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The circum-Mediterranean anorogenic Cenozoic igneous province

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Abstract

During the Cenozoic widespread *anorogenic* magmatism, unrelated to recent supra-subduction zone modification of its mantle source, developed within the Mediterranean and surrounding regions; this is referred to collectively as the CiMACI (Circum-Mediterranean Anorogenic Cenozoic Igneous) province. On the basis of a comprehensive review of published and new major and trace element and Sr–Nd–Pb isotopic data (more than 7800 samples) for the magmatic rocks, a common sub-lithospheric mantle source component is identified for most of the region. This has geochemical affinities to the source of HIMU oceanic island basalts and to the European Asthenospheric Reservoir (EAR) and the Low Velocity Component (LVC) of previous workers; we refer to this as the Common Mantle Reservoir (CMR).

Global and local seismic tomography studies of the mantle beneath the CiMACI province have revealed a range of P- and S-wave velocity anomalies, some of which have been related to the presence of mantle plumes. Detailed local tomography experiments in the Massif Central of France and the Eifel region of central Germany suggest that, locally, there are diapiric upwellings rooted within the upper mantle which induce adiabatic decompression melting and magma generation. These velocity anomalies can be interpreted as evidence of mantle temperatures up to 150 °C hotter than the ambient mantle. However, the seismic attenuation could also be attributable to the presence of fluid or partial melt and significant thermal anomalies are not necessary to explain the petrogenesis of the magmas.

A model is proposed in which the geochemical and isotopic characteristics of the sub-lithospheric mantle beneath the CiMACI province reflect the introduction of recycled crustal components (derived from both oceanic and continental lithosphere) into the ambient depleted upper mantle. This sub-lithospheric mantle is subsequently partially melted in a variety of geodynamic settings related to lithospheric extension, continental collision and orogenic collapse, and contemporaneous subduction, slab roll-back and slab-window formation.

On the basis of this in-depth geochemical and petrological study of the CiMACI province, we consider that there is no need to invoke the involvement of anomalously hot mantle (i.e., the presence of a single or multiple deep mantle plumes) in the petrogenesis of the magmas. If, however, we adopt a more permissive definition of "mantle plume", allowing it to encompass passive, diapiric upwellings of the upper mantle, then we can relate the CiMACI province magmatism to multiple upper mantle plumes upwelling at various times during the Cenozoic. To avoid confusion we recommend that such upper mantle plumes are referred to as diapiric instabilities.

Keywords: Cenozoic; magmatism; Europe; mantle plumes; HIMU; basalt

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1. Introduction

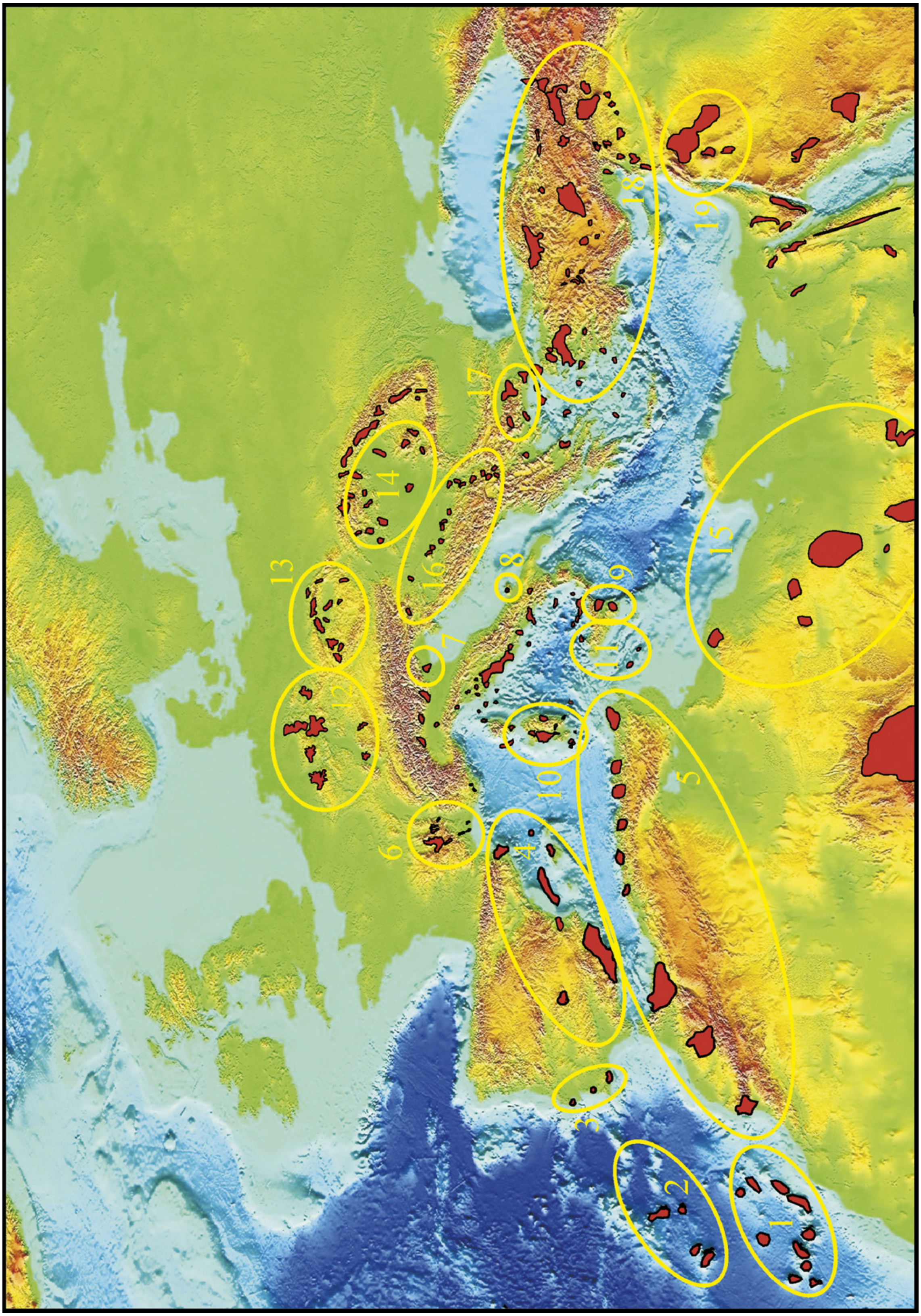
During the last 60–70 Myr the region now occupied by the Mediterranean Sea and the surrounding continental areas has experienced extensive igneous activity (e.g., Wilson and Downes, 1991, 2006; Wilson and Bianchini, 1999; Lustrino, 2000a,c, 2003; Peccerillo, 2005; Fig. 1). The magmatism reflects the response of the upper mantle to the complex geodynamic evolution of this area, characterized by diachronous continent–continent collision, west-, north- and east-directed subduction of oceanic plates, slab roll-back, delamination and detachment of sub-continental lithosphere, development of lithospheric fault systems, passive upwelling of asthenospheric mantle and, locally, ascent of small-scale plume-like instabilities from the Transition Zone at the base of the upper mantle. In some areas mantle partial melting was also accompanied by crustal anatexis (e.g., Zeck et al., 1998; Turner et al., 1999; Fourcade et al., 2001; Zeck and Williams, 2002; Lustrino et al., 2004b).

The Cenozoic magmatic rocks of the circum-Mediterranean region are mostly effusive with minor pyroclastic deposits and high-level intrusions. About 60% are basic to intermediate in terms of their SiO_2 content (SiO_2 ~45–57 wt.%), with the remainder represented by more differentiated and ultrabasic compositions. Very rare occurrences of CaO-rich rocks in Germany (e.g., Schleicher et al., 1990; Riley et al., 1996, 1999), France (e.g., Lenoir et al., 2000; Chazot et al., 2003), Morocco (e.g., Ibhi et al., 2002; Wagner et al., 2003), Libya (Conticelli et al., 1995) and Italy (e.g., Stoppa et al., 2003) have been classified as carbonatites. However, at least for the Italian examples, there remains some doubt about whether these are the crystallization products of true carbonatitic magmas or limestone metasomatised by siliceous melts (e.g., Peccerillo, 1998, 2004).

In many areas (e.g., Spain, France, Algeria, Morocco, Sardinia, Carpathian–Pannonian region, Bulgaria, Serbia, Turkey) there is a temporal sequence of two distinct types of igneous activity, with different geochemical and petrographic characteristics. The first type is mainly of Oligocene to Miocene age, but locally extends from late Cretaceous–Paleogene (e.g., E Serbia: Karamata et al., 1997; Clark and Ulrich, 2004; eastern Rhodopes: Yanev, 2003) to Pliocene (e.g., Harghita Mts., E Carpathians: Pécskay et al., 1995; Mason et al., 1998) and Pleistocene (E Anatolia, Turkey: Pearce et al., 1990; Yilmaz, 1990) times. This typically displays the geochemical characteristics of magmas generated in subduction-related (*orogenic*) tectonic settings, although in some cases the apparent *orogenic* geochemical signature may be partially related to the effects of lithospheric contamination of mantle-derived magmas *en route* to the surface (e.g., Mason et al., 1996; Turner et al., 1999; Coulon et al., 2002; Harangi et al., 2006). The distinct features of this circum-Mediterranean *orogenic* magmatic cycle are calc-alkaline affinities for subalkaline magmas and a potassic to ultrapotassic character for alkaline magmas; enrichment of Large Ion Lithophile (LIL) over High Field Strength (HFS) elements is nearly ubiquitous. The most abundant subalkaline lithologies are intermediate in composition (basaltic andesites and andesites), although basic (picritic basalts) and more differentiated rock types (dacites and rhyolites) have been found. The alkaline igneous rocks are critically SiO_2-saturated (e.g., shoshonites), SiO_2-undersaturated (e.g., tephrites to tephri-phonolites) to strongly SiO_2-undersaturated (e.g., leucitites and hauynophyres). Rare peralkaline rocks (e.g., comendites; Morra et al., 1994; Lustrino et al., 2004b; Mann et al., 2006) and some extreme compositions (e.g., lamproites, lamprophyres) (e.g., Conticelli, 1998; Turner et al., 1999; Altherr et al., 2004; Cvetkovic et al., 2004; Prelevic et al., 2005; Çoban and Flower, 2006; Peccerillo and Martinotti, 2006) have also been reported.

The second type of igneous activity follows immediately after the *orogenic* cycle, or after a hiatus of ~5–15 Myr. In rare cases an overlap between the two types of magmatic activity has been recorded (e.g., E Carpathians: Pécskay et al., 1995; Morocco: Maury et al., 2000; Duggen et al., 2005; E Rhodopes: Marchev et al., 1998a,b). This activity is mainly Pliocene to Recent in age, but locally rare Paleocene–Eocene (E Serbia: Jovanovic et al., 2001; Cvetkovic et al., 2004), Oligocene (E Rhodopes: Marchev et al., 1998b) and

Fig. 1. Digital topography of circum-Mediterranean area from NOAA (http://www.ngdc.noaa.gov/mgg/image/2minrelief.html) showing the locations of Cenozoic igneous provinces (with both *anorogenic* and *orogenic* geochemical signatures) in red. Yellow numbers refer to the *anorogenic* provinces discussed in this paper: 1) = Canary Islands; 2) = Madeira Islands; 3) = Portugal (Serra de Monchique, Sines, Sintra, Gorringe Bank); 4) = Spain (Betic Cordillera, Valencia Gulf, Calatrava and Olot–Garrotxa areas); 5) = Maghrebian Africa (Morocco, Algeria and Tunisia); 6) = France (Massif Central, Aix-en-Provence, Toulon and Cap d'Agde); 7) = Veneto Province (Italy); 8) Punta delle Pietre Nere (Italy); 9) Mt. Etna–Hyblean Plateau (Italy); 10) Sardinia (Italy); 11) = Ustica (Italy) and Sicily Channel islands (Pantelleria and Linosa; Italy); 12) = Germany (Eifel, Siebengebirge, Westerwald, Vogelsberg, Hessian Deptression, Rhön, Heldburg, Upper Palatinate, Kaiserstuhl, Bad Urach and Hegau); 13) = Bohemian Massif (Czech Republic, Poland and eastern Germany); 14) = Pannonian Basin (Hungary, Slovakia, Austria and Romania); 15) Libya; 16) = Serbia (former Yugoslavia); 17) = Bulgaria–Macedonia; 18) = Turkey; 19) Mashrek (Jordan, Israel, Syria, Lebanon and Palestine).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Miocene (Morocco: Duggen et al., 2005; Pannonian Basin: Balogh et al., 1994; Harangi, 2001a; E Thrace, Turkey: Yilmaz and Polat, 1998; W and E Anatolia, Turkey: Yilmaz, 1990; Aldanmaz et al., 2000; Agostini et al., 2007) examples also occur. Rare subalkaline products have tholeiitic affinities, whilst the dominant alkaline rocks are mostly sodic with $Na_2O/K_2O \sim 1$, although occasionally potassic rock types have been observed (e.g., Calatrava district, central Spain, Cebrià and Lopez-Ruiz, 1995; Eifel district, Germany, Mertes and Schmincke, 1985). The most abundant subalkaline lithologies are basic–intermediate in composition (tholeiitic basalts and basaltic andesites) with rarer more diferentiated rock types (mostly rhyolites). The alkaline rocks are essentially ultrabasic to intermediate in terms of SiO_2 content, and range from transitional (olivine- and hypersthene-normative) to mildly (relatively low normative nepheline contents) and strongly alkaline (high normative nepheline and, in some cases, also normative leucite) types (alkali basalts, basanites, tephrites, hawaiites, mugearites, benmoreites). We refer to this type of magmatism subsequently as *anorogenic*.

In rare cases the *orogenic* style of igneous activity postdates the *anorogenic* cycle (e.g., the Eocene calc-alkaline plutonic and subvolcanic rocks of the Periadriatic magmatic province in the eastern Alps were intruded after the *anorogenic* magmatism of the Paleocene–Eocene Veneto volcanic province; Macera et al., 2003, 2004). In a few regions an *anorogenic* volcanic cycle is sandwiched in between two periods of igneous activity with *orogenic* affinities (e.g., E. Serbia: Cvetkovic et al., 2004 and references therein). In other regions, *orogenic* (calc-alkaline, high-K calc-alkaline and shoshonitic) igneous activity was not followed by the *anorogenic* type (sodic alkaline and/or tholeiitic) but by more strongly potassic magmatism. Examples include: calc-alkaline, shoshonitic and ultrapotassic Oligocene volcanic rocks from the NW Alps (e.g., Venturelli et al., 1984; Callegari et al., 2004), the Miocene–Quaternary Tuscan and Roman Provinces (e.g., Conticelli et al., 2002, 2004; Poli, 2004; Peccerillo, 2005), Pliocene to recent volcanic products of the Aeolian Islands in Italy (e.g., Francalanci et al., 2004; Peccerillo, 2005), and the Miocene–Pliocene Hellenic arc in the Aegean Sea (e.g., Altherr and Siebel, 2002; Pe-Piper et al., 2002). Elsewhere, for example in the French Massif Central (Wilson et al., 1995a; Legendre et al., 2001), Sicily and the Sicily Channel in southern Italy (Beccaluva et al., 1998; Civetta et al., 1998; Trua et al., 1998, 2003; Avanzinelli et al., 2004; Rotolo et al., 2006), Libya (Conticelli et al., 1995; Woolley, 2001, Lustrino, 2005b) and the Mashrek (Middle East; Shaw et al., 2003, Bertrand et al., 2003; Lustrino and Sharkov, 2006; Weinstein et al., 2006) only *anorogenic* magmatism has been recorded. In some cases evolution of mantle sources from *orogenic* s.l. (*sensu lato*) to *anorogenic* s.l. has been proposed within the same volcanic district (e.g., Pontine Islands, central-southern Italy; Cadoux et al., 2005; Gourougou and Guilliz volcanic centers, northern Morocco; Duggen et al., 2005; Central Anatolia, Turkey, Wilson et al., 1997), whereas in other cases evolution from *anorogenic* to *orogenic* magmatism at the same volcano has been reported (e.g., Mt. Etna; Schiano et al., 2001).

The aim of this review is to develop a general model for the petrogenesis of the *anorogenic* magmatism within the Circum-Mediterranean Anorogenic Cenozoic Igneous Province (hereafter referred to as the CiMACI province). We have specifically excluded magmatism which is clearly linked to active or very recent subduction (i.e. the *orogenic* type). Whilst it could be argued that both the *orogenic* and *anorogenic* magmatism should be discussed in parallel, to do so would have made this review impossibly long. Additionally a systematic overview of the orogenic magmatism of Europe has recently been completed by Harangi et al. (2006) and we consider that there is no need to duplicate this effort.

We have compiled and synthesised all the available geochemical, geological, geophysical and petrological data from the literature, together with our own unpublished data, for the CiMACI province. The complete database of bulk-rock geochemical and Sr–Nd–Pb isotope analyses (~7800 samples) is available on request from the first author or may be downloaded from the *Earth Science Reviews* data repository (doi:10.1016/j.earscirev.2006.09.002). We believe that this will provide a valuable resource for researchers interested in intra-plate magmatism.

Our analysis of these data has allowed the identification of a dominant, sub-lithospheric magma source region beneath this extensive magmatic province which we refer to as the Common Mantle Reservoir (CMR); this shares many similarities with previously identified mantle source components including the European Asthenospheric Reservoir (EAR; Cebrià and Wilson, 1995), the Low Velocity Component (LVC; Hoernle et al., 1995), Prevalent Mantle (PreMa, Wörner et al., 1986), Focus Zone (FoZo; Hart et al., 1992) and Component A (Wilson and Downes, 1991).

2. What do the terms "anorogenic" and "orogenic" mean in the context of magma generation processes?

From a tectonic point of view, the settings of many of the circum-Mediterranean Cenozoic igneous districts (Spain, southern France, Maghrebian Africa, Italy, Pannonian Basin, Serbia, Bulgaria, Turkey) are *orogenic* (i.e., associated with oceanic lithosphere subduction or continent–continent collision; e.g., Carminati and Doglioni, 2004). Since the early Cretaceous opening of the South Atlantic Ocean, Africa started to move toward the Eurasian block resulting in subduction of the Tethys Ocean. Africa collided with Eurasia diachronously because of the existence of several microcontinent blocks in between the two large plates. This resulted in a tectonic setting characterized by contemporaneous continent–continent collision, oceanic subduction beneath continental lithosphere, opening of back-arc basins and transtensional fault system development (e.g., Doglioni et al., 1999; Rosenbaum et al., 2002; Carminati and Doglioni, 2004). Remnants of the former Tethys Ocean are still present in the central Mediterranean Sea and are now being subducted beneath the Calabrian–Peloritani Arc system (Ionian Sea; central Mediterranean) and the Aegean subduction system. The age of the oceanic crust in the western part of the Mediterranean is younger (<30 Ma) than that in the central and eastern sectors, where the old Tethyan oceanic lithosphere is still present.

The spectrum of igneous rocks emplaced since the late Cretaceous in the circum-Mediterranean region is dynamically linked with the Alpine Orogeny, and, therefore, at one level, we might conceptually consider them all to be *orogenic* or at least "Alpine Orogeny-related". The only exception is the igneous activity of the Atlantic oceanic islands of the Canary Islands and Madeira which are considered part of the CiMACI province for the purpose of this review. Additionally, as noted above, the apparent *orogenic* geochemical signatures of a number of circum-Mediterranean Cenozoic igneous rocks may actually reflect the effects of shallow-level crustal contamination of mantle-derived magmas (i.e., Assimilation–Fractional Crystallization (AFC)-like processes; e.g., Mason et al., 1996; Di Battistini et al., 1998) or generation from lithospheric mantle sources which had experienced supra-subduction zone metasomatism in the distant geological past (e.g. Wilson and Downes, 1991; Tamburelli et al., 2000; Wilson and Patterson, 2001; Peccerillo and Lustrino, 2005).

Consequently, before we can start to draw any conclusions about the petrogenesis of the CiMACI magmatic rocks we must establish some ground rules for distinguishing an *anorogenic* from an *orogenic* geochemical signature. Superficially, this may appear to be a difficult task, not because of the absence of geochemical tools, but rather because of the large number of different geochemical/petrological discriminants which have been proposed in the literature. For example, Wilson and Bianchini (1999) considered that *orogenic* and *anorogenic* magmas could be effectively discriminated using combinations of trace element ratios (e.g. Th/Yb *vs.* Ta/Yb and Th/Zr *vs.* Nb/Zr). Unfortunately, high quality bulk-rock trace element data for these elements are not always available for the magmatic rocks of the CiMACI province. Additionally, in many cases, there is actually a continuous range in chemical composition between *anorogenic* and *orogenic* igneous rock types.

We stress that use of the term *orogenic* does not necessarily imply a recent subduction-related geodynamic setting. Indeed an orogenic geochemical signature may, in some cases, reflect nothing more than the effects of crustal contamination of mantle-derived magmas or derivation of the magmas from lithospheric mantle sources metasomatically enriched by subduction fluids carrying a sediment signature, sometime in the geological past. Similarly, our use of the term *anorogenic* does not necessarily imply a geodynamic setting far away from contemporaneous subduction zones, but rather that the mantle sources of the magmas were not modified by crustal recycling at these subduction zones.

As noted above, the transition from igneous activity with *orogenic* geochemical characteristics to that with *anorogenic* characteristics is relatively common throughout the entire circum-Mediterranean area (e.g., Lustrino, 2003; Beccaluva et al., 2007b). To understand the reasons for, and the causes of, this shift, and the much rarer transition from *anorogenic* to *orogenic* magmatic activity in individual volcanic regions within the CiMACI province is beyond the scope of this study. In some areas the change in chemical composition from *orogenic* to *anorogenic* can be related to "slab detachment" or the development of "slab windows" (e.g., Coulon et al., 2002; Duggen et al., 2005; Agostini et al., 2007), whereas in other cases the involvement of different mantle sources (asthenospheric mantle slightly modified by subduction-related metasomatism and lithospheric mantle strongly modified during ancient times) has been proposed (e.g., Lustrino et al., 2004b, in press).

In general, only the most primitive mafic magmatic rocks (basalts s.l.) can provide information about the nature of their mantle source; consequently, we focus on the geochemical characteristics of such rock types in subsequent sections. To determine the *anorogenic vs. orogenic* affinity of a particular suite, following Wilson and Bianchini (1999), we stress the importance of basing the classification on comprehensive major and trace element and Sr–Nd–Pb isotope data sets.

Mafic magmatic rocks which we classify as *anorogenic* in this review typically have the following characteristics:

(1) They are geochemically and isotopically similar to oceanic island basalts (OIB), despite their relatively large range of variation in chemical composition;
(2) Compared to *orogenic* magma series, they typically have Na_2O/K_2O weight ratios ≥ 1;
(3) On primitive mantle-normalised trace element variation diagrams, basalts with an *anorogenic* affinity have a peak at Nb–Ta and a trough at K, whereas those of *orogenic* affinity have a distinct Nb–Ta trough, and enrichment in the fluid mobile elements Th, Ba, Sr, Pb and K. The negative potassium anomaly characteristic of the *anorogenic* basalts has been suggested to reflect the presence of a residual K-bearing phase in their mantle source, such as amphibole or phlogopite (e.g., Wilson and Downes, 1991); however, it is also possible that the source itself is depleted in K and Large Ion Lithophile Elements (LILE). The trace element patterns of the more differentiated magmas are often more complex to interpret because of the combined effects of crustal assimilation and fractional crystallisation; therefore, for tectonic discrimination purposes, only mafic magmas with >6–7 wt.% MgO are considered. Even for the more primitive magmas, interpretation of the trace element signature can be ambiguous. For example, negative Nb–Ta and positive Pb anomalies can indicate either shallow-level crustal contamination, the introduction of subducted sediment into the mantle source of the magmas or fluid-induced enrichment of the elements adjacent to Nb–Ta in the trace element pattern;
(4) Primitive *anorogenic* mafic alkaline magmas have Nd–Sr isotope compositions similar to those of the European Asthenospheric Reservoir (EAR; Cebrià and Wilson, 1995; Granet et al., 1995) and the Low Velocity Component (LVC; Hoernle et al., 1995). The subalkaline (tholeiitic) basalts typically plot between the EAR and the depleted mantle source of mid-ocean ridge basalts (DM). Pb isotope compositions commonly plot along the Northern Hemisphere Reference Line (NHRL) between HIMU (high-μ) OIB and Depleted Mantle (DM) [see Hofmann (1997) and Stracke et al. (2005) for the definition of these components].

In the following discussion we have subdivided the CiMACI province into western, central and eastern sub-provinces. Within each we discuss, systematically, the age relationships, geochemical and Sr–Nd–Pb isotope characteristics of the Cenozoic *anorogenic* magmatism and its relationship (if any) to associated *orogenic* activity. The data are presented in a standard format (Figs. 2–12) to facilitate comparison from region to region. Where there are preferred petrogenetic models for individual volcanic provinces, these are discussed briefly. The complete data set is then evaluated holistically in the context of a number of different petrogenetic models.

3. Western CiMACI province

The present western Mediterranean Sea started to form some 30 Myr ago, in a general compressive tectonic regime related to Eurasia–Africa convergence (e.g., Doglioni et al., 1999). Convergence was accompanied by volcanic and minor plutonic activity in Spain, France, Italy, Algeria and Morocco (e.g., Wilson and Bianchini, 1999; Lustrino, 2000a; Savelli, 2002; Duggen et al., 2005). The E–SE coast of Spain and the Provençal coast of France were in the hinterland of a subduction system which had been active since the Oligocene, along which oceanic lithosphere of possible Cretaceous age (Tethyan or Mesogean) was subducted beneath the southern European continental margin as a consequence of the relative movements between Africa and Europe (e.g., Catalano et al., 2001). During the evolution of this subduction system, magmatic activity of *orogenic* geochemical affinity occurred along the E–SE Spanish and French coasts. After a period of time ranging from few kyr to several Myr, *anorogenic* igneous activity developed, but spanning a much wider area. This activity has a completely different geochemical signature, reflecting the involvement of different mantle sources. However, from a geodynamic and tectonic point of view, the magmas emplaced during this second stage are still clearly related to Africa–Eurasia convergence and the Alpine Orogeny and, therefore, as noted above, should not strictly be called *intra-plate*, as is common practice in the literature. The same considerations can applied to the Cenozoic igneous activity of Sardinia, Morocco and Algeria in the western Mediterranean Sea.

In a general sense, the western-central Mediterranean area has been characterized by roughly continuous westward-directed subduction of oceanic lithosphere and contemporaneous eastwards slab roll-back throughout the Tertiary (e.g. Wilson and Bianchini, 1999). In detail, the subduction polarity varies from north-west (in the central Mediterranean) to south-west (in peninsular Italy) to north (along the African margin, e.g., Kabiles) to east (in the westernmost Mediterranea, e.g., Alboran Sea). Two main continental micro-plates (Balearic Islands, Sardinia–Corsica) plus other minor lithospheric fragments (e.g., Calabria, Peloritani Mts. in Sicily, Kabiles in northwest Africa and the Alboran block) have been isolated by the formation of back-arc basins (Ligurian–Provençal and Tyrrhenian Basins; Carminati et al., 1998; Gueguen et al., 1998; Doglioni et al., 1999). The only active subduction at the present day appears to be along the Calabrian–Peloritani Arc, although Gutscher et al. (2002) also proposed active subduction near Gibraltar in the westernmost Mediterranean Sea.

Tertiary–Recent intra-plate volcanic activity also occurs to the west of the Straits of Gibraltar in the Canary Islands and Madeira, within the Atlantic Ocean. Whilst these oceanic volcanic provinces are not strictly part of CiMACI, we begin our detailed discussion of the Western CiMACI sub-province with a brief overview of their geochemical characteristics. Since both the western Canary Islands and Madeira are located on oceanic lithosphere, these data provide an important baseline for characterising magmas derived from sub-lithospheric mantle sources which have not been subsequently modified during their transit through continental lithosphere.

3.1. Canary islands

The Canary Islands are an E–W trending chain of oceanic islands situated close to the continental margin of NW Africa (Fig. 1). There are seven main islands which show an approximate east (Fuerteventura ~24 Ma) to west (Hierro ~1.1 Ma) age progression in subaerial volcanism. Together with the Selvagen Islands and seamounts to the NE (Lars, Anika and Dacia seamounts), the Canary Islands form an 800 km long by 450 km wide volcanic belt which decreases in age from the NE (~68 Ma, Lars Seamount) to the SW, interpreted as a hotspot track (Geldmacher et al., 2001, 2005). The islands are constructed on oceanic lithosphere which formed during the opening of the Central Atlantic Ocean at ~180–200 Ma (Hoernle, 1998). They are considered here to represent the westernmost volcanic activity of the CiMACI province. Their geochemical characterisation is important in developing models for CiMACI province magmatism as this is one of the few regions in which we can probably discount the involvement of ancient enriched lithospheric mantle in the petrogenesis of the magmas. The westernmost islands (La Palma and Hierro) clearly rest on oceanic crust, while the eastern islands (Lanzarote and Fuerteventura) are located on thicker lithosphere in the Jurassic Magnetic Quiet Zone (Roest et al., 1992). The ocean–continent transition zone appears to be located to the east of the Canary Islands on the basis of lithospheric mantle xenolith studies (Neumann et al., 2004).

Magmatic activity is dominated by basaltic lavas; more differentiated compositions (tephrites, trachytes, phonolites, rhyodacites, rhyolites, pantellerites and comendites) and ultrabasic rock types (melilitites, nephelinites, basanites) also occur (Fig. 2). The mafic magmatism ranges from hypersthene-normative tholeiitic basalts to strongly silica-undersaturated nephelinites, but is dominated by TiO_2-rich alkali basalts (Schmincke, 1982). In Fuerteventura the oldest volcanic complex is intruded by a series of rock types that include nepheline syenites and carbonatites (Le Bas et al., 1986; Balogh et al., 1999; Hoernle et al., 2002). The most significant eruptions of tholeiitic basalt have occurred only on the eastern islands (Lanzarote) during the Plio-Quaternary (Lundstrom et al., 2003).

Magmatism in individual Canary Island volcanoes is typically divided into an older, shield-building stage, which, after a period of erosion and quiescence, is followed by a second phase of activity. There is a very irregular westwards decrease in the age of the oldest exposed lavas in the shield-building stage from ~20 Ma in Lanzarote and Fuerteventura to 1.1 Ma in La Palma and Hierro (Schmincke, 1982; Balogh et al., 1999; Anguita and Hernán, 2000). Mantle xenoliths are common in the alkali basaltic lavas and dykes of the younger period of volcanic activity on each island (e.g., Neumann et al., 2004).

The most primitive mafic volcanic rocks are characterized by a restricted range of Sr–Nd isotope compositions ($^{87}Sr/^{86}Sr$ = 0.7028–0.7038; $^{143}Nd/^{144}Nd$ = 0.51274–0.51304), plotting mainly within the EAR/LVC field (Fig. 2). Pb isotopic compositions define a coherent linear trend between HIMU and DM mantle components, parallel to the Northern Hemisphere Reference Line (Hart, 1984) ($^{206}Pb/^{204}Pb$ = 18.76–20.21; $^{208}Pb/^{204}Pb$ = 38.69–39.92; Δ7/4 = −11.3 to +8.0; Δ8/4 = −43.1 to + 55.8; Fig. 2). The $^{206}Pb/^{204}Pb$ ratios include some of the most radiogenic in the entire CiMACI province together with Maghrebian African volcanic rocks (see below). The trace element characteristics of the most primitive mafic lavas are similar to those of HIMU-OIB (Fig. 2).

