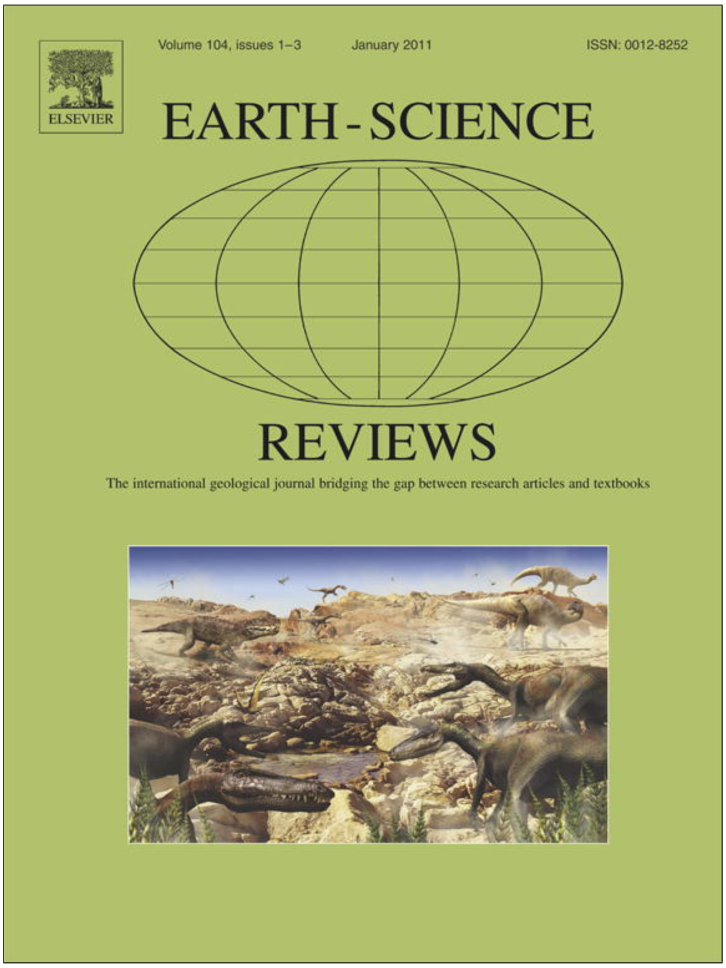
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The Central-Western Mediterranean: Anomalous igneous activity in an anomalous collisional tectonic setting

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ABSTRACT

The central-western Mediterranean area is a key region for understanding the complex interaction between igneous activity and tectonics. In this review, the specific geochemical character of several 'subduction-related' Cenozoic igneous provinces are described with a view to identifying the processes responsible for the modifications of their sources. Different petrogenetic models are reviewed in the light of competing geological and geodynamic scenarios proposed in the literature.

Plutonic rocks occur almost exclusively in the Eocene–Oligocene Periadriatic Province of the Alps while relatively minor plutonic bodies (mostly Miocene in age) crop out in N Morocco, S Spain and N Algeria. Igneous activity is otherwise confined to lava flows and dykes accompanied by relatively greater volumes of pyroclastic (often ignimbritic) products. Overall, the igneous activity spanned a wide temporal range, from middle Eocene (such as the Periadriatic Province) to the present (as in the Neapolitan of southern Italy). The magmatic products are mostly SiO_2-oversaturated, showing calcalkaline to high-K calcalcaline affinity, except in some areas (as in peninsular Italy) where potassic to ultrapotassic compositions prevail. The ultrapotassic magmas (which include leucitites to leucite-phonolites) are dominantly SiO_2-undersaturated, although rare, SiO_2-saturated (i.e., leucite-free lamproites) appear over much of this region, examples being in the Betics (southeast Spain), the northwest Alps, northeast Corsica (France), Tuscany (northwest Italy), southeast Tyrrhenian Sea (Cornacya Seamount) and possibly in the Tell region (northeast Algeria).

Excepted for the Alpine case, subduction-related igneous activity is strictly linked to the formation of the Mediterranean Sea. This Sea, at least in its central and western sectors, is made up of several young (<30 Ma) V-shaped back-arc basins plus several dispersed continental fragments, originally in crustal continuity with the European plate (Sardinia, Corsica, Balearic Islands, Kabylies, Calabria, Peloritani Mountains). The bulk of igneous activity in the central-western Mediterranean is believed to have tapped mantle 'wedge' regions, metasomatized by pressure-related dehydration of the subducting slabs. The presence of subduction-related igneous rocks with a wide range of chemical composition has been related to the interplay of several factors among which the pre-metasomatic composition of the mantle wedges (i.e., fertile vs. refractory mineralogy), the composition of the subducting plate (i.e., the type and amount of sediment cover and the alteration state of the crust), the variable thermo-baric conditions of magma formation, coupled with variable molar concentrations of CO_2 and H_2O in the fluid phase released by the subducting plates are the most important.

Compared to classic collisional settings (e.g., Himalayas), the central-western Mediterranean area shows a range of unusual geological and magmatological features. These include: a) the rapid formation of extensional basins in an overall compressional setting related to Africa-Europe convergence; b) centrifugal wave of both compressive and extensional tectonics starting from a 'pivotal' region around the Gulf of Lyon; c) the development of concomitant Cenozoic subduction zones with different subduction and tectonic transport directions; d) subduction 'inversion' events (e.g., currently along the Maghrebian coast and in northern Sicily, previously at the southern paleo-European margin); e) a repeated temporal pattern whereby subduction-related magmatic activity gives way to magmas of intraplate geochemical type; f) the late-stage appearance of magmas with collision-related 'exotic' (potassic to ultrapotassic) compositions, generally absent from simple

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subduction settings; g) the relative scarcity of typical calcalkaline magmas along the Italian peninsula; h) the absence of igneous activity where it might well be expected (e.g., above the hanging-wall of the Late Cretaceous–Eocene Adria–Europe subduction system in the Alps); i) voluminous production of subduction-related magmas coeval with extensional tectonic régimes (e.g., during Oligo-Miocene Sardinian Trough formation).

To summarize, these salient central-western Mediterranean features, characterizing a late-stage of the classic 'Wilson Cycle' offer a 'template' for interpreting magmatic compositions in analogous settings elsewhere.

Contents

1. Introduction

Magmatism and its geodynamic associations in the central-western Mediterranean (hereafter CWM) has been the focus of numerous geochemical, petrological, and geophysical studies (e.g., Wilson and Bianchini, 1999; Carminati and Doglioni, 2004; Schmid et al., 2004; Guerrera et al., 2005; Peccerillo, 2005; Harangi et al., 2006; Beccaluva et al., 2007a; Lustrino and Wilson, 2007; Duggen et al., 2008; Avanzinelli et al., 2009; Conticelli et al., 2009a; Carminati et al., accepted for publication, and references therein). Despite the massive geochemical and petrologic database that now exists, a considerable number of highly controversial questions remain, not surprisingly, given the overall geodynamic complexity of the region.

The range of tectonic processes recorded in this area (continental rifting and drifting, lithospheric boudinage, back-arc basin opening, formation of volcanic arcs and orogens) are generally grouped under the 'rubric' of 'Alpine–Himalayan orogeny-related' effects, ultimately linked to the pre-, syn-, and post-collisional effects of converging Africa and Eurasia (and the associated Africa-derived micro-plates).

In the CWM, the Alpine–Himalayan orogeny has been plausibly subdivided geographically and temporally into a number of second-order tectonic phases and associated orogenies, which include: the Alpine (s.s.), Pyrenean, Apennine, Betic, Rifian, Tellian and Atlas Belts (Fig. 1; e.g., Carminati and Doglioni, 2004; Boccaletti et al., 2005; Chaoulan et al., 2008; Frizon de Lamotte et al., 2009; Carminati et al., accepted for publication). Despite their common association with the Africa–Eurasia collision, several key differences exist, such as the polarities and ages of subduction systems, the variable rates of subduction rollback, the involvement of both continental and oceanic fragments in subduction/obduction processes, the depth of basal décollement, the wide variety of associated magmatic products, the foredeep development and its association with molasse or flyschoid deposits. All these issues are still controversial and are critical to a sound understanding of the relationship between tectonics and magmatism. The main differences between the inferred geodynamic models hinge on the relative motions of Africa and Eurasia (whether they are of peripheral significance or fundamental to shaping of the CWM), and the role of micro-continents such as Iberia, Adria, AlKaPeCa, Briançonnais, Sesia–Lanzo and Nebbio (e.g., Lustrino et al., 2009, Carminati et al., accepted for publication, and references therein). No consensus exists also on the geodynamic significance of oceanic or thinned crustal 'salients' located between these two major plates, on their respective subduction polarities, (as for the Liguride, Ionian, Lucanian and Valais Oceans), and the dynamics of both upper and lower mantle regions. For example, are deep mantle plumes a major factor in determining magma compositions, as compared to the effects of passive asthenospheric decompression in response to lithosphere extension? And how plumes may be significant in the context of palinspastic plate reconstruction with respect to commonly expected geochemical signatures? The latter have all been exhaustively used as source 'fingerprints' to characterize 'subduction-related' or 'intraplate' tectonic settings. Additionally, the presence of 'exotic' lithologies such as carbonatites, lamprophyres, kamafugites and lamproites, have been commonly used as 'proofs' *pro* or *contra* the existence of thermal anomalies, upper mantle structures and physical distribution models of chemical heterogeneities (e.g., Lustrino, 2010, and references therein).

Here, we provide an overview of CWM magmatism in relation to its diverse geodynamic settings with the aim of critically evaluating competing possible interpretations of the geochemical data. The immense quantity of seismic, structural, petrologic, and geochemical data compiled over at least 5 decades and the ongoing magmatic and tectonic activity of this area, make it an ideal target to investigate the relationships between magmatism and tectonic setting on a scale ranging from individual microplate motions to the entire Africa–Eurasia collision zone. A compiled electronic database containing more than 7000 whole-rock analyses (Appendix 1) may be downloaded from the Earth-Science Reviews data repository (Supplementary data).

2. Geodynamic and igneous evolution of the Central-Western Mediterranean

A complete review of all tectonic and petrologic models pertaining to specific districts in the CWM is beyond the scope of this paper.

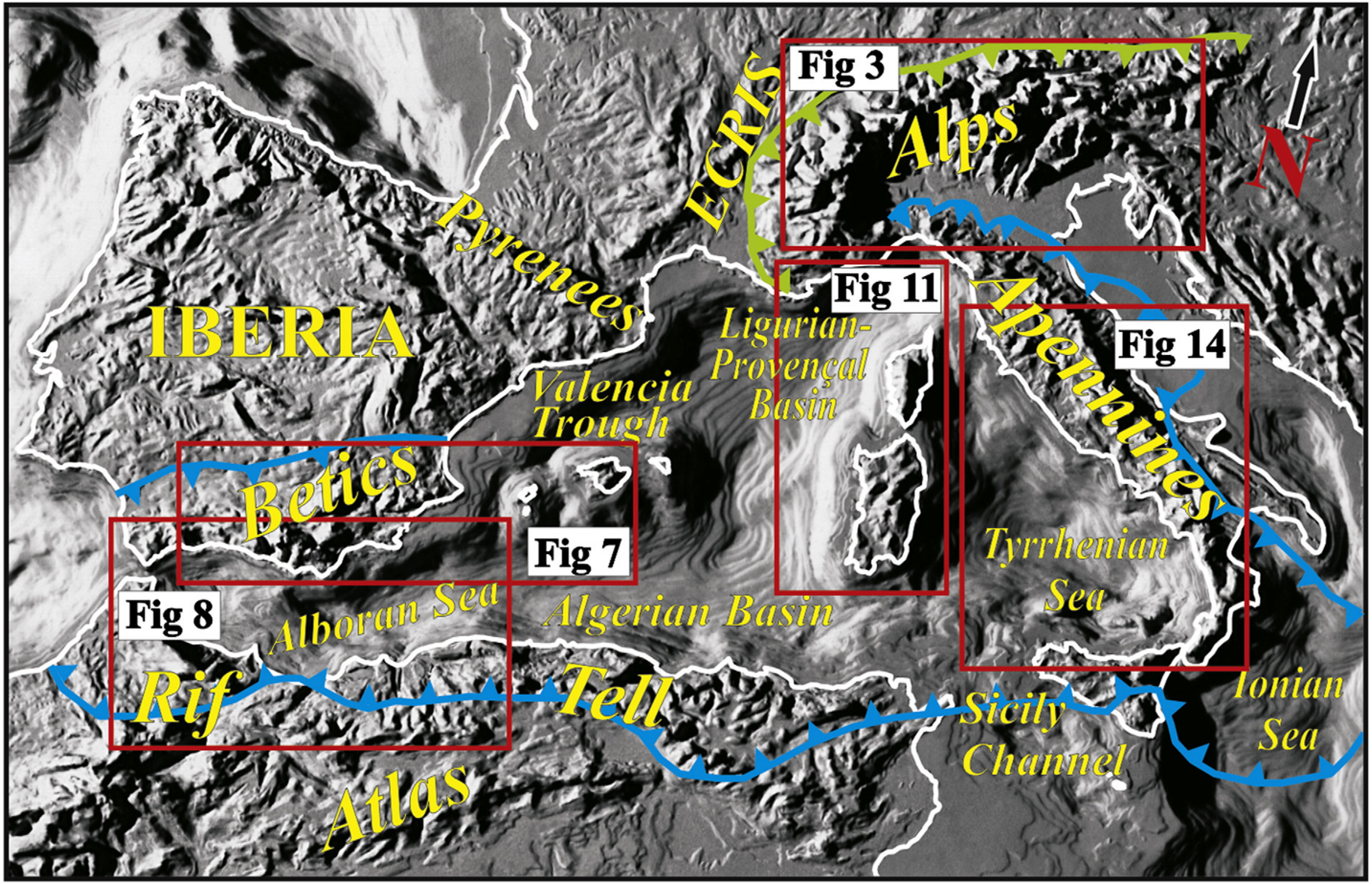

Fig. 1. Central-Western Mediterranean area digital elevation and bathymetric map investigated in this review, showing the most prominent mountain chains (Alps, Pyrenees, Apennines, Tell, Rif and Betics) and basins (Valencia Trough, Alboran Sea, Algerian Basin, Ligurian–Provençal Basin, Tyrrhenian Sea, Sicily Channel and Ionian Sea). ECRIS = European Cenozoic Rift System. Rectangles indicate close up of the various igneous districts.

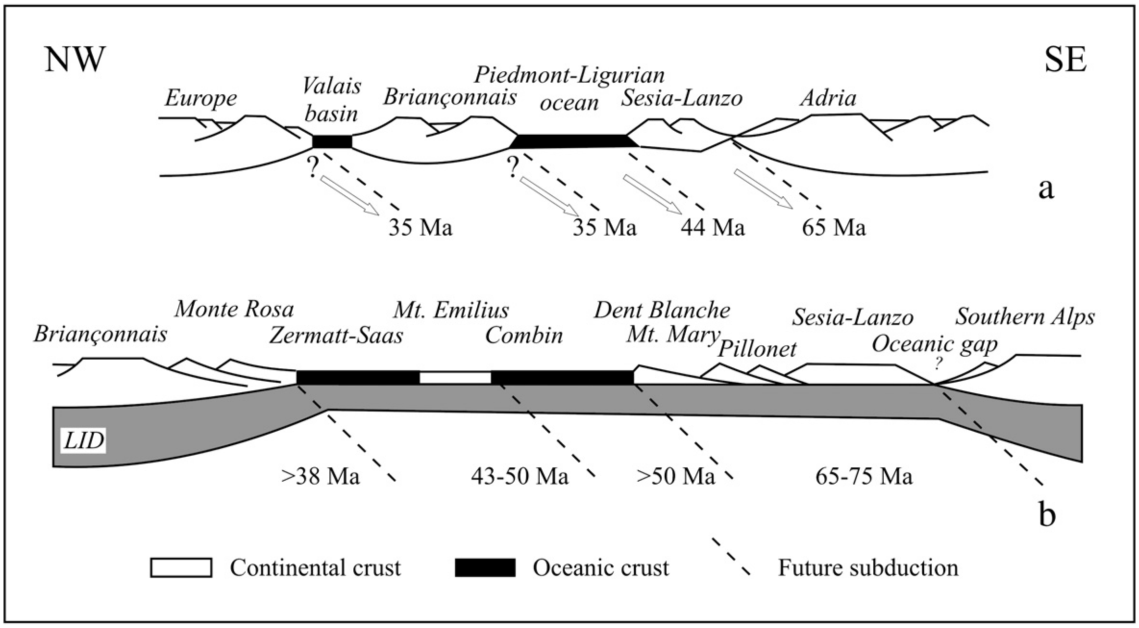

Fig. 2. Cretaceous–Eocene paleostructural restoration of Alpine Tethys in Western Alps, modified from Dal Piaz et al. (2001). A = Reconstruction after Rubatto et al. (1998). B = Reconstruction after Dal Piaz et al. (2001).

The data are evaluated within a geodynamic framework, with emphasis on those models presented in the recent literature (most containing comprehensive bibliographies). The main problem is that, in such complex areas, geological interpretations (e.g., tectonic direction of nappes, types of faults, age of igneous activity and chemical-mineralogical composition of igneous and metamorphic rocks) are still equivocal. As examples, some tectonic units in the Alps and in northeast Corsica have been alternatively interpreted as thrusts or back-thrusts associated to a subduction system with completely opposite direction; the ages of some igneous rocks in southern Spain have both been considered the true age of magma emplacement or resetting ages due to younger metamorphic overprint; the chemical and isotopic composition of igneous rocks has been alternatively considered as the derivation from “normal” or “depleted” or “exotic” mantle sources, on the basis of artificial pre-concepts that can mine the scientific validity of the models themselves.

Before providing a more detailed description of the geology and magmatism of the CWM, we summarize below some of its most significant geodynamic characteristics.

- The CWM is a geologically young area, mostly developed during the last 30 Myr;
- The geological structure and the igneous activity developed within this area is essentially related to the relative movements of two large plates (Africa and Europe) plus an unknown number of smaller continental and oceanic plates;
- Of particular relevance is the existence of an African promontory (called Adria or Apulia), about 1200 km by 400 km wide, elongated

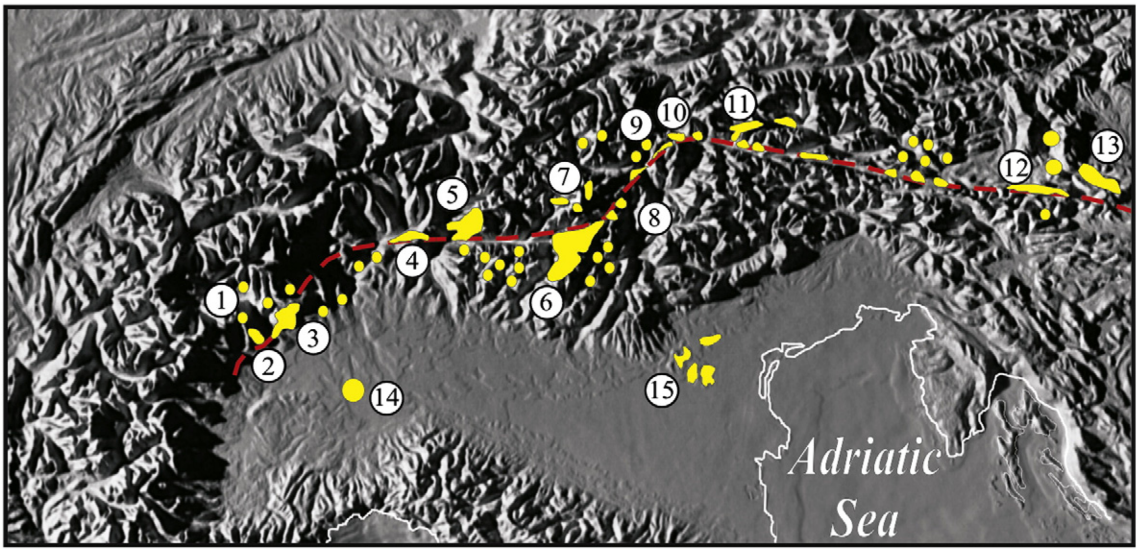

Fig. 3. Digital elevation model of topography and bathymetry in the Alpine region. The main Cenozoic igneous rock outcrops and buried bodies, mostly associated with the Periadriatic dextral transcurrent fault system (dashed line, red in the on-line version of the article) are shown. 1 = Lamprophyric/lamproitic and calcalkaline dykes (~30 Ma); 2 = Traversella (~32–29 Ma); 3 = Biella (~31 Ma); 4 = Novate (~26–24 Ma); 5 = Bergell (~33–26 Ma); 6 = Adamello (~42–27 Ma); 7 = Gran Zebru, Mare and Grunsee; 8 = Rumo and Samoclevo; 9 = Merano; 10 = Rensen (~29 Ma); 11 = Rieserferner (~32 Ma); 12 = Karawanken; 13 = Pohorje (~19 Ma); 14 = Tertiary Mortara Volcano (~30 Ma; buried under ~5000 m thick sediments of the Po Plain); 15 = Veneto Volcanic Province with within-plate geochemical characteristics (~60–25 Ma).

in northwest–southeast direction, that is alternatively considered to be in crustal continuity with the African mainland or separated from the latter by an oceanic plate (called Ionian or Mesogean Ocean);

- Two main mountain chains (plus several other minor belts) developed during the Cenozoic in the CWM, namely the Alps and the Apennines. Both are associated to subduction of oceanic plates followed by continent–continent collision.
- The first orogenesis formed the main Alpine Chain and its continuation in northwest Corsica and in the Betics (southeast Spain). Its formation ranges from Early Cretaceous to Eocene (subduction phase, roughly southeast-directed) and from Eocene–Oligocene to Present (continent–continent collision phase).
- The second orogenesis is termed Apennine–Maghrebide Orogenesis. It started ~50–45 Ma ago when the Alpine Orogenesis was undergoing its continental stage and is associated to a subduction system with polarity (northeastward) nearly opposite to the Alpine subduction system. The Apennine–Maghrebide subduction system involved the recycling of oceanic lithosphere of the Mesogean/Ionian Ocean. The Apennine–Maghrebide Orogenesis is still active, involving both continent–continent collision and oceanic lithosphere subduction.

As briefly anticipated above, the present configuration of the CWM is a direct consequence of the African–Eurasian relative plate motion and their subsequent diachronous collision. The latter was preceded by subduction of neo-Tethyan oceanic lithosphere (the earliest, 'eo-Alpine' phase) essentially lacking evidence of pre-collisional subduction-related igneous activity. The relative motions of Africa, Adria, and stable Eurasia have been attributed to the Early Cretaceous opening of the southern Atlantic Ocean, as indicated by the magnetic anomaly patterns of ocean crust contiguous with continental South America and

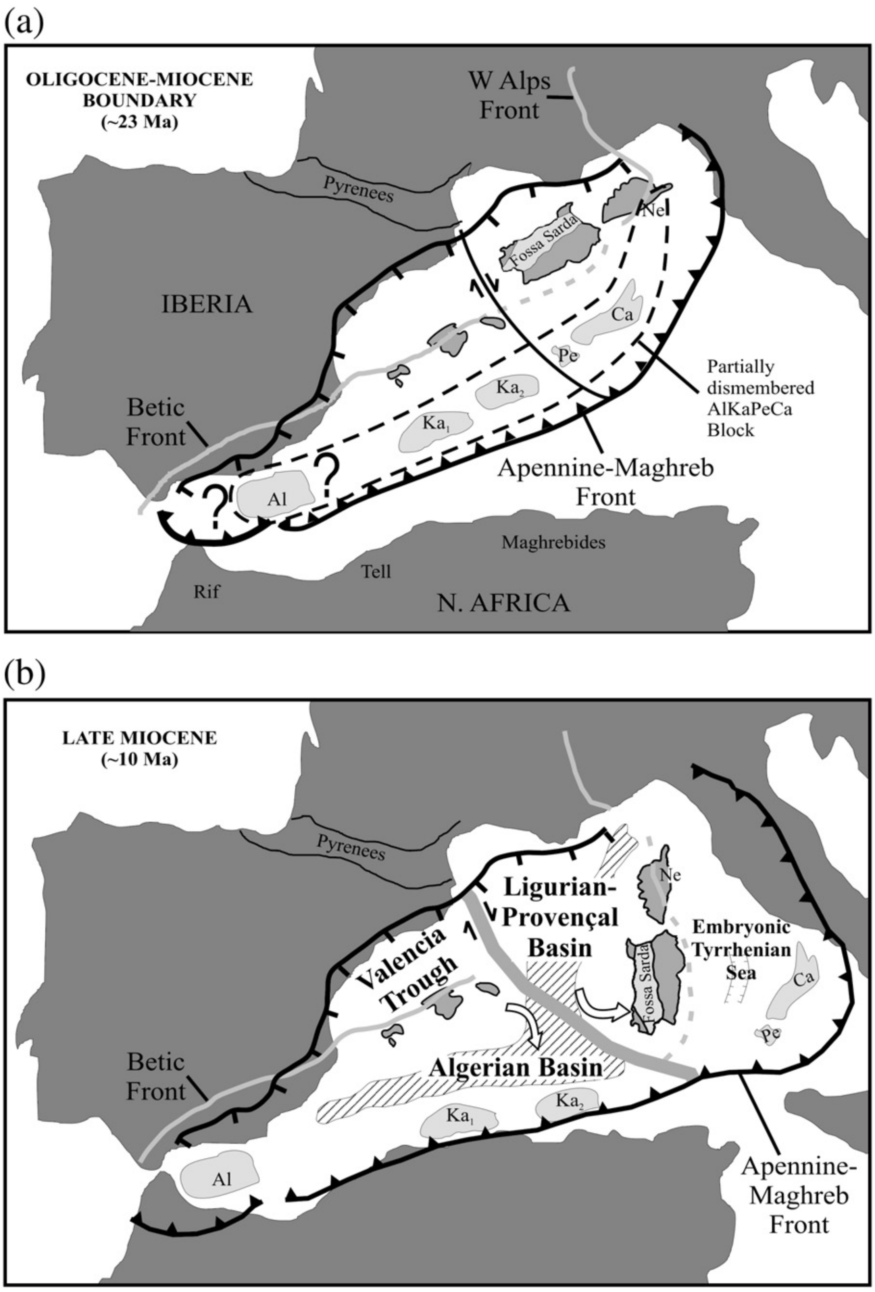

Fig. 4. Simplified geodynamic cartoons illustrating the Late Oligocene–Early Miocene (a) and Late Miocene (b) tectonic evolution of the Sardinia–Corsica block, modified from Gueguen et al. (1998). Al = Alborán Block; Ka_1 = Greater Kabylies; Ka_2 = Lesser Kabylies; Pe = Peloritani Mountains; Ca = Calabria. The Alborán Block palinspastic reconstruction is the least constrained among the various continental blocks. Alternative views indicate this as a completely separated block with respect to the Rif orogen (see text for further details).

southwest Africa, and also signaled by the voluminous Paraná–Etendeka–Angola Large Igneous Province (e.g., Piccirillo and Melfi, 1988; Marzoli et al., 1999; Lustrino et al., 2005; Eagles, 2007).

