Phantom plumes in Europe and the circum-Mediterranean region

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Abstract

Anorogenic magmatism of the circum-Mediterranean area (Tyrrhenian Sea, Sardinia, Sicily Channel and Middle East) and of continental Europe (French Massif Central, Eifel, Bohemian Massif and Pannonian Basin) has been proposed to be related to the presence of one or more mantle plumes. Such conclusions based on geochemical data and seismic tomography are not fully justified because: *1)* a given chemical and isotopic composition of a magma can be explained by different petrogenetic models; *2)* a given petrogenetic process can produce magmas with different chemical and isotopic composition; *3)* tomographic studies do not furnish unique results (i.e., different models give different results); *4)* the commonly adopted interpretation of seismic wave velocity anomalies exclusively in terms of temperature is not unique – velocities are dependent also on other parameters such as composition, melting, anisotropy and anelasticity. Tomography and geochemistry are powerful tools but must be used in an interdisciplinary way, in combination with geodynamics and structural geology. Alone they cannot provide conclusive evidence for or against the existence of mantle plumes.

The existence of large and/or extensive thermal anomalies under Europe is here considered unnecessary, because other models, based on the existence of upper mantle heterogeneity, can explain the major, trace-element, and isotopic variability of the magmas. Volcanism in central Europe (the French Massif Central, Germany and the Bohemian Massif) is concentrated in Cenozoic rifted areas and is here interpreted as the result of passive asthenosphere upwelling driven by decompression. Similarly, anorogenic magmatism in Sardinia, the Tyrrhenian Sea and the Pannonian Basin is explained as the result lithospheric stretching in a back-arc geodynamic setting. The most important factors determining the locus and, in part, the

geochemical characteristics of magmatic activity are the Moho and the lithosphere/asthenosphere boundary depths. Where both are shallowed by tectonic processes (e.g., in rift zones or back-arc basins) passive upwelling of asthenospheric mantle can explain the magmatic activity.

1. Introduction

The most popular model for the Earth's mantle proposed in the last few decades requires a sub-continental upper mantle structure with variably enriched lithospheric mantle and a relatively depleted asthenospheric mantle. On the other hand, the oceanic lithospheric and asthenospheric mantles are considered to be more similar in chemical composition, and to have a generally depleted character. The depleted character of the uppermost oceanic mantle was deduced indirectly from the composition of magmas emplaced along mid-ocean ridges (MORB). Some "anomalous" compositions, mostly recorded on punctiform magmatic manifestations, and known as OIB (Ocean Island Basalts) are considered related not to the upper mantle but rather to deep mantle sources, possibly ascending from the core-mantle boundary or the upper/lower mantle transition zone (e.g., Wilson, 1963; Olson et al., 1987; Courtillot et al., 2003). The difference between MORB and OIB has often been related to different compositions between the two mantle sources: relatively depleted for the first and relatively enriched (or even pristine) for the second. This dichotomy was based on the assumption of a nearly total separation between upper and lower mantle with the presence of two separate systems of mantle convection divided by the 660-km seismic discontinuity (e.g., Karato, 1997; Tackley, 2000).

Since the 1990s global tomographic models (e.g., van Der Hilst et al., 1997; Bijwaard et al., 1998) provided insights in the distribution of seismic wave velocities in the whole mantle. Fast and slow velocity anomalies were interpreted as evidence for cold and hot bodies in the mantle. Although partial evidence for this interpretation is available for the upper mantle, no clear proof is available for the lower mantle. A major inference from the results of global tomography analyses is that a total separation between upper and lower mantle reservoirs (one of the major requirements to invoke the peculiar geochemical signature for OIB magmatism) is no longer a-priori acceptable. In fact, it has been proposed, on the basis of seismic tomography models, that subducting slabs cross the 660 km discontinuity (e.g., Grand et al., 1997; Grand, 2002) and may bend and accumulate within the D'' layer, at the core-mantle boundary (Hutko et al., 2006). Nonetheless, the plume model is still based on such geochemical distinctions.

In order to explain OIB magmatism with the mantle plume model, geochemists propose a contrasting model: from one side, plume-advocates invoke isolated sources (i.e., a closed system that has never contributed to magma production, and consequently considered to be undegassed and with high $^3He/^4He$ ratios). But, at the same time, such a reservoir cannot be considered to be a closed system because it must allow the entrance of subducted oceanic crust where it should be stored for at least 2 Ga. This requirement is necessary to explain the high uranogenic Pb isotopic ratios observed (e.g., $^{206}Pb/^{204}Pb > 21$; e.g., Hofmann and White, 1982).

In this chapter we discuss the main problems arising from an uncritical use of geochemical data and tomography images to propose the existence of hot mantle upwellings from the core-mantle boundary or shallower depths (e.g., the 660 km discontinuity) under "hot spots". An upwelling of hot mantle is commonly called a mantle plume. The difference between the potential temperature (T_p; i.e., the temperature that a volume of the mantle would have if brought to the surface adiabatically without melting; McKenzie and Bickle, 1988) of "normal" asthenosphere (with T_p ~1280 °C) and mantle plume material can be as high as 300 °C (e.g., Richards et al., 1989; Griffiths and Campbell, 1990), even if its ΔT_p can be reduced to as low as 100 °C (e.g., McKenzie and Bicke, 1988; White and McKenzie, 1989). The requirement of such a large ΔT_p (i.e., the existence of extremely hot upwelling volumes of mantle) has been proposed to explain the huge volumes of magma emplaced in continental large igneous provinces such as the Deccan, Paranà-Etendeka and Siberian traps, Kerguelen and Ontong Java etc. (e.g., Saunders et al., 1992; Mahoney and Coffin, 1997). These geochemical and geophysical models are based on the assumption that the source regions of large igneous provinces (volumes of magmas of the order of some million km^3 produced in relatively short time, ~1-2 Ma) are entirely peridotitic. However, during the last decade, new models have suggested the presence of lithologies (eclogites, pyroxenites, garnet granulites etc.) with solidus temperatures several hundred degrees lower than peridotic mantle (e.g., Cordery et al., 1997; Hirschmann, 2000; Yaxley, 2000; Kogiso et al., 2004). At least in some cases, enhanced melt productivity could be the consequence of chemical anomalies (e.g., the presence of low temperature melting point assemblages) rather than thermal anomalies (as proposed in mantle plume models). As an example of questionable plumes, defined on the basis of geochemical and tomography analyses, we discuss some volcanic areas in the European and circum-Mediterranean area.

The main conclusions of this review can be summarized: 1) geochemical data must be used with care because *a)* a given chemical and isotopic composition can be explained by different models, and *b)* a single petrogenetic model can produce magmas with different chemical and isotopic composition; 2) tomographic studies do not furnish unique results and different studies give different results; 3) the presence of deeply rooted mantle plumes or the presence of anomalously hot upper mantle is unlikely and unnecessary to propose in the circum-Mediterranean area; 4) most Cenozoic magmatism in the circum-Mediterranean area can be considered related to passive upwelling of asthenospheric material as a consequence of lithospheric thinning in rift or back-arc areas.

2. Geodynamic setting

European continental structure is the result of several orogenic cycles dating back at least to the Precambrian. Cenozoic evolution of the western Mediterranean area occurred in the framework of convergence between Africa and Europe which led, during the lower Cretaceous-Paleocene, to the consumption of the Tethyan ocean(s) (see Polino et al, 1990, and Schmid et al., 1996, for a discussion on Cretaceous paleogeography) along an E-SE dipping subduction zone (with European- Iberian lithosphere sinking below the African plate) running continuously from the Alps to the Betics via Corsica and Balearics (Fig. 1). There, European-Iberian lithosphere sank below the African plate. The Tertiary-Present diachronous collisional stage was possibly associated with slab detachment events and the presence of continental microplates between Africa and Europe (Doglioni, 1991; Von Blanckenburg and Davies, 1995; Schmid et al, 1996; Carminati et al., 1998).

During the Cenozoic, a rift system (the Rhine, Rhone and Bresse grabens) developed from the North Sea to the Mediterranean (Ziegler 1992) contemporaneously with the development of the Alpine Orogeny (Fig. 2). Since the Oligocene, the onset of west-dipping subduction of the Adriatic plate beneath the European plate drove the development of the Apennine and Maghrebide belts. The fast (up to 5 times faster than Africa-Europe convergence) radial rollback of the subducting plate resulted in the opening of two diachronous back-arc basins, the Lower Miocene-Langhian Algerian-Provencal Basin and the Langhian-Present Tyrrhenian Sea. The rollback of the subduction zones also caused partial disruption of the Alpine-Betic chain due to back-arc extension and the migration and rotation of segments of this chain (e.g., Corsica, Calabria, Kabilies). these are now entrained within the Apennines-Maghrebides belt or located in the western Mediterranean back-arc basin (Fig. 1).

The Balkan, Aegean and Anatolian areas have been characterized, since the Mesozoic, by the development of a poly-phased double vergence orogenic belt (the Dinarides, Hellenides and

Taurides). This orogen is the result of at least two or three subduction zones, as shown by the occurrence of two distinct oceanic sutures (the Vardar-Izmir-Ankara and Sub-Pelagonian ophiolites). These represent one or two (depending on the reconstruction) branches of the Mesozoic Tethyan ocean and the present oceanic subduction of the Ionian and Aegean Sea (e.g., Carminati and Doglioni, 2004; Fig. 1). The E-dipping subduction of the Adriatic lithosphere beneath the Dinarides-Hellenides started as early as the Cretaceous and in the Tertiary became the main process acting in the Eastern Mediterranean area. In the Tertiary, widespread extension overprinted compressional structures in the Dinarides-Hellenides-Taurides orogen. Similarly to the western Mediterranean, boudinage of the pre-existing Alps and Dinarides orogens occurred also in the Pannonian basin, which is a back-arc basin related to the eastward-retreating, westward-dipping Carpathian subduction zone, which has been active from the Oligocene to the Present (Fig. 1).

3. Key parameters of basic magmas

Widespread volcanic activity accompanied the Cenozoic evolution of continental Europe, northern Africa, Mashrek (Middle East) and the Mediterranean Sea. Cenozoic igneous rocks emplaced within this large region show an extremely wide range of chemical compositions which can be grouped into: (a) sodic, mildly alkaline compositions, often associated with tholeiitic rocks with geochemical characteristics resembling magmas emplaced in oceanic intraplate tectonic settings (e.g., composition OIB-like); (b) ocean-floor rocks, whose compositions vary from N (Normal)-MORB to E (Enriched)-MORB and low-K calcalkaline basalts and andesites; (c) calcalkaline rocks with geochemical characteristics resembling magmas emplaced in subduction-related settings; (d) potassic to ultrapotassic alkaline rocks with mildly to strongly SiO_2- undersaturated compositions (e) rare exotic compositions such as lamproites, lamprophyres and carbonatites (Wilson and Downes, 1991; Wilson and Bianchini, 1999; Conticelli and Melluso, 2004; Peccerillo, 2005; Beccaluva et al., 2007; Lustrino and Wilson, 2007). The Sr-Nd-Pb isotopic compositions of these products comprise virtually all the known worldwide reservoirs, the extreme heterogeneous compositions of the mantle sources of this sector of the European-Mediterranean lithosphere/asthenosphere system reflecting its complex geodynamic history (e.g., Lustrino and Wilson, 2007).

This review mainly focuses on the first type of product (i.e., Na-alkaline and tholeiitic magmas) because in many cases these have been interpreted as the result of partial melting of an anomalously hot mantle. The origin of this type of igneous activity (called “anorogenic”) has often been related to the presence of deep mantle plumes that interact with upper mantle in several ways. Many of the petrological studies proposing the existence of a single giant mantle plume or the presence of several smaller plumes are based on tomographic images that depict anomalously low-velocity (interpreted as hot) regions at variable depths.

The first problem is to try to define what an "anorogenic" magma is. This topic has been approached by several authors (e.g., Wilson and Bianchini, 1999; Wilson and Downes, 2006; Lustrino and Wilson, 2007). Proposals are mostly based on inter-elemental incompatible trace element ratios (e.g., Ba/Nb, La/Nb, Th/Yb etc.), major element ratios (e.g., Na_2O/K_2O) and isotopic systematics (mostly $^{87}Sr/^{86}Sr$, $^{143}Nd/^{144}Nd$ and, to lesser extent, $^{207}Pb/^{204}Pb$). However it must be noted that no clear geochemical distinction exists between "anorogenic" and "orogenic" magmas, because of the existence of common transitional or "hybrid" compositions. Following Lustrino and Wilson (2007), the mafic "anorogenic" volcanic rocks considered in this review have the following geochemical characteristics: 1) most of the *anorogenic* rocks are mildly to strongly alkaline (basanites, nephelinites, hawaiites and alkali basalts) plus rarer tholeiitic basalts and basaltic andesites. The SiO_2 content of the most primitive mafic rocks is 40-52 wt.%, and the sum of Na_2O+K_2O is mostly 3-10 wt.%. In general, these rocks are sodic with Na_2O/K_2O in the range 1.4-5. The TiO_2 contents of the most mafic samples (MgO >7 wt%) ranges from ~1 to ~ 6 wt.%; 2) most of the primitive "anorogenic" rocks have primitive mantle-normalized trace-element patterns resembling those of end-member HIMU OIB. The most common feature of the trace-element patterns are peaks at the HFSE (Nb-Ta-Hf-Zr) and negative K and Pb anomalies; 3) the Sr and Nd isotopic compositions are mostly confined to the depleted quadrant, with $^{87}Sr/^{86}Sr$ lower than present-day Bulk Silicate Earth estimates (BSE = 0.70445), and $^{143}Nd/^{144}Nd$ higher than Chondritic Uniform Reservoir values (ChUR = 0.51264). $^{206}Pb/^{204}Pb$ ratios are in general relatively radiogenic (18.8-20.4), whereas $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ are highly variable (Lustrino and Wilson, 2007).