Sr–Nd–Pb–O–Os isotope studies have demonstrated the involvement of both lithospheric and asthenospheric mantle source components in the petrogenesis of the Canary Island magmas (e.g.,

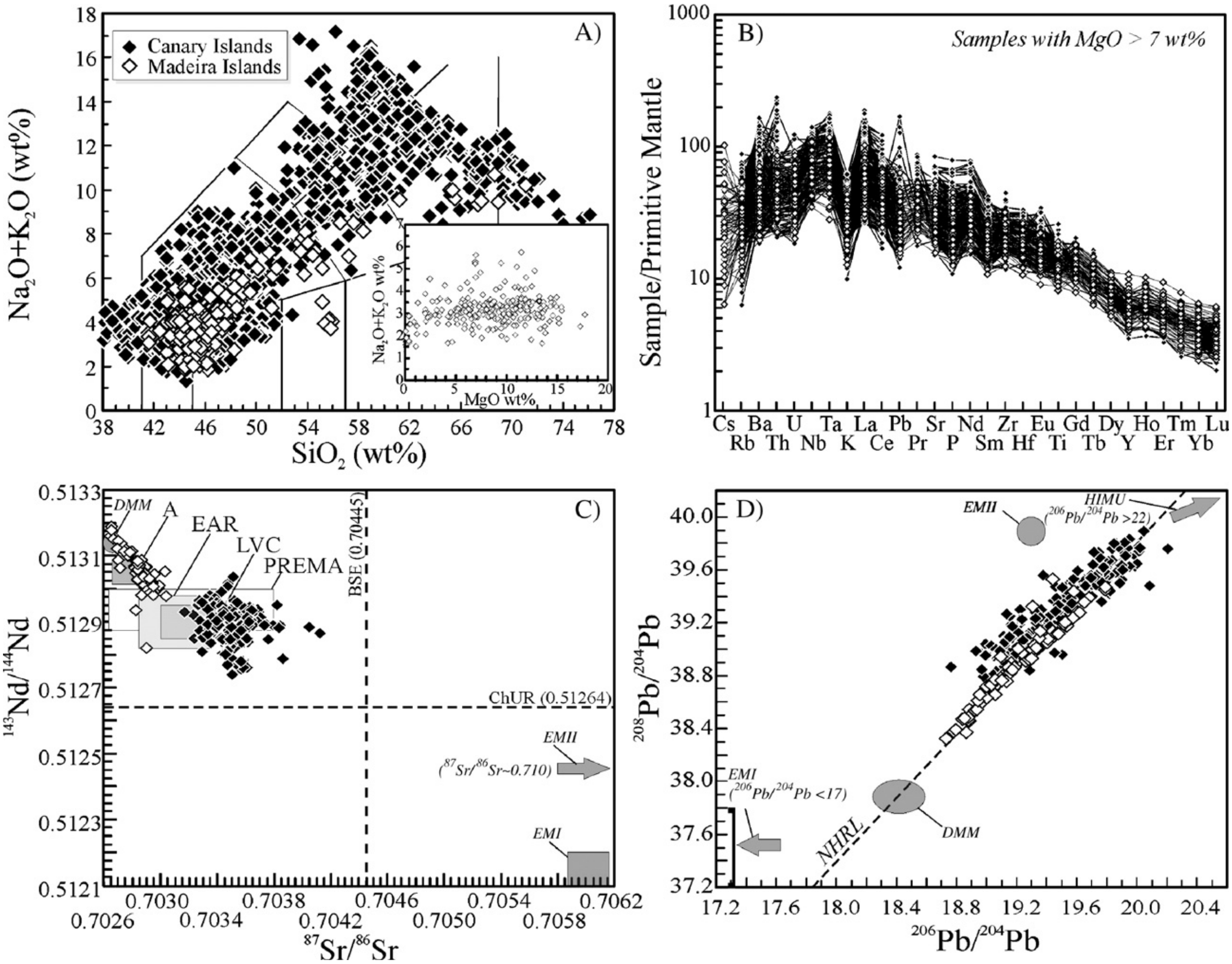


Fig. 2. A) Total alkali *vs.* silica diagram (TAS; le Maitre, 2002); B) primitive mantle (Sun and McDonough, 1989) normalised incompatible element diagram; C) $^{143}Nd/^{144}Nd$ *vs.* $^{87}Sr/^{86}Sr$ diagram; D) $^{208}Pb/^{204}Pb$ *vs.* $^{206}Pb/^{204}Pb$ diagram for igneous rocks from the Canary Islands and seamounts (black diamonds) and Madeira Islands and seamounts (white diamonds). Data for the Canary Islands and seamounts are from the GEOROC database (http://georoc.mpch-mainz.gwdg.de/georoc/) and Geldmacher et al. (2005). References for Madeira islands and seamounts: Aires-Barros et al., 1980; Halliday et al., 1992; Mata et al., 1998; Geldmacher and Hoernle, 2000; Geldmacher et al., 2005; Schwarz et al., 2005. Abbreviations: DMM = Depleted MORB Mantle (Kramers and Tolstkhin, 1997; Workman and Hart, 2005); EMI = Enriched Mantle type 1 (Lustrino and Dallai, 2003); EMII = Enriched Mantle type 2 (Workman et al., 2004); HIMU = high-μ (where $\mu = ^{238}U/^{204}Pb$; Hofmann, 2004); A = Component A (Wilson and Downes, 1991); EAR = European Asthenospheric Reservoir (Granet et al., 1995; Wilson and Bianchini, 1999); LVC = Low Velocity Component (Hoernle et al., 1995); PREMA = Prevalent Mantle (Wörner et al., 1986).

Hoernle et al., 1991; Thirlwall et al., 1997; Widom et al., 1999). During the shield-building stage, the volcanic rocks have HIMU-like Sr–Nd–Pb isotopic characteristics (e.g., Simonsen et al., 2000); these have been interpreted in terms of derivation of the parental magmas from a mantle plume dominated by <2 Ga recycled oceanic crust by Thirlwall et al. (1997). Rocks of the post-erosional cycle on the eastern islands are characterized by less radiogenic Sr and Pb and more radiogenic Nd isotopic compositions, interpreted to reflect the involvement of a depleted mantle (DM) source component in the plume (Hoernle et al., 1991). During both the shield-building and post-erosional cycles, nephelinites and basanites dominate the early stages of activity. More silica-saturated basalts occur at the peak of the cycle, and alkali basalts, nephelinites and more evolved alkaline lavas dominate the waning stages of activity (Hoernle and Schmincke, 1993a,b). The more silica-saturated mafic and evolved volcanic rocks have more radiogenic Sr and less radiogenic Nd and Pb (i.e., more EM-like) isotopic compositions than the associated basanites, interpreted to reflect interaction of asthenosphere-derived melts with the overlying lithosphere (e.g., Widom et al., 1999). Lundstrom et al. (2003) have used U-series disequilibria studies to demonstrate that tholeiitic basalts in the eastern Canary

Islands could be generated by mixing between plume-derived basanites and a high-silica melt component derived from mantle lithosphere metasomatised during the earlier volcanic stages.

Magmatic activity at individual volcanic islands and seamounts is long-lived and continued for ~27 Myr in the Selvagen Islands (Geldmacher et al., 2001) and Fuerteventura (Le Bas et al., 1986) and ~40 Myr at Dacia seamount (Geldmacher et al., 2005). These exceptionally long intervals of volcanic activity at a single volcanic centre, coupled with 1) the presence of long periods of magmatic quiescence in almost all the islands (up to 7 Myr gaps); 2) the long volcanic record, probably up to 80 Myr on the oldest islands; 3) the irregular westward progression of the volcanic activity, 4) the presence of recent magmatic activity in almost all of the Canary Islands, 5) the absence of any geoid anomaly under the Canaries (e.g., Jung and Rabinowitz, 1986; Watts, 1994) and 6) the very high flexural rigidity for the Canary Islands' lithosphere (Filmer and McNutt, 1989), are difficult to explain in terms of a classical mantle plume model (Anguita and Hernán, 2000). Nevertheless, Geldmacher et al. (2005) have proposed the presence of an upper mantle plume, rooted in the mantle Transition Zone (410–660 km depth), to explain the formation of the ~800 km long Canary Island volcanic chain, in combination with edge-driven convection processes close to the continental margin of Africa. Anguita and Hernán (2000) proposed a model involving a weak thermal anomaly (upper mantle residue from a fossil Triassic mantle plume, anchored at shallow sub-lithospheric depths), coupled with a propagating fracture system (that geographically and magmatically connects the Canary Islands volcanism with that of the Atlas Chain in Morocco; see below) and tectonic forces which were alternatively distensive (allowing formation and uprising of magma) and compressive (responsible for the strong uplift recorded in almost all the islands). The persistence of a Mesozoic thermal anomaly to the present-day is, however, highly unlikely.

3.2. Madeira

The Madeira archipelago is located in the eastern North Atlantic Ocean to the north of the Canary Islands (Fig. 1). It forms the southwestern end of a 700 km long, age-progressive, chain of volcanic islands and seamounts which intersects the Azores–Gibraltar Fracture Zone in the vicinity of the Ormonde seamount, some 200 km to the west of the Serra de Monchique complex in southern Portugal (Geldmacher and Hoernle, 2000). The volcanic islands of the archipelago, are constructed on 140 Ma oceanic crust and range in age from 0–5 Ma old (Madeira and Desertas Islands) to 14 Ma (Porto Santo). Madeira thus appears to represent the present day location of a mantle hotspot which has been active for more than 70 Myr. Geldmacher et al. (2005) estimated an average age progression of ~1.2±0.2 cm/a for the entire history of the chain, consistent with the migration of the African plate over a near stationary hotspot. The 70–72 Ma old Serra de Monchique complex (Section 3.3.1) could be the earliest expression of the hotspot before it was displaced by ~200 km by right lateral motion along the Azores–Gibraltar Fracture Zone.

The subaerial evolution of Madeira has been divided into a shield-building stage from >4.6 to 3.9 Ma and a post-erosional stage <0.7 Ma (Geldmacher and Hoernle, 2000). The most primitive mafic magmas range in composition from transitional tholeiites to basanites; more evolved magmas (hawaiites–trachytes) also occur, consistent with shallow-level magma differentiation processes (Fig. 2). The primitive mantle-normalised trace element patterns of the most primitive mafic magmas (MgO >7 wt.%) are remarkably homogeneous with distinctive depletions in K, Rb, Ba and Pb and enrichment in Nb–Ta; these are the distinctive trace element characteristics of HIMU-OIB. The depletion in K and Pb of the Madeira basalts is a primary feature of the magmas and not an effect of subaerial weathering. Ratios of more to less incompatible elements (e.g. Ba/Ce, La/Yb, Nb/Zr, Zr/Y) appear to increase with decreasing age, reflecting decreasing degrees of melting with decreasing age. Heavy REE abundances are consistent with the presence of residual garnet in the mantle source during partial melting.

Nd–Sr isotope compositions are predominantly DMM-like ($^{87}Sr/^{86}Sr$ = 0.7026–0.7031 = $^{143}Nd/^{144}Nd$ = 0.51298–0.51320), similar to present day Atlantic MORB, and form a well-correlated array extending towards the EAR field (Fig. 2). Pb isotopic compositions ($^{206}Pb/^{204}Pb$ = 18.7–19.8; $^{208}Pb/^{204}Pb$ = 38.3–39.4) define a tight array along the NHRL which could be interpreted in terms of mixing between DMM and HIMU components; however $^{207}Pb/^{204}Pb$ compositions plot below the NHRL (not shown) with $\Delta 7/4$ values as low as −7.2 in those samples with the highest $^{206}Pb/^{204}Pb$ ratios (Geldmacher and Hoernle, 2000) and are inconsistent with such a simple mixing model. $^{87}Sr/^{86}Sr$ decreases and $^{143}Nd/^{144}Nd$ increases with decreasing age from the early shield stage to the post-erosional stage; all Pb isotope ratios become less radiogenic with decreasing age.

A number of different models have been put forward to explain the magmatism. Halliday et al. (1992) proposed that the magmas are derived from a metasomatically enriched mantle source at the base of the oceanic lithosphere; metasomatism was inferred to date from the Mesozoic age of formation of the oceanic crust on which Madeira sits. In contrast Hoernle et al. (1995) argued for magma generation linked to interaction between a HIMU mantle plume and normal MORB-source mantle in the asthenosphere. Mata et al. (1998) and Widom et al. (1999) also favoured mantle plume models. On the basis of an Os isotope study of Holocene basalts from both Madeira and the Canary Islands, Widom et al. (1999) suggested that the plume source contains 25–35% relatively young (~1.2 Ga) recycled oceanic crust which mixes with ambient depleted asthenospheric mantle containing a component of Palaeozoic recycled oceanic crust. Geldmacher and Hoernle (2000) proposed that during the early shield stage this recycled component comprised hydrothermally altered MORB (~880 Ma old according to Geldmacher and Hoernle, 2001), whereas during the post-erosional stage it is derived from recycled lower oceanic crust and mantle lithosphere.

3.3. Iberia

Cenozoic igneous activity in the Iberian Peninsula can be subdivided chronologically and geographically into two main episodes and areas: late Mesozoic–early Cenozoic alkaline igneous activity, mostly concentrated in Portugal, and a Neogene igneous cycle in Spain.

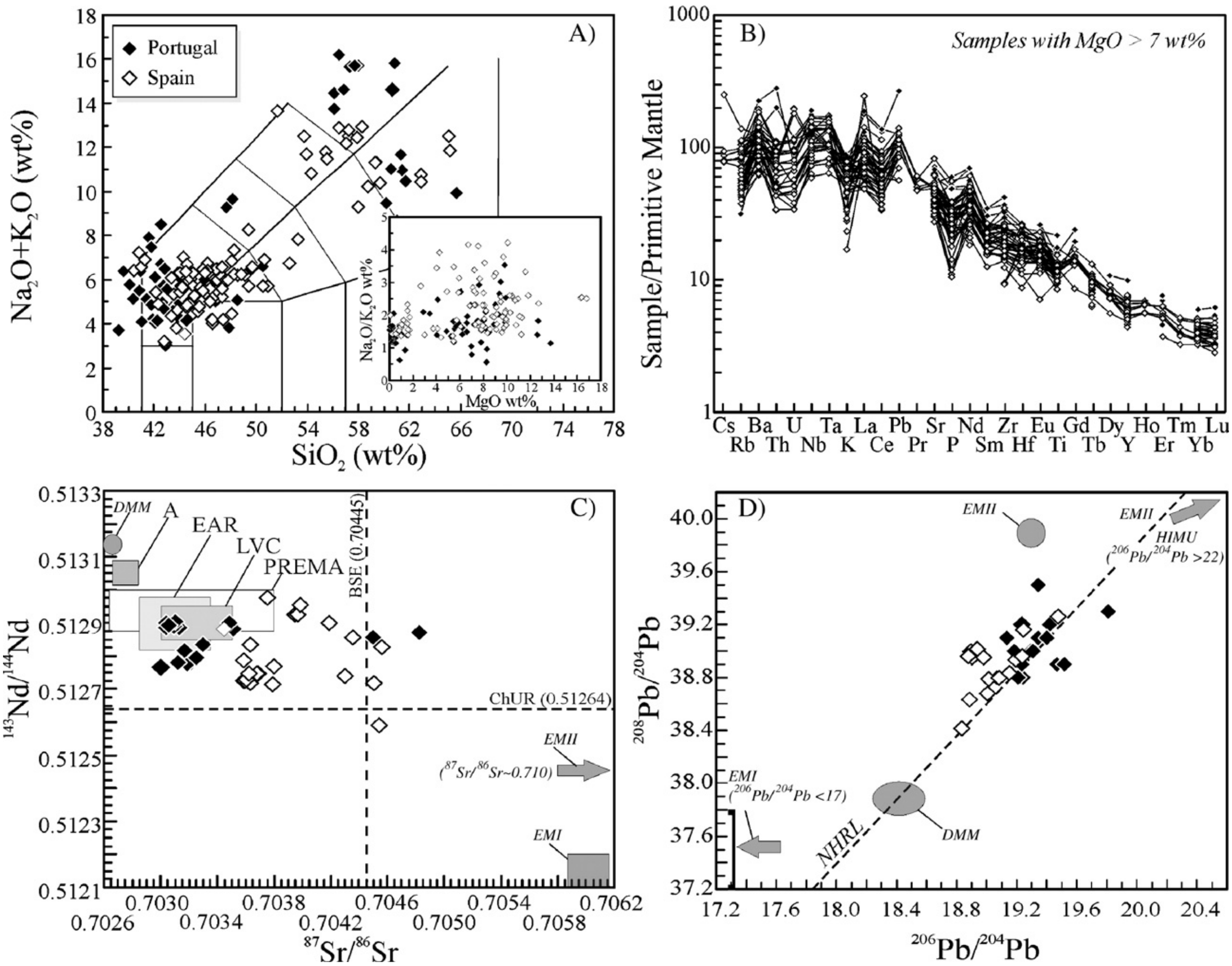


Fig. 3. A) Total alkali *vs.* silica diagram (TAS; le Maitre, 2002); B) primitive mantle (Sun and McDonough, 1989) normalised incompatible element diagram; C) $^{143}Nd/^{144}Nd$ *vs.* $^{87}Sr/^{86}Sr$ diagram; D) $^{208}Pb/^{204}Pb$ *vs.* $^{206}Pb/^{204}Pb$ diagram for igneous rocks from Portugal (black diamonds) and Spain (white diamonds). Data sources: Portugal (Rock, 1982; Bernard-Griffiths et al., 1997; Tavares Martins, 1999); Spain (Alibert et al., 1983; Araña et al., 1983; Ancochea et al., 1984; Lopez Ruiz and Badiola, 1985; Capedri et al., 1989; Aparicio et al., 1991; Martì et al., 1992; Aparicio and Garcia, 1995; Cebrià and Lopez-Ruiz, 1995; Gimeno Torrente, 1995; Diaz et al., 1996; Neumann et al., 1999; Turner et al., 1999; Cebrià et al., 2000; Ancochea and Huertas, 2002; Beccaluva et al., 2004; Acosta et al., 2004; Duggen et al., 2005; this study). Abbreviations as in Fig. 2.

3.3.1. Portugal

The existence of a late Mesozoic–early Cenozoic Iberian alkaline igneous province was originally proposed by Rock (1982) who studied ~80–60 Ma old igneous rocks mostly from mainland Portugal (e.g., Serrra de Monchique, Sines, Sintra, Lisbon) but also from offshore areas (Mt. Ormonde and Gorringe Bank in the N Atlantic, offshore SW Portugal; Cornen, 1982; Féraud et al., 1982, 1986; Bernard-Griffiths et al., 1997; Tavares Martins, 1999). The only locality which has been studied in any detail is the Serra de Monchique complex in S Portugal (~72±2 Ma; Rock, 1978; Bernard-Griffiths et al., 1997). The emplacement of this complex is roughly coeval with the late stages of the Bay of Biscay opening (~80–70 Ma). The most abundant lithologies are nepheline syenites plus rare basic–ultrabasic volcanic and plutonic rocks (alkali gabbros, shonkinites, basanites, ankaramites and alkaline lamprophyres; Fig. 3). The least altered samples are characterized by low $^{87}Sr/^{86}Sr$ (0.70304–0.70325) and relatively high $^{143}Nd/^{144}Nd$ (0.51276–0.51290; Fig. 3). $^{206}Pb/^{204}Pb$ isotopic ratios are in the range 19.07 to 19.81 whereas $\Delta 7/4$ and $\Delta 8/4$ range from −2.8 to +8.2 and from −32.7 to +42.1, respectively (Fig. 3).

The origin of the Serra de Monchique complex has been related to lithospheric extension and North Atlantic passive margin formation during the late Cretaceous, without any evidence of the involvement of a mantle plume (Rock, 1982; Bernard-Griffiths et al., 1997). Recently, however, Geldmacher et al. (2005) have proposed a genetic link between the Serra de Monchique magmatism and that of the Madeira volcanic province, relating both to the activity of a mantle plume anchored in the mantle Transition Zone, whose track is represented by the NE–SW alignment of Madeira, Mt. Ormonde and Serra de Monchique. The absence of significant geochemical differences between the onshore (Serra de Monchique) and offshore igneous complexes (Mt. Ormonde) suggests an asthenospheric mantle source for the parental magmas (Bernard-Griffiths et al., 1997). The slightly less radiogenic Nd isotopic composition of the Serra de Monchique rocks compared to those of Mt. Ormonde, however, has been interpreted as an effect of contamination of asthenospheric mantle partial melts with the continental lithospheric mantle (Bernard-Griffiths et al., 1997).

3.3.2. Spain

During Neogene, igneous activity in Iberia was confined to Spain, concentrated in four sectors: Internal Betics (SE Spain; ~34–2 Ma), Valencia Trough (~24–0.01 Ma), Calatrava Volcanic Province (~9–2 Ma) and Olot–Garrotxa, NE Spain (~10–0.01 Ma). A fifth sector can be identified in the Alboran Basin (westernmost Mediterranean Sea, between the southern coast of Spain and northern Morocco) where upper Miocene (~12.1–6.1 Ma) dacites, rhyolites and granites occur, together with significant volumes of tholeiitic to calc-alkaline basalts, basaltic andesites and andesites (e.g., Alboran Island, Yusuf Ridge, Al Mansour Seamount; Hoernle et al. 1999; Duggen et al., 2003, 2004, 2005; Gill et al. 2004). Some of the mafic lavas in the Alboran Basin may have affinities to boninites (Gill et al., 2004). Duggen et al. (2004) have proposed that the coastal volcanic complexes of Cabo de Gata (S Spain) and Ras Tarf and Trois Furches (N Africa) should be considered part of the Alboran Basin volcanic province.

In SE Spain (Internal Betics) and in the Alboran Basin the igneous rocks have extremely variable geochemical compositions ranging from tholeiitic to calc-alkaline, high-K calc-alkaline and ultrapotassic (shoshonitic to lamproitic; Zeck, 1997; Benito et al., 1999; Turner et al., 1999; Duggen et al., 2003, 2004, 2005; Gill et al., 2004). These belong to the *orogenic* association and are, therefore, not discussed further here. The only volcanism with *anorogenic* geochemical characteristics of SE Spain occurs at Tallante (Dupuy et al., 1986; Capedri et al., 1989; Beccaluva et al., 2004; Duggen et al., 2005). Here, silica-poor, Na-rich volcanic rocks (mostly hawaiites) have been dated at ~2.3–2.9 Ma by Bellon et al. (1983) and Duggen et al. (2005). Mantle xenoliths are commonly associated with these lavas (Dupuy et al., 1986; Capedri et al., 1989; Beccaluva et al., 2004). The most mafic rocks from Tallante show relative enrichments of Nb–Ta and primitive mantle-normalised trace element patterns broadly resembling those of HIMU-OIB (Fig. 3). Some differences from the trace element patterns of end-member HIMU-OIB (e.g., variable U–Th anomalies, absence of Pb troughs and less pronounced Nb–Ta peaks) have been related to interaction with lamproitic components in the sub-continental lithosphere (Duggen et al., 2005). Initial $^{87}Sr/^{86}Sr$ ratios are relatively low (0.7034–0.7046) and initial $^{143}Nd/^{144}Nd$ ratios, with the exception of one sample (0.51259), are high (0.51272–0.51298; Capedri et al., 1989; Turner et al., 1999; Beccaluva et al., 2004; Duggen et al., 2005). Pb isotopic ratios are generally quite radiogenic (e.g., $^{206}Pb/^{204}Pb$ ranges from 18.83 to 19.47; $\Delta 7/4$ from −0.7 to +12.6 and $\Delta 8/4$ from +3.1 to +52.6; Turner et al., 1999; Duggen et al., 2005; Fig. 3).

Offshore eastern Spain (Valencia Trough) magmatic activity is clearly related to extensional tectonics. The western Mediterranean back-arc basins started to open during the late Oligocene (~30 Ma; Séranne, 1999), with the formation of the late Oligocene–early Miocene Alboran, Valencia and Ligurian–Provençal sub-basins, followed in time and space eastward by the formation of the middle to late Miocene Balearic and Algerian Basins, and the late Miocene–Pleistocene formation of the Tyrrhenian Sea (Gueguen et al., 1998). The opening of the Ligurian–Provençal Basin marked the separation of the Sardinia–Corsica micro-plate from Provence as a consequence of counterclockwise rotation about a pole located near the city of Genoa in NW Italy. The splitting of Sardinia and Corsica from Provence was induced by roll-back of a W–NW dipping slab associated with back-arc basin opening (Doglioni et al., 1997; Faccenna et al., 1997; Séranne, 1999) and oceanic crust formation. During the early to middle Miocene (~24–19 Ma), calc-alkaline volcanism developed in the central to eastern offshore areas of the Valencia Trough and on Mallorca Island (Mitjavila et al., 1990), and has been found in drill cores in the Gulf of Valencia (up to 1200 m of volcanic rocks of Miocene age; Lanaja, 1987). As the magmatic products have *orogenic* geochemical characteristics they are not discussed further here.

From Middle Miocene to Holocene times (~10–0.01 Ma) a new volcanic cycle developed in the Valencia Trough and in SE Spain. Its products are mainly represented by relatively undifferentiated alkaline rocks with *anorogenic* geochemical characteristics outcropping both on land (Picasent and Cofrentes) and offshore (Columbretes island; Aparicio et al., 1991; Martì et al., 1992; Ancochea and Huertas, 2002; Muñoz et al., 2005). Olivine–nephelinites, basanites, alkali basalts and hawaiites are the commonest lithologies, followed by phonolites (Fig. 3). The mafic magmas have sodic (Na_2O/K_2O >1) geochemical characteristics, with strongly HREE-depleted trace element patterns. Comprehensive data on the Sr–Nd–Pb isotopic systematics of this volcanic cycle are, unfortunately, lacking.

Mafic undersaturated alkaline lavas of late Miocene–Pleistocene age also occur in central Spain (Calatrava Volcanic Province; Ciudad Real) over an area of about 4000 km^2 (Lopez-Ruiz et al., 1993; Cebrià and Lopez-Ruiz, 1995). This district includes both undersaturated (alkali basalts and nephelinites) and strongly undersaturated sodic to potassic alkaline igneous rocks (melilitites and leucitites; Fig. 3).

The volcanic rocks of NE Spain (Olot–Garrotxa) are emplaced within extensional basins and graben-type structures oriented NE–SW (following the main direction of the Cenozoic European rift system; Ziegler, 1992; Dèzes et al., 2004) and NW–SE (the trend of the onshore continuation of the North Balearic Fault Zone). These include leucite basanites, nepheline basanites and alkali olivine basalts (Cebrià et al., 2000), plus rare trachytes (Martì et al., 1992; Diaz et al., 1996; Fig. 3). Geochemically, the Olot–Garrotxa volcanic rocks are remarkedly homogeneous, with HIMU-OIB-like primitive mantle-normalised trace element patterns (convex patterns with positive peaks at Nb–Ta; Fig. 3); the positive Ba anomalies are, however, more typical of EMI OIB. Sr–Nd isotopic compositions (Fig. 3) indicate derivation from a time-integrated LILE and LREE depleted mantle source ($^{87}Sr/^{86}Sr$ = 0.7036–0.7038; $^{143}Nd/^{144}Nd$ = 0.51271–0.51283). Pb isotopic compositions fall in the range $^{206}Pb/^{204}Pb$ = 18.88–19.47.

Decompression melting of both lithospheric and asthenospheric mantle sources is considered to be the main cause of volcanism in the low volcanicity rift setting of the Gulf of Valencia (Martì et al., 1992). The relatively variable isotopic signature of the Olot–Garrotxa mafic igneous rocks has been related by Cebrià et al. (2000) to at least two distinct mantle sources: an asthenospheric mantle source with a composition resembling EAR and an enriched lithospheric mantle characterized by the presence of a K-bearing phase. Cebrià et al. (2000) proposed a model in which a channelized mantle plume (whose first products were the Jurassic igneous rocks of the Central Atlantic Plume Province; Hames et al., 2003) provided the sub-lithospheric component of the Olot–Garrotxa magmas, while the lithospheric component was enriched by subduction fluids in Permo-Carboniferous times. For the Calatrava volcanic district (central Spain) trace element abundances suggest (as in the Olot–Garrotxa area) mixing of partial melts from the lithospheric mantle with partial melts of an asthenospheric mantle component with affinities to HIMU-OIB source mantle (Cebrià and Lopez-Ruiz, 1995). A temporal sequence of potassic leucitites (emplaced at ~8.7–6.4 Ma) followed by a sodic basaltic suite (~4.6–1.7 Ma) at Calatrava has been explained by a HIMU-like mantle diapir (mantle plume?) triggering lithospheric mantle partial melting (in particular easily fusible, phlogopite-rich, metasomatic veins or streaks, followed by partial melting of the asthenospheric mantle and the production of sodic HIMU-OIB-like magmas (Lopez-Ruiz et al., 1993; Cebrià and Lopez-Ruiz, 1995). Duggen et al. (2005) proposed an origin for the Tallante volcanism from a plume-contaminated, sub-lithospheric, mantle source, whose partial melts were modified by interaction with sub-continental lithospheric mantle melts with lamproitic affinity. Interaction of sub-lithospheric mantle melts with metasomatised lithosphere would have shifted the *anorogenic* geochemical fingerprint of the sub-lithospheric melts towards subduction-related signatures. Upwelling of the plume-contaminated sub-lithospheric mantle source was inferred to be the result of delamination or detachment of the sub-continental lithospheric mantle (Duggen et al., 2005).

3.4. Maghrebian Africa (Morocco, Algeria)

Cenozoic igneous activity in NW Africa is concentrated mostly along the Maghreb margin (~1200 km long, ~50 km wide), in Morocco and Algeria, together with very rare outcrops in Tunisia (Fig. 1). This activity is mostly represented by volcanic rocks characterized by a temporal sequence from *orogenic* compositions (mostly Miocene calc-alkaline to high-K calc-alkaline, potassic and ultrapotassic) to transitional *anorogenic* products (late Miocene–Quaternary in age), mostly with sodic alkaline affinity (e.g., Louni-Hacini et al., 1995; El Bakkali et al., 1998; El Azzouzi et al., 1999; Maury et al., 2000; Coulon et al., 2002; Duggen et al., 2003, 2005). The *anorogenic* igneous rocks are moderately to strongly SiO_2-undersaturated and show incompatible trace element ratios (e.g., La/Nb <1 and positive HFSE anomalies in primitive mantle-normalised diagrams) typical of magmas emplaced in intra-plate settings

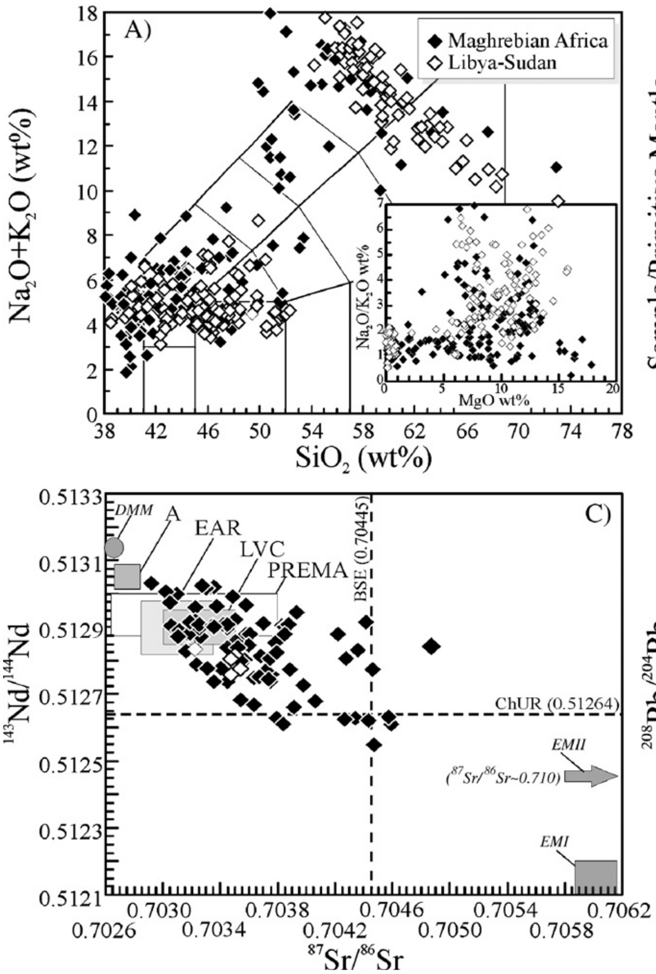


Fig. 4. A) Total alkali *vs.* silica diagram (TAS; le Maitre, 2002); B) primitive mantle (Sun and McDonough, 1989) normalised incompatible element diagram; C) $^{143}Nd/^{144}Nd$ *vs.* $^{87}Sr/^{86}Sr$ diagram; D) $^{208}Pb/^{204}Pb$ *vs.* $^{206}Pb/^{204}Pb$ diagram for igneous rocks from Maghrebian Africa (black diamonds) and Libya (white diamonds). Data sources: Maghrebian Africa (Hernandez and Lepvrier, 1979; Allègre et al., 1981; Berrahma and Hernandez, 1985; Rachdi et al., 1985; Bouabdli et al., 1988; Mokhtari and Velde, 1988; Velde and Rachdi, 1988; Bernard-Griffiths et al., 1991; Bouabdli and Liotard, 1992; Dautria et al., 1992; Berrahma et al., 1993; Semroud et al., 1994; Belanteur et al., 1995; Rachdi et al., 1997; El Bakkali et al., 1998; Maza et al., 1998; El Azzouzi et al., 1999; Ait-Hamou et al., 2000; Coulon et al., 2002; Wagner et al., 2003; Duggen et al., 2005; this paper); Libya (Piccoli, 1970; Almond, 1974; Almond et al., 1974; Busrewil and Wadsworth, 1980a, 1980b; Andre et al., 1991; Atherton et al., 1991; Busrewil and Oun, 1991; Flinn et al., 1991; Conticelli et al., 1995; this study). Abbreviations as in Fig. 2.

(Fig. 4). The most typical rock-types are basanites and nephelinites, followed by hawaiites, alkali basalts, tephri-phonolites and phonolites. Sr–Nd isotopic ratios suggest derivation of the magmas by mixing of partial melts from both lithospheric and EAR-like mantle sources ($^{87}Sr/^{86}Sr$ = 0.7029–0.7049, $^{143}Nd/^{144}Nd$ = 0.51255–0.51303; Fig. 4). A recent Pb isotopic study of these rocks shows a large range in $^{206}Pb/^{204}Pb$ (18.32–20.76), $^{208}Pb/^{204}Pb$ (38.71–40.43) Δ7/4 (−7.7 to +10.5) and Δ8/4 (−63.0 to +101.2; Duggen et al., 2005).