During its northward drift, Adria overrode the northeastern margin of the Liguride–Piedmontese Ocean (Alpine Tethys) between the Early Cretaceous and Early Eocene (~135–55 Ma). Eocene collision between Adria and Europe was accompanied by local continental subduction, as evidenced by the Early Eocene HP/LT metamorphism of the Briançonnais micro-continent and Sesia–Lanzo continental slice (Dercourt et al., 1986; Rubatto et al., 1998; Stampfli et al., 1998, 2002; Bucher et al., 2004; Fig. 2). The Late Cretaceous–Eocene Alpine eclogites and continental rocks metamorphosed in eclogite-facies (e.g., Berger and Bousquet, 2008, and references therein) have been interpreted as the exhumed remnants of an 'Alpine' branch of the Tethys, including allochtonous slices of extended continental crust subducted southeastwards beneath the Adriatic plate between ~70 and 40 Ma; Berger and Bousquet, 2008; Fig. 2). This process of continent–continent collision and eclogitization continued through the Late Eocene and may be still-ongoing, at least in the eastern Alps. Metamorphic products of the 'meso-Alpine' collision, initiated between the Late Eocene and Early Oligocene, as testified by a Barrovian overprint ranging from very low grade to greenschist and amphibolite facies (Spalla et al., 1996; Bousquet et al., 2008; Fig. 2). The number of micro-continental plates–the Iberia/Briançonnais, Adria, AlKaPeCa, and Sesia–Lanzo–and oceanic basins–such as the Liguride–Piedmontese, Valais and Mesogean Oceans–reflects the diachronous collision of Africa and Eurasia, as also indicated by the regional pattern of tectonic, metamorphic and magmatic events (e.g., Froitzheim et al., 1996; Doglioni et al., 1998; Rubatto et al., 1999; Lustrino, 2000a; Faccenna et al., 2002, 2004; Carminati and Doglioni, 2004; Carminati et al., accepted for publication, and references therein). Tomographic imagery reveals the presence of distinct slabs and slab fragments down to depths of 660 km, where they seem to stagnate and coalesce (e.g., Piromallo and Morelli, 2003).

As also noted above, a large block, 'Adria' or 'Apulia', sometimes referred to as the African 'finger', characterized the northern margin of Africa (e.g., Channell et al., 1979; Stampfli et al., 1998; Wortmann et al., 2001; Oldow et al., 2002), and it has been explained in terms of two competing models. The first model contends that Adria drifted northwards following its detachment from Africa, allowing for opening of the Mesogean Ocean, whose present-day remnant is the Ionian Sea. The exact timing of such detachment is not known basin opening being variously inferred to have occurred during Early Cretaceous (e.g., Dercourt et al., 1986; Schmid et al., 2008), Late Triassic (e.g., Hsu, 1977), or Late Permian/Early Triassic time (e.g., Catalano et al., 2000; Muttoni et al., 2001). There is little or no direct information on the composition of Ionian Sea basement as it is buried under several thousand meters of sediments (e.g., Catalano et al., 2000). An alternative interpretation contends that the Adria microplate was a Florida-like promontory that remained connected to the African mainland (e.g., Muttoni et al., 2001, and references therein) in the region of present-day Tunisia. This model has been supported by pre-Oligocene plate reconstructions whereby the micro-continent was confined to the west by the Ligurian (or Liguride) Ocean and to the southeast by the Ionian Ocean.

Intermittent igneous activity only appeared after total consumption of oceanic lithosphere, near-coeval with HP/LT metamorphism, along the evolving Alpine Chain, along with development of the Periadriatic or 'Insubric' Fault System (Fig. 3). The latter is an orogen-parallel, dextral-transpressive, mylonitic-to-cataclasitic belt marking the boundary between the Southalpine and Austroalpine domains (e.g., Schmid et al., 1989; Carminati et al., accepted for publication, and references therein). This structure crosscuts the upper plate of the Tertiary orogen, eventually reaching the lower horizons of European subducting lithosphere (e.g., Fig. 3 of Schmid et al., 2004). The Alpine subduction system, with north-vergent thrusts and Adria–Africa-verging back-thrusts, is generally viewed as an originally single co-linear system between the Alps and the Betics, passing through Corsica and southeastern Iberia (e.g., Doglioni et al., 1998).

An important change in Africa-Eurasia kinematics occurred during the Late Eocene, when the collision progressed from a 'soft' to a 'hard' phase (e.g., Garzanti et al., 2008). Northwest-southeast-directed movements between Adria and Eurasia shifted to northeast-southwest (e.g., Dewey et al., 1989; Rosenbaum et al., 2002; Aerden and Sayab, 2008), probably marking the 'hard' collision stage reaching the Alps, following total consumption of the neo-Tethyan Liguride–Piedmontese Oceans. Along the eastern margin of Iberia, a block comprising the present-day Alborán, Kabylies, Peloritani Mountains and Calabria (AlKaPeCa) collided with Iberia, following final, Middle Eocene consumption of the Liguride Ocean (Fig. 4). At this point, the Adria–Eurasia suture would have ceased to be a significant plate boundary, as consumption of the remaining Mesogean Ocean would have been initiated via northwest-directed subduction, either in response to 'lateral expulsion' of crustal wedges (e.g., Mantovani, 2005 and references therein) or eastward asthenospheric mantle flow, as proposed by Doglioni et al. (1999b).

Immediately after, or during the last stages of the southeast-directed Alpine subduction, involving consumption of Liguride–Piedmontese oceanic lithosphere beneath Adria, Apennine–Maghrebide subduction began in the embryonic CMW area. The timing of this event has been variously placed in the Late Cretaceous (e.g., Faccenna et al., 1997; Argnani, 2007), Late Eocene (Lustrino et al., 2009), Early Oligocene (e.g., Benedicto et al., 1996), or Late Oligocene (e.g., Doglioni et al., 1999b). Apennine–Maghrebide subduction was initially associated with a northeast–southwest-trending trench along the southeast border of 'Greater Iberia' (as illustrated by animation in Carminati et al., accepted for publication), the latter including Iberia itself plus the Sardinian, Corsican and the AlKaPeCa microplates as located prior to post-Oligocene drifting (Fig. 4).

About 10–15 Myrs after initiation of Apennine–Maghrebide subduction, continental rifting started along the eastern margin of Greater Iberia, as expressed by northwest–southeast trending grabens. The first stages of extension in the upper plate occurred both within the pre-existing Betic Cordillera (i.e., southwest extension of the Alps) and within its foreland along the southern European paleo-margin, with initial opening of the Ligurian–Provençal Basin and the Valencia Trough (Doglioni et al., 1999b; Fig. 4). This suggests that the northwest-directed Apennine subduction was oblique to the pre-existing Alpine–Betic Orogen (Doglioni et al., 1998).

Several 'V-shaped' basins opened in response to back-arc extension in the overriding plate, coeval with roll-back of the Apennine–Maghrebide subduction system (Scandone, 1979; Malinverno and Ryan, 1986; Doglioni et al., 1998; Gueguen et al., 1998; Faccenna et al., 2001, 2004; Sartori, 2004; Schettino and Turco, 2006; Mauffret et al., 2007; Lustrino et al., 2009; Carminati et al., accepted for publication; Figs. 1 and 4). The timing of basin opening decreased from the central-northern area (Ligurian–Provençal Basin; ~30–20 Ma) towards the southwest (Valencia Trough, ~30–20 Ma; Alborán Basin, ~20–0 Ma), the south (Algerian Basin, ~25–20 Ma) and the southeast (Tyrrhenian Sea; ~12–0 Ma; Scandone, 1979; Malinverno and Ryan, 1986; Doglioni et al., 1998; Gueguen et al., 1998; Faccenna et al., 2001, 2004; Sartori, 2004; Schettino and Turco, 2006; Mauffret et al., 2007; Lustrino et al., 2009; Fig. 1).

The Ligurian–Provençal Basin (Fig. 1) opened in response to the ~60° counter-clockwise rotation of Sardinia–Corsica with respect to a pole approximately located in the Gulf of Genoa (e.g., Speranza et al., 2002; Gattacceca et al., 2007, and references therein; Fig. 4) near-coeval with opening of the Valencia Trough, in the latter case the Balearic Rise undergoing a smaller (~10°) clockwise rotation (Schettino and Turco, 2006, and references therein). In the Ligurian–Provençal and Algerian Basins, and east Alborán Basin, the presence of oceanic basement has been inferred from paleomagnetic and seismic evidence, beneath several thousand metres of sediment

(Rollet et al., 2002; Ayala et al., 2003; Schettino and Turco, 2006; Booth-Rea et al., 2007; Duggen et al., 2008 and references therein).

The two other major basins of the CWM region are the Alborán Basin and the Tyrrhenian Sea (Fig. 1). The kinematics of Alborán Basin opening is controversial. One model invokes the westward retreat of the Betic-Rif subduction hinge, facilitated by the presence of pre-existing oceanic lithosphere (e.g., Gutscher et al., 2002; Duggen et al., 2004, 2005, 2008 and references therein), while another suggests that basin opening was an effect of extensional orogenic collapse and radial thrusting around the Gibraltar arc, following post-collisional detachment of a lithospheric root (Platt and Vissers, 1989; Platt et al., 2003b). Yet a third model contends that the Betic and Rif mountain chains developed as independent systems, the first related to the Alpine tectonic system, with northwest-directed transport, and the second to Apennine–Maghrebide tectonics, with south to southwest-directed transport (Carminati et al., accepted for publication).

In the eastern part of the CWM, the Tyrrhenian Sea was initiated in the Middle to Late Miocene as a consequence of southeast-directed rollback of the Apennine subduction hinge, in this case facilitated by the presence of the Mesogean/Ionian oceanic 'corridor' (e.g., Gueguen et al., 1998; Mattei et al., 2007; Carminati et al., accepted for publication).

The origin of the Algerian Basin has been considered to reflect either south-directed subduction rollback, leading to continental collision and south-directed tectonic transport, and thrusting over North Africa, of the Greater and Lesser Kabylies (Fig. 1; e.g., Carminati et al., 1998; Gueguen et al., 1998; Schettino and Turco, 2006) or to east–west opening in response to west- and eastward migration of the Alborán and Calabrian–Peloritani belts, respectively (e.g., Mauffret et al., 2004).

As the northwest-directed Apennine–Maghrebide subduction progressed (Fig. 1), diffuse subduction-related magmatic activity (mostly volcanic) appeared: 1) along the southeast coast of France in Provence (Ivaldi et al., 2003; Beccaluva et al., 2005); 2) onshore and offshore Corsica (Ottaviani-Spella et al., 1996, 2001; Rossi et al., 1998); 3) in Sardinia (Morra et al., 1997; Franciosi et al., 2003; Lustrino et al., 2004a,b, 2008, 2009); 4) within and on either side of the Valencia Trough, on the Spanish mainland and the Balearic Islands (Wadsworth and Adams, 1989; Martì et al., 1992); 5) in the Betic Cordillera (Turner et al., 1999; Duggen et al., 2005, 2008; Doblas et al., 2007; Conticelli et al., 2009a); 6) along the Rifian belt (Northern Morocco and northwest Algeria; Maury et al., 2000; Coulon et al., 2002); 7) along the Tellian belt (Northern Algeria and Tunisia; Kaminskey et al., 1993; Semroud et al., 1994; Belanteur et al., 1995; Fourcade et al., 2001; Talbi et al., 2005) and 8) along the central-southern Apennines (e.g., Peccerillo, 1985, 2005; Beccaluva et al., 1991; Conticelli et al., 2002, 2004, 2007, 2009; Avanzinelli et al., 2009; Carminati et al., accepted for publication).

Late Miocene uplift led to closure of the Mediterranean–Atlantic marine gateways and subsequent desiccation of the Mediterranean Sea (the so-called Messinian Salinity Crisis; Duggen et al., 2003, 2004), matched by a drastic change in C-isotopic composition of Central Mediterranean marine carbonates (e.g., Brandano et al., 2010). This has also been explained as a response to the Africa–Eurasia collision.

3. Igneous activity in the Central-Western Mediterranean

During the Cenozoic, igneous activity initiated in the CWM. This paper focuses on those igneous rocks exhibiting unambiguous subduction-related geochemical character. These are grouped below in terms of igneous provinces. A review of geological, geochronological, petrographical and geochemical data is followed by a discussion of petrological character, and current petrogenetic models. The term 'province' is used to group igneous associations that form a discrete temporal-spatial entity, most likely of common provenance.

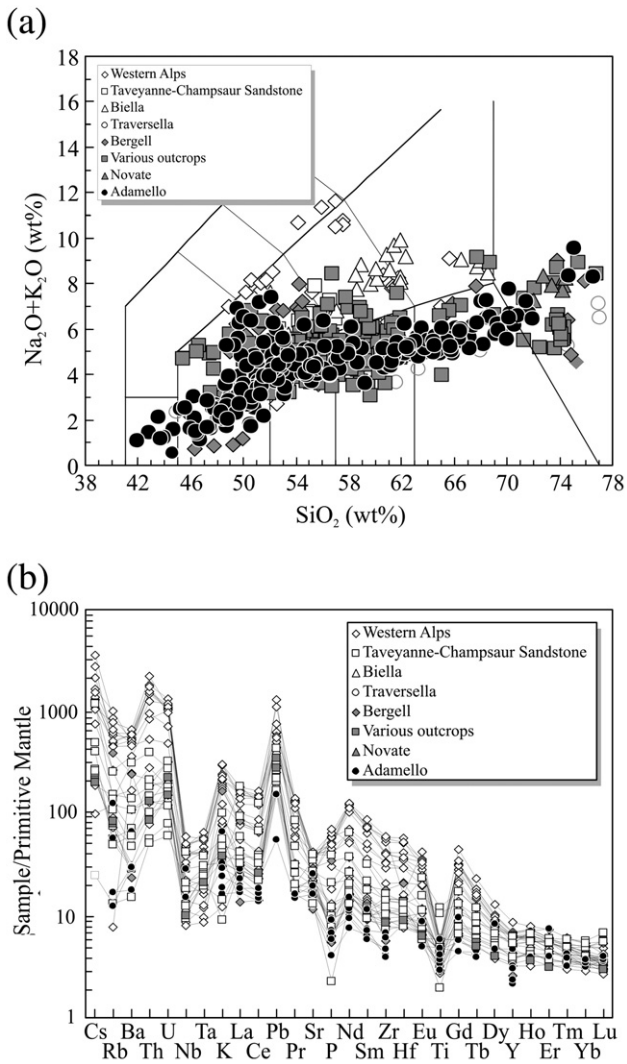

Fig. 5. (a) Total Alkali vs. Silica diagram (after Le Bas et al., 1986) for the igneous rocks of the Periadriatic line (data given in the electronic appendix). (b) Primitive mantle-normalized diagrams (normalization factors after Sun and McDonough, 1989) only for the most mafic (Mg# >0.60) igneous rocks of the Periadriatic line. Data sources are as follows: *Western Alps* (Venturelli et al., 1984; Callegari et al., 2004; Owen, 2008; Conticelli et al., 2009a), *Taveyanne* (Ruffini et al., 1997; Boyet et al., 2001), *Biella* (Bigioggero et al., 1994; Romer et al., 1996; von Blanckenburg et al., 1998), *Traversella* (von Blanckenburg et al., 1998; van Marcke de Lummen and Vander Auwera, 1990), *Bergell* (Diethelm, 1985, 1990; Bellieni et al., 1991; von Blanckenburg et al., 1992, 1998), *Various outcrops* (Beccaluva et al., 1979, 1983; Deutsch, 1984; Dal Piaz et al., 1988; Martin et al., 1993; Altherr et al., 1995; Romer et al., 1996; von Blanckenburg et al., 1998; Mattioli et al., 2002; Trepmann et al., 2004; Macera et al., 2008), *Novate* (von Blanckenburg et al., 1998), *Adamello* (Dupuy et al., 1982; Macera et al., 1983; Ulmer et al., 1983; Bigazzi et al., 1986; Kagami et al., 1991; Blundy and Sparks, 1992; Thompson et al., 2002; Mayer et al., 2003), *Zinsnock* (Bellieni, 1980, Bellieni et al., 1996), *Rieserferner* (Bellieni et al., 1991; Bellieni, unpublished data), *Rensen* (Bigazzi et al., 1986; Bellieni et al., 1991; Bellieni, unpublished data), *Pohorje* (Altherr et al., 1995; Zupancic, 1996; von Blanckenburg et al., 1998).

4. Periadriatic (Insubric) Province

Middle Eocene–Late Oligocene igneous rocks are intermittently exposed along the ~700 km-long 'S' shaped Periadriatic Fault System, striking roughly east–west through the entire Alpine Chain (Fig. 3). According to some authors (Pamic et al., 2002) this province is only part of a larger, ~1700 km-long, magmatic province, that continues eastwards to include the Dinarides and Hellenides. The linear distribution of plutonic rocks is particularly striking, pointing to regional-scale tectonic control on magma ascent. Structural investigations of the plutons and their host lithologies suggest that magma

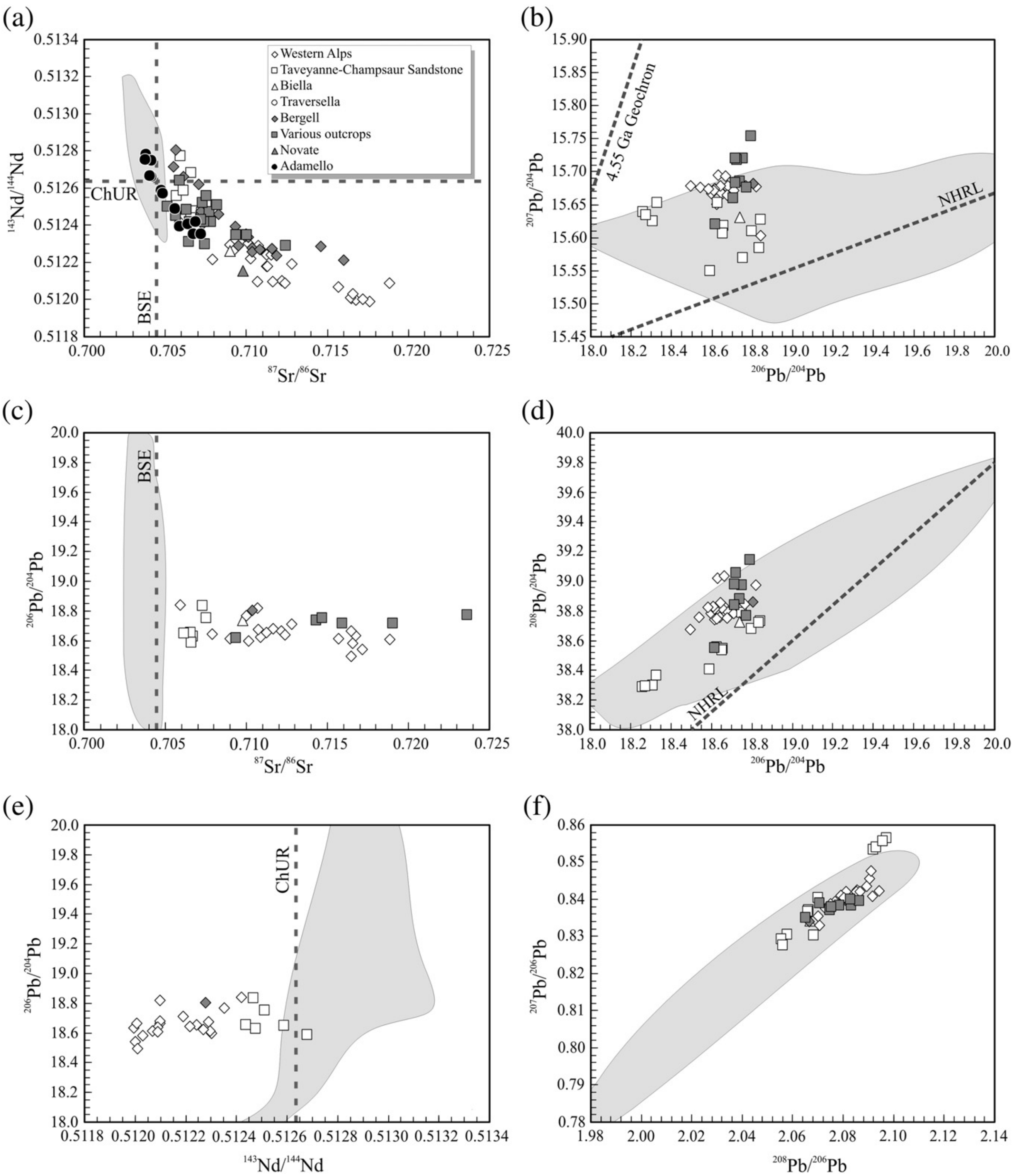

Fig. 6. Plots of (a) ^{87}Sr/^{86}Sr$_{(i)}$ vs. ^{143}Nd/^{144}Nd$_{(i)}$; (b) ^{207}Pb/^{204}Pb$_{(i)}$ vs. ^{206}Pb/^{204}Pb$_{(i)}$; (c) ^{206}Pb/^{204}Pb$_{(i)}$ vs. ^{87}Sr/^{86}Sr$_{(i)}$; (d) ^{208}Pb/^{204}Pb$_{(i)}$ vs. ^{206}Pb/^{204}Pb$_{(i)}$; (e) ^{206}Pb/^{204}Pb$_{(i)}$ vs. ^{143}Nd/^{144}Nd$_{(i)}$; and (f) ^{207}Pb/^{206}Pb$_{(i)}$ vs. ^{208}Pb/^{206}Pb$_{(i)}$ for the Periadriatic plutons and dykes (data given in the electronic appendix). BSE = Present-day Bulk Silicate Earth estimate (^{87}Sr/^{86}Sr = 0.70445). ChUR = Present-day Chondritic Uniform Reservoir estimate (^{143}Nd/^{144}Nd = 0.51264). NHRL = Northern Hemisphere Reference Line (Hart, 1984). The grey field indicates compositions of the Circum-Mediterranean Anorogenic Cenozoic Igneous (CiMACI) rocks (Lustrino and Wilson, 2007). References as for Fig. 5.

ascent occurred during transpressive displacements along the main fault system (Martin et al., 1993; Rosenberg et al., 1995; Berger et al., 1996; Davidson et al., 1996; Steenken et al., 2000; Rosenberg, 2004; Stipp et al., 2004; Wagner et al., 2006). Radiometric ages (Rb/Sr, ^{40}Ar/^{39}Ar, K/Ar, U–Th–Pb) of the igneous rocks range from ~42 to 24 Ma, with a peak in the Early Oligocene (~34–28 Ma; von Blanckenburg and Davies, 1995; Rosenberg, 2004, and references therein). Intrusive activity in the eastern sector of the fault persisted to younger ages (~18–16 Ma; Fodor et al., 2008), but this aspect has been linked to extensional tectonics associated with Pannonian Basin opening (e.g., Trajanova et al., 2008; Fodor et al., 2008, and references therein). Lithologies are otherwise heterogeneous in terms of petrography (plutonics significantly exceeding eruptives), geochemical character (SiO_2-rich exceeding basic), and serial character-abundant calcalkaline and high-K calcalkaline types associated with (relatively minor) potassic and ultrapotassic types such as lamprophyres and lamproites (Venturelli et al., 1984; von Blanckenburg and Davies, 1995; Callegari et al., 2004; Owen, 2008; Prelević et al., 2008, 2010; Conticelli et al., 2009a, and references therein; Fig. 5). In order of decreasing size: the Adamello batholith (~42–27 Ma; Rosenberg, 2004 for review), the Bergell (also known as Bregaglia; ~33–28 Ma; Oberli et al., 2004), the Pohorje (~19–16 Ma; Fodor et al., 2008), the Rieserferner (also known as Vedrette di Ries; ~32 Ma; Romer and Siegesmund, 2003) and the Biella (also known as Valle del Cervo; ~31 Ma; Romer et al., 1996) are the most prominent intrusive bodies. Smaller plutons include the Karawanken, Traversella, Rensen, Altenberg and Zinsnock (also known as Cima di Villa) bodies along with numerous thin tonalitic sheets (or 'lamellae') smeared along the Periadriatic fault. In addition, numerous dykes of similar composition, nearly coeval or slightly younger than the plutons, along the Alpine axial zone, define an elongate east–west area with a significant north–south extension, hence a more diffuse map-scale distribution than that of the plutons (Rosenberg, 2004). Eruptive lithologies have probably been substantially removed by erosion, judging from their abundance within ~40–30 Myr-old clasts in the Alpine molasses (e.g., Ruffini et al., 1997; Brügel et al., 2000). This may be a major factor contributing to the apparent paucity of exposed *in situ* eruptives, as compared to those associated with other orogens (e.g., the European Variscides, North American Cordillera, the Andes, etc.).

The plutons mostly show I-type character with 'TTG' (Tonalite–Trondhjemite–Granodiorite) affinity, and include minor granites, syenites, diorites and gabbros. Many of the Periadriatic plutons represent mantle- and crust-derived melts (e.g., Blundy and Sparks, 1992; von Blanckenburg and Davies, 1995; Tiepolo et al., 2002), as evidenced by the range of $^{87}Sr/^{86}Sr_{(i)}$ isotopic ratios (~0.704–0.716) and $^{143}Nd/^{144}Nd_{(i)}$ values (~0.5128–0.5121; Fig. 6) which, with the exception of the most mafic Adamello samples, plot in the enriched Sr–Nd isotopic quadrant. According to von Blanckenburg and Davies (1995), strong incompatible element enrichments in mafic dykes of the Adamello batholith preclude asthenospheric mantle sources, assuming such enrichments would have been diluted in the convecting mantle regions. This is a classical approach that is based on unconstrained parameters. Indeed, the presence of small percentages of melts in the asthenospheric mantle (as evidenced by the reduction of V_P and V_S in seismic images) cannot be considered as proof of a fully convecting asthenospheric mantle. Very commonly the asthenosphere has been considered to be the source region for MORB (Mid Ocean Ridge Basalt) and therefore considered as isotopically and chemically depleted. Both the implicit assumptions, that the asthenosphere is homogeneous (well-stirred) and depleted, are not necessarily correct, as has been suggested by Anderson (2007).

Exclusively crustal sources are only inferred for rare, very small leucogranitic bodies of the 'Southern Steep Belt' of the Central Alps, immediately to the north and within the Periadriatic Fault mylonites (e.g., Novate leucogranite, Fig. 3; Liati et al., 2000; Ciancaleoni and Marquer, 2006). These S-type granites are interpreted to have formed by fluid-buffered melting during high-grade metamorphism in the Late Oligocene to Early Miocene (~25 Ma; Liati et al., 2000; Berger et al., 2005), hence shortly after the emplacement of the more voluminous I-type magmas. Whereas the latter may be assumed to have formed at the base of a thickened continental crust, and migrated upwards to the surface via mylonitic belts (e.g., as in the Central Alps; Rosenberg et al., 1995; Berger et al., 1996; Rosenberg, 2004), leucogranite magmas are viscous and tend to crystallize within or close to their migmatitic source regions, forming pods and veins, and rarely (e.g. Novate Granite) stocks of more than a few km^2 in diameter.

Despite an extensive range of geochemical and petrographic character, the $^{206}Pb/^{204}Pb_{(i)}$ isotopic ratios cluster around 18.6–18.8, irrespective of the large range of Sr–Nd values (Fig. 6). Only the volcaniclastic sandstones of Taveyanne (northwest Alps) show significantly lower $^{206}Pb/^{204}Pb_{(i)}$ values (~18.3; Boyet et al., 2001); Sr–Nd isotopic data are unfortunately not available for these samples. $^{207}Pb/^{204}Pb_{(i)}$ ratios for both Periadriatic plutons and dykes fall within ~15.55 to ~15.75, whereas $^{208}Pb/^{204}Pb_{(i)}$ ratios range between ~38.3 and ~39.2 (no data published for the Adamello massif to the best of the authors' knowledge). All the samples plot to the right of the 4.55 Ga geochron, well above the Northern Hemisphere Reference Line (HNRL; Hart; 1984), thereby conforming to the definition of DUPAL affinity. Hart (1984) originally believed that the DUPAL anomaly was confined to oceanic lithosphere in the southern hemisphere. It was commonly defined to the Δ7/4 and Δ8/4 values which represent the vertical deviation from the best fit of the Northern Hemisphere oceanic basalts, defining the NHRL. Positive Δ7/4 and Δ8/4 values indicate higher $^{207}Pb/^{204}Pb$ values for a given $^{206}Pb/^{204}Pb$, compared to the NHRL. The Periadriatic plutons show Δ7/4 and Δ8/4 values ranging from ~+4 to ~+80 and from ~+30 to ~+87, respectively. Discussion of the various models proposed to explain the origin of such vertical shifts is outside the scope of this paper, but we note that the DUPAL anomaly has been variously ascribed to recycling of ancient sediments (Hawkesworth et al., 1986), delamination of cratonic lithospheric mantle (e.g., Mahoney et al., 1992) and/or lower continental crust (e.g., Escrig et al., 2004).