4. Geochemistry-Tomography-Geodynamics: is there an obvious link?

The classification of a rock as "anorogenic" is particularly important in very complex areas such as the circum-Mediterranean area. Indeed if we want to investigate the composition of the upper mantle and its response to tectonics, we must exclude magmas that show clear effects of crustal contamination at shallow levels or that have suffered metasomatic modifications along subduction margins. Crustal contamination of melts in magma chambers at crustal levels is relatively easy to detect on the basis of petrographic observations (e.g., the presence of quartz xenocrysts or crustal/mantle xenoliths) and on the basis of some key geochemical parameters (e.g., compatible and incompatible trace elements and Sr-Nd isotopic variation with MgO, SiO_2 etc.). More complex is the identification of crustal contamination at deeper levels in the mantle.

What is important to stress here is that virtually all the igneous rocks reflect in their chemistry the effects of interaction between mantle (i.e., peridotitic) and recycled crustal (i.e., pyroxenitic/eclogitic) lithologies (e.g., Hofmann and White, 1982; Hirschmann and Stolper, 1996; Kogiso et al., 2003, 2004; Meibom and Anderson, 2003; Lustrino, 2005, 2006; Stracke et al., 2005).

The presence of crustal lithologies at mantle depths obviously influences the composition of the partial melts of a mixed source. Crustal lithologies have a solidus temperature about 50-300 °C lower than peridotitic mantle at similar pressures. This means that, in an adiabatically upwelling mantle, crustal lithologies may melt several tens of km deeper than ambient mantle. No consensus has been so far reached about the effective role of pyroxenitic/eclogitic material in basalt petrogenesis (e.g., Hirschmann et al., 2003; Kogiso et al., 2003; Keshav et al., 2004; Sobolev et al., 2005; Huang and Frey, 2005; Lustrino, 2006).

Neither mixing between peridotite and bulk clinopyroxenite, nor between peridotite and clinopyroxenite partial melts provides a satisfactory explanation for the range of chemical compositions observed in basaltic magma. A complex style of interaction between peridotite and clinopyroxenite partial melts involving assimilation of some phases and stabilization of other minerals is more likely (e.g., Arai and Abe, 1995; Yaxley, 2000; Lustrino, 2005). The effects of such chromatographic styles of metasomatism are hard to model geochemically and are still not understood in detail.

Regarding geophysics, the results of tomographic models are regularly proposed as compelling evidence for the existence of mantle plumes. The basic idea is that low-velocity anomalies in tomographic images reflect relatively high temperature zones. There are two main limitations in the use of tomography to detect plumes. The first and more obvious is that different studies frequently yield different results and lead to different conclusions being drawn. The case of the so-called Iceland plume is emblematic. Bijwaard and Spakman (1999) and Zhao (2001) imaged with global tomography a narrow low-velocity anomaly below Iceland extending down to the core-mantle boundary. Alternative tomographic results provided by Foulger et al. (2000), Ritsema et al. (1999) and Montelli et al. (2004) showed that below Iceland a low-velocity anomaly interpretable as a plume can be resolved clearly only in the upper-mantle. Discrepancies are related to the different analysis approaches and, more importantly, to the different graphical presentations and interpretations of the results. Particularly important is the style of colour used in cross-sections (Anderson, 1999). For example, Bijwaard and Spakman (1999) and Zhao (2001) were able to illustrate a continuous low-velocity anomaly extending down to the lower mantle by saturating their colour scales at the red end at a velocity anomaly of ~ 0.5%. Then, anomalies of ~ 5% in the upper mantle and 0.5% in the lower mantle are all shown as the same colour. On the other hand, if the colour scale is saturated at ~ 2% as, for example, used by Ritsema et al. (1999), the strong upper mantle anomalies appear red but the weak lower mantle anomalies only yellow or orange. Then, there is no visual impression of a continuous structure. In fact, most tomographic models for the mantle beneath Iceland basically agree that there are strong anomalies in the upper mantle and only weak ones in the lower mantle. The weak lower mantle anomalies vary from model to model because they are close to the noise level and not robust enough to be well repeatable between different studies.

Surface waves are well suited for investigating the uppermost mantle on a global scale. Although narrow plume stems may be theoretically too narrow to be imaged by global tomography, large ponds of plume material (of the order of 1,000 km in width) beneath the lithosphere are detectable. Body waves are regularly used to investigate the whole mantle, although teleseismic tomography may be used for detailed analyses of the upper few hundred km of the mantle beneath particular areas. Whole-mantle (or global) tomography uses large global data sets, to derive 3D models of the entire mantle. The main pitfalls of such models derive from errors due to uneven distribution of rays from earthquakes and seismometers throughout the world and from weak resolving power (of the order of a few hundred km). Such a low resolution should encourage caution when interpreting low-velocity bodies in the lower mantle as plumes (Foulger and Natland, 2003).

The second limitation is more subtle. As discussed, Earth scientists commonly interpret low-velocity bodies in tomographic images as hot zones and high-velocity bodies as cold zones. It is, however, often neglected that other factors also affect the velocity of seismic waves. These include pressure, rock composition, melting, anisotropy and anelasticity. Consequently, anomalously-low-velocity bodies in the mantle may not automatically be interpreted as hot material and assumed to indicate the presence of a plume. For example, Anderson (2006) argues that high-density eclogitic material may be characterized by lower V_s compared to peridotitic material at the same depth, thus resulting in “red zones” in tomographic images and giving the impression of the existence of hot mantle. High-velocity bodies are generally interpreted as indication of a subducting slab that is colder than the adjoining lithosphere due to the fact that thermal re-equilibration is slower than slab sinking. Although such an interpretation may not be absolutely certain, the correspondence, worldwide, of high-velocity bodies with sub-crustal earthquake alignments down to the bottom of the transition zone (Wadati-Benioff planes) provides independent supporting evidence for this interpretation. Such independent evidence is, however, missing for the low-velocity bodies beneath presumed hot spots and elsewhere. It may be concluded that, at the moment, there is insufficient evidence for mantle plumes in seismology alone.

Caution is thus needed when interpreting geochemical data and tomographic results in terms of geodynamic features such as hot spots. A further problem with such an interpretation derives from the fact that, in the recent literature, a clear and unique definition of hot spot is not available. In his seminal work Morgan (1971) defined a hot spot as a zone of anomalous intraplate volcanism underlain by a narrow upwelling plume originating from the deep mantle. Such a definition has been progressively modified to account for unexpected observations and presently the hot spot family consists of a wide variety of processes and geometries, generating (in the better cases only semantic) confusion. Concepts like fossil plumes (e.g., Stein and Hofmann, 1992; Rotolo et al., 2006), dying plumes (Davaille and Vatteville, 2005), recycled plume heads (Gasperini et al., 2000), tabular plumes (Hoernle et al., 1995), finger-like plumes (e.g., Granet et al., 1995), baby plumes (Ritter, 2006), channelled plumes (e.g., Oyarzun et al., 1997), plumes with thoroidal chemical compositions (Mahoney et al., 1992), head-free plumes (e.g., Ritter, 2006), relatively cold plumes (Garfunkel, 1989; Hanguita and Hernan, 2000) etc. have been proposed, trying to explain geophysical and geochemical features not compatible with the original definition of plumes. Courtillot et al. (2003) tried to eliminate such problems and divided hot spots into three main categories, depending on whether their origin is at the core-mantle boundary, at the base of the upper mantle or in the lithosphere. It is evident that the original model of Morgan (1971) has been so enlarged that the use of the terms “hot spot” and “mantle plume” may have lost significance since they comprise processes from lithospheric plate tectonics to deep mantle dynamics. Anderson (2005), scoring worldwide proposed mantle plumes, concluded that there is no place on Earth that fulfils all the five characteristics required by Courtillot et al. (2003) to identify a mantle plume (i.e., the presence of a hot spot track, a large igneous province at the start of this track, high buoyancy flux, high $^{3}He/^{4}He$ ratios and low seismic shear-wave velocity at 500 km depth).

In particular, plume-supporting geochemists have attributed the high $^{3}He/^{4}He$ ratios of some OIB to the involvement of an undegassed (i.e., primitive) lower mantle source component.

However $^3He/^4He$ isotope systematics cannot be considered to be clear evidence of derivation from a primitive (lower mantle) source, as shown by several studies (e.g., Anderson, 1998; Gautheron et al., 2005; Meibom et al., 2003, 2005; Parman et al., 2005). The most important criticisms of $^3He/^4He$ systematic use as "proof" of lower mantle derivation are: 1) often MORB show higher ^3He than OIB, an effect opposite to that expected since the sources of MORB are much more depleted (and, therefore, degassed) than the proposed lower-mantle sources of HIMU-OIB; 2) high ^3He could be related not only to undegassed sources but also to recycled oceanic sediments where cosmic dust (rich in ^3He) accumulated; 3) an undegassed reservoir would have high ^3He but also high ^4He (derived from U and Th decay) therefore it is not obvious that an undegassed (i.e., primitive) mantle reservoir should be characterized by high $^3He/^4He$; 4) the bulk partition coefficient for He is not known in detail and recent experimental studies show that it is lower than that of U+Th (the parent isotopes of ^4He). This means that depleted (i.e., degassed) mantle sources evolve with higher He/(U+Th) compared with primitive (i.e., undegassed) mantle, therefore leading to higher $^3He/^4He$, associated with low total helium contents.

5. Mantle plumes imaged in the circum-Mediterranean area

Regarding the circum-Mediterranean area, the presence of a mantle plume has been suggested based on some geochemical characteristics of mafic magmas [e.g., unradiogenic $^{87}Sr/^{86}Sr$, radiogenic $^{143}Nd/^{144}Nd$, radiogenic $^{206}Pb/^{204}Pb$, low Large Ion Lithophile Elements (LILE)/High Field Strength Elements (HFSE) ratios etc.] and on tomographic images. We suggest that this approach may lead to wrong conclusions since comparable geochemical features characterize also magmas derived from passive upwelling of upper mantle and tomographic images can be interpreted in different ways. Below we review a series of case studies of mantle plumes whose existence has been postulated on geochemical and tomographic grounds.

5.1 TYRRHENIAN SEA

A classical example is from the circum-Tyrrhenian area where the presence of a deep mantle plume has been hypothesized mostly on geochemical (e.g., Bell et al., 2004) and geophysical/geological basis (Locardi and Nicolich, 2005). Bell et al. (2004) proposed the existence of a mantle plume on the basis of a pronounced isotopic polarity along the length of

Italy (i.e., northward increase in $^{87}Sr/^{86}Sr$, $\delta^{18}O$ and $\delta^{13}C$ and decrease in $^{143}Nd/^{144}Nd$, $^{206}Pb/^{204}Pb$ and $^{3}He/^{4}He$ isotopic ratios). They interpreted this as mixing between two end-members, one depleted and the other enriched. The enriched end-member (called ITalian Enriched Mantle, ITEM; Bell et al., 2004) is considered to represent a deep mantle reservoir sited at the D" core-mantle boundary layer and isolated from mantle convection. These authors discussed the opening of the Mediterranean region along the SW-ward continuation of the Rhine-Rhone rift system. They concluded that the geochemical signature of the circum-Tyrrhenian lavas [i.e., the presence of hypothetical geochemical end-members like EM-I (Enriched Mantle type I), FOZO (Focal Zone) or ITEM] is the expression of melts originating in the deep mantle.

According to Locardi and Nicolich (2005) an eastward migrating deep-seated thermal plume would be responsible for: 1) opening of several basins and the reduction of lithospheric thickness; 2) fragmentation of the western Mediterranean; 3) subsequent rotations and collision of microplates and 4) production of magmas with an extremely wide range of chemical compositions in a short time interval. Bell et al. (2004) and Locardi and Nicolich (2005) ignored the existence of subduction beneath Italy. In particular, Locardi and Nicolich (2005) interpreted the indisputable seismically active belt in southern Italy as the effect of a convective cell associated with hot asthenosphere inducing stress and seismic activity at the interface with the neighbouring cooler mantle.