Within the Maghrebian margin of Africa there are rare occurrences of older igneous rocks (late Cretaceous–Tertiary; Woolley, 2001), all of alkaline composition. All these rocks are SiO_2-undersaturated to strongly SiO_2-undersaturated (e.g., nephelinites, monchiquites, melilitites) and have Na_2O/K_2O ratios > 1. The most important of these sub-provinces are the: Taourirt area (~67–37 Ma) at the NE end of the Middle Atlas (Morocco), mostly made up of alkaline lamprophyre (camptonite and monchiquite) sills and dykes, plus rarer olivine nephelinite lava flows and nepheline syenite intrusions (Wagner et al., 2003); Tamazert massif (~45–35 Ma) on the northern margin of the High Atlas (Morocco), made up mostly of a nepheline syenite intrusion plus much rarer carbonatites, basanites, phonolites, nephelinites and alkaline lamprophyre dykes (monchiquites; Bouabdli et al., 1988; Bernard-Griffiths et al., 1991); the Rekkame province (~50–32 Ma) south of the Tamazert massif (Morocco), characterized by melanephelinites, basanites and rare phonolites (Rachdi et al., 1997); Djebel Saghro (~10–2.8 Ma), a large stratovolcano in the Anti-Atlas mountains (Morocco), composed of a wide spectrum of igneous rocks, ranging from olivine nephelinites to basanites, hawaiites, phonotephrites and phonolites (Berrahma et al., 1993); Siroua volcano (~11–2 Ma), the southernmost Tertiary volcanic outcrop in Morocco, characterized by rare basanites and hawaiites and voluminous evolved rock types (mostly trachytes and phonolites; Berrahma and Hernandez, 1985; Berrahma and Delaloye, 1989; Bondi et al., 2002). The Djebel Saghro and Siroua volcanoes lie along a major fault zone running SW–NE that continues into the North Atlantic Ocean towards the Canary Islands. Far from the Maghrebian margin, in the Algerian Sahara, there are igneous rocks related to the Ahaggar (Hoggar) swell of Oligocene to Quaternary age, characterized by sodic alkaline SiO_2-undersaturated compositions (mostly alkali basalts and trachytes, plus rare essexites, monzonites, syenites and peralkaline rhyolites) and even rarer tholeiitic rocks (melagabbros and leucogabbros; e.g., Allègre et al., 1981; Dautrià et al., 1992; Maza et al., 1998; Ait-Hamou et al., 2000; Liegeois et al., 2005).

The origin of the Maghrebian Africa *anorogenic* magmatism has been related to processes involving lithospheric delamination and/or slab detachment in a post-collisional tectonic setting (e.g., Maury et al., 2000; Coulon et al., 2002; Duggen et al., 2005). According to the slab-detachment model, the asthenosphere underlying the downgoing plate (part of the Tethys oceanic lithosphere) flows up into the widening gap above the sinking detached slab. Maury et al. (2000) and Duggen et al. (2005) consider that the geochemical signature of the Maghrebian Africa *anorogenic* igneous rocks supports an origin from an enriched asthenospheric (HIMU-OIB-like) mantle source similar to that which feeds the other central European volcanic provinces (see below). This asthenospheric source closely resembles EAR. Maury et al. (2000) and Coulon et al. (2002) considered that it represents a channelised sub-lithospheric mantle plume extending from the Cape Verde Islands to Central Europe, similar to models proposed by a number of other authors (e.g., Oyarzun et al., 1997, Cebrià et al., 2000, Wilson and Patterson, 2001, Macera et al., 2004). El Bakkali et al. (1998) attributed the magmatism to the existence of an enriched, sub-lithospheric mantle source without any involvement of a mantle plume; they related the sequence of volcanism (from *orogenic* to *anorogenic*) to a succession of extensional, compressional and strike-slip fault tectonics. Duggen et al. (2005) developed a hybrid model: they excluded any possibility of derivation of the Maghrebian alkaline lavas directly from mantle plumes; however, they proposed an origin from an asthenospheric "plume-contaminated" mantle source, forced to upwell as consequence of lithosphere delamination. Anguita and Hernán (2000) favoured a more complex model requiring the existence of a weak thermal anomaly (fossil mantle plume) combined with lithospheric thin-spots (c.f., Thompson and Gibson, 1991) and regional tectonic forces. They related the Maghrebian magmatic activity to that of the Canary Islands.

The petrogenetic model proposed by Bouabdli et al. (1988) for the Eocene Tamazert massif alkaline lamprophyres and related rocks involves low-degree partial melting (~3%) of a carbonated, amphibole-bearing lherzolite mantle source, later suggested by Bernard-Griffiths et al. (1991) to be a Sr–Nd isotopically depleted mantle plume (with both HIMU and EMI geochemical characteristics), combined with variable amounts of contamination from ancient (~1 Ga) lithospheric mantle. A similar carbonated, amphibole ± mica-bearing lherzolitic source was suggested for the Taourirt area by Wagner et al. (2003). The enriched geochemical characteristics of the volcanic rocks of this area suggest that this mantle source was only recently metasomatised. The range of variation in the isotopic compositions of the silicate (lamprophyres, nephelinites and basanites) and carbonatitic rocks from Taourirt were explained by Wagner et al. (2003) in terms of mixing of three mantle end-members (HIMU, EMI, EMII). These authors considered that the presence of carbonatites was indicative of mantle plume activity on the basis of the axiom that carbonatites imply the presence of a mantle plume, as proposed by Bell and Tilton (2001).

The incompatible trace element and Sr–Nd–Pb isotopic compositions of the magmatic rocks from Ahaggar have been attributed to mixtures of partial melts from both HIMU- and EMI-like mantle sources, the first being located in the asthenospheric mantle (probably coming from some form of mantle plume), and the second located in Pan-African age lithospheric mantle (Allègre et al., 1981; Maza et al., 1998; Ait-Hamou et al., 2000). Liegeois et al. (2005), however, have recently discounted the existence of a mantle plume beneath the Ahaggar massif, relating the magmatism to a process of lithospheric delamination as a consequence of Africa-Europe collision, and subsequent asthenospheric upwelling and partial melting due to pressure release.

3.5. France

Cenozoic igneous activity in France ranges in age from ~65 to 0.004 Ma and is mostly confined to the European Continental Rift System (ECRiS) in the northern foreland of the Alpine orogenic belt (e.g.,

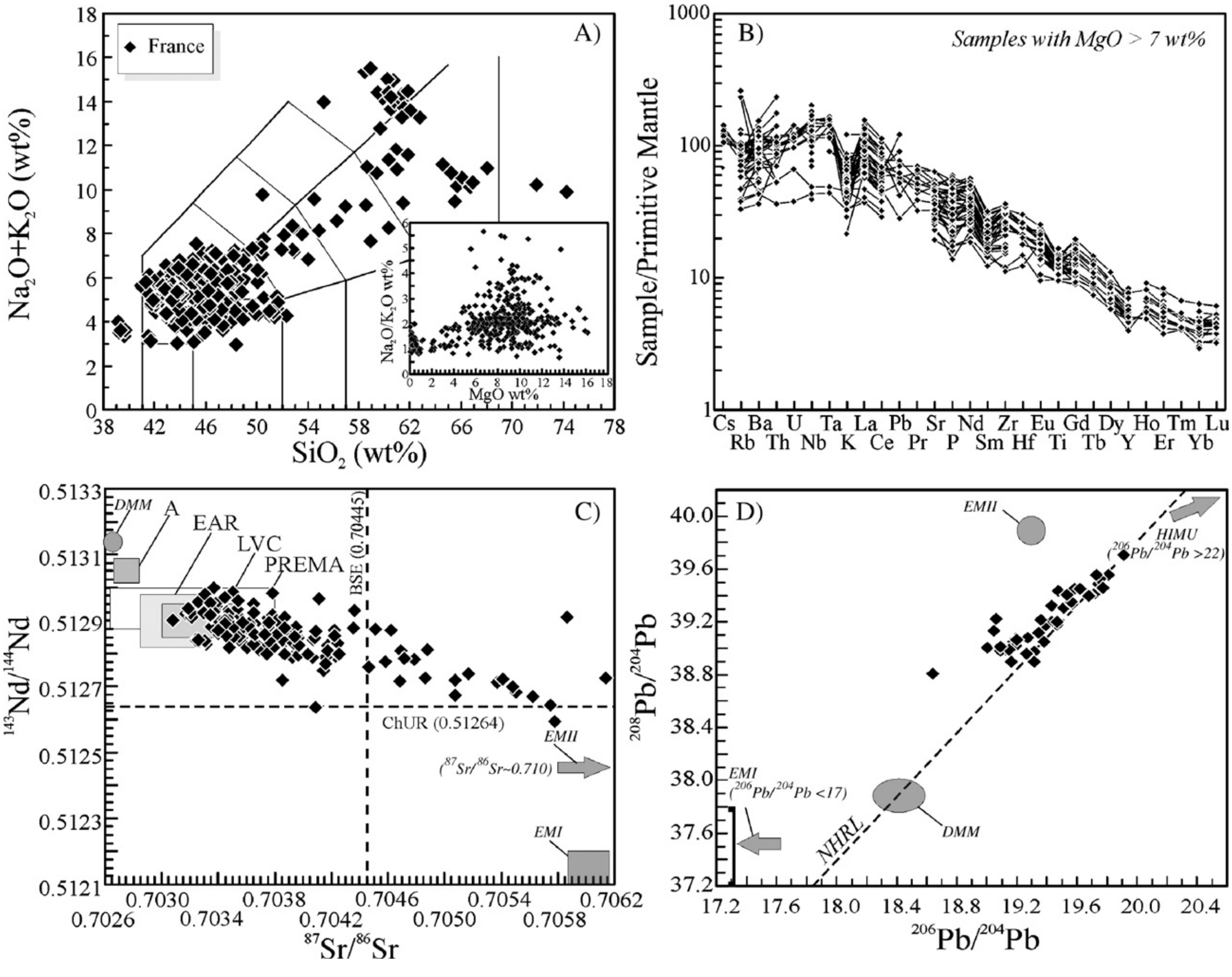


Fig. 5. A) Total alkali *vs.* silica diagram (TAS; le Maitre, 2002); B) primitive mantle (Sun and McDonough, 1989) normalised incompatible element diagram; C) $^{143}Nd/^{144}Nd$ *vs.* $^{87}Sr/^{86}Sr$ diagram; D) $^{208}Pb/^{204}Pb$ *vs.* $^{206}Pb/^{204}Pb$ diagram for igneous rocks from France. Data sources: Wimmenauer, 1964; Baubron and Demange, 1982; Alibert et al., 1983; Condomines et al., 1982; Downes, 1984, 1987; Chauvel and Jahn, 1984; Dautria and Liotard, 1990; Briot et al., 1991; Liotard et al., 1988, 1991, 1995, 1999; Wilson and Downes, 1991; Wilson et al., 1995a; Sørensen et al., 1999; Lenoir et al., 2000; Legendre et al., 2001; Wilson and Patterson, 2001; Pilet et al., 2002, 2005; Dautria et al., 2004; this study. Abbreviations as in Fig. 2.

Downes, 1987; Wilson and Downes, 1991; Ziegler, 1992; Michon and Merle, 2001; Nehlig et al., 2003; Dèzes et al., 2004; Fig. 1). The main volcanic areas are located in the Massif Central; these include: Cantal (~13–2 Ma), the largest Tertiary volcano in Europe (~8000 km^2), characterized by both strongly and weakly alkaline magma series (basanite–tephrite–phonolite and alkali basalt–hawaiite–trachyte–rhyolite; Chauvel and Jahn, 1984; Downes, 1984, 1987; Wilson and Downes, 1991; Wilson et al., 1995a; Legendre et al., 2001; Wilson and Patterson, 2001); Mont Dore (~3–0.25 Ma), comprising both SiO_2-saturated and SiO_2-undersaturated magma series lavas with rhyolitic ignimbrites followed by hawaiites, tephrites, trachytes and phonolites (Feraud et al., 1990; Briot et al., 1991; Gourgaud and Villemant, 1992; Sørensen et al., 1999); Aubrac (~7.5–5 Ma), a large basaltic plateau made up of mildly to strongly sodic alkaline lavas (mostly basanites, alkali basalts, hawaiites, mugearites, phonotephrites; Baubron and Demange, 1982; Chauvel and Jahn, 1984; De Goer de Herve et al., 1991); Chaine des Puys (0.15–0.04 Ma), made up entirely of mildly sodic alkaline volcanic rocks (basanites, alkali basalts, hawaiites, mugearites, benmoreites, trachytes; Condomines et al., 1982; Chauvel and Jahn, 1984); Velay (~14–0.25 Ma), characterized by sodic alkaline volcanic rocks (alkali basalts, trachytes and phonolites; Roger et al., 1999; Chazot et al., 2003; Dautria et al., 2004); Forez (~62–15 Ma) characterized by strongly SiO_2-undersaturated alkaline rocks (basanites, nephelinites, melilitites; Alibert et al., 1983; Lenoir et al., 2000). With the exception of a few occurrences, most of the Massif Central volcanic activity developed from the Miocene onwards and peaked between ~10 and 2 Ma. Extremely limited outcrops of Eocene (~47–46 Ma) volcanic breccias with alkaline lamprophyric and K-basanitic compositions have been reported by Liotard et al. (1991) in the region of Languedoc, southern France.

As noted above, both SiO_2-undersaturated (basanites, tephrites, phonolites) and SiO_2-saturated (alkali basalts, hawaiites, trachytes, rhyolites) were erupted, mainly with sodic affinity (Fig. 5). The most primitive mafic alkaline rocks often entrain mantle xenoliths (e.g., Werling and Altherr, 1997; Zangana et al., 1999; Downes, 2001). Magma mixing and minor crustal contamination, via AFC processes, frequently influence the geochemical characteristics of the intermediate to evolved products (e.g., Briot et al., 1991; Wilson et al., 1995a; Dautria et al., 2004). The large variation in Sr–Nd–Pb isotopic composition of the magmatic rocks ($^{87}Sr/^{86}Sr$ = 0.7031–0.7061, $^{143}Nd/^{144}Nd$ = 0.51260–0.51300, $^{206}Pb/^{204}Pb$ = 19.01–19.91, $^{208}Pb/^{204}Pb$ = 38.81–39.71, Δ7/4 = −1.2 to +12.1; Δ8/4 = −6.9 to +65.4; Fig. 5) has been considered to reflect mixing of partial melts of asthenospheric (EAR-like) and enriched lithospheric mantle sources, combined with both upper and lower crustal contamination in the petrogenesis of the more evolved magma types (Chauvel and Jahn, 1984; Downes, 1987; Briot et al., 1991; Wilson and Downes, 1991; Wilson et al., 1995a; Dautria et al., 2004). Highly radiogenic $^{87}Sr/^{86}Sr$ values (0.7087–0.7119) are recorded only in highly evolved rocks (phonolites), considered to reflect crustal contamination at shallow levels (Dautria et al., 2004). The trace element characteristics of the most primitive sodic basalts (inferred to be asthenosphere-derived) are similar to those of HIMU-OIB with distinctive negative K anomalies (Fig. 5). Rare potassic magma types (leucitites and leucite nephelinites), assumed by Wilson and Downes (1991) to be lithospheric partial melts, have much flatter trace element patterns from Cs–Pb and no negative K anomaly.

During the late stages of Ligurian–Provençal basin opening, volcanic activity also occurred along the French Mediterranean margin in three districts: Aix-en-Provence (mildly to strongly alkaline basanites, nephelinites, leucite–nephelinites and transitional basalts of Miocene age; ~18.2 Ma); Toulon (tholeiitic basalts; ~6.7–5.8 Ma) and Cap d'Agde (transitional basalts; ~1.3–0.6 Ma; Dautria and Liotard, 1990; Liotard et al., 1999; Wilson and Patterson, 2001; Fig. 5).

Mantle xenoliths are relatively common in the most primitive mafic alkaline volcanic rocks of the Massif Central (e.g., Jakni et al., 1996; Downes, 2001). With few exceptions, the $^{87}Sr/^{86}Sr$ isotopic ratios of these xenoliths fall in a narrow range from 0.70308 to 0.70337, with $^{143}Nd/^{144}Nd$ varying from MORB-like values (~0.51300) to EAR-like values (~0.51286; Liotard et al., 1999; Zangana et al., 1999; Wilson and Patterson, 2001). The Sr and Nd isotopic compositions show only limited overlap (within the EAR field) with the range of French Massif Central lavas indicating quite clearly that the mechanical boundary layer of the lithosphere (from which the xenoliths are derived) is not the dominant magma source component.

The topographic elevation of the Massif Central combined with the intense Cenozoic magmatic activity, the strongly deformed texture of mantle xenoliths entrained within alkaline basic igneous rocks from the periphery of the massif, the presence of a strong negative Bouguer gravity anomaly, and the identification of a finger-like mantle low-velocity anomaly by seismic tomography, suggest the presence of an active mantle diapir (mantle plume) beneath the main volcanic fields coming from the mantle Transition Zone (410–660 km; Hoernle et al., 1995; Granet et al., 1995; Sobolev et al., 1997; Zeyen et al., 1997; Zangana et al., 1999; Wilson and Patterson, 2001). More recently, however, interpretations of the evolution of the European Cenozoic rift system have emphasised the role of passive rifting in response to the build-up of Pyrenean and Alpine collision-related compressional stresses in magma generation processes (Dèzes et al., 2004). Such stresses, upon reaching a critical value, caused westward escape of the Massif Central block, inducing both northward and southward rift propagation. Nevertheless, Dèzes et al. (2004) still consider that a mantle plume-like upwelling of the asthenosphere is the cause of thermal weakening of the Alpine foreland lithosphere in France and Germany.

The Miocene–Quaternary volcanic cycle in the Aix-en-Provence, Toulon and Cap d'Agde areas has been related by Nehlig et al. (2003) to the same thermal anomaly which caused the Massif Central Paleocene–Quaternary volcanism. Liotard et al. (1999) considered that the trace element and Sr–Nd–Pb isotopic compositions of these magmatic rocks reflect mixing of three end-member components: ambient asthenosphere (EAR component) with high $^{206}Pb/^{204}Pb$ (~20) and low $^{87}Sr/^{86}Sr$ (~0.7030–0.7034), lithospheric mantle [with slightly lower $^{206}Pb/^{204}Pb$ (~18.75) and similar $^{87}Sr/^{86}Sr$] and a sedimentary component [with relatively low $^{206}Pb/^{204}Pb$ (18.25) and high $^{87}Sr/^{86}Sr$ (0.7065)].

3.6. Italy

Since Paleocene times, the area surrounding the Tyrrhenian Sea has been the locus of a wide variety of igneous activity (Fig. 1). Most of the magmatic rocks (e.g., the Oligo-Miocene igneous cycle of Sardinia, the Oligocene potassic–ultrapotassic rocks of NW Alps, the Oligocene calc-alkaline plutonic rocks of NE Alps, the Plio-Quaternary Tuscan and Roman magmatic provinces, the Plio-Quaternary rocks of the Aeolian Archipelago, the Plio-Quaternary rocks of Mt. Vulture) have geochemical characteristics compatible with magmas emplaced in *orogenic* settings (e.g., Venturelli et al., 1984; Beccaluva et al., 1991; Bigioggero et al., 1994; Conticelli et al., 2004; Francalanci et al., 2004; Peccerillo, 2005; Peccerillo and Lustrino, 2005) and, therefore, will not be discussed further here.

In the Tyrrhenian Sea, basalts dredged from two diachronous basins (Vavilov Basin, ~8–2 Ma; Marsili Basin, ~2–0.1 Ma; Argnani and Savelli, 1999; Trua et al., 2004) vary between E-MORB and calc-alkaline basalt in composition (Sborshchikov and Al'mukhamedov, 1992; Savelli and Gasparotto, 1994; Trua et al., 2002, 2004). These basalts are thought to be derived by partial melting of mantle sources variably metasomatised by slab-derived fluids during Alpine subduction of oceanic and continental African and European lithosphere.

During the Paleogene, anorogenic magmatism occurred in the Veneto Volcanic Province in NE Italy and at Punta delle Pietre Nere in Apulia (central-eastern Italy). A rare occurrence of Eocene alkaline lamprophyric magmatism has been reported in drill holes from SW Sardinia (Maccioni and Marchi, 1994). The Plio-Quaternary anorogenic magmatism of the circum Tyrrhenian Sea area (Mt. Etna and Hyblean Mts. in Sicily, Sardinia, Ustica Island in SE Tyrrhenian Sea and Pantelleria and Linosa Islands in the Sicily Channel) has been studied in the most detail (e.g., Peccerillo and Lustrino, 2005 and references therein).

3.6.1. Veneto volcanic province

Magmatism within the ~2000 km^2 Veneto Volcanic Province occurred over a large time span of ~35 Myr (from late Paleocene to late Eocene; ~60–25 Ma). Activity is concentrated in four main volcanic districts characterized by the eruption of predominantly *anorogenic* basic lavas (mostly alkali basalts, hawaiites, benmoreites, trachytes; Lessini Mts., Marostica Hill, Berici Hills); only rarely are acidic rock types dominant (quartz trachytes and rhyolites; Euganean Hills; De Vecchi and Sedea, 1995; Milani et al., 1999; Macera et al., 2003; Beccaluva et al., 2007; Fig. 6). The most primitive mafic volcanic rocks often exhume mantle xenoliths (e.g., Beccaluva et al., 2001b). Sr isotope compositions are relatively unradiogenic ($^{87}Sr/^{86}Sr$ = 0.7031–0.7039), whereas $^{143}Nd/^{144}Nd$ ratios are highly radiogenic ($^{143}Nd/^{144}Nd$ = 0.51298–0.51279). $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ isotopic ratios are variably radiogenic (18.79–19.79 and 38.81–39.49, respectively); Δ7/4 ranges from −1.8 to +12.3 and Δ8/4 from −9.0 to +4.

The *anorogenic* geochemical characteristics of the magmatism are somewhat inconsistent with the regional geodynamic setting — an extensional system to the south of the main Alpine compressional tectonic zone. On the basis of seismic low velocity anomalies between 100 and 250 km depth, Piromallo and Morelli (2003) and Macera et al. (2003) proposed the existence of a mantle plume beneath the Veneto Volcanic Province. The apparent paradox of a mantle plume in a collisional tectonic setting was explained by Macera et al. (2003) as a consequence of slab detachment, with deeper mantle material upwelling through a slab window. Most of the Veneto volcanic rocks have HIMU-like geochemical and Sr–Nd isotope characteristics (Fig. 6), consistent with the involvement of an EAR-like mantle plume component in their petrogenesis.

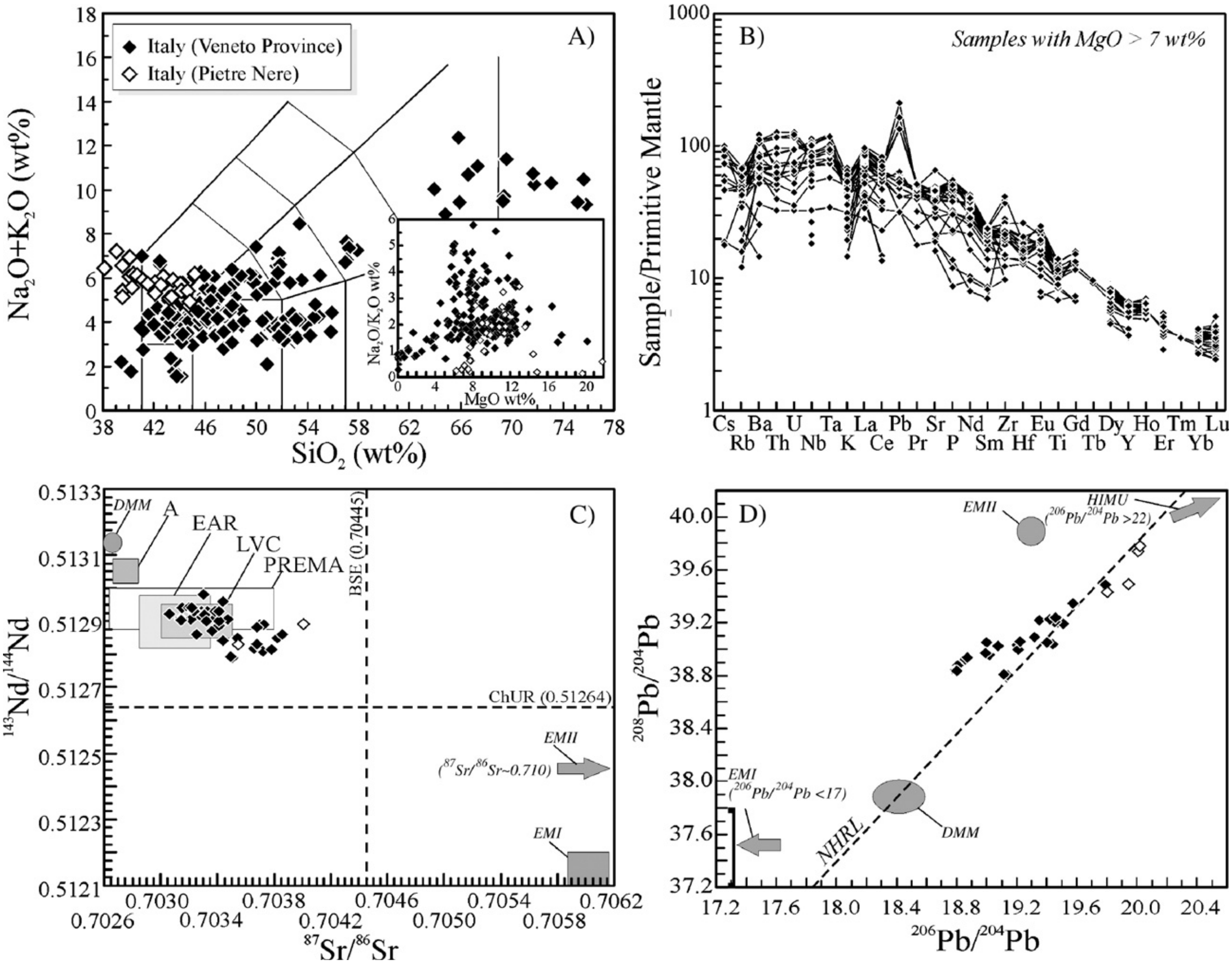


Fig. 6. A) Total alkali *vs.* silica diagram (TAS; le Maitre, 2002); B) primitive mantle (Sun and McDonough, 1989) normalised incompatible element diagram; C) $^{143}Nd/^{144}Nd$ *vs.* $^{87}Sr/^{86}Sr$ diagram; D) $^{208}Pb/^{204}Pb$ *vs.* $^{206}Pb/^{204}Pb$ diagram for igneous rocks from Veneto, Italy (black diamonds) and Punta delle Pietre Nere, Italy (white diamonds). Data sources: Veneto (De Vecchi et al., 1976; De Vecchi and Sedea, 1995; Milani et al., 1999; Macera et al., 2003; Beccaluva et al., 2007a); Punta delle Pietre Nere (Vollmer, 1976; Vollmer and Hawkesworth, 1980; De Fino et al., 1981; De Astis et al., 2006). Abbreviations as in Fig. 2.

De Vecchi and Sedea (1995), however, proposed a different scenario in which the Veneto igneous activity and the Cenozoic E–W extensional tectonics of the southern Alpine foreland were a response to N–S compression throughout the Alps, without requiring any mantle plume involvement.

3.6.2. Punta delle Pietre Nere — Mt Queglia

The Punta delle Pietre Nere is a small outcrop (~800 m^2) on the Adria Plate, along the northern coast of the Apulian region (Fig. 1) of Paleocene (62–56 Ma; Vollmer, 1976; Bigazzi et al., 1996; De Astis et al., 2006) intrusive mafic alkaline rocks (mela-syenite and mela-gabbro; Fig. 6) with lamprophyric affinity (De Fino et al., 1981). Vollmer (1991) proposed that the magmatism represents the first manifestation of a mantle hot spot which later gave rise to the Roman Comagmatic Province. $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ isotopic ratios show little variation (0.7035–0.7040 and 0.51289–0.51283, respectively; Fig. 6). $^{206}Pb/^{204}Pb$ ratios (19.80–20.14) are amongst the most radiogenic in the CiMACI province, while $^{208}Pb/^{204}Pb$ ranges from 39.43 to 39.80; Δ7/4 is positive (+4.8 to +6.8) while Δ8/4 is negative (−5.1 to −24.1; Vollmer, 1976; De Astis et al., 2006).

A poorly known occurrence of ultramafic lamprophyre, represented by a small dike swarm, has been reported in central Italy (Mt. Queglia, Abruzzo region; Barbieri and Ferrini, 1984; Durazzo et al., 1984). No geochronological data are available; however, since the dykes cut limestones of Eocene age, their age must be Paleogene or younger. The $^{87}Sr/^{86}Sr$ isotopic composition of the least altered samples is ~0.70344. Recently,

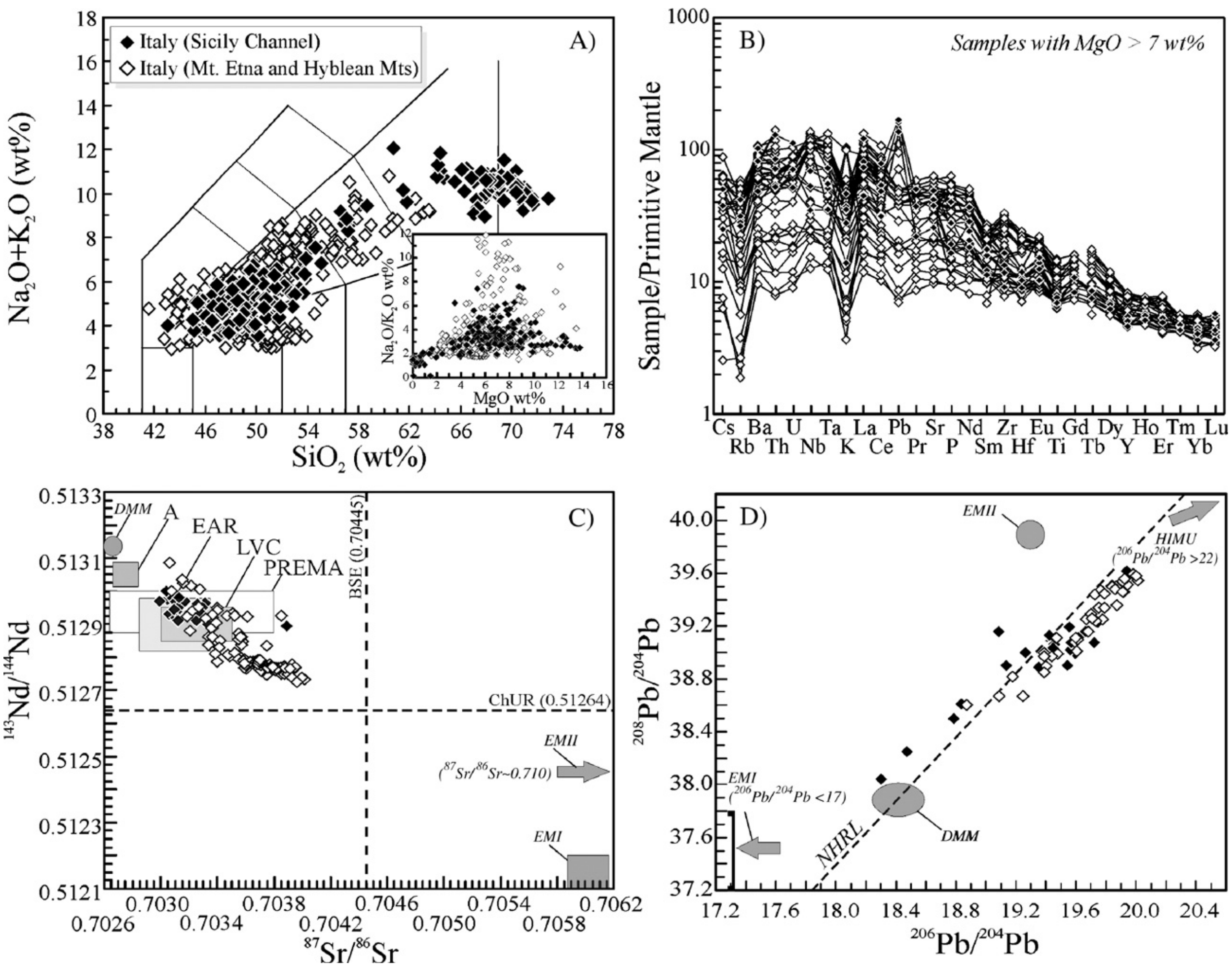


Fig. 7. A) Total alkali *vs.* silica diagram (TAS; le Maitre, 2002); B) primitive mantle (Sun and McDonough, 1989) normalised incompatible element diagram; C) $^{143}Nd/^{144}Nd$ *vs.* $^{87}Sr/^{86}Sr$ diagram; D) $^{208}Pb/^{204}Pb$ *vs.* $^{206}Pb/^{204}Pb$ diagram for igneous rocks from Mt. Etna and Hyblean Mts. (Italy; white diamonds), and the islands of Ustica, Pantelleria and Linosa in the Sicily Channel (Italy; black diamonds). Data sources: Mt. Etna (Carter and Civetta, 1977; Cristofolini et al., 1984, 1991; Tanguy and Clocchiatti, 1984; Armienti et al., 1988, 1989, 1994, 2004; Corsaro and Cristofolini, 1993, 1996; Finocchiaro, 1994; Treuil and Joron, 1994; Condomines et al., 1995; Tonarini et al., 1995; D'Orazio et al., 1997; Tanguy et al., 1997; Clocchiatti et al., 1998 this study); Hyblean Mts. (Carter and Civetta, 1977; Tonarini et al., 1996; Beccaluva et al., 1998; Trua et al., 1998; Bianchini et al., 1998, 1999; Di Grande et al., 2002); Ustica (Cinque et al., 1988; Trua et al., 2003; Peccerillo, 2005); Pantelleria (Romano, 1968; Civetta et al., 1984, 1998; Esperança and Crisci, 1995; Avanzinelli et al., 2004; Ferla and Meli, 2006); Linosa (Barberi et al., 1969; Calanchi et al., 1989; Rossi et al., 1996; Civetta et al., 1998; Bindi et al., 2002; Peccerillo, 2005); Sicily Channel (Beccaluva et al., 1981; Calanchi et al., 1989; Rotolo et al., 2007). Abbreviations as in Fig. 2.