Apart from the Periadriatic calcalkaline lithologies (representing the bulk of magmatic activity in the region), some dykes of 'exotic' chemical and mineralogical character intrude the plutons and appear to reflect a progressive increase in K content from the southeastern (mostly tholeiitic to calcalkaline) to the Central Alps (mostly high-K calcalkaline) and further to the northwest (mostly shoshonitic to ultrapotassic), as observed by Beccaluva et al. (1983). Despite its obvious significance, the petrogenetic affinity of the potassic and ultrapotassic dykes in the Alps remains controversial. They have indeed been referred to as lamprophyres (minettes, spessartites, kersantites; e.g., Venturelli et al., 1984; Owen, 2008) or lamproites (e.g., Peccerillo and Martinotti, 2006; Prelević et al., 2010). While preferring the term 'lamproite', Conticelli et al. (2009a) agree with Owen (2008) that they could strictly speaking be classified as lamprophyres from a mineralogical and chemical point of view. The potassic dykes of the northwestern Alps are mostly mafic in composition (e.g., MgO ~6–16 wt.%) and show high Cr and Ni content, indicating that they probably equilibrated with a peridotitic source. Thus, their unusually high K_2O and LILE content, coupled with radiogenic Sr ($^{87}Sr/^{86}Sr_{(i)}$ ~0.706–0.719) and the unradiogenic Nd ($^{143}Nd/^{144}Nd_{(i)}$ ~0.5120–0.5124; Fig. 6) isotopic compositions may be reasonably considered to be mantle source characteristics rather than acquired within the magma supply system (e.g., Venturelli et al, 1984; Peccerillo and Martinotti, 2006; Owen, 2008; Conticelli et al., 2009a; Prelević et al., 2010). The lead isotopes plot in the field defined by the other Periadriatic plutons, with lamproites clustering towards less radiogenic $^{206}Pb/^{204}Pb_{(i)}$ isotopic ratios (Fig. 6). Overall, the northwestern Alpine dykes (high-K calcalkaline, shoshonitic and lamprophyric/lamproitic) show incompatible trace element patterns enriched in Rb, Th, U, Pb and LREE and depleted in HFSE (Nb, Ta, Hf, Zr and Ti) in primitive mantle-normalized diagrams (Fig. 5). The entire geochemical and isotopic data set for the more primitive compositions (Fig. 5; dykes and plutonic rocks) suggests a relatively enriched mantle source for these magmas. Such features are common among post-collisional magmas, reflecting an ubiquitous crustal component in their mantle sources (Prelević et al., 2008).

Key trace element ratios (e.g., low Ce/Pb, Nb/U and U/Th) have been interpreted by several authors to suggest sediment input into the mantle (e.g., Venturelli et al., 1984; Peccerillo and Martinotti, 2006; Owen, 2008; Conticelli et al., 2009a), as proposed for many lamproites world-wide (e.g., Foley et al., 1987; Conticelli, 1998; Turner et al., 1999; Murphy et al., 2002; Duggen et al., 2005; Prelević et al., 2005). Input of crustal components would have occurred during the eo-Alpine phase (Early Cretaceous–Middle Eocene), prior to the Europe–Adria collision.

The apparent paucity of subduction-related igneous activity during the eo-Alpine orogenic phase and the fact that such activity is largely confined to the earliest stages of continental collision, remains a matter of debate. Von Blanckenburg and Davies (1995) explained this as an effect of relatively slow, pre-Oligocene Adria–Eurasia convergence (<1 cm/yr). Their 'slab break-off' model may be able to explain both the linear distribution of Periadriatic intrusives, the brief and nearly coeval emplacement of most plutons, and their apparent mantle-related magmatic sources. A near-horizontal rupture in a syn-collisional (post-subduction) down-going slab is best explained as due to the low density, hence difficulty in subducting, of continental crust, generating tensional forces in the relatively high-density oceanic lithosphere. A tendency for slab tearing along the transition between subducting continental and oceanic crust is thus to be expected (von Blanckenburg and Davies, 1995).

Interestingly, a 'slab break-off' model has also been proposed by Macera et al. (2008) to explain the Late Paleocene–Oligocene intraplate basalts of the Veneto Province with geochemical character, exposed in the southeastern Alps, both to the north and south of the Adamello batholith (Fig. 3). This activity (~65–25 Ma) pre-dates or is mostly coeval with the bulk of the syncollisional Periadriatic plutons. Whereas Periadriatic magmatism occurred over a wide area during a relatively short time (~12 Myr), the activity of the Veneto Province is confined to a small area and lasted much longer (~35–40 Myr).

Geodynamic models of the Periadriatic and Veneto magmas are by no means unequivocal, having varied significantly during the past thirty years. Deep-seated melting of the subducting European plate was originally suggested to be the cause of slab break-off and also the source of the uprising Periadriatic Plutons (Dietrich, 1976), implying that one and the same orogenic source fed both magmatic systems. This interpretation was followed by the proposal of a temporary hiatus in convergence during the Oligocene (an 'orogenic lull'), expressed by significant extension within the Alpine orogen, allowing for decompression melting and emplacement of the Periadriatic plutons (Laubscher, 1983). Structural studies of these plutons and their host lithologies in the early 1990s showed that these magmas were emplaced during north-south shortening within the axial part of the Alpine chain (e.g., Rosenberg, 2004, and references therein). This scenario was both compatible and probably contemporaneous with the slab break-off model of von Blanckenburg and Davies (1995).

The latter model dominate current petrogenetic interpretations, and is validated by seismic tomographic imagery (Lippitsch et al., 2003). Model variants may be needed to explain asthenospheric mantle sources of differing depth, for example, in the Veneto volcanic Province (Macera et al., 2003, 2008). In contrast to von Blanckenburg and Davies (1995), Macera et al. (2008) suggested that a subducting slab might be weakened by a small-scale mantle plume, facilitating its rupture and the ascent of 'plume-type' magmas. Moreover, they provide a simple explanation for the association of 'anorogenic' (intraplate) magmatism of Veneto in a post-subduction orogenic setting. On the other hand, following Houseman et al. (1981), von Blanckenburg et al. (1998) pointed out that thermal boundary layer detachment (TBLD) could equally explain the formation of deep-seated thermal anomalies capable of generating Periadriatic-type calcalkaline magmas. Such delamination would be expected to result in widespread magmatic activity, possibly inconsistent with the linear distribution of the Periadriatic plutons.

Application of the break-off model to the Veneto Province implies a transition from subduction-related (calcalkaline) to intraplate type activity (Macera et al., 2003), prior to and during slab detachment, the resulting 'window' providing a pathway for asthenospheric magmas to the surface. This is inferred to have occurred between 55 and 45 Ma (Macera et al., 2008), in contrast to geological indications (Barbieri et al., 1981, 1991). The absolute northward displacement rate of Africa between 60 and 30 Ma amounts to about 3 cm/yr (O'Connor and le Roex, 1992). If similar rates are assumed for motion of the Adriatic plate it is difficult to explain such a stationary location for a single magma source, rather than its expected southward migration. Therefore, such models are, by no means, exclusive in explaining Tertiary Alpine magmatic patterns.

5. Betic–Rifian Province

During the transition from Mesozoic to Cenozoic time, the Betic–Rif Cordillera (Fig. 1) began to acquire its present-day arc-like shape (e.g., Michard et al., 2002; Martinez-Martinez et al., 2006). This is represented by a Cenozoic thrust-and-fold belt, whose tectonic transport ranges from northwest to west and southwest. The internal units of these two belts, those closer to the Mediterranean, share many similarities, as represented by the Alborán domain, a presumed fragment of the now dismembered AlKaPeCa micro-continent ('Al' denoting the Alborán terrane; Bouillin et al., 1986; Fig. 4). This micro-continent is believed to have migrated south-westward, concomitant

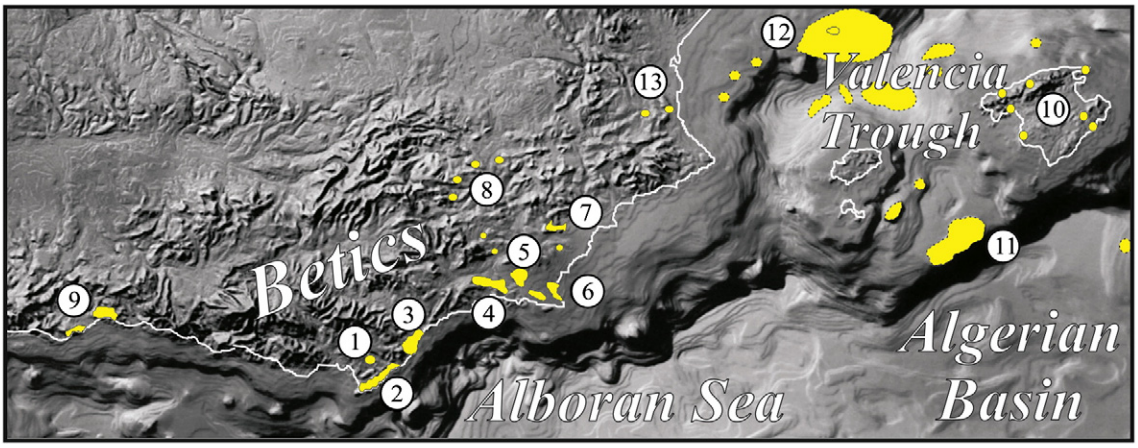

Fig. 7. Digital elevation model of the Betics–Valencia Trough and Balearic Islands region showing the main occurrences, onshore and offshore, of Cenozoic igneous activity. 1 = Cerro El Hoyazo (~9.7–6.2 Ma); 2 = Cabo de Gata (~14.4–6.6 Ma); 3 = Vera (~7.5–6.4 Ma); 4 = Mazarron (~8.9–6.8 Ma); 5 = Tallante–Cartagena (~2.9–2.3 Ma); 6 = Mar Menor (~18.5 Ma); 7 = Zeneta–Fortuna (~8.1–7.1 Ma); 8 = Jumilla, Cancarix, Calasparra, Mula, Barqueros (~8.1–6.8 Ma); 9 = Malaga (~33.6–17.4 Ma); 10 = Mallorca (volumetrically scarce and scattered Miocene pyroclastic rocks); 11 = Emile Baudot Seamount (with whithin-plate geochemical characteristics; ~1.5 Ma); 12 = Columbretes Islands volcanic field (with whitin-plate geochemical characteristics; ~1–0.3 Ma); 13 = Cofrentes and Picasent (with within-plate geochemical characteristics; ~2–1.3 Ma). Submarine outcrops are bounded with broken lines and their areal extent is only roughly estimated.

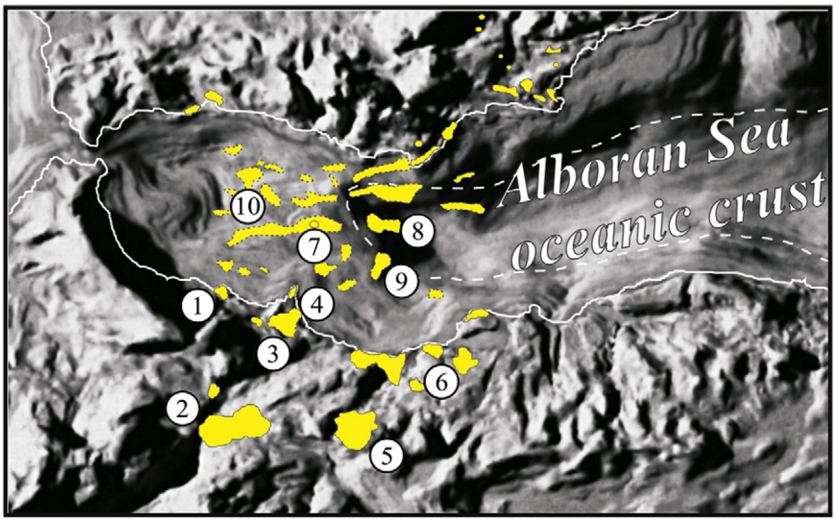

Fig. 8. Digital elevation model of the Alborán Sea-Rif region showing the main onshore and offshore Cenozoic igneous rock associations. 1 = Ras Tarf (~15–12 Ma); 2 = Guilliz (~6.9–0.6 Ma); 3 = Gourougou (~7.6–3.7 Ma); 4 = Trois Furches (~6.6–2.6 Ma); 5 = Oujda (~5.6–1.5 Ma); 6 = Oranie (~11.0–0.8 Ma); 7 = Alborán Ridge and Alborán Island (9.6–7.5 Ma); 8 = Al Mansour Seamount (8.7 Ma); 9 = Yusuf Ridge (10.7 Ma); 10 = Djibouti Bank.

with the opening of the Valencia Trough and Ligurian–Provençal Basin (Fig. 4). Tectonic units of the Alborán terrane are thrust southwestward over the Maghrebian Flysch nappes in the Rif and northwestward in the central-eastern Betics onto the Iberian margin (e.g., Michard et al., 2002; Martinez-Martinez et al., 2006). The present-day boundary between Africa and western Eurasia appears to correspond to the transpressional dextral Azores–Gibraltar fault (e.g., McClusky et al., 2003), although some authors (e.g., Mantovani et al., 2007) suggest the involvement of several separate microplates (including, for example, the Morocco micro-plate, as distinct from the larger African–'Nubian'-plate). In both the Betic and Rif Belts, a rapid phase of extension during the early Miocene was followed by rapid unroofing (>3 km/Myr) as recorded in the peridotite-dominated core complexes of Ronda (southern Spain) and Beni Bousera (northern Morocco; e.g., Zeck, 1997). An alternative geodynamic model (Doglioni et al., 1999a) contends that there is no direct tectonic relation between the Betic and Rif Belts. According to this model, the Betics represent a southwestward prolongation of an originally southeast-dipping Alpine subduction front, whereas the Rif (and Tell) belts would be part of the northward-dipping Apennine subduction front, possibly connected with the Italian Apennine Chain. This model is illustrated in an animated reconstruction of the last 50 Myr geodynamic evolution of the CWM in Carminati et al. (accepted for publication).

Three main groups of models were proposed for the origin and evolution of the Betic-Rif region (e.g., Michard et al., 2002; Doblas et al., 2007): the first model suggests ensialic development of a collisional orogen, unrelated to oceanic subduction (e.g., Turner et al., 1999; Platt et al., 2003b; Doblas et al., 2007, and references therein); a second model involves the development of a collisional orogen along a single northwest-dipping subduction zone, linking the Betics to the northern Apennines (e.g., Lonergan and White, 1997 and references therein); and a third group includes the development of a collisional orogen involving Cretaceous–Eocene southeast-dipping subduction (a southwestward prolongation of the eo-Alpine subduction beneath the Western Alps), followed by Oligocene to Present subduction (with notable changes in subduction polarity and tectonic transport associated with Apennine–Maghrebian subduction), separated by an Alborán micro-continent (e.g., Duggen et al., 2004, 2005, 2008; Guerrera et al., 2005, and references therein). Duggen et al. (2005, 2008) proposed a synthetic solution that combines westward roll-back of an east-dipping oceanic slab, causing sub-continental lithospheric mantle to 'peel off' beneath the margins of southern Iberia and northwest Africa ('continental-edge' delamination). Oblique extension within the Alborán Basin, would have cross-cut the Betic orogen and developed both in hinterland (the Alborán Sea) and foreland (the Valencia and Algerian Basins), the latter considered as evidence for the existence for two independent phenomena (Doglioni et al., 1999a).

In the following, we focus on the geochemical character of the Late Eocene to Early Pliocene (~38 to ~4.5 Ma) magmas emplaced during the evolution of the Betic-Rif Belt (e.g., Turner et al., 1999; Duggen et al., 2004, 2005, 2008; Doblas et al., 2007; Conticelli et al., 2009a; Figs. 7 and 8), while avoiding consideration of the Late Miocene to Quaternary intraplate lavas, whose petrogenesis has been discussed in detail elsewhere (Lustrino and Wilson, 2007; Cebriá et al., 2009; Duggen et al., 2009, and references therein).

An Early Miocene dyke swarm of arc-tholeiitic affinity is exposed in the Malaga area, southwest of the Betics. This has been interpreted to have formed in a pre-collisional, subduction-related (possibly back-arc) extensional setting (Torres-Roldàn et al., 1986; Duggen et al., 2004; Fig. 9). Recent $^{40}Ar/^{39}Ar$ age data point to a bimodal age distribution, $^{40}Ar/^{39}Ar$ ages reported by Turner et al. (1999) and Duggen et al. (2004) lying between ~38 and 30 Ma being interpreted as intrusive ages. K/Ar and additional $^{40}Ar/^{39}Ar$ ages (Torres-Roldàn et al., 1986, Turner et al., 1999; Duggen et al., 2004), between ~23 and 18 Ma, are inferred to reflect thermal resetting in response to metamorphism (Duggen et al., 2004). The latter may be related to Early Miocene collision of the Alborán terrane with Iberian margin (Duggen et al., 2004). Igneous rocks of strikingly different composition (most likely of crustal origin) were also emplaced in the Early Miocene. These include: 1) cordierite-bearing dacites at the Mar Menor in southern Spain (~18.5 Ma $^{40}Ar/^{39}Ar$ age; Duggen et al., 2004), 2) leucogranites (~20.4–18.8 Ma, Rb–Sr age; Zeck et al., 1989), and 3) anatectic dykes associated with Early Miocene emplacement of ultramafic massifs such as Ronda; (e.g., Platt et al., 2003a).

For a better understanding of Miocene to Recent igneous activity in this area it is useful to distinguish between magmatism in the Alborán Basin from that along the continental margins of Iberia and Africa. The former includes Early to Late Miocene activity in the present-day Alborán Sea (as represented by Alborán Island and basement highs such as the Alborán Ridge, Djibouti Bank, Cabliers Banks, Yusuf Ridge and the Al Mansour Seamount), and in uplifted and eroded volcanic complexes in southern Spain (Cabo de Gata, Aguilas block), Morocco (Ras Tarf, Trois Furches) and Algeria (M'Sirda and Sahel d'Oran; see Fig. 1 in Duggen et al., 2008 for an overview). Duggen et al. (2004, 2008), Fig. 9) showed that the Alborán

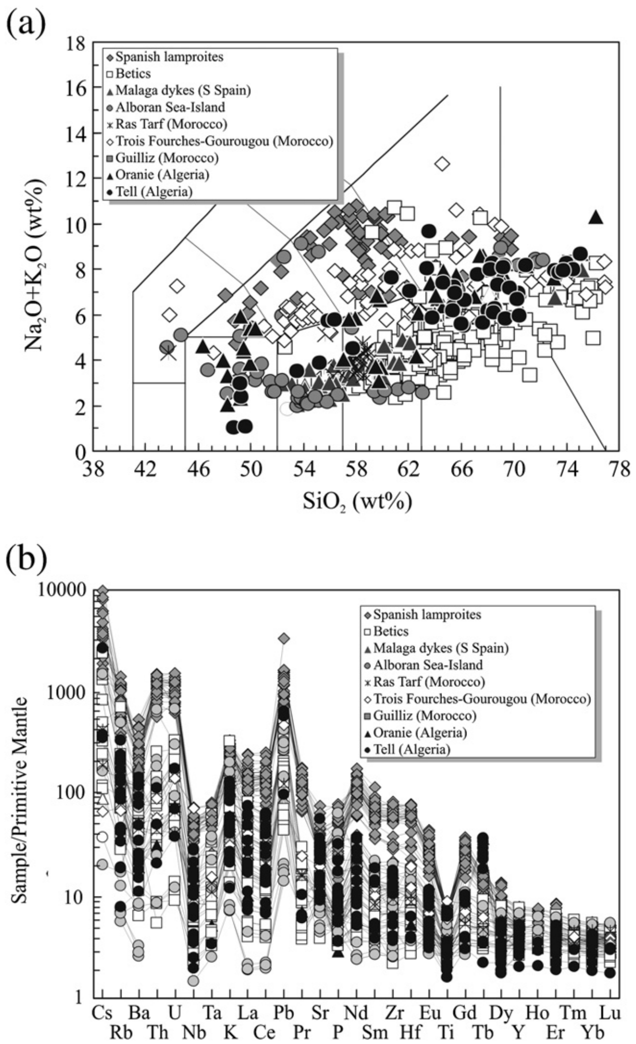

Fig. 9. (a) Total Alkali vs. Silica diagram (after Le Bas et al., 1986) for subduction-related igneous rocks of the Betics–Rif–Tell region (data given in the electronic appendix). (b) Primitive mantle-normalized incompatible element diagram (normalization factors after Sun and McDonough, 1989) only for the most mafic (Mg# > 0.60) igneous rocks of the Betics–Rif–Tell region. Data sources are as follows: *Spanish lamproites* (Nixon et al., 1984; Venturelli et al., 1984; Nelson et al., 1986; Toscani et al., 1995; Benito et al., 1999; Turner et al., 1999; Duggen et al., 2005; Prelević and Foley, 2007; Conticelli et al., 2009a), *Betics* (Munksgaard, 1984; Toscani et al., 1990; Fernandez-Soler, 1996; Turner et al., 1999; Zeck et al., 1998, 1999; Duggen et al., 2004, 2005; Conticelli et al., 2009a), *Malaga dykes* (Torres-Roldàn et al., 1986; Turner et al., 1999; Duggen et al., 2004), *Alborán Sea/Island* (Fernandez-Soler, 1996; Gill et al., 2004; Duggen et al., 2008), *Ras Tarf* (El Bakkali et al., 1998; El Azzouzi et al., 1999; Duggen et al., 2004), *Trois Fourches-Gourougou* (El Bakkali et al., 1998; El Azzouzi et al., 1999; Duggen et al., 2004, 2005), *Guilliz* (Duggen et al., 2009), *Oranie* (Hernandez and Lepvrier, 1979; Belanteur et al., 1995; Louni-Hacini et al., 1995; Coulon et al., 2002), *Tell* (Semroud et al., 1994; Fourcade et al., 2001).

Basin volcanism can be subdivided into a keel-shaped zone of LREE-depleted (mainly tholeiitic) lava series in the centre of region (including Alborán Island, the Yusuf Ridge, and Al Mansour Seamount, dated at ~12.1–8.7 Ma) and an arcuate zone of LREE-enriched (mainly calcalkaline) lavas (~15–6 Ma) surrounding the LREE-depleted keel-shaped zone. The LREE-enriched arc is mostly sub-parallel to the Betic–Rif orogen, such that the consequent geochemical zonation complements the distinctive north-south symmetry and west-east asymmetry of the Alborán system. The Miocene LREE-depleted, mainly tholeiitic central Alborán Sea lavas are typical of a frontal arc and appear to reflect Miocence infiltration of a depleted source by slab-derived hydrous fluids (Fig. 9). The origin of the Miocene LREE-enriched, mainly calcalkaline, magmas is still unclear but may either result from 1) crustal contamination of LREE-depleted, arc-tholeiites like those from the central Alborán Sea or 2) LREE-enriched, mantle-derived calcalkaline melts variably contaminated by crustal material during their ascent (Duggen et al., 2008). Based on the geochemical data for igneous rocks in southeastern Spain, Turner et al (1999) argued against subduction beneath the Betics and attributed the calcalkaline rocks to convective removal of a lithospheric keel (following Houseman et al., 1981, and others). This model, now considered questionable by many workers, invokes shallow-level crustal contamination of adiabatic partial melts of MORB-like affinity in developing 'secondary' calcalkaline geochemical character. However, based on major and trace element and O-Sr-Nd-Pb-isotope data, most of the Alboran Sea lavas are clearly not anatectic melts (Duggen et al., 2004, 2008), as proposed by Zeck et al. (1998), and cannot represent crustal-contaminated MORB melts as instead proposed by Turner et al. (1999). The geochemistry of the Alborán Basin lavas and their spatial-temporal distribution is difficult to explain with detachment/delamination geodynamic models but are more plausibly explained in terms of east-directed Miocene subduction of oceanic lithosphere — see Duggen et al. (2008) for a more detailed discussion of geodynamic implications of Alborán Sea volcanism.

Late Miocene to Early Pliocene igneous activity at the margins of southern Iberia and northwest Africa is mostly of high-K calcalkaline type, with shoshonites and ultrapotassic (also lamproitic) compositions being significantly more potassic than those from the Alborán Basin (Fig. 9). These K-rich magmas are often associated with the younger, Late Miocene to Quaternary, intraplate basanites, alkali basalts and their respective derivatives (e.g., Turner et al., 1999; Coulon et al., 2002; Duggen et al., 2005; Cebriá et al., 2009). While conforming to the transition from subduction-related and K-rich orogenic magmas to magmas of intraplate geochemical affinity (as observed elsewhere), the latter present an as yet unresolved problem. The temporal change in geochemical composition has been attributed to a major re-organisation of Western Mediterranean upper mantle structure during the Miocene to Pliocene transition (Duggen et al., 2003, 2005). The origin of Late Miocene to Early Pliocene potassic magmas in southwest Spain (Murcia, Cartagena, Vera and Hoyazo; ~8.1–6.3 Ma) and northwest Africa (Gourougou, Guilliz in northern Morocco and M'Sirda and Sahel d'Oran in northwest Algeria; ~11–4.8 Ma) has commonly been linked to partial melting of sub-continental mantle lithosphere, metasomatized during Miocene or earlier subduction (for example that in the Alborán Basin or the Late Cretaceous to Oligocene Alpine orogeny; Nelson, 1992; Louni-Hacini et al., 1995; El Bakkali et al., 1998; Benito et al., 1999; Turner et al., 1999; Coulon et al., 2002; Duggen et al., 2003, 2005; Gill et al., 2004; Conticelli et al., 2009a).

Magmas emplaced in the Betic, Alborán Basin, and Rif regions show a larger range of Sr–Nd isotopic compositions compared with the Periadriatic plutons and dykes, $^{87}Sr/^{86}Sr_{(i)}$ ranging between ~0.7044 and ~0.7023 and $^{143}Nd/^{144}Nd_{(i)}$ between 0.5132 and 0.5120 (Fig. 10). As for the Periadriatic Province, lamproites in the Murcia, Almeria and Albacete areas are characterized by strongly radiogenic Sr and unradiogenic Nd, their isotopic compositions overlapping with those of crustal origin in the Tell Atlas (see below). Another similarity with the subduction-related Periadriatic magmas is the relatively homogeneous $^{206}Pb/^{204}Pb_{(i)}$ isotopic composition, which fall mostly between 18.70 and 18.85 (Fig. 10). Only few samples from Guilliz (Morocco) reach $^{206}Pb/^{204}Pb_{(i)}$ values as high as 19, whereas the few samples from Oranie (northwest Algeria) exhibit relatively low $^{206}Pb/^{204}Pb_{(i)}$ (18.70–18.54). Ratios of $^{207}Pb/^{204}Pb_{(i)}$ range from ~15.58 to ~15.74, while $^{208}Pb/^{204}Pb_{(i)}$ ranges from ~38.6 to 39.4 (Fig. 10). These samples are therefore characterised by positive Δ7/4 and Δ8/4 values (+6 to +21 and +22 to +85, respectively).

