional phase at 16-12 Ma) (Gröger et al., 2008) controlled the generation and emplacement of the intrusive bodies (Figs. 2, 4); rapid emplacement is suggested by the presence of garnet-bearing andesite and dacite bodies. The event was coeval with the 12 Ma differential rotation and decoupling from the main body of Transcarpathia (NE-ALCAPA) that occurred at or near the Hernád fault (Márton et al, 2007) (Fig. 4). The mechanism of magma generation was decompression melting of the local heterogeneous mantle lithosphere and lower crust, as suggested by low mineral oxygen isotope values (Papp et al., 2005). This may be the result of oblique convergence of Tisza–Dacia with the NW–SE striking European margin, evidenced by eastward thrusting in the external Miocene thrust belt (Mațenco and Bertoti, 2000).

5. 2b. Eastern part of Transylvanian Basin

This area is a continuation of the previous area and consists of calc-alkaline volcanism that occurred along the easternmost margin of the rigid Dacia block, in the front of European Platform, forming the Călimani-Gurghiu-North Harghita volcanic chain, known for its diminishing age and volume southwards at 10-3.9 Ma (Szakács and Seghedi, 1995). It marks the end of subduction-related magmatism along the post-collision front of the European convergent plate margin (Mason et al, 1996, 1998).

The rocks show homogeneous $^{87}Sr/^{86}Sr$, but a linear trend of Th/Y vs Nb/Y that reflects a common mantle source considered to be the metasomatized lithospheric mantle wedge (Fig. 6b). Fractionation or AFC are characteristic for each main volcanic area, suggestive of lower to middle crust magma chamber processes (Mason et al., 1996) (Figs. 6b, 7b).

The locations of the eruption centers in the CGNH chain are concentrated at intersections of crustal fault system that propagated from N to the S along the arc at the eastern boundary of Dacia, suggesting NNW–SSE striking sinistral transtensional faulting (Fielitz and Seghedi, 2005). The relatively large volume of magma in the CGNH is difficult to relate exclusively with a transtensional fault mechanism, as in the northern part of the Transylvanian Basin. Its generation may be associated with asthenosphere uprise, explained by progressive break-off of the Miocene subducted slab (Seghedi et al., 2004a) (Figs. 4, 10). The evidence is the present high heat flow corresponding exclusively to the volcanic area (Tari et al, 1999; Demetrescu et al., 2001). Along-arc temporal distribution of the volcanism has been already explained as gradual slab detachment following an oblique subduction stage (Mason et al. 1998; Seghedi et al., 1998; Wortel and Spakman, 2000).

5. 2c. South-Eastern part of Transylvanian Basin

The South Harghita volcanic area is the continuation of the CGNH volcanic chain. Here at ca. 3 Ma following a time-gap, magma compositions changed suddenly to adakite-like calc-alkaline and continued until recent times (< 0.03 Ma). This volcanism was interrupted at 1.6-1.2 Ma by simultaneous generation of Na- and K-alkalic varieties in nearby areas, suggestive of various sources and melting mechanisms (Downes et al., 1995; Mason et al., 1998; Seghedi et. al., 2004a, b). The specific geochemistry is revealed by higher Nb/Y and Th/Y ratios and lower $^{87}Sr/^{86}Sr$ as compared to the CGNH chain (Figs. 6b, 7b).

This complex magmatism is situated in front of the Moesian platform and was associated with two main geodynamic events: (a) slab-pull and steepening, with opening of a tear-window in the vertical Vrancea lithospheric block hanging into the asthenospheric mantle (forming adakite-like calc-alkaline magmas) and (b) inversion tectonics along reactivated fault systems that allowed decompression melting of asthenospheric and lithospheric sources, thus generating the Na- and K-alkalic magmas. A detailed explanation and a profile interpretation were given recently (Seghedi et al., 2010, Fig. 6a).

5. 3. Areas with Miocene to recent magmatism situated at the southern boundary of CPR

5. 3a. Sava Zone

This area coincides with the Sava depression bounded by a E-W system of faults (Pamić, 1998, Pamić and Balen, 2001) bordering the southern part of Tisza-Dacia block (Fig. 2). Small volume calc-alkaline volcanic rocks found in this depression, mostly described from boreholes, in several successive periods between 22.8 and 7.4 Ma, with K-alkalic rocks at 17.5-15.4 and 9.8-7.4 Ma (Pécskay et al., 2006). The calc-alkaline rocks show high $^{87}Sr/^{86}Sr$, large SiO_2 range and low Nb/Y, without significant AFC (Figs. 8a, b). The K-alkalic rocks have higher Nb/Y and Th/Y that reflect a different source considered to be a metasomatized lithospheric mantle wedge (Fig. 8b). Balen and Pamić (2001) suggested slab-breakoff for their generation. During Miocene the area acted as a strike-slip wrenching at the boundaries of the former blocks (Tari, 2002) and we suggest that it was reactivated several times in transpressional to transtensional mode, generating magmas via decompression melting of heterogeneous lithospheric mantle, sometimes influenced by AFC.

5. 3b. Vardar Zone

Miocene-Pliocene magmatism characterizes the Serbian part of the Vardar zone and its extension southward to FYROM and Greece. Ultrapotassic, shoshonitic and high K-alkalic magmas were erupted at 23-21 Ma; at 12.9-10 Ma shoshonitic magmas were generated. Magmatism ended with K-alkalic and ultrapotassic magmas at 9.1-1.5 Ma, becoming younger southward. Following Cvetković et al. (2004) and Prelević et al. (2005), we separated low SiO_2 shoshonitic and ultrapotassic rocks that, due to their special mineralogical features (kamafugite affinities), are not properly distinguished in the TAS diagram from those that fall normally in the alkalic and ultra-alkalic fields. The kamafugite rocks with lower $^{87}Sr/^{86}Sr$, higher Nb/Y and lower Th/Y reflect a different mantle source enriched during Mesozoic subduction events which corresponds to the Western Vardar, since the other group belongs to the eastern Vardar (Figs. 8a, b) (Prelević et al., 2005; Schmid et al., 2008).
The small volume magmas, were generated in a metasomatized depleted mantle, more enriched for K-alkalic rocks compared with ultra-potassic during various transtensional events that affected the whole Vardar zone during Oligocene-Miocene times (Prelević et al., 2005, 2007). The calc-alkaline dominantly crust-derived intrusions formed at 20-16 Ma, contemporaneous with volcanic activity, were induced by extensional exhumation located in the footwall of core complexes in the Vardar zone. This local extensional area seems to have accommodated the extension in the Pannonian basin during Miocene times (Cvetković et al., 2007; Fodor et al, 2008; Schefer et al., 2010; Koroneos et al., 2010; Mațenco et al., 2010b).

6. Discussion

6.1. Extension and magma generation

Extension tectonics related to magmatism in this review are mainly of two types: core502 complex type and transtensional faulting. Core-complex type extension was initially illustrated for the whole lithosphere, showing the implication of crustal flow (e.g. Buck, 1991; Wernike, 1992). The concept evolved mainly toward understanding crustal flow involving extension and thinning of thick and hot crust, leading to the formation of a large diversity of metamorphic core complexes cored by migmatite domes that are bounded by vertical or usually low angle faults (Fayon et al., 2004; Dunkl and Frisch, 2002). Models of crustal partial melting related with crustal flow have been also developed (e.g. Corti et al., 1993; Ray et al., 2009), however such models rarely imply that types of magmatism can be generated at lithosphere scale. Tari et al., (1999) recognized core-complex type extension at the scale of Pannonian basin, remarking on its association with magmatism. Roşu et al. (2004) and Seghedi et al. (1998, 2004a, 2007) related the magmatism in the Apuseni Mts. area with extension, however without explaining the specific implication of core-complex type extension. In the Balkan, Aegean or Anatolian areas, many authors recognize the temporal association of extension and core-complex massifs with widespread volcanic activity (e.g. Aldanmaz et al., 2000, Bonev et al., 2006; Bozkurt, 2004; Dilek and Whitney, 2000; Dilek and Altatuniak, 2007; Karaöglu et al., 2010; Ersoy et al, 2010; Marchev et al., 2004). In most cases the models suggest partial melting formed by asthenospheric upwelling that caused mantle delamination.

We suggest that major extension associated with crustal flow was crucial in magma generation on both sides of the Moho boundary, along with asthenosphere upwelling. Extension caused heating of the crust and allowed generation of crustal partial melts which, in association with the presence of subduction-related mantle lithospheric magmas, made the crust more ductile and facilitated the tectonic processes (Figs. 4, 9, 10).