Beccaluva et al. (2007a) reported the first Pb isotopc ratios for the Mt. Queglia lamprophyres; these are the most radiogenic within the entire CiMACI province: $^{206}Pb/^{204}Pb \sim 20.4–21.5$; $^{208}Pb/^{204}Pb \sim 40.5–41.5$.

3.6.3. Mt. Etna

The volcanic activity of Mt. Etna started about 0.6 Ma ago (Gillot et al., 1994) characterized by the eruption of tholeiitic to transitional lavas, and evolved with time towards more sodic alkaline magma compositions (Tanguy et al., 1997; Schiano et al., 2001; Armienti et al., 2004; Corsaro and Pompilio, 2004). With the exception of rare picritic magma eruptions (e.g., Coltelli et al., 2005), the most recent eruptive products (younger than 250 ka) have a mildly alkaline (sodic) affinity and relatively evolved compositions (hawaiites, mugearites, benmoreites, trachytes), whereas the oldest products (580–250 ka) are tholeiitic basalts (e.g., Tanguy et al., 1997; Corsaro and Pompilio, 2004; Fig. 7). The scarcity of near-primitive melts is mainly attributable to shallow-level fractionation processes (D'Orazio et al., 1997; Armienti et al., 2004; Corsaro and Pompilio, 2004). Moderate assimilation of crustal lithologies and heterogeneity of the mantle source is suggested by the variability of Sr and Nd isotope and trace element ratios ($^{87}Sr/^{86}Sr$ = 0.7030–0.7038; $^{143}Nd/^{144}Nd$ = 0.51293–0.51283; $^{206}Pb/^{204}Pb$ = 19.47–20.01; $^{208}Pb/^{204}Pb$ = 39.11–39.59; $\Delta 7/4$ = −3.4 to +4.5; $\Delta 8/4$ = −27.7 to −3.6), which are similar to the EAR mantle reservoir and HIMU-OIB (Fig. 7).

Volcanic activity at Mt. Etna has been considered to be related to asthenospheric flow through a slab window created as consequence of differential roll-back of the subducted Ionian (Mesogean) oceanic lithosphere relative to the neighbouring African and Adrian plates: passive upwelling of asthenospheric mantle through such a slab-window would cause partial melting as consequence of adiabatic decompression (e.g., Gvirtzman and Nur, 1999; Doglioni et al., 2001; Armienti et al., 2004). Such asthenospheric "suction" might also mobilise subduction-related fluids coming from the neighbouring Ionian oceanic plate (e.g., Tonarini et al., 2001). However, in contrast with this model (that implies a shallow origin for Mt. Etna melts), Montelli et al. (2004) have suggested the presence of a deep mantle plume beneath Mt. Etna, on the basis of a detailed global seismic tomographic study.

3.6.4. Hyblean Plateau

Unlike most of the other circum-Mediterranean volcanic provinces, the Hyblean Plateau represents a long-lived magmatic system (~200 Myr), whose early products (known only from drill cores) are late Triassic basaltic rocks while the youngest are Pleistocene (Longaretti and Rocchi, 1990; Beccaluva et al., 1998; Trua et al., 1998; Di Grande et al., 2002). The Cenozoic volcanic products include rare strongly alkaline sodic mafic rocks (nephelinites and ankaratrites; ~1%), transitional basalts (~3%), mildly alkaline sodic rocks (basanites, alkali basalts and hawaiites; ~32%) and abundant tholeiitic basalts (~62%; Di Grande et al., 2002; Fig. 7). The most primitive mafic alkaline rocks often entrain crustal and mantle xenoliths (e.g., Tonarini et al., 1996; Scribano et al., 2006). The trace element (e.g., relatively high Nb–Ta contents) and Sr–Nd–Pb isotopic characteristics ($^{87}Sr/^{86}Sr$ = 0.7027–0.7036; $^{143}Nd/^{144}Nd$ = 0.51325–0.51290; $^{206}Pb/^{204}Pb$ = 18.87–20.00; $^{207}Pb/^{204}Pb$ = 15.57–15.70; $^{208}Pb/^{204}Pb$ = 38.60–39.58; $\Delta 7/4$ = −3.0 to +8.6; $\Delta 8/4$ = −31.6 to +0.4) are EAR-like with similarities to HIMU-OIB (Fig. 7). Widely varying normalised abundances of the most incompatible elements reflect varying degrees of partial melting of a heterogeneous mantle source.

The origin of the most recent volcanic activity in the Hyblean Plateau has been related to extensional tectonics and the development of pull-apart basins, with subsequent adiabatic decompressive melting (Trua et al., 1997; Beccaluva et al., 1998). On the basis of major and trace element and Sr–Nd–Pb isotopic constraints, Beccaluva et al. (1998) proposed that the Miocene–Pliocene volcanism of the Hyblean Plateau was caused by partial melting of a depleted mantle source, variably modified by asthenosphere-derived metasomatising melts or fluids, with both alkali–silicate and carbonatitic compositions, capable of stabilizing metasomatic phases (e.g., amphibole, phlogopite, apatite and carbonate), which partly replaced the original paragenesis.

3.6.5. Sardinia

In Sardinia (Fig. 1) extensional tectonics resulted in the development of a rift system (Campidano graben in the SW; N–S oriented fault system in the Sarrabus area in the SE) and in intra-plate magmatism which lasted from ~6.6 to ~0.1 Ma (Lustrino et al., 2004b, 2007). Recent $^{40}Ar/^{39}Ar$ data (Lustrino et al., in press) suggest that the onset of this magmatic phase may have occurred much earlier, in the middle Miocene (~11.8 Ma). Both alkaline (basanites, alkali basalts, hawaiites, mugearites, benmoreites, trachytes and phonolites) and subalkaline (tholeiitic basalts, basaltic andesites, andesites, dacites and rhyolites) magmatic rock types are present (Fig. 8). The alkaline rocks are mildly to strongly alkaline, with sodic affinity. The most primitive mafic alkaline rocks often exhume abundant mantle xenoliths (e.g., Lustrino et al., 1999, 2004a; Beccaluva et al., 2001a). The subalkaline rocks are less primitive than their alkaline counterparts and are tholeiitic in character (Lustrino et al., 2000, 2004b).

The middle Miocene–Quaternary mafic volcanic rocks of Sardinia have been subdivided geochemically into two types (UPV and RPV) by Lustrino et al. (2000). The UPV (Unradiogenic Pb Volcanics) type is by far the most abundant (>99%) and is characterized by unradiogenic Pb isotope compositions ($^{206}Pb/^{204}Pb$ 17.36–18.04), Fig. 8. The much rarer RPV (Radiogenic Pb Volcanics) type has significantly more radiogenic Pb isotope compositions ($^{206}Pb/^{204}Pb$ 18.84–19.42). These two magma types occur in distinct areas, and define a significant geochemical discontinuity, roughly running E–W across the southern part of the island, in which the RPV rocks are concentrated (Lustrino et al., 2004b, 2007). With the exception of the most evolved samples (dacites and rhyolites), the $^{87}Sr/^{86}Sr$ isotopic composition of the UPV rocks ranges from 0.7042 to 0.7055, and $^{143}Nd/^{144}Nd$ from 0.51263–0.51231, plotting in the crustal quadrant of the Nd–Sr isotope diagram. Also with the exception of the most evolved samples (trachytes), the RPV rocks have $^{87}Sr/^{86}Sr$ in the range

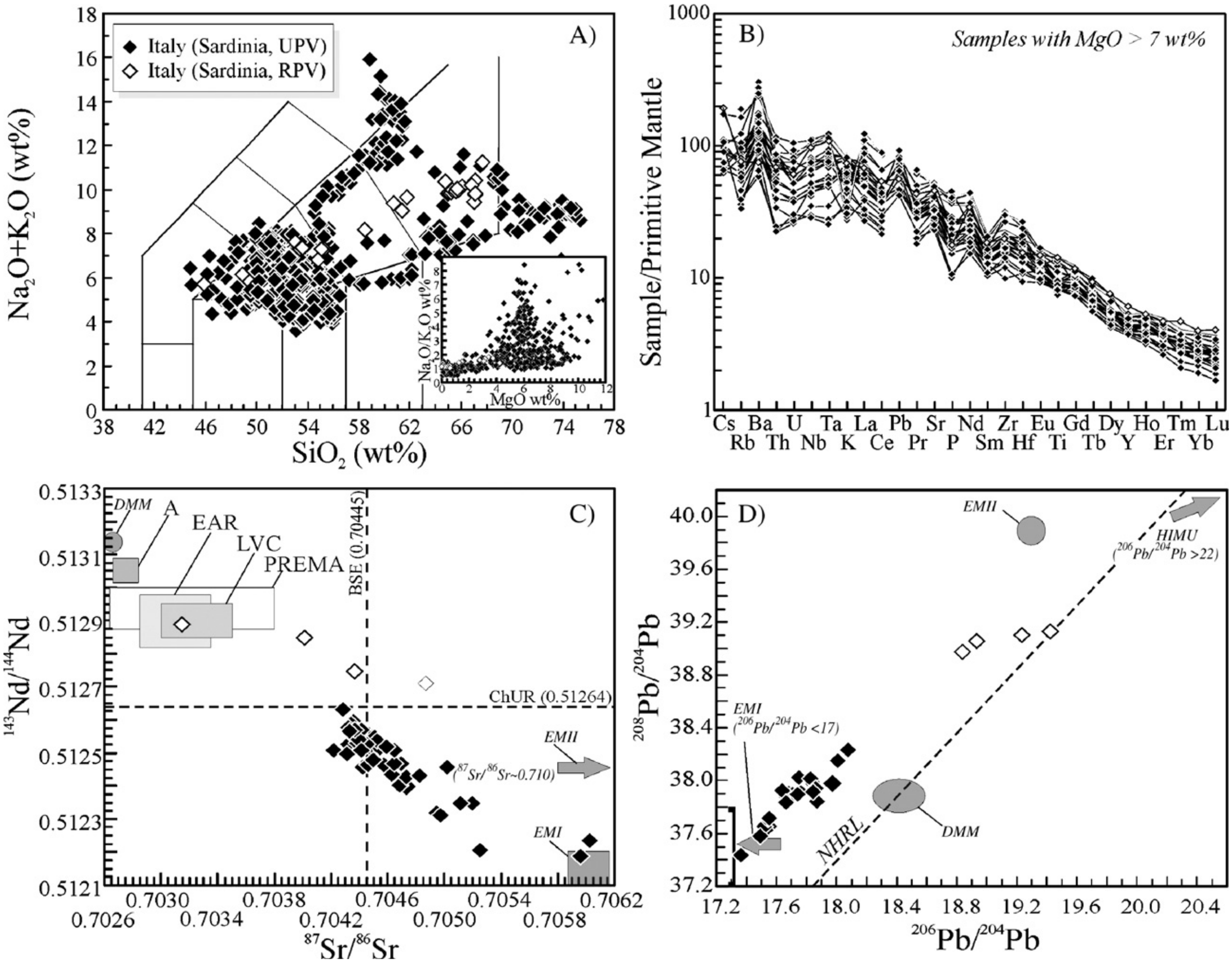


Fig. 8. A) Total alkali *vs.* silica diagram (TAS; le Maitre, 2002); B) primitive mantle (Sun and McDonough, 1989) normalised incompatible element diagram; C) $^{143}Nd/^{144}Nd$ *vs.* $^{87}Sr/^{86}Sr$ diagram; D) $^{208}Pb/^{204}Pb$ *vs.* $^{206}Pb/^{204}Pb$ diagram for igneous rocks from Sardinia. UPV = unradiogenic Pb volcanic group (black diamonds; Lustrino et al., 2000); RPV = radiogenic Pb volcanic group (white diamonds; Lustrino et al., 2000). Data sources: UPV (Brotzu et al., 1969; Beccaluva et al., 1973, 1974, 1975; Deriu et al., 1973; Assorgia et al., 1976; Cioni et al., 1982; Rutter, 1987; Di Battistini et al., 1990; Montanini et al., 1994; Lustrino et al., 1996, 2000, 2002, 2004a, 2007; Gasperini et al., 2000; Godano, 2000; Lustrino, 1999, 2000b; Fedele et al., in press; this study); RPV (Brotzu et al., 1975; Lustrino et al., 2000, 2007). Abbreviations as in Fig. 2.

0.70312–0.7043 and $^{143}Nd/^{144}Nd$ from 0.51289–0.51274 (Gasperini et al., 2000; Lustrino et al., 2000, 2002, 2004b, 2007).

The UPV rocks of Sardinia have the least radiogenic $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ isotopic compositions of the entire Cenozoic European Volcanic Province (17.37–18.07 and 37.44–38.23, respectively; Fig. 8). These samples show also the highest Δ7/4 (+11.7 to +20.7) and very high Δ8/4 (+60.8 to +97.1). Isotopically, they closely resemble the EMI mantle end-member composition (Lustrino and Dallai, 2003); this signature is believed to be a source feature related to mixing of depleted mantle with low $^{87}Sr/^{86}Sr$ and high $^{143}Nd/^{144}Nd$ with partial melts of ancient lower crust (e.g., Lustrino et al., 2004b, 2007; Lustrino, 2005a). On the other hand, Gasperini et al. (2000) proposed that the parental magmas in the Logudoro area (northern Sardinia) originated from a deep mantle plume mantle source modified by subduction of a large oceanic plateau which included rocks rich in cumulate plagioclase. The rarer RPV basanites have HIMU-like geochemical characteristics similar to those of other CiMACI basalts. Their origin is still not well constrained; however, Lustrino et al. (2004b, 2007) and Peccerillo and Lustrino (2005a,b) considered the hypothesis of an active upwelling of deep mantle highly improbable.

3.6.6. Ustica

The island of Ustica is located to the NW of Sicily, in the southern Tyrrhenian Sea, and is made up of

Quaternary volcanic rocks with Na-alkaline affinity which range in composition from alkali basalts to rarer trachytes (Cinque et al., 1988; Trua et al., 2003; Peccerillo, 2005; Fig. 7). Ustica lavas have a small range in $^{87}Sr/^{86}Sr$ (0.7030–0.7033) and $^{143}Nd/^{144}Nd$ (0.51300–0.51290) and a slightly wider range of compositions for $^{206}Pb/^{204}Pb$ (18.84–19.56), $^{208}Pb/^{204}Pb$ (38.61–39.19), Δ7/4 (+3.3 to +7.4) and Δ8/4 (−7.8 to +20.8); Trua et al., 2003; Peccerillo, 2005; Fig. 7).

On the basis of their trace element and Sr–Nd isotopic compositions, Cinque et al. (1988) and Trua et al. (2003) proposed that the source of the Ustica alkali basalts was slightly contaminated by a component of slab origin. It has been suggested that the volcanic activity of Ustica (together with that of the Prometeo seamount and Mt. Etna) can be explained without the need to invoke a deep-rooted mantle plume, but considering instead flow of African mantle below Sicily, channeled along the SW edge of the Ionian oceanic plate which is undergoing subduction roll-back (e.g., Gvirtzman and Nur, 1999; Doglioni et al., 2001; Trua et al., 2003).

3.6.7. Sicily channel

The islands of Pantelleria and Linosa (Fig. 1) lie in a transtensive zone of post-Miocene NW-trending horsts and grabens located between Sicily and Tunisia (Sicily Channel), possibly related with the Plio-Pleistocene rift system which developed in SW Sardinia (the Campidano Graben). The Pantelleria products (~118–4 ka) have sodic affinity and are alkaline to peralkaline (Esperança and Crisci, 1995; Civetta et al., 1998; Avanzinelli et al., 2004; Fig. 7). Basalt and hawaiite flows are volumetrically scarce with respect to more evolved compositions such as trachytes and rhyolites. Variable trace element ratios and elevated $^{207}Pb/^{204}Pb$ are believed to be indicators of the involvement of ancient enriched mantle source components (which resemble the enriched lithosphere sampled by circum-Tyrrhenian potassic magmas) in addition to DMM-HIMU (± EMI) mantle sources (Esperança and Crisci, 1995; Civetta et al., 1998). With the exception of one sample, the total range of $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ is 0.7030–0.7033 and 0.51300–0.51292, respectively. Compared to the limited range of Sr–Nd isotope compositions, the Pb isotopic ratios show a much wider range ($^{206}Pb/^{204}Pb$ = 18.30–19.94, $^{208}Pb/^{204}Pb$ = 38.04–39.62: Δ7/4 = −6.4 to +11.6; Δ8/4 = −39.4 to +45.8); no substantial differences in terms of Sr–Nd–Pb isotopic ratios are evident between trachytes/pantellerites and mafic rocks.

The volcanic products of the island of Linosa (1.06–0.53 Ma; Rossi et al., 1996) are mainly of sodic alkaline to slightly transitional character and are mainly represented by basalts and hawaiites; more evolved lithologies (mugearites, benmoreites, trachytes) have also been found (Fig. 7). Trace element abundances (Rossi et al., 1996; Bindi et al., 2002; Peccerillo, 2005) suggest a within-plate tectonic setting, with bell-shaped patterns in primitive mantle-normalised multi-elemental diagrams, peaking at Nb with relatively low LILE abundances, within the HIMU-OIB compositional field (Fig. 7). Sr–Nd–Pb isotopic ratios (Civetta et al., 1998) range as follows: $^{87}Sr/^{86}Sr$ from 0.7030 to 0.7031, $^{143}Nd/^{144}Nd$ from 0.51298 to 0.51295, $^{206}Pb/^{204}Pb$ from 19.28 to 19.43, $^{207}Pb/^{204}Pb$ from 15.60 to 15.62, $^{208}Pb/^{204}Pb$ from 38.89 to 39.02, Δ7/4 from +1.5 to +2.8, Δ8/4 from −12.5 to −4.3.

On the basis of major and trace element and Sr–Nd–Pb isotopic constraints, Civetta et al. (1998) proposed that the magmatism of Pantelleria, as well as that of Mt. Etna, is related to the activity of a mantle plume, consistent with the model of Hoernle et al. (1995). On the other hand, Esperança and Crisci (1995) prefer a petrogenetic model in which a lithospheric mantle source, enriched and metasomatised by asthenospheric partial melts, undergoes partial melting in an extensional tectonic regime without any mantle plume involvement.

Geochemical data on the Sicily Channel seamounts (e.g., Graham and Nameless Banks, Tetide, Anfitrite, Galatea and Cimotoe seamounts, plus other minor volcanic centers) are limited (e.g., Beccaluva et al., 1981; Calanchi et al., 1989). However, a recent study (Rotolo et al., 2006) allows comparison of the composition of some of these seamounts (Graham and Nameless Banks, plus two unnamed seamounts E and SE of Pantelleria Island) with the other Sicily Channel volcanic rocks. Rotolo et al. (2006) report major and trace element and Sr–Nd–Pb isotopic data for dredged samples from this area. Rock types range from hawaiites to alkali basalts, basanites, mugearites and tholeiitic basalts, in order of abundance. Volcanic rocks from Graham and Nameless Banks have similar geochemical characteristics, broadly resembling HIMU-OIB; small differences are probably related to sea-water alteration (Rotolo et al., 2006). As a whole, the Sicily Channel seamounts have $^{87}Sr/^{86}Sr$ ranging from 0.70308 to 0.70478, $^{143}Nd/^{144}Nd$ from 0.51312 to 0.51295, $^{206}Pb/^{204}Pb$ from 19.15 to 19.81, $^{207}Pb/^{204}Pb$ from 15.66 to 15.70, $^{208}Pb/^{204}Pb$ from 38.95 to 39.62, Δ7/4 from + 3.9 to +13.0, Δ8/4 from −13.0 to +19.4. Rotolo et al. (2006) have proposed the presence of a fossil plume head beneath the area, including the Sicily Channel, Mt. Etna, the Hyblean

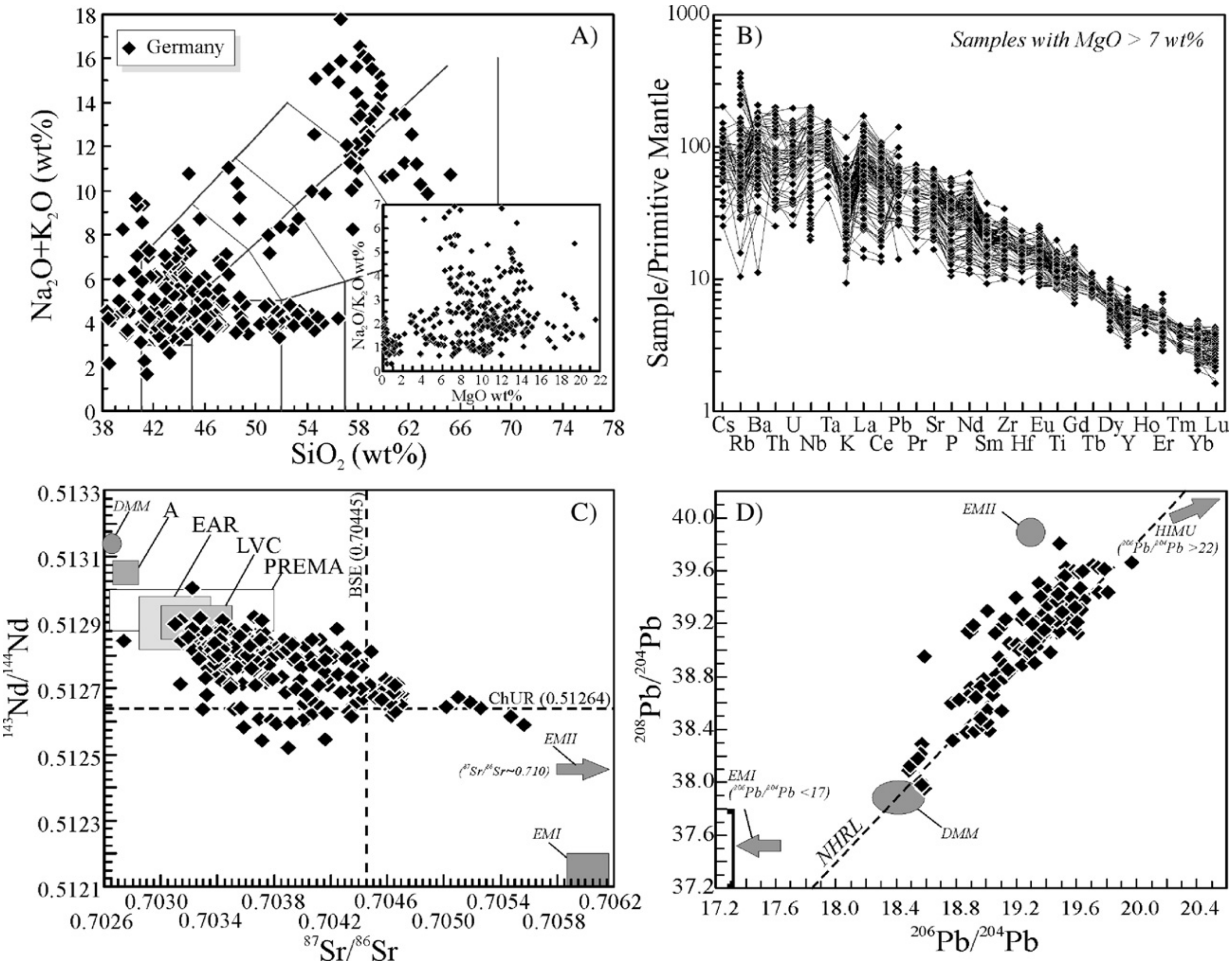


Fig. 9. A) Total alkali *vs.* silica diagram (TAS; le Maitre, 2002); B) primitive mantle (Sun and McDonough, 1989) normalised incompatible element diagram; C) $^{143}Nd/^{144}Nd$ *vs.* $^{87}Sr/^{86}Sr$ diagram; D) $^{208}Pb/^{204}Pb$ *vs.* $^{206}Pb/^{204}Pb$ diagram for igneous rocks from Germany. Data sources: Wimmenauer, 1964; Hoefs and Wedepohl, 1968; Duda and Schmincke, 1978; Alibert et al., 1983; Wörner et al., 1983, 1986, 1987; Wörner and Schmincke, 1984; Mertes and Schmicke, 1985; Wedepohl, 1985, 2000; Harmon et al., 1987; Hubberten et al., 1988; Kramm and Wedepohl, 1990; Schleicher et al., 1990, 1991; Ehrenberg et al., 1994; Wedepohl et al., 1994; Hegner et al., 1995; Jung, 1995; Wilson et al., 1995b; Dunworth and Wilson, 1998; Jung and Masberg, 1998; Riley et al., 1999; Schreiber et al., 1999; Wedepohl and Baumann, 1999; Jung and Hoernes, 2000; Meyer et al., 2002; Bogaard and Wörner, 2003; Haase et al., 2004; Jung et al., 2005, 2006; Fekiacova et al., in press-a,b. Abbreviations as in Fig. 2.

Plateau, Ustica Island and the westernmost island of the Aeolian Archipelago, Alicudi). Their model is based on the relatively homogeneous EAR-like major and trace element and Sr–Nd–Pb isotopic characteristics of the magmatic rocks.

3.7. Germany

The Tertiary igneous activity of Germany (Fig. 1) is probably the best studied within the CiMACI province in terms of its geochemistry, volcanology and petrology. Whilst there is a broad correlation with regions of Cenozoic rifting (e.g., Ziegler, 1992; Dèzes et al., 2004), the main volcanic areas are associated with an E–W trending tectonic suture zone in the Variscan basement, within an area of Cenozoic topgraphic uplift, the Rhenish Massif. The main graben structure, the NNE–SSW trending Rhine graben, is essentially non-magmatic with the exception of the Kaiserstuhl volcano in its southern part (Wilson and Downes, 1991). The formation of the Rhine Graben commenced in the late Eocene with extensional climaxes in the middle Eocene–early Oligocene (~42–31 Ma) and late Oligocene–early Miocene (~25–20 Ma; Ziegler, 1992). In contrast, the main uplift of the Rhenish Massif occurred in the late Oligocene (Meyer et al., 1983), postdating the onset of graben formation and the earliest Cenozoic (~60 Ma) magmatic activity in Germany (Keller et al., 2002). As in

France, the German volcanic fields are located in the northern foreland of the Alpine orogenic belt (Fig. 1).

The most important volcanic areas are from W to E: Eifel (~45–24 Ma and 0.5–0.01 Ma), Siebengebirge (~28–6 Ma), Westerwald (~32–0.4 Ma), Vogelsberg (~21–9 Ma), Hessian Depression (~21–8 Ma), Rhön (26–11 Ma), Heldburg (~42–11 Ma) and Upper Palatinate (~29–19 Ma); these ages are based on K–Ar data from Lippolt (1983). Three other small igneous districts occur in the southern part of Germany: Kaiserstuhl (~18–13 Ma), Bad Urach (~17–16 Ma) and Hegau (~15–7 Ma). Magmatic activity was long-lasting (~45 Myr) and peaked between ~30 and 10 Ma, covering a total area of ~13,000 km^2. Fekiacova et al. (in press-a), in a recent Ar–Ar geochronology study, have demonstrated that there are two middle–late Eocene peaks in the volcanic activity of the Hocheifel from 44–39 Ma and 37–35 Ma. These authors also dated the earliest magmatism in the Upper Rhine Graben (59–47 Ma) which appears to pre-date the onset of rifting.

The dominant rock types are mildly alkaline basic rocks of sodic affinity (basanites, alkali basalts, hawaiites) plus rarer differentiated products (trachytes, phonolites); olivine tholeiites and quartz tholeiites (the latter mostly confined to the Vogelsberg and the Hessian Depression) are present locally (Wedepohl et al., 1994; Wedepohl and Baumann, 1999; Wedepohl, 2000; Fekiacova et al., in press-b; Fig. 9). The most primitive mafic alkaline rocks often entrain mantle xenoliths (e.g., Franz et al., 1997; Shaw and Klugel, 2002; Witt-Eickschen et al., 2003). With few exceptions (e.g., Urach–Hegau) the differentiated products are the earliest within a particular volcanic area.

During the Cenozoic, magmatic activity within the Eifel (Hocheifel) region was characterized by eruption of intermediate, mildly alkaline sodic magmas (predominantly nephelinites and basanites, but also including more evolved hawaiites, mugearites, benmoreites and rare trachytes; Jung et al., 2006; Fekiacova et al., in press-b). Quaternary volcanism within the Eifel, however, is predominantly potassic with $K_2O/Na_2O > 1$, although sodic magmas (basanites, olivine nephelinities) occur locally in the West Eifel volcanic field (Schmincke, in press). During Quaternary times the production of differentiates exceeded the volume of basic to intermediate rocks, the latter characterized by strongly silica-undersaturated basanitic, nephelinitic, leucititic and melilititic compositions (Mertes and Schmicke, 1985; Wörner et al., 1986; Wedepohl et al., 1994; Bogaard, 1995). The differentiated rocks of Hocheifel show Sr–Nd–Pb isotopic evidence for assimilation of lower crustal material (e.g., Jung et al., 2006). Volcanism in the Siebengebirge is predominantly of more differentiated magma types (mostly trachytes plus benmoreites; Wedepohl et al., 1994). In Westerwald (the second largest outcrop of Cenozoic volcanic rocks in Germany; ~1000 km^2) the volcanism is mostly basaltic with very rare phonolites and trachytes (Haase et al., 2004). The Vogelsberg, the largest Cenozoic volcano in Germany (~2500 km^2), is situated just to the east of the Rhenish Massif, close to the triple junction of the Rhine, Ruhr and Leine Grabens. Here volcanic activity started in the Aquitanian (21 Ma) but the peak was around 17–16 Ma. The first products are rare phonolites and trachytes, followed by alkali basalts, nepheline basanites and very minor hawaiites, mugearites, latites and trachytes. Intercalated with these rocks are olivine to quartz tholeiites (Wedepohl et al., 1994; Jung and Masberg, 1998; Bogaard and Wörner, 2003). Late Cretaceous (~69–68 Ma) camptonitic lamprophyres found in drill cores in the southern Vogelsberg (Bogaard and Wörner, 2003) are similar to late Cretaceous ultramafic alkaline lamprophyre dykes in Lusatia (easternmost Germany; Renno et al., 2003), the late Cretaceous alkaline province of Iberia (characterized by the presence of camptonitic and monchiquitic dykes; Azambre et al., 1992; Rossy et al., 1992), the late Cretaceous alkaline lamprophyres (camptonites and monchiquites) of the Taourirt Massif, Morocco (~67 Ma; Wagner et al., 2003), the late Cretaceous (~71–68 Ma) monchiquitic dykes of the NE Alps (Calceranica and Corvara–Val Badia; Lucchini et al., 1983; Galassi et al., 1994), the Eocene camptonites of southern Sardinia (Maccioni and Marchi, 1994), and of the French Pyrenees (Les Corbieres, South France; Vitrac-Michard et al., 1977), north-eastern Spain (San Feliu; Velde and Turnon, 1970), and southern France (Languedoc; Liotard et al., 1991).

In the Hessian Depression, alkali basalts, limburgites, olivine nephelinites and quartz tholeiites occur (Kramm and Wedepohl, 1990). The Rhön and Heldburg volcanic zones cover an area of ~1500 km^2 (Jung, 1995); the main rock types are olivine nephelinites, nepheline basanites, tephrites and alkali basalts, plus rarer phonolites, trachytes, benmoreites and quartz-normative tholeiitic basalts (Wedepohl et al., 1994; Jung, 1995). The Upper Palatinate district is located at the SW end of the Ohre Rift in the Bohemian Massif and its description is, therefore, deferred to Section 4.1. The Urach–Hegau volcanic district comprises a bimodal olivine melilitite–phonolite suite, including olivine melilite nephelinite, phonolite and carbonatite (Wedepohl et al., 1994; Wilson et al., 1995b; Hegner and Vennemann, 1997). Kaiserstuhl is located within the upper (southern) Rhine Graben and is an alkaline carbonatite complex made up of carbonatites, tephrites, olivine nephelinites and phonolites (Hubberten et al., 1988; Schleicher et al., 1990, 1991).

