Tectonophysics xxx (2012) xxx–xxx

Contents lists available at SciVerse ScienceDirect

Tectonophysics

journal homepage: www.elsevier.com/locate/tecto

Geodynamic evolution of the central and western Mediterranean: Tectonics vs. igneous petrology constraints

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ARTICLE INFO

Article history:
Received 18 October 2011
Received in revised form 9 January 2012
Accepted 19 January 2012
Available online xxxx

Keywords:
Mediterranean geodynamics
Tertiary
Subduction flip
Magmatism
Petrology

ABSTRACT

We present a geodynamic reconstruction of the Central–Western Mediterranean and neighboring areas during the last 50 Myr, including magmatological and tectonic observations. This area was interested by different styles of evolution and polarity of subduction zones influenced by the fragmented Mesozoic and Early Cenozoic paleogeography between Africa and Eurasia. Both oceanic and continental lithospheric plates were diachronously consumed along plate boundaries. The hinge of subducting slabs converged toward the upper plate in the double-vergent thick-skinned Alps–Betics and Dinarides, characterized by two slowly-subsiding foredeeps. The hinge diverged from the upper plate in the single-vergent thin-skinned Apennines–Maghrebides and Carpathians orogens, characterized by a single fast-subsiding foredeep. The retreating lithosphere deficit was compensated by asthenosphere upwelling and by the opening of several back-arc basins (the Ligurian–Provençal, Valencia Trough, Northern Algerian, Tyrrhenian and Pannonian basins). In our reconstruction, the W-directed Apennines–Maghrebides and Carpathians subductions nucleated along the retro-belt of the Alps and the Dinarides, respectively. The wide chemical composition of the igneous rocks emplaced during this tectonic evolution confirms a strong heterogeneity of the Mediterranean upper mantle and of the subducting plates. In the Apennine–Maghrebide and Carpathian systems the subduction-related igneous activity (mostly medium- to high-K calcalkaline melts) is commonly followed in time by mildly sodic alkaline and tholeiitic melts. The magmatic evolution of the Mediterranean area cannot be easily reconciled with simple magmatological models proposed for the Pacific subductions. This is most probably due to synchronous occurrence of several subduction zones that strongly perturbed the chemical composition of the upper mantle in the Mediterranean region and, above all, to the presence of ancient modifications related to past orogeneses. The classical approach of using the geochemical composition of igneous rocks to infer the coeval tectonic setting characteristics cannot be used in geologically complex systems like the Mediterranean area.

1. Introduction

A vast scientific literature is available for the Cenozoic geological and magmatological evolution of the Central–Western Mediterranean (Figs. 1 and 2), with nearly all the most important geological formations, structures, igneous and metamorphic rocks having been described in detail. Also the crustal and the upper mantle structure have been investigated using seismic and seismological data, showing a complex scenario, governed by several subduction zones and rifting environments (e.g., Amato et al., 1993; Giacomuzzi et al., 2011; Piromallo and Morelli, 2003; Wortel and Spakman, 2000). The present-day and the past stress state has been investigated via in-situ measurements, seismological and classical structural geology analyses. The past and present plate kinematics of the region has been constrained by reconstructions based mostly on paleomagnetism and, in recent years, on space geodesy (mainly GPS). This huge amount of knowledge, together with field geology, and the presence of igneous activity with peculiar geochemical and petrographic characteristics, have been condensed in a wealth of geodynamic evolutionary models (e.g., Boccaletti and Guazzone, 1974; Carminati et al., 1998a, 1998b; Chalouan et al., 2008; Channel and Mareschal, 1989; Csontos and Voros, 2004; Doglioni, 1991; Faccenna et al., 1997; Gueguen et al., 1998; Malinverno and Ryan, 1986; Mauffret, 2007; Rosenbaum et al., 2002a, 2002b; Schmid et al., 2008; Tari, 2002; Wortel and Spakman, 2000). Why, thus, another study on the Central–Western Mediterranean geology?