Duggen et al. (2003, 2004, 2005) explained K-rich magma genesis at the southern Iberian and northwest African margins in terms of perturbation and partial melting of metasomatized, sub-continental lithosphere, associated with geodynamic evolution during the

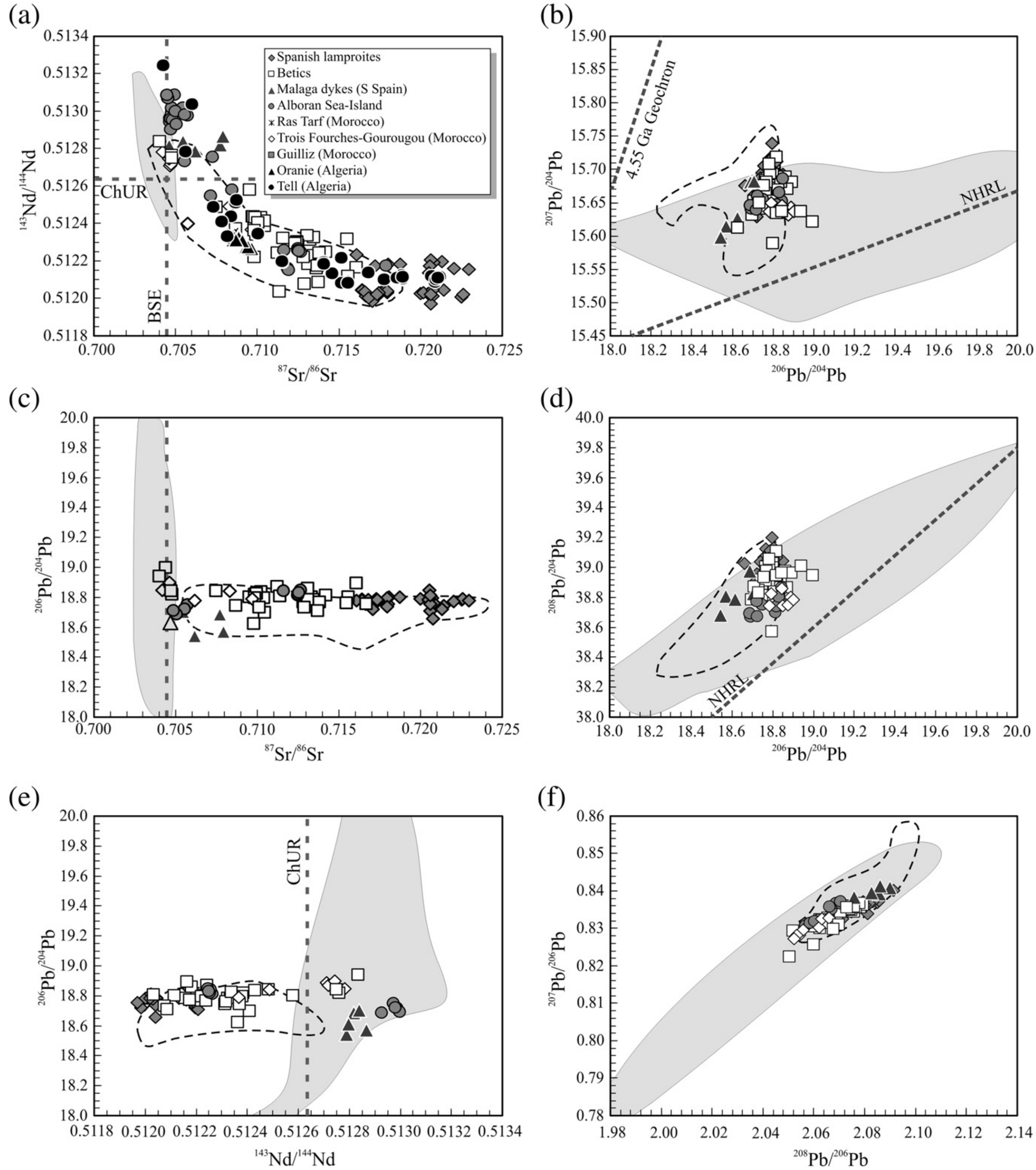

Fig. 10. Plots of (a) $^{87}Sr/^{86}Sr_{(i)}$ vs. $^{143}Nd/^{144}Nd_{(i)}$; (b) $^{207}Pb/^{204}Pb_{(i)}$ vs. $^{206}Pb/^{204}Pb_{(i)}$; (c) $^{206}Pb/^{204}Pb_{(i)}$ vs. $^{87}Sr/^{86}Sr_{(i)}$; (d) $^{208}Pb/^{204}Pb_{(i)}$ vs. $^{206}Pb/^{204}Pb_{(i)}$; (e) $^{206}Pb/^{204}Pb_{(i)}$ vs. $^{143}Nd/^{144}Nd_{(i)}$; and (f) $^{207}Pb/^{206}Pb_{(i)}$ vs. $^{208}Pb/^{206}Pb_{(i)}$ for the Betics–Rif–Tell region volcanic and plutonic rocks (data given in the electronic appendix). BSE = Present-day Bulk Silicate Earth estimate ($^{87}Sr/^{86}Sr = 0.70445$). ChUR = Present-day Chondritic Uniform Reservoir estimate ($^{143}Nd/^{144}Nd = 0.51264$). NHRL = Northern Hemisphere Reference Line (Hart, 1984). The grey field indicates compositions of the Circum-Mediterranean Anorogenic Cenozoic Igneous (CiMACI) rocks (Lustrino and Wilson, 2007). The dashed line delineates the field of the Periadriatic igneous rocks. References as for Fig. 9.

Miocene to Pliocene time. Other authors have also argued that such magmas on the northwestern African margin result from partial melting of continental lithosphere, either in response to extensional and compressional strike-slip faults (e.g. El Bakkali et al., 1998) or linked to the detachment of subducted oceanic lithosphere following hot asthenospheric upwelling (e.g., Coulon et al., 2002). Duggen et al. (2005) further proposed that 'continental-edge' delamination of sub-continental lithosphere associated with 'upwelling' of sublithospheric mantle accounts for the close spatial-temporal association of K-rich, HFSE-depleted and HFSE-enriched melts, and the occurrence of hybrid magmas of subduction-related and intraplate magmas (e.g., in the Gourougou and Guilliz volcanic centres).

6. Valencia Trough Province

The Valencia Trough, lying offshore from southeast Spain (Fig. 1) shares more tectonic and magmatic features with the Ligurian–Provençal Basin than with the Alborán Sea and Betic–Rif Belt. Both the Valencia and Ligurian–Provençal Basins were clearly initiated as back-arc basins, developed above the hanging-wall of the retreating Apennine–Maghrebide subduction system (e.g., Gueguen et al., 1998; Carminati and Doglioni, 2004). During the Early to Middle Miocene (~24–19 Ma), calcalkaline andesitic-dacitic volcanism developed within central to eastern sectors of the Valencia Trough and in the island of Mallorca (Wadsworth and Adams, 1989; Martì et al., 1992). Some 1200 m of these lavas — buried beneath more than 2000 m of sediments northeast of the Columbretes Archipelago (about half-way between Valencia and Mallorca) were sampled by ocean floor drilling (Lanaja, 1987). Despite the huge volume of magma in evidence, there are only few data pertaining to their petrographic, geochemical and Sr–Nd–Pb isotopic character (Martì et al., 1992). As for the Betic-Rif region (and elsewhere), this area also reflects a transition from 'subduction-related' to 'intraplate-type' magmatic character (e.g., Lustrino and Wilson, 2007; Lustrino et al., 2007b, and references therein).

7. Tell Province

East of the Rif thrust-and-fold belt zone, the northern coast of Algeria and Tunisia is characterized by a series of discrete Miocene–Pliocene volcano-plutonic complexes. These were mostly emplaced following a major episode of Alpine compression (~15–20 Ma), post-dating Tethyan oceanic crust subduction and the Middle to Late Miocene southward transport of the Tellian nappes (e.g., Mickus and Jallouli, 1999; Benaouali-Mebarek et al., 2006, and references therein). According to recent seismologic and structural studies, the present north Algerian margin could mark the locus of new subduction, showing opposite polarity to that of subduction responsible for the southward-directed emplacement of the Tellian and Kabylian terranes (e.g., Billi et al., 2007; Kherroubi et al., 2009). In other words, after southward tectonic transport related to the opening of the Northern Algerian Basin, the Maghreb margin would be expected to accommodate southward subduction of thin Neogene (possibly oceanic) Northern Algerian Basin lithosphere beneath thicker continental lithosphere of the Tell.

The Tell is a roughly WSW–ENE-oriented chain composed from north to south by the Kabylies crustal wedge, a flysch domain corresponding to the original cover of the Maghrebian Tethys, and an external zone corresponding to the North African paleo-margin. These three main units were thrust via a north-dipping subduction system over the African foreland, the ensialic Atlas Chain (Frizon de Lamotte et al., 2006). The Greater and Lesser Kabylies, the 'Ka' block of AlKaPeCa (e.g., Bouillin et al., 1986; Guerrera et al., 2005; Fig. 4), represent terranes of Eurasian affinity probably separated from Greater Iberia, during Early Miocene. The AlKaPeCa block separated in response to rifting and back-arc extension from stable Eurasia to collide with Africa during the Tellian orogeny (the local name of the greater Apennine–Maghrebide orogeny), delineating the Miocene suture of the Maghrebian Tethys (e.g., Bouillin et al., 1986; Benaouali-Mebarek et al., 2006). The connection between formation of the Northern Algerian Basin and the Kabylies collision with North Africa is still debated. Indeed, it was proposed that opening of this Basin was similar to that of the Ligurian–Provençal Basin, i.e., related to subduction rollback and concomitant back-arc basin spreading (e.g., Vergés and Sabat, 1999; Rosenbaum and Lister, 2004). An alternative view contends that formation of the Northern Algerian Basin is unrelated to the Kabylies collision, the formation of the Basin having occurred at ~16 Ma, following the ~18 Ma Kabylies–Africa collision (e.g., Mauffret et al., 2004). According to this interpretation, the Northern Algeria basin would be linked to east–west-directed extension (e.g., Mauffret et al., 2004), associated with west-directed drifting of the Alborán micro-plate and east-directed drift of the Peloritani Mountains–Calabria blocks, ('Pe' and 'Ca' of AlKaPeCa, respectively; Fig. 4).

The age of Tell igneous activity varies but clusters around ~16–15 Ma and, in contrast to observations by Laouar et al (2005), there is little or no systematic variation with longitude (e.g., Bellon, 1981). The oldest plutonic and volcanic products, intermediate to acid calcalkaline rocks (Bellon, 1981; Semroud et al., 1994; Belanteur et al., 1995; Fourcade et al., 2001; Laouar et al., 2005; Fig. 9) are of Late Oligocene to Middle Miocene age (~24–15 Ma; Lesser Kabylie; ~19–15 Ma; Greater Kabylie). Their Sr isotopic compositions are variable and plot above the BSE estimate ($^{87}Sr/^{86}Sr_{(i)} = 0.7077–0.7220$; Fig. 10). In a few cases (e.g., the Filfila peraluminous cordierite-bearing granite), $^{87}Sr/^{86}Sr_{(i)}$ ratios are as high as 0.7528 (not reported in Fig. 10), likely reflecting partial melting of local meta-sediments (Fourcade et al., 2001). Nd isotopic ratios are lower than the ChUR estimate ($^{143}Nd/^{144}Nd(i) = 0.5121–0.5125$) Fig. 10). Only three mafic (gabbroic) samples from the Lesser Kabylie Cap Bougauroun show more depleted isotopic compositions, particularly with respect to Nd ($^{87}Sr/^{86}Sr_{(i)} = 0.7043–0.7060$; $^{143}Nd/^{144}Nd_{(i)} = 0.51324–0.51278$; Fourcade et al., 2001). In the Greater and Lesser Kabylies the igneous activity persisted intermittently until the Late Miocene (~9 Ma), with emplacement of relatively uniform volcanic and plutonic products. The presence of two very small outcrops of quartz-normative 'lamproite' (~11–9 Ma), near the city of Constantine in northern Algeria, is notable

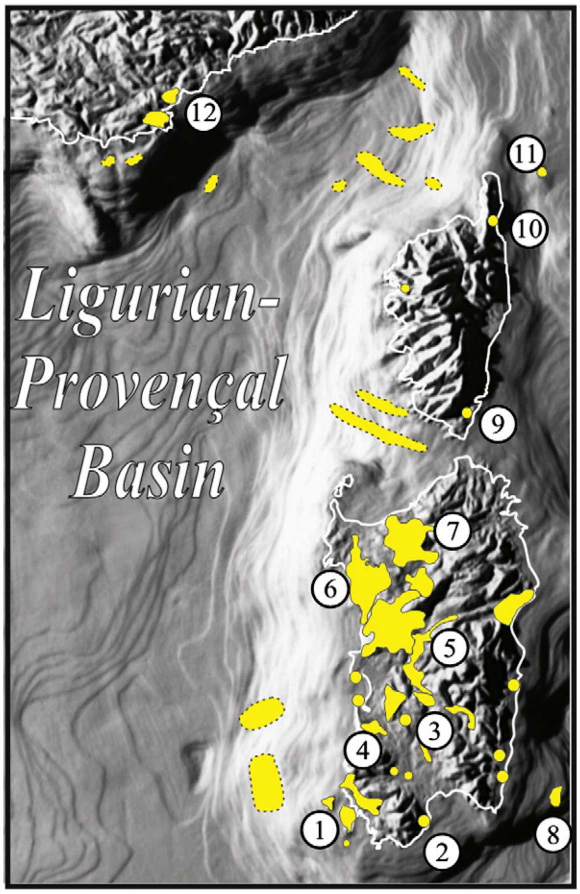

Fig. 11. Digital elevation model of the Sardinia–Corsica–Esterel region. 1 = Sulcis (S. Pietro and S. Antioco Islands, Narcao; ~28–15 Ma); 2 = Sarroch (~24–22 Ma); 3 = Marmilla (~19–17 Ma); 4 = Arcuentu (~30–17 Ma); 5 = Marghine–Barigadu (~21–19 Ma); 6 = Logudoro–Bosano (~38–16 Ma); 7 = Anglona (~21–18 Ma); 8 = Cornacya Seamount (~12.6 Ma); 9 = Balistra–Tre Paduli (~19 Ma); 10 = Sisco (~15–14 Ma); 11 = Capraia Island (~7.7–4.8 Ma); 12 = Esterel region (~32–24 Ma). Submarine outcrops are bounded with broken lines and their areal extent is only roughly estimated.

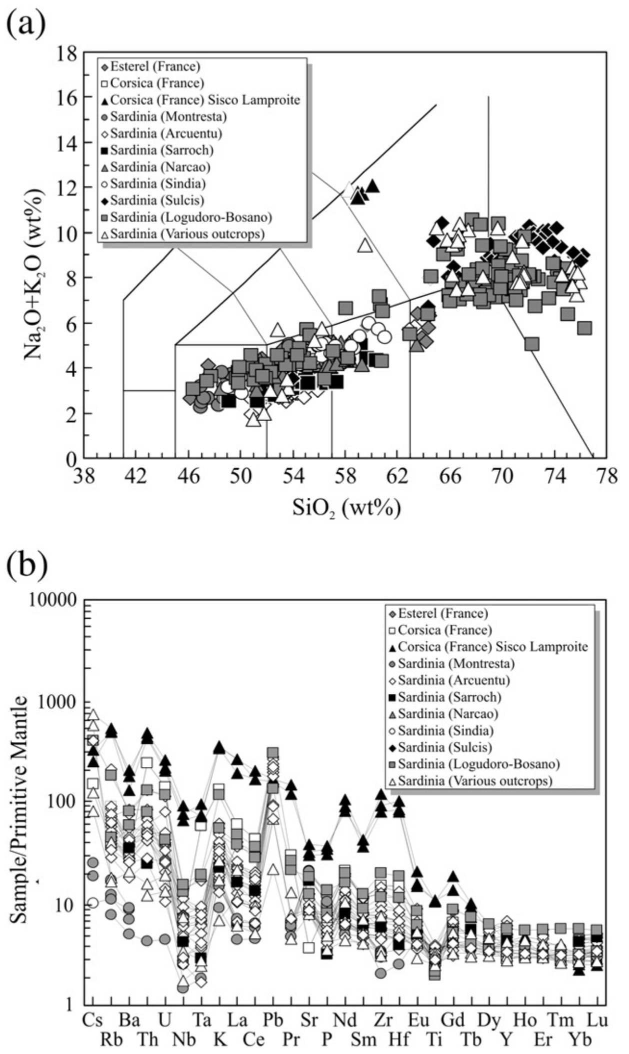

Fig. 12. (a) Total Alkali vs. Silica diagram (after Le Bas et al., 1986) for the subduction-related igneous rocks in the Sardinia–Corsica–Esterel region (data given in the electronic appendix). (b) Primitive mantle-normalized incompatible element diagram (normalization factors after Sun and McDonough, 1989) only for the most mafic (Mg# >0.60) igneous rocks of Sardinia–Corsica–Esterel region. Data sources are as follows: *Esterel* (Beccaluva et al., 1994, 2005; Ivaldi et al., 2003), *Corsica* (Ottaviani-Spella et al., 1996; Rossi et al., 1998), *Corsica* (*Sisco lamproite*; Peccerillo et al., 1988; Conticelli et al., 2002, 2009a), *Sardinia-Montresta* (Morra et al., 1997; Franciosi et al., 2003), *Sardinia-Arcuentu* (Dostal et al., 1982; Brotzu et al., 1997b; Downes et al., 2001; Franciosi et al., 2003; Beccaluva et al., 2005), *Sardinia-Sarroch* (Conte, 1997), *Sardinia-Narcao* (Dostal et al., 1982; Brotzu et al., 1997a), *Sardinia-Sindia* (Lonis et al., 1997), *Sardinia-Sulcis* (Araña et al., 1974; Morra et al., 1994; Conte et al., 2010), *Sardinia-Logudoro-Bosano* (Dupuy et al., 1974; Dostal et al., 1982; Rossi et al., 1998; Beccaluva et al., 2005), *Sardinia-Various outcrops* (Dostal et al., 1982; Caron and Orgeval, 1996; Beccaluva et al., 2005; Mattioli et al., 2000; Lustrino et al., 2004a, b, 2009).

(Bellon and Brousse, 1977; Bellon, 1981; Kaminskey et al., 1993), although some doubts exist concerning their classification (Woolley, 2001). The Constantine lamproites have never as yet been considered in reviews of Mediterranean lamproite petrogenesis (e.g., Prelević et al., 2008).

To the east, in Tunisia, igneous activity may be divided into two main phases. The first occurring in the Archipelago of La Galite, northwest Tunis (e.g., Talbi et al., 2005; Belayouni et al., 2010) and within two volcano-plutonic complexes (west of Tunis, Nefza and Mogods (Talbi et al., 2005). The products of the first phase resemble subduction-related Kabylies magmas in terms of both composition (mostly intermediate to evolved, the latter metaluminous to peraluminous in character) and age (~15–10 Ma; Bellon, 1981; Juteau et al., 1986; Talbi et al., 2005). Sr isotopic compositions are relatively radiogenic ($^{87}Sr/^{86}Sr_{(i)}$ ranging from 0.7075 to 0.7197) while the few published $^{143}Nd/^{144}Nd_{(i)}$ ratios (for the Isle La Galite) are relatively unradiogenic (averaging 0.5122; Fig. 10). The second phase is markedly different in terms of its very limited areal extent (confined to two localities, near Nefza and Mogods), mode of emplacement (only dykes with no plutonic or pyroclastic components), geochemical composition (mostly basic, SiO_2-undersaturated sodic alkali basalts), and age (~8–5.2 Ma; Bellon, 1981; Talbi et al., 2005, and references therein). Their areal extent may be more significant than currently estimated, at least in the Nefza region, as suggested by Bouguer gravity and aeromagnetic anomalies (Jallouli et al., 2003). The change from 'subduction-related' to 'intraplate-type' magma compositions, is essentially the same as that observed in the Betics, northern-Morocco, the Valencia Trough, Sardinia, and southern France (e.g., Lustrino et al., 2007b). Although not intensively studied, the presence of older igneous products (~20–14 Ma) represented by alkali basalts drilled in the Gulf of Hammamet southeast of Tunis (Laridhi Ouazaa, 1994) is of particular interest.

In summary, the Tell Province igneous rocks show near-exclusive subduction-related character, presumably tapping mantle wedge regions modified by fluids and/or melts released from a (currently immobile) north-dipping slab. The occasional appearance of garnet- and cordierite-bearing meta-sedimentary xenoliths, the peraluminous composition and extremely radiogenic $^{87}Sr/^{86}Sr_{(i)}$ (up to 0.7528), and unradiogenic $^{143}Nd/^{144}Nd_{(i)}$ (down to 0.5121) isotopic ratios, high $\delta^{18}O$ (up to +11‰) and relatively low $\delta^{34}S$ (down to −33‰) of some SiO_2-rich rocks have been attributed either to partial melting of pure meta-sedimentary flysch sources, or to shallow-level crustal contamination of mantle-derived melts (e.g., Semroud et al., 1994; Fourcade et al., 2001; Laouar et al., 2005). As far as we are aware, no Pb isotopic data for the Tell lithologies have yet been published.

8. Sardinia–Provençal–Corsica Province

As for the Alborán Sea, the Betic-Rif Belts, Valencia Trough, and the Tell Belt, two distinct Cenozoic magmatic episodes affected Sardinia (e.g., Lustrino et al., 2004a,b and references therein; Fig. 11). The first phase has been dated at ~38 to ~15 Ma (Beccaluva et al., 1985a; Lecca et al., 1997; Morra et al., 1994, 1997; Lustrino et al., 2009, and references therein), peaking at around ~22–18 Ma, and mostly comprising rocks with calcalkaline and arc-tholeiitic character along with subsidiary, late-stage high-K calcalkaline variants in the Middle Miocene. Dacitic to rhyolitic ignimbrites prevail, followed by andesitic, basaltic andesitic and minor High-Mg Basalts and High-Al Basalts (Brotzu et al., 1997a,b, Morra et al., 1997; Mattioli et al., 2000; Downes et al., 2001; Franciosi et al., 2003; Conte et al., 2010). Their compositional character is typically HFSE-depleted, indicative of subduction-related provenance, with K_2O/Na_2O ratios mostly >1 (Fig. 12). The most evolved rocks include also peralkaline trachytes and rhyolites, with comendites — from their type locality, Commenda, Sulcis, in southwest Sardinia (Araña et al., 1974; Morra et al., 1994; Lustrino et al., 2004a,b; Fig. 12). The second igneous episode, not covered in detail here, comprises ~12–0.1 Myr-old volcanics showing intraplate geochemical character (Lustrino et al., 2000, 2007a,b).

Even the isotopic ratios of the least differentiated (i.e., MgO> 6 wt.%) subduction-related rocks cover a wide compositional range ($^{87}Sr/^{86}Sr_{(i)}$ = 0.704–0.711; $^{143}Nd/^{144}Nd_{(i)}$ = 0.5121–0.5127; $^{206}Pb/^{204}Pb_{(i)}$ = 18.4–18.7; $^{207}Pb/^{204}Pb_{(i)}$ = 15.62–15.64; $^{208}Pb/^{204}Pb_{(i)}$ = 38.4–38.9, 'unfiltered' data being projected in Fig. 13). Coupled with $\delta^{18}O_{cpx}$ values ranging from +6.06‰ to +7.44‰ (Downes et al., 2001; Lustrino et al., 2008), these isotopic compositions indicate a complex range of crustal contamination processes, whether via slabs or slab-derived fluids and/or melts or at shallow levels via AFC processes in ascending mantle-derived magmas. Thus, three compositional end-members have been invoked to explain

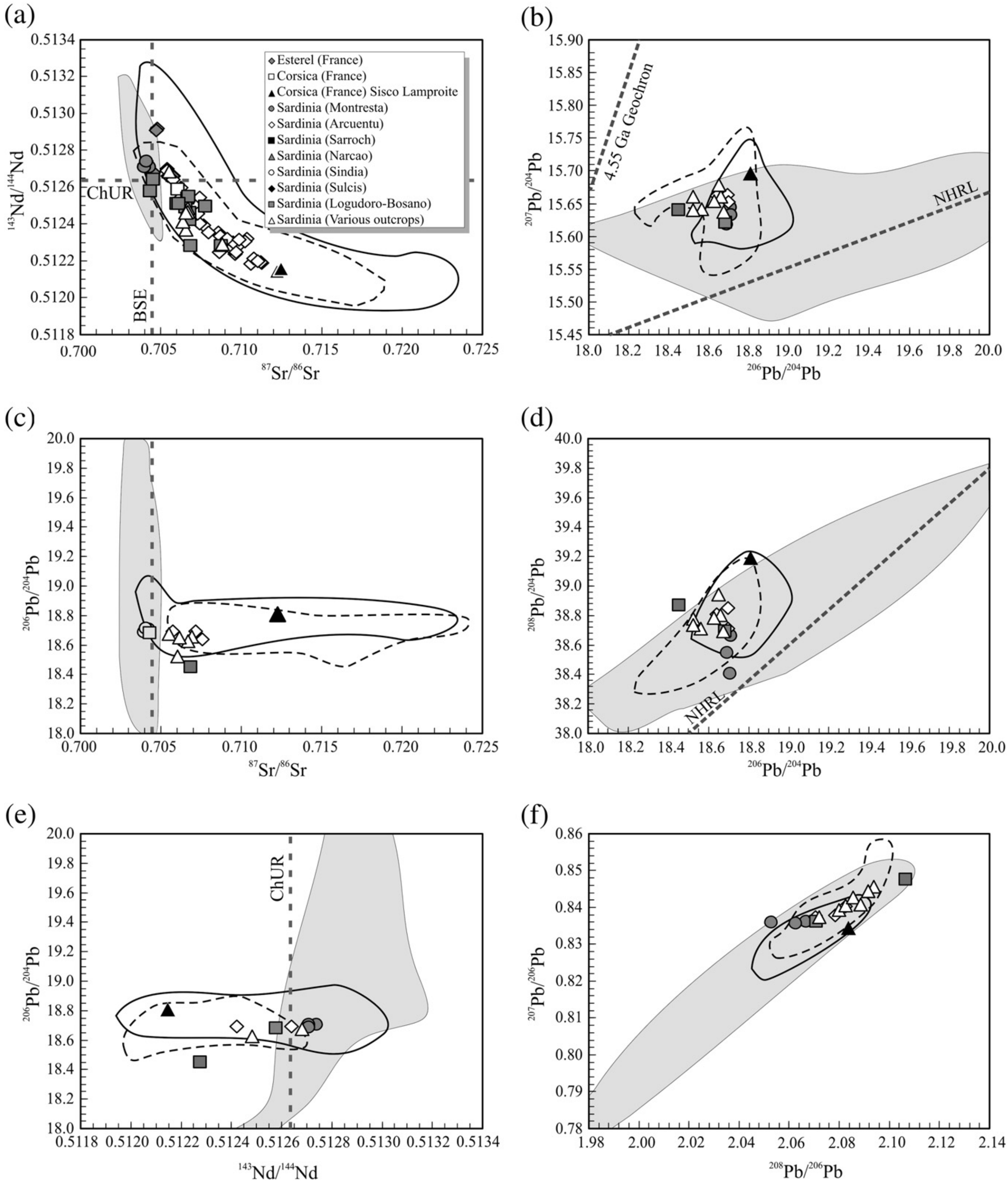

Fig. 13. Plots of (a) $^{87}Sr/^{86}Sr_{(i)}$ vs. $^{143}Nd/^{144}Nd_{(i)}$; (b) $^{207}Pb/^{204}Pb_{(i)}$ vs. $^{206}Pb/^{204}Pb_{(i)}$; (c) $^{206}Pb/^{204}Pb_{(i)}$ vs. $^{87}Sr/^{86}Sr_{(i)}$; (d) $^{208}Pb/^{204}Pb_{(i)}$ vs. $^{206}Pb/^{204}Pb_{(i)}$; (e) $^{206}Pb/^{204}Pb_{(i)}$ vs. $^{143}Nd/^{144}Nd_{(i)}$; and (f) $^{207}Pb/^{206}Pb_{(i)}$ vs. $^{208}Pb/^{206}Pb_{(i)}$ for the Sardinia–Corsica–Esterel region (data given in the electronic appendix). NHRL = Northern Hemisphere Reference Line (Hart, 1984). BSE = Present-day Bulk Silicate Earth estimate ($^{87}Sr/^{86}Sr = 0.70445$). ChUR = Present-day Chondritic Uniform Reservoir estimate ($^{143}Nd/^{144}Nd = 0.51264$). The grey field indicates compositions of the Circum-Mediterranean Anorogenic Cenozoic Igneous (CiMACI) rocks (Lustrino and Wilson, 2007). The dashed line delineates the field of the Periadriatic igneous rocks. The continuous line delineates the field of the Betics–Rif–Tell subduction-related igneous rocks. References as for Fig. 12.

the isotopic character of primitive Late Eocene to Middle Miocene magmatic rocks of Sardinia (Beccaluva et al., 1989; Franciosi et al., 2003; Lustrino et al., 2004a,b, and references therein): 1) a depleted MORB-like mantle wedge); 2) a component derived from altered (subducted) oceanic crust; 3) a subducted oceanic sediment component. The reaction of fluid phases from both oceanic crust and sediments, rather than melts released from the subducting plate, is favoured (e.g., Franciosi et al., 2003). Most of the intermediate to evolved compositions (andesite, dacite, rhyolite) are considered to represent products of magma mixing and contamination within the crust.