Geological and geophysical observations, together with new experimental data, demonstrate coupling of the lower crust and upper mantle beneath major transcurrent faults (e.g. Vauchez et al., 1998). In these cases, lithosphere deformation leads to small volume magma production that often accommodates and assists the lithospheric deformation (e.g. Vaughan and Scarrow, 2003). Initiation of faulting in heterogeneous enriched lithospheric mantle is followed by adiabatic decompression melting and ascent of magma-driven fractures to shallow levels similar to magmatism along the North and South Transylvanian basin. Transtension-related magmatism in the CPR is of small volume, at the margins of a strong lithospheric block (Dacia), mirroring the local lithosphere composition and consisting of a large variety of magmas: calc-alkaline, K-alkalic (shoshonitic) and ultrapotassic. Calc535 alkaline magma generation (e.g. north and east of the Transylvanian basin) was the result of catastrophic oblique collision that led to crustal fragmentation, transtensional faulting (Gröeger et al., 2008; Fielitz and Seghedi, 2005), exhumation and final collision in the East Carpathians (Maţenco et al., 2007) and eventually by progressive break-off of the Miocene subducted slab (e.g. Mason et al. 1998). Sometimes Na-alkalic asthenospheric magmas are also extruded along such transtensional fault systems. Generation of Na-alkalic asthenospheric magmas is characteristic for various geodynamic situations (see Lustrino and Wilson, 2007) and was controversially explained at the scale of the CPR (see Harangi and Lenkey, 2007; Seghedi et al., 2004b). Here we link it mostly with post-collisional lithospheric mantle perturbation (as the one generated by Adria push) that allowed small degree asthenospheric melting. In the case of Central Slovakian volcanic field, the small volume intermittent Na-alkalic volcanism showing the longest interval of activity, from 8 Ma to recent times, a mantle plume scenario should be considered.

6.2 Temporal variation and magma generation

Temporal variation is a useful indicator of the degree and depth of partial melting during the evolution of a magmatic province. The most significant observation is the Pannonian basin example (1b in Fig. 2) that shows an age range from 22 Ma to recent times (Fig. 3). Here crustal melting occurred during catastrophic lithospheric extension, long before asthenosphere melting or uprise followed by mantle lithosphere melting. The explosive eruption of crustal magmas indicates the presence of a large volume of volatiles that helped to reduce the solidus temperature during partial melting. Initially the andesites have been derived from sub-continental mantle lithosphere with melting at progressively shallower levels (crust /mantle boundary) in the period of maximum extension with systematic change and diminishing volume. The partial melts of the mantle lithosphere rising upwards stalled and ponded at the base of the crust that usually is less dense, forming underplating magma reservoirs that further fractionated or mixed with the crustal melts. In the next phase the small degree melting of the deeper asthenospheric mantle produced Na-alkalic magmas. In time the large amount of melting in the lithosphere diminished and asthenosphere uprise initiated melting, showing small volume and intermittent eruption rates. This is a similar to the Basin and Range (Bradshaw et al., 1993) where a decrease in magma generation suggests relaxation of geotherms following extension that subsequently requires the melting zone to reach greater depths (Na-alkalic magmas). The estimated magma volumes and eruption rates correlate with the amount of extension that was most significant in the Pannonian area, and less significant in Transcarpathia and the Apuseni Mountains (Fig. 3) (Lexa et al., 2010).

6.3. Petro-chemistry and magma generation

Geochemistry and isotope geology are crucial in understanding the way magmas have been formed and reached the surface. They provide a simple way to separate the crustal, lithospheric and asthenospheric magma sources. Correlation with petrographic data is essential, as for example the presence of garnet-bearing calc-alkaline rocks that implies high pressure crystallization from a hydrous source (lower crust or upper mantle), followed by rapid magma ascent to the surface (Harangi et al, 2001, 2007). Garnet-bearing magmas are suggested to be generated mostly via partial melting of the metasedimentary crust in areas of anomalously high temperature, which produces peraluminous felsic magmas which generally crystallize at depth as leucogranitic plutons (e.g. Clemens, 2003). More rarely, these viscous and relatively cool magmas may rise and extrude onto the Earth‘s surface, however they require an extensional setting to get quickly to the surface, in order for the garnet to survive (e.g. Harangi et al., 2001). A variable delay (up to 3 Ma) between magma production and lava extrusion found in crustal-derived Miocene garnet-bearing rocks in the Betic area, by using U-Pb SHRIMP geochronology shows that is difficult to constrain the real age of crustal melting and judging the data should be done with caution (Cesare et al., 2008).
Variation of SiO_2 vs. alkali and $^{87}Sr/^{86}Sr$ are essential to evaluate magma chamber processes, as FC or AFC. Trace element variability, e.g. Nb/Y vs Th/Y, suggests that there are no important differences between crustal sources (with highest Th/Y, as most metasomatized) and a subduction-related lithospheric mantle source (Figs. 7a, b, 8b). During the peak periods of extension we expect higher degrees of partial melting to correlate with large magma volumes, whereas lower degrees of melting form smaller volumes. The correlation between high volume magma and low Nb/Y and high Th/Y may suggest a high crustal and mantle potential temperature, the volume of melting increasing with both the amount of extension

and the mantle potential temperature. In contrast magmas showing small volumes have high Nb/Y that do not require high mantle potential temperature for decompression melting. In the CPR from the Miocene to the present there were two regions of melting: one was situated in the mantle lithosphere and another in the asthenosphere, the latter always following the main periods of extension.

7. Conclusion

This review of the tectonic setting, timing and geochemistry of post-collisional volcanism in the CPR leads us to constrain the role of orogenic processes on melt production and migration as follows:

- The geodynamic record of the CPR may be explained as the result of continuous northward Aegean subduction and accretion since the Cretaceous, followed by widespread extension, sometimes with exhumation of metamorphic rocks as a result of west-ward slab retreat (roll-back) of several blocks in a small landlocked basin since the late Miocene. In the Aegean and Anatolian areas where slab retreat was southward, similar extension and lower-middle crust exhumation occurred (e.g. Dilek et al., 2009; Jolivet and Brun, 2010; Tirel et al., 2009; van Hinsenberg et al, 2010).
- The magmatism is post-collisional and the absence of the proper arc-type volcanism makes the term ―back-arc‖ in the CPR inappropriate. Extension was the main process during which magmas were generated. The main tectonic mechanisms imply both core-complex extension and transtensional faulting. Blocks with weak lithosphere (e.g. Alcapa, Tisza) tended to break and disintegrate: magmas formed above and at the boundaries (e.g. Pannonian basin). Blocks with relatively strong lithosphere (Dacia) formed magmas at destructive boundaries along transtensional faults (e.g. at margins of the Transylvanian Basin).
- Magmatism mirrors the composition of its source (mantle lithosphere, crust and asthenosphere) and occurred during catastrophic tectonic events, acting as a lubricant for lithospheric block extension.
- Decompression melting of a hydrated and/or metasomatized mantle was the main magma-generating mechanism.
- Magmatic activity influenced the extension dynamics. Magma generation and tectonic activity were contemporaneous. Magmatism enhances lithospheric weakness in several ways: Decompression mantle melting and further underplating by asthenosphere uprise and assisting crustal melting correlate with increased amount of extension. K-alkalic and large volume of calc-alkaline magmas are formed. Decrease in magma volume following extension marks the transition from lithosphere melting to asthenosphere melting. Magmatism enhances lithospheric weakness and controls timing and location of transtensional faulting: calc-alkaline magmas are generated via crust and mantle melting; slab breakoff or delamination associated with asthenosphere uprise enhanced the magma production. Small volume K-alkalic, ultra-K and Na635 alkalic asthenospheric magmas used transtensional faults to aid eruption.

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Figure caption

Fig. 1. Simplified map (after Harangi et al., 2006) with the distribution of Tertiary to Quaternary volcanic rocks in Europe. The geodynamic system is controlled by subduction of the Africa megaplate with Adria promotory (that separates west Mediterranean area -WM from the east Mediterranean area -EM) with the European megaplate. Miocene and Oligocene volcanic rocks are separated only in Carpathian-Pannonian region and Balkan area. Dark blue areas indicate oceanic crust. Abbreviations: ECRIS, European Cenozoic Rift System; BAR, Betic-Alboran-Rif province (Ab, Alboran; Be, Betic; RTG, Gourougou-Trois Furches-Ras Tarf (Rif); Or, Oranie; Ca, Calatrava; Ol, Olot); CM, Central Mediterranean (Sa, Sardinia; Si, Sisco; Tu, Tuscany; Rp, Roman province; Ca, Campania; Vu, Vulture; Va, Vavilov; Ma, Marsili; Us, Ustica; Ai, Aeolian Islands; Et, Etna; Hy, Iyblei; Pa, Pantelleria); PIL, Periadriatic-Insubric Line (Be, Bergell; Ad, Adamello; Ve, Veneto); CPR, Carpathian-Pannonian region (WC, western Carpathians; EC, eastern Carpathians; Ap, Apuseni); Balkans -BK; (S-Sava Zone; VZ- Vardar zone; Di-Dinarides ,He-Hellenides); Rhodope (R- Rhodope Th-Thrace); AA, Aegean–Anatolia (Sa, Santorini; WA, Western Anatolia; Ku, Kula; Af, Afyon; Ko, Konya; Ga, Galatia; CA, Central Anatolia).

Fig.2. Sketch map of the Carpathian-Pannonian region (simplified after Schmid et al., 2008) showing the Miocene-Quaternary magmatism (simplified after Pécskay et al., 2006) without Miocene-Quaternary sedimentary cover. The regions where volcanism has developed volcanic edifices are encircled: 1a. Styrian basin, 1b. Western Carpathians and Pannonian basin, 1c. Transcarpathian basin, 1d. Zărand basin and Apuseni Mountains belonging to the main Pannonian basin system; 2a Northern part of Transylvanian basin, 2b Eastern part of Transylvanian basin, 2c South-eastern part of Transylvanian basin, belonging to the Transylvanian basin system; 3a Sava zone and 3b. Vardar zone that compose the southern boundaries of CPR system. Time intervals of volcanic activity for each region are given.