In drawing their conclusions, Bell et al. (2004) and Locardi and Nicolich (2005) did not take in consideration some basic features of Cenozoic Italian magmatism, in particular: 1) from Oligocene until Recent times, volcanic activity with a subduction-like geochemical signature has occurred from the NW (Sardinia) through the central Tyrrhenian Sea (Magnaghi and Vavilov seamounts) to the SE (Aeolian Archipelago; e.g., Lustrino, 2000; Savelli, 2002; Peccerillo, 2005); 2) the chemical composition of upper Miocene-Quaternary igneous rocks emplaced along the western (Sardinia) and eastern (Italian peninsula) branch of the Tyrrhenian Sea is completely different (Peccerillo, 2005; Lustrino et al., 2007); 3) the composition of Italian peninsula volcanic rocks (mostly potassic to ultrapotassic) has never been found in oceanic intraplate tectonic settings (i.e., no OIB show potassic to ultrapotassic compositions); 4) the depth of the Tyrrhenian Sea crust is very high compared to the depth of oceanic crust of a similar age (Malinverno, 1981; Spadini et al., 1995), in disagreement with the hypothesis that doming would be induced by a mantle heat anomaly; 5) the calculated T_p of the Tyrrhenian Sea exceeds only by a few tens °C that of a mantle normal conditions (i.e., T_p ~1320 °C vs. 1280

°C, respectively; Zito et al., 2003; Cella et al., 2006) ; 6) numerical modelling suggests that the development of the Apennine belt cannot be explained by the occurrence of a mantle plume but requires tectonic forces such as those in subduction settings (e.g., Carminati et al., 1999); 7) sub-crustal earthquakes indicate the existence of a slab at least below the Northern Apennines and the Calabrian Arc down to a depth of 500 km (Amato and Selvaggi, 1991; Giardini and Velonà, 1991; Selvaggi and Chiarabba, 1995; Carminati et al., 2002); 8) the off-scraping of sediments previously deposited on continental lithosphere and their accretion in the Apennines thin-skinned accretionary wedge suggests the subduction of continental crust for some 170 km in the Northern Apennines (Carminati et al., 2005) and some 280 km in the Southern Apennines (Scrocca et al., 2005). These values may be added to previous oceanic lithosphere subduction to obtain a total subduction of more than 200 km and 500 km in the two areas, whereas below Calabria, some 770 km of oceanic lithosphere were subducted since 23 Ma (Gueguen et al., 1998). These estimates are consistent with the predictions of tomographic models showing high-velocity anomalies beneath Italy with comparable lengths (Piromallo and Morelli, 2003).

This evidence suggest that the opening of the Tyrrhenian sea was the result of a south-eastward radial roll-back of the north-west directed Apennines-Maghrebides subduction zone which started some 30 Ma ago (e.g., Carminati et al., 1998; Gueguen et al., 1998). This is shown in Fig. 1.

5.2 MIDDLE EAST AND THE SICILY CHANNEL

In other cases the presence of “fossil” plume heads has been proposed on the basis of relative constancy of $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ isotopic ratios of Israeli basalts erupted during the Meso-Cenozoic (e.g., Stein and Hofmann, 1992) or submarine volcanism in the Sicily Channel rift (Rotolo et al., 2006). In particular, Stein and Hofmann (1992) proposed the existence of a fossil plume head stagnating beneath the lithospheric mantle at least since the Proterozoic because Mesozoic/Cenozoic Israeli lavas show a relatively homogeneous Sr-Nd isotopic composition. However this model is not able to explain the strong isotopic difference between Miocene-Pliocene Israeli lavas and coeval igneous rocks from neighbouring areas (e.g., Lebanon, Syria and Jordan; Shaw et al., 2003; Krienitz et al., 2006; Lustrino and Sharkov, 2006). The differences in composition between lavas emplaced a few tens of km away from each other can be more easily explained with shallower (lithospheric) mantle sources activated by decompression along the Dead Sea transform fault. Rotolo et al. (2006) proposed the occurrence of a fossil plume head beneath an area at least as wide as Sicily on the basis of a common HIMU/FOZO component in the volcanic rocks of Pantelleria and Linosa islands in the Sicily Channel, Mt. Etna and Hyblean Mts. in Sicily, Ustica island (NW Sicily) and Alicudi (one of the Aeolian Archipelago islands).

The relatively constant HIMU-FOZO-like trace element and Sr-Nd-Pb isotopic character of the Sicily Channel lavas fall within the chemical range of the CMR (Common Mantle Reservoir) as defined by Lustrino and Wilson (2007). The CMR is believed to represent the "average composition" of the upper asthenospheric mantle in the entire circum-Mediterranean area, without any relationship to deep-seated mantle sources (Lustrino and Wilson, 2007).

5.3 SARDINIA

Concerning the magmatism of Sardinia the presence of a mantle plume head recycled back into the mantle has been hypothesized. Gasperini et al. (2000) proposed for the Pleistocene volcanic rocks of northern Sardinia (Logudoro area) an origin from a mantle source represented by recycled oceanic plateaux. Since these authors relate the existence of oceanic plateaux (unusually thick regions of ocean floor) to anomalous production of basaltic magma as a consequence of an anomalously hot mantle, they concluded that these lavas reflect the existence of plume heads recycled back into the mantle. The timing, location and mechanism are unspecified. This model has been strongly criticized on geochemical, petrological and geological grounds by Lustrino et al. (2002, 2004, 2007). In particular, the geochemical peculiarity of igneous rocks of Sardinia (e.g., relatively high SiO_2, low CaO and CaO/Al_2O_3, relatively high Ni, relatively low HFSE, low HREE, high Ba/Nb and La/Nb, slightly high $^{87}Sr/^{86}Sr$, unradiogenic $^{143}Nd/^{144}Nd$ and $^{206}Pb/^{204}Pb$ ratios) is not explainable by anomalously hot mantle but requires the presence of heterogeneous, metasomatized shallow (lithospheric) mantle sources. In particular, this geochemical peculiarity is thought to be related to an orthopyroxene-rich lithospheric mantle source. The origin of this enrichment in orthopyroxene would be a consequence of SiO_2-rich melt derived from delaminated and detached ancient lower-continental crust reacting with mantle peridotite (for more details see Lustrino, 2005 and Lustrino et al., 2007).

5.4 MEDITERRANEAN SEA AND CENTRAL-WESTERN EUROPE

An extreme example of a mantle plume model is the work of Hoernle et al. (1995) who, on the basis of a geochemical review, coupled with a tomographic investigation, proposed the existence of a large tabular-shaped mantle plume beneath the entire circum-Mediterranean area whose stem would be located beneath the Canary archipelago. As concerns the geochemical database, Hoernle et al. (1995) used not only the composition of typical “anorogenic” magmas (such as those from the French Massif Central, the Rhenish Massif, the Rhine and Eger grabens and the Pannonian Basin, Sicily) but also the composition of potassic/ultrapotassic lavas from the Roman Comagmatic Province and the calcalkaline to ultrapotassic lavas from Aeolian lavas. This is despite the fact that those compositions (never recorded among OIB) are classically considered to be generated in subduction-related settings (e.g., Conticelli et al., 2002, 2004; Peccerillo, 2005; Peccerillo and Lustrino, 2005).

Other authors proposed models resembling that of Hoernle et al. (1995). Among these, the papers of Oyarzun et al. (1997) and Macera et al. (2003) should be mentioned. Again, these authors proposed a process of plume channeling with the stem of the hypothetical thermal anomaly centred beneath the Canary archipelago or the Cape Verde islands, moving then north-eastward through Portugal, central-south-eastern Spain, central-southern France, central-northern Germany, up to the Bohemian Massif and eventually ending in north-eastern Italy (Veneto volcanic province). This model is based on two considerations: 1) most of the mafic rocks from these areas show relatively uniform and common geochemical compositions; 2) these compositions resemble HIMU-FOZO end members. Since these end-members have been commonly associated with deep mantle sources, the reasoning: “uniform HIMU-FOZO composition = origin from a common mantle source = origin from a deep mantle plume” is at work. The hypothesis of plume channeling cannot be considered reliable for at least four reasons: 1) there is no age progression of the volcanic rocks from the SE (Cape Verde-Canaries) to the NW (Bohemian Massif and Veneto Province), as it would be expected in a NW-moving channelled plume; 2) tomographic investigations fail to image clearly a low-velocity zone connecting the Canaries to the Bohemian Massif (e.g., Piromallo and Morelli, 2003); 3) the channelling of hot mantle material beneath the lithospheric thermal boundary layer would necessarily imply thermal erosion of the base of the lithosphere, resulting in the development of geochemical heterogeneities. This prediction contrasts with the overall homogeneous compositions of the Cenozoic lavas emplaced between the Canaries andthe Bohemian Massif; 4) the HIMU-FOZO-CMR-like chemical composition of the Cenozoic anorogenic volcanic rocks from this area (and worldwide in general) can be explained also by invoking much shallower mantle sources (e.g., Stracke et al., 2005; Lustrino and Wilson, 2007).

5.5. FRENCH MASSIF CENTRAL

The Paleocene-Quaternary volcanic activity recorded in the French Massif Central, partially associated with the Limagne-Forez graben that developed in the Oligocene, has been often related to the presence of a mantle plume (e.g., Froidevaux et al., 1974; Granet et al., 1995; Sobolev et al., 1997; Wilson and Patterson, 2001). The strongest evidence supporting a mantle plume origin for the circum-Mediterranean anorogenic Cenozoic igneous province magmas is provided by the correlation, in some areas, between the incompatible trace element content and Sr-Nd-Pb isotope characteristics of the most primitive mafic magmas and those of oceanic island basalts (HIMU-OIB). There is also evidence for contemporaneous regional basement uplift, lithospheric thinning and seismic tomography images show finger-like upper mantle low velocity anomalies beneath the volcanic fields.

On the other hand, contrasting interpretations of the evolution of the European Cenozoic rift system (Fig. 2) have emphasised the role of passive rifting in response to the build-up of Pyrenean and Alpine collision-related compressional stresses in magma generation processes (Dèzes et al., 2004). Such stresses, upon reaching a critical value, may have caused rifting in the European foreland. In this scenario, anomalous uplift of the Massif Central area may be interpreted as shoulder uplift related to the Rhone and Bresse rifts. Rifting in the Limagne-Forez graben, may have further contributed to flexural uplift of the adjacent areas. This scenario is consistent with the Neogene-Recent uplift of the Massif Central (Granet et al., 1995). Minor Cretaceous uplift may be simply related to foreland propagation of compressional stresses related to the Alpine and Pyrenean orogeny. Within this rift-related hypothesis, the magmatism may be considered to be the result of passive upwelling of asthenospheric material and consequent partial melting. This interpretation is consistent with the tomographic results that show that the low velocity anomaly is limited to the upper mantle. A rifting related nature for the magmatism may moreover explain the absence of the hot spot traces that should be evident, given the non-negligible relative motion of the European plate with respect to the underlying mantle since the Paleocene (Fekiakova et al., 2006b). The oldest magmatic activity that occurred before the main formation of the Limagne-Forez graben cannot be easily explained by this rift-related hypothesis. However the very small volume of magma produced during this phase is also contrary to what is expected from a mantle plume.

5.6 EIFEL AND NEIGHBORING AREAS

The Rhenish Massif (located at the junction of the Rhine, Leine, and Rhur grabens, close to the Eifel area) suffered a ~300-m domal uplift of the Hercynian basement since the early Miocene, i.e., 20–40 Ma after the onset of Rhine-graben rifting. Uplift was associated with volcanism that shifted to the Eifel volcanic fields west of the Rhine in the Quaternary. The last eruption occurred ~11,000 years ago (Lippolt 1983).

The existence of a mantle plume beneath northern Germany (Rhenish Massif) has been inferred by analogy with the Cenozoic igneous activity of the French Massif Central where tomographic studies have indicated the presence of a finger-like low velocity anomaly extending down to transition zone depths (410-660 km; Granet et al, 1995; Sobolev et al., 1997; Wilson and Patterson, 2001). Additional supporting evidence for an Eifel plume is provided by ~250 m of uplift during the Quaternary. Ritter et al. (2001) provided the first detailed images of mantle structure beneath Eifel based on high-resolution teleseismic tomography. The top of the velocity anomaly is well-constrained at a depth of 50-60 km (Ritter, 2006), corresponding to the base of the lithosphere. Pilidou et al. (2005) obtained high resolution Rayleigh-wave tomography images showing low-velocity anomalies beneath the Eifel volcanoes down to about 400 km. The low velocity anomalies imaged are, however, smaller in magnitude than those observed in other potential hot spot areas (e.g., Azores and Iceland). In their global tomography model, Montelli et al. (2004) imaged a low velocity anomaly only down to 650 km. The upper mantle character of the low-velocity anomaly, if interpreted as hot, may be consistent with a mantle upwelling related to intraplate extensional stresses. This hypothesis is consistent with the correlation argued to exist between the timing of magmatic activity and changes in the regional stress field (Wilson and Bianchini 1999). According to this scenario, the main volcanic phases may be associated with periods of compressional stress relaxation in the foreland of the Alpine orogenic belt.