One of the major problems for a full understanding of Mediterranean geodynamics is that, with few exceptions, the available scientific literature represents the results of single discipline investigations rather than multidisciplinary approaches. The result is that the various geochemical, petrological, structural and tomographic models do not fully take into consideration the constraints evidenced by the other disciplines. In this work we integrate all the available pieces of information in a geodynamic reconstruction focussing on

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doi:10.1016/j.tecto.2012.01.026

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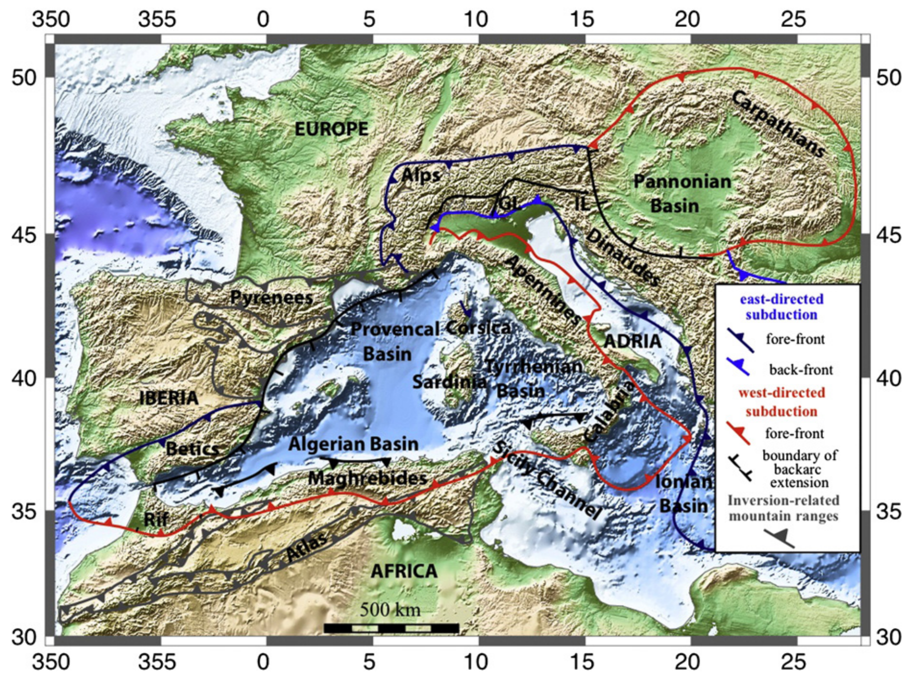

Fig. 1. Simplified present-day geodynamic scenario of the Central–Western Mediterranean region superimposed on the topography and bathymetry. GL: Giudicarie Lineament; IL: Insubric Line.

the Central–Western Mediterranean area, from the Gibraltar Straits to Western Greece, comprising Maghrebian Africa and Central–Eastern European domains (Rhine Graben to Pannonian–Carpathians). Carminati et al. (2010) presented fifty plane-section views of the post-50 Myr evolution of the Central–Western Mediterranean, highlighting the most important structural constraints and all the main igneous rock districts. Starting from this model, we present two evolutionary cross section views, one roughly NW–SE directed, passing through NE Spain to the Ionian Sea, and another from Southern France to the Carpathians, passing through the Adriatic Sea (see the movie associated with this manuscript, downloadable from the journal website). The cross sections presented here at 1 Myr interval are the first that take into consideration the full geological, metamorphic and igneous petrology of the investigated area and have a temporal continuity sufficient to evaluate their feasibility. A brief description of the main tectonic and magmatic features of the area is also provided, with special reference to the geology of the Central–Western Mediterranean area. For further details on tectonics and