Fig 3. Simplified and updated spatial and temporal distribution (after Pécskay et al., 2006) distinguishing the development of magmatism as related to the main sedimentary basins (Pannonian and Transylvanian) generated during the main Miocene-Quaternary tectonic events in CPR. The events include tectonic escape and extension of the main ALCAPA and Tisza-Dacia blocks, resulting in a simultaneous breakup and various time of collision, last one in the East Carpathians and Pliocene-Quaternary inversion tectonics derived from the push of Adria. The contemporaneous magmatism situated at the southern limit of CPR that includes the Sava and Vardar zones ophiolitic accretionary prisms is shown.

Fig. 4. Sketch map of the Carpathian-Pannonian region (simplified after Schmid et al., 2008) showing the Miocene-Quaternary geodynamic developments in CPR with Miocene-Quaternary sedimentary cover. Time intervals of geodynamic processes that led to final collision in different parts of the system showing a shift from west to east during major extensional processes (wide rift and core-complex types), main rotational events and young inversion processes due to Adria push are given. The curved arrows indicate the major rotation processes during the extension. Map legend as in the Fig. 2.

Abbreviations: SB- Styrian basin; MHL-Mid Hungarian line; PB- Pannonian basin; TcB1142 Transcarpathian basin; H. fault-Hernad fault; D.V. fault –Dragoş Voda fault system; S-TS1143 South Transylvanian fault system. A-A‘, B-B‘, C-C‘ and D-D‘ are the profile lines that will be shown in figures 9 and 10. Metamorphic core complex of lower-middle crust are shown by oval shape with interrupted lines (after Tari et al., 1999; Csontos and Vörös, 2004; Soták et al., 2000; Dunkl and Frisch, 2002). Fault and fracture pattern that governed the Miocene–

Quaternary volcanism and basinal subsidence along a transtensional corridor in the CGH1148 mountain region is after Fielitz and Seghedi (2005). Continuous oval shape shows the Vrancea seismogenic area.
Fig. 5. SiO_2 vs. $Na_2O + K_2O$ (Le Bas et al, 1986) for Carpathian–Pannonian region volcanic rocks. (1) Calc-alkaline, (2) Na-alkalic, (3) K-alkalic and (4) Ultra-K have been separated. Mantle type rocks show low silica and crustal-derived rocks show high-silica content. Data from Ali et al., 2011, submitted; Downes et al., 1995a, b; Embey-Isztin et al., 1993; Harangi et al., 1995a, b, 2001a, 2005, 2007; Klébesz et al., 2009; Lukács et al., 2009, 2010; Mason et al, 1996, 1998; Nițoi et al., 2002; Ntaflos et al., 2007; Papp et al., 2005; Pécskay et al., 1995a, b, 2006; Póka et al., 2004; Roșu et al., 2004; Salters et al, 1988; Seghedi et al., 1995, 2001, 2004a, 2004b, 2005, 2007, 2008; Trua et al., 2006; Tschegg et al., 2010; Vinkler et al., 2007. Average continental crust after Rudnick and Fountain (1995) and average local continental crust from Mason et al. (1998).
Fig. 6a. SiO_2 vs $Na_2O + K_2O$ and SiO_2 vs $^{87}Sr/^{86}Sr$ diagrams for 1a. Styrian basin, 1b. Western Carpathians and Pannonian basin, 1c. Transcarpathian basin including sign legend. Data as in Fig. 5. Symbols are shown.
Fig. 6b. SiO_2 vs $Na_2O + K_2O$ and SiO_2 vs. $^{87}Sr/^{86}Sr$ diagrams for 1d. Zărand basin and Apuseni Mountains, 2a Northern part of Transylvanian basin, 2b, c Eastern and South-eastern part of Transylvanian basin. Data as in Fig. 5 and symbols as in Fig. 6a.
Fig. 7a. Nb/Y vs. Age (Ma), SiO_2 and Th/Y diagrams for 1a. Styrian basin, 1b. Western Carpathians and Pannonian basin, 1c. Transcarpathian basin. Data as in Fig. 5 and symbols as in Fig. 6a.
Fig. 7b. Nb/Y vs. Age (Ma), SiO_2 and Th/Y diagrams for 1d. Zărand basin and Apuseni Mountains, 2a Northern part of Transylvanian basin, 2b, c Eastern and South-eastern part of Transylvanian basin. Data as in Fig. 5 and symbols as in Fig. 6a.
Fig. 8a. SiO_2 vs. $Na_2O + K_2O$ and SiO_2 vs. $^{87}Sr/^{86}Sr$ diagrams for Sava zone and Vardar zone including sign legend. Data from Balen and Pamić, 2001; Cvetković et al., 2004; Pamić et al., 1995; Pamić and Balen, 2001; Prelević et al., 2005, 2007; Yanev et al., 2008. Symbols are shown.
Fig. 8b. Nb/Y vs. Age (Ma), SiO_2 and Th/Y diagrams for Sava zone and Vardar zone. Data and symbols as in Fig. 8a.
Fig. 9. Interpretative profiles for the style of magma generation in the Pannonian and West Carpathians- as wide rift system A-A‘ (after Tari et al., 1999) and for Transcarpathian basin- B-B‘, showing core-complex extension and rotation mechanism and crust-mantle lithosphere melting. In both cases asthenosphere upwelling is crucial for magma generation. Abbreviations in A-A‘ profile: 1. Initiation of the volcanism as crustal-derived rhyolitic explosive volcanism, followed by 2. andesitic lithosphere-derived mostly effusive volcanism and further mixing with crustal-derived melts and 3. Na-alkalic volcanism, as both explosive and effusive (see also Fig. 3).
Fig.10. Interpretative profiles for the style of magma generation in the Zărand basin and Apuseni - as core-complex extension and rotation system C-C‘ (after Tari et al., 1999) and for East Carpathians- D-D‘ (after Mațenco et al., 2010a), showing slab detachment following an oblique subduction stage and mantle lithosphere melting. In both cases asthenosphere upwelling is crucial for magma generation. Abbreviations in C-C‘ profile: 1. Initiation of the volcanism as crust-derived volcanism, followed by 2. normal and adakite-like calc-alkaline rocks, as mostly effusive volcanism and 3. Small volume transitional basaltic andesites,

suggesting asthenosphere-derived magmas mixed with lithospheric magmas (see also discussion at Apuseni area).

Figure

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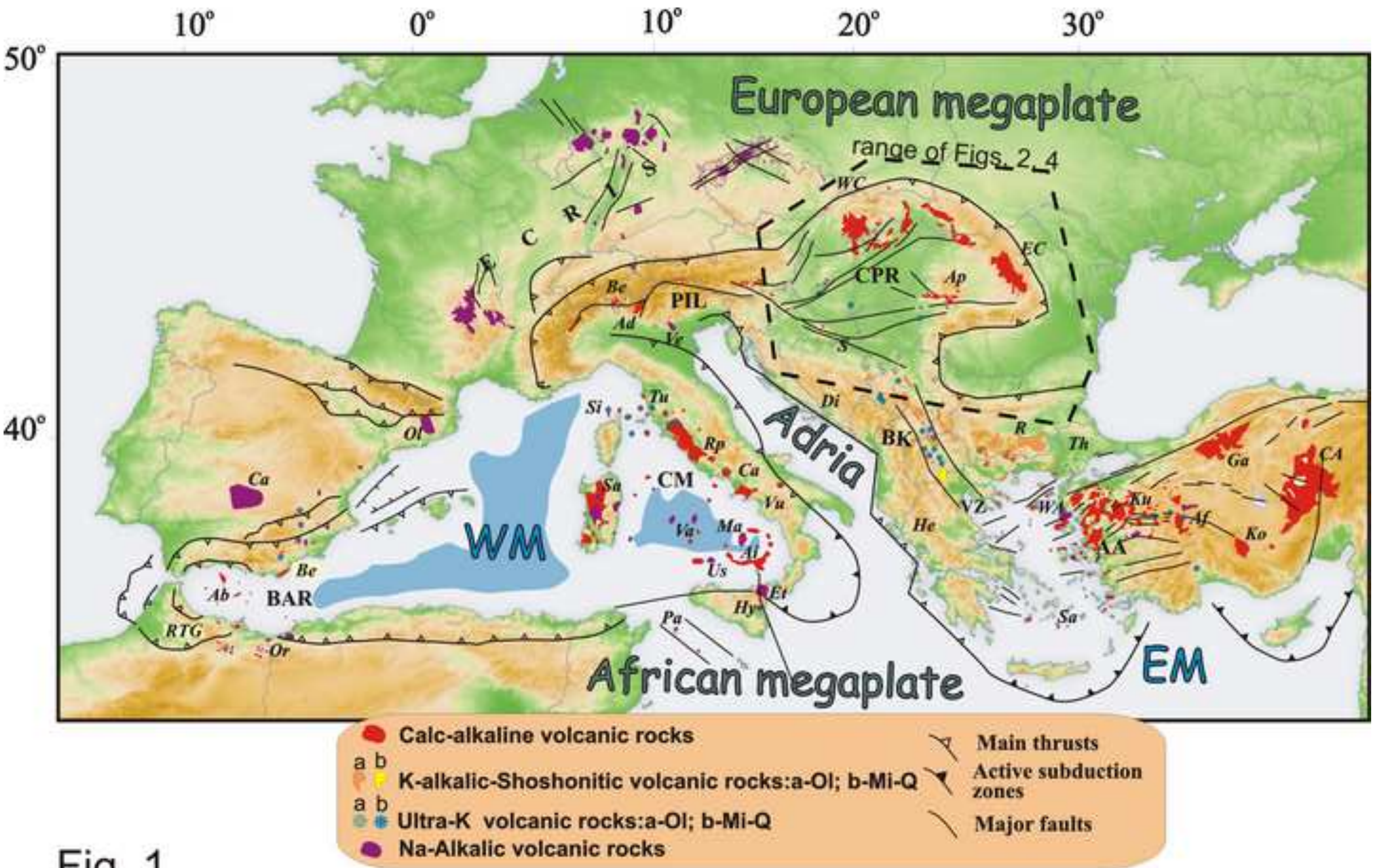