Ritter (2006) has attributed the magnitude of the low-velocity anomaly to a mantle excess temperature of ~100-200 °C combined with ~1% partial melt. An apparent “hole” in the shear-wave velocity anomaly at ~ 200 km depth could be caused by the onset of partial melting (Ritter, 2006). A shear-wave splitting analysis of the Eifel area (Walker et al., 2005) shows a first-order parabolic pattern in fast polarization azimuth around the hot spot. This feature suggests a lattice preferred orientation of olivine fast axes in the asthenosphere. The preferred orientation was interpreted as a result of the interaction between the slow WSW absolute motion of the Eurasian plate and a mantle upwelling beneath the Eifel volcanic fields. The parabolic pattern is similar to that observed beneath Hawaii. It has been concluded, however, that the anisotropy beneath the Eifel hot spot and surrounding Rhenish Massif is mostly contained in the asthenosphere. Walker et al. (2005) suggest that the Eifel upwelling is sporadic. In our opinion, this is more likely to be the result of varying crustal intraplate stresses that periodically change and facilitate sporadic eruption rather than the result of varying a low excess upwelling temperature.

According to Goes et al. (1999), a mantle plume beneath central Europe can be traced in tomographic images down to lower mantle depths (~2000 km). In any case, the absence of an E-W age-progressive hot spot track with the oldest magmatism confined to the easternmost sectors (Heldburg and Rhon) and the youngest to the west (Eifel; Wedepohl and Baumann, 1999) is inconsistent with the kinematics of the European plate (which moved eastwards in consequence to the opening of the Atlantic Ocean). If we consider that anorogenic volcanic activity in central Europe extends eastward up to the Bohemian Massif (where volcanic rocks of the Eger graben show an age interval of ~77 to ~0.26 Ma) the absence of the hot spot track is even harder to explain in the context of a mantle plume model. Additionally, Fekiacova et al. (2006) estimate that at ~40 Ma the Tertiary Hocheifel volcanic field was ~1000 km SW of its current geographic position, making a geodynamic link with the present-day seismic structure beneath the Eifel impossible.

Several authors have drawn conclusions against the hypothesis of a deep mantle plume or small plumes or diapirs in this area. Bogaard and Wörner (2003) and Jung et al. (2005) favoured passive upwelling of the asthenosphere as the main cause of magma generation. Meyer et al. (2002) proposed as a potential mantle magma source of the Rhön massif the metasomatically overprinted sub-continental lithosphere forced to partially melt by crustal extensional processes. Fekiacova et al. (2006) exclude any genetic relationship between the Eocene (~44-35 Ma) and Quaternary volcanic activity in the Eifel area, and relate the origin of the Tertiary Hocheifel volcanic field to Rhine graben formation. Haase et al. (2004) modelled the petrogenesis of the Westerwald magmas in terms of partial melting of the thermal boundary layer at the base of the lithospheric mantle, as a consequence of adiabatic decompression during lithospheric thinning.

5.7 PANNONIAN BASIN

The Pannonian basin and the Carpathian arc show a geodynamic evolution similar to the Tyrrhenian basin and Apenninic arc (Fig. 1). In both areas a retreating subduction system developed during Cenozoic, roughly coeval with two geochemically, petrographically and volcanologically contrasting types of igneous activity: the older with orogenic geochemical characteristics, and a younger with more or less anorogenic geochemical characteristics (e.g., Seghedi et al., 2005; Harangi et al., 2006; Harangi and Lenkey, 2007). Also the age and shape of the Carpathian belt can be compared to that of the Apennine chain in Italy (e.g., Carminati and Doglioni, 2004; Fig. 1).

The Neogene anorogenic magmatic products of the Pannonian basin have been explained as the result of a mantle plume finger impingement beneath lithosphere (e.g., Embey-Isztin and Dobosi, 1995; Wilson and Patterson, 2001; Seghedi et al., 2004). In particular, despite the relatively small volume of basaltic magma, Embey-Isztin and Dobosi (1995) proposed the involvement of a mantle plume to explain the origin of the Neogene volcanic rocks of the Pannonian Basin, linking the extension of the lithosphere to an active rifting process (i.e., extension of the lithosphere initiated by a mantle plume). Seghedi et al. (2004) proposed that mantle partial melting was triggered by the upwelling of finger-like upper mantle plumes, similar to those proposed beneath the Massif Central (France) and the Eifel district in Germany.

Peak anorogenic volcanism took place at some 6 Ma in the Pannonian basin, about 5 Ma after the end of the main extensional phases (Harangi and Lenkey, 2007 and references therein), although the first anorogenic volcanic rocks were emplaced at ~11 Ma, coeval with the last extensional phases (Harangi, 2001). Moreover, most of the alkaline anorogenic volcanism in the Pannonian basin is offset from the areas with major lithospheric thinning (i.e., they are localized in areas with relatively thick lithosphere). These two observations were taken as evidence against any role of lithospheric extension in the genesis of anorogenic magmatism in the Pannonian basin. Notwithstanding these two observations, Harangi and Lenkey (2007) argue that the presence of a mantle plume beneath this area is highly unlikely. According to these authors, the magmatic activity is related to the presence of a relatively high temperature regime, a consequence of the ~17-11 Ma lithospheric thinning. The time lapse between extension and peak of magmatism is readily explained by the thermal inertia of lithosphere, that remains close to the solidus temperature as consequence of local perturbance caused by mantle flow. On the basis of geophysical, geological and petrological observations (e.g., presence of a high-velocity anomaly at ~500 km depth, the absence of regional doming and the ongoing subsidence in many parts of the Pannonian Basin and the low volume of magma produced), Harangi and Lenkey (2007) excluded the possibility of the existence of a mantle plume beneath the Pannonian basin.

6. Discussion and concluding remarks

Notwithstanding the extremely wide range of composition of the Cenozoic anorogenic products emplaced in the circum-Mediterranean area, some common geochemical key-parameters can be highlighted (Tab. 1; Fig. 3). In particular, Lustrino and Wilson (2007) proposed the existence of a Common Mantle Reservoir (CMR) characterized by a restricted range of major and trace element composition as well as Sr-Nd-Pb isotopic ratios. However, the existence of Cenozoic anorogenic volcanic rocks with relatively common geochemical features (resembling CMR) cannot be interpreted directly as requiring a physically continuous mantle source region. Rather, the presence of homogeneous geochemical composition suggests the existence of common petrogenetic processes. Indeed, a continuous (upper) mantle source able to feed all Cenozoic volcanic activity from Canaries to the west to Turkey to the east and from Libya-Maghrebian Africa to the south to the Bohemian Massif and Germany to the north is hardly imaginable.

If one postulates the existence of different regions in the upper mantle able to produce melts with relatively homogeneous geochemical characteristics (Tab. 1; Fig. 3), then the ultimate origin of magmatism should be addressed. Is it related to the presence of anomalously hot mantle rising from deep (e.g., upper/lower mantle boundary) or very deep (e.g., lower mantle/core boundary) sources? Or is it related to shallow mantle processes, mostly linked with lithospheric extension causing asthenospheric passive upwelling and consequential adiabatic partial melting? We favour the second hypothesis, where magmatism is unrelated to any heat excess of mantle sources but is, rather, controlled by plate kinematics (e.g., Anderson, 2001). The volcanism in central Europe (French Massif Central, Germany and Bohemian Massif) is concentrated in Cenozoic rift areas and can be easily interpreted (as we do) as the result of passive asthenosphere upwelling. Similarly, anorogenic magmatism in Sardinia, Tyrrhenian Sea and Pannonian Basin may be explained as the result of lithospheric stretching in a back-arc geodynamic setting.

Plume-advocates may argue that most of the European Cenozoic Rift System (ECRIS) is amagmatic (i.e., lacking significant volcanic activity; Fig. 2). Consequently, a magmatogenetic process related only to lithospheric thinning and passive upwelling of asthenospheric mantle is unreliable. Nevertheless, we can explain this vast amagmatic zone in the central portion of the ECRIS with a plume-free model. According to McKenzie and Bickle (1988) and White and McKenzie (1989), huge volumes of magmas derived from partial melting of upper mantle without heat excess (e.g., with T_p ~1280 °C) can develop only under thin lithosphere (~70 km thick) in association with large stretching factors (i.e., β >5). Unfortunately, little is known about the origin of igneous activity characterized by much lower volumes of magmas. ECRIS is characterized by β <5 (β >5 is associated with continental rifts evolving to pure oceanic crust), and in this case the melt thickness generated by adiabatic decompression of asthenospheric mantle with normal potential temperature (T_p ~1280 °C) is very limited (White and McKenzie, 1989). Anyway, the volumes of igneous rocks in the volcanic districts considered here are at least two to four orders of magnitude smaller than the volumes of magmas emplaced in continental (e.g., Paraná-Etendeka, Deccan, Siberian traps) and oceanic large igneous provinces (e.g., Kerguelen, Ontong Java). This means that lithospheric extension alone cannot explain the extreme melt productivity of mantle sources in correspondence of large igneous provinces, but it can be considered a potential mechanism able to feed the much less voluminous Cenozoic anorogenic volcanic activity of the circum-Mediterranean area. It is also worth noting that most of the products of this area are sodic alkaline lavas (Lustrino and Wilson, 2007), considered to arise from low degrees of partial melting of mantle sources (generally <5 %), opposite to the bulk of the products of large igneous provinces (tholeiitic basalts and basaltic andesites), considered to be the products of larger degrees of melting (generally ~10%; Jacques and Green, 1980; Falloon et al., 1988, 1999; Schwab and Johnston, 2001).

For this reason, we suggest that the key factor determining partial melting in rift zones and its upwelling to the Earth's surface is lithospheric thickness. From Fig. 4 is clear that igneous activity in ECRIS is concentrated where lithospheric thickness is lower (around 60-80 km) and is virtually absent where it reaches maximum values (up to 140 km). This is evidence for an effective role of lithospheric stretching in promoting adiabatic decompression of asthenospheric mantle where the lithosphere is relatively thin. It could be argued that the shallow lithospheric/asthenospheric boundary below the Massif Central and Eifel could be due to thermal thinning induced by a rising plume rather than to stretching. However, the correspondence between minimum lithospheric and crustal thicknesses (Fig. 4) in these areas excludes such an interpretation. The fact that relatively large lithospheric thicknesses occur below the upper Rhine Graben in central Germany and small thicknesses below the Rhenish Massif and the Massif Central is, in our opinion, related to the interaction between rifting and post-Variscan lithospheric geometry. Central Germany coincided with internal sectors of the Variscan Orogeny (e.g., Saxo-Thuringian Zone) and was likely characterized by post-Variscan lithospheric thicknesses significantly larger than at the northern and southern boundaries of the chain. The superposition of a similar amount of Cenozoic stretching in internal and external sectors of the Variscan chain may explain the resulting differences in the lithospheric thickness. Accordingly, igneous activity in central Europe is mostly confined to the northern (Rhenish Massif) and the southern (French Massif Central) borders of the Variscan belt.

We conclude that in strongly rifted areas (e.g., the Massif Central, Eifel and the Eger graben in the Bohemian Massif), lithospheric stretching can induce passive partial melting of asthenospheric mantle. According to this process, the sub-lithospheric mantle is sucked upward as consequence of rift development. Following the conclusions of Wilson and Patterson (2001) and Lustrino and Wilson (2007), in order to avoid confusion we suggest calling such passive upwellings not "mantle plumes" but "upper mantle diapiric instabilities". According to McKenzie and Bickle (1988) and White and McKenzie (1989), large volumes of lava can be produced in the presence of limited heat excess related to thermal plumes. We propose that the large volume of magmas produced in the French Massif Central and in the Eifel-Rhön-Vogelsberg areas (but still much smaller than large igneous provinces) are instead related to low temperature melting point lithologies in the upper mantle and not to anomalously hot mantle sources. Indeed, the melt productivity of a mantle reservoir can be enhanced, not only by 100-200 °C hotter sources, but also by crustal lithologies with low solidus temperatures (generally subducted and/or delaminated crustal slices that have solidus up to 300 °C lower than peridotitic assemblages at similar pressures; e.g., Hirschmann, 2000).

Petrological and tomographic methods and interpretations normally used to suggest the existence of mantle plumes are not conclusive. Moreover, frequently local or regional geodynamic and tectonic evolution can explain anorogenic magmatism without the involvement of deep-seated plumes. Our final conclusion is that a detailed knowledge of the regional geology, petrology, geophysics and geodynamics and a fully interdisciplinary approach is necessary before proposing large-scale processes on the basis of scant and unconstrained assumptions.