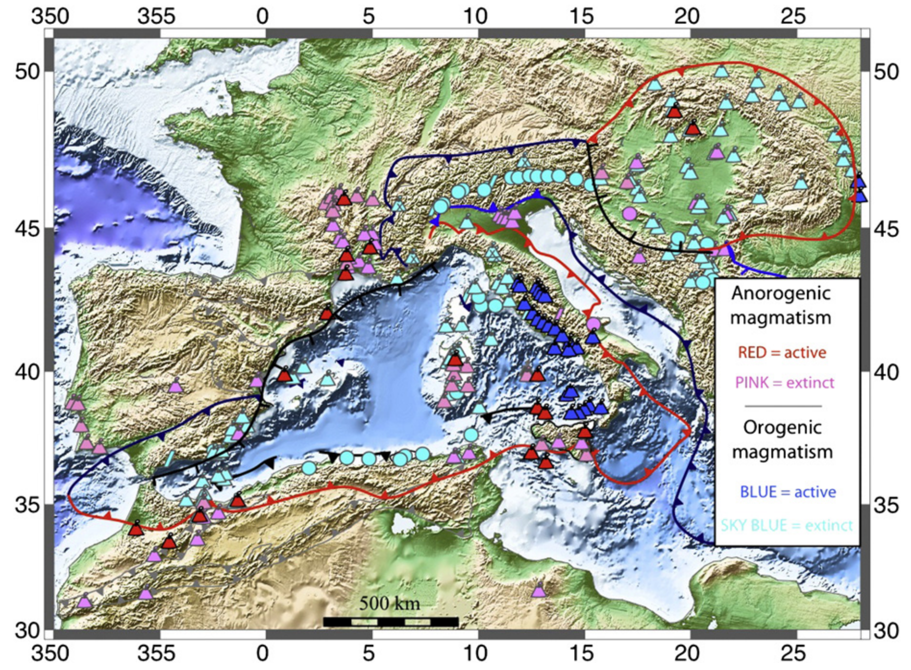

Fig. 2. Distribution of Tertiary magmatism in the Central–Western Mediterranean region. Triangles: volcanics and pyroclastics; Triangles with crosses: volcaniclastics; Circles: plutons; Slashes: dykes. Red symbols: active "anorogenic" igneous rocks; Pink symbols: extinct "anorogenic" igneous rocks; Blue symbols: active "subduction-related" igneous rocks; Sky Blue symbols: fossil "subduction-related" igneous rocks. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

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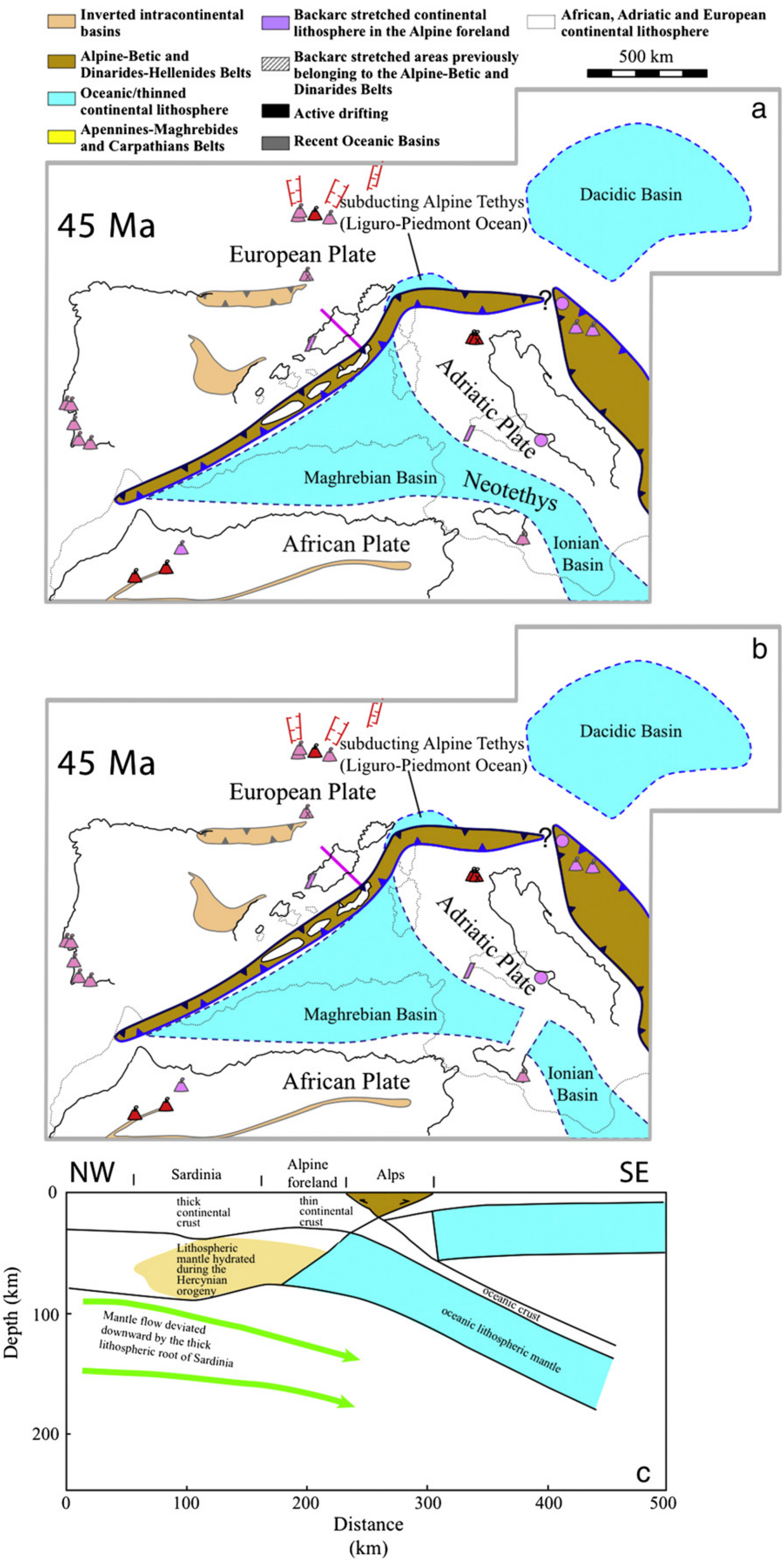