Fig. 1

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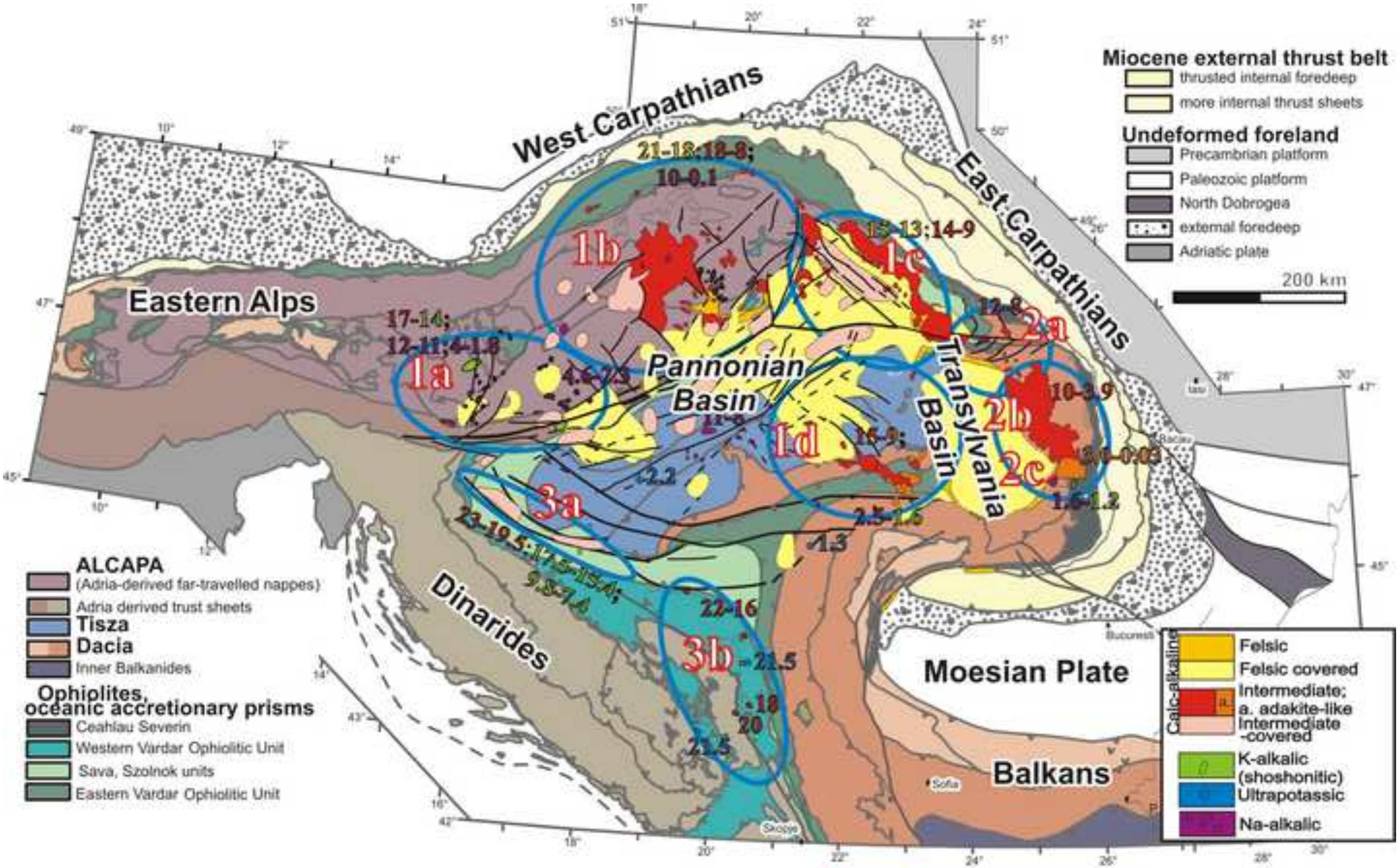

Fig. 2

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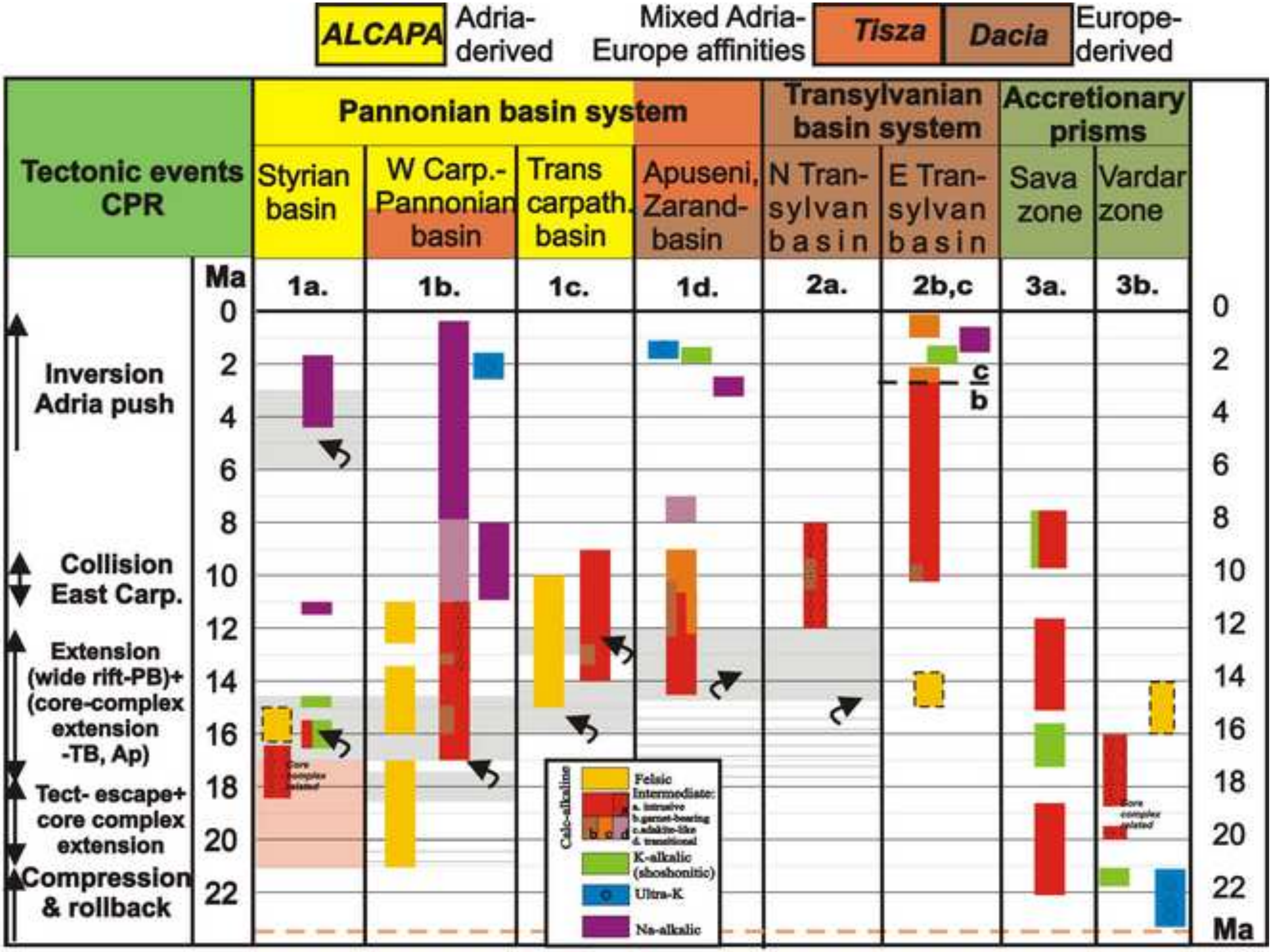