The Cenozoic volcanic rocks of Germany exhibit a large range in initial $^{87}Sr/^{86}Sr$ (0.7027–0.7093) and $^{143}Nd/^{144}Nd$ (0.51300–0.51252); the most radiogenic Sr and unradiogenic Nd isotope compositions have been related to crustal contamination. The most mafic samples plot between EAR-like and Bulk Earth compositions (Fig. 9) with the Quaternary potassic Eifel lavas having the most radiogenic Sr-isotope compositions. The rare quartz tholeiitic rocks are characterized by less radiogenic Nd and slightly more radiogenic Sr isotopic compositions compared to the sodic alkaline basic rocks. Pb isotopic ratios span a relatively wide range ($^{206}Pb/^{204}Pb$ = 18.49–19.97, $^{207}Pb/^{204}Pb$ = 15.27–15.70 and $^{208}Pb/^{204}Pb$ = 37.68–39.81; $\Delta 7/4 = -1.0$ to $+12.3$; $\Delta 8/4 = -15.1$ to $+84.9$), consistent with the involvement of both DMM, HIMU and EMI-like mantle source components in the petrogenesis of the most primitive magmas.

The geochemical characteristics of the quartz tholeiites (including the correlation of Sr–Nd isotopic compositions with bulk-rock compositional parameters such as MgO, SiO_2 and Nb contents) has been related to magma chamber contamination at lower crustal depths via AFC processes (e.g., Kramm and Wedepohl, 1990; Jung and Masberg, 1998; Wedepohl, 2000; Haase et al., 2004). Their Pb isotopic ratios indicate the involvement of distinct mantle source components (DM, HIMU, EMI) coupled with some crustal contamination at shallow levels (Wörner et al., 1986; Hegner et al., 1995; Wedepohl and Baumann, 1999; Wedepohl, 2000; Bogaard and Wörner, 2003; Haase et al., 2004; Fig. 9).

The existence of a mantle plume beneath northern Germany (Rhenish Massif) has been inferred by analogy with the Cenozoic igneous activity of the French Massif Central (e.g., Granet et al., 1995; Wedepohl, 2000), where tomographic studies have indicated the presence of a finger-like low velocity anomaly extending down to Transition Zone depths (410–660 km; Granet et al, 1995; Sobolev et al., 1997; Wilson and Patterson, 2001). Raikes and Bonjer (1983) had previously identified a slow P-wave velocity anomaly (~3%) beneath the Eifel between 50 and 200 km depth using teleseismic tomography. Ritter et al. (2001) provided the first detailed images of this structure based on a high-resolution local seismic tomographic experiment. The top of the velocity anomaly is well-constrained at a depth of 50–60 km (Ritter, in press), corresponding to the base of the lithosphere, while the 410 km seismic discontinuity is depressed by ~20 km, consistent with a thermal anomaly upwelling from the Transition Zone. Additional supporting evidence for an Eifel plume is provided by ~250 m of dynamic uplift during the Quaternary. Ritter (in press) has attributed the magnitude of the low velocity anomaly to a mantle excess temperature of ~100–200 K combined with ~1% partial melt, similar to estimates based on seismic data for the Massif Central (~150 K; Sobolev et al., 1997), but much lower than those for the Hawaiian plume (~250–300 K; Li et al., 2000). The presence of water (in the form of hydrous minerals such as amphibole or phlogopite) in the upwelling mantle may also contribute to the velocity reduction. An apparent "hole" in the shear wave velocity anomaly at ~200 km depth could be caused by the onset of partial melting (Ritter, in press). Farnetani and Samuel (2005) have shown by numerical simulations, that such upper mantle plume-like structures are a feasible form of upper mantle convection. Also Jung et al. (2006) proposed the involvement of metasomatic fluids or melts from a rising diapir or plume for the Hocheifel. However in this case the low mantle potential temperatures inferred (<1200 °C) suggest that the mantle plume is not thermally anomalous.

The generation of the most primitive mafic magmas has been related to adiabatic decompression-induced partial melting of hotter than ambient asthenosphere (Wedepohl et al., 1994). A number of different models have been proposed to explain the presence of a thermal anomaly in the upper mantle, mostly involving some form of mantle plume. Wedepohl and Baumann (1999) proposed the upwelling of an almost stationary, ultrafast, mantle plume to explain the Eocene–Quaternary volcanic activity in central Germany, and discounted any major contribution to the magmatism from the lithospheric mantle. According to Goes et al. (1999), a mantle plume beneath central Europe can be traced in tomographic images down to lower mantle depths (~2000 km). The eastward movement of the European plate (as consequence of the opening of the Atlantic Ocean) should, however, have resulted in an age-progressive hot spot track with the older magmatism confined to the easternmost sectors (Heldburg and Rhön) and the youngest in the west (Eifel; Wedepohl and Baumann, 1999). This is clearly inconsistent with the ages of the volcanic fields, which suggests a more complex geodynamic setting. Fekiacova et al. (in press-b) have estimated that at ~40 Ma the Hocheifel volcanic field was ~1000 km SW of its current geographic position, making a geodynamic link with the present-day plume structure beneath the Eifel impossible. Wilson et al. (1995b) also proposed the indirect involvement of a mantle plume to explain the petrogenesis of melilitites in the Urach and Hegau volcanic fields of southern Germany; their model involved partial melting of a metasomatised thermal boundary layer (TBL) at the base of the lithosphere which has been enriched by infiltration of partial melts from an isotopically distinct mantle plume source with HIMU geochemical characteristics.

Anderson (2005) has argued against the presence of a mantle plume below the Eifel, rooted either at the core/mantle boundary (~2900 km in depth) or in the Transition Zone (410–660 km), favouring a shallow (plate tectonic) explanation for the magmatism. Similarly, Bogaard and Wörner (2003) discounted the possibility that the overall HIMU-OIB character of the Vogelsberg volcanic rocks could be related to a large, deep mantle plume (as proposed, among others, by Hoernle et al., 1995 and Wedepohl and Baumann, 1999), or to small plumes or diapirs from a shallow asthenospheric mantle reservoir (e.g., Wilson and Patterson, 2001). Both Jung and Masberg (1998) and Bogaard and Wörner (2003) favour passive upwelling of the asthenosphere as the main cause of magma generation. Similarly, Fekiacova et al. (in press-b) excluded any genetic relationship between the Eocene (~44–35 Ma) and Quaternary volcanic activity in the Eifel area, and related the origin of the Tertiary Hocheifel volcanic field to extension-induced decompression associated with Rhine Graben formation. Haase et al. (2004) modeled the petrogenesis of the Westerwald magmas in terms of partial melting of the thermal boundary layer at the base of the lithospheric mantle, as a consequence of adiabatic decompression during lithospheric thinning. Additionally, the lack of Tertiary uplift in the Vogelsberg and Hessian Depression areas does not support the presence of a mantle plume, which would be expected to induce lithosphere doming.

Recent noble gas (Ne–He–Ar) isotope studies of spinel peridotite facies mantle xenoliths from the Eifel (Gautheron et al., 2005; Buikin et al., 2005) have reached totally opposite conclusions concerning the existence of a deep mantle plume beneath the Rhenish Massif. While both sets of authors agree that the remarkably constant He isotope composition of the xenoliths (R/R_A ~6) could be a consequence of fluid/melt migration in areas of Tertiary–Quaternary volcanism, only Buikin et al. (2005) find any support for a lower mantle rare gas component (primarily Ar).

4. Central CiMACI province

The central sector of the CiMACI province experienced igneous activity in the Bohemian Massif (e.g., Ulrych et al., 2002), the Carpathian–Pannonian region (Austria, Poland, Slovakia, Romania, Hungary, Ukraine; Embey-Isztin et al., 1993; Harangi et al., 2001; Seghedi et al., 2001, 2004a,b, 2005), north Africa (Libya; Conticelli et al., 1995; Woolley, 2001; Lustrino, 2005b), Serbia–Macedonia (e.g., Balen and Pamic, 2001; Boev and Yanev, 2001; Altherr et al., 2004; Cvetkovic et al., 2004) and Greece (W Thrace; e.g., Pe-Piper and Piper, 2001; Eleftheriadis et al., 2003; Perugini et al., 2004). In the last case, only subduction-related igneous activity is recorded, therefore Greece is excluded from further discussion in this review.

4.1. Bohemian Massif

Magmatism in the Bohemian Massif (Czech Republic, Poland and eastern Germany) represents the easternmost extension of the Central European Volcanic Province (e.g., Wilson and Downes, 1991; Bendl et al., 1993; Ulrych et al., 2002, 2003). Magmatic activity occurred predominantly within and on the flanks of a major NE-trending rift zone (the Ohre Rift, known also as the Eger Graben), and the NW-trending Labe Volcano–Tectonic Zone. These rifts are part of the Cenozoic rift system which developed throughout central Europe from France (Limagne Graben) to Germany and Bohemia in the northern foreland of the Alps (e.g., Dèzes et al., 2004; Fig. 1). The K–Ar ages of igneous rocks emplaced in the Ohre Rift range from late Cretaceous to Quaternary (~77–0.26 Ma); however the peak of volcanic activity is mostly Eocene to Miocene (~42–16 Ma; Ulrych et al., 2002, 2003).

The main rock types are sodic alkaline SiO_2-undersaturated types. These include primitive alkali olivine basalts, basanites, nephelinites, ankaratrites, melilitites and more differentiated compositions such as tephriphonolites, phonolites and trachytes (e.g., Alibert et al., 1983; Bendl et al., 1993; Vaneckova et al., 1993; Ulrych et al., 2002; Fig. 10). Mantle xenoliths are relatively abundant in the most primitive mafic rock types (Blusztajn and Shimizu, 1994; Ulrych et al., 2000c; Knittel, 2003).

The major and trace element characteristics of the magmatic rocks are similar to those of alkaline magmas emplaced in intra-plate tectonic settings worldwide, with a distinct *anorogenic* geochemical signature. With the exception of one sample, initial $^{87}Sr/^{86}Sr$ isotopic ratios range from 0.7031 to 0.7047 and include some of the most depleted compositions in the CiMACI province; initial $^{143}Nd/^{144}Nd$ ranges from 0.51301 to 0.51267 (Alibert et al., 1983, 1987; Blusztajn and Hart, 1989; Bendl et al., 1993; Ulrych et al., 2002; Fig. 10). Pb isotope data are available only for Lower Silesia

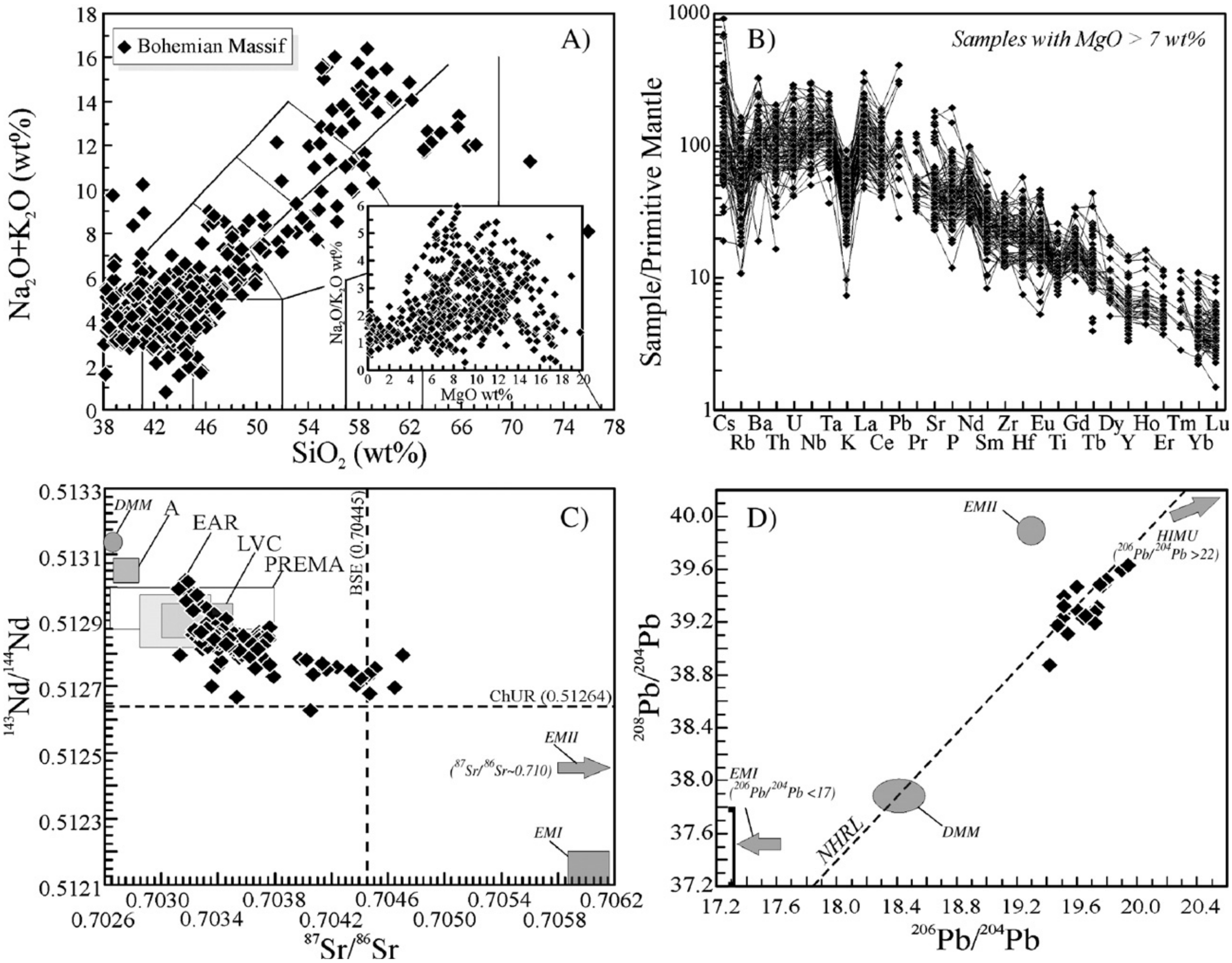


Fig. 10. A) Total alkali *vs.* silica diagram (TAS; le Maitre, 2002); B) primitive mantle (Sun and McDonough, 1989) normalised incompatible element diagram; C) $^{143}Nd/^{144}Nd$ *vs.* $^{87}Sr/^{86}Sr$ diagram; D) $^{208}Pb/^{204}Pb$ *vs.* $^{206}Pb/^{204}Pb$ diagram for igneous rocks from Bohemian Massif. Data sources: Wimmenauer, 1964; Shrbeny, 1979, 1980, 1982; Alibert et al., 1983, 1987; Ulrych et al., 1988, 1998, 2000a,b, 2002, 2003; Blusztajn and Hart, 1989; Bendl et al., 1993; Vaneckova et al., 1993; Wilson et al., 1995b; Ulrych and Pivec, 1997; Pivec et al., 2003, 2004; Ulrych, 1998; Randa et al., 2003; this study. Abbreviations as in Fig. 2.

(Poland) basalts (Blusztajn and Hart, 1989) and indicate a strong HIMU affinity ($^{206}Pb/^{204}Pb$ = 19.42–19.94; $^{207}Pb/^{204}Pb$ = 15.62–15.68; $^{208}Pb/^{204}Pb$ = 38.87–39.63; $\Delta7/4 = -1.2$ to $+7.1$; $\Delta8/4 = -27.6$ to $+16.9$). The trace element characteristics of the most primitive mafic magmas are also similar to those of HIMU-OIB.

The lower $^{87}Sr/^{86}Sr$ and higher $^{143}Nd/^{144}Nd$ isotopic ratios of the Bohemian Massif mafic lavas, compared to those of other Cenozoic European mafic volcanic rocks from France and Germany, suggest that in this part of the CiMACI province the magmas may be predominantly sampling an end-member mantle component. In this context, the mantle beneath the Bohemian Massif and Lower Silesia may be less enriched than that beneath the French Massif Central and northern Germany. The HIMU-like characteristics of the mafic lavas have been attributed to recycling of subducted oceanic crust in the upper mantle (e.g., Alibert et al., 1987). The trace element and isotopic similarity of the Cenozoic Bohemian Massif lavas with Permo-Carboniferous volcanic rocks from the same area led Ulrych et al. (2002) to propose that the HIMU-like source already existed in Permian times and was generated by Devonian subduction-related metasomatism of the mantle lithosphere.

4.2. Pannonian Basin

The geodynamic evolution of the Carpathian–Pannonian Basin region, in the central-eastern sector of the CiMACI province shows some similarities with that of the central and western Mediterranean (Fig. 1). It is

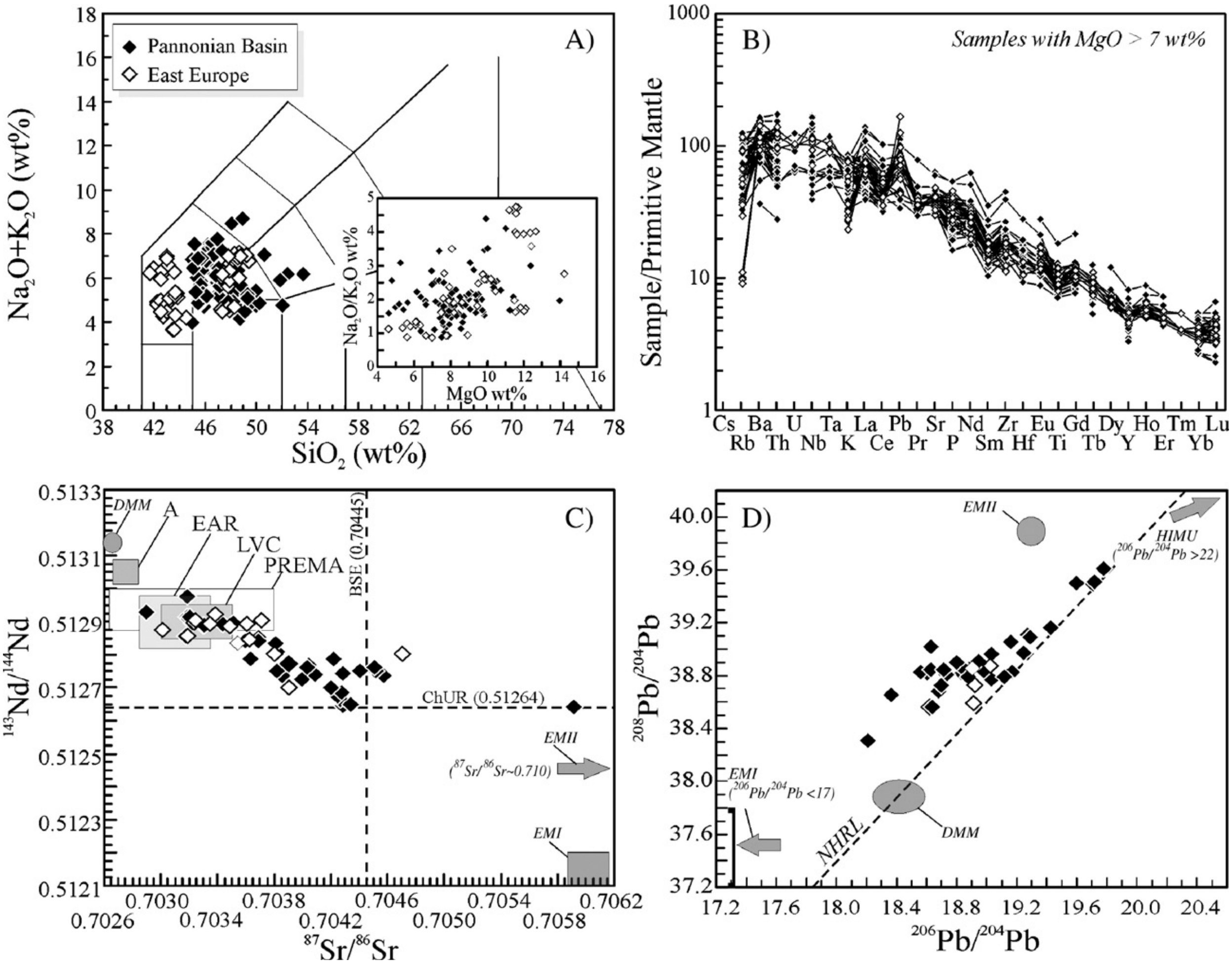


Fig. 11. A) Total alkali *vs.* silica diagram (TAS; le Maitre, 2002); B) primitive mantle (Sun and McDonough, 1989) normalised incompatible element diagram; C) $^{143}Nd/^{144}Nd$ *vs.* $^{87}Sr/^{86}Sr$ diagram; D) $^{208}Pb/^{204}Pb$ *vs.* $^{206}Pb/^{204}Pb$ diagram for igneous rocks from Pannonian Basin (black diamonds) and East Europe (Serbia, Bulgaria and Macedonia; white diamonds). Data sources: Pannonian Basin (Salters et al., 1988; Embey-Isztin et al., 1993; Ivan and Hovorka, 1993; Harangi et al., 1994, 1995a,b; Dobosi et al., 1995; Downes et al., 1995a,b; Rosenbaum et al., 1997; Seghedi et al., 2004b); East Europe (Marchev et al., 1992, 1997, 1998b; Jovanovic et al., 2001; Cvetkovic et al., 2004). Abbreviations as in Fig. 2.

characterized by an arcuate orogenic belt (the Carpathian arc) surrounding a Neogene extensional basin (Pannonian Basin) which has thin lithosphere (50–80 km) and high heat flow (Harangi and Lenkey, 2007, and references therein). The Pannonian Basin is commonly described as a back-arc basin, although the tectonic setting is actually considerably more complex. The geodynamic setting of the Carpathian–Pannonian Basin region can also be compared to that of the Apennine Chain in Italy (Fig. 1). The only difference is that the resultant back-arc basin reached an oceanisation stage in the Tyrrhenian Sea (Vavilov and Marsili oceanic basins) whereas it stopped at a thinned continental crustal stage in the Pannonian Basin (e.g., Doglioni et al., 1998; Seghedi et al., 2004a). The presence of dismembered continental slices in between the back-arc basins (the lithospheric boudinage process proposed by Gueguen et al., 1998) is another common feature of the two areas (e.g., Carminati and Doglioni, 2004). In this context, the Sardinia–Corsica continental block finds its counterpart in the Apuseni Mts. in the Pannonian Basin. The main phase of extension in the Pannonian Basin occurred from 17 to 11 Ma (Horvath, 1993) concurrent with continental collision in the northern Carpathians. Active subduction beneath most of the the Eastern Carpathians ceased during the Late Miocene (~ 11–10 Ma).

The inner part of the Carpathian arc is characterized by a belt of syn- and post-collisional calc-alkaline volcanic complexes (16.5–0.02 Ma; Downes et al., 1995a,b) with *orogenic* geochemical characteristics. Alkaline basaltic volcanism s.l. of *anorogenic* affinity (including nephelinites, basanites, alkali basalts and trachybasalts; Szabò et al., 1992; Embey-Isztin et al., 1993; Seghedi et al., 2004a,b, 2005) subsequently occurred throughout the region; this generally postdates the *orogenic* volcanism in the inner Carpathian Arc and the Pannonian Basin with the exception of the Persani Mountains in the SE Carpathians where *orogenic* and *anorogenic* volcanism are contemporaneous (Seghedi et al., 2004b). Harangi (2001a) has also recognised a trachyandesite–trachyte volcanic complex with minor alkali basalts beneath the sedimentary fill of the Little Hungarian Plain.

The earliest mafic alkaline volcanism (*anorogenic*) commenced at ~11 Ma in the western Pannonian Basin during the late stages of extension in an area in which the lithosphere was not significantly thinned (Harangi and Lenkey, 2007). The main phase of alkaline mafic magmatism lasted from ~7 Ma to 4–2 Ma, forming a series of monogenetic volcanic fields; the peak of the alkaline basaltic volcanism occurred during the Pliocene (~5–3 Ma). There appears to have been a hiatus in magmatic activity from ~10 to 7 Ma during a regional compression event (Fodor et al., 1999). The basalts s.l. have high Mg-numbers (>0.62) and often contain mantle xenoliths, suggesting minimal differentiation *en route* to the surface (e.g., Embey-Isztin et al., 1989; Downes et al., 1992; Downes and Vaselli, 1995; Vaselli et al., 1995; Bali et al., 2002; Szabò et al., 2004). The onset of *anorogenic* volcanism is broadly coeval with a change in the geochemical characteristics of the contemporaneous calc-alkaline volcanism in the Carpathian arc.

The *anorogenic* magmatism of the Pannonian Basin occurred within an overall orogenic tectonic setting, even though it is locally associated with extensional tectonics (e.g., Embey-Isztin et al., 1993). A detailed discussion of the relationship between extensional tectonics and the alkaline mafic magmatism can be found in Harangi and Lenkey (2007). The *anorogenic* volcanism clearly post-dates the main syn-rift extensional stage of the Pannonian Basin in the Middle Miocene (~17–13 Ma; Horvath, 1993; Embey-Isztin and Dobosi, 1995; Harangi, 2001b).

The *anorogenic* igneous rocks show some geochemical characteristics that deviate from those of magmas emplaced in typical intra-plate tectonic settings (Embey-Isztin et al., 1993; Embey-Isztin and Dobosi, 1995; Downes et al., 1995a). These "anomalies" have been related to varying degrees of *orogenic* heritage within the mantle source of the magmas, related to previous subduction cycles (e.g., Little Hungarian Plain and Persani Mts.; Harangi et al., 1994, 1995a; Downes et al., 1995a). A spectrum of compositions exists between true *anorogenic* magma types (i.e., without any evidence of source contamination by previous subduction processes) and *transitional–anorogenic* types (i.e., compositions that resemble typical HIMU-OIB compositions but with some evidence of source modification by subduction-related fluids).

Reviews of the spatial and temporal distribution of the Neogene–Quaternary volcanism can be found in Szabò et al. (1992), Pécskay et al. (1995), Seghedi et al. (2004a, 2004b, 2005) and Harangi and Lenkey (2007). From W to E the most important volcanic districts are: Miocene–Pliocene (~11.5–1.7 Ma) igneous activity in the Styrian Basin and Burgenland, including alkali basalts, nepheline basanites and tephrites (e.g., Embey-Isztin et al., 1993; Embey-Isztin and Dobosi, 1995; Dobosi et al., 1998); Miocene–Pliocene (~7.5–2.3 Ma) pyroclastic rocks in the Trans-Danubian Central Range (Balaton area; Embey-Isztin et al., 1993; Dobosi et al., 1998); late Miocene (~11–10 Ma) trachyandesites and alkali trachytes with minor alkali basalts and Pliocene (~5–3 Ma) alkali basalts in the Danube Basin (Little Hungarian Plain; Embey-Isztin et al., 1993; Harangi et al., 1994, 1995a; Harangi, 2001a); Miocene (8.0–6.8 Ma) and rare Quaternary (~0.5 Ma) nepheline basanites and alkali basalts from Stiavnica stratovolcano (central Slovakia; Ivan and Hovorka, 1993; Dobosi et al., 1995, 1998); late Miocene (~11–10 Ma) alkali basalts from Kecel (Pécskay et al., 1995) and late Miocene–Pleistocene (7.7–0.5 Ma) alkali basalt volcanism in southern Slovakia–northern Hungary (Nograd–Gomor; Ivan and Hovorka, 1993; Dobosi et al., 1995, 1998; Konecny et al., 1995); upper Pliocene (~2 Ma) olivine–leucitites of Bar (Harangi et al., 1995b); Pliocene (~2.5 Ma) alkali basaltic and hawaiitic rocks of Banat, west of the Apuseni Mts. (Downes et al., 1995a; Dobosi et al., 1998) and Quaternary (~1.2–0.6 Ma; Panaiotu et al., 2004) alkali basalts and hawaiites of the Persani Mts. in the SW of the Transylvanian Basin (Downes et al., 1995a). Predating this phase of Miocene–Pleistocene igneous activity, with mostly Na-alkaline characteristics, are lower Eocene (~58–48 Ma) small-volume intrusions of mafic alkaline magmas (basanites) at Poiana Rusca (Romania; Downes et al., 1995b).

The Pannonian Basin volcanic rocks exhibit a relatively large range of radiogenic isotope compositions; with the exception of one sample, $^{87}Sr/^{86}Sr$ ranges from 0.7029 to 0.7046, $^{143}Nd/^{144}Nd$ from 0.51298 to 0.51264, $^{206}Pb/^{204}Pb$ from 18.21 to 19.71, $^{207}Pb/^{204}Pb$ from 15.57 to 15.67, $^{208}Pb/^{204}Pb$ from 38.31 to 39.61; $\Delta 7/4$ from +1.0 to +16.7; $\Delta 8/4$ from +4.2 to +86.5; Fig. 11, ranging from EAR-like to Bulk Earth values. This heterogeneity might reflect minor crustal contamination of some of the magmas *en route* to the surface, in addition to mixing between asthenosphere-derived magmas and partial melts of enriched lithospheric mantle variably metasomatised during previous episodes of subduction.

The Pannonian Basin *anorogenic* igneous rocks are well characterized from a geochemical point of view, and several models have been proposed to explain their petrogenesis. There is a consensus of opinion that most of the primary magmas originated from an asthenospheric mantle source and were then partially modified as they passed through the lithosphere (both the lithospheric mantle and the lower crust; e.g., Embey-Isztin et al., 1993; Downes et al., 1995a; Embey-Isztin and Dobosi, 1995; Seghedi et al., 2004b). Distinctive geochemical characteristics (in particular, high $^{207}Pb/^{204}Pb$ for a given $^{206}Pb/^{204}Pb$) in a few areas (e.g., Lake Balaton, Little Hungarian Plain and Persani Mts.), compared to most of the Pannonian Basin, have been linked to metasomatic modification of the mantle source of the magmas by subduction-related fluids (e.g., Embey-Isztin et al., 1993; Downes et al., 1995a; Girbacea and Frisch, 1998). Perhaps one of the most contentious issues is the role, if any, of a mantle plume in the petrogenesis of the magmas. Seghedi et al. (2004b) proposed that mantle partial melting was triggered by the upwelling of several finger-like upper mantle plumes, similar to those imaged by seismic tomography beneath the French Massif Central and the Eifel district of Germany (see Sections 3.5 and 3.7). Harangi and Lenkey (2007), however, rule out mantle plume involvement based on the small volumes and scattered occurrence of the *anorogenic* volcanism, the lack of evidence for basement uplift and the presence of a high-velocity region (inferred to be cold subducted material at the base of the upper mantle; Piromallo et al., 2001) beneath the Pannonian Basin.

4.3. Libya

Relatively little is known about the Cenozoic magmatism of Libya (Fig. 1). Volcanic rocks cover a total area of more than 65,000 km^2 (Woller and Fediuk, 1980; Busrewil and Esson, 1991; Busrewil and Oun, 1991; Conticelli et al., 1995; Woolley, 2001). The main volcanic fields are Gharyan, Jabal Fezzan, Jabal as Sawda, Al Haruj al Aswad and Jabal Nuquay. With the exception of a few studies (e.g., Conticelli et al., 1995) on the igneous rocks in SE Libya (Uwaynat area), near the border with Egypt and Sudan, there are no modern petrological studies. The northernmost outcrop is the Gharyan area (~3000 km^2); this is an extensive Eocene–Quaternary (~55–1 Ma) volcanic field composed mostly of plateau lavas (transitional basalts and andesitic basalts; ~55–50 Ma) plus much rarer mafic alkaline lavas (basanites), intermediate alkaline lavas (hawaiites, mugearites; 12–1 Ma) and volumetrically insignificant differentiated products (phonolites; ~40 Ma; Piccoli and Spadea, 1964; Piccoli, 1970; Almond, 1974; Almond et al., 1974; Bausch, 1978; Busrewil and Wadsworth, 1980b, Lustrino, 2005b; Fig 4). The only isotopic data available are those from Uwaynat; initial Sr isotopic ratios of ~28–26 Ma old basic rocks (olivine mela-nephelinite, basanites and alkali basalts) and carbonatites range from 0.7032 to 0.7037, while $^{143}Nd/^{144}Nd$ ranges from 0.51284 to 0.51276 (Conticelli et al., 1995; Fig. 4). Andre et al. (1991) reported $^{87}Sr/^{86}Sr$ data for the most differentiated compositions of the same complex (syenites, alkali rhyolites, trachytes, alkali feldspar granites) dated at ~61–43 Ma.

Most of the publications dealing with the Cenozoic magmatism of Libya are purely descriptive, focussing on the petrography and geochemistry. Only one paper presents a petrogenetic model, that of Conticelli et al. (1995) for the Upper Oligocene igneous rocks of the Uwaynat area. These authors relate the Uwayanat basic and ultrabasic magmatism to the north-eastward migration of the African plate above an intra-continental hotspot.