Fig. 3. Geodynamic reconstructions (a, b) and related cross section (c) at 45 Ma. Panel a) shows the reconstruction adopted in this work, characterized by a continuous Ionian Ocean. Panel b) shows an alternative reconstruction with two oceanic basins (Maghrebian and Ionian), separated by a continental to shallow-water corridor. The trace of the section is shown in panel a.

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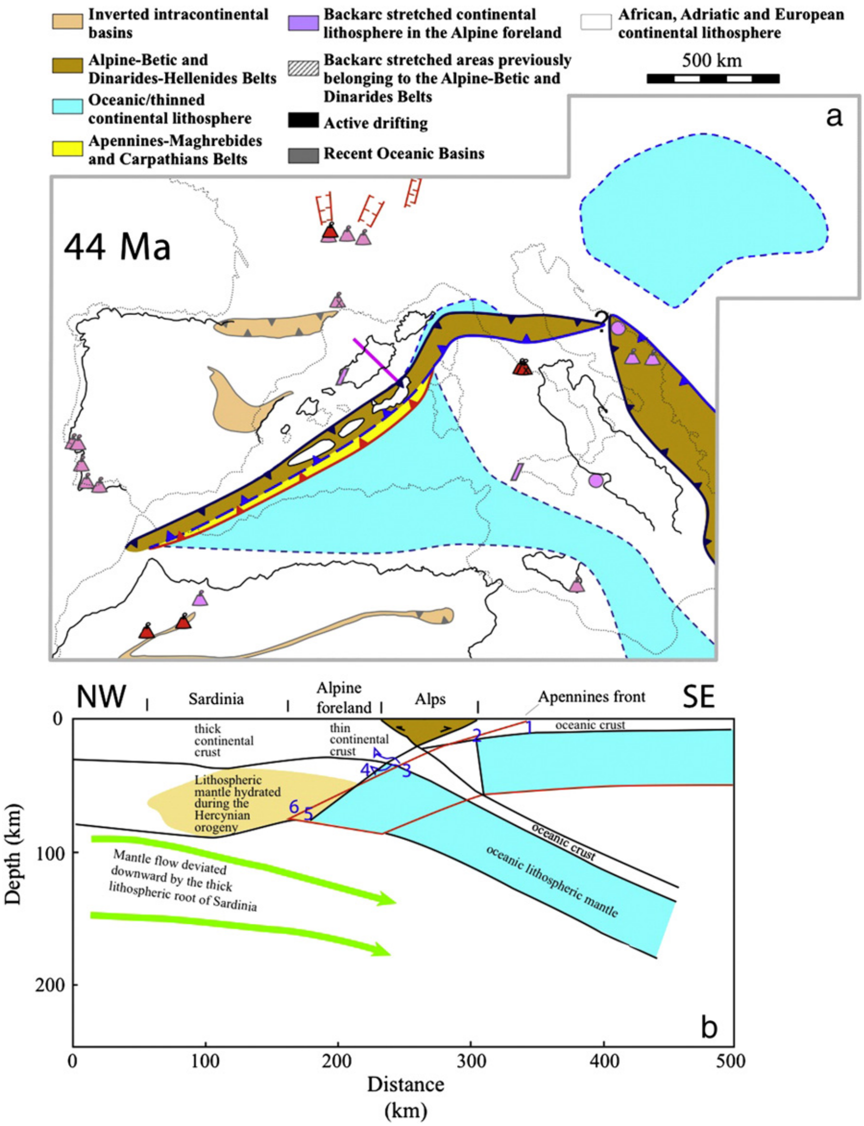

Fig. 4. Geodynamic reconstruction (a) and related cross section (b) at 44 Ma. The trace of the section is shown in panel a.

magmatism, we refer the readers to recent thematic issues on this argument (e.g., Beccaluva et al., 2007; Beltrando et al., 2010; Cavazza et al., 2004; Schmid et al., 2008, and references therein).