Fig.3

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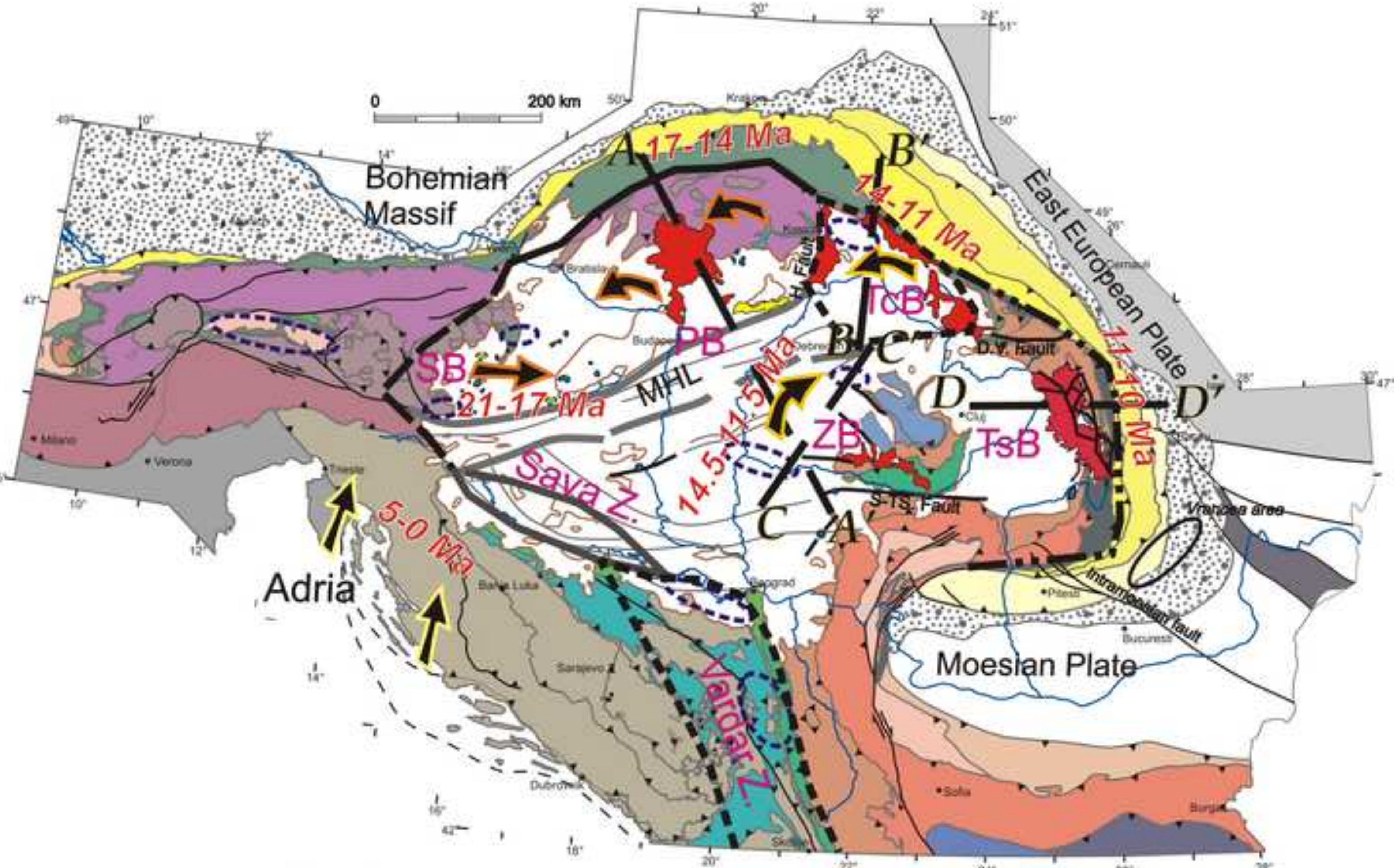

Fig. 4

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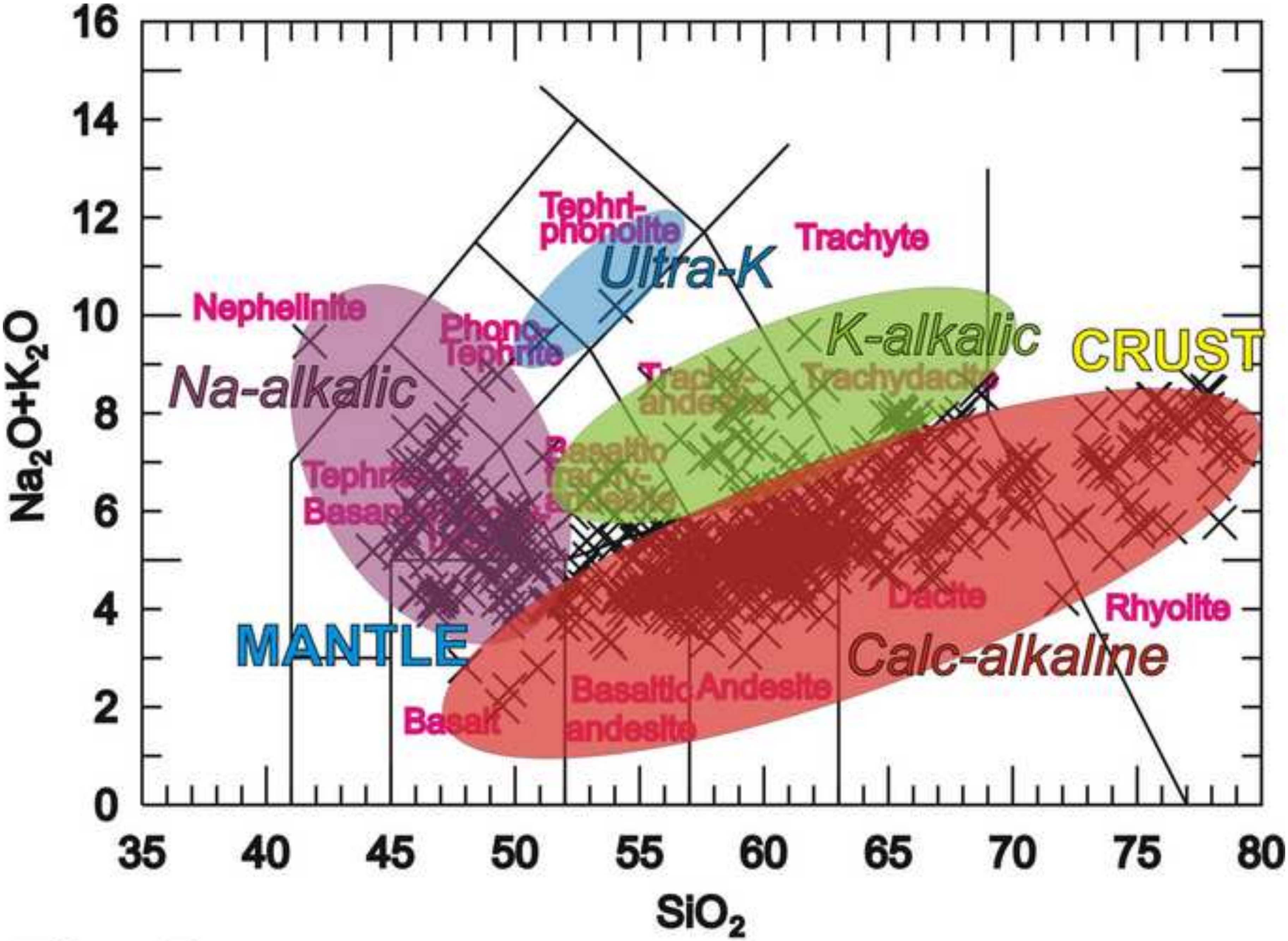

Fig. 5

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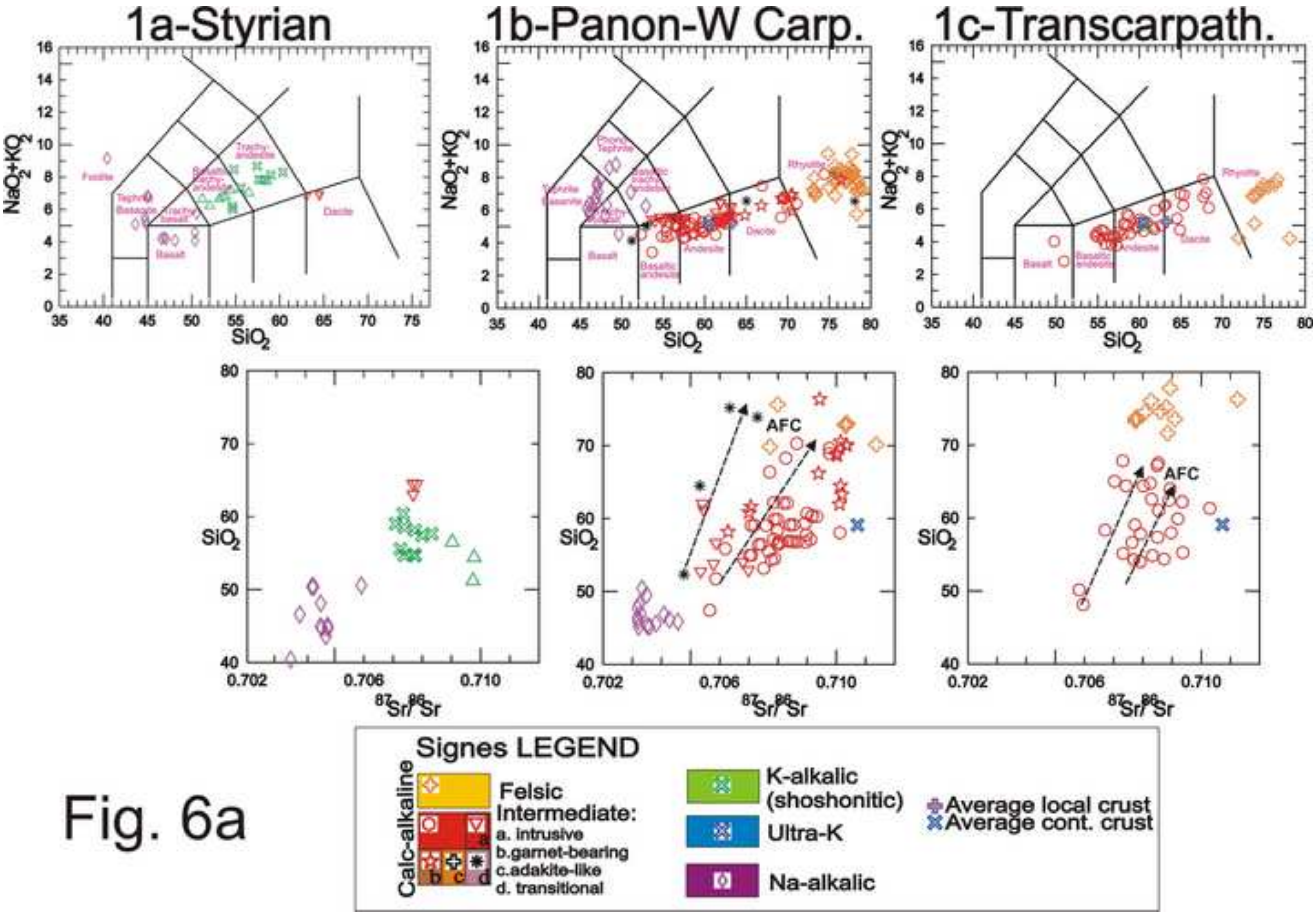