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Figure captions

Fig. 1: Evolution of the Mediterranean and adjacent areas since 45 Ma. (a) At about 45 Ma the Alps were linked to the Betics down to the Gibraltar Strait. The Alps-Betics chain developed on top of a south-eastward dipping subduction zone that consumed an ocean to the northwest of the belt. Along the Dinarides and Hellenides an east-to-northeast-dipping subduction zone was active. Corsica and Sardinia were attached to the Iberian peninsula. (b) At about 30 Ma the Apennines-Maghrebides subduction developed along the Alps-Betics retrobelt, where oceanic or thinned pre-existing continental lithosphere were present to the east. The subduction zone was characterised by a fast, mainly southeastward, rollback. Similarly, the Carpathians started to develop along the Dinarides retrobelt (i.e., the Balkans). The fronts of the Alps-Betics orogen are cross-cut by the Apennines-related subduction backarc extension, which drove the opening of the Liguro-Provençal basin, the Valencia trough and the North Algerian basin. Between 19 and 15 Ma Corsica and Sardinia were separated from Iberia and rotated counterclockwise to their present-day position. (c) At 15 Ma the Liguro-Provençal basin, the Valencia trough and the North Algerian basin were almost completely opened and active extension jumped to the east of Corsica and Sardinia, leading to the opening of the Tyrrhenian back-arc basin. Similarly, the Carpathians migrated eastward due to the rollback of the subduction zone and the Pannonian back-arc basin was generated. The Dinarides subduction

slowed down, due to the presence of the thick Adriatic continental lithosphere to the west, whereas to the south, the Hellenic subduction was very active due to the presence in the footwall plate of Ionian oceanic lithosphere. (d) At present four subduction zones are active in the Mediterranean: the west-directed Apennines-Maghrebides, the west-directed Carpathians, the NE-directed Dinarides-Hellenides-Taurides and the SE-directed Alps. Modified after Carminati and Doglioni (2004).

Fig. 2: Map of the main features of the European Cenozoic rift system in the Alpine and Pyrenean foreland, showing Cenozoic faults (black lines), rift-related sedimentary basins (light grey) and volcanic fields (black). Dashed barbed lines: Alpine deformation front. BG: Bresse Graben, EG: Eger Graben, LG: Limagne Graben, LRG: Lower Rhine Graben, URG: Upper Rhine Graben. Modified after Dezes et al. (2004).

Fig. 3: Primitive mantle-normalized diagrams (normalizing factors after Sun and McDonough, 1989) of several primitive (MgO >8 wt%) anorogenic volcanic districts of the circum-Mediterranean area. References in Lustrino and Wilson (2007). The field of the least differentiated (MgO >5 wt%) volcanic rocks from St. Helena Island (considered to represent typical HIMU-OIB) has been plotted for comparison (references in Lustrino and Sharkov, 2007).

Fig. 4: (a) Moho depth below central-western Europe and the central Mediterranean (modified after Dezes and Ziegler, 2002). (b) Lithospheric thickness below central-western Europe (modified after Babuska and Plomerová, 1992). MC: Massif Central; RM: Rhenish Massif.

Table caption

Tab. 1: Average composition of several anorogenic volcanic districts of the circum-Mediterranean area. Only samples with MgO >8 wt% have been used for statistics. ST. DEV. = Standard Deviation (2σ); *n* = number of samples. References in Lustrino and Wilson (2007).

		SiO_2	TiO_2	Al_2O_3	Fe_2O_3	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	Na_2O/K_2O	$^{87}Sr/^{86}Sr$	εSr	$^{143}Nd/^{144}Nd$	εNd	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$
	AVERAGE	**44.47**	**2.57**	**13.98**	**11.96**	**0.18**	**9.80**	**10.88**	**3.66**	**1.74**	**0.76**	**2.45**	**0.703971**	**-6.8**	**0.512772**	**2.6**	**19.13**	**15.65**	**38.98**
SPAIN	ST. DEV.	2.61	0.38	2.03	1.07	0.02	2.27	1.71	0.68	0.56	0.32	1.33	0.000395	5.6	0.000060	1.2	0.11	0.03	0.18
	n	*74*	*74*	*74*	*74*	*74*	*74*	*74*	*74*	*74*	*74*	*74*	*11*	*11*	*5*	*5*	*2*	*2*	*2*
	AVERAGE	**41.87**	**3.08**	**12.26**	**12.58**	**0.20**	**11.47**	**12.61**	**3.22**	**1.7**	**1.0**	**2.50**	**0.703586**	**-12.3**	**0.512852**	**4.1**	**19.69**	**15.63**	**39.51**
MAGHREBIAN AFRICA	ST. DEV.	3.05	0.61	2.52	1.60	0.05	3.52	1.92	1.03	0.9	0.5	1.53	0.000444	6.3	0.000097	1.9	0.69	0.04	0.51
	n	*117*	*117*	*117*	*117*	*117*	*117*	*117*	*117*	*117*	*114*	*117*	*42*	*42*	*39*	*39*	*17*	*17*	*17*
	AVERAGE	**44.84**	**2.87**	**13.61**	**12.28**	**0.19**	**9.84**	**10.52**	**3.40**	**1.7**	**0.8**	**2.25**	**0.703573**	**-12.5**	**0.512887**	**4.9**	**19.43**	**15.62**	**39.21**
FRANCE	ST. DEV.	2.30	0.36	1.18	0.70	0.02	1.94	1.25	0.60	0.5	0.2	0.89	0.000285	4.0	0.000054	0.9	0.27	0.02	0.21
	n	*259*	*259*	*259*	*259*	*259*	*259*	*259*	*259*	*259*	*259*	*259*	*96*	*96*	*77*	*77*	*29*	*29*	*29*
	AVERAGE	**50.20**	**2.06**	**15.60**	**10.07**	**0.14**	**8.14**	**7.77**	**3.73**	**1.8**	**0.5**	**2.39**	**0.704463**	**0.2**	**0.512515**	**-2.4**	**17.89**	**15.59**	**37.98**
ITALY (SARDINIA UPV)	ST. DEV.	1.79	0.37	0.82	0.58	0.03	0.86	0.80	0.71	0.5	0.2	1.33	0.000116	1.6	0.000059	1.2	0.11	0.01	0.15
	n	*109*	*109*	*109*	*109*	*109*	*109*	*109*	*109*	*109*	*109*	*109*	*12*	*12*	*7*	*7*	*4*	*4*	*4*
ITALY (SARDINIA RPV)	1 sample	**45.78**	**3.13**	**15.21**	**11.69**	**0.16**	**7.55**	**10.37**	**3.49**	**2.23**	**0.39**	**1.56**	**0.704010**	**-6.2**	**0.512850**	**4.1**	**19.23**	**15.64**	**39.10**
	AVERAGE	**47.01**	**2.19**	**16.00**	**11.24**	**0.17**	**8.97**	**9.32**	**3.42**	**1.11**	**0.58**	**3.49**	**0.703302**	**-16.3**	**0.513002**	**7.1**	**19.47**	**15.65**	**39.12**
ITALY (USTICA AND SICILY CHANNEL)	ST. DEV.	1.48	0.56	1.36	1.06	0.03	2.06	1.07	0.44	0.36	0.23	1.34	0.000415	5.9	0.000052	1.0	0.17	0.03	0.18
	n	*49*	*49*	*49*	*49*	*49*	*49*	*49*	*49*	*49*	*49*	*49*	*16*	*16*	*10*	*10*	*9*	*9*	*9*
	AVERAGE	**48.25**	**1.51**	**16.26**	**10.90**	**0.16**	**8.33**	**10.12**	**3.19**	**0.9**	**0.4**	**4.74**	**0.703444**	**-14.2**	**0.512875**	**4.6**	**19.83**	**15.64**	**39.46**
ITALY (MT. ETNA)	ST. DEV.	1.32	0.13	1.84	0.96	0.02	1.50	1.02	0.44	0.4	0.1	2.67	0.000161	2.3	0.000049	1.0	0.17	0.01	0.14
	n	*20*	*20*	*20*	*20*	*20*	*20*	*20*	*20*	*20*	*20*	*20*	*98*	*98*	*38*	*38*	*8*	*8*	*8*
	AVERAGE	**47.17**	**2.15**	**15.38**	**11.28**	**0.17**	**8.65**	**10.03**	**3.58**	**0.87**	**0.73**	**7.67**	**0.703068**	**-19.6**	**0.513021**	**7.4**	**19.70**	**15.64**	**39.25**
ITALY (HYBLEAN MTS.)	ST. DEV.	3.06	0.51	0.91	1.94	0.02	1.37	1.34	1.27	0.68	0.39	6.57	0.000194	2.8	0.000069	1.4	0.17	0.03	0.21
	n	*77*	*77*	*77*	*77*	*77*	*77*	*77*	*77*	*77*	*77*	*77*	*40*	*40*	*29*	*29*	*24*	*24*	*24*
	AVERAGE	**46.03**	**2.54**	**13.85**	**12.17**	**0.16**	**9.94**	**10.16**	**2.93**	**1.34**	**0.88**	**2.47**	**0.703341**	**-15.7**	**0.512912**	**5.3**	**19.30**	**15.63**	**39.09**
ITALY (VENETO PROVINCE)	ST. DEV.	3.00	0.44	1.08	0.96	0.03	1.93	1.63	0.73	0.48	0.32	1.11	0.000167	2.4	0.000036	0.7	0.23	0.02	0.18
	n	*128*	*128*	*128*	*128*	*128*	*128*	*128*	*128*	*128*	*128*	*128*	*25*	*25*	*25*	*25*	*18*	*18*	*18*
	AVERAGE	**42.63**	**2.67**	**13.28**	**11.22**	**0.20**	**9.81**	**12.74**	**3.32**	**3.03**	**0.65**	**1.23**	**0.704339**	**-1.6**	**0.512744**	**2.0**	**19.17**	**15.63**	**39.41**
GERMANY (EIFEL)	ST. DEV.	1.68	0.23	1.19	0.52	0.01	1.72	1.47	0.56	0.77	0.19	0.66	0.000353	5.0	0.000080	1.6	0.26	0.01	0.24
	n	*64*	*64*	*64*	*64*	*64*	*64*	*64*	*64*	*64*	*64*	*64*	*26*	*26*	*19*	*19*	*6*	*6*	*6*
	AVERAGE	**43.30**	**2.94**	**12.92**	**12.61**	**0.18**	**10.92**	**12.10**	**3.01**	**1.25**	**0.75**	**2.82**	**0.703557**	**-12.7**	**0.512825**	**3.6**	**19.40**	**15.60**	**39.15**
GERMANY (RHON)	ST. DEV.	1.56	0.46	0.89	0.91	0.02	1.23	0.72	0.62	0.53	0.22	1.20	0.000250	3.6	0.000021	0.4	0.09	0.02	0.09
	n	*8*	*8*	*8*	*8*	*8*	*8*	*8*	*8*	*8*	*8*	*8*	*45*	*45*	*36*	*36*	*6*	*6*	*6*
	AVERAGE	**45.08**	**2.56**	**12.88**	**12.15**	**0.18**	**11.61**	**10.80**	**2.96**	**1.14**	**0.65**	**3.13**	**0.703482**	**-13.7**	**0.512797**	**3.1**	**19.21**	**15.62**	**38.97**
GERMANY (VOGELSBERG)	ST. DEV.	3.47	0.38	0.95	0.79	0.02	2.19	1.43	0.39	0.46	0.18	1.55	0.000269	3.8	0.000081	1.6	0.19	0.03	0.25
	n	*43*	*43*	*43*	*43*	*43*	*43*	*43*	*43*	*43*	*43*	*43*	*35*	*35*	*32*	*32*	*10*	*10*	*10*
	AVERAGE	**44.64**	**2.63**	**12.92**	**12.36**	**0.19**	**11.30**	**10.86**	**3.05**	**1.39**	**0.67**	**2.40**	**0.703590**	**-12.2**	**0.512832**	**3.7**	**19.42**	**15.60**	**39.15**
GERMANY (WESTERWALD)	ST. DEV.	1.73	0.38	0.75	0.71	0.01	1.65	0.93	0.64	0.41	0.25	1.00	0.000226	3.2	0.000025	0.5	0.20	0.02	0.28
	n	*55*	*55*	*55*	*55*	*55*	*55*	*55*	*55*	*55*	*55*	*55*	*17*	*17*	*13*	*13*	*9*	*9*	*9*
	AVERAGE	**41.86**	**3.24**	**11.77**	**13.19**	**0.20**	**11.12**	**13.62**	**2.98**	**1.19**	**0.82**	**2.92**	**0.703396**	**-15.0**	**0.512863**	**4.3**	**19.66**	**15.62**	**39.32**
BOHEMIAN MASSIF	ST. DEV.	2.49	0.87	1.67	1.34	0.03	2.97	2.05	0.69	0.47	0.27	1.32	0.000155	2.2	0.000053	1.0	0.14	0.10	0.18
	n	*254*	*254*	*254*	*254*	*254*	*254*	*254*	*254*	*254*	*254*	*254*	*79*	*79*	*75*	*75*	*18*	*18*	*18*
	AVERAGE	**46.60**	**2.25**	**15.42**	**10.46**	**0.16**	**9.30**	**9.32**	**3.88**	**1.92**	**0.68**	**2.25**	**0.703748**	**-10.0**	**0.512809**	**3.3**	**19.03**	**15.62**	**38.91**
PANNONIAN BASIN	ST. DEV.	1.72	0.24	1.00	0.95	0.02	1.54	0.90	0.68	0.55	0.18	0.95	0.000648	9.2	0.000101	2.0	0.24	0.02	0.16
	n	*37*	*37*	*37*	*37*	*37*	*37*	*37*	*37*	*37*	*37*	*37*	*19*	*19*	*18*	*18*	*15*	*15*	*15*
	AVERAGE	**46.22**	**2.52**	**14.48**	**12.21**	**0.15**	**8.90**	**9.93**	**3.45**	**1.50**	**0.63**	**2.55**	**0.703515**	**-13.3**	**0.512914**	**5.3**	**19.25**	**15.66**	**39.16**
TURKEY	ST. DEV.	2.00	0.48	1.14	1.61	0.04	1.44	0.91	0.76	0.54	0.34	0.93	0.000384	5.4	0.000069	1.3	0.06	0.02	0.05
	n	*99*	*99*	*99*	*99*	*99*	*99*	*99*	*99*	*99*	*99*	*99*	*21*	*21*	*17*	*17*	*3*	*3*	*3*
	AVERAGE	**45.70**	**2.47**	**14.57**	**12.83**	**0.17**	**8.84**	**9.68**	**3.68**	**1.26**	**0.80**	**3.23**	**0.703299**	**-16.3**	**0.512884**	**4.8**	**19.10**	**15.61**	**38.87**
MASHREK	ST. DEV.	1.81	0.46	1.04	1.05	0.03	1.18	1.19	0.67	0.62	0.48	0.91	0.000230	3.3	0.000048	0.9	0.14	0.03	0.16
	n	*235*	*231*	*231*	*231*	*231*	*231*	*231*	*231*	*231*	*231*	*231*	*56*	*56*	*56*	*56*	*38*	*38*	*38*