The 100 cross-sections presented here (2 × 50) are only a simplified view of a much more complex scenario, controlled not only by plate (or sub-plate) scale kinematics, but also by inherited structures (e.g., Variscan suture zones and structural grain, Mesozoic Tethyan rifting and related discontinuities). The complexity of the beginning of the Apennine–Maghrebide subduction system is investigated at the latitude of Sardinia. We show that shape and behavior of the subducting slab as well as its kinematics and the magmatism associated with subduction may have been potentially influenced by the occurrence of the Variscan orogeny in this region. In addition, we discuss how the occurrence of multiple contemporaneous subductions could have chemically perturbed the mantle in the Mediterranean region, rendering the interpretation of magmatism in terms of active geodynamic processes a difficult task.

2. Tectonic and magmatological evolution of the Central–Western Mediterranean

The Central–Western Mediterranean is floored by sub-basins (Alboran, Valencia, Provençal, Algerian and Tyrrhenian basins), which developed essentially during the last 40–30 Myr. This area is geologically younger than the Eastern Mediterranean, which is floored possibly by Mesozoic oceanic crust, with a thick sedimentary blanket, or by thinned continental crust (Robertson and Dixon, 1984). The Central-Western Mediterranean basins are younger moving from west to east (Rehault et al., 1984). The geological evolution of this area is connected with the relative movements of three main plates (Africa, Adria and Europe; Fig. 3) plus an unknown number of smaller continental terranes and oceanic or transitional basins. Paradoxically, the development of several basins occurred in a context of relative convergence between Africa and Europe (e.g., Durand et al., 1999). The maximum amount of North–South Africa/Europe relative motion at the Tunisia longitude was about 135 km in the last 23 Ma, more than five times shorter with respect to the eastward migration of the Apennines arc which moved eastward more than 700 km during the last 23 Ma (Gueguen et al., 1998). For this reason we speculate that the roughly E-directed migration of the Apennine–Maghrebide arc is not a consequence of the relative N–S relative convergence between Africa and Europe, but it is rather a consequence of the Apennine–Maghrebide subduction rollback, as broadly discussed below.

The mountain belts bordering the basins of the Mediterranean region can be divided in two end-members. The Alps–Betics and Dinarides are double-vergent and thick-skinned belts, bounded by