Fig. 6a

Figure

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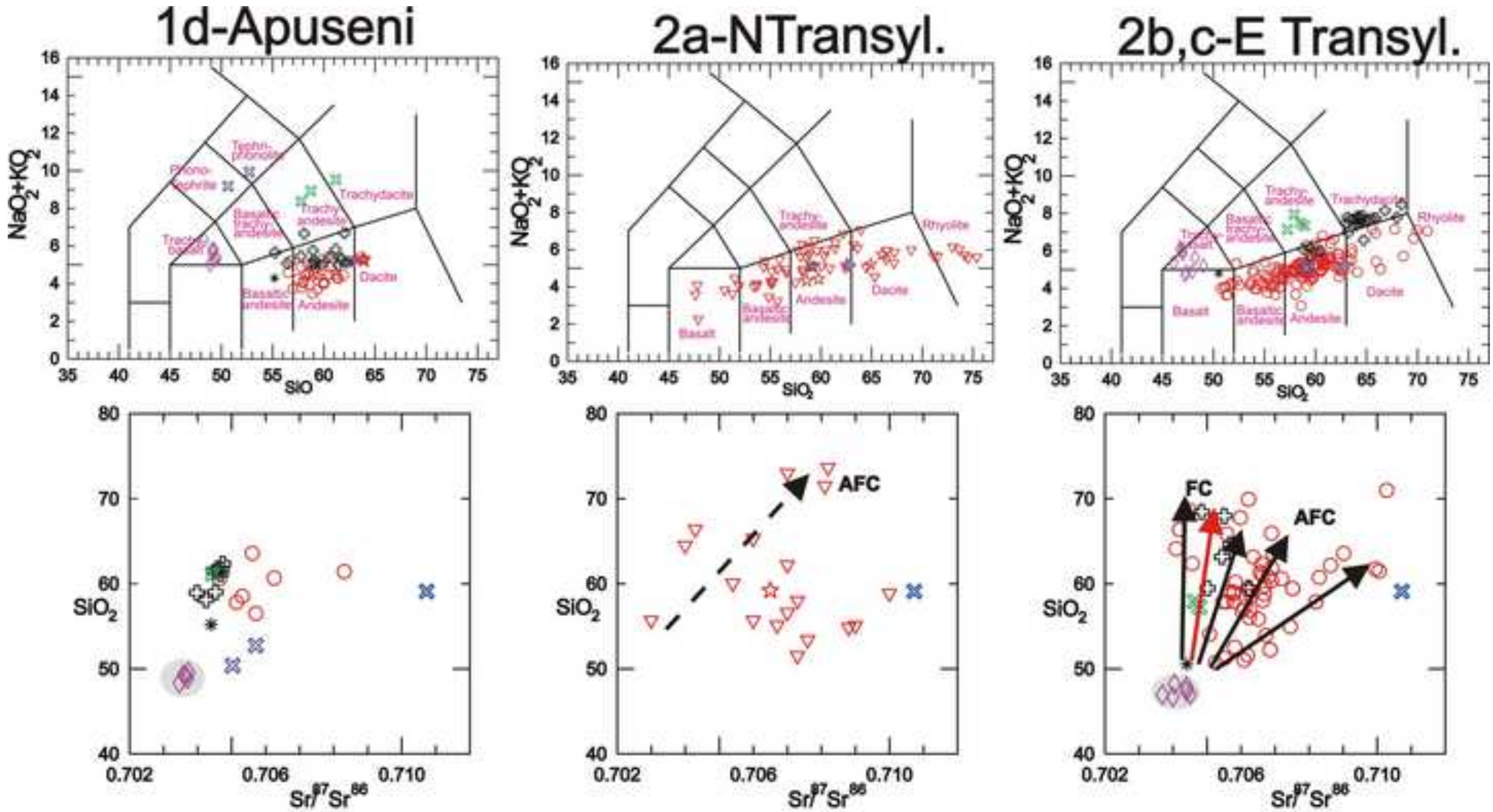

Fig. 6b

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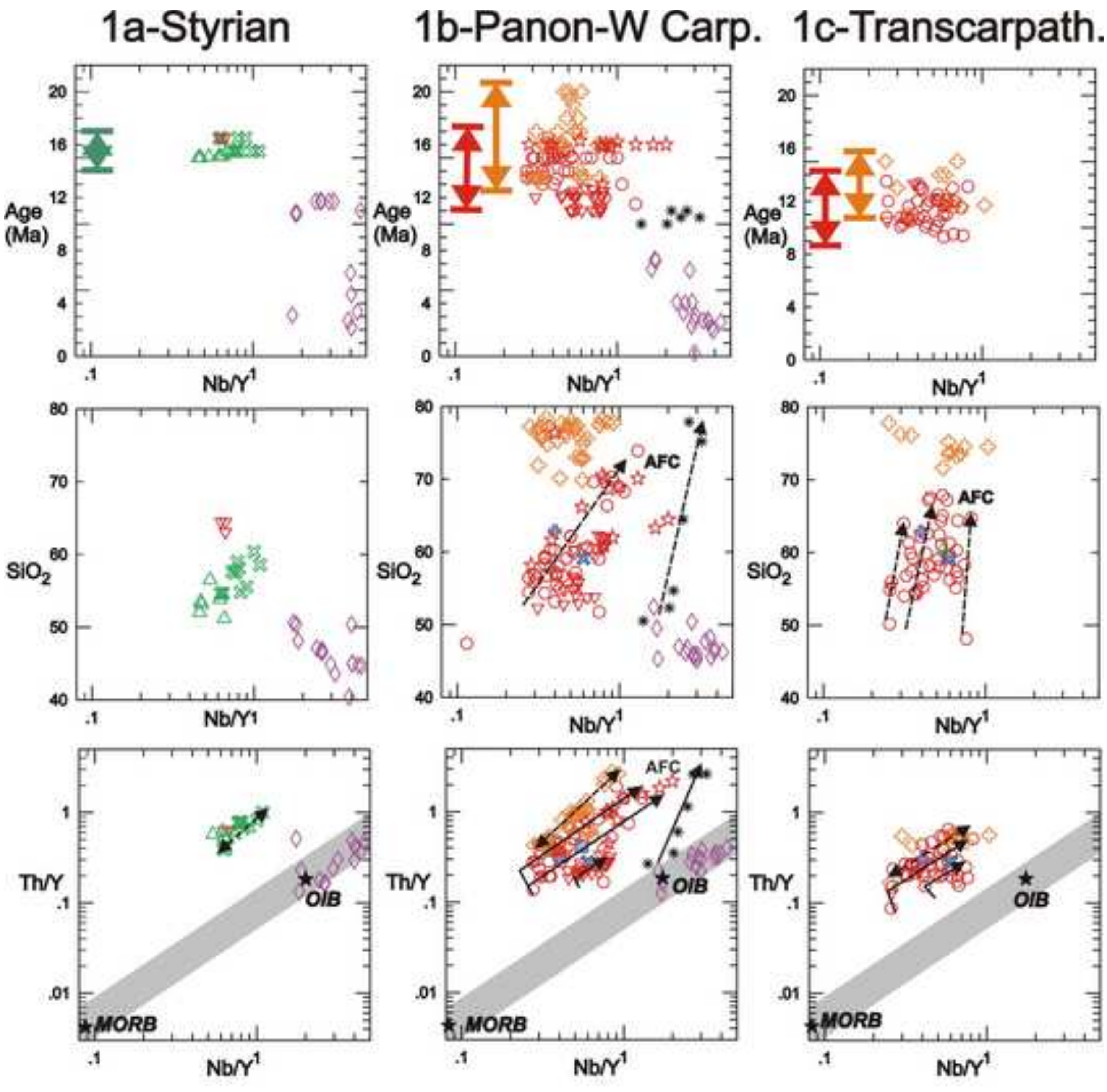