Table 1 (Lustrino and Carminati P4, 2007)

		Rb	Sr	Ba	Sc	V	Cr	Co	Ni	Y	Zr	Nb	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu	Th	Pb	U	Hf	Ta	Cs
	AVERAGE	48	937	801	22	233	298	58	185	28	222	78	61.72	118.44	15.80	53.13	12.11	3.24	10.39	1.04	6.96	1.13	3.36	0.33	2.10	0.28	6.23	6.00	1.45	9.89	4.84	2.00
SPAIN	ST. DEV.	42	295	194	3	29	158	35	95	5	75	22	23.48	40.00	-	23.49	3.85	1.04	2.27	0.72	1.42	-	0.63	-	0.47	0.03	1.98	-	0.65	19.07	2.69	-
	n	*69*	*69*	*69*	*4*	*54*	*67*	*64*	*69*	*68*	*68*	*40*	*69*	*70*	*1*	*43*	*23*	*37*	*20*	*16*	*20*	*1*	*2*	*1*	*36*	*22*	*23*	*1*	*4*	*18*	*18*	*1*
	AVERAGE	44	1333	984	24	253	373	52	191	29	276	109	65.59	128.68	14.90	54.83	9.19	3.44	7.28	1.10	4.93	0.90	2.29	0.28	1.70	0.27	7.06	2.79	2.15	5.91	6.17	0.55
MAGHREBIAN AFRICA	ST. DEV.	25	459	343	6	44	220	14	100	6	83	35	26.47	48.26	5.15	19.06	3.14	3.70	2.49	0.30	1.07	0.18	0.41	0.05	0.37	0.21	2.91	1.04	0.86	1.56	2.81	0.33
	n	*91*	*94*	*90*	*23*	*82*	*87*	*30*	*90*	*82*	*84*	*83*	*39*	*36*	*10*	*39*	*31*	*38*	*14*	*21*	*27*	*10*	*26*	*10*	*38*	*22*	*28*	*6*	*12*	*20*	*8*	*7*
	AVERAGE	54	892	666	20	208	269	48	192	28	287	93	59.79	115.19	13.79	50.85	9.48	2.95	8.58	1.13	6.01	1.10	2.72	0.39	2.16	0.32	10.58	4.11	2.14	6.27	5.26	0.77
FRANCE	ST. DEV.	23	240	205	5	55	140	5	92	4	67	23	19.45	36.37	3.43	13.98	2.09	0.58	1.70	0.19	0.90	0.16	0.40	0.06	0.32	0.05	3.48	1.04	0.55	1.48	1.85	0.23
	n	*230*	*240*	*235*	*186*	*180*	*221*	*187*	*223*	*205*	*212*	*212*	*135*	*135*	*80*	*130*	*136*	*136*	*99*	*82*	*109*	*81*	*98*	*56*	*136*	*118*	*221*	*71*	*33*	*55*	*41*	*14*
	AVERAGE	38	794	959	17	181	297	46	210	21	203	42	40.52	78.79	9.10	34.25	6.68	2.11	6.11	0.79	4.13	0.76	1.80	0.22	1.34	0.18	5.11	4.60	0.99	5.21	2.49	0.56
ITALY (SARDINIA UPV)	ST. DEV.	16	184	366	3	21	77	6	54	3	52	14	13.17	27.97	3.36	8.27	1.16	0.31	1.67	0.15	0.54	0.09	0.23	0.04	0.17	0.03	1.88	1.13	0.53	1.27	1.14	0.32
	n	*93*	*93*	*81*	*34*	*80*	*81*	*61*	*81*	*90*	*80*	*92*	*81*	*81*	*9*	*75*	*10*	*11*	*10*	*7*	*10*	*9*	*10*	*4*	*10*	*9*	*11*	*11*	*5*	*9*	*6*	*2*
ITALY (SARDINIA RPV)	1 sample	49	970	528	20	254	212	44	126	28	223	70	47.10	96.40	12.10	46.60	8.27	2.53	7.02	1.01	5.66	0.88	2.26	0.35	1.97	0.30	5.90	4.00	1.30	4.80	4.51	1.6
	AVERAGE	22	609	319	27	216	300	47	182	25	200	45	35.24	63.24	9.23	32.68	6.33	2.20	7.13	0.93	5.20	0.90	2.33	0.51	1.98	0.32	4.50	1.00	1.24	4.23	3.02	2.10
ITALY (USTICA AND SICILY CHANNEL)	ST. DEV.	8	195	125	3	37	90	14	83	4	59	16	13.01	21.67	4.60	13.81	2.68	0.82	2.63	0.36	1.58	0.24	0.54	0.22	0.36	0.07	1.89	0.00	0.49	1.98	1.96	1.68
	n	*48*	*48*	*33*	*26*	*30*	*33*	*32*	*30*	*44*	*43*	*33*	*36*	*33*	*4*	*8*	*9*	*9*	*4*	*5*	*4*	*4*	*4*	*8*	*9*	*9*	*11*	*3*	*5*	*9*	*9*	*4*
	AVERAGE	36	1039	574	27	250	97	41	55	22	195	36	52.11	101.13	12.70	40.23	8.48	2.66	6.60	0.88	4.80	0.83	2.11	0.31	2.15	0.26	7.32	-	2.15	4.25	2.61	0.74
ITALY (MT. ETNA)	ST. DEV.	16	285	171	3	60	116	6	68	1	43	15	13.62	32.07	-	16.20	1.01	0.48	-	0.10	-	-	-	-	0.23	0.05	2.13	-	0.60	0.79	0.59	0.24
	n	*88*	*88*	*65*	*61*	*7*	*62*	*62*	*63*	*6*	*63*	*5*	*65*	*83*	*1*	*26*	*52*	*62*	*1*	*62*	*1*	*1*	*1*	*1*	*51*	*6*	*61*	-	*61*	*62*	*62*	*59*
	AVERAGE	16	757	423	29	214	309	47	218	30	198	58	47.85	96.19	12.32	46.19	8.81	3.06	8.07	2.13	6.11	1.51	2.81	0.82	2.26	0.35	5.27	3.21	1.29	4.58	3.33	0.27
ITALY (HYBLEAN MTS.)	ST. DEV.	11	434	249	5	38	56	3	28	5	83	35	29.79	60.63	7.87	27.73	3.42	1.25	2.07	1.94	1.55	0.68	0.58	0.75	0.52	0.05	3.20	2.03	0.82	1.73	1.86	0.15
	n	*84*	*84*	*83*	*15*	*73*	*73*	*73*	*73*	*73*	*73*	*73*	*73*	*78*	*20*	*30*	*43*	*43*	*18*	*37*	*26*	*27*	*26*	*27*	*36*	*32*	*73*	*22*	*30*	*32*	*30*	*25*
	AVERAGE	30	829	486	-	217	290	46	190	25	238	54	43.01	89.36	10.59	37.57	8.39	2.73	7.91	1.04	5.49	0.94	2.31	0.31	1.72	0.26	5.61	3.11	1.37	5.68	3.95	0.49
ITALY (VENETO PROVINCE)	ST. DEV.	12	245	200	-	35	118	6	61	4	88	21	14.59	30.85	3.26	13.84	2.26	0.73	2.29	0.22	0.98	0.13	0.38	0.05	0.23	0.05	1.64	0.57	0.48	1.14	0.81	0.12
	n	*117*	*117*	*114*	-	*40*	*116*	*40*	*116*	*117*	*116*	*116*	*114*	*115*	*42*	*108*	*54*	*64*	*45*	*33*	*54*	*42*	*55*	*33*	*54*	*54*	*49*	*5*	*40*	*9*	*5*	*5*
	AVERAGE	70	991	1102	28	319	369	48	189	28	239	100	83.10	153.33	-	64.36	9.84	2.72	7.49	1.00	4.71	0.79	2.05	-	1.81	0.35	8.95	3.00	2.36	5.59	5.40	1.60
GERMANY (EIFEL)	ST. DEV.	22	193	197	5	14	153	14	86	3	30	22	33.90	37.25	-	10.57	2.72	0.26	-	0.10	-	-	-	-	0.14	0.07	3.25	-	0.85	0.38	0.80	-
	n	*34*	*34*	*14*	*2*	*2*	*14*	*14*	*13*	*12*	*13*	*12*	*2*	*3*	-	*17*	*17*	*3*	*1*	*2*	*1*	*1*	*1*	-	*3*	*2*	*2*	*1*	*2*	*2*	*2*	*1*
	AVERAGE	52	893	733	31	298	365	56	211	27	263	84	60.03	118.63	-	52.68	10.53	3.08	8.03	-	5.64	-	2.19	-	1.76	0.20	-	3.88	-	-	-	-
GERMANY (RHON)	ST. DEV.	18	121	119	4	60	169	3	102	5	62	17	15.56	29.44	-	13.78	2.82	0.87	1.11	-	0.64	-	0.58	-	0.21	0.01	-	1.05	-	-	-	-
	n	*14*	*14*	*8*	*4*	*8*	*8*	*4*	*8*	*8*	*8*	*8*	*8*	*8*	-	*8*	*8*	*8*	*8*	-	*8*	-	*8*	-	*8*	*4*	-	*8*	-	-	-	-
	AVERAGE	66	841	683	24	234	425	52	271	24	207	63	47.17	96.87	10.19	40.56	8.68	2.52	6.47	0.95	4.83	0.86	2.22	0.29	1.71	0.24	5.56	2.89	1.33	4.84	4.37	0.65
GERMANY (VOGELSBERG)	ST. DEV.	54	224	185	4	38	114	5	78	4	47	18	14.71	29.24	3.52	10.33	1.51	0.33	0.92	0.12	0.44	0.07	0.26	0.03	0.28	0.05	1.66	0.59	0.42	1.08	1.11	0.25
	n	*39*	*39*	*36*	*36*	*36*	*36*	*36*	*36*	*36*	*36*	*36*	*35*	*35*	*23*	*35*	*33*	*35*	*35*	*23*	*35*	*23*	*35*	*23*	*35*	*35*	*22*	*22*	*22*	*18*	*18*	*23*
	AVERAGE	45	789	599	23	241	320	50	209	24	260	79	50.52	101.58	12.71	47.57	9.02	2.72	7.78	1.04	5.40	0.95	2.33	0.30	1.80	0.25	5.91	3.41	1.75	6.15	4.61	0.55
GERMANY (WESTERWALD)	ST. DEV.	20	167	169	3	36	126	6	77	3	58	24	15.24	28.32	3.24	11.09	1.65	0.45	1.20	0.14	0.65	0.11	0.27	0.03	0.19	0.03	1.90	0.84	0.54	1.34	1.32	0.32
	n	*18*	*18*	*14*	*14*	*14*	*14*	*14*	*14*	*14*	*14*	*14*	*14*	*14*	*13*	*14*	*14*	*14*	*13*	*14*	*13*	*13*	*13*	*13*	*14*	*14*	*14*	*14*	*14*	*14*	*14*	*14*
	AVERAGE	42	998	799	26	245	328	51	205	26	267	102	77.25	141.63	14.36	61.91	10.77	3.17	9.16	1.34	6.43	1.00	2.39	0.83	2.14	0.61	10.50	3.08	1.17	8.24	6.02	1.35
BOHEMIAN MASSIF	ST. DEV.	21	328	260	7	78	234	9	125	6	106	35	30.12	50.95	3.49	19.19	2.99	0.85	2.07	0.60	1.62	0.23	0.36	0.33	0.85	2.78	3.83	2.13	1.08	3.48	2.18	1.48
	n	*173*	*173*	*168*	*124*	*108*	*152*	*120*	*158*	*148*	*151*	*148*	*87*	*89*	*39*	*94*	*95*	*78*	*71*	*44*	*52*	*46*	*45*	*4*	*89*	*78*	*125*	*63*	*115*	*54*	*54*	*53*
	AVERAGE	58	927	848	21	209	242	41	180	23	237	75	57.40	104.17	9.70	42.31	7.76	2.39	6.01	0.83	4.63	0.86	2.18	-	2.14	0.35	8.31	3.68	1.07	5.50	5.37	-
PANNONIAN BASIN	ST. DEV.	18	256	268	5	21	87	6	75	4	82	26	22.61	35.72	-	14.69	2.59	0.60	0.03	0.21	0.28	-	0.10	-	0.40	0.10	3.30	1.86	0.49	1.59	2.29	-
	n	*33*	*31*	*34*	*14*	*30*	*34*	*20*	*29*	*33*	*34*	*30*	*34*	*30*	*1*	*19*	*17*	*17*	*2*	*9*	*2*	*1*	*2*	-	*16*	*12*	*21*	*8*	*4*	*13*	*8*	-
	AVERAGE	25	826	426	23	208	288	53	163	24	209	60	33.22	67.67	8.92	34.09	7.30	2.34	6.70	1.00	4.85	0.91	2.08	0.34	1.62	0.23	4.63	2.22	1.38	5.14	3.36	0.78
TURKEY	ST. DEV.	15	245	220	3	24	150	13	72	4	53	25	9.61	17.62	2.13	8.16	1.41	0.45	1.03	0.56	0.87	0.21	0.36	0.28	0.31	0.03	1.37	1.51	0.55	0.86	1.05	0.39
	n	*76*	*87*	*85*	*56*	*83*	*85*	*55*	*84*	*84*	*85*	*85*	*50*	*50*	*30*	*51*	*51*	*50*	*47*	*45*	*47*	*45*	*47*	*29*	*50*	*49*	*48*	*46*	*45*	*45*	*45*	*21*
	AVERAGE	17	953	427	20	192	253	52	196	27	199	47	41.19	82.72	9.71	35.91	8.21	2.64	6.67	0.90	4.37	0.78	1.84	0.24	1.46	0.20	3.22	2.01	0.97	4.19	2.52	0.22
MASHREK	ST. DEV.	9	435	272	3	21	74	7	64	27	66	26	27.20	44.59	3.54	16.98	2.47	0.73	1.87	0.25	0.68	0.15	0.23	0.04	0.21	0.03	1.79	1.25	0.49	1.03	1.11	0.10
	n	*146*	*155*	*160*	*89*	*130*	*169*	*95*	*170*	*157*	*151*	*121*	*133*	*117*	*32*	*87*	*49*	*89*	*44*	*57*	*37*	*35*	*36*	*31*	*49*	*45*	*101*	*72*	*58*	*54*	*54*	*23*