4.4. Serbia

Anorogenic igneous activity initiated in Paleocene–Eocene times in SW Romania and E Serbia, following a late Cretaceous–Paleocene phase of subduction-related, calc-alkaline igneous activity (e.g., Karamata et al., 1997; Clark and Ulrich, 2004; Fig. 11) related to the final closure of the Vardar Ocean by subduction beneath the southern European active continental margin (e.g., Dimitrijevic, 1997, 2001). The magmatism is of basic-ultrabasic Na-alkaline type (mostly basanites and tephrites; Fig. 11) very similar to that of Poiana Rusca in the Pannonian Basin (Downes et al., 1995b). Following this volumetrically insignificant phase of *anorogenic* igneous activity, a new phase of magmatic activity with high-K calc-alkaline, shoshonitic and ultrapotassic characteristics developed in the same area during the Oligo-Miocene and Pliocene (Prelevic et al., 2005). This Oligocene–Pliocene episode postdates the most recent subduction beneath the region by ~30 Myr, eliminating the possibility that it is directly related to active subduction (Cvetkovic et al., 2004; Prelevic et al, 2005). Similar potassic to ultrapotassic igneous activity (Oligo-Miocene and Pliocene in age), outropping a few km to the southeast in Macedonia, was considered to have a “within-plate” signature by Boev and Yanev (2001). Sr and Nd isotopic data are available for the

Serbian volcanic rocks; these indicate derivation of the magmas from an isotopically heterogeneous mantle source ($^{87}Sr/^{86}Sr$ = 0.7030–0.7047; $^{143}Nd/^{144}Nd$ = 0.51287–0.51270; Jovanovic et al., 2001; Cvetkovic et al., 2004).

The Paleocene–Eocene *anorogenic* volcanic rocks of Serbia show no geochemical evidence of derivation from mantle sources metasomatised during the previous phase of subduction, and have been related to asthenospheric mantle sources without any crustal interaction (Jovanovic et al., 2001), or to partial melting of the deepest part of the lithosphere (TBL; Cvetkovic et al., 2004). The latter authors relate magma generation to passive lithospheric extension after the closure of the Vardar Ocean, and do not consider any mantle plume involvement likely. The Oligo-Miocene and Pliocene potassic and ultrapotassic volcanics have many similarities to those of the Roman Volcanic Province in Italy (Section 3.6). Prelevic et al. (2005) favour a lithospheric origin for the parental magmas linked to collapse of the Dinaride orogen.

5. Eastern CiMACI province

Substantial amounts of Cenozoic volcanic activity have occurred in the eastern part of the CiMACI province in Bulgaria (e.g., Marchev et al., 1998a,b, 2004), Turkey (e.g., Pearce et al., 1990; Wilson et al., 1997; Keskin et al., 1998; Aldanmaz et al., 2000; Alici et al., 2002; Keskin, 2003; Agostini et al., 2007), Jordan (e.g., Bertrand et al., 2003; Shaw et al., 2003), Israel (Stein and Hofmann, 1992; Weinstein et al., 2006) and Syria (Mouty et al., 1992; Sharkov et al., 1994; Krienitz et al., 2006; Lustrino and Sharkov, 2006).

5.1. Bulgaria–Macedonia

During the Oligocene, extensive igneous activity occurred in the Rhodope Massif, mostly concentrated in Bugaria and Macedonia (Boev and Yanev, 2001; Marchev et al., 2004) with both *orogenic* and *anorogenic* geochemical characteristics. This igneous activity can be traced northeastward through Serbia, to Austria and the peri-Adriatic lineament. An early Oligocene *orogenic* (calc-alkaline and shoshonitic) igneous episode (~33–29 Ma) developed in response to the late Cretaceous–Paleogene collision of Africa and Eurasia, following the closure of the Tethys Ocean. This *orogenic* igneous activity (mostly volcanic with few high-level intrusions) started at the Eocene–Oligocene boundary (~33 Ma) with the production of high-Ba trachybasalts, followed by absarokites, shoshonites, calc-alkaline and high-K calc-alkaline basalts (Marchev et al., 2004). Following this episode, in the S Rhodope Massif (Bulgaria) there are remnants of upper Oligocene (~28–26 Ma) igneous activity characterized by rare basic alkaline products (basanites, trachybasalts, alkali basalts, lamprophyres with camptonitic affinity; e.g., Kotopouli and Pe-Piper, 1989; Marchev et al., 1998a,b, 2004; Fig. 11). These scarce outcrops occur close to the town of Krumovgrad, near the border with Greece. The Na_2O/K_2O ratio of these volcanic rocks varies considerably in the different localities (0.9–2.3) and primitive mantle-normalised trace element diagrams exhibit strong positive Nb–Ta and Ti peaks. There is little published isotopic data for the *anorogenic* rocks; $^{87}Sr/^{86}Sr$ is low and ranges from 0.7032 to 0.7037; $^{143}Nd/^{144}Nd$ is conversely high (0.51290–0.51284), and Pb is midly radiogenic ($^{206}Pb/^{204}Pb$ = 18.91–19.02; $^{208}Pb/^{204}Pb$ = 38.59–38.87; $\Delta7/4 = -2.3$ to $+9.7$; $\Delta8/4 = +10.3$ to $+36.9$; Marchev et al., 1998a,b; Jovanovic et al., 2001; Cvetkovic et al., 2004 Fig. 11).

In north (Moesian Platform) and central (Balkanides–Srednogorie Zone) Bulgaria, Miocene alkaline volcanism has been reported by Marchev et al. (1992). This occurs in a narrow zone ~120 km long, oriented approximately NNE–SSW. In the Moesian Platform the Miocene rocks have been dated at ~23–16 Ma and are represented by lava domes, flows and dykes with basanitic composition (Marchev et al., 1992). These are fairly primitive mafic magmatic rocks (Mg# ranging from 66 to 72) which are strongly SiO_2-undersaturated (normative nepheline up to 15 wt.%). No isotopic data are available.

Cenozoic magmatism within Macedonia shows some similarities with the Bulgarian igneous activity. An *orogenic* igneous episode also occurred here during the Oligocene (~33–25 Ma) but continued in rare cases up to the Burdigalian (~16 Ma; Plavica locality) and late Miocene to late Pliocene (~6.5–1.8 Ma; Boev and Yanev, 2001). Boev and Yanev (2001) report the occurrence of "within-plate potassic mafic volcanism" during the upper Miocene–lower Pliocene (~9.5–5.5 Ma) in the Vardar zone. The most common lithologies are leucite–tephrites and lamprophyres, plus absarokites and shoshonites to latites. In trace element discrimination diagrams these rocks plot in the fields of within-plate basalts, but, unlike most of the CiMACI rocks, they are potassic to ultrapotassic in character.

The geochemical signatures of the upper Oligocene *anorogenic* igneous rocks of Bulgaria indicate limited or no modification of mantle sources by the previous subduction cycle (e.g., Marchev et al., 1998b, 2004). The mantle source shares many geochemical similarities

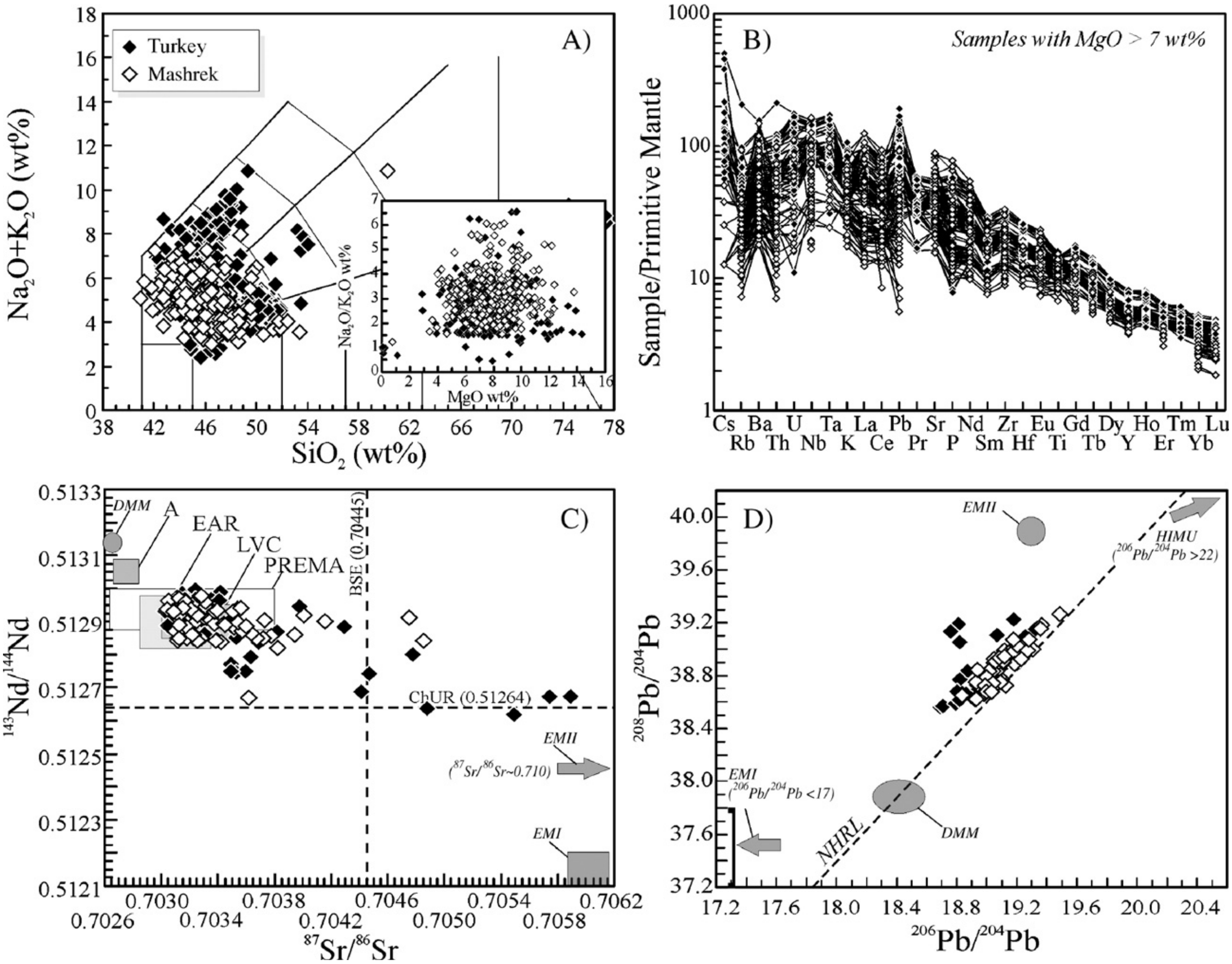


Fig. 12. A) Total alkali *vs.* silica diagram (TAS; le Maitre, 2002); B) primitive mantle (Sun and McDonough, 1989) normalised incompatible elements diagram; C) $^{143}Nd/^{144}Nd$ *vs.* $^{87}Sr/^{86}$7Sr diagram; D) $^{208}Pb/^{204}Pb$ *vs.* $^{206}Pb/^{204}Pb$ diagram for igneous rocks from Turkey (black diamonds) and Mashrek (Jordan, Israel, Syria, Lebanon and Palestine; white diamonds). Data sources: Turkey (Innocenti et al., 1982, 2005; Çapan et al., 1987; Pearce et al., 1990; Guleç, 1991; Aydar et al., 1995; Notsu et al., 1995; Parlak et al., 1997, 2001; Polat et al., 1997; Wilson et al., 1997; Genç, 1998; Yilmaz and Polat, 1998; Aldanmaz et al., 2000, 2006; Aldanmaz, 2002; Alici et al., 2001, 2002; Alici Sen et al., 2004; Agostini et al., 2007; this study); Mashrek (Barberi et al., 1979; Altherr et al., 1990; Mouty et al., 1992; Stein and Hofmann, 1992; Sharkov et al., 1994, 1996; Weinstein et al., 1994, 2006; Fediuk and Al Fugha, 1999; Bertrand et al., 2003; Ibrahim et al., 2003; Shaw et al., 2003; Abdel-Rahman and Nassar, 2004; Ibrahim and Al-Malabeh, 2006; Krienitz et al., 2006; Lustrino and Sharkov, 2006). Abbreviations as in Fig. 2.

with the EAR. Marchev et al. (2004) have proposed that magmatic activity was triggered by convective removal of the base of the lithosphere and subsequent mantle diapirism. According to Boev and Yanev (2001), the potassic–ultrapotassic "within-plate" Neogene rocks of Macedonia were derived from a metasomatised (phlogopite-rich) lithospheric mantle source which underwent partial melting as a consequence of asthenospheric upwelling.

5.2. Turkey

Turkey provides one of the best examples within the CiMACI province in which there is a clear transition from *orogenic* to *anorogenic* igneous activity (e.g., Innocenti et al., 1982, 2005; Pearce et al., 1990; Tonarini et al., 2005; Wilson et al., 1997; Tankut et al, 1998; Agostini et al., 2007). Here northward subduction of the Afro-Arabian plate beneath the Eurasian plate was followed by continent–continent collision during the middle–late Miocene (e.g., Yurur and Chorowicz, 1998; Carminati and Doglioni, 2004). There is a temporal succession from igneous rocks with subduction-related geochemical signatures (e.g., calc-alkaline to potassic and ultrapotassic compositions, radiogenic $^{87}Sr/^{86}Sr$, unradiogenic $^{143}Nd/^{144}Nd$, high LILE/HFSE ratios) followed by Na-alkaline (and rarer tholeiitic) volcanic rocks, some of which contain mantle xenoliths (e.g., Pearce et al., 1990; Yilmaz, 1990; Buket and Temel, 1998; Yilmaz and Polat, 1998; Aldanmaz, 2002; Aldanmaz et al., 2005, 2006; Alici et al., 2001, 2002, 2004; Keskin, 2003; Fig. 12). Crustal contamination has, in some cases, obscured the original mantle signature of the *anorogenic* basaltic magmas (e.g., quartz tholeiites have been interpreted as the products of lower crustal contamination of parental OIB-like Na-alkaline basaltic magmas; Alici et al., 2004).

A detailed treatment of the Cenozoic volcanic activity of Turkey is clearly beyond the scope of this paper. In particular, given the complex tectonic evolution of this area it would be necessary to make a distinction between the "anorogenic" volcanic activity in W Anatolia (e.g., Biga, Ezine, Kula) and that in the Thrace/Aegean area (e.g., Psathoura, Chios, Kalogeri, Samos, Patmos), central (e.g., Galatia) and E Anatolia (e.g., Karacadag) and the Karasu Valley, at the northernmost end of the Dead Sea Transform Fault. For simplicity, we have treated all the *anorogenic* volcanic rocks of Turkey as a single group.

Large volumes of trachyandesitic–dacitic lava flows and pyroclastics of Miocene age are associated with small volumes of alkali basalt lava flows in the Galatia volcanic province of northwest Central Anatolia (Wilson et al., 1997). The volcanism postdates continental collision, occurring in a transtensional tectonic setting associated with movement along the North Anatolian Fault zone. The alkali basalts were erupted during two distinct time periods in the early Miocene (19–17 Ma) and late Miocene (>10 Ma). These two groups of alkali basalts are inferred to have been derived from different mantle sources, based on their Sr–Nd isotope and geochemical characteristics. The late Miocene basalts were derived from a more depleted mantle source than the early Miocene basalts, which were generated by partial melting of an incompatible element enriched, subduction-modified, mantle source. The depleted source component was inferred by Wilson et al. (1997) to reside within the asthenosphere and to have some affinity with the source of HIMU oceanic island basalts.

A similar distinction between two main periods of igneous activity has been proposed by Agostini et al. (2007) for W Anatolia and the Aegean Sea/Thrace. According to these authors, the oldest products (late Miocene; Aliaga, Foça, Urla, Samos, Chios, Patmos and Psathoura districts in westernmost Anatolia and the Aegean Sea) exhibit a wide range of isotopic and trace element compositions, linked to the presence of a subduction-related component in their petrogenesis. On the other hand, the youngest products (Pliocene–Holocene; Kula, Biga, Kalogeri and Thrace) have Sr–Nd isotope and trace element characteristics more typical of *anorogenic* volcanism with no subduction-related fingerprint. The Quaternary alkaline volcanism of the Kula region in western Anatolia is closely associated with an extensional tectonic regime which developed during the Miocene and is still active (Alici et al., 2002; Innocenti et al., 2005). The magmatic rocks are typically sodic, SiO_2-undersaturated, and range in composition from basanite to tephrite and phonotephrite. They show significant enrichment in HSFE (e.g., Nb and Ta) and LILE, with high Nb/Y ratios (<3), typical of within-plate alkaline volcanic rocks. However, the Rb, Ba, Sr, Nb and Ta contents and Rb/Nb and K/Nb ratios of the Kula lavas are considerably higher than in typical OIB, suggesting some contribution from the lithospheric mantle in their petrogenesis. The Sr–Nd–Pb isotopic compositions of the lavas ($^{87}Sr/^{86}Sr$ = 0.7030–0.7035, $^{143}Nd/^{144}Nd$ = 0.51300–0.51277, $^{206}Pb/^{204}Pb$ = 18.69–19.06, $^{207}Pb/^{204}Pb$ = 15.61–15.68, $^{208}Pb/^{204}Pb$ = 38.56–39.11; Δ7/4 = +7.4 to +12.5; Δ8/4 = +16.6 to +43.8; Fig. 12) indicate a contribution from two mantle source components: a dominant HIMU-like asthenospheric component and a more enriched lithospheric mantle component (Seyitoglu et al., 1997; Alici et al., 2002). Crustal contamination does not appear to play a significant role in the petrogenesis of the more evolved magmas.

Quaternary basaltic volcanism in central and eastern Anatolia also exhibits transitional calc-alkaline to alkaline characteristics, with significant enrichment in LILE (Ba, Rb, Pb, Sr) relative to the HFSE (Nb, Ta; Innocenti et al., 1982; Pearce et al., 1990; Alici et al., 2004). Geodynamically, the magmatism is linked to extensional tectonics, and post-dates active subduction beneath the area. The temporal transition from *orogenic* to *anorogenic* volcanic activity has been related to slab-detachment processes (e.g., Innocenti et al., 1982) or to differential movement between the faster moving Greek lithosphere with respect to slower Cyprus–Anatolian lithosphere, producing a "window" in the slab (e.g., Doglioni et al., 2002; Agostini et al., 2007). The subduction-related geochemical characteristics are most likely inherited from mantle lithosphere modified by slab fluids released during northward subduction of the Afro-Arabian plate beneath the Eurasian plate during Eocene to Miocene times (e.g., Pearce et al., 1990).

The *anorogenic* rocks of Turkey, as a whole, exhibit a relatively small range of Sr–Nd isotopic ratios ($^{87}Sr/^{86}Sr$ = 0.7030–0.7059, $^{143}Nd/^{144}Nd$ = 0.51300–0.51262; Pearce, 1990; Guleç, 1991; Notsu et al., 1995; Aldanmaz et al., 2000; Alici et al., 2002; Fig. 12).

In addition to Kula, Pb isotopic data are available only for the Karasu Valley in SE Turkey ($^{206}Pb/^{204}Pb$ = 18.76–19.33, $^{207}Pb/^{204}Pb$ = 15.64–15.73, $^{208}Pb/^{204}Pb$ = 39.05–39.22; Δ7/4 = +6.2 to +19.5; Δ8/4 = +19.1 to +82.9), whose activity should be included in the Mashrek area, being emplaced along the northernmost sector of the Dead Sea Transform (see discussion in Lustrino and Sharkov, 2006). The Sr–Nd–Pb isotopic compositions are consistent with derivation of the magmas from an EAR-like mantle source. With the exception of the work of Erçan et al. (1991) and Pearce et al. (1990), nothing is known about the largest *anorogenic* shield volcano of the CiMACI (Karacadag volcano; ~8–10,000 km^2). The Pleistocene–Holocene volcanic rocks are exclusively basic-ultrabasic Na-alkaline volcanic rocks (mostly basanites and alkali basalts).

Polat et al. (1997) have discounted the involvement of a mantle plume in the petrogenesis of magmas erupted in the East Anatolian and Dead Sea Fault zones of southern Turkey, where alkali olivine basalts were erupted within transtensional depressions. Magma generation is instead related to low-degree partial melting of an enriched asthenospheric mantle source (phlogopite±amphibole-bearing garnet lherzolite), followed by melt migration through the lithosphere along strike-slip faults (Polat et al., 1997; Alici et al., 2001).

5.3. Mashrek (Jordan, Israel, Syria, Lebanon, Palestine)

An extensive late Eocene-late Pleistoscene (~40–0.05 Ma) intra-plate volcanic province occurs within the territories of Jordan, Israel, Syria, Lebanon and Palestine (Barberi et al., 1979; Ilani et al., 2001; Mouty et al., 1992; Sharkov et al., 1994; Bertrand et al., 2003; Shaw et al., 2003; Krienitz et al., 2006; Lustrino and Sharkov, 2006; Weinstein et al., 2006; Fig. 1); for simplicity in subsequent discussions this will be refered to as the Mashrek region. The magmatism occurs in a transtensional tectonic setting intimately linked with the development of the Red Sea rift system (e.g., Lyberis, 1988). Individual volcanic fields can be very large (e.g. Harrat Ash Shaam; ~50,000 km^2; Bertrand et al., 2003; Shaw et al., 2003; Weinstein et al., 2006) and mostly made up of flood basalts. About 95% of the published age determinations are <14 Ma, suggesting that most of the volcanic activity is confined to Miocene–Quaternary times (Ilani et al., 2001). The main rock-types are mostly sodic alkali basalts, basanites and hawaiites; more rarely tholeiitic basalts also occur (Barberi et al., 1979; Ibrahim et al., 2003; Shaw et al., 2003; Weinstein et al., 2006; Fig. 12). The most primitive mafic alkaline rocks often contain mantle xenoliths (e.g., Sharkov et al., 1996; Nasir and Safarjalani, 2000; Bilal and Touret, 2001).

Sr–Nd–Pb isotopic ratios exhibit a relatively narrow range: $^{87}Sr/^{86}Sr$ = 0.7031–0.7039; $^{143}Nd/^{144}Nd$ = 0.51298–0.51267; $^{206}Pb/^{204}Pb$ = 18.84–19.49; $^{207}Pb/^{204}Pb$ = 15.56–15.65; $^{208}Pb/^{204}Pb$ = 38.63–39.27; Δ7/4 = −1.3 to +10.3; Δ8/4 = −3.2 to +31.6 (Altherr et al., 1990; Stein and Hofmann, 1992; Bertrand et al., 2003; Shaw et al., 2003; Lustrino and Sharkov, 2006; Weinstein et al., 2006), falling largely within the EAR field (Fig. 12). Krienitz et al. (2006) report highly radiogenic Sr isotopic ratios (up to 0.7049) for the Syrian basalts and attributed these values to the effects of up to 25% upper crustal contamination of primitive magmas. We believe that these conclusions are improbable and consider that the strongly radiogenic Sr isotope compositions are the effects of alteration of the samples, clearly evidenced by the petrographic study of the same authors. It is important to note that similar samples studied by Lustrino and Sharkov (2006) gave similar radiogenic Sr values (up to 0.7048) before acid leaching which were reduced to 0.7039 after leaching. The Sr–Nd–Pb isotopic variations of the Jordanian basalts have been explained by Shaw et al. (2003) in terms of polybaric melting of a heterogeneous mantle source. On the basis of incompatible trace element and isotopic differences between the intra-plate volcanism of Jordan and Yemen, further to the south, Bertrand et al. (2003) and Shaw et al. (2003) proposed only a minor role, if any, for the Afar mantle plume in the genesis of the Harrat Ash Shaam volcanic field, relating the Mashrek volcanic activity to adiabatic partial melting of both asthenospheric and lithospheric mantle as consequence of lithospheric extension. Weinstein et al. (2006) suggested a lithospheric mantle origin for the north-westernmost Harrat Ash Shaam volcanic field, unrelated to deep mantle plumes. On the other hand, Stein and Hofmann (1992) proposed that the distinctive HIMU-like isotopic and trace element characterstics of the Israeli basalts (e.g. Nb/U = 43±9; Ce/Pb = 26±6; $^{87}Sr/^{86}Sr$ = 0.7029–0.7033, ϵ_{Nd} = +3.9 to +5.9; $^{206}Pb/^{204}Pb$ = 18.88–19.99; $^{207}Pb/^{204}Pb$ = 15.58–15.70; $^{208}Pb/^{204}Pb$ = 38.42–39.57) reflect partial melting of the fossilized head of a late Paleozoic–early Mesozoic mantle plume, trapped at the base of the continental lithosphere. It has been suggested that this plume head has been tapped a number of times during the past 250 Myr, producing basanites, alkali basalts and rare tholeiitic basalts with remarkably homogeneous major and trace element and Sr–Nd–Pb isotope

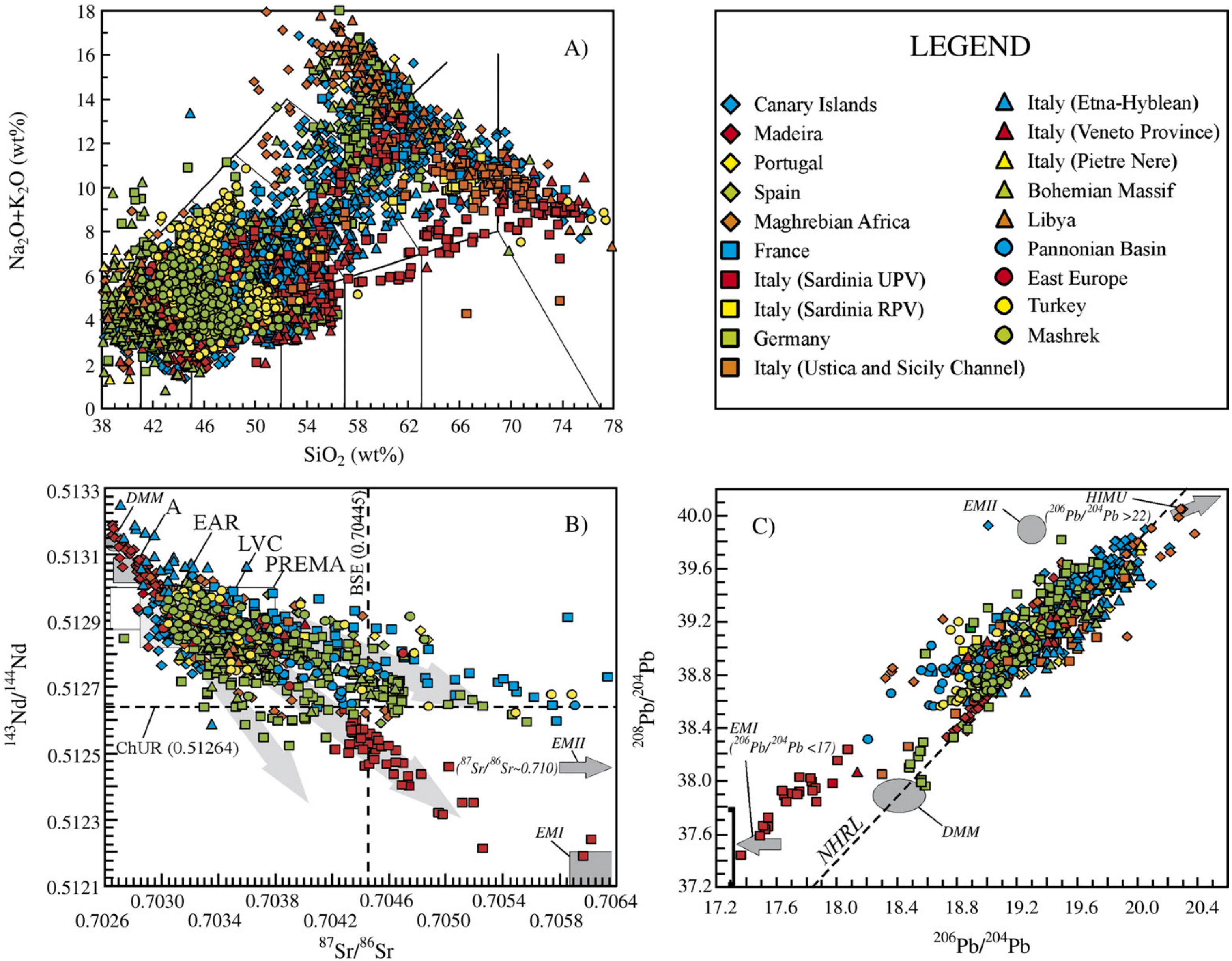


Fig. 13. a) Total alkali *vs.* silica diagram (TAS; le Maitre, 2002); b) $^{143}Nd/^{144}Nd$ *vs.* $^{87}Sr/^{86}Sr$ diagram; c) $^{208}Pb/^{204}Pb$ *vs.* $^{206}Pb/^{204}Pb$ diagram for the entire CiMACI province, including both primitive and differentiated magmatic rocks. For sources of data see the captions of Figs. 2–12.

compositions (Garfunkel, 1989; Mouty et al., 1992; Stein and Hofmann, 1992; Wilson et al., 2000). Garfunkel (1989), in contrast, favoured a model involving hotter than normal mantle underlying the Mashrek during the Cenozoic, and proposed that the magmatism of this area is caused by a system of intermittently active mantle plumes, similar to that envisaged by Granet et al. (1995) for Central European Volcanic Province (Sections 3.5 and 3.7), notwithstanding the low heat flow characteristics of the area.

6. Discussion

6.1. Geochemical characteristics of anorogenic mafic magmas in the cimaci Province

This study is based on a compilation of about 7800 whole-rock analyses from the published literature, combined with unpublished data of the authors (295 samples) from a number of circum-Mediterranean volcanic areas (Spain, Algeria, Italy, Poland, Turkey, France, Czech Republic). Clearly great care must be taken when using such an extensive data set for petrogenetic interpretation as the major and trace element and Sr–Nd–Pb isotopic analyses were obtained with a range of methods over more than 30 years. Thus, this does not represent an internally consistent data set. Nevertheless, we have confidence that most of the data we have selected are of good quality.

Among the CiMACI rocks, ultrabasic to basic compositions (SiO_2 <52 wt.%) form about 60% of the population, the remainder being intermediate to acid (SiO_2 >52 wt.%; Fig. 13). Most of the analysed samples are volcanic to sub-volcanic; plutonic rocks form <5% of the dataset.

The classification of a CiMACI rock as *anorogenic* is based on the following criteria.

6.1.1. Major elements

Most of the *anorogenic* rocks are mildly to strongly alkaline basanites, tephrites, nephelinites, hawaiites and alkali basalts. The SiO_2 content of the most primitive mafic rocks is between 40 and 52 wt.%, and the sum of Na_2O +K_2O is mostly between 3 and 10 wt.%. Subalkaline basalts are generally much rarer than alkaline rocks, and in some volcanic districts they are totally absent. These subalkaline basalts have tholeiitic affinity, characterized by iron enrichment in the initial stages of differentiation. In general, the CiMACI rocks are sodic with Na_2O/ K_2O in the range of 1.4–5. The TiO_2 contents of the most mafic samples (MgO >7 wt.%) ranges from ~1 to ~6 wt.%.

6.1.2. Trace elements

Most of the primitive (MgO >7 wt.%) *anorogenic* rocks have relatively homogeneous incompatible trace element compositions, with primitive mantle-normalised trace element patterns (Figs. 2–12) resembling those of end-member HIMU-OIB, such as those from St. Helena (South Atlantic Ocean) and Rurutu (South Pacific Ocean, French Polynesia). The most common feature of the trace element patterns are peaks at the HFSE (Nb–Ta–Hf–Zr) and negative K anomalies; LILE (Cs, Ba, Sr) are variably enriched whereas Rb is commonly depleted. Incompatible trace element ratios in the most primitive (MgO >7 wt.%) samples show relatively large ranges (e.g., Ba/Nb 3–13; La/Nb = 0.4–1.1; Ce/Pb = 10–50; Nb/U = 20–60; Zr/Nb = 2–6; Th/Ta = 0.7–2.2; Ta/Yb = 1–3). This suggests that the mantle source from which the magmas are derived must be heterogeneous. The less differentiated tholeiitic rocks (e.g., those from Germany) show no substantial differences from the less differentiated alkaline types in this respect. Only the quartz-bearing tholeiitic rocks and the most evolved compositions deviate significantly from the typical inter-element ratios of the alkaline rocks; as outlined in section 3.7, these deviations are most likely related to variable degrees of crustal contamination combined with fractional crystallisation and are not considered a primary mantle signature. Since similar trace element characteristics are evident in the most primitive mafic rocks from both the continental and oceanic (e.g. Canary Islands, Madeira) sectors of the CiMACI province, it seems likely that these reflect those of a sub-lithospheric mantle source.

The distinctive HIMU-OIB-like trace element patterns of the most primitive mafic CiMACI rocks have been attributed to the presence of residual garnet and hydrous phases (e.g., amphibole, phlogopite) in their mantle source (e.g., Wilson and Patterson, 2001; Wilson and Downes, 2006). Amphibole is the most likely phase to account for the marked depletion of Ba relative to Rb and the distinctive K trough in the primitive mantle-normalised trace element variation diagrams (Figs. 2–12). Phlogopite is stable to ~200 km depth along a normal mantle adiabat (with a potential temperature of ~1300 °C), whereas amphibole (K-richerite) may be stable up to at least 350 km (Tronnes, 2002). The upper temperature stability of both phlogopite and K-richterite is ~1400 °C, suggesting that mantle thermal gradients are not significantly elevated if there is a hydrous phase in the source.