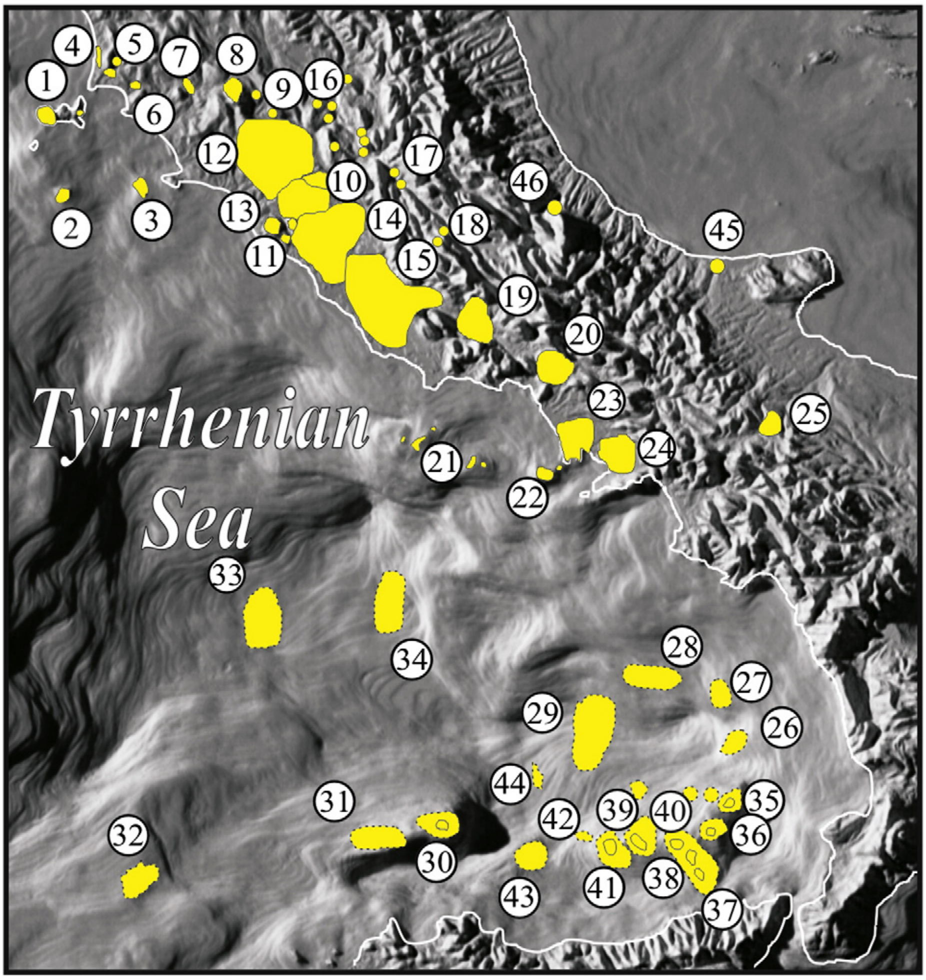

Fig. 14. Digital elevation model of Italy–Tyrrhenian Sea. *Localities from 1 to 11 = Tuscan Magmatic Province:* 1 = Elba Island (~8.5–5.8 Ma); 2 = Montecristo Island (~7 Ma); 3 = Giglio Island (~5 Ma); 4 = S. Vincenzo (~4.5 Ma); 5 = Campiglia (~5.9–4.3 Ma); 6 = Gavorrano (~4.9–4.3 Ma); 7 = Roccastrada (~2.5 Ma); 8 = Mount Amiata (~0.3–0.2 Ma); 9 = Radicofani (~1.3 Ma) and Torre Alfina (~0.9 Ma); 10 = Cimini Mountains (~1.3–0.9 Ma); 11 = Tolfa–Manziana–Cerite (~3.5 Ma). *Localities from 12 to 15 and 19 to 24 = Roman Magmatic Province:* 12 = Vulsini Mountains (~0.6–0.15 Ma); 13 = Vico (~0.4–0.1 Ma); 14 = Sabatini Mountains (~0.8–0.04 Ma); 15 = Alban Hills (~0.6–0.02 Ma); *Localities from 16 to 18 = Intra-Apennine Province:* 16 = S. Venanzo (~0.26 Ma); 17 = Cupaello (~0.64 Ma); 18 = Grotta del Cervo (~0.5 Ma). 19 = Ernici Mountains (~0.7–0.1 Ma); 20 = Mount Roccamonfina (~0.6–0.1 Ma); 21 = Pontine Islands (Ponza, Ventotene, Zannone, Santo Stefano, Palmarola; ~4.2–0.1 Ma); 22 = Ischia–Procida–Vivara Islands (<0.15 Ma); 23 = Phlegrean Fields (<0.2 Ma); 24 = Somma–Vesuvius Complex (<0.03 Ma); 25 = Mount Vulture (~0.8–0.1 Ma); 26 = Lametini Seamount; 27 = Alcione Seamount 28 = Palinuro Seamount; 29 = Marsili Seamount (<0.7 Ma); 30 = Ustica Island (~0.7–0.1 Ma); 31 = Anchise Seamount; 32 = Aceste Seamount; 33 = Magnaghi Seamount (~3.0–2.7 Ma); 34 = Vavilov Seamount (~0.7–0.1 Ma). *Localities from 35 to 41 = Aeolian Islands:* 35 = Stromboli; 36 = Panarea; 37 = Vulcano; 38 = Lipari; 39 = Salina; 40 = Filicudi; 41 = Alicudi. 42 = Eolo Seamount; 43 = Enarete Seamount; 44 = Glauco Smt; 45 = Pietre Nere (mela-syenite with lamprophyric affinity and within-plate geochemical signature; ~62–55 Ma); 46 = Pescosansonesco (lamprophyric dyke; ~70 Ma). Submarine outcrops are bounded with broken lines and their areal extent is only roughly estimated.

The Late Eocene to Middle Miocene activity in Sardinia appeared largely in the central-western sector of the Island, along the *Fossa Sarda* Graben, the easternmost branch of the Ligurian–Provençal Basin that cuts the island from north to south for about 220 km with an average width of 30–40 km (Cherchi et al., 2008 and references therein; Figs. 4 and 11). The volcanic activity (peaking at ~22–18 Ma) is roughly coeval with the maximum extension rate of the Ligurian–Provençal Basin and maximum angular velocity of rotation of the Corsica–Sardinia block (~21–16 Ma; Speranza et al., 2002; Gattacceca et al., 2007). The origin of magmatic activity can thus plausibly be adiabatic partial melting at temperatures exceeding a H_2O-undersaturated mantle wedge peridotite solidus (contaminated by northwest-dipping slab-derived fluids) during western Mediterranean extension associated with Ligurian–Provençal back-arc opening. The slab may be taken to represent Mesogean (Ionian) oceanic lithosphere (e.g., Beccaluva et al., 1989; Morra et al., 1997; Franciosi et al., 2003; Lustrino et al., 2009; Carminati et al., accepted for publication).

Compared to the Periadriatic plutons and dykes and the Betic-Rif samples, the Sardinian igneous rocks show a relatively narrow range of Sr and Nd isotopic compositions, given the absence of lamproites, which are elsewhere typically characterized by more extreme Sr and Nd isotopic compositions. The strikingly low radiogenic $^{206}Pb/^{204}Pb_{(i)}$ ratios of Sardinian rocks (ranging from 18.45 to 18.70; Fig. 13), lie between those of the northwest Alpine Taveyanne volcaniclastic sandstones and most Periadriatic and Betic-Rif igneous lithologies, Δ7/4 and Δ8/4 values ranging between ~+10 to ~+77 and from ~+17 to ~+94, respectively.

Calcalkaline and high-K calcalkaline rocks (andesites, microdiorites, dacites and rhyolites, often emplaced as ignimbritic flows) of Early Oligocene to Early Miocene age (~34–16 Ma) also appear on the Provençal coast in the Nice-Cap D'Ail area, in the Esterel region (e.g., Ivaldi et al., 2003; Beccaluva et al., 2005), southeast Corsica (e.g., Ottaviani-Spella et al., 1996, 2001), and offshore western Corsica (Rossi et al., 1998; Fig. 11). The areal extent of exposed Oligo-Miocene igneous rocks in the Provençe and southeastern Corsica is more than three orders of magnitude smaller than that of coeval igneous rocks in Sardinia. The smaller volume implied is possibly related to the relatively low stretching rate at loci closer to the rotation hinge near the Gulf of Genoa.

The magmatic rocks of southern France, Corsica and Sardinia show striking petrographic and geochemical similarities and, for this reason, similar mantle sources (variably modified by crustal components) have been inferred (e.g., Beccaluva et al., 1994). As far as we know, Sr, Nd and Pb isotopic data are lacking for the scarce Early Miocene volcanic rocks of Corsica, and only few Sr–Nd analyses for the Esterel microdiorites (locally called esterellites) and andesites, basaltic andesites and dacites have been published (Beccaluva et al., 2005). The Esterel igneous rocks show less radiogenic Sr and more radiogenic Nd compared with the Sardinian ones ($^{87}Sr/^{86}Sr_{(i)}$ = 0.7045–0.7058; $^{143}Nd/^{144}Nd_{(i)}$ = 0.51292–0.51266; Fig. 13). In northeasternmost Corsica an important outrop of lamproitic dykes is present near the village of Sisco. This will be treated in the Tyrrhenian Sea-related magmatism paragraph on the basis of geochronological, geological, petrographical and geochemical grounds.

In summary, dacitic to rhyolitic ignimbrites are the prevailing Late Eocene–Middle Miocene magmatic products in Sardinia and the conjugate French margin, followed by andesitic, basaltic andesitic and basaltic lavas with arc-tholeiitic to calcalkaline character. In contrast to the Alpine and Alborán case studies, true 'potassic' and 'ultrapotassic' rocks are absent in this area.

9. Tyrrhenian Sea Province

The Tyrrhenian Sea Province (taken here to include peninsular Italy and Sicily; Figs. 1 and 14) is characterized by the most diverse range of igneous rock types, with compositions ranging from subalkaline (both tholeiitic and calcalkaline) to alkaline (potassic), ultra-alkaline (ultrapotassic, lamproitic, kamafugitic) and carbonatitic (Peccerillo, 1985, 2005; Beccaluva et al., 1991; Conticelli et al., 2002, 2004, 2007; 2009b; Peccerillo and Lustrino, 2005; D'Orazio et al., 2007; Lustrino and Carminati, 2007; Avanzinelli et al., 2009; Rosatelli et al., 2010; Carminati et al., accepted for publication). Igneous activity peaked essentially during Pleistocene time (e.g., Peccerillo, 2005, and references therein). 'Anorogenic' basalts appear mostly in Sicily (Mount Etna and Hyblean Mountains) and the Sicily Channel (e.g., Pantelleria and Linosa Islands, plus scattered seamounts), with minor occurrences northwest of Sicily (Ustica Island and surrounding seamounts and other submerged volcanic fields). These, and similar basalts in Sardinia are considered as beyond the scope of this paper, having been covered in the recent review of 'anorogenic' Cenozoic igneous rocks by Lustrino and Wilson (2007).

After the Early to Middle Miocene, the Alpine orogenic wedge of western Corsica began to undergo extensional deformation with the formation of intramontane troughs and relatively sparse igneous activity (~14 Myr-old Sisco lamproite) to the northeast. Early Tortonian (~11–9 Ma) syn-rift sediments on the eastern margins of Corsica and Sardinia signal the initial opening of the northern Tyrrhenian Sea (Rosenbaum and Lister, 2004, and references therein). A top-to-the-east extensional shear allowed the change from ambient, relatively cold conditions during the Early to Middle Miocene to relatively high-temperatures associated with the Late Miocene exhumation of the high-pressure Giglio massif (e.g., Rossetti et al., 1999). This change in thermal regime has been explained either as a transition from syn- to post-tectonic conditions (e.g., Rossetti et al., 1999) or to continuous syn-tectonic conditions, subjected to west-directed subduction, until Pliocene, of Adriatic continental slices associated with delamination of continental lithosphere (e.g., Argnani, 2002). The formation of the Corsica basin, in the northern Tyrrhenian Sea, is associated with the inception of west-dipping extensional faults which appear to young eastward, consistent with the eastward younging of sediments in this basin (Argnani, 2002, and references therein). These features are considered to be linked to the first stages of formation of the Apennine Chain, developed at these latitudes under ensialic settings (e.g., Boccaletti et al., 2005). The eastward tectonic motion of the northwest-southeast striking Apennine Chain appears to have led to the thrusting of remnants of Liguride–Piedmontese oceanic lithosphere above the western margin of the Adria micro-plate. The overall volume of continental and oceanic slices involved in the thrusting (Triassic limestones structurally uplifted up to 10 km in the Gran Sasso Thrust, central Apennines) has been considered to reflect a lithospheric- rather than crustal-scale thrusting, at least from the Pliocene (Boccaletti et al., 2005).

The birth of the Tyrrhenian Sea is generally related to a southeastward shift of the axis of the Oligo-Miocene Ligurian–Provençal extension, to the east of Corsica–Sardinia. Both basins result from extension of the upper plate of the retreating subduction zone, the remnant of which is currently represented by the Calabrian arc (e.g., Cocchi et al., 2008, and references therein). During the Middle Miocene (~Langhian) the Sardinia–Corsica block stopped its counterclockwise rotation. Extension involving the 'Pe' and 'Ca' portions of the former AlKaPeCa terrane (Peloritani Mountains and Calabria), continued to migrate southeastwards (e.g., Gueguen et al., 1998). After splitting from southeastern Europe, the AlKaPeCa block started to fragment into individual blocks (Fig. 4), moving centrifugally towards external sectors of the CWM system: the Alborán (Al block moved west- and southwestwards, the Kabylies (Ka) block southward, and the Peloritani Mountains and Calabria (PeCa) block east- and southeastward. As previously noted, the Alborán block evolution has been modelled also in a completely different way (see movie in Carminati et al., accepted for publication). The southeastward drift trajectory of the Peloritani Mountains and Calabria terranes follows that of an inferred oceanic corridor, where relatively cold, dense Mesogean oceanic lithosphere was bounded by the continental lithosphere of Adria to the north and the Pelagonian (N African) block to the south. The east- to-southeastward drift of the Peloritani Mountains–Calabria followed the direction of slab rollback, bounded by the retreating Calabrian arc-foreac, allowing for the formation of the southern Tyrrhenian Sea.

The Tyrrhenian Sea can be subdivided into two main sectors, roughly separated by the lithospheric discontinuity at latitude 41° N (e.g., Serri, 1990), each of which evolved separately. The northern sector opened at relatively low velocities (~1–1.5 cm/yr) whereas the southern sector opened much faster (~4.5–5 cm/yr; Gueguen et al., 1998; Argnani and Savelli, 1999). The northeast–southwest-trending Selli fault system bounds the northwestern edge of the southern Tyrrhenian basin, connecting the abyssal plain of the Tyrrhenian Sea on the east with the Sardinian passive margin to the west (Doglioni et al., 2004; Cocchi et al., 2008). The Selli line developed entirely to the south of 41° N. Igneous activity associated with these two sectors also has distinct geochemical characteristics, as described below. The tectonic stress field in the southern Tyrrhenian Sea varied considerably with roughly north–south compression west of the Aeolian archipelago and northwest–southeast extension to the east (e.g., Pondrelli et al., 2004). These differences correspond to the respective difference of the plates around the southern Tyrrhenian: the Pelagonian, relatively thick and hardly subductable continental crust of mainland Sicily (west of the Aeolian archipelago), and the relatively thin, cold, dense and readily subductable Ionian lithosphere southeast of the Calabrian Arc. Between the two domains, a transitional zone is marked by a north–south transtensional discontinuity from the Aeolian islands to Mount Etna in Sicily (Pondrelli et al., 2004, and references therein).

Subduction-related igneous activity of this area may be grouped into six main districts (Fig. 14), which are described below:

1) The Tuscan–Corsican district (~14–0.2 Ma; Poli et al., 2003; Poli, 2004; Peccerillo, 2005, Conticelli et al., 2009a). Avanzinelli et al (2009) grouped the mantle-derived rocks in two distinct Provinces Corsican and Tuscan. The Corsican Province is large in area but yields only scarce igneous products at Sisco, (northeastern

Corsica), Capraia Island (part of the Tuscan archipelago) and Cornacya seamount in the southwest Tyrrhenian Sea. The Tuscan Province comprises the remainder of the Tuscan archipelago and the igneous rocks of the administrative regions of Tuscany and northern Latium. In this review we have grouped these two provinces into a single Tuscan–Corsican district.

The oldest Middle Miocene (~14.2 Ma; Civetta et al., 1978) lamproitic products are volumetrically insignificant and occur as dykes at Sisco. These lithologies have been tentatively linked petrogenetically with slightly younger (~12.6 Ma), strongly altered rocks dredged southeast of Sardinia in the southwestern Tyrrhenian Sea, along the flanks of the Cornacya seamount, which show shoshonitic to lamproitic affinity (Mascle et al., 2001). Following a possible hiatus of 5–6 Myr, this Middle Miocene activity was followed in the Late Miocene (~8.5–5 Ma; Ferrara et al., 1985) by granitoid intrusions (mostly monzogranite) and minor volcanic potassic to ultrapotassic, mafic to intermediate eruptions, whose products have been dredged from seamount flanks (such as the 7.5–6.5 Myr-old Vercelli seamount) and sampled from the Tuscan Archipelago (Elba, Capraia, Montecristo and Giglio Islands; Dini et al., 2002; Poli et al., 2003; Peccerillo, 2005; Gasparon et al., 2008). The remaining igneous activity is mostly represented by Plio-Quaternary basic to acid intrusive and eruptive products, either exposed or recorded at depth, in northwestern Tuscany (Campiglia, San Vincenzo, Orciatico, Montecatini Val di Cecina, Larderello, Monteverdi, Roccastrada, Mount Amiata, Radicofani, Torre Alfina) and northwest Latium (Tolfa, Manziana, Cerite). This is the most complex district in terms of the range of coexisting lithologies and their geochemical variation — the latter ranging from purely crustal anatectic types (peraluminous and also cordierite–garnet-bearing rhyolites and granites) to calcalkaline, high-K calcalkaline, potassic, ultrapotassic and lamproitic types (Fig. 15). The Tuscan lamproites are by far the most interesting from a petrologic standpoint. Although ultrapotassic in character, they differ from other ultrapotassic Italian lithologies in

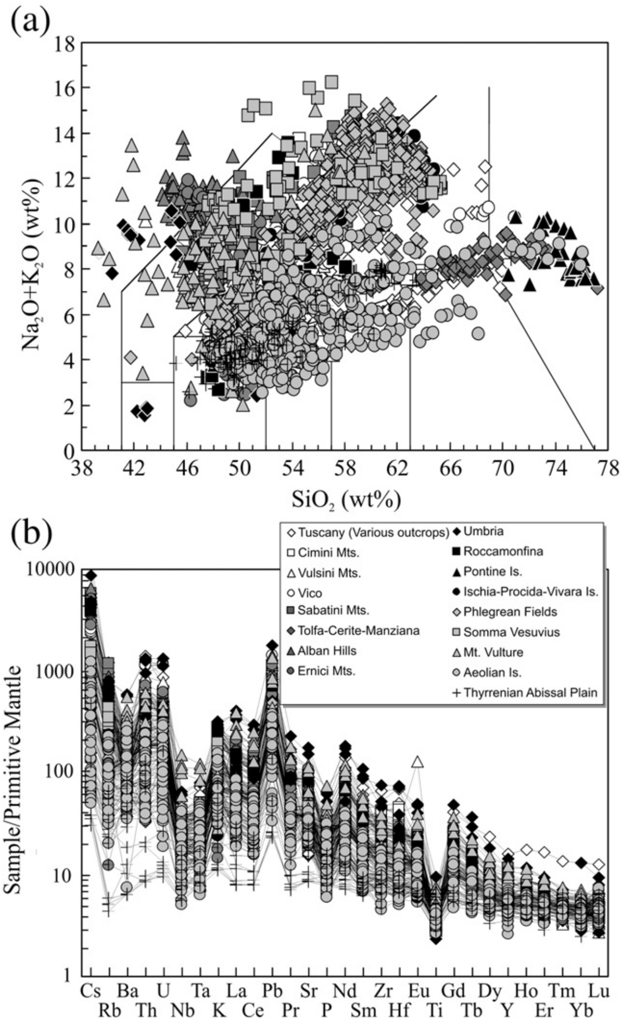

Fig. 15. (a) Total Alkali vs. Silica diagram (after Le Bas et al., 1986) for the subduction-related igneous rocks of peninsular Italy, the Aeolian Islands and Tyrrhenian Sea (data given in the electronic appendix). (b) Primitive mantle-normalized incompatible element diagram (normalization factors after Sun and McDonough, 1989) only for the most mafic (Mg# >0.60) igneous rocks of peninsular Italy, Aeolian Islands and Tyrrhenian Sea. Data sources are as follows: *Tuscany* (Vollmer, 1976, 1977; Peccerillo et al., 1987, 1988; Conticelli et al., 1992a, b, 2001, 2002, 2007, 2009a; Feldstein et al., 1994; Conticelli, 1998; Gasperini et al., 2002; Chelazzi et al., 2006; Gagnevin et al., 2007; Cadoux and Pinti, 2009), *Cimini Mountains* (Vollmer, 1976; Conticelli et al., 2002, 2007, 2009a; Gasperini et al., 2002; Perini et al., 2003); *Vulsini Mountains* (Cundari, 1979; Vollmer and Hawkesworth, 1980; Civetta et al., 1984; Rogers et al., 1985; Conticelli et al., 1986, 1987, 1991, 2002, 2007; Conticelli, 1989; Varekamp and Kalamarides, 1989; Conticelli and Peccerillo, 1992; Di Battistini et al., 1998, 2001; Gasperini et al., 2002; Martelli et al., 2004), *Vico* (Vollmer, 1976; Barbieri et al., 1988; Gasperini et al., 2002; Perini et al., 1997, 2004), *Sabatini Mountains* (Vollmer, 1976; Cundari, 1979; Conticelli et al., 1997, 2002, 2007; Gasperini et al., 2002), *Tolfa-Cerite-Manziana* (Clausen and Holm, 1990; Pinarelli, 1991; De Rita et al., 1994, 1997), *Alban Hills* (Vollmer, 1976; Vollmer and Hawkesworth, 1980; Peccerillo et al., 1984; Ferrara et al., 1985; D'Antonio et al., 1996; Gaeta, 1998; Conticelli et al., 2002; Martelli et al., 2004; Freda et al., 2006; Gaeta et al., 2006; Giordano et al., 2006; Boari et al., 2009; Marra et al., 2009), *Ernici Mountains* (Civetta et al., 1979, 1981; D'Antonio et al., 1996; Conticelli et al., 2002; Gasperini et al., 2002; Boari and Conticelli, 2007; Frezzotti et al., 2007; Boari et al., 2009), *Umbria* (Peccerillo et al., 1988; Stoppa, 1988; Cundari and Ferguson, 1991; Conticelli and Peccerillo, 1992; Stoppa and Lupini, 1993; Stoppa and Cundari, 1995, 1998; Stoppa and Woolley, 1997; Castorina et al., 2000; Barbieri et al., 2002; Conticelli et al., 2002; Stoppa et al., 2002, 2003a, b, 2005; Stoppa and Sharygin, 2009), *Roccamonfina* (Vollmer, 1976; Carter et al., 1978; Vollmer and Hawkesworth, 1980; Beccaluva et al., 1991; Di Girolamo et al., 1991; Giannetti and Ellam, 1994; D'Antonio et al., 1996; Conticelli et al., 2002, 2007, 2009b; Martelli et al., 2004; Rouchon et al., 2008), *Pontine Islands* (Ponza, Ventotene and Palmarola; D'Antonio and Di Girolamo, 1994; D'Antonio et al., 1996, 1999; Conte and Dolfi, 2002; Conticelli et al., 2002; Cadoux et al., 2005, 2007), *Ischia–Procida-Vivara Islands* (Hawkesworth and Vollmer, 1979; Vollmer and Hawkesworth, 1980; Poli et al., 1987, 1989; Crisci et al., 1989; Civetta et al., 1991a, b; D'Antonio and Di Girolamo, 1994; Di Girolamo et al., 1995; D'Antonio et al., 1996, 1999, 2007; Conticelli et al., 2002; Brown et al., 2008), *Phlegrean Fields* (Di Girolamo, 1968, 1970; Di Girolamo and Stanzione, 1973; Di Girolamo et al., 1973, 1984; Vollmer, 1976; Albini et al., 1977; Ghiara et al., 1977, 1979; Barberi et al., 1978, 1979a, b; Di Girolamo and Rolandi, 1979; Vollmer and Hawkesworth, 1980; Carbone et al., 1984; Lirer et al., 1987; Rosi and Sbrana, 1987; Beccaluva et al., 1991; Civetta et al., 1991b, 1997; Orsi et al., 1992; Scarpati et al., 1993; D'Antonio and Di Girolamo, 1994; Melluso et al., 1995; Wohletz et al., 1995; D'Antonio et al., 1996, 1999, 2007; De Vita et al., 1999; Pappalardo et al., 1999, 2002, 2008; Ricci, 2000; Lustrino et al., 2002a; De Astis et al., 2004; Martelli et al., 2004; D'Oriano et al., 2005; Piochi et al., 2005; Fedele et al., 2006; Mastrolorenzo and Pappalardo, 2006; Aulinas et al., 2008; Pabst et al., 2008; Tonarini et al., 2009), *Somma–Vesuvius* (Vollmer, 1976; Hawkesworth and Vollmer, 1979; Vollmer and Hawkesworth, 1980; Cortini and Hermes, 1981; Cortini and Van Calsteren, 1985; Joron et al., 1987; Civetta et al., 1991a; Civetta and Santacroce, 1992a,b; Caprarelli et al., 1993; Santacroce et al., 1993a,b; Villemant et al., 1993; Cioni et al., 1995a,b; D'Antonio et al., 1996; Ayuso et al., 1998; Landi et al., 1999; Marianelli et al., 1999, 2005; Somma et al., 2001; Cortini et al., 2004; Mastrolorenzo and Pappalardo, 2006; Paone, 2006, 2008; Piochi et al., 2006; Di Renzo et al., 2007; Aulinas et al., 2008; Marturano et al., 2009), *Mount Vulture* (Vollmer, 1976; Hawkesworth and Vollmer, 1979; Vollmer and Hawkesworth, 1980; De Fino et al., 1982, 1986; Melluso et al., 1996; Stoppa and Woolley, 1997; Stoppa and Principe, 1998; Rosatelli et al., 2000; Beccaluva et al., 2002; Conticelli et al., 2002; D'Orazio et al., 2007; Stoppa et al., 2003a,b, 2009), *Aeolian Islands* (Alicudi, Filicudi, Vulcano, Stromboli, Panarea, Salina, Lipari; Barberi et al., 1974; Keller, 1974, 1980a,b; Pichler, 1980; Rosi, 1980; Villari, 1980a,b; Cortini, 1981; Dupuy et al., 1981; Luais, 1988; De Fino et al., 1988, 1991; Ellam et al., 1988, 1989; Ellam and Harmon, 1990; Francalanci et al., 1988, 1989, 1993a,b, 2004a,b; Crisci et al., 1991; Esperança et al., 1992; Peccerillo and Wu, 1992; Francalanci and Santo, 1993; Gillot and Keller, 1993; Hornig-Kjarsgaard et al., 1993; Clocchiatti et al., 1994; Bonaccorso et al., 1996; De Astis et al., 1997, 2000; Del Moro et al., 1998; Gioncada et al., 1998, 2003; Clift and Blusztajn, 1999; Peccerillo, 1999; Gertisser and Keller, 2000; Rosi et al., 2000; Santo, 2000; Metrich et al., 2001; Tonarini et al., 2001a,b; De Rosa et al., 2003a,b; Corsaro et al., 2004; Frezzotti et al., 2004; Santo et al., 2004; Peccerillo et al., 1993, 2004a,b; Landi et al., 2006; Tommasini et al., 2007; Araña and Frazzetta, 2008; Cimarelli et al., 2008; Martelli et al., 2008; Landi et al., 2009) *Tyrrhenian abyssal plain* (Keller and Leiber, 1974; Selli et al., 1977; Colantoni et al., 1981; Beccaluva et al., 1985b, 1990; Robin et al., 1987; Bertrand et al., 1990; Sborshchikov and Al'mukhamedov, 1992; Savelli and Gasparotto, 1994; Gasperini et al., 2002; Trua et al., 2002, 2010).