Table 1 (Lustrino and Carminati P4, 2007)

A)

B)

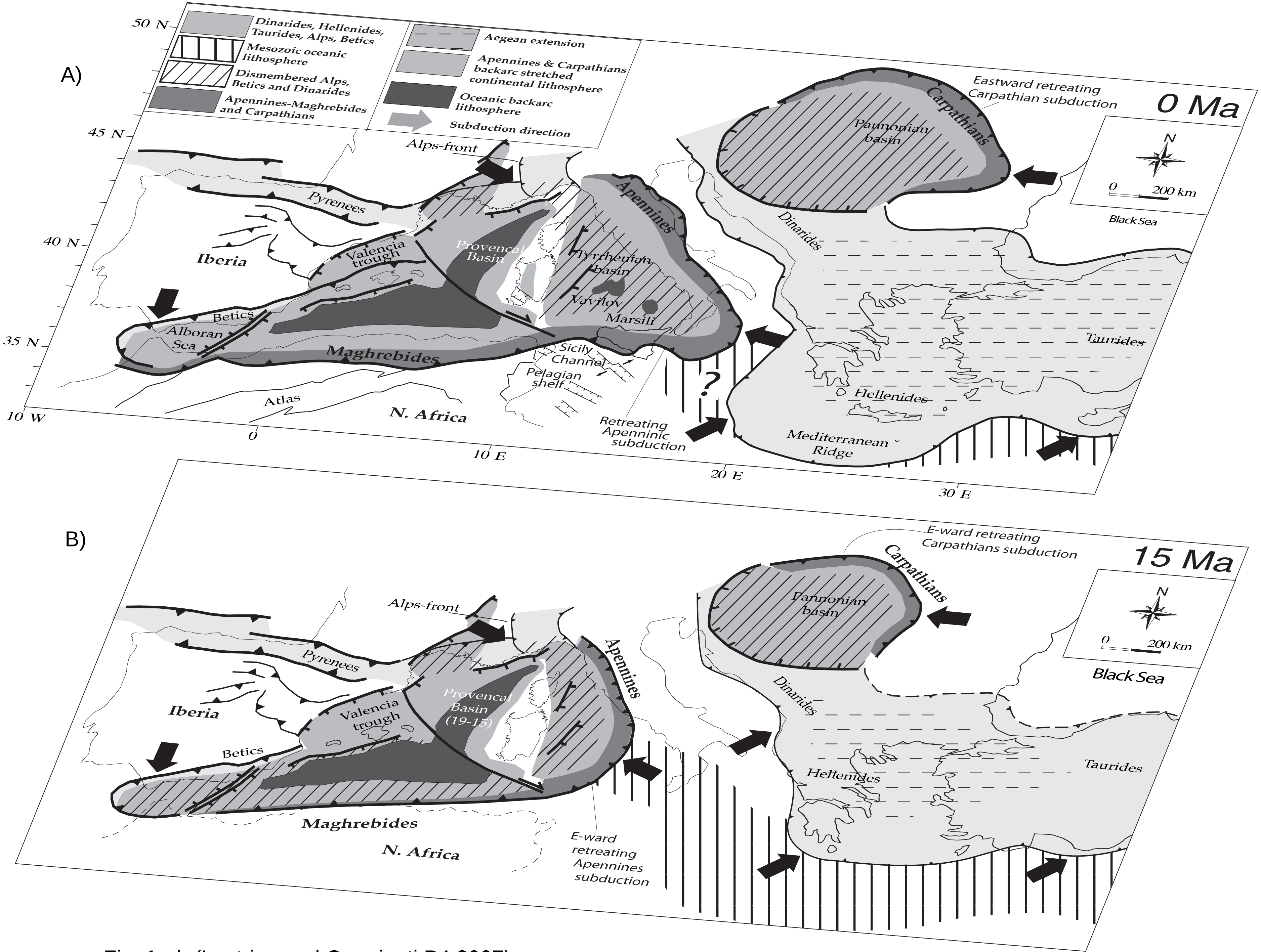
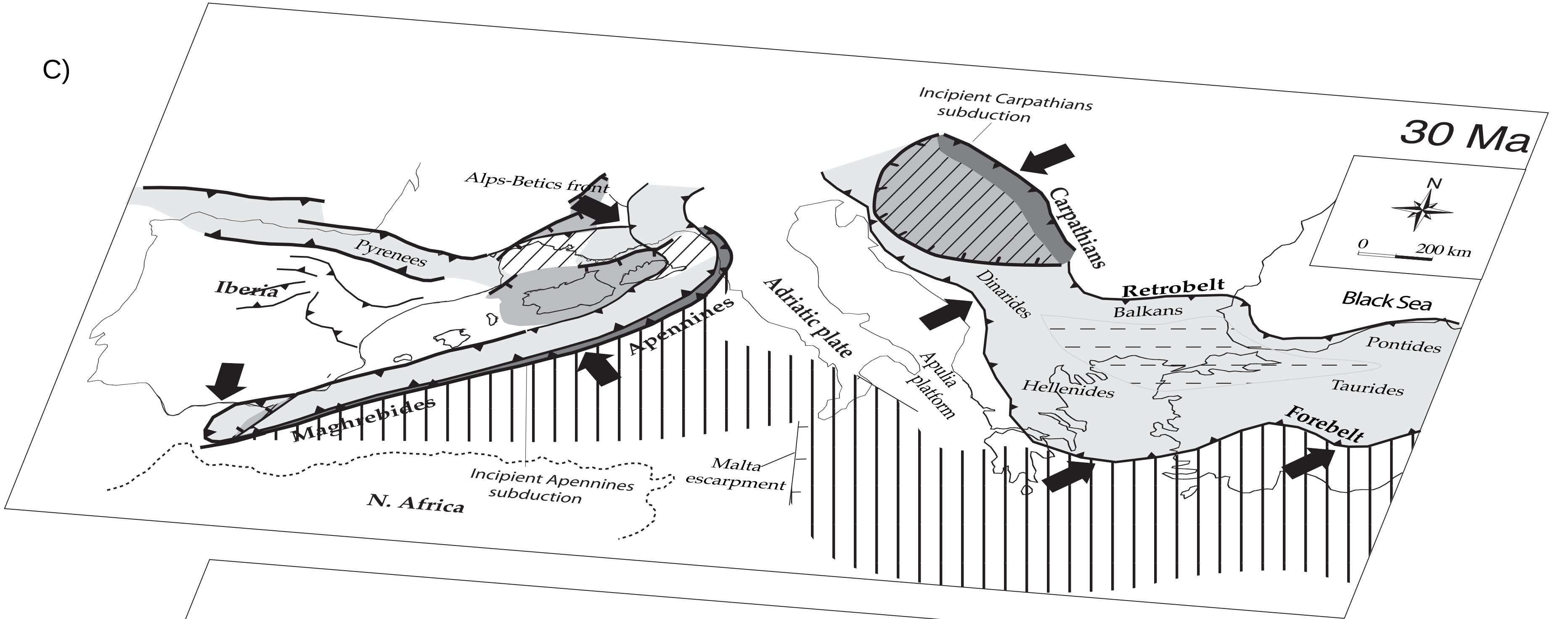


Fig. 1a-b (Lustrino and Carminati P4 2007)

C)



D)

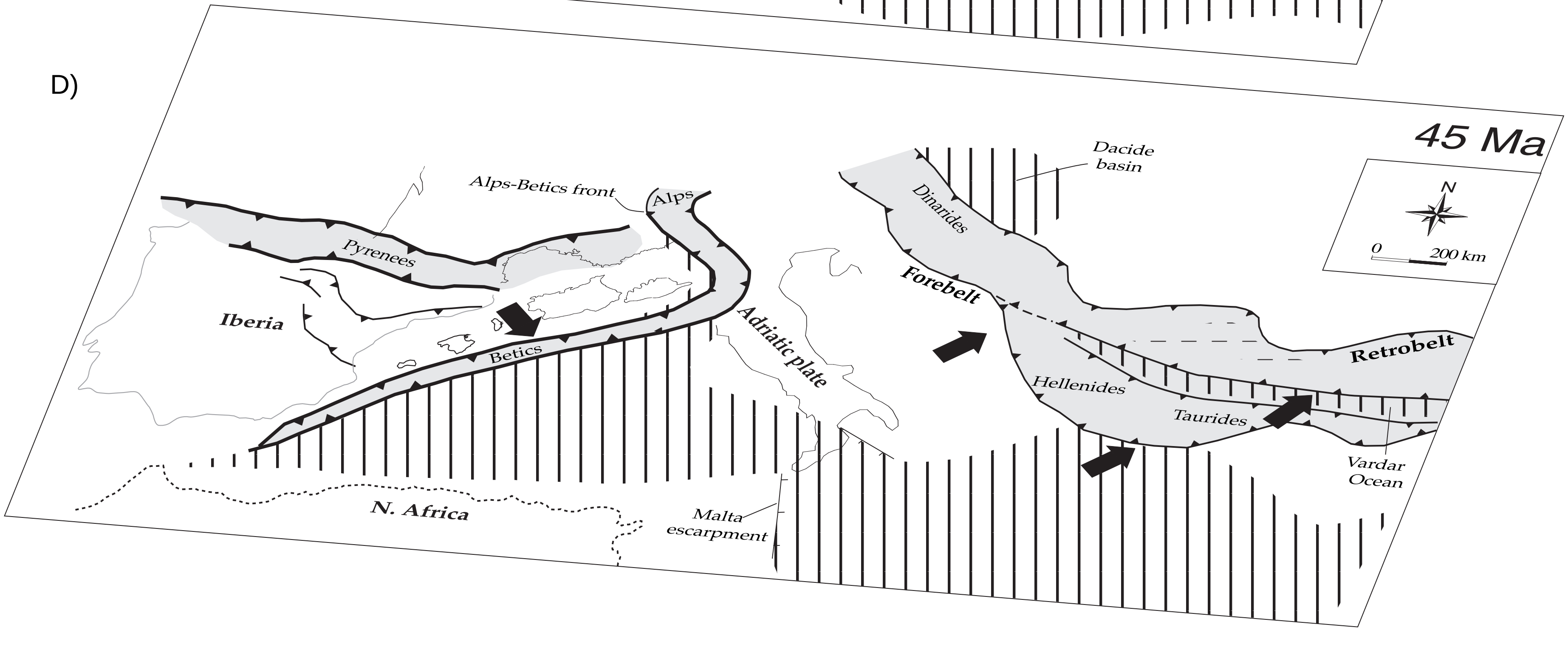


Fig. 1c-d (Lustrino and Carminati P4 2007)