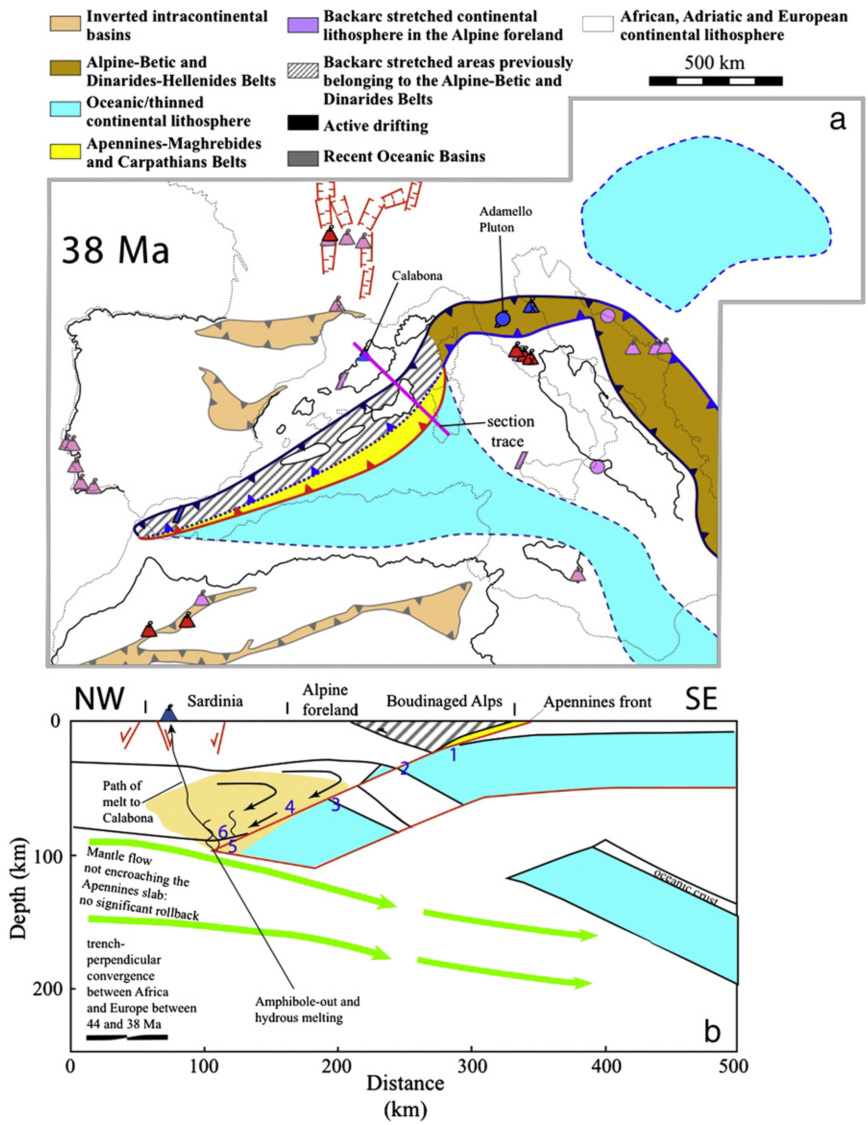

Fig. 5. Geodynamic reconstruction (a) and related cross section (b) at 38 Ma. The trace of the section is shown in panel a.

two slowly-subsiding foredeeps (e.g., Kummerow et al., 2004). The single-vergent thin-skinned (e.g., Bally et al., 1986; Scrocca et al., 2005) Apennines–Maghrebides and Carpathians belts are, on the contrary, characterized by a single fast-subsiding foredeep. Other relevant differences between Alpine–Betic–Dinaride and Apennine–Maghrebide–Carpathian chains are, respectively (Carminati et al., 2004, 2010): 1) subduction hinge moving towards vs. moving away from the upper plate; 2) higher metamorphic grade vs. lower metamorphic grade; 3) thickened lithosphere vs. shallow asthenosphere in the upper plate; 4) no back-arc basin and no syn-subduction magmatism vs. very wide back-arc basins and abundant syn-subduction arc-tholeiitic to calcalkaline and potassic/ultrapotassic magmatism; 5) SiO_2-oversaturated vs. SiO_2-saturated to strongly SiO_2-undersaturated compositions of igneous rocks.

The tectonic and magmatological evolution described in this paper is depicted in Figs. 3–10 and in the attached movie showing paleogeographic maps and cross sections through the study area referred to the last 50 Myr. Before discussing them, it is necessary to briefly highlight limitations and problems related with our reconstructions.

It is finally stressed that intraplate mountain belts occur in the Mediterranean and Circum-Mediterranean area (Atlas, Iberian Chain, Pyrenees). These belts developed from the inversion of former sedimentary basins and were not associated with subduction processes. The detailed treatment of intraplate belts is beyond the scope of this work, although they are represented in the movie. The reader is referred, among others, to the following papers for a detailed description of their evolution: Guiraud (1998) and Frizon de Lamotte et al. (2000, 2009) for the Atlas; Guimerà et al. (2004) for the Iberian Chain; Vergés et al. (2002) and Lacombe and Jolivet (2005) for the Pyrenees.

2.1. The movie: some caveats

As already discussed by Carminati et al. (2010) the scale of the movie and the complexity of the evolution of the region forced us to present some features in a schematic way: 1) the symbol of normal faults indicates the occurrence and direction of stretching in an area, rather than the existence of a real fault. 2) At present, geophysical and geological data suggest that the shape of the contractional belt fronts