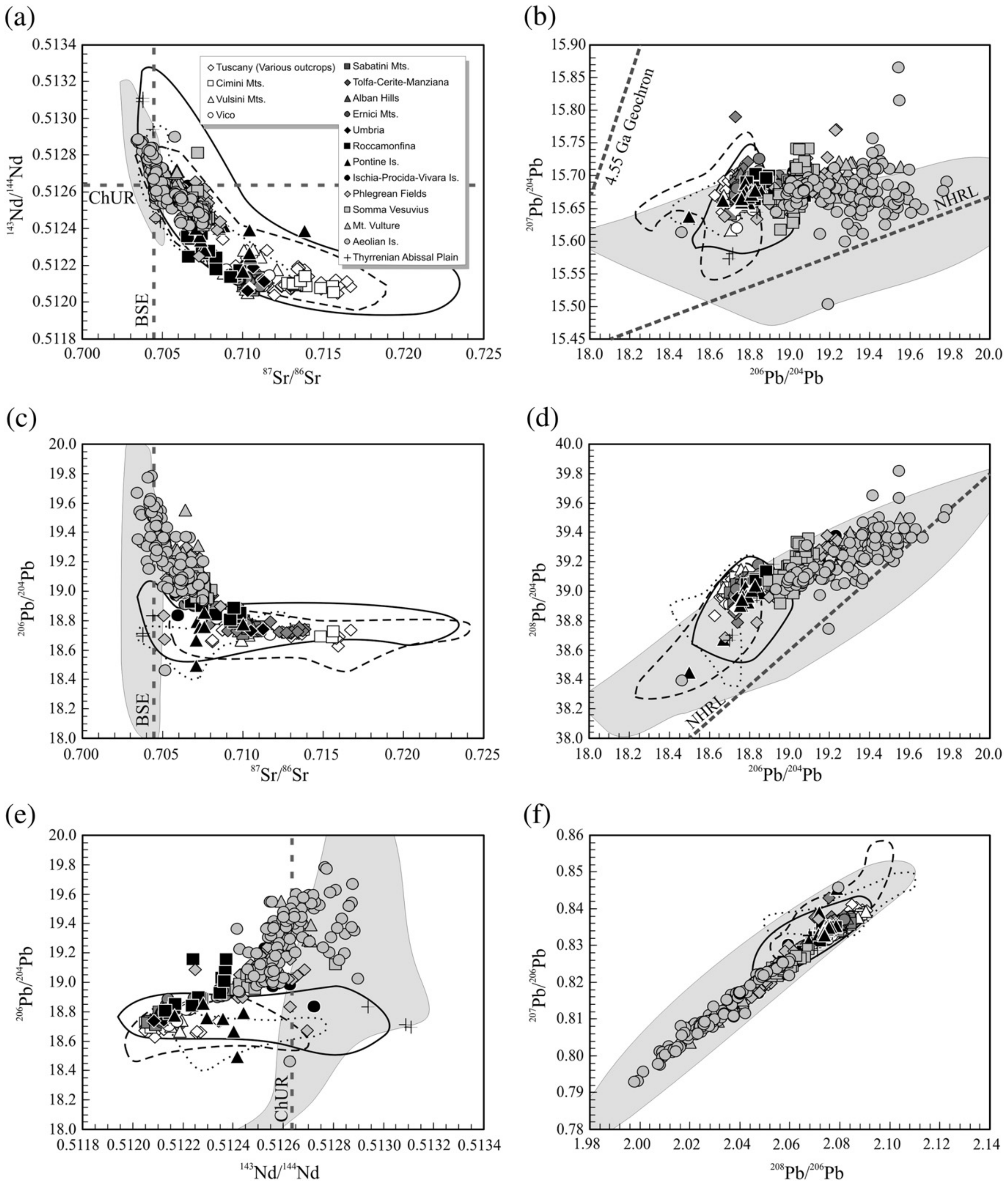

Fig. 16. Plots of (a) $^{87}Sr/^{86}Sr_{(i)}$ vs. $^{143}Nd/^{144}Nd_{(i)}$; (b) $^{207}Pb/^{204}Pb_{(i)}$ vs. $^{206}Pb/^{204}Pb_{(i)}$; (c) $^{206}Pb/^{204}Pb_{(i)}$ vs. $^{87}Sr/^{86}Sr_{(i)}$; (d) $^{208}Pb/^{204}Pb_{(i)}$ vs. $^{206}Pb/^{204}Pb_{(i)}$; (e) $^{206}Pb/^{204}Pb_{(i)}$ vs. $^{143}Nd/^{144}Nd_{(i)}$; and (f) $^{207}Pb/^{206}Pb_{(i)}$ vs. $^{208}Pb/^{206}Pb_{(i)}$ for peninsular Italy, the Aeolian Islands and Tyrrhenian Sea rocks (data given in the electronic appendix). Present-day BSE = Bulk Silicate Earth estimate ($^{87}Sr/^{86}Sr = 0.70445$). ChUR = Present-day Chondritic Uniform Reservoir estimate ($^{143}Nd/^{144}Nd = 0.51264$). NHRL = Northern Hemisphere Reference Line (Hart, 1984). The grey field indicates compositions of the Circum-Mediterranean Anorogenic Cenozoic Igneous (CiMACI) rocks (Lustrino and Wilson, 2007). The dashed line delineates the field of the Periadriatic igneous rocks. The continuous line delineates the field of the Betics–Rif–Tell subduction-related igneous rocks. The dotted line delineates the field of the Sardinia–Corsica–France subduction-related igneous rocks. References as for Fig. 15.

lacking either modal or normative leucite, and having relatively high SiO_2 (mostly 55–60 wt.%), MgO (>7 wt.%), Cr (>400 ppm) and Ni (>150 ppm) contents and high Mg# (>72), while their Sr–Nd isotopic compositions resemble those of upper continental crustal samples ($^{87}Sr/^{86}Sr_{(i)} > 0.716$; $^{143}Nd/^{144}Nd_{(i)} < 0.5121$; Fig. 16). The Sisco lamproites have similarly extreme isotopic compositions (i.e., $^{87}Sr/^{86}Sr_{(i)}>0.712$; $^{143}Nd/^{144}Nd_{(i)}$ ~0.5121; Conticelli et al., 2009a and references therein; Fig. 13), likewise suggesting crustal-like isotopic compositions in melts equilibrated with mantle sources. A conventional explanation for this would be in terms of subduction-related crustal contamination of the lithospheric mantle in the overriding plate (e.g., Conticelli and Peccerillo, 1992; Conticelli, 1998; Peccerillo, 2005; Avanzinelli et al., 2009; Conticelli et al., 2009a). Such interaction between crustal components and ambient mantle would be expected to produce a network of metasomatic veins characterized by phlogopite–orthopyroxene–olivine parageneses in a matrix of unreacted harzburgitic peridotite. Exclusive or near-exclusive partial melting of such metasomatic assemblages under high X_{H2O} could plausibly produce melts of lamproitic composition (see Avanzinelli et al., 2009 for more detailed discussion of this process). However, the timing of infiltration of crustal components into otherwise pristine mantle remains a subject of debate. The two existing model alternatively favour coincidence with the Late Cretaceous to Middle Eocene (eo-alpine) phase (Peccerillo and Martinotti, 2006) or Late Cenozoic (Miocene to Pliocene time; Conticelli et al., 2009a). Isotopic data for Tuscan igneous rocks plot in the enriched quadrant of a Sr–Nd isotopic diagram ($^{87}Sr/^{86}Sr_{(i)}$ 0.708–0.716; $^{143}Nd/^{144}Nd_{(i)}$ 0.5123–0.5120; Fig. 16) and show relatively uniform ratios of $^{206}Pb/^{204}Pb_{(i)}$ (~18.65–18.74), $^{207}Pb/^{204}Pb$ (~15.64–15.70) and $^{207}Pb/^{204}Pb_{(i)}$ (~38.91–39.15; Fig. 16). Δ7/4 and Δ8/4 values range, respectively, from ~+13 to ~+18 and from ~+69 to ~+87. These values are nearly indistinguishable from those of the subduction-related provinces shown before.

2) The Roman district (~4.2–0 Ma; Beccaluva et al., 1991; Conticelli et al., 2002, 2004, 2009b; Marra et al., 2004; Peccerillo, 2005; Peccerillo and Lustrino, 2005; Avanzinelli et al., 2009). From north to south this district includes the volcanoes of Vulsini Mountains, Cimini Mountains-Vico. Sabatini Mountains, Alban Hills, Ernici Mountains, Mount Roccamonfina, Pontine Islands, Procida, Vivara and Ischia Islands, Mount Vesuvius, and Phlegrean Fields. The unusual mineralogical and chemical character of these volcanoes (i.e., the relative abundance of leucite and strongly potassic to ultrapotassic character) was pointed out more than a century ago (Washington, 1906). As for the Tuscan district, detailed petrologic studies led some authors to identify several characteristic sub-districts. However, for the sake of clarity and simplicity, they are considered as a single one in this review. The activity in the Roman district began during the Early Pliocene (~4.2–1 Ma) with eruptions of acid magmas, mostly trachytes and rhyolites, in the Pontine Islands (Cadoux et al., 2005). According to some authors these SiO_2-oversaturated magmas are more closely associated to the Tuscan anatectic district rather than to the Roman district (e.g., Conticelli et al., 2002, 2004; Peccerillo, 2005; Avanzinelli et al., 2009). Thick (up to 1800 m) Late Pliocene (~2 Myr-old) volcanic successions, mostly of two-pyroxene andesites and basaltic andesites (Barbieri et al., 1979) appeared in the Campanian Plain to the northwest of Naples, representing the second important volcanic episode in the Roman district before the main phase of magmatic activity commenced during the Early Pleistocene (~0.8 Ma), continuing to Recent time (e.g., Peccerillo, 2005, and references therein). Abundant volcanic rocks are buried offshore the Gaeta Gulf, east of the Pontine Islands, just south of the 41° N latitude, under few tens of meters Volturno river sediments at ~250 m b.s.l. (de Alteriis et al., 2006).

The principal eruptive types are mostly potassic and ultrapotassic in character. The potassic rocks, mostly comprising potassic trachybasalts and shoshonites, are typically SiO_2-saturated, therefore characterized by the absence of leucite. The ultrapotassic lavas are mildly to strongly SiO_2-undersaturated, often showing high abundances of modal leucite. The main lithotypes are leucitites, tephrites to phonolites (Peccerillo, 2005; Avanzinelli et al., 2009; Conticelli et al., 2009b, and references therein; Fig. 15). A variety of rock associations (e.g., calcalkaline, shoshonitic and ultrapotassic) are sometimes observed in the same volcano (e.g., Peccerillo, 2005, and references therein). The origin of potassic and ultrapotassic magmas in the Roman district has been ascribed mostly to partial melting of heterogeneous mantle sources, infiltrated by phlogopite-rich veins resulting from the interaction of slab-derived melts and fluids with ambient mantle (Peccerillo, 1985, 1999, 2005; Beccaluva et al., 1991; Conticelli et al., 2002, 2004, 2009b; Avanzinelli et al., 2009, and references therein). In particular, the leucite-bearing magmas have been modelled in terms of interaction of sedimentary components (essentially marls) introduced to the mantle via subduction (e.g., Peccerillo, 1985; Beccaluva et al., 1991; Avanzinelli et al., 2009, and references therein). Given the CaO-rich composition of kamafugites and leucite-bearing ultrapotassic rocks in this domain, the mantle source is unlikely to be harzburgitic (as proposed for lamproitic melts; e.g., Conticelli and Peccerillo, 1992; Conticelli, 1998; Prelević and Foley, 2007) but it is consistent with more 'fertile' lherzolitic to wehrlitic peridotites. Experimental studies indicate that partial melting of such mantle facies under high X_{CO2} would be expected to produce magmas resembling kamafugites and other SiO_2-undersaturated ultrapotassic types, even at relatively low pressures (see discussion in Avanzinelli et al., 2009). Accordingly, enrichment in Ca and CO_2 would be considered to reflect the infiltration of $CaCO_3$-rich recycled sediments into the mantle wedge (see also Frezzotti et al., 2009). Nearly 80 years ago Alfred Rittmann (1933) proposed that the SiO_2-poor and alkali-rich (i.e., 'potassic') composition of the Somma–Vesuvius lavas was an effect of large-scale assimilation of carbonate country-rocks by magmas of deeper origin. Such hypotheses were later discarded on the basis of petrologic arguments by Savelli (1967) who argued that heat loss caused by the digestion of significant amounts of limestone would have precluded any such magmatic process. However, recent geochemical and petrologic investigations (e.g., Pappalardo et al., 2004; Piochi et al., 2006; Gaeta et al., 2009), as well as experimental studies on sedimentary carbonate–silicate magma interaction (Iacono Marziano et al., 2008; Mollo et al., 2010), appear to sustain Rittmann's original hypothesis. These studies evidence that carbonate digestion could, after all, produce the 'exotic' compositions of the Italian lavas, according to the simplified reaction: $CaCO_3^{solid} + SiO_2^{melt} + MgO^{melt} + FeO^{melt} + Al_2O_3^{melt} \rightarrow (Di\text{-}Hd\text{-}CaTs)_{ss}^{solid} + CO_2^{fluid}$ (Mollo et al., 2010), the contaminated magma being characterized by lower SiO_2 and higher alkalies (K_2O) contents as compared to the original magma. Sr–Nd isotopic composition of the Roman Magmatic Province rocks as a whole define a smooth hyperbola in plots of $^{87}Sr/^{86}Sr_{(i)}$ versus $^{143}Nd/^{144}Nd_{(i)}$ ratios (Fig. 16), $^{87}Sr/^{86}Sr_{(i)}$ ranging from 0.7046 (in the Phlegrean Fields) to 0.7056 (Cimini Mountains) and with $^{143}Nd/^{144}Nd_{(i)}$ ranging from 0.5128 (Procida Island) to 0.5120 (Vulsini Mountains). The compositions of each volcanic domain plot in a relatively restricted, well-defined Sr–Nd ratio field. Regarding Pb isotopic ratios, magmas in the Roman district show much larger ranges of compositions compared to other subduction-related provinces such as those of the Alps, Betics, Rif, Tell, Sardinia–Corsica, and Esterel. Indeed, with the exception of one sample, $^{206}Pb/^{204}Pb_{(i)}$ ratios in the 'Roman' magmas range between 18.63 and 19.30, $^{207}Pb/^{204}Pb_{(i)}$ between 15.62 to 15.77 and $^{208}Pb/^{204}Pb_{(i)}$ between from 38.68 and 39.39 (Fig. 16). Moreover, each volcanic domain plots in relatively well-defined Pb isotopic fields, Δ7/4 and Δ8/4 values ranging, respectively, from ~+7 to ~+19 and ~+36 to ~+89. The Somma–Vesuvius, Phlegrean Fields, Ischia–Procida Islands, and most of the Roccamonfina and Ernici Mountains compositions plot towards higher $^{206}Pb/^{204}Pb_{(i)}$ and $^{208}Pb/^{204}Pb_{(i)}$, outside Pb isotopic fields defined by the other CWM subduction-related rocks, but with overlapping $^{207}Pb/^{204}Pb_{(i)}$ (Fig. 16).

3) The Umbria–Latium–Abruzzo district (~0.6.0.4 Ma; Stoppa and Lavecchia, 1992; Castorina et al., 2000; Stoppa et al., 2003a,b; Rosatelli et al., 2010). This ultra-alkaline district includes numerous monogenetic volcanoes of unusual chemical and mineralogical composition. They are confined to a relatively limited area located along major extensional faults running nearly in the middle of the Italian peninsula. Typical mineral assemblages include melilite, leucite, kalsilite, monticellite, wollastonite, perovskite and Ba–Sr-rich calcite, the main rock types comprising melilitolite, melilitite, Ca-carbonatite, and kamafugite (with coppaelite and venanzite types occuring near the villages of Copaello and San Venanzo; Fig. 15). The most remarkable characteristic of this domain is its strongly SiO_2-undersaturated character, high Mg# (up to 92), high CaO (up to 40 wt.%), high LILE/HFSE and Rb/Sr ratios, with low Al_2O_3 (from 4 to 11 wt.%) and Nb contents. $^{87}Sr/^{86}Sr_{(i)}$ isotopic ratios are strongly radiogenic (~0.710–0.711), and $^{143}Nd/^{144}Nd_{(i)}$ are strongly unradiogenic (down to ~0.5120; Fig. 16). The few available Pb isotopic data indicate mildly radiogenic $^{206}Pb/^{204}Pb_{(i)}$ (~18.7), $^{207}Pb/^{204}Pb_{(i)}$ (~15.65) and $^{208}Pb/^{204}Pb_{(i)}$ (~38.8; Fig. 16), Δ7/4 and Δ8/4 values ranging from ~+13 to ~+14 and from ~+61 to ~+67, respectively. It is notable that the wollastonite- and melilite-bearing Colle Fabbri and Ricetto rocks are considered either as paralavas (pyrometamorphic rocks) produced either by partial melting of marly sediments due to coal fire (Melluso et al., 2003, 2005) or, as truly igneous rocks, resulting from assimilation of pelitic Ca-rich country rocks by melilititic magma (Stoppa et al., 2009). The rare carbonate-rich rocks have been interpreted as magmas of carbonatitic type (e.g., Stoppa and Woolley, 1997; Stoppa et al., 2005; Rosatelli et al., 2010, and references therein) or sedimentary limestone-silicate magma hybrid product (Peccerillo, 2004, 2005 and references therein).

4) The Mount Vulture district (~0.8–0.1 Ma; i.a., Melluso et al., 1996; Beccaluva et al., 2002; De Astis et al., 2006). This volcano, along with small monogenetic centres along the Ofanto line in south-eastern Italy, represents the Apulian Region, as defined by Washington (1906) or the Lucanian Magmatic Province of Conticelli et al. (2004). The Mount Vulture volcano consists almost exclusively of pyroclastic rocks, with minor lava flows and domes, and it is located close to the boundary of the southern Apennine east-verging thrusts and the western margin of the Apulian (Adria) foreland (e.g., Schiattarella et al., 2005 and references therein). The Mount Vulture igneous rocks are mostly basanite and trachy-phonolite with minor foidite, tephrite, phono–tephrite and melilitite (Fig. 15). These are less potassic than those of the Roman domain, the main feldspatoid being haüyne rather than leucite. A different petrogenetic model to that invoked for the Roman domain draws attention to their relatively high HFSE contents and lower LILE/HFSE ratios Sr–Nd compared to lavas in the Roman domain (Peccerillo, 2005). A small carbonatitic outcrop, referred to as a massive alvikite flow (D'Orazio et al., 2007, 2008), or welded tuff (Stoppa et al., 2008), was recently discovered in this domain, after being initially interpreted as a travertine slab (Giannandrea et al., 2004). The Mount Vulture isotopic compositions plot close to the Sr–Nd depleted field, with $^{87}Sr/^{86}Sr_{(i)}$ ratios ranging from ~0.7052 to ~0.7064 and $^{143}Nd/^{144}Nd_{(i)}$ ranging from ~0.51272 to ~0.51258 (Fig. 16). Mount Vulture $^{206}Pb/^{204}Pb_{(i)}$ isotopic ratios are also particularly radiogenic (~19.16–19.55) compared to most other Italian magmas, but $^{207}Pb/^{204}Pb_{(i)}$ (15.67–15.72) and $^{208}Pb/^{204}Pb_{(i)}$ (39.12–39.40) ratios overlap with those of Somma–Vesuvius, Phlegrean Fields, Ischia–Procida Islands, Roccamonfina and the Ernici Mountains (Fig. 16). Δ7/4 and Δ8/4 range from ~+9 to ~+14 and from ~+23 to ~+44, respectively.
Models involving near-horizontal rupture of subducting slabs (slab break-off model; e.g., De Astis et al., 2006; Bianchini et al., 2008), near vertical slab rupture (slab windows model; e.g., Schiattarella et al., 2005; D'Orazio et al., 2007), or more complex scenarios involving horizontal detachment of subducted slab in post-collisional conditions, coupled with vertically-propagating slab tearing and differential slab roll-back (Rosenbaum et al., 2008), have been alternatively proposed in the literature to explain the origin of the igneous activity at Mount Vulture and the Roman district in general. In both cases, a toroidal mantle flow would have put in contact uncontaminated African (Adriatic) foreland mantle sources with subduction-related fluids and melts related to the Apennine subduction system, generating a hybrid mantle source with subduction-related geochemical features, but also with still partially preserved pre-metasomatic compositions (ranging from OIB- to DMM-like; e.g., Peccerillo, 2005; D'Orazio et al., 2007; Bianchini et al., 2008; Rosenbaum et al., 2008; Avanzinelli et al., 2009).

5) The Aeolian Islands district (~1.3–0 Ma, including seamounts; e.g., De Astis et al., 2000; Calanchi et al., 2002; Gioncada et al., 2003; Francalanci et al., 2004a,b, 2007). Out of the seven subaerial volcanic edifices – Filicudi, Alicudi, Panarea, Lipari, Vulcano, Salina and Stromboli – only Stromboli and Vulcano are currently active. In general, igneous rocks range from arc-tholeiite, calcalkaline, high-K calcalkaline, potassic (shoshonitic) to ultrapotassic (leucititic) series, with intermediate to evolved compositions prevailing. The Aeolian islands, with several seamounts – from west to east, Eolo, Enarete, Sisifo, Marsili, Palinuro, Glabro, Alcione and Lametini – form a ring-like volcanic chain (Beccaluva et al., 1982; Francalanci et al., 2007, and references therein). The Aeolian volcanoes were erupted onto ~15–20 km-thick continental crust, above a seismically active, steeply northwest-dipping (~60–70°) Benioff–Wadati plane, a part of the Ionian (Mesogean Ocean) subduction system (e.g., Chiarabba et al., 2008 and references therein). The Aeolian activity shows a spatial-temporal association with the voluminous activity of the neighbouring Mount Etna, characterized by 'anorogenic' igneous activity (Lustrino and Wilson, 2007; Viccaro and Cristofolini, 2008, and references therein). This apparent anomaly reflects the proximity of the Aeolian activity (i.e., toward the trench) to that associated with the exotic terrane comprising Calabria (the 'toe' of Italy) and the Peloritani Mountains (on the northeastern-most edge of Sicily), the latter separated, respectively, from the Apennine–Maghrebian belt by the Sangineto and Taormina tectonic lineaments (e.g., Elter et al., 2004, and references therein). The Peloritani Mountains and Calabria blocks ('Pe' and 'Ca' in AlKaPeCa; Fig. 4), were detached concomitantly from Sardinia–Corsica, while the latter drifted away from southern Europe (Bouillin et al., 1986; Gueguen et al., 1998; Lustrino, 2000a; Guerrera et al., 2005; Lustrino et al., 2009, Carminati et al., accepted for publication, and references therein). The igneous activity starts with emplacement of calcalkaline products (Sisifo seamount), succeeded by shoshonitic (Eolo and Enarete seamounts) and finally ultrapotassic magmas (Francalanci et al., 2007 and references therein). Such a general increase in potassic character of the magma is recorded in several islands (e.g., Vulcano and Stromboli), but this is not a rule for the entire district (e.g., Peccerillo, 2005; Francalanci et al., 2007). The origin of igneous activity in the Aeolian district is considered to be related to metamorphic reactions within the Ionian subducting plate at depths of ~100 to 250 km. These slab dehydration processes metasomatize the mantle wedge, lowering its solidus temperature and therefore favouring partial melting (e.g., Peccerillo, 2005; Francalanci et al., 2007; Chiarabba et al., 2008 and references therein).

6) The Tyrrhenian Sea district (~4.3–0.1 Ma; Vavilov and Marsili basins and seamounts; i.a., Beccaluva et al., 1990; Sborshchikov and Al'mukhamedov, 1992; Trua et al., 2002, 2004, 2007, 2010). The oceanic crust is confined to the Vavilov and Marsili sub-basins. In the Vavilov basin (~4.3–3.5 Ma) two large volcanic structures have been identified: the Vavilov seamount (~0.7–0.1 Ma) and the

Magnaghi seamount (~3.0–2.7 Ma; Trua et al., 2004, 2007). A single large volcano, the Marsili seamount (<0.7 Ma), has been identified in the Marsili basin (~1.9–1.7 Ma). Geochemical data have been acquired for igneous rocks sampled from scattered sites such as the Cornacya seamount — offshore southeast Sardinia (~12.6 Ma; Mascle et al., 2001), the Vercelli, Baronie and Etruschi seamounts (offshore east Sardinia; e.g., Trua et al., 2007 and references therein), Aceste, Anchise and Prometeo seamounts, and Ustica Island (offshore northwest Sicily; e.g., Trua et al., 2003, 2004, 2007 and references therein), and from ODP Sites 650 and 655 and DSDP Site 373 (Marsili basin), ODP Site 654 (offshore east Sardinia), ODP Site 651 (Vavilov basin), and in the vicinity of the Aeolian Islands (Beccaluva et al., 1985b, 1990; Francalanci et al., 2007; Trua et al., 2007, and references therein). In the Tyrrhenian Sea domain we consider data only for Plio-Quaternary samples associated with the formation of oceanic crust of the Vavilov and Marsili basins. The composition of the Marsili seamount lavas range from T — (Transitional) and E — (Enriched) type MORB to calcalkaline basaltic and high-K andesitic types (Fig. 15). Whole rock analyses for the Magnaghi and Vavilov seamounts are scarce but indicate mildly sodic hawaiites, moderately enriched in Nb (e.g., Robin et al., 1987). These compositions are anomalous with respect to typical oceanic crust of (e.g.) North Atlantic type and suggest the existence of sources affected by slab-derived melts and/or fluids. Four published Sr–Nd–Pb isotopic analyses for Tyrrhenian Sea rocks show ranges of $^{87}Sr/^{86}Sr_{(i)}$ from 0.7037 to 0.7058, $^{143}Nd/^{144}Nd_{(i)}$ from 0.51311 to 0.51244, $^{206}Pb/^{204}Pb_{(i)}$, from 18.70 to 18.92, $^{207}Pb/^{204}Pb_{(i)}$ from 15.57 to 15.68, and $^{208}Pb/^{204}Pb_{(i)}$ from 38.67 to 39.18 (Fig. 16). Δ7/4 and Δ8/4 range from ~ +59 to ~+13 and ~+43 to ~+68, respectively. The observed positive correlation of $^{206}Pb/^{204}Pb$ vs. $^{87}Sr/^{86}Sr$ and the negative correlation of $^{206}Pb/^{204}Pb$ vs. $^{143}Nd/^{144}Nd$ contrast with patterns observed in most other Italian rocks (Fig. 16).

10. Discussion

Compared to most subduction- and collision-related tectonic settings, the geological features of the CWM and the compositional heterogeneity of igneous products are quite complex. On the basis of geological and geochemical data synthesized in this paper, the key characteristics of the CWM may be summarized as follows:

1) The CWM developed within an overall shortening field generated by the collision of Africa with Eurasia. Several small, generally V-shaped, extensional basins opened progressively, explained in terms of a range of highly contrasting mechanisms. These include: back-arc sea-floor spreading (e.g., the Ligurian–Provençal Basin, Valencia Trough and Tyrrhenian Sea), anomalous mantle upwelling (e.g., Ligurian–Provençal Basin and Tyrrhenian Sea), eastward asthenospheric flow, lateral expulsion of crustal wedges (e.g., Sardinia, northwest and northeast Tell Chain and northeast Sicily), slab break-off (Tyrrhenian Sea), perturbation of accumulated subducted material at the ~660 km discontinuity, rapid subduction rollback, interactions of detached lithospheric roots with ductile upper mantle (e.g., Alborán Sea), and periodic reversals of subduction polarity (Northern Algerian Basin, southern Tyrrhenian Sea and Ligurian–Provencal Basin). At present, active continental rifting, apparently unassociated with back-arc spreading (e.g., Civile et al., 2008 and references therein) still continues, as observed in the Sicily Channel, despite the above-noted regional shortening field.
2) Currently active and inactive subduction zones have recycled both continental and oceanic lithosphere, and show significant variations of polarity in space and time, in general from south-southeast [e.g., Early Cretaceous–Eocene central-western Alps (northern Italy, Suisse) and Betic Chain (Spain)], to the east [e.g., Late Cretaceous–Eocene western Alps (Italy, France) and northeast Corsica (in present-day coordinates; Early Miocene–Present Alborán Sea (Spain–Morocco)], northwest [e.g., Late Eocene–Middle Miocene arc in Sardinia–Corsica; Middle Miocene–Present Calabrian Arc (Italy)], N [e.g., Middle Miocene–Present Peloritani (Sicily, Italy), Tellian (Algeria–Tunisia) and Rifian Belts (Morocco–northwestern Algeria)] and southwest [Apennines (peninsular Italy)] (Carminati and Doglioni, 2004; Carminati et al., accepted for publication, and references therein).
3) In general, magmatic districts exhibit temporal sequences involving a progression from subduction- and collision-related episodes to anorogenic (intraplate) magmatic activity. This pattern is observed in the Betic-Rif system (e.g., Turner et al., 1999; Maury et al., 2000; Coulon et al., 2002; Duggen et al., 2005, and references therein), the Valencia Trough (e.g., Martì et al., 1992), Sardinia (e.g., Lustrino et al., 2004a,b, 2007b), Provence–Languedoc (e.g., Liotard et al., 1999; Beccaluva et al., 2005) and Tell (Tunisia; e.g., Maury et al., 2000; Talbi et al., 2005). Elsewhere, the activity is exclusively subduction-related, in some cases showing the effects of continental collision, anorogenic activity being largely absent (e.g., Late Miocene–Present peninsular Italy; Peccerillo, 1985, 2005; Beccaluva et al., 1991; Conticelli et al., 2007; Avanzinelli et al., 2009), Middle Eocene–Late Oligocene western, central and eastern Alps (e.g., Venturelli et al., 1984; von Blanckenburg and Davies, 1995; Callegari et al., 2004; Rosenberg, 2004). However, some magmas do exhibit intraplate-like character only, as in the Pliocene to Present Pantelleria–Linosa Islands in the Sicily Channel (Lustrino and Wilson, 2007; Di Bella et al., 2008, and references therein). Finally, complex, coeval igneous associations, for example, within narrow areas in Sicily and within the Alpine Chain, may show diverse geochemical characteristics. The intraplate-type Pleistocene to Present activity of Mount Etna (e.g., Clocchiatti et al., 1998; Armienti et al., 2004; Peccerillo, 2005; Viccaro and Cristofolini, 2008) occurs only a few tens of km south of the coeval Aeolian Island volcanic district, which shows subduction-related geochemical character (e.g., Francalanci et al., 2004a,b, 2007, Peccerillo, 2005; Tommasini et al., 2007). Likewise, the Late Paleocene to Oligocene intraplate activity in the Veneto Area, in northeastern Italy (Fig. 3; e.g., Beccaluva et al., 2007b; Macera et al., 2003, 2008), occurs only a few tens of km south of the roughly coeval collision-related Periadriatic Fault activity (e.g., Beccaluva et al., 1983; von Blanckenburg and Davies, 1995; Rosenberg, 2004).
4) In a few cases, magmas of unambiguously intraplate affinity, emplaced in foreland settings or distal to active subduction zones may show unexpectedly potassic character, expressed by modal leucite and/or analcime pseudomorphs over leucite. Examples include CWM as well as central-eastern European Cenozoic igneous domains: Late Miocene–Late Pliocene Calatrava leucitites (central Spain; e.g., Cebriá and Lopez-Ruiz, 1995), Quaternary leucite basanites from Olot–Garrotxa (northeast Spain; Cebriá et al., 2000), Pliocene Cantal leucite nephelinites and leucite basanites (French Massif Central; e.g., Wilson and Downes, 1991), Late Pliocene Kecel–Bar leucitites (Pannonian Basin; Harangi et al., 1995), Pleistocene Eifel leucitites (Germany; Mertes and Schmincke, 1985), Pliocene Montiferro analcime basanites (Sardinia, Italy; Fedele et al., 2007). These potassic lithologies are distinct from the (invariably HFSE-depleted) collision-related potassic/ultrapotassic magmas that characterize peninsular Italy, and are clearly unrelated to coeval or recent subduction, or collisional processes (e.g., Conticelli et al., 2004, 2009b; Peccerillo, 2005; Lustrino and Wilson, 2007, and references therein).
5) Most of the peninsular Italy volcanic rocks (e.g., Peccerillo, 2005, and reference therein) are SiO_2-saturated to SiO_2-undersaturated and potassic to ultrapotassic in character. The relative scarcity of such magmas in 'normal' subduction settings (usually characterized by calcalkaline to high-K calcalkaline magmatism). On the

basis of both geological and geophysical data (see below) several authors have proposed the existence of active upwelling from the deep mantle (e.g., Gasperini et al., 2002; Bell et al., 2006; Cadoux et al., 2007), effectively excluding any role for subduction *per se* beneath the Apennine Chain (e.g., Lavecchia and Stoppa, 1996; Locardi and Nicolich, 2005; Lavecchia and Creati, 2006; Centamore and Rossi, 2008).

6) In some areas where 'subduction-type' igneous activity might be expected (e.g., above the hanging-wall of the Late Cretaceous to Eocene Alpine Adria–European subduction system) there is little or no sign of igneous activity. The bulk of subduction-related activity in the Alps occurred during the Oligocene (~34–28 Ma), following the complete subduction of Alpine Tethian lithosphere beneath the Adria plate (e.g., von Blanckenburg and Davies, 1995; Garzanti and Malusà, 2008, and references therein). On the other hand, in areas where significant lithospheric extension is recorded, as represented, for example, by the Oligo-Miocene Sardinian Trough formation (Cherchi and Montadert, 1982; Faccenna et al., 2002; Cherchi et al., 2008), voluminous subduction-type eruptives have been recorded (e.g., Brotzu et al., 1997a,b; Morra et al., 1997; Franciosi et al., 2003; Lustrino et al., 2004a,b, 2009, and references therein).

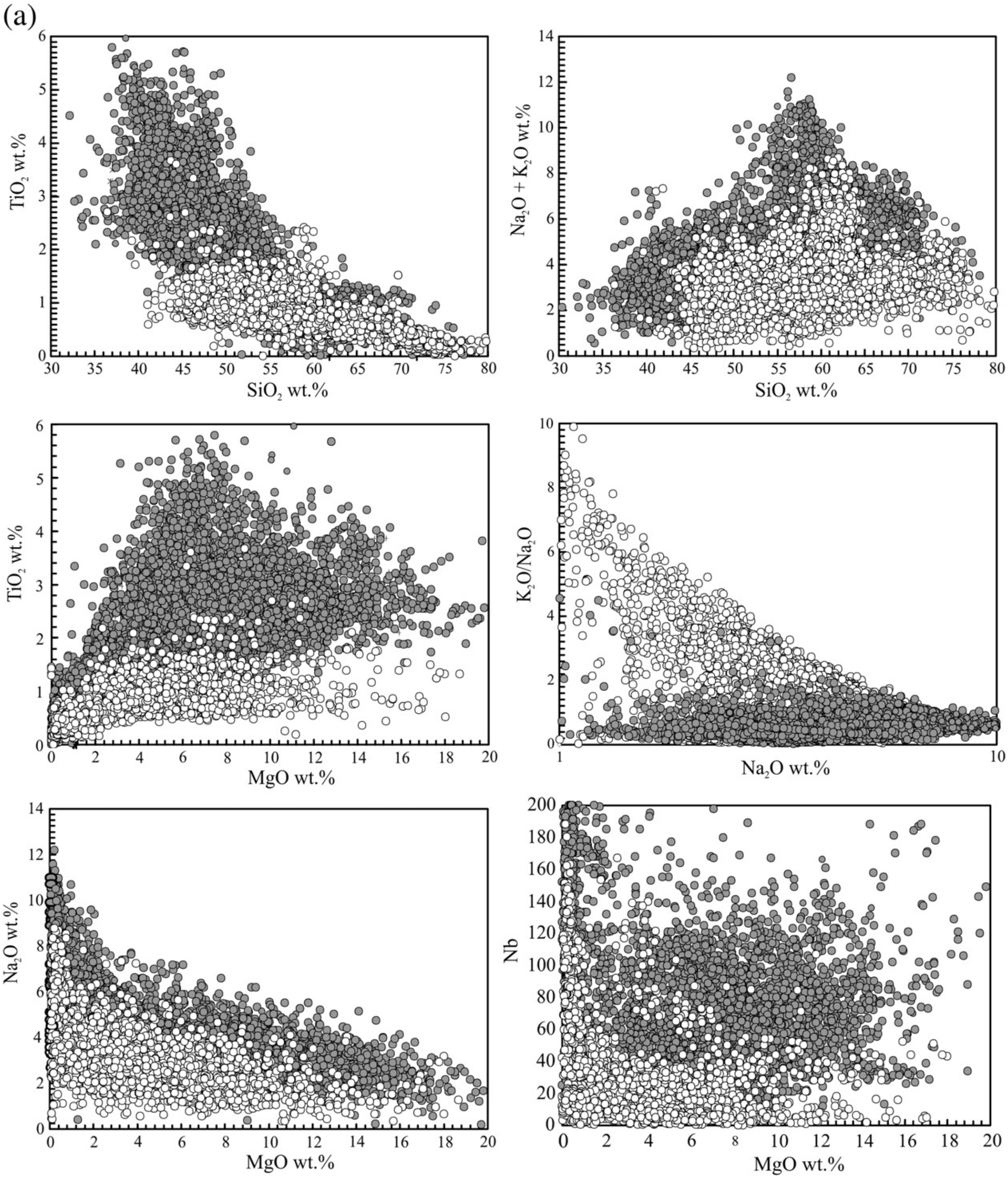

Fig. 17. Selected major and trace element variation diagrams and Sr–Nd isotopic ratios of subduction-related rocks from central-western Mediterranean (white circles; references in captions of Figs. 5, 9, 12 and 15) compared to the Circum-Mediterranean Anorogenic Cenozoic Igneous (CiMACI) rocks (grey circles; references in Lustrino and Wilson, 2007).

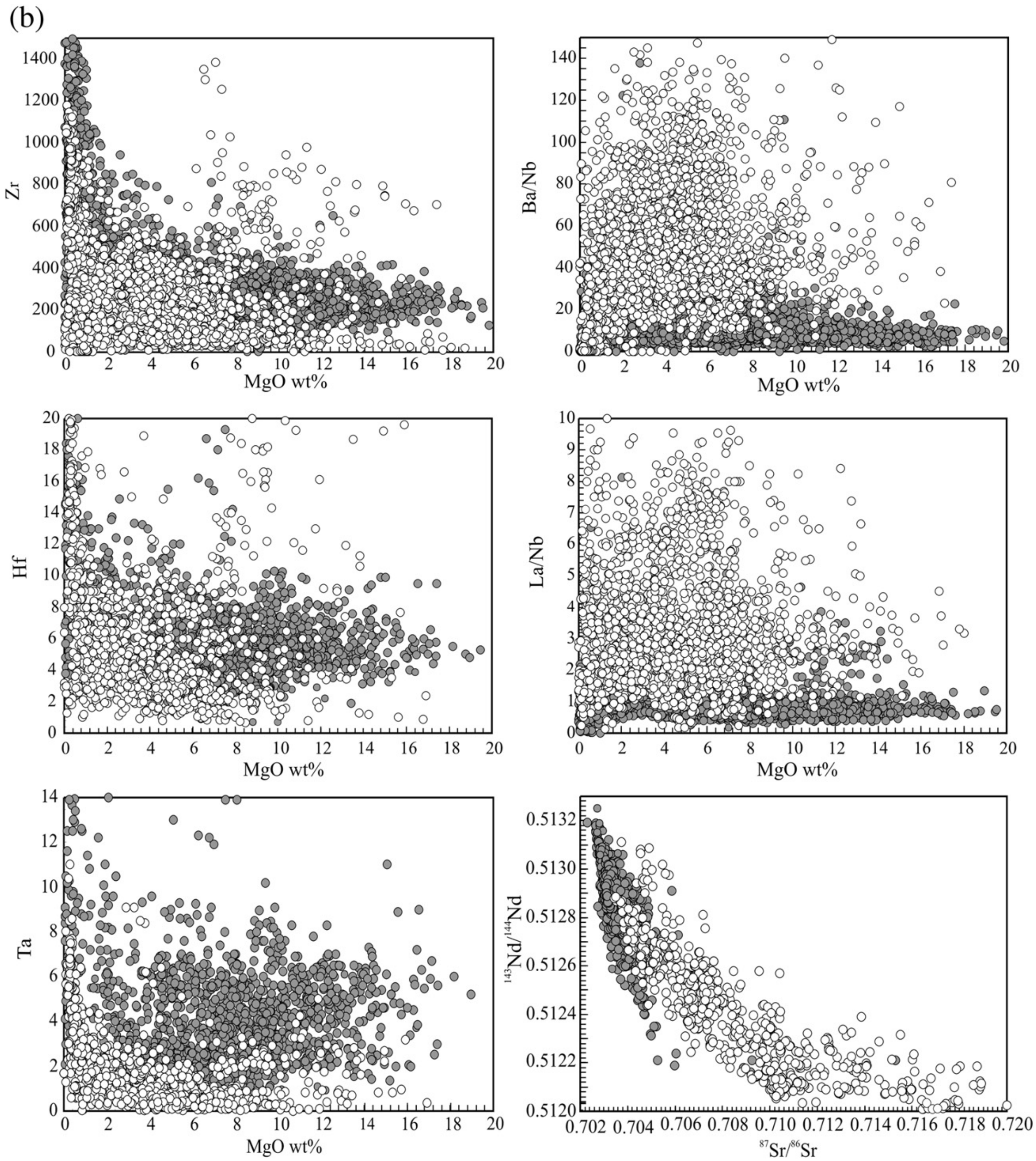

Fig. 17 (*continued*).

11. The pitfalls of translating geochemical terms into geological concepts

Most controversial proposals for explaining CWM igneous activity stem from the relatively different significance assigned to key major and trace element abundances, and specific radiogenic (e.g., Rb–Sr, Sm–Nd, U–Th–Pb, Lu–Hf, Re–Os) and stable (e.g., O, He, C) isotope systematics. Another contentious area pertains to the notion of unequivocal geochemical characterizations of geodynamic activity, e.g., 'orogenic' or 'subduction-related', 'anorogenic', 'intraplate-like', 'plume-related, etc. (e.g., Wilson and Bianchini, 1999; Lustrino, 2000b; Wilson and Downes, 2006; Harangi et al., 2006; Lustrino and Carminati, 2007; Lustrino and Wilson, 2007). In particular, the prevalence of 'plume' models in recent decades has been sustained by many researchers by hypothesizing active mantle upwellings — whether from the core-mantle boundary layer, at 2900 km, or from the upper mantle/lower mantle transition zone, at ~660 km–largely on the basis of geochemical arguments.

Clearly, the composition of igneous rocks associated with active or recently active subduction zones, orogenies, mid-ocean ridges, and back-arc basins can be unambiguously characterized as such. However, the origin of their diagnostic patterns, are still poorly understood in many cases. With regard to the composition of subduction-related magmas, even those clearly associated with active subduction fronts, their variation is a result of factors that are often still hard to constrain. These may include: 1) compositional variation of mantle wedge regions before they have been affected by subduction-related metasomatism, e.g., either refractory (melt-depleted) or fertile in character; 2) variations in the composition of subducted material, for example distinctions between continental vs. pelagic sediments, altered vs. pristine oceanic crust, Si- or Ca-rich lithologies, and so on; 3) the common effects of crustal contamination of mantle melts at shallow depths (e.g., AFC-type processes), especially where oceanic lithosphere under-thrust continental plates; 4) different types of metasomatic modification of the mantle wedge — these may consist of partial melts of the down-going, eclogitized slab and/or sedimentary component, the composition of the fluids released by slab dehydration reactions with or without slab melting; 5) the depths at which the main dehydration reactions occur, typically at pressures of ~3 kbar, but conceivably ranging from ~70 to more than 200 km; 6) the physical state of metasomes, potentially occurring as pervasive networks of enriched veins or massive fronts of orthopyroxene-rich peridotite; 7) the occurrence of 'anorogenic' and/or 'transitional' magmas in back-arc basins; 8) the potential of ancient metasomatic effects recorded in the lithospheric mantle.

Bearing in mind all such *caveats*, we advance a series of necessary considerations on the geochemical character of CWM subduction-related and intraplate-like magmas. These considerations are based on the large database for 1) intraplate-like (anorogenic) Cenozoic igneous rocks (more than 8400 whole-rock analyses) for the entire circum-Mediterranean area (CiMACI Province; Lustrino and Wilson, 2007) and 2) a new database of subduction-related rocks (comprising more than 7000 whole rock analyses) available in the electronic appendix of this paper. The reader should pay particular attention to our definition of 'subduction-related' affinity on geochemical grounds only. There are many exceptions to overly strict definitions and therefore such criteria cannot be considered as universally valid.

The following basic points need to be stressed, when attempting to designate a rock as 'subduction-related' in the CWM:

a) Subduction-related rocks are typically richer in SiO_2 compared to the intraplate-like igneous rocks. Generally SiO_2 is >45 wt.% for most of the less evolved subduction-related rocks, compared to SiO_2 <50 wt.% for the majority of less evolved intraplate-like rocks; Fig. 17). In other words, ultrabasic rocks among the subduction-related group are rare or absent. Moreover, with the exception of a few MgO-rich samples (e.g., lamproites), the bulk of the subduction-related rocks CWM have MgO <9 wt.%, in contrast to about half of the CiMACI rocks, which show MgO >7 wt.% (in some cases up to ~20 wt.% MgO; Fig. 17).
b) Among alkaline compositions, the subduction-related rocks show a potassic to ultrapotassic composition with K_2O/Na_2O ratios ranging from ~1 to ~10 in the most differentiated terms. By comparison, K_2O/Na_2O ratios of the CiMACI rocks rarely exceed 1.5 (Fig. 17). According to the IUGS-endorsed definition (Le Maitre, 1989), the term ultrapotassic can only be applied where K_2O/Na_2O ratio is >3, although the classification of Foley et al. (1987) requires K_2O/Na_2O >2 and MgO >3 wt.%. K-rich magmas imply the presence of K-bearing phases entering the partial melt during mantle anatexis, usually taken to be phlogopite. It is noteworthy that this phase is very rare (<<1% in volume) in typical intraplate-basalt borne mantle xenoliths (e.g., Lustrino et al., 1999; Turner et al., 1999; Beccaluva et al., 2004; Galán et al., 2008; Rampone et al., 2009; Bianchini et al., 2010; Villaseca et al., 2010), and even more rare in mantle xenoliths associated with subduction-related rocks (e.g., Conticelli and Peccerillo, 1990).

   Typical subalkaline subduction-related rocks are invariably calcalkaline in character (lacking the marked increase in Fe content with progressive fractionation, attributed to early precipitation of magnetite in relatively high fO_2 conditions (e.g., Kuno, 1968; Miyashiro, 1974). Early definitions of 'calcalkaline' (e.g., Peacock, 1931) had no particular geodynamic significance, referred simply to magmas whose CaO and $K_2O + Na_2O$ contents were more or less the same within a SiO_2 range of 56 to 61 wt.%. They were typically considered to conform Hawaiian or Icelandic tholeiites (e.g., Peacock, 1931; Hatch et al., 1961), in contrast to the current view that the term is nearly a synonymous of 'subduction-related' (e.g., Peccerillo and Taylor, 1976). It should also be noted that 'calcalkaline' character as defined by Kuno (1968), Miyashiro (1974) and Peccerillo and Taylor (1976) does not conform to Peacock's (1931) criteria with respect to CaO and total alkali content. For more detailed discussion of the geodynamic significance of 'calcalkaline' affinity the reader is referred to reviews by Sheth et al. (2002) and Arculus (2003). This represents a clear case where caution must be adopted when using specific terms based on unclear criteria.
c) The subduction-related rocks are characterized by relatively low TiO_2 (<1.5 wt.%, often <1.1 wt.%) over a broad spectrum of MgO content, as compared to consistently higher values of the same oxide in rocks of intraplate affinity like those of the CiMACI Province (TiO_2>1.2 wt.%, up to 5 wt.%), with limited overlap between the two rock groups (Fig. 17). The relatively low Ti contents are a source feature that can be related to the low K_D values and solubilities of this element (and all the other HFSE) in fluids released from downgoing slabs (e.g., Saunders et al., 1980; Tatsumi et al., 1986; Ryerson and Watson, 1987; Foley and Wheller, 1990). Noteworthy, the Ti trough often refers to early precipitation of Fe–Ti oxide, favoured by generally high fO_2 conditions rather than to a direct mantle source message.
d) Following from the above, subduction-related rocks generally show lower Na_2O (<4.0 wt.%) contents in more mafic (MgO >5 wt.%) magmas, again contrasting volcanics of the CiMACI Province for equivalent MgO content (Fig. 17). On the other hand, variation in K_2O contents is significantly greater, values ranging from near-zero to ~8 wt.% (not shown). Comparing concentration ranges of alkalies in subduction-related and intraplate-like rocks is handicapped, given the considerable variation in the extent of fractionation of these oxides. Other major elements (e.g., Al, Ca, Fe P) are not especially diagnostic in distinguishing 'subduction-related' from 'anorogenic' magmas.
e) As incompatible elements, both LILE and LREE contents will vary as a function of the degree of partial melting, but the high LILE/HFSE and LREE/HFSE ratios of subduction-related rocks, as compared to intraplate magmas are highly distinctive (e.g., Ba/Nb is generally >20 and La/Nb is generally >1.4 for the subduction-related rocks, while the same ratios are generally <20 and <1.3 for the CiMACI rocks; Fig. 17). Some subduction-related lavas show strong negative Ba anomalies, as reflected in their relatively low Ba/Nb ratios (e.g., Peccerillo, 2005).
f) It is often assumed that HFSE (i.e., Nb, Ta, Hf, Zr, Ti) depletions are exclusive to subduction-related magmas as compared to those of intraplate affinity. This may not, however, be taken as a rigid distinction. While, indeed, subduction-related magmas typically show Ta–Nb troughs compared with intraplate types in primitive mantle-normalized diagrams, Hf and Zr may lack such 'troughs' and even show 'peaks'. Niobium content is generally lower than 25 ppm in samples with MgO >5 wt.%, compared to Nb 10–120 ppm in CiMACI Province rocks with MgO >5 wt.% (Fig. 17). Fig. 17 shows the larger overlap between subduction-related and CiMACI rocks in terms of Zr and Hf.
g) The Sr isotopic composition of subduction-related, as compared to 'anorogenic' rocks, is generally more radiogenic. CWM subduction-related lithologies show $^{87}Sr/^{86}Sr_{(i)}$ ratios greater than 0.7035 (irrespective of MgO content), more than 90% of the samples in the database showing $^{87}Sr/^{86}Sr_{(i)}$ ratios in excess of the estimated BSE (Bulk Silicate Earth) average of 0.70445. In contrast, more than

95% of CiMACI $^{87}Sr/^{86}Sr_{(i)}$ ratios fall between 0.7027 and 0.7047 (Fig. 17).

h) Subduction-related rocks generally show less radiogenic $^{143}Nd/^{144}Nd_{(i)}$ isotopic ratios than those of anorogenic affinity. More than 95% of the CWM subduction-related rocks, across the spectrum of MgO contents, have $^{143}Nd/^{144}Nd_{(i)}$ <0.5128, ranging down to 0.5120. Again, the contrast with CiMACI rocks is striking, more than 95% of these latter showing $^{143}Nd/^{144}Nd_{(i)}$ ratios above the ChUR (Chondritic Uniform Reservoir) estimate (0.51264), falling between 0.5132 and 0.5126 (Fig. 17). In Fig. 17 a small subset of subduction-related samples plot in an anomalous field of relatively radiogenic Sr ($^{87}Sr/^{86}Sr_{(i)}$>BSE), but strongly radiogenic Nd ($^{143}Nd/^{144}Nd_{(i)}$ >0.5129). Most of these are from the Alborán Island or adjacent seamounts. Anomalous compositions also include gabbros from the Tell and dredged basalts from the Vavilov basin.

i) The two magma types, subduction-related and anorogenic, are not so clearly distinguished in terms of their lead isotopic compositions. More than 95% of the subduction-related rocks of the CWM have $^{206}Pb/^{204}Pb_{(i)}$ comprised between 18.6 and 19.5, indistinguishable from the values of more of the 95% of the CiMACI rocks ($^{206}Pb/^{204}Pb_{(i)}$ = 18.7–19.7; Fig. 17). Also $^{208}Pb/^{204}Pb_{(i)}$ is indistinguishable between the CWM subduction-related rocks and the CiMACI rocks (38.5–39.4 and 38.5–39.7, respectively). The subduction-related rocks seem to have slightly more radiogenic thorogenic lead ($^{208}Pb/^{204}Pb$) for a given $^{206}Pb/^{204}Pb$ ratio. More than 90% of the CWM subduction-related rocks have $^{207}Pb/^{204}Pb_{(i)}$ >15.65 (up to 15.71), whereas more than 95% of the CiMACI rocks have $^{207}Pb/^{204}Pb_{(i)}$ <15.66 (down to ~15.49). This feature (radiogenic $^{207}Pb/^{204}Pb$) appears to be typical of subduction-related rocks world-wide and has been related to relatively fast isotopic decay of a ^{235}U-enriched mantle source modified by the interaction with ancient continental components. This takes place because ^{235}U decays into ^{207}Pb with a half-time of $7.04*10^8$ yr, much shorter compared to the half-time of ^{238}U decaying into ^{206}Pb with a half-time of $4.47*10^9$ yr. In order to have ^{207}Pb excess with "normal" ^{206}Pb, it is not possible to invoke the derivation from a source that evolved with a single-stage μ (where $\mu = {}^{238}U/{}^{204}Pb$ isotopic ratio; Zartman and Doe, 1981; Kramers and Tolstikhin, 1997; Murphy et al., 2002). A more complex multi-stage evolutionary process is therefore necessary to explain the $^{207}Pb/^{204}Pb$ excess of subduction-related magmas, involving fractionation of U/Pb ratios during partial melting processes, recycling of crustal material and insulation of modified volumes of mantle for several Gyrs. In conclusion, the $^{207}Pb/^{204}Pb$ ratio might be considered to be the most diagnostic Pb isotopic parameter.

j) Concerning oxygen isotopes, in view of the susceptibility of $\delta^{18}O$ values to secondary deuteric processes and late-stage percolating fluids in evolved magma systems (e.g., Fourcade et al., 2001; Downes et al., 2001; Eiler, 2001), only laser ablation measurements on fresh and unaltered fresh separates of mafic phases or fresh glasses should be prioritized. Most of the oxygen isotope data in the literature on CWM subduction-related rocks were obtained for whole-rock aliquots, thereby rendering their petrologic significance to be qualitative, at best. However, recent years have seen a significant increase in high quality laser fluorination analyses of oxygen isotope compositions in mineral separates such as olivine, pyroxenes and feldspars, at least for the Italian and Betic–Alborán–Rif samples (e.g., Downes et al., 2001; Dallai et al., 2004; Duggen et al., 2004, 2008; Peccerillo et al., 2004a,b; Lustrino et al., 2008). $\delta^{18}O$ values in olivine separates of about +5.2‰ are considered to reflect a mantle source approximating DMM (Depleted MORB Mantle) character, with little or no sign of crustal contamination during magma ascent (e.g., Mattey et al., 1994; Eiler, 2001). In contrast, higher $\delta^{18}O$ values measured in mafic phases (the oxygen isotopic fractionation between olivine and clinopyroxene $\delta^{18}O_{cpx} = \delta^{18}O_{ol} + 0.4$‰ at ~1200 °C; Mattey et al., 1994) have been ascribed to crustal contamination at mantle depths.

In summary, it is clear that unambiguous diagnostic geochemical or mineralogical tools for characterizing subduction-related rocks are lacking. The analogous conclusion appears to apply to magmas of intraplate affinity too (Lustrino and Wilson, 2007; Lustrino and Carminati, 2007).

12. The role of subduction vs. upper mantle active upraise: the Tyrrhenian Sea case study

Bearing the above observations and discussions in mind, this review is concluded with a case study of the Tyrrhenian Sea and contiguous regions. Three key features remain the subject of intense debate: 1) the geochemical significance of potassic–ultrapotassic magmatism along peninsular Italy, 2) the geodynamic mechanisms giving rise to the Tyrrhenian Sea, and 3) the associated igneous activity (e.g., Peccerillo, 1985, 2005; Beccaluva et al., 1991; Peccerillo and Lustrino, 2005; Conticelli et al., 2002, 2007, 2009b; Avanzinelli et al., 2009; Carminati et al., accepted for publication). Given the lack of any consensus (after several decades of intense geological, geochemical and geophysical study), we will attempt a critical review of some recent key papers (i.e., Gasperini et al., 2002; Locardi and Nicolich, 2005; Bell et al., 2006; Lavecchia and Creati, 2006; Scalera, 2006; Cadoux et al., 2007). These share in common the thesis (based essentially on geochemical arguments) that magmas of peninsular Italy and the southeastern part of the Tyrrhenian Sea exhibit at least one characteristic source component, located in the lower mantle. We do not exclude 'mantle plume' models as a potential factor for magma genesis, but we emphasize that there is no necessity to invoke such a large-scale process (transport of mass from very deep to very shallow levels) to explain the composition and timing of igneous activity associated with the Tyrrhenian Sea, specifically, and CWM in general. Reconciling large-scale mantle process with the peculiar, small-scale spatial and temporal heterogeneity of magmatism in the Tyrrhenian region is certainly a major challenge.

On the base of geochemical data and tomographic imagery, Piromallo et al. (2008) proposed a lengthy (~70 Myr) and widespread period of igneous activity – extending from the Cape Verde Islands (North Atlantic) northwards to central and eastern Europe (Germany, Silesia, the Pannonian Basin) and including Libya (Northern Africa) – resulting from a common mantle source, associated with a hypothetical 'Central Atlantic Plume' (Piromallo et al., 2008). However, this model is based on some relatively unconstrained assumptions, namely that: a) similar geochemical compositions of igneous rocks denote the presence of similar physical mantle reservoir or reservoirs; b) certain isotopic ratios or incompatible trace element ratios (the so-called geochemical 'fingerprints') in igneous rocks can provide unequivocal proof of a source derived from the deep mantle; c) tomographic images depicting fast and slow V_s and V_p seismic anomalies, respectively, represent clear evidence of cold and warm regions of the mantle. Here we suggest that none of these assumptions represent unequivocal criteria upon which to base a robust geodynamic model.

On a relatively smaller scale, Bell et al. (2006), exclusively on the basis of Sr–Nd–Pb isotope data, have proposed that most igneous activity in Italy can be attributed to the presence of a mantle plume rather than, as is more conventionally believed, collisional tectonics and Mesogean Oceanic subduction. In particular, the 'plume' would be at least 1000 km in diameter, extending westwards beneath Sardinia and Corsica, to the northwest under the northwestern Alps and in the east underlying peninsular Italy (Bell et al., 2006). Partial melting of the isotopically heterogeneous plume head is considered to explain many of the features that characterize Italian magmatism, while reconciling it with associated extensional tectonics, lithosphere thinning, high heat flow, and prolific CO_2 emissions that characterize the region (Bell et al., 2006).

Cadoux et al (2007) also claim to have identified a 'unique lower mantle source' for Plio-Quaternary volcanics in southern Italy, on the by statistical analysis of data for mildly to strongly evolved lava samples, some of which have been interpreted as products of pure crustal melts or hybrid mixtures of upper mantle and crustal components (e.g., Poli, 2004; Peccerillo, 2005). This represents a first-order problem — that of interpreting mantle source characteristics from data for highly evolved or strongly crustally-contaminated melts (albeit of mantle provenance). It has been demonstrated elsewhere (e.g., Peccerillo, 2005; Conticelli et al., 2007, 2009a; Avanzinelli et al., 2009) that evolved and mafic magmas from Tuscany and the northern part of the Roman Province are not genetically related and that other processes (e.g., crustal contamination of mantle-derived melts via wall-rock reaction or mixing of crustal and mantle melts) played an important role (Feldstein et al., 1994; Poli, 2004; Peccerillo, 2005).

Lavecchia and Creati (2006) propose a slightly different model, invoking the presence of a 'wet' rather than a 'hot' plume. This plume (with an anomalous triangular shape in plain view) would not be characterized by a thermal anomaly but by anomalous local enrichment in H_2O- and CO_2, thereby sufficiently reducing viscosity, to allow upwelling. The volatiles (and the plume itself) are postulated to tap the core-mantle boundary (Lavecchia and Creati, 2006). Rapid dilation (up to 3 cm/yr) of the plume head on reaching the lithospheric boundary would be asymmetric towards the east, as a consequence of ambient large-scale eastward-directed upper mantle circulation. As a yet more radical proposal, Lavecchia and Creati (2006) contend that «the Apennines and the Maghrebides are not related to subduction, but represent a recent example of plume-induced orogenesis». Locardi and Nicolich (2005) arrived at a similar conclusion, proposing an eastward migrating, deep-seated thermal plume ('asthenolith') to explain the progressive opening of basins in the CWM, the reduction of lithospheric thickness, the rotation and collision of several microplates, and the genesis of compositionally diverse magmas. Each of these models appears to follow the original concept of Wezel (1982).

Cadoux et al. (2007) assume that the asthenosphere comprises «ubiquitous depleted upper mantle» and, in contrast to recent S-wave tomography models (Panza et al., 2007a), that the asthenosphere beneath Italy is «particularly cold, likely as a result of the multiple Mediterranean subduction systems». They therefore assumed *a priori* that magma genesis is not triggered in the upper mantle and, not surprisingly, that it taps sources convecting upwards from the lower mantle.

Bell et al. (2006) support this kind of model by assuming that so-called FOZO and EM1 mantle isotopic components are intrinsic to the lower mantle, such that *only* deep mantle upwelling can give rise to mantle melting. FOZO and EM1 are terms commonly used in the geochemical jargon. FOZO is an acronym indicating FOcal ZOne, corresponding to a geochemical composition proposed on the basis of Sr–Nd–Pb–He isotopic ratios that should represent a physical region of the Earth located in the lower mantle, near the core-mantle boundary (Hart et al., 1992). On the other hand, EM1 is an acronym indicating Enriched Mantle type 1, a hypothetical geochemical end-member with very peculiar Sr–Nd–Pb isotopic characteristics whose origin is still strongly debated (e.g., see discussion in Lustrino and Dallai, 2003). This approach pre-ordains the final conclusion and flies in the face of numerous geochemical studies that show the opposite (e.g., Stracke et al., 2005; Willbold and Stracke, 2010).

Given that the upper mantle is both compositionally and thermally heterogeneous (e.g., Meibom and Anderson, 2003; Peccerillo and Lustrino, 2005; Anderson, 2007, and references therein), it is reasonable to surmise that many mantle regions have been cooled for long periods by subduction, yet are able to yield prolific arc, back-arc and continental basalts (Anderson, 2007). Pursuing this line of thinking, Lustrino (in press) argued that the sub-lithospheric mantle is geochemically homogeneous, but significantly different from MORB mantle, as widely assumed. EM1-type basalts are found both in continental and oceanic areas, in intraplate settings and along passive continental margins, associated with variable (from very low to huge) volumes of magma production, emplaced in areas without any substantial lithospheric doming or excess heat flow or in areas near subduction zones (see discussion in Lustrino and Dallai, 2003). Together with EM2-type basalts, the EM1-type basalts simply indicate derivation from a mantle source contaminated by continental (mainly upper and lower, respectively) crust (Lustrino et al., 2004a,b; Willbold and Stracke, in press), without any indication for the presence of mantle plumes.

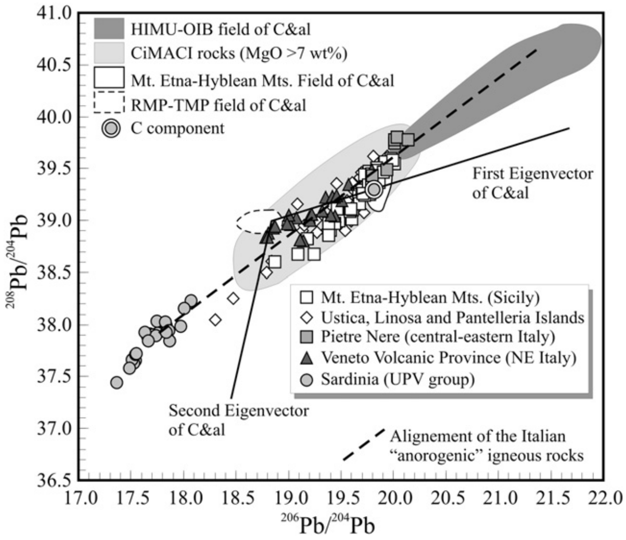

Fig. 18. $^{208}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ diagram. The field of HIMU-OIBs is from Cadoux et al. (2007; C&al). The field of CiMACI (Circum-Mediterranean Anorogenic Cenozoic Igneous) Province mafic rocks is from Lustrino and Wilson (2007). Also shown are the fields of Mount Etna-Hyblean Mountains (Sicily), Aeolian Islands and RMP (Roman magmatic Province) and TMP (Tuscan Magmatic Province) igneous rocks as given by Cadoux et al. (2007). The C component (Hanan and Graham, 1996) the so-called Common Component (also known as FOZO and PHEM) is taken by some workers as evidence for lower mantle involvement given it is a common presence in ocean island magmas, sometimes also showing high $^3He/^4He$ ratios (this is not the case for Italy, however). For definitions of mantle isotopic end-member components see Hofmann (1997) and Stracke et al. (2005). First and second eigenvectors as defined by Cadoux et al. (2007) are shown by thick continuous lines, while thick broken lines delineate isotopic compositions of the Italian 'anorogenic' mafic rocks.

Cadoux et al. (2007) consider only one model to explain the geochemistry of peninsular Italy and southern Tyrrhenian Sea volcanism, assuming that FOZO isotopic character (or '*C* component', where C stands for 'Common'; Hanan and Graham, 1996) can be assumed to indicate a lower mantle source. This notion, however, has long been discarded by numerous workers, including those who originally defined FOZO itself (Stracke et al., 2005), postulating that such assumptions are *ad hoc* and unsupported. Stracke et al. (2005) stated that all the hypothetical compositions (among which FOZO is only one) simply reflect interaction with subducted crust rather than lower mantle. In orogenic and arc environments, contributions from delaminated crust, in addition to subducted components, are expected (see also Lustrino, 2005 for a discussion of crust recycling and its effects on basalt genesis). Carlson et al. (2006) point out that the FOZO component observed in Siberia and ocean-island basalts is prevalent in the upper mantle under both continents and ocean basins. The long-lived presence of this source beneath Siberia shows that it is unnecessary to appeal to the lower mantle, or to a mechanism, such as a plume, in order to bring this type of mantle into play only during flood-basaltic episodes.

Any contribution of the upper mantle is neglected in Cadoux et al.'s model because they assume that the upper mantle has a homogeneous DMM composition. As described above southern Italy has been the locus of oceanic diachronous back-arc basin opening, subduction and obduction of both oceanic and continental lithosphere throughout much of the Phanerozoic (see above). Virtually every kind of igneous activity records successive mantle depletion and enrichment events (e.g., Doglioni et al., 1998; Lustrino, 2000a; Carminati and Doglioni, 2004, and references therein). It is unrealistic to assume that upper mantle beneath this region shows homogeneous DMM character, and is distinct from that supplying MORB magmas to mature spreading mid-ocean ridges. Igneous basement of the Tyrrhenian abyssal plain is clearly distinguishable from MORB (e.g., Trua et al., 2004, 2007, 2010), and, by no means, homogeneous or depleted. This is supported by seismic tomographic imagery based on non-linear inversion and therefore independent from other reference models (Piromallo and Morelli, 2003; Panza et al., 2007a,b), revealing chemically and rheologically heterogeneous sub-Tyrrhenian upper mantle. Major velocity anomalies are consistent with sizeable partial melt bodies under the Marsili oceanic basin at depths shallower than 100 km, evidence for deeply-rooted plumes having not been observed (Panza et al., 2007b; Chiarabba et al., 2008).

The model of Cadoux et al. (2007) rests on a limited Pb isotopic data set (seven mean values). In their Fig. 3 they plot data from coeval rocks from Sardinia, but they ignore them in their discussion and conclusions. They ignore critical data from highly relevant igneous districts such as Ustica, Pantelleria and Linosa Islands. Their statistical conclusions are based on small, incomplete and highly filtered data subsets and must be judged accordingly. Assuming the (obvious) implication that a hypothetical lower mantle source would leave its geochemical mark on all igneous activity in the region, Fig. 18 plots $^{208}Pb/^{204}Pb_{(i)}$ vs. $^{206}Pb/^{204}Pb_{(i)}$ variation, depicting both the field of original data from Cadoux et al. (2007) along with the database from Lustrino and Wilson (2007), all fitting the definition of intraplate-like igneous rocks as proposed by Wilson and Bianchini (1999) and Lustrino and Wilson (2007). In contrast, the data used by Cadoux et al. (2007), cover a relatively restricted isotopic range for each district, substantially less than the compositional variation recorded in the extant database.

On the basis of their mean values, Cadoux et al. (2007) claim that the bulk of Italian igneous activity is aligned on a trend connecting the Roman and Tuscan Magmatic Provinces, the Aeolian Islands, and the igneous activity in Sicily (Francalanci et al., 1993a,b, 2004a,b). Interestingly, this alignment intersects closely with the Pb isotopic composition of hypothetical 'C' component (Hanan and Graham, 1996) and is clearly distinct from the fields of HIMU-OIBs. If the compositional range represented by the Italian 'anorogenic' database is included, an additional – and highly significant – trend comes to light, connecting the relatively unradiogenic isotopic compositions of Sardinia (the UPV group; Lustrino et al., 2000, 2002a,b, 2007a,b) to the most radiogenic compositions of Pietre Nere and Mount Etna (Fig. 18; references in Lustrino and Wilson, 2007; Conticelli et al., 2007), and partly overlapping with the HIMU-OIB field. The expanded database thus changes the slope of the trend so that it does not point towards 'C', as concluded by Cadoux et al. (2007), but towards typical HIMU-OIB compositions. Even if the oldest (i.e., pre-Pliocene) rocks are discarded, the picture remains the same.

As a concluding observation, the existence of deep mantle upwelling, as proposed by Bell et al. (2006) on the basis of isotopic data, appears to require an 'adjunct' hypothesis to explain the absence of a 'large igneous province' (the classic 'smoking gun' of a mantle plume; see discussion in www.mantleplumes.org). This feature has been rationalized by the fact that any Tyrrhenian Sea 'plume' is still on its way up, and not yet able to produce voluminous basaltic magmas. According to Lavecchia and Creati (2006), active subduction beneath Calabria and northeast Sicily is a remnant of formerly southeast-vergent Alpine subduction, deflected southeastward by a hypothetical wet mantle coming from the core-mantle boundary. The eastward mantle push is the 'wet' plume«impinged beneath the pre-existing southeast-dipping Tethyan slab, that might have caused the progressive bending and overturning of the deeper portion of such a slab»(see Fig. 5.2 of Lavecchia and Creati, 2006). Scalera (2006) offered an alternative proposal, that the subduction beneath Calabria might be interpreted as a deep mantle plume tongue ejected from the transition zone. Somewhat similarly, Locardi and Nicolich (2005) interpreted Calabrian subduction as an effect of hot asthenospheric convection cells, inducing stress and seismic activity at the interface with a contiguous cooler mantle.

13. Concluding remarks

This paper reviews the main post-Eocene magmatic districts of the central-western Mediterranean Sea that host igneous rocks with subduction-related geochemical signature. With the exception of the Alpine Chain, igneous rocks of the CWM are younger than ~38–35 Myr.

Several V-shaped basins, possibly floored by true oceanic crust, occur in the central-western Mediterranean. The opening of these basins is typically associated with widespread, compositionally variable igneous activity of both subduction-related, intraplate-like affinity and also hybrids of these, often occurring in a fairly close proximity. Usually, the intraplate-like – or an anorogenic – igneous activity follows calcalkaline, potassic and ultrapotassic subduction-related activity after a period of a few to several Myrs. With few exceptions (e.g., southeast Spain), the shift in igneous rock geochemistry is abrupt and sharp, implying tapping of markedly different mantle sources by partial melting products. Such compositional shifts are likely to reflect major spatial-temporal changes in upper mantle structure of the type expected when a subducting slab of oceanic lithosphere breaks-off (i.e., via near-horizontal rupturing or 'tearing'), during and following continent-continent, or continent-arc collision, or with concomitant opening of asthenospheric slab windows (i.e., nearly vertical rupturing), allowing for passive upwelling of hot (because deep) sub-lithospheric mantle, uncontaminated by the effects of subduction. Alternatively, delamination/detachment of sub-continental lithosphere megaliths, in response to mantle wedge perturbations resulting from subduction and/or continental collision, may be able to explain abrupt changes in chemical composition of magmas produced by sudden decompression.

In general, the age of the initial opening of the central-western Mediterranean basins and their associated igneous activity decreases from the northwest sectors (Ligurian–Provençal Basin) towards the southwest (Alborán Sea), south (Algerian Basin) and southeast (Tyrrhenian Sea), involving three main directions, perhaps reflecting 'centrifugal' compressional stress fields, compensated subsequentially by post-collision extension.

The compositions of Cenozoic subduction-related CWM magmas range broadly, in order of decreasing abundance, from calcalkaline and high-K calcalkaline to potassic, ultrapotassic and arc-tholeiitic affinity. On the Italian peninsula, however, where subduction would be expected to have involved recycling of pelitic and carbonate-rich sediments, the potassic and ultrapotassic products are by far the most common subduction-related compositions.

In comparison with classic collisional tectonic settings, the central-western Mediterranean can be considered anomalous, for several reasons, in terms of both tectonic and igneous processes, as summarized below:

a) Several relatively small asymmetric basins developed in an overall compressional stress regime. These basins are typically associated with strong arcuate fold and thrust belts, with traces of subducted slabs identified both as 'fossil' or still-ongoing features.

b) In three cases at least, there is evidence for sudden inversions of subduction polarity: during late Eocene–Oligocene along the southern European paleo-continental margin, along the present-day coasts of northern Algeria and along the southernmost edge of the Tyrrhenian Sea.
c) Subduction vergence ranges from south-southeast [e.g., Early Cretaceous–Eocene central-western Alps (northern Italy, Suisse) and Betic Chain (Spain)] to east [e.g., Late Cretaceous–Eocene western Alps (Italy, France) and Corsica (France); Early Miocene–Present Alborán Sea (Spain–Morocco)], northwest [e.g., Late Eocene–Middle Miocene arc in Sardinia–Corsica; Middle Miocene–Present Calabrian Arc (Italy)], N [e.g., Middle Miocene–Present Peloritani (Sicily, Italy, Miocene Eastern Alps), Tellian (Algeria–Tunisia) and Rifian Belts (Morocco–northwestern Algeria)] and southwest [Apennines (peninsular Italy)].
d) A temporal sequence of igneous activity, comprising two sharply contrasting types of igneous products (in terms of mineralogy, major and trace element and isotopic composition, volcanological style, and petrography) is recorded in nearly all districts as, for example: Betics, Rif, Valencia Trough, Sardinia, Provence–Languedoc and Tell. In other cases only subduction-related igneous activity is recorded, as in Late Miocene to present-day peninsular Italy and the Middle Eocene to Late Oligocene western, central and eastern Alps. In other cases, complex igneous associations are recorded within relatively small areas (as in the Pleistocene to present-day Mount Etna region, with igneous activity of intraplate type occurring a few tens of km south of the coeval Aeolian islands—characterized by markedly subduction-related igneous activity; the Late Paleocene to Oligocene activity in the Veneto Area, a few tens of km south of the near-coeval magmatic activity along the Periadriatic Fault System. Eventually, only activity of intraplate-type is evident as, for example the Pliocene to present-day Pantelleria–Linosa Islands in the Sicily Channel, or the long-standing activity recorded on the southeastern border of Sicily—in the Late Triassic to Pleistocene, Hyblean Mountains).
e) Potassic to ultrapotassic magmas that are generally assumed to form within syn- and post-collisional mantle wedge regions (as in peninsular Italy) are generated well away from currently active subduction zones, although the latter were clearly implicated during pre-collisional stages of their genesis. However, K-rich magmas such as those in central Spain (the leucitites of Calatrava), the northeast Spain leucite basanites of Olot–Garrotxa, the analcime/leucite-bearing basanites of central Sardinia and, to the north of our investigated area, the Eifel leucitites in northwestern Germany and leucite nephelinites of the French Massif Central appear to represent a distinct phenomenon.
f) The bulk of activity in peninsular Italy is characterized by potassic to ultrapotassic compositions, relatively scarce in active subduction systems (where calcalkaline to high-K calcalkaline magmas prevail) unless these are progressively involved in collisions with continental plates (Italy, the Sunda-Banda arc of Indonesia, and the Mindoro collision in the Philippines, being good examples).
g) Igneous activity is apparently absent along the Alpine Chain during the lengthy period of southward oceanic lithosphere subduction beneath the Adria plate. The bulk of 'subduction-related' activity in this area is, indeed, syn- to post-collisional.

Comparison within more than 15,000 whole rock analyses for the circum-Mediterranean region made it possible to identify the geochemical similarities and differences of 'anorogenic' (or intraplate-like) and 'orogenic' (or subduction-related, collisional) lithologies: 1) The latter are generally richer in SiO_2 (SiO_2 is mostly >45 wt.%) compared to the intraplate-like lithologies (SiO_2 mostly <50 wt.% for most of the less evolved compositions); 2) except for a few MgO-rich samples such as the lamproites, the bulk of 'subduction-related' compositions have MgO <9 wt.%, in contrast to about half the CiMACI rocks showing MgO >7 wt.% (up to ~20 wt.%); 3) among alkaline compositions, 'subduction-related' magmas show potassic to ultrapotassic compositions with K_2O/Na_2O between ~1 and ~10, irrespective of late-stage differentiation effects. By comparison, CiMACI Province magmas rarely exceed K_2O/Na_2O ratios of 1.5; 4). Among subalkaline compositions, the 'subduction-related' compositions are essentially calcalkaline in character, whereas 'anorogenic' types are tholeiitic; 5) 'subduction-related' rocks show depletions in TiO_2 (<1.5 wt.%, often <1.1 wt.%) for a wide range of MgO contents, compared to those of the CiMACI Province (TiO_2 >1.2 wt.%, up to 5 wt.%); 6) 'subduction-related' rocks are characterized by lower Na_2O (<4.0 wt.%) in samples with MgO >5 wt.%, in contrast to values of CiMACI Province rocks for equivalent MgO contents, characterized by 5–6 wt.% Na_2Ot; 7) other major elements (e.g., Al, Ca, Fe P) are not statistically diagnostic in distinguishing 'subduction-related' from 'anorogenic' magmatic affinity; 8) while LILE and LREE contents vary considerably, both as a function of partial melt fraction and the effects of metasomatic enrichments, LILE/HFSE and LREE/HFSE ratios are significantly higher in 'subduction-related' rather than in 'intraplate-like' magmas (e.g., Ba/Nb >20 and La/Nb >1.4 as compared to <20 and <1.3, respectively); 9) 'subduction-related' rocks often show distinct Ta–Nb troughs in primitive mantle-normalized diagrams compared to 'intraplate-like' compositions, although the Hf and Zr pair may lack 'troughs' and even show relative peaks; 10) as a whole, irrespective of MgO content, CWM 'subduction-related' samples show $^{87}Sr/^{86}Sr_{(i)}$ >0.7035, more than 90% of which having $^{87}Sr/^{86}Sr_{(i)}$ exceeding the average estimate of BSE, 0.70445. In contrast, more than 95% of CiMACI rocks show $^{87}Sr/^{86}Sr_{(i)}$ between 0.7027 and 0.7047; 11) more than 95% of the 'subduction-related' rocks have $^{143}Nd/^{144}Nd_{(i)}$ <0.5128, down to 0.5120. In contrast, more than 95% of CiMACI rocks $^{143}Nd/^{144}Nd_{(i)}$ lie above ChUR estimate (0.51264), between 0.5132 and 0.5126; 12) More than 95% of the subduction-related CWM rocks have $^{206}Pb/^{204}Pb_{(i)}$ ratios between 18.6 and 19.5, near-indistinguishable from those of more of the 95% of CiMACI rocks ($^{206}Pb/^{204}Pb_{(i)} = 18.7$–19.7). Also $^{208}Pb/^{204}Pb_{(i)}$ compositions are indistinguishable from those of both the 'subduction-related' CWM and CiMACI rocks (38.5–39.4 and 38.5–39.7, respectively). The 'subduction-related' compositions seem to have slightly higher thorogenic lead ratios ($^{208}Pb/^{204}Pb$) for a given $^{206}Pb/^{204}Pb$ ratio, more than 90% having $^{207}Pb/^{204}Pb_{(i)}$ >15.65 (up to 15.71), with more than 95% of CiMACI $^{207}Pb/^{204}Pb_{(i)}$ ratios <15.66 (down to ~15.49).

In our view the most promising approach to improve our understanding of the geodynamic situation of a given CWM area and its igneous evolution is by a multi-disciplinary approach integrating state-of-the art igneous rock geochemistry and geochronology, geophysical constraints and other robust geological information.

In most regions the problem of distinguishing coeval subduction signatures from those imposed by metasomatism during earlier subduction processes is hard to constrain (e.g., Peccerillo, 1999; Tamburelli et al., 2000; Kovacs and Szabò, 2008). An integrated approach of geochemical, geophysical and geological tools/parameters/facts is the most promising way to investigate if the origin of igneous rocks with a subduction-related geochemical signature is related to an active subduction zone.

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