Both negative and positive K spikes in primitive mantle-normalised diagrams have been related to the presence of phlogopite and/or amphibole in the mantle source

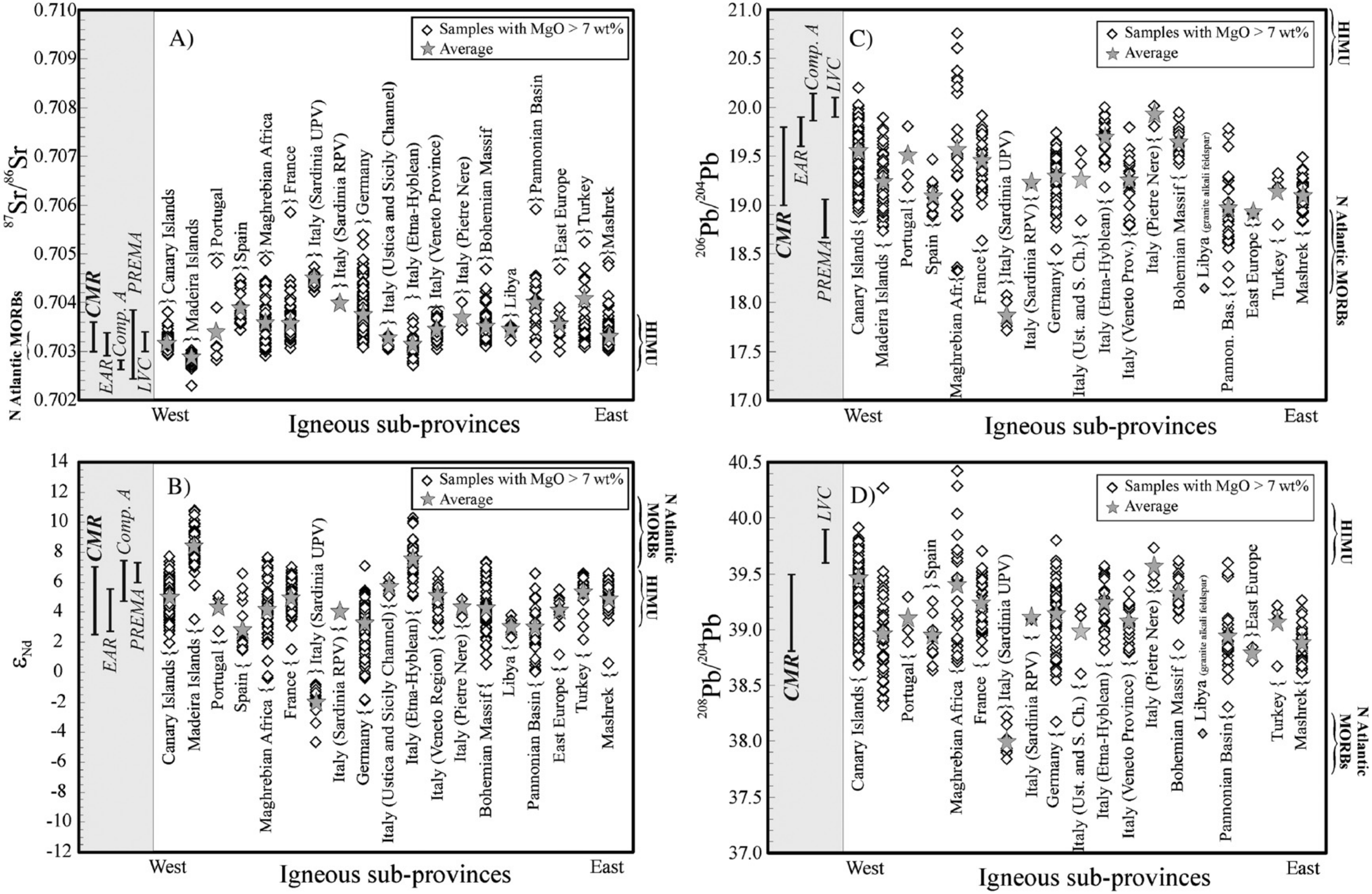


Fig. 14. a) ^{87}Sr/^{86}7Sr; b) ϵ_{Nd}; c) ^{206}Pb/^{204}Pb; d) ^{208}Pb/^{204}Pb *vs.* geographic location for the most primitive (MgO > 7 wt.%) mafic magmatic rocks of the CiMACI province. Stars = average composition for each district. For data sources see the captions of Figs. 2–12. EAR = European Asthenospheric Reservoir (Granet et al., 1995); Comp. A = Component A (Wilson and Downes, 1991); PREMA = Prevalent Mantle (Wörner et al., 1986); LVC = Low Velocity Component (Hoernle et al., 1995); CMR = Common Mantle Reservoir.

(Lustrino, 2001). In the first case (negative spikes), the hydrous phases are thought to remain in the residual mineralogy, thus retaining K in their lattice (e.g., Wilson et al., 1995b; Weinstein et al., 2006). In the second case (positive spikes and potassic melt compositions), these phases are thought to participate in the melting process (e.g., Cebrià and Lopez-Ruiz, 1995; Schiano et al., 2004). For the Hocheifel volcanic field in Germany (Section 3.7), Fekiacova et al. (in press-b) have concluded that the negative K anomaly is a source characteristic, whereas according to Jung et al. (2006) it is related to the presence of residual amphibole in the source. In their modelling of the transition from subduction-related to intra-plate magmatism in the western Mediterranean, Duggen et al. (2005) demonstrated that whilst K is a compatible element in phlogopite, the bulk-rock mineral-melt partition coefficient is still <1. Additionally the stabilization of even small amounts of phlogopite require significant amounts of K in the mantle source. Consequently it seems likely that depletion in K and other fluid–mobile trace elements (e.g. Rb, Th, U and Pb) might be a distinctive feature of the source of those CiMACI magmas with HIMU-OIB-like characteristics.

The occurrence of volcanic rocks with hybrid geochemical affinities intermediate between *anorogenic* and *orogenic* compositions (e.g., SE Spain, Morocco, Sardinia, Pannonian Basin, Serbia, Turkey; Pearce et al., 1990; Duggen et al., 2005; Seghedi et al., 2005; Lustrino et al., 2007) is relatively common within the CiMACI province. Most of these “hybrid” trace element compositions occur in areas in which there is a temporal transition from volcanic activity with *orogenic* to *anorogenic* characteristics. The hybrid composition has commonly been related to interaction between sub-lithospheric melts (with true *anorogenic* geochemical characteristics) and subcontinental lithospheric mantle (metasomatised by earlier subduction-related fluids).

6.1.3. Sr–Nd–Pb isotopic compositions

The Sr and Nd isotopic compositions of the CiMACI rocks are mostly confined to the depleted quadrant of the Sr–Nd isotope diagram (Fig. 13) with $^{87}Sr/^{86}Sr$ lower than present-day Bulk Silicate Earth estimates (BSE = 0.70445), and $^{143}Nd/^{144}Nd$ higher than Chondritic Uniform Reservoir values (ChUR = 0.51264). With some rare exceptions (e.g., UPV Sardinian group, Section 3.6.5), $^{206}Pb/^{204}Pb$ ratios are relatively radiogenic (18.8–20.4), whereas $^{207}Pb/^{204}Pb$ is highly variable. Whilst some of the variation in $^{207}Pb/^{204}Pb$ is undoubtedly a consequence of instrumental mass bias and insufficient mass bias correction in different laboratories, some of it is real; $^{208}Pb/^{204}Pb$ is similarly variable but shows a good correlation with $^{206}Pb/^{204}Pb$ (R^2 = 0.85; Figs. 2–13).

The Sr–Nd–Pb range of the least differentiated CiMACI rocks (with MgO >7 wt.%) is: $^{87}Sr/^{86}Sr$ = 0.7023–0.7049; $^{143}Nd/^{144}Nd$ = 0.51319–0.51240 (ϵ_{Nd} = +10.8 to −4.7); $^{206}Pb/^{204}Pb$ = 17.71–20.76; $^{207}Pb/^{204}Pb$ = 15.49–15.72; $^{208}Pb/^{204}Pb$ = 37.84–40.43 (Figs. 14–15). Δ7/4 and Δ8/4 values show a relatively large range: from −8.5 to +20.5 and from −42.2 to +99.8, respectively. A comprehensive dataset involving other isotopic systems commonly utilized in petrological studies (e.g., Lu/Hf, Re/Os, O, He, C) is currently available only for a few localities, and does not allow statistical treatment of the dataset.

Most of the CiMACI rocks (>90% of the dataset) plot within the geochemical ranges outlined above (Figs. 13–14). Whilst the majority of samples are of sodic affinity (Na_2O >K_2O, weight ratio), slightly potassic to ultrapotassic compositions (Na_2O/K_2O down to 0.7) occur in central Spain (Calatrava igneous province; leucitites; Cebrià and Lopez-Ruiz, 1995), NE Spain (Olot–Garrotxa–Ampurdan; leucitites to trachytes; Cebrià et al., 2000), France (Cantal; leucite nephelinites and basanites; Wilson and Downes, 1991; Wilson and Patterson, 2001), Sardinia (Montiferro basanite; Fedele et al., in press), Pannonian Basin (Bar leucitites; Harangi et al., 1995b) and Germany (Eifel leucitites; Mertes and Schmicke, 1985; Wörner et al., 1986). Rare CIPW quartz-normative basic rocks occur in Sardinia (Lustrino et al., 2002, 2007), Sicily (Hyblean Mts. and Mt. Etna; Beccaluva et al., 1998; Trua et al., 1998), southern France (Bas–Languedoc; Dautria and Liotard, 1990), Germany (e.g., Vogelsberg and Westerwald; Bogaard and Wörner, 2003; Haase et al., 2004), SE Turkey (e.g., Karasu Valley; Alici et al., 2001) and Jordan (Harrat Ash Shaam; Ibrahim et al., 2003). It is probable that the quartz-normative character of some of these basalts (e.g. Germany and SE Turkey) is related to lower crustal contamination of mantle-derived magmas. A role for crustal contamination has also been proposed in the petrogenesis of the more differentiated members of the weakly alkaline magma series in the French Massif Central (e.g., Downes, 1987; Wilson et al., 1995a,b; Dautria et al., 2004), the Tertiary Hocheifel volcanic field in Germany (Jung et al., 2006; Fekiacova et al., in press-a,b) and the alkaline lamprophyres of Morocco (e.g., Bernard-Griffiths et al., 1991). The effect of crustal contamination is reflected in more radiogenic Sr and less radiogenic Nd isotopic ratios and a shift in $^{206}Pb/^{204}Pb$ isotopes towards less radiogenic compositions.

Exceptions also exist when considering the incompatible trace element characteristics of individual igneous

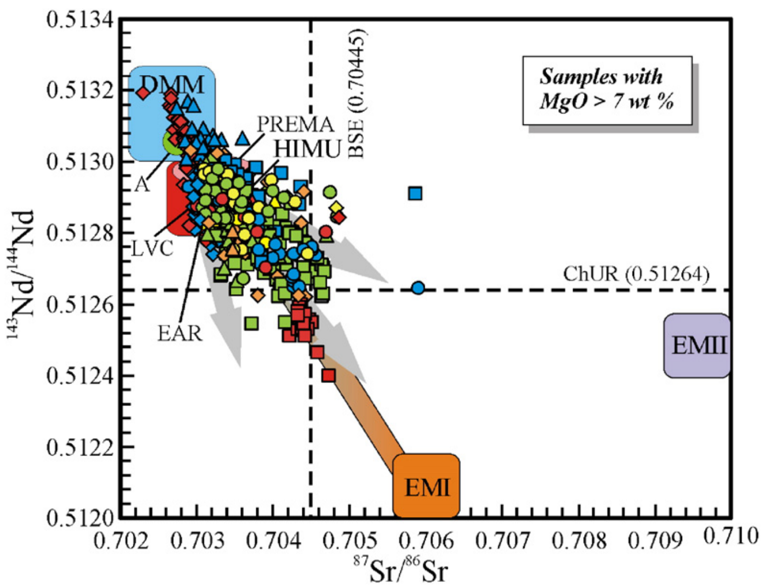


Fig. 15. Variation of $^{143}Nd/^{144}Nd$ *vs.* $^{87}Sr/^{86}7Sr$ for the most primitive (MgO > 7 wt.%) mafic magmatic rocks of the CiMACI province. For data sources see the captions of Figs. 2–12. DMM = Depleted MORB Mantle [average North Atlantic compositions (PETDB database; http://beta.www.petdb.org/); Kramers and Tolstkhin, 1997]; HIMU = high MU [(MU = $\mu = {}^{238}U/{}^{204}Pb$) Zindler and Hart, 1986]; EMI = Enriched Mantle type I (Zindler and Hart, 1986; Lustrino and Dallai, 2003); EMII = Enriched Mantle type II (Zindler and Hart, 1986) EAR = European Asthenospheric Reservoir (Granet et al., 1995); Comp. A = Component A (Wilson and Downes, 1991); PREMA = Prevalent Mantle (Wörner et al., 1986); LVC = Low Velocity Component (Hoernle et al., 1995). BSE = Bulk Silicate Earth estimate; ChUR = Chondritic Uniform Reservoir. Most of the CiMACI rocks are clustered in a relatively restricted Sr–Nd isotopic range (depleted quadrant). Only few mafic rocks have $^{143}Nd/^{144}Nd$ ratios lower than ChUR and virtually all of the CiMACI rocks have $^{87}Sr/^{86}Sr$ ratios lower that BSE. Grey arrows indicate the effect of interaction of asthenosphere-derived melts with lithospheric components characterized by radiogenic Sr and unradiogenic Nd isotopic compositions (pointing towards EMI and EMII end-members).

districts within the CiMACI province in more detail. The most striking example is provided by the middle Miocene Pleistocene magmatic rocks of Sardinia (e.g., Lustrino et al., 2007): most of the analysed samples (the UPV group of Lustrino et al., 2000) are characterized by much higher Ba/Nb (15–65), La/Nb (0.8–1.7), Sr/Nd (15–40), Th/Ta (1.4–5) and lower Ce/Pb (3–20) in the MgO 11–4 wt.% range compared to other CiMACI rocks. Other localities that deviate from the generally uniform incompatible trace-element composition of the CiMACI province include: Mt. Etna (Sicily) and Linosa Island (Sicily Channel), which are characterized by high Th/Ta (2.3–4.4), high La/Nb (1.2–1.8) and high Zr/Nb (6–14); some Canary Island rocks characterized by anomalously high La/Nb (1.9–3.5), high Ba/Nb (19–26) and high Zr/Nb (6–11); some Bohemian Massif rocks with high La/Nb (1.9–2.6); some French Massif Central rocks with high Ba/Nb (13–25) and Pliocene igneous rocks from the Eastern Rif (Morocco) which exhibit positive peaks at Ba rather than at Nb in primitive mantle-normalised trace element diagrams.

What emerges from Sr–Nd–Pb isotope and incompatible trace element survey of the primitive mafic igneous rocks from the CiMACI province is that there is a range of source components involved in the petrogenesis of the magmas, of both mantle and crustal origin. Whilst these can be decribed in terms of the mantle end-members DMM, HIMU, EMI, EMII (Hofmann, 1997), as discussed below, there is no implication that any of these end-members necessarily exists as a physical entity within the upper mantle (e.g., Armienti and Gasperini, in press). There is a general consensus that the upper mantle is made up not only of peridotitic rocks but also contains dispersed chemical and lithological heterogeneities (e.g., Allegre and Turcotte, 1986; Meibom and Anderson, 2003; Lustrino, 2006); the range in chemical composition of the CiMACI rocks undoubtedly reflects the presence of variably distributed and variable size heterogeneities.

6.2. Constraints on mantle plume involvement in the CiMACI province magmatism

A number of authors (e.g., Wörner et al., 1986; Wilson and Downes, 1991, 1992; Cebrià and Wilson, 1995; Granet et al., 1995; Hoernle et al., 1995; Wilson and Patterson, 2001) have proposed the existence of a common, and geochemically relatively uniform, mantle source for the most primitive mafic magmas of the CiMACI province. Some authors have related this source to the upwelling of a lower mantle plume beneath Europe and the eastern North Atlantic (e.g., Hoernle et al., 1995; Goes et al., 1999; Bell et al., 2004), and variously associated this plume with the Canary Islands (e.g., Cebrià and Lopez-Ruiz, 1995; Oyarzun et al., 1997), the Cape Verdes (Macera et al., 2003) and Iceland (e.g., Bijwaard and Spakman, 1999; Wilson and Patterson, 2001). Alternatively this source has been envisaged as a layer of anomalous mantle in the Transition Zone (410–660 km depth) from which small-scale, finger-like, upper mantle plumes rise as convective instabilities (e.g., Granet et al., 1995; Wilson and Patterson, 2001; Macera et al., 2004). Froidevaux et al. (1974) were amongst the first authors to propose the existence of a deep mantle plume beneath the French Massif Central, although they were dubious about the presence of a significant thermal anomaly.

It is possible that an ancient “fossil” mantle plume head, unable to penetrate thick lithosphere (and therefore trapped and stored beneath it) could provide a source for subsequent intra-plate magmatism. Such an enriched reservoir (i.e. a fossilised mantle plume head) could have been intermittently tapped beneath the CiMACI province during the Cenozoic; the process would be similar to that proposed for the Cenozoic evolution of the Mashrek (Israel, Jordan, Syria, Lebanon) by Stein and Hofmann (1992) and Wilson et al. (2000). A similar model has been recently proposed by Rotolo et al. (2006) to explain the volcanism of Sicily and the Sicily Channel because of the geochemical similarities of the mafic volcanic rocks to HIMU-FOZO mantle compositions. However, there are problems with this model in that the highest volume of magmatism (e.g., Hyblean Mts. and Mt. Etna) occurs on the thickest lithosphere and the smallest (e.g., Sicily Channel) on the thinnest part.

Wilson and Downes (1992) and Lustrino et al. (2000) have suggested that the HIMU-like characteristics of the sodic alkaline CiMACI mafic magmas could, in some cases, be related to the recycling of ancient oceanic crust (Hercynian age or even older; >300 Ma) in the upper mantle. Sampling of such an enriched mantle source clearly does not require the involvement of deep mantle plumes to trigger partial melting; this can be caused by adiabatic decompression of ambient upper mantle in response to lithospheric extension or its convective destabilization, without the need for a thermal anomaly.

Tomographic images of the mantle beneath the CiMACI province can be quite variable according to the type of data (global *vs.* local) and data reduction methods employed. On the basis of the presently available data there is no unequivocal evidence for the presence of a regionally extensive, thermally anomalous, mantle plume head (or heads) beneath the CiMACI province (e.g., Courtillot et al., 2003; Piromallo and Morelli, 2003; Ritsema and Allen, 2003; Montelli et al.,2004; Zhao, 2004). High resolution local tomography experiments in the Massif Central and Eifel (e.g., Granet et al., 1995; Ritter et al., 2001) do, however, provide compelling evidence for the presence of narrow (50–100 km diameter), finger-like convective instabilities in the upper mantle which appear to originate from the Transition Zone.

The common, relatively uniform, mantle reservoir sampled intermittently during the Cenozoic by virtually all the primitive mafic magmas of the CiMACI province has been alternatively called PREMA (PREvalent MAntle; Wörner et al., 1986), Component A (Wilson and Downes, 1991, 1992), LVC (Low Velocity Component; Hoernle et al., 1995), EAR (European Asthenospheric Reservoir; Cebrià and Wilson, 1995; Granet et al., 1995) or CEA (Central European Anomaly; Goes et al., 1999).

Here we propose a new acronym (CMR: Common Mantle Reservoir) to encompass the much larger area of the upper mantle considered in this study. Whilst the merits of adding yet another acronym to the “mantle alphabet” is debatable we feel that CMR is an appropriate term. Most of the previous acronyms for this reservoir were defined based on the geochemical and Sr–Nd–Pb isotope characteristics of European primitive mafic volcanic rocks, whereas the CMR is based on a significantly larger dataset, more that 50% of which comprises North African, Central Atlantic and Arabian plate volcanic rocks.

In order to define the isotopic composition of CMR, we have chosen only those samples showing the following characteristics: 1) MgO >7 wt.%; 2) Nb/Nb* >1 [where Nb/Nb* is $Nb_N/((K_N*La_N)$^0.5) and the subscript “N” refers to the primitive mantle-normalised abundance]; 3) Pb/Pb* <1 [where Pb/Pb* is $Pb_N/((Ce_N*Pr_N)$^0.5)]; 4) $U_N/Nb_N<1$; 5) $Nb_N/K_N>1$; 6) $Ce_N/Pb/_N<1$; 7) $Pb_N/Pr_N<1$. The rationale behind this choice is that sub-lithospheric anorogenic melts should have significant positive Nb anomalies and negative Pb anomalies, as is commonly observed for lavas from ocean islands such as the Canaries and Madeira.

We propose the following isotopic composition for the CMR: $^{87}Sr/^{86}Sr$ = 0.7030–0.7037; $^{143}Nd/^{144}Nd$ = 0.51300–0.51279; $^{206}Pb/^{204}Pb$ = 18.95–19.85; $^{207}Pb/^{204}Pb$ = 15.55–15.65; $^{208}Pb/^{204}Pb$ = 38.80–39.60; $^{207}Pb/^{206}Pb$ = 0.78–0.84; $^{208}Pb/^{206}Pb$ = 1.99–2.05; $\Delta7/4=-4$ to $+8$; $\Delta8/4=-15$ to $+27$. Whilst these isotopic ranges inevitably include some sampling biases (for example, the negative $\Delta7/4$ values are mostly

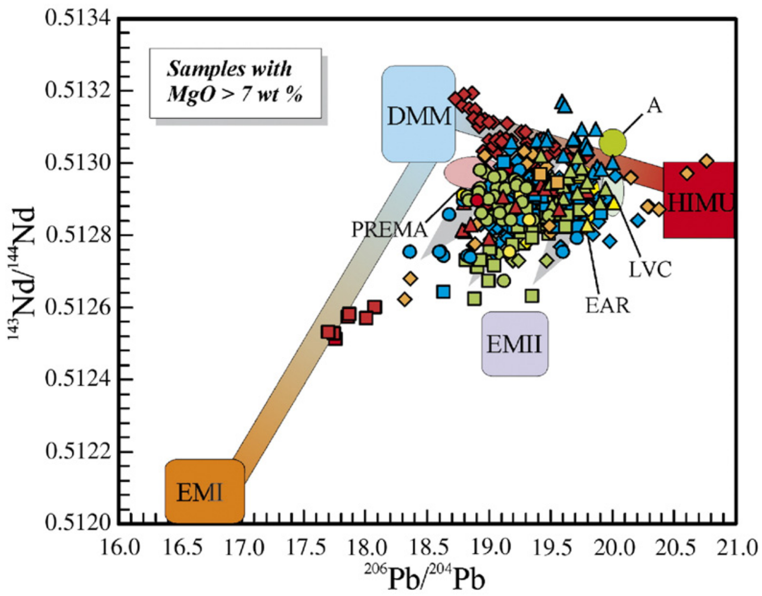


Fig. 16. $^{143}Nd/^{144}Nd$ *vs.* $^{206}Pb/^{204}Pb$ diagram for the most primitive (MgO > 7 wt.%) mafic magmatic rocks of the CiMACI province. References to data sources are given in the captions of Figs. 2–12. DMM = Depleted MORB Mantle [average North Atlantic compositions (PETDB database; http://beta.www.petdb.org/); Kramers and Tolstkhin, 1997]; HIMU = high MU [(MU = $\mu = ^{238}U/^{204}Pb$) Zindler and Hart, 1986]; EMI = Enriched Mantle type I (Zindler and Hart, 1986; Lustrino and Dallai, 2003); EMII = Enriched Mantle type II (Zindler and Hart, 1986) EAR = European Asthenospheric Reservoir (Granet et al., 1995); Comp. A = Component A (Wilson and Downes, 1991); PREMA = Prevalent Mantle (Wörner et al., 1986); LVC = Low Velocity Component (Hoernle et al., 1995). The range of Nd and Pb isotopic compositions of the most mafic CiMACI rocks indicates the involvement of at least four mantle components: mixing between HIMU-like and DMM-like compositions is not able to explain the spread of isotopic data. The involvement of EM components is therefore required. Only the Sardinian rocks seem to reflect a mixing between DMM and EMI end-members, with very limited involvement of HIMU. Grey arrows indicate the effects of interaction of HIMU-DMM melts with lithospheric components characterized by unradiogenic Nd and mildly radiogenic Pb isotopic compositions (pointing towards EMI and EMII end-members).

confined to the Canary Islands and Madeira for which a wealth of isotopic data is available), we consider that they are broadly representative of the CMR.

The inferred Sr–Nd–Pb isotopic compositions of the various mantle reservoirs are plotted in Figs. 13–17. In terms of these diagrams it is clear that the CMR is always more or less "polluted" by other components, as shown by the grey arrows in Figs. 13 and 15–17. These "contaminant" components are generically called EM (Enriched Mantle) and appear to reflect the involvement of crustal lithologies (or partial melts/fluids thereof) at mantle depths (e.g., Wilson and Downes, 1991, 1992). Based on the present state of knowledge, it is not clear what these EM components effectively represent in a physical sense. For example, some "anomalous" EM compositions represented by potassic volcanic rocks (leucitites and leucite nephelinites) in the Calatrava district (Spain), Massif Central (France) and Eifel region (Germany) have been related to derivation of the magmas from phlogopite-rich domains within the lithospheric mantle (e.g., Wörner et al., 1986; Wilson and Downes, 1991, 1992; Cebrià and Lopez-Ruiz, 1995). However, other magmatic rocks, also attributed to lithosphere derivation (e.g. Sardinian middle Miocene-Quaternary volcanism), have sodic rather than potassic characteristics, suggesting that K-metasomatism of the lithosphere is not a uniform characteristic of the Circum-Mediterranean region.

In evaluating a potential role for deep mantle plumes to explain the geochemical characteristics of the CMR two major assumptions need to be challenged.

6.2.1. *Can deep mantle plumes be easily traced in seismic tomographic images?*

There remains considerable controversy about the existence of deep mantle plumes (e.g. Maruyama et al., 2004; Foulger et al., 2005; Foulger and Jurdy, in press; www.mantleplumes.org). It is not appropriate here to review all the arguments for or against; nevertheless, it is important to identify some of the key issues. Firstly, the low resolution of global seismic tomography images, even at upper mantle depths (Hoernle et al., 1995; Ritsema and Allen, 2003), limits their use in identifying mantle plumes. Local seismic tomography experiments (involving the use of closely spaced mobile arrays of seismometers; e.g., Granet et al. 1995; Ritter et al., 2001)

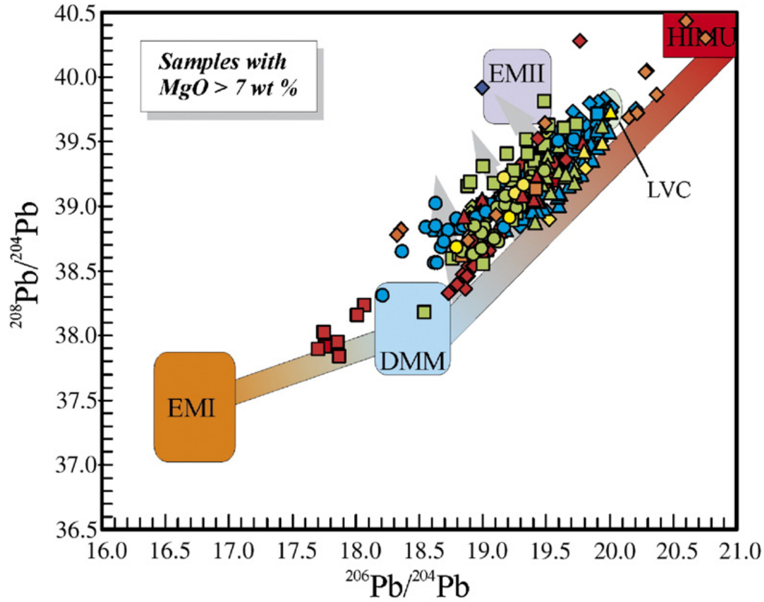


Fig. 17. Variation of $^{208}Pb/^{204}Pb$ *vs.* $^{206}Pb/^{204}Pb$ for the most primitive (MgO>7 wt.%) mafic magmatic rocks of the CiMACI province. For data sources see the captions of Figs. 2–12. DMM = Depleted MORB Mantle [average North Atlantic compositions (PETDB database; http://beta.www.petdb.org/); Kramers and Tolstkhin, 1997]; HIMU = high MU [(MU = $\mu = ^{238}U/^{204}Pb$) Zindler and Hart, 1986]; EMI = Enriched Mantle type I (Zindler and Hart, 1986; Lustrino and Dallai, 2003); EMII = Enriched Mantle type II (Zindler and Hart, 1986) EAR = European Asthenospheric Reservoir (Granet et al., 1995); Comp. A = Component A (Wilson and Downes, 1991); PREMA = Prevalent Mantle (Wörner et al., 1986); LVC = Low Velocity Component (Hoernle et al., 1995). The Pb isotopic composition of the most mafic CiMACI rocks requires the involvement of at least four mantle end-members: mixing between HIMU-like and DMM-like compositions is not able to explain the spread of isotopic data. The involvement of EM components is therefore required. As in Figs. 15 and 16, only the Sardinian rocks seem to reflect a mixing between DMM and EMI end-members, with very limited involvement of HIMU. Grey arrows indicate the effect of mixing between HIMU-DMM melts and EMII-like lithospheric components.

provide much better resolution within the upper mantle (<400 km depth). The problem is that, within the CiMACI province, such detailed studies are, at present, limited to the Massif Central and Eifel volcanic areas only.

Interpretation of seismic tomography images is predicated on the assumption that seismic velocity anomalies (v_p, v_s) can be converted to an equivalent temperature contrast in the mantle; in particular slow v_p and/or v_s anomalies (calculated with respect to a mantle reference model) are considered to reflect the presence of anomalously hot mantle (i.e., a mantle plume). It is worth noting, however, that low values of v_s can also reflect the presence of a small amount of partial melt (e.g., the presence of 1% melt in mantle can reduce v_s by up to 3.3%; Faul et al., 1994), mineralogical modification of the mantle (e.g., the presence of phlogopite), or presence of fluids such as H_2O or CO_2 (e.g., Mei and Kohlstedt, 2000; Ritter, in press).

Ritter et al. (2001) and Keyser et al. (2002) have reported variable S-wave velocity reductions in the upper mantle beneath the Eifel region, based upon the same high-resolution local tomography data set. The velocity contrast is depth dependent from $\sim -3\%$ at ~ 50 km to $\sim -0.5\%$ at ~ 150 km; between ~ 170 and 240 km there is no negative velocity anomaly (but rather a slight positive anomaly), while at ~ 260–440 km the anomaly in v_s is less than -1%. The relatively small velocity perturbation (v_s velocity contrast generally less than -1%), and the presence of a "hole", clearly does not provide strong support for upwelling of a mantle plume from the lower mantle. The data are, however, consistent with a diapir-like upwelling from the Transition Zone. A more recent, high-resolution, S-wave velocity model for the upper mantle beneath the Eifel (Pilidou et al., 2005) suggests a different velocity profile with depth: $v_s \sim -2\%$ down to 100 km; $\sim -1\%$ at 100–150 km; $\sim -2\%$ at 150–300 km; $\sim -1\%$ at 300–350 km; $\sim 0\%$ at 350–400 km. The differences between the two velocity models reflect the different data sets used (i.e., local *vs.* global tomography); nevertheless, both are in agreement that there is a slow velocity amomaly in the upper mantle beneath the Eifel. Montelli et al. (2004), on the basis of a global finite-frequency tomography study, consider that the P-wave velocity anomaly beneath the Eifel is robust to 650 km depth (i.e., to the base of the upper mantle). According to Ritter et al. (2001), the maximum v_p anomaly of -2% down to depths of at least 400 km could be equivalent to about a 150–200 °C excess temperature.

In summary, although seismic tomography studies provide the only evidence we have for the presence of thermally and/or chemically anomalous mantle beneath the CiMACI province, we must interpret the resultant images cautiously. As noted above (e.g., Lustrino and Carminati, in press), different inversions of the same data sets (e.g., based on different reference models or data processing algorithms) can give contrasting results. Additionally, the v_p and v_s velocity anomalies are relatively small (generally less than −2%) and cannot be related exclusively to temperature variations, but also to the presence of partial melt/fluid, the mineralogy and bulk-rock chemistry of the mantle.

Piromallo et al. (2001) demonstrated the presence of an extensive region of fast seismic velocities in the Transition Zone beneath the CiMACI province. This could be interpreted as a region of subducted slabs of oceanic lithosphere which have accumulated in the base of the upper mantle during the Alpine (or earlier) collisional orogeny. If this interpretation is correct then the Transition Zone is not likely to be a source of thermally anomalous mantle plumes (i.e., hotter than ambient mantle), but it could be a rich source of volatiles (particularly water). The presence of a high velocity Transition Zone also renders the direct lower mantle feeding of the slow velocity anomalies observed in the upper mantle highly unlikely.

6.2.2. Do deep mantle plumes have distinctive geochemical fingerprints?

According to much of the current geochemical and petrological literature, mantle plumes anchored at the core−mantle boundary at depths of ~2900 km can be identified on the basis of the radiogenic and rare gas isotope systematics of the magmas derived from them. This assumption is highly equivocal and may lead to conflicting conclusions.