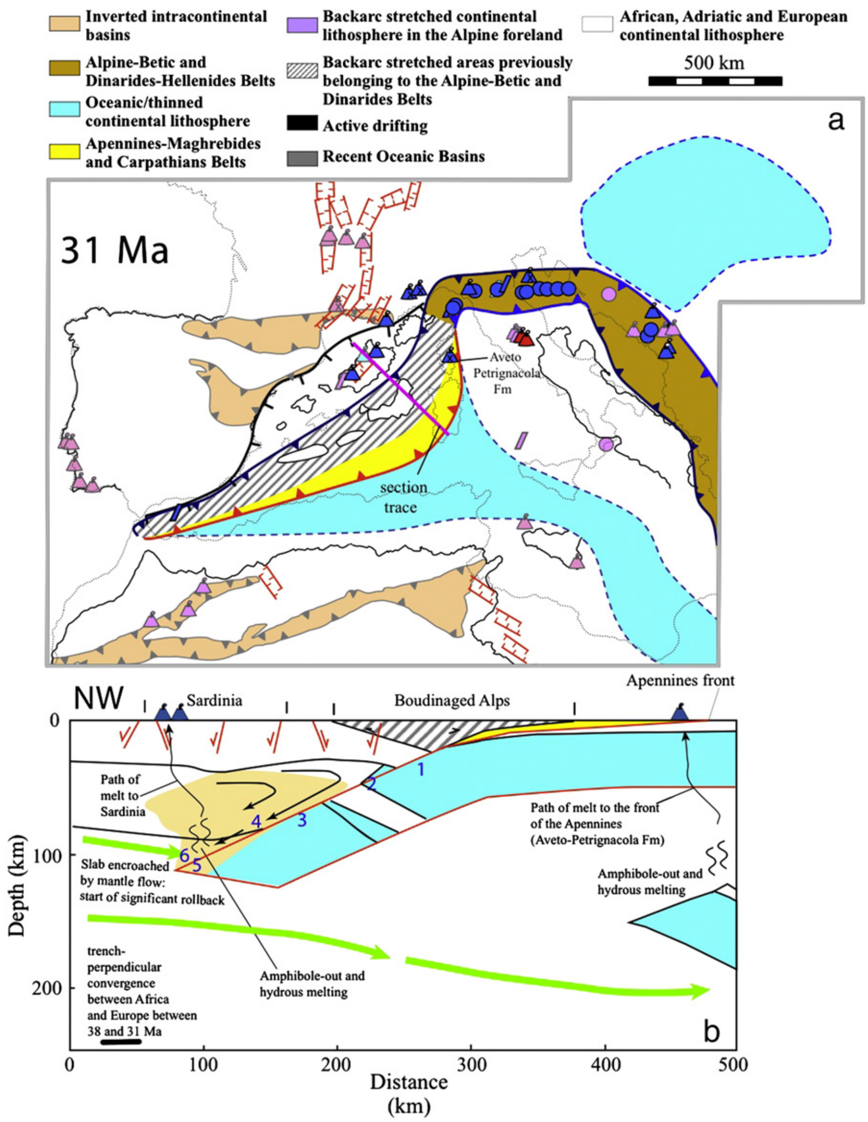

Fig. 6. Geodynamic reconstruction (a) and related cross section (b) at 31 Ma. The trace of the section is shown in panel a.

is characterized by salients and recesses and is highly segmented. The fronts were likely segmented also in the past but, owing to the lack of data to constrain such lateral variations for past times, they are shown as linear features in the movie. 3) The symbols related to magmatism are not proportional to the volume of associated rocks.

In this work, the relative motions between Africa and Europe were reconstructed using the model by Rosenbaum et al. (2002b). In the movie, Europe is kept fixed, while Africa and Adria are moved coherently. The reason for this choice is discussed later. The proposed geodynamic scenarios (e.g., position of continents; fronts of Alps and Dinarides and their connection in the past) are a function of the adopted plate kinematic reconstruction. Different kinematic reconstructions (e.g., Albarello et al., 1995; Dewey et al., 1989; Mazzoli and Helman, 1994; Schmid et al., 2008; Stampfli and Hochard, 2009; Stampfli et al., 2002) would have implied different reconstructions.

Also the age of the igneous activity is problematic. The geochronological data are incomplete and almost no systematic coverage exists. Most of the ages are based on few and old K–Ar method (particularly for the Veneto volcanic district), while detailed $^{40}Ar/^{39}Ar$ ages are few. In submerged basins (e.g., Tyrrhenian Sea) or continental subsided basins (e.g., Northern Sardinia) only rocks associated with the most recent activity have been dated. In these cases, available data provide only a lower bound for the age of magmatism. In other cases the erosion of the volcanic material (particularly for the Western Alps) may prevent the precise dating of the main magmatic periods of activity.