Fig.7a

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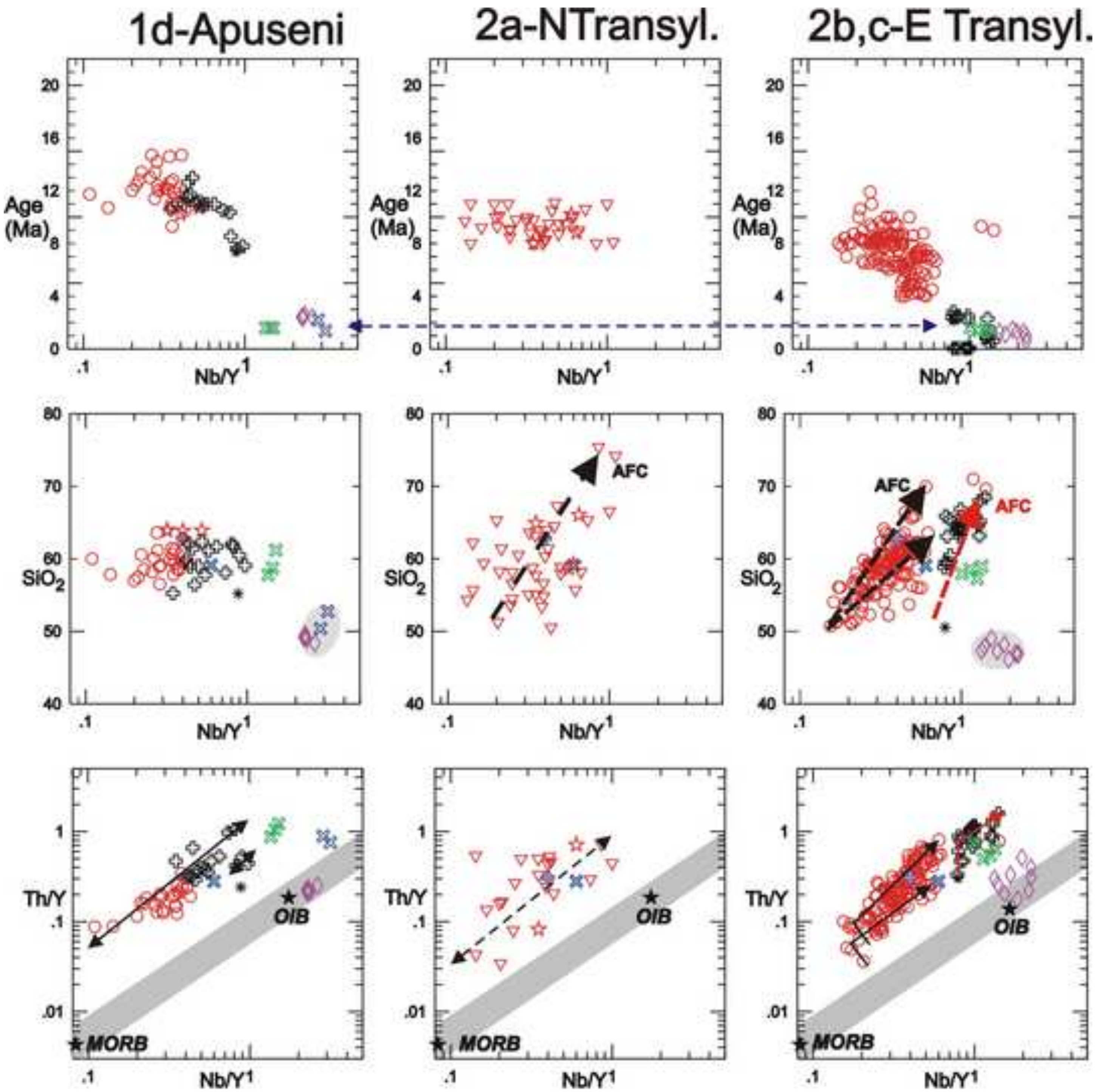

Fig. 7b

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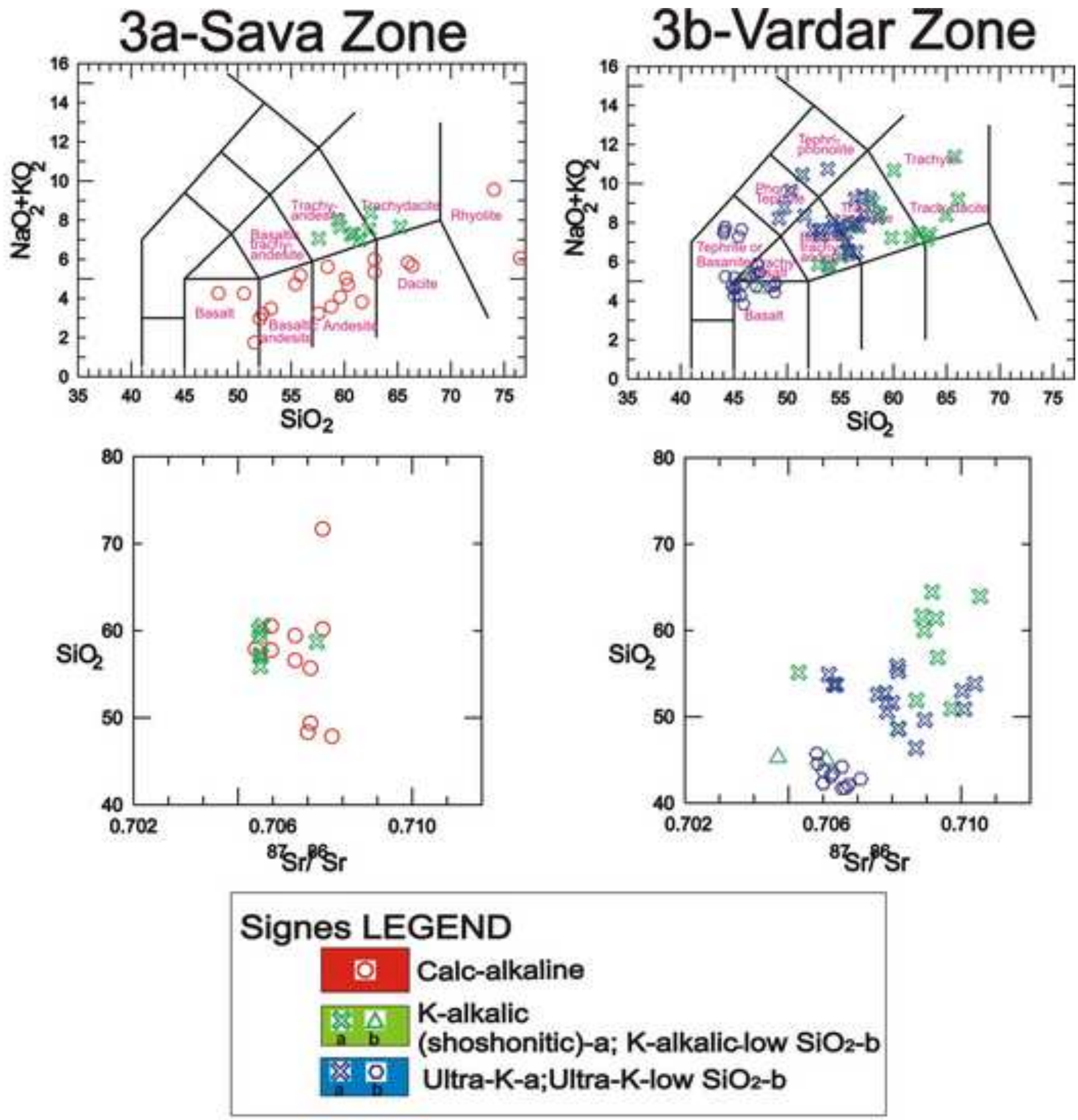

Fig. 8a

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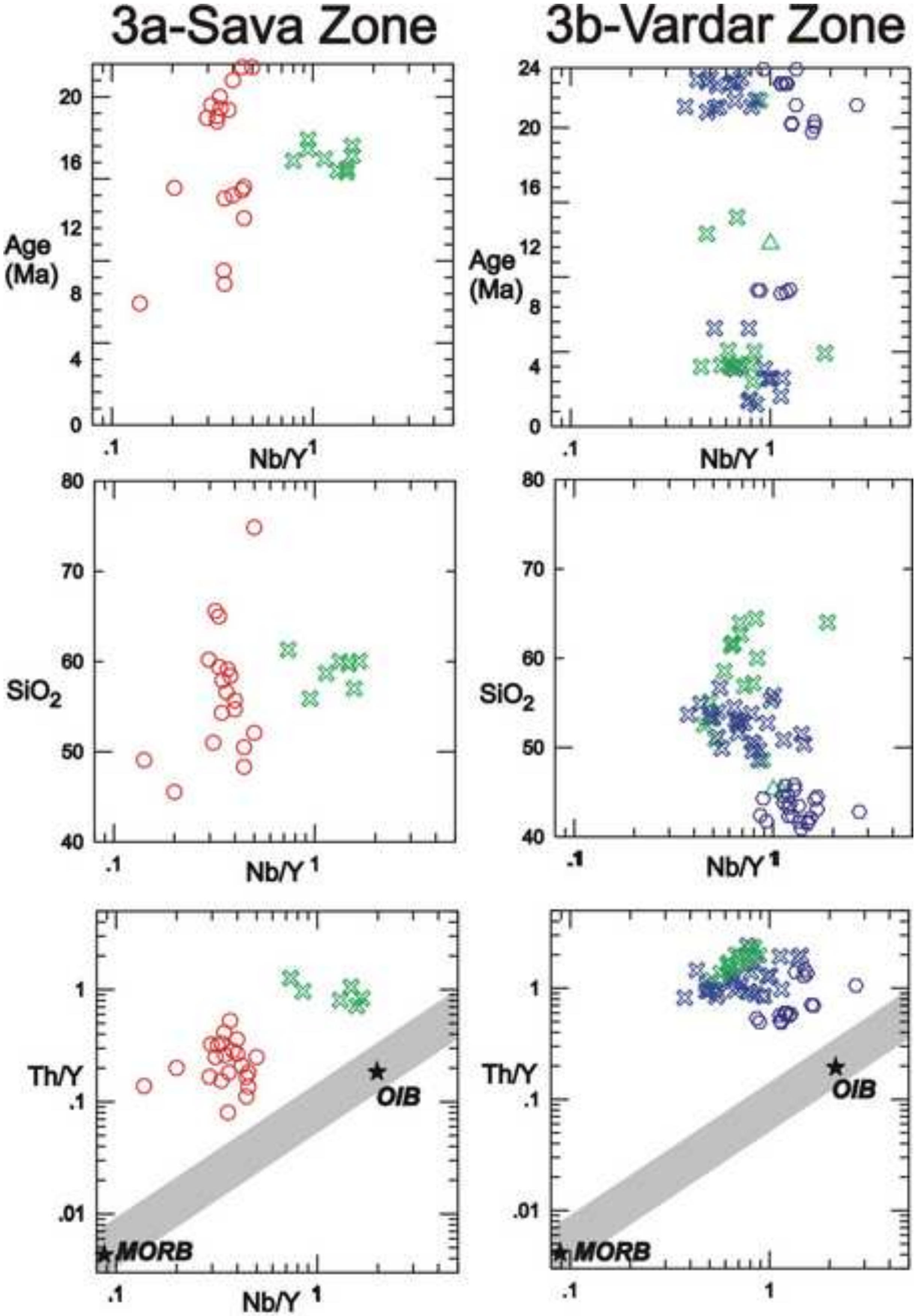