The $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ isotopic compositions of the most primitive mafic CiMACI rocks have previously been considered indicative of a "plume-signature" (e.g., Wilson and Patterson, 2001) because of their similarity to those of HIMU-OIB. The other mantle reservoir that can evolve with a broadly similar Sr–Nd isotopic composition (usually referred to as being isotopically "depleted") is the DMM (Depleted MORB Mantle). Most of the basalts emplaced along mid-ocean ridges are characterized by low $^{87}Sr/^{86}Sr$ isotopic ratios (down to 0.7025) and high $^{143}Nd/^{144}Nd$ (up to 0.5133). The Nd–Sr isotope characteristics of DMM are generally considered to be the consequence of a long history of melt extraction from the upper mantle, with preferential removal of Rb and Nd compared to Sr and Sm, respectively (e.g. Hofmann, 1997, 2004; Workman and Hart, 2005). The time-integrated isotopic evolution of mantle sources with extremely low Rb/Sr and Nd/Sm ratios leads to strongly retarded radiogenic ingrowth of ^{87}Sr and strongly enhanced ingrowth of ^{143}Nd. Here the first paradox is apparent: the depleted upper mantle (DMM) is characterized by nearly the same Sr–Nd isotopic composition as an alleged deep and undegassed mantle component (the source of HIMU-OIB). The lowermost mantle is generally inferred to contain material never, or only rarely, involved in partial melting processes; therefore, it should evolve with higher Rb/Sr and Nd/Sm ratios than the DMM source. However, what we see is that the DMM and HIMU sources have more or less similar Sr–Nd isotopic compositions, both plotting in the depleted Nd–Sr isotopic quadrant (although DMM is more depleted than HIMU).

Recent studies of the He and other noble gas (Ne, Ar) isotope systematics of peridotite xenoliths from the European mantle lithosphere have provided conflicting evidence for the involvement of lower mantle plumes in the petrogenesis of the Cenozoic magmas (Section 3.7; Buikin et al., 2005; Gautheron et al., 2005). The xenoliths have remarkably constant He isotope compositions of 6–7 R_A (where R_A is the $^3He/^4He$ atmospheric ratio), close to that of MORB-source mantle (~8 R_A) but much lower than that of typical plume-related OIB (e.g. Hawaii, Iceland) of up to 40 R_A. Dunai and Baur (1995) attributed this constant He isotope composition to a mixture of MORB-source helium and that derived from crustal materials recycled into the mantle during the Hercynian orogeny at ~300 Ma. Gautheron et al., (2005) present convincing arguments that the He isotope characteristics of the continental lithosphere are the product of a steady-state process in which metasomatic fluids from the asthenosphere permeate the lithosphere; according to their model ^4He is generated in the lithosphere by decay of U and Th whereas the ^3He is derived from the asthenosphere.

Another piece of evidence often used to argue for the involvement of deep mantle plumes in the petrogenesis of intra-plate basalts (both continental and oceanic) is the extremely radiogenic $^{206}Pb/^{204}Pb$ isotopic ratios (>21) of the most extreme HIMU-OIB. Such Pb isotopic compositions have been attributed to the storage of ancient subducted oceanic crust, evolving with a high $^{238}U/^{204}Pb$ ratio (μ value), in the lower mantle for periods of time in excess of 1–2 Gyr (e.g., Hofmann, 1997). Most of the CiMACI basalts approach the $^{206}Pb/^{204}Pb$ range (typically 19–20) of end-member HIMU-OIB, but never reach such strongly radiogenic values. Thirlwall (1997) argued that variably radiogenic Pb in OIB can result from the continuous incorporation (over the course of geological time) into the mantle of high-μ recycled oceanic crust, resulting in a range of OIB sources with progressively less extreme Pb isotope signatures with decreasing age. OIB sources with $^{206}Pb/^{204}Pb$ values in the range 19–20 are referred to in this model as "young HIMU". Such young HIMU mantle sources should have low $^{207}Pb/^{204}Pb$ for a given $^{206}Pb/^{204}Pb$, and consequently negative $\Delta7/4$ values. Stracke et al. (2005) have argued that the apparent ubiquity of a FOZO-like component in the upper mantle is a natural consequence of the continuous recycling of subducted oceanic crust. Within the CiMACI data set negative $\Delta7/4$ values are relatively rare and mostly confined to samples from the oceanic islands of the Canary Islands and Madeira. Making appropriate allowance for the likely errors in calculated $\Delta7/4$ values (probably at least ±2), this suggests that much of the Pb isotope variability of the CiMACI basalts is likely to be the consequence of interaction between asthenosphere-derived magmas and enriched sub-continental lithosphere.

The strongest evidence supporting a mantle plume origin for the CiMACI magmas is provided by the correlation, in some areas, between the geochemical and Sr–Nd–Pb isotope characteristics of the most primitive mafic magmas and those of oceanic island basalts (OIB), evidence for contemporaneous regional basement uplift, lithospheric thinning and local seismic tomography images showing finger-like upper mantle low velocity anomalies beneath the volcanic fields. The French Massif Central may be considered a type locality in this respect (e.g., Wilson and Patterson, 2001). Here there is strong support for the upwelling of an upper mantle plume below the base of the lithosphere, based on both local and global seismic tomography images (e.g., Granet et al., 1995; Goes et al., 1999), the geochemical characteristics of the Cenozoic mafic alkaline lavas (e.g., Wilson and Downes, 1991; Hoernle et al., 1995; Granet et al., 1995; Wilson and Bianchini, 1999; Wilson and Patterson, 2001) and petrographic/textural studies of mantle xenoliths entrained within the lavas (e.g., Xu et al., 1998; Zangana et al., 1999; Downes, 2001). Most of the plume-models which have been proposed for the CiMACI province are effectively based on the similarities between the various igneous districts and the Massif Central. Unfortunately, the axiom: "Massif Central Cenozoic mafic magmas = partial melts of anomalously hot upper mantle plume" has led several authors to propose the existence of mantle plumes beneath other areas without strong geophysical constraints for the existence of low velocity (i.e., potentially hotter than ambient) mantle.

One of the strongest arguments against a deep mantle plume model for the CiMACI magmatism is the generally small volume of erupted magma in individual areas (and the scarcity of high-degree partial melts, e.g., tholeiitic basalts). With the exception of the Harrat Ash Shaam volcanic province in Syria, Jordan and Saudi Arabia, there are no regionally extensive continental flood basalt sequences, conventionally interpreted as the products of mantle plume heads impacting on the base of the lithosphere. Moreover, the common association of the most voluminous Cenozoic CiMACI volcanic activity with extensional tectonics, and with reactivated ancient structural dislocations within the lithosphere (e.g., Hercynian terrane boundaries within the northern foreland of the Alps; Wilson and Downes, 1991), suggest that lithospheric extension has played a significant role in the petrogenesis of the CiMACI magmas (e.g., Wilson and Downes, 1992; Bogaard and Wörner, 2003; Lustrino and Sharkov, 2006). However, the presence of a mostly amagmatic N–S trending graben system in the Alpine foreland (Rhine, Limagne and Bresse grabens) requires a more complex model than simple extensional thinning of the European lithosphere.

It is worth noting that Lustrino and Carminati (in press) explain the lack of magmatic activity in most of the ECRiS in terms of lithospheric thickness: where the lithosphere/asthenosphere boundary is very deep (up to 140 km) magmatic activity is absent. Only where the lithospheric thickness is reduced to 60–80 km (and where this is associated also with a shallow Moho) does magmatic activity in the ECRiS become important (e.g., French Massif Central and Eifel). In summary, the lack of extensive tholeiitic flood basalt activity, the existence of typically small volume (with some exceptions) individual alkaline volcanic centers (explained by low-degree partial melting of mantle sources which need not be anomalously hot), the presence of magmatic activity associated with large scale lithospheric discontinuities, the absence of systematic age progressions of magmatic activity and the lack of evidence for regional basement uplift in many areas before the onset of igneous activity, argues against the involvement of a deep mantle plume (or plumes) in the CiMACI province. Nevertheless, the existence of largely amagmatic graben systems in the Alpine foreland, clearly indicates that parental magmas cannot solely be generated by lithospheric extension and adiabatic decompression of the upper mantle.

When comparing the geochemical characteristics of the Cenozoic volcanic rocks of the Massif Central with those of other CiMACI provinces for which there is no seismic tomographic or geological evidence (e.g., no doming before magmatic activity) for the existence of a mantle plume, another problem arises. Many authors (e.g., Wilson and Downes, 1991; Oyarzun et al., 1997; Wilson and Patterson, 2001; Macera et al., 2004) have noted that the Cenozoic alkaline mafic magmatic rocks of the Canary Islands, Iberian peninsula (Portugal and Spain), France (French Massif Central and Bas–Languedoc), Germany, Bohemian Massif (Poland, Czech Republic), Italy (Veneto

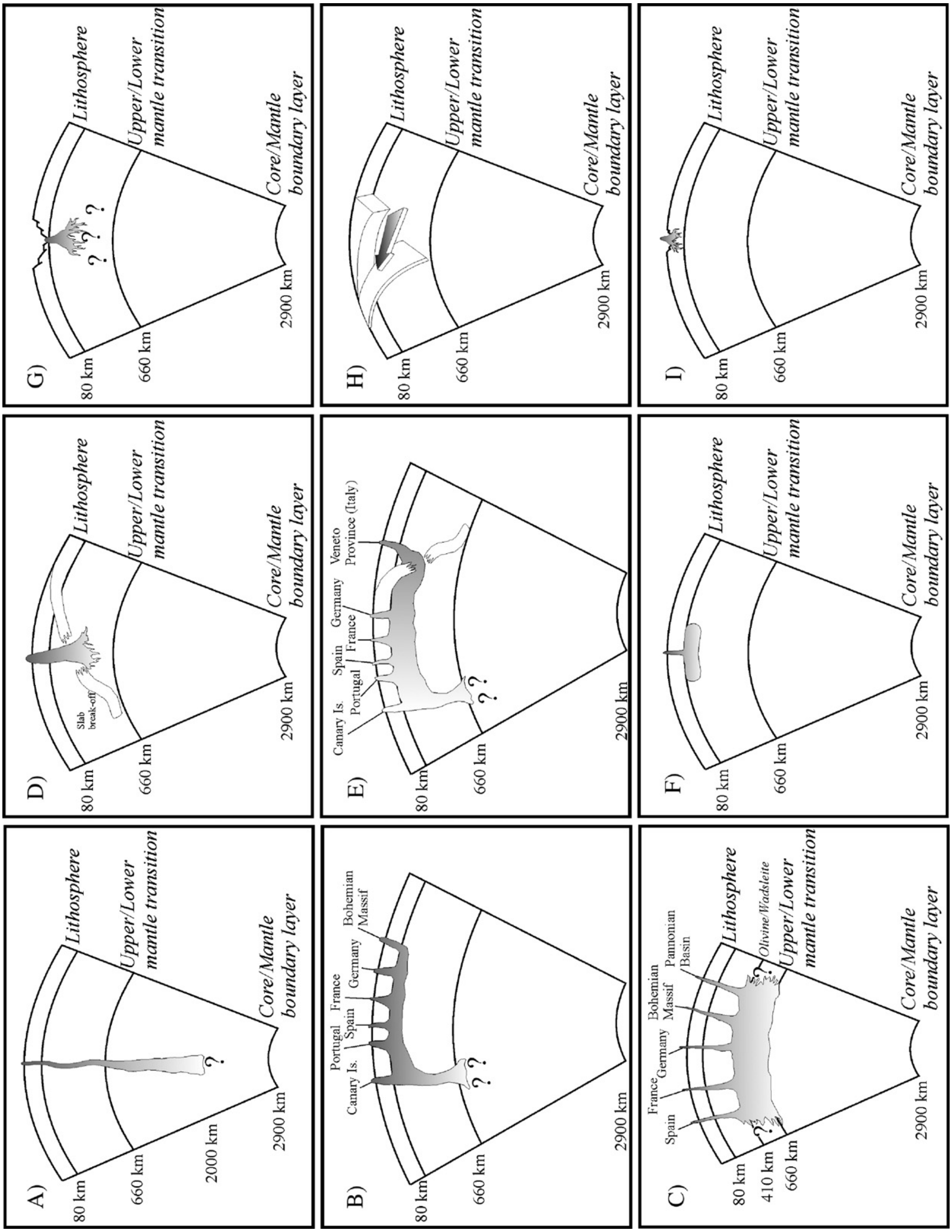
Volcanic Province) and Pannonian Basin (Hungary, Austria, Slovakia) are virtually indistinguishable from each other in terms of their incompatible trace element abundances and ratios and Sr–Nd–Pb isotopic compositions. A much closer investigation of all the CiMACI volcanic sub-provinces, however, indicates the existence of significant local heterogeneity, at least in Sr–Nd–Pb isotope signature. In Fig. 14 we compare all the available Sr–Nd–Pb isotope data for the most mafic (MgO >7 wt.%) CiMACI province magmatic rocks. With the exception of most Sardinian middle Miocene-Quaternary rocks, the average $^{87}Sr/^{86}Sr$ is surprisingly constant, considering the regional extent of the CiMACI province. The average Nd isotopic composition (expressed as ϵ_{Nd}) of the individual CiMACI sub-provinces is also confined to a relatively narrow range (from +5.8 to +2.2, with the exception of most of the Sardinian volcanic rocks). However, when Pb isotopes are taken into consideration, the differences between the individual CiMACI provinces become more marked (average $^{206}Pb/^{204}Pb$ ranges from ~19.0 to ~19.8, and $^{208}Pb/^{204}Pb$ from ~38.8 to ~39.5). The distinctive geochemistry of the Sardinian volcanic rocks is explained more fully in Lustrino (2005a), and Lustrino et al. (2000, 2004a,b, 2007).

The complete isotopic range (including the Sardinian rocks) of CiMACI igneous rocks with MgO >7 wt.% is: $^{87}Sr/^{86}Sr$ = 0.7026–0.7054; ϵ_{Nd} = +10.5 to −4.5; $^{206}Pb/^{204}Pb$ = 17.7–20.4; $^{207}Pb/^{204}Pb$ = 15.50–15.69; $^{208}Pb/^{204}Pb$ = 37.8–39.9 (Fig. 14). Whilst some of this relatively wide range of isotopic compositions can undoubtedly be explained by crustal contamination in high-level magma chambers, it is clear that the parental magmas must have been derived from relatively heterogeneous mantle sources with variable depletion/enrichment histories. Crustal contamination is a relatively common process, supported by both petrographic (e.g., presence of crustal xenoliths) and geochemical data [e.g., the quartz-normative magma types (quartz tholeiites) are typically those with more radiogenic Sr and more unradiogenic Nd isotopic composition in a given province, suggesting assimilation of older crustal rocks]. The presence of crustal rocks with a wide range of ages and compositions in the CiMACI province can account for the spread of isotopic compositions. However, if we focus our attention only on the most primitive mafic (MgO >7 wt.%) SiO_2-undersaturated–saturated compositions (i.e., CIPW nepheline- or hypersthene-normative), a wide range of Sr–Nd–Pb isotope compositions still exist. This strongly argues for the presence of isotopic heterogeneities at mantle depths.

In Figs 15, 16 and 17 the $^{87}Sr/^{86}Sr$–$^{143}Nd/^{144}Nd$, $^{206}Pb/^{204}Pb$–$^{143}Nd/^{144}Nd$, and $^{206}Pb/^{204}Pb$–$^{208}Pb/^{204}Pb$ compositions of the most primitive CiMACI rocks are plotted together with the hypothetical mantle end-members DMM, HIMU, EMI and EMII. The DMM composition is based on the isotopic composition of North Atlantic MORB from the PETDB database and from the estimate of Kramers and Tolstikhin (1997). It is important to note that the $^{206}Pb/^{204}Pb$ isotopic composition of DMM is usually considered to be much less radiogenic (generally ~17–17.5; Hofmann, 2004) than average north Atlantic MORB (here assumed to be ~18.5); HIMU is from Zindler and Hart (1986); EMII is from Zindler and Hart (1986) and Workman et al. (2004); EMI is from Lustrino and Dallai (2003). The inferred range of compositions of the hypothetical European mantle plume source (PREMA, EAR, LVC, Component A) is shown for comparison. What emerges from these isotopic comparisons is:

(1) Most of the CiMACI primitive magma compositions appear to indicate the involvement of both DMM and HIMU mantle source components in their petrogenesis;
(2) The majority of the primitive magma compositions also require the involvement of EM components to explain the range of Sr–Nd–Pb isotopic data;
(3) n a small number of cases (e.g., Sardinia, Italy), mixing between DMM and EMI mantle sources (or partial melts thereof), in the apparent absence of any HIMU-like component, is clearly identifiable.

Our major problem lies in elucidating what the CMR component actually represents. We believe that it is not necessary to invoke the upwelling of a lower mantle plume during the Cenozoic to explain the existence of this component in the upper mantle beneath the CiMACI province. In our preferred model the CMR represents the ambient sub-lithospheric mantle, locally contaminated with recycled crustal lithologies, introduced, for example, by subduction along previously active collision zones or as consequence of lithospheric delamination. Having precluded the involvement of deep-rooted thermal anomalies (classical mantle plume concept), the other potential candidate for the CMR could be the thermal boundary layer (TBL) at the base of the lithosphere. However it is highly improbable that such a relatively small volume of the mantle (i.e. the TBL) could be the only source of such widespread and long-lasting igneous activity. This does not mean that the TBL is not involved in the genesis of the CiMACI rocks (indeed as noted by Wilson et al. (1995b) the dominant magma generation zone lies close to the base of the


A)
80 km
660 km
2000 km
2900 km
Lithosphere
Upper/Lower mantle transition
Core/Mantle boundary layer
B)
80 km
660 km
2900 km
Canary Is.
Portugal
Spain
France
Germany
Bohemian Massif
C)
80 km
410 km
660 km
2900 km
Spain
France
Germany
Bohemian Massif
Pannonian Basin
Olivine/Wadsleite
Lithosphere
Upper/Lower mantle transition
Core/Mantle boundary layer
D)
80 km
660 km
2900 km
Slab break-off
Lithosphere
Upper/Lower mantle transition
Core/Mantle boundary layer
E)
80 km
660 km
2900 km
Canary Is.
Portugal
Spain
France
Germany
Veneto Province (Italy)
F)
80 km
660 km
2900 km
Lithosphere
Upper/Lower mantle transition
Core/Mantle boundary layer
G)
80 km
660 km
2900 km
Lithosphere
Upper/Lower mantle transition
Core/Mantle boundary layer
H)
80 km
660 km
2900 km
Lithosphere
Upper/Lower mantle transition
Core/Mantle boundary layer
I)
80 km
660 km
2900 km
Lithosphere
Upper/Lower mantle transition
Core/Mantle boundary layer

lithosphere), but simply that there must be another, more volumetrically important, magma source in the sub-lithospheric mantle.

The presence of recycled crustal lithologies in the CMR can provide the EMI and HIMU components. Their age is indeterminate; they could have been subducted during the Mesozoic–Cenozoic closure of the Neo-Tethys Ocean, or during the earlier Hercynian orogenic cycle in the mid–late Carboniferous, as proposed by Wilson and Downes (1991) and Lustrino et al. (2000). In the latter case this would require that the sub-lithospheric mantle has remained physically coupled to the overlying lithosphere for the past 300 Myr. This model, however, has some limitations: heterogeneities within the ambient asthenospheric mantle (caused by contamination with subducted and/or delaminated crustal lithologies) might be expected to result in a much wider range of compositions than actually observed.

7. Concluding remarks

During the Cenozoic, the Circum-Mediterranean Anorogenic Cenozoic Igneous (CiMACI) province was characterized by igneous activity with an extremely wide range of chemical compositions. Both *orogenic* (subduction-related) and *anorogenic* (intra-plate) igneous rocks were emplaced over a wide region contemporaneous with Alpine collisional tectonics induced by Eurasia–Africa convergence. The geochemical characteristics of the most primitive mafic magmas cannot always be used to define a precise geodynamic setting because of the previous geochemical "heritage" of both the lithospheric and asthenospheric mantle beneath the region. A precise definition of the terms *anorogenic* and *orogenic* cannot be given, because of the existence of a large number of variables. The term *anorogenic* cannot be used as an exact synonym of "intra-plate" because CiMACI igneous activity frequently occurs along passive continental margins (e.g., Portugal), in back-arc basins (e.g., Sardinia, Pannonian Basin, W Anatolia, Aegean Sea), along ancient suture zones affected by Cenozoic rifting (e.g., Germany, France), along strike-slip fault systems (e.g., Mashrek) or in a hinterland position above active or fossil subduction zones (e.g., Turkey, Serbia). Moreover, in a number of areas *anorogenic* lavas are closely associated, both spatially and temporally, with *orogenic* lavas (e.g., Maghrebian Africa, SE Spain, SE France, Sardinia, Carpathian–Pannonian area, Serbia, E Turkey). Consequently, the term *orogenic* cannot be used as synonym of "subduction-related", because the geochemical characteristics of some *orogenic* magmatic rocks appear to be related to ancient subduction-fluid modification of the mantle lithosphere, rather than to contemporaneous subduction.

This detailed review of >7800 whole-rock analyses of magmatic rocks from the CiMACI province has

Fig. 18. Schematic representation of the various petrological and geodynamic models proposed in the literature to explain the origin of individual igneous districts within the CiMACI province. a) *Deep mantle plume* from lower mantle (proposed to explain the origin of the Eifel magmatic field in Germany; Goes et al., 1999); b) *Thin spot model*, proposed to explain the origin of several European Cenozoic volcanic regions. Two variations of this model are known: the first considers that the stem of the plume is centred below the Canary Islands (e.g., Cebrià and Lopez-Ruiz, 1995), whereas the second model considers that the stem is centred beneath Iceland (Wilson and Patterson, 2001); c) *Finger-like upper mantle plumes* proposed to explain the origin of a number of the European Cenozoic volcanic provinces. All these plumes are linked in the Transition Zone to a single upper mantle reservoir (the European Asthenospheric Reservoir; EAR) located between the 410 and the 660 km discontinuities (Hoernle et al., 1995; Granet et al., 1995); d) *Slab break-off model* proposed to explain the common temporal transition of igneous activity from *orogenic* to *anorogenic* geochemical characteristics. According to this model, the *anorogenic* magmatism is caused by adiabatic decompression of asthenospheric mantle replacing the detached slab of oceanic lithosphere (Maury et al., 2000; Coulon et al., 2002; Keskin, 2003); e) *Slab break-off coupled with an upper mantle plume and lithospheric thin spots*, proposed to explain the igneous activity of the Veneto area in NE Italy. This model explains the paradox of the existence of igneous activity with *anorogenic* geochemical characteristics in an active subduction-related tectonic setting (Macera et al., 2003); f) *Fossil plume model*, proposed to explain the igneous activity of Israel. According to this model, the prolonged and geochemically uniform Mesozoic/Cenozoic igneous activity in Israel and Jordan is related to partial melting of a common mantle reservoir represented by a fossil plume head unable to penetrate thick lithosphere (Stein and Hofmann, 1992; Wilson et al., 2000). A similar model (fossil plume head) has also been proposed to explain the volcanic activity in Sicily and the Sicily Channel (Rotolo et al., 2007); g) *Passive rifting model*, proposed to explain the origin of several igneous districts in Turkey, Jordan, Israel, Syria and Germany. According to this model, magmatism is triggered by adiabatic decompression induced melting of the asthenosphere activated by lithospheric thinning (Altherr et al., 1990; Shaw et al., 2003; Lustrino and Sharkov, 2006); h) *Slab-window model*, proposed to explain the igneous activity of Mt. Etna and Ustica Island (Italy). According to this model, magmatism is related to an asthenospheric window developing between two parts of a subducting plate with different rheological characteristics: a thicker and a lighter continental plate (barely subductable) and a thinner and denser (easily subductable) oceanic plate (Gvirtzman and Nur, 1999; Trua et al., 2003; Armienti et al., 2004); A modified version of this model involves *slab roll-back and lithospheric delamination*, proposed for the continental margins of S Iberia and NW Africa by Duggen et al. (2005). In this model, upwelling of sub-lithospheric mantle and the generation of *anorogenic* lavas is restricted to the continental margins of S Iberia and NW Africa; i) *Lithospheric model*, proposed to explain the igneous activity of Sardinia and NW Harrat Ash Shaam (Israel). According to this model, volcanic activity is related to decompression melting of lithospheric mantle variably metasomatised during earlier tectono-thermal events e.g. Pan-African/Hercynian (or even older) orogenesis (Lustrino et al., 2004b, 2007; Weinstein et al., 2006).

allowed us to identify a relatively common Sr–Nd–Pb isotopic and trace element signature in the most primitive mafic magmas. The source of these magmas, the CMR (Common Mantle Reservoir), is defined largely on the basis of its Sr–Nd–Pb isotopic characteristics and has many similarities with previously identified mantle reservoirs in the circum-Mediterranean area (e.g., EAR, LVC, PreMa, Component A). The relatively large range of incompatible trace element ratios and contents observed in the most primitive CiMACI rocks indicate the existence of a number of different magma sources, which could reside in both the lithospheric and sub-lithospheric mantle. Partial melts of the CMR are likely to have interacted locally with metasomatically enriched domains in the sub-continental lithospheric mantle, where they acquired distinctive (often potassic) compositions. In rare cases the CiMACI rocks are considered to be pure partial melts of the lithospheric mantle without any thermal or geochemical contribution from the sub-lithospheric mantle.

The sub-lithospheric mantle beneath the circum-Mediterranean area is frequently considered to have a variable geochemical composition, ranging from strongly depleted to highly enriched. In a number of cases the term “enriched” has been used with very different meanings. Additionally there is sometimes confusion between terms such as “plume”, “sub-lithospheric mantle”, “asthenosphere” and “upper mantle”. The petrological definition of the base of the lithospheric mantle (i.e., the TBL) can also be ambiguous; in some areas the TBL can acquire an asthenospheric (DMM) geochemical signature by thermal capture, whilst still being part of the lithosphere from a rheological point of view.

A number of different models have been proposed to explain the origin and the evolution of the most primitive mafic CiMACI magmas. A series of cartoons indicating the range of possible models for magma generation in individual CiMACI volcanic provinces is shown in Fig. 18. The various models are as follows.

Fig. 18a: Mantle plume anchored in the lower mantle (possibly near the core–mantle boundary), with a near vertical stem. This model has been proposed to explain the geochemical characteristics of Cenozoic igneous rocks from Germany (e.g., Goes et al., 1999; Wedepohl and Baumann, 1999).

Fig. 18b: Channelised mantle plume (a modified version of the *thin spot* model of Thompson and Gibson, 1991). In this case the stem of the plume is strongly curved, and the root of the plume is anchored in the eastern North Atlantic Ocean (more or less at the present-day location of the Cape Verde or Canary Islands). This model relates the CiMACI activity to a long-lived plume system associated with the 200 Ma Central Atlantic Magmatic Province (e.g., Oyarzun et al., 1997; Hames et al., 2003). It has been used to explain the Cenozoic magmatism in central Spain (e.g., Cebrià and Lopez-Ruiz, 1995) and the Paleogene igneous activity in NW Italy (Veneto Province; Beccaluva et al., 2007a). A slightly different, but analogous, scenario was proposed by Bijwaard and Spakman (1999) and Wilson and Patterson (2001) who suggested that the CiMACI igneous activity could be related to outflow from the Iceland mantle plume which was channelled towards the SE. Hoernle et al. (1995) also proposed a similar concept, hypothesising a mantle plume channelled from the eastern North Atlantic.

Fig. 18c: Multiple mantle plumes anchored in the upper mantle Transition Zone at 410–660 km depth — the “mantle hot finger” model of Granet et al. (1995) and Wilson and Patterson (2001). In this case there is no need to invoke the presence of a lower mantle plume. The finger-like convective instabilities are generated within the upper mantle, and there is no requirement that they are thermally anomalous. This model can explain the paradox of the relatively small volumes of volcanic activity in the individual CiMACI provinces, and has been proposed to explain the Cenozoic magmatic activity in France, Germany and the Bohemian Massif (e.g., Granet et al., 1995; Wilson and Patterson, 2001; Ritter et al., 2001).

Fig. 18d: Shallow mantle upwelling induced by subducted slab break-off. This model is often invoked to explain the temporal transition from igneous activity with *orogenic* to that with *anorogenic* geochemical characteristics, e.g., in Maghrebian Africa, Pannonian Basin and Turkey (e.g., Maury et al., 2000; Coulon et al., 2002; Keskin, 2003). Asthenospheric partial melting is caused by adiabatic decompression at normal mantle potential temperatures as the mantle beneath the slab rises through the slab-window created by the rupture of the downgoing slab.

Fig. 18e: Deep-rooted mantle plume channelised along sub-lithospheric discontinuities but also rising through a slab-window. This model is a hydrid of models (b) and (d), and has been proposed to explain the igneous activity of the Veneto Volcanic Province in northern Italy (e.g., Macera et al., 2003) and the northern Pannonian Basin (Harangi et al., 2006).

Fig. 18f: Fossil plume head ponded at the base of the lithosphere, and intermittently partially melted during the past 250 Myr by adiabatic decompression (lithospheric thinning). This model has been invoked to explain the petrogenesis of continental flood basalts in Jordan, Syria and Israel (e.g., Stein and Hofmann, 1992; Wilson et al., 2000). A similar model involving a fossil plume head ponded at the base of thinned lithosphere has ben proposed to explain the volcanic activity in Sicily and the Sicily Channel (Rotolo et al., 2006).

Fig. 18g: Shallow asthenospheric mantle adiabatically decompresed by lithospheric thinning, e.g. passive rifting in a trans-tensional tectonic setting. A possible example is the magmatism along the Dead Sea fault zone in S Turkey, Jordan, Israel and Syria (e.g., Altherr et al., 1990; Shaw et al., 2003; Lustrino and Sharkov, 2006).

Fig. 18h: Adiabatic decompression of asthenospheric mantle as consequence of the formation a slab-window (e.g., Gvirtzman and Nur, 1999). According to this model, the magmatic activity is structurally (but not geochemically) related to a subduction system: slab roll-back or simple lithospheric delamination creates low-pressure regions into which asthenospheric mantle flows is "sucked". This model has been proposed to explain the volcanism of Ustica Island (southern Tyrrhenian Sea) and Mt. Etna (Sicily; e.g., Trua et al., 2003; Armienti et al., 2004; Faccenna et al., 2005) and W Anatolia (Innocenti et al., 2005; Agostini et al., 2007). A variant of this model involves delamination of bands of sub-continental lithosphere caused by slab roll-back and steepening of subducted oceanic lithosphere. This model has been proposed for the southeastern Carpathians (Girbacea and Frisch, 1998) and for the continental margins of S Iberia and NW Africa by Duggen et al. (2005).

Fig. 18i: Partial melting of lithospheric mantle metasomatised during earlier orogenic cycles by subduction-zone fluids/partial melts (e.g., Weinstein et al., 2006) or by lower crustal lithologies which have subsided into the upper mantle under their own weight (e.g., Lustrino et al., 2007). In the first case, Israeli Plio-Pleistocene alkali volcanism has been related to partial melting of heterogeneous lithospheric mantle containing pyroxenite veins formed during Late Proterozoic Pan-African events as consequence of slab-derived fluid fluxing. On the other hand, the model proposed for most of the middle Miocene-Quaternary volcanism of Sardinia is based on the assumption that after any major orogenic cycle, as a consequence of crustal thrust stacking, mafic lower crustal rocks are metamorphosed into granulitic/eclogitic assemblages, with consequent increase in density and gravitative instability. The high density of the granulite–eclogite facies lower crust may force the lowermost part of the lithospheric keel to founder into the asthenosphere. During such delamination and detachment, the lower crust may partially melt and produce magmas with a tonalitic–trondjemitic–granodioritic composition that metasomatise the overlying mantle (formerly asthenospheric mantle that has replaced the lithospheric mantle and is now accreted to the base of the lithosphere). This model satisfactorily explains the distinctive Sr–Nd–Pb isotopic and trace element characteristics of most of the middle Miocene-Quaternary volcanic rocks of Sardinia (e.g., Lustrino et al., 2000, 2007; Lustrino, 2005a).

The above models can be grouped into two main sub-types:

(1) Models that require active asthenospheric (or even deeper) mantle convection (i.e., mantle plumes);
(2) Models that rely strongly on lithospheric extension (or delamination and detachment) to induce passive, adiabatic, decompression melting of both asthenospheric and lithospheric upper mantle.

Neither sub-type can explain the complete range of magmatism in the CiMACI province. Probably the best case study explaining the complexity of the petrogenetic processes is represented by the Canary Islands igneous activity (e.g., Anguita and Hernán, 2000).

We believe that there is no need to invoke the presence of anomalously hot mantle (reflecting the involvement of a lower mantle plume actively upwelling from a thermal boundary layer at the core–mantle boundary) beneath the CiMACI province to explain both the regional distribution and geochemical characteristics of the *anorogenic* volcanism (Lustrino and Carminati, in press). If, however, we adopt a more permissive definition of "mantle plume", allowing it to encompass *passive*, diapiric upwellings of the upper mantle (e.g., the mantle "hot fingers" model of Wilson and Patterson, 2001), then we can relate the CiMACI province magmatism to multiple upper mantle plumes upwelling at various times during the Cenozoic. To avoid confusion we recommend that such upper mantle plumes are referred to as diapiric instabilities.

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Appendix A. Supplementary data

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