Furthermore, the distinction between the crude or aseptic observation of a natural phenomenon (e.g., the strike of a fault, the areal outcrop of a formation and the classification of an igneous rock) and the interpretation of the geological–geochemical message is subtle. Different interpretations of the same geological features have led to partially or completely different conclusions from different authors (e.g., see discussion in Lustrino et al., 2011). The existence of active rather than passive movements of the upper mantle, the engine itself responsible for the main geological structures (fold-and-thrust belts, basin opening) and, above all, the capacity to interpret the geochemical signals of the igneous rocks (e.g., whole-rock composition and isotopic ratios) into geological constraints are far for being fully understood. The same basic field geology observations (e.g., the vergence of a thrust) can be interpreted in completely different ways (e.g., as foreland-vergent thrust, with implication on subduction polarity or rather as a back-thrust of a larger orogen, implying completely opposite polarity of subduction). The interference between different strain fields induced by different geodynamic processes as, for example, recorded along the Betic Chain in SE Spain or in the

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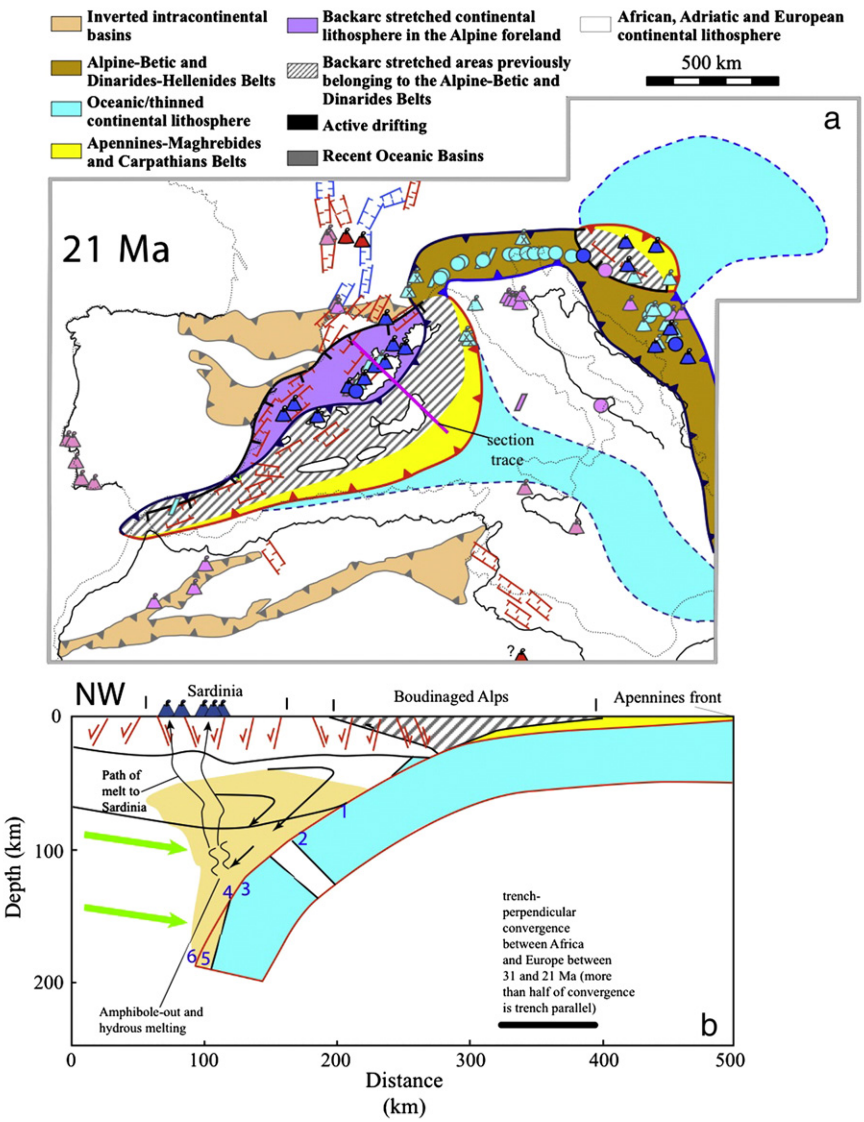

Fig. 7. Geodynamic reconstruction (a) and related cross section (b) at 21 Ma. The trace of the section is shown in panel a.

Sicily Channel, contribute to render more complex and far from the objectivity any kind of interpretation.

Despite our cross sections are based on a large number of geological constraints, they must clearly be considered as an interpretative model. In particular, the shape (dip) of subducting plate for past times is completely unknown. We chose to mimic present-day geometry of subduction zones, with shallow slabs in the