Fig. 8b

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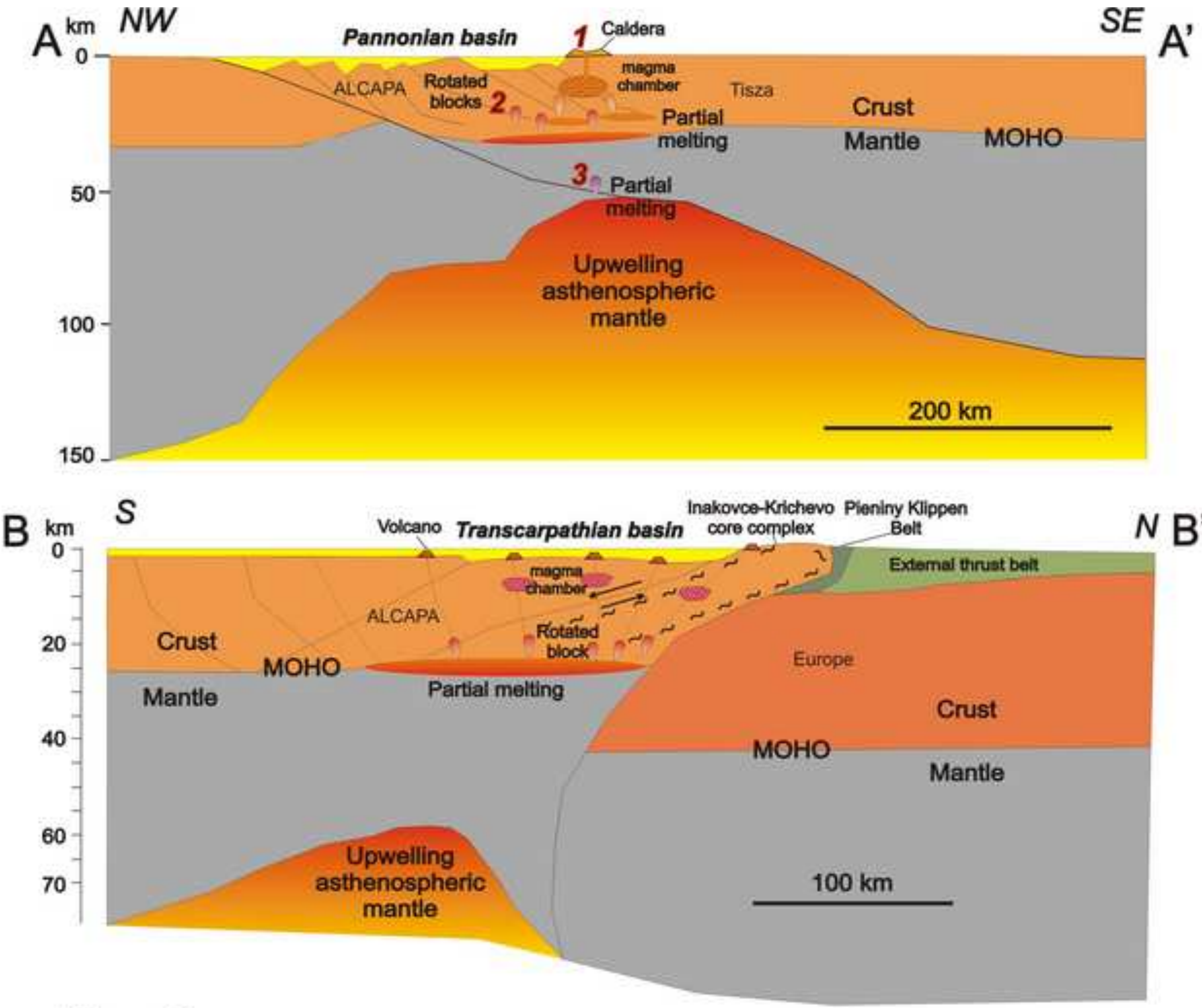

Fig. 9

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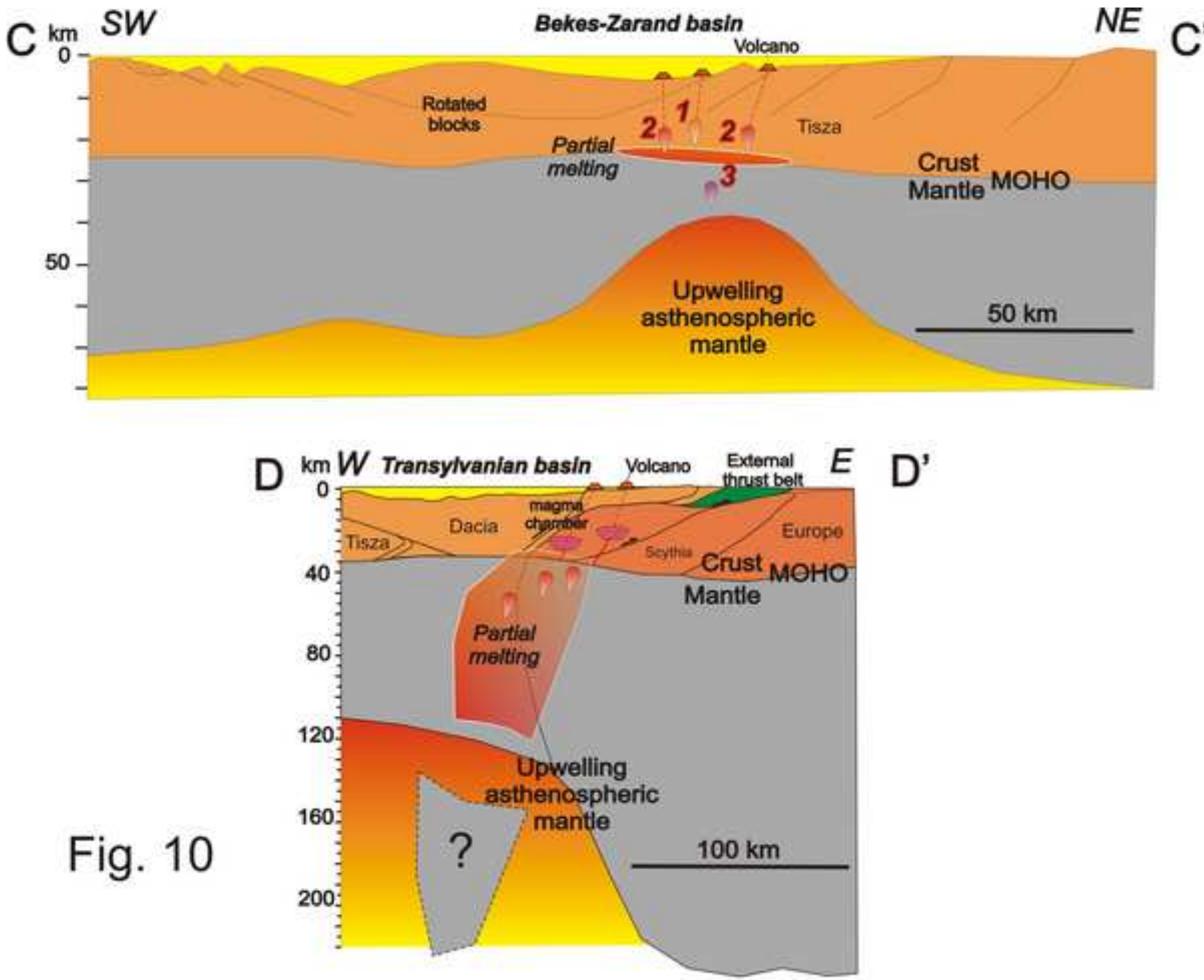

Fig. 10