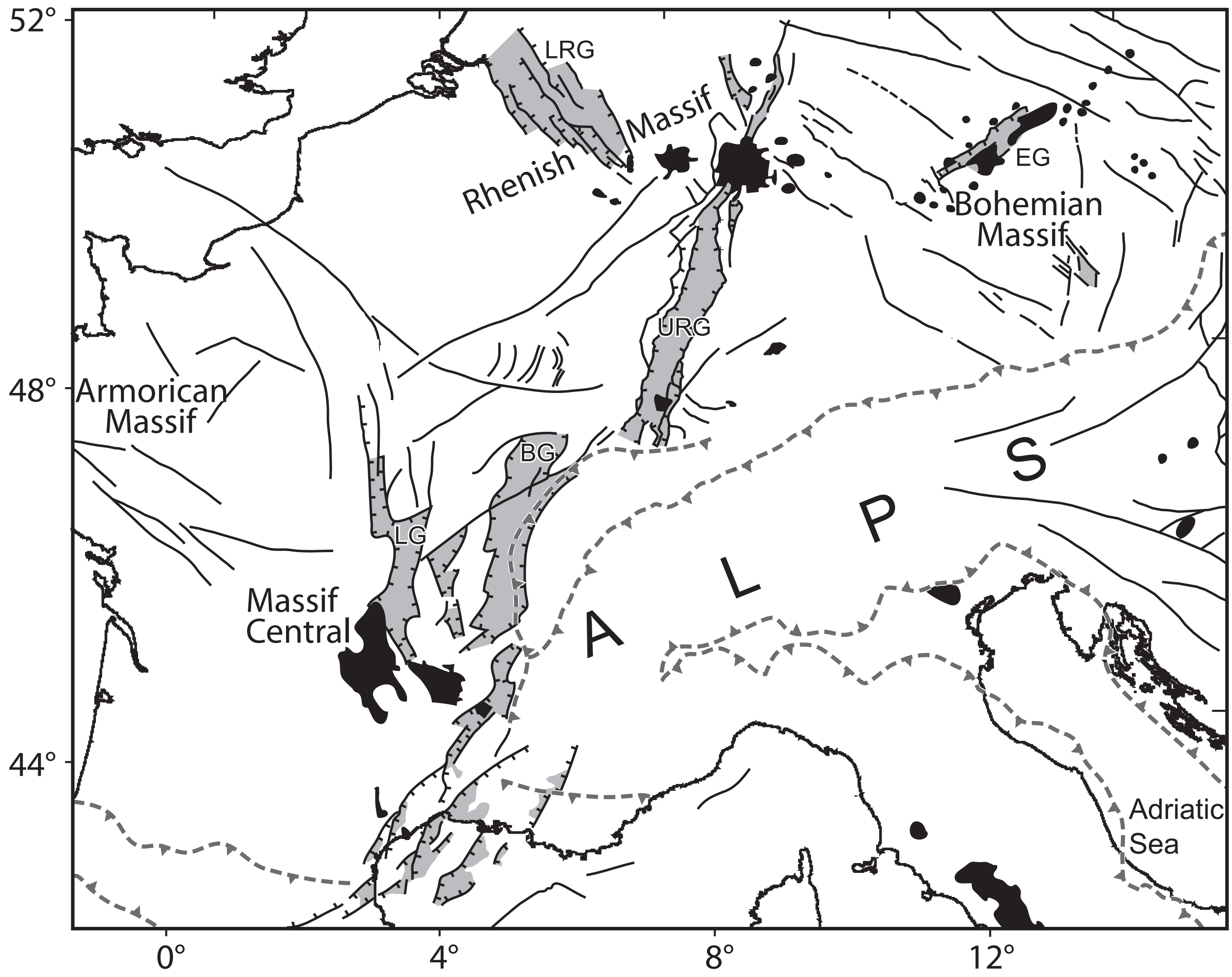
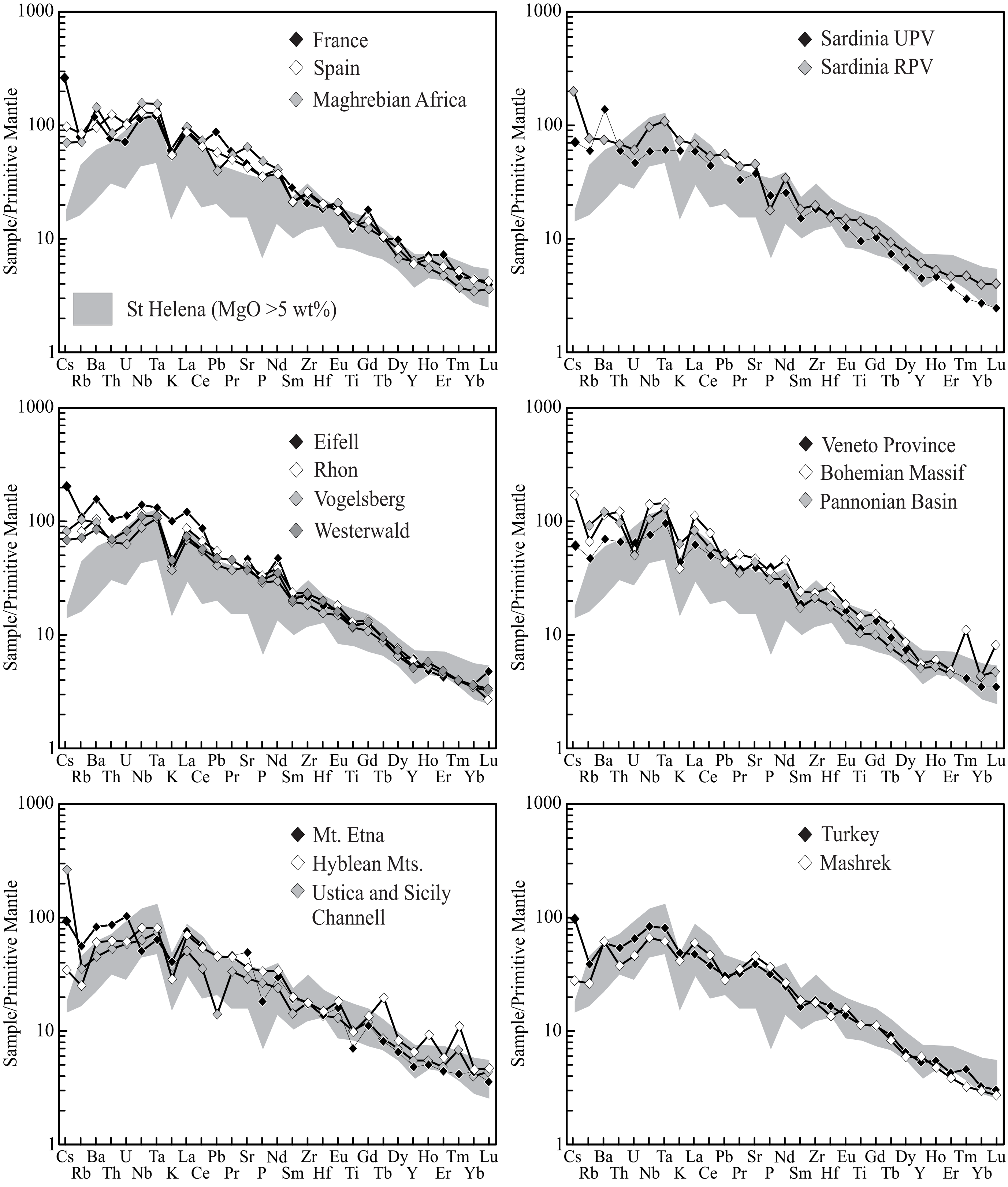


Fig. 2 (Lustrino and Carminati P4 2007)



Lustrino and Carminati
Fig. 3

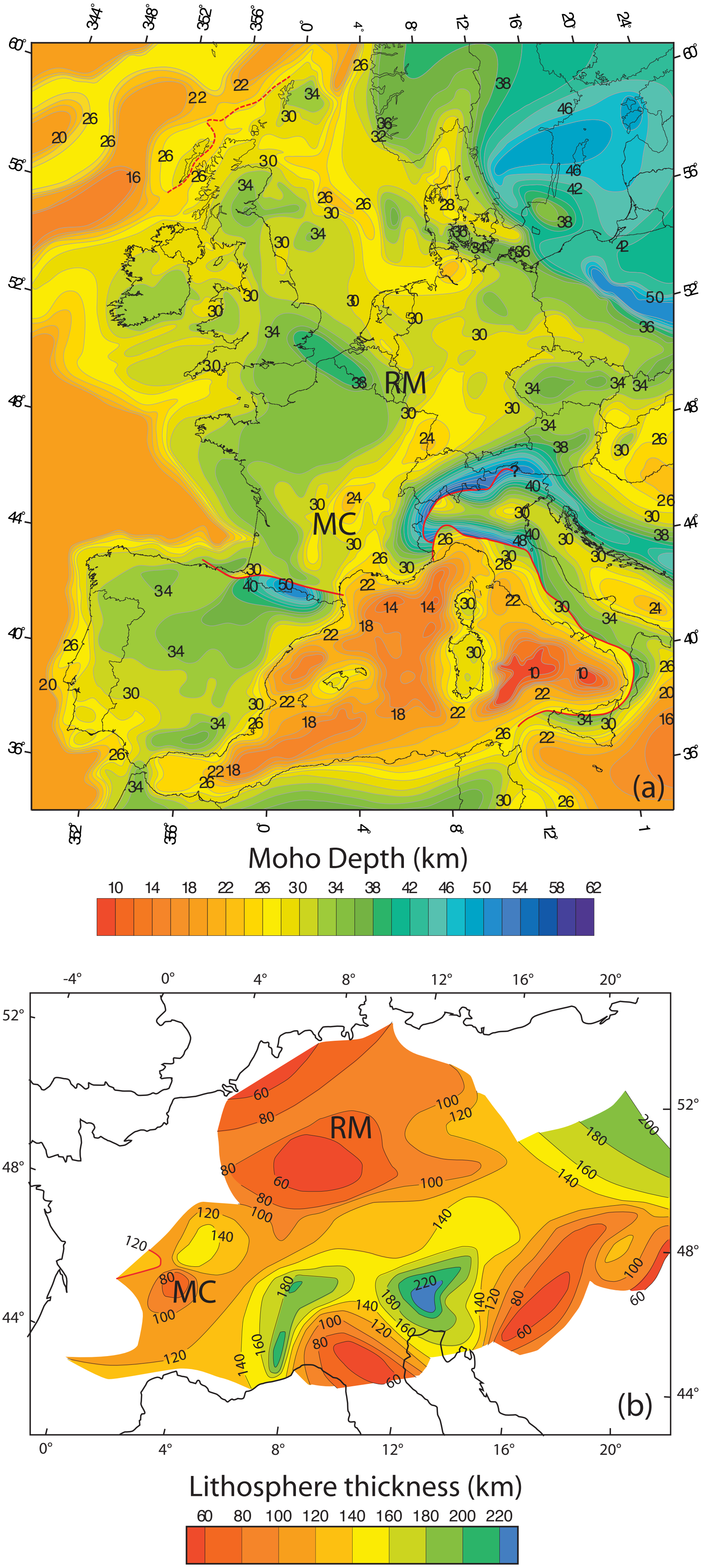


Fig. 4 (Lustrino and Carminati P4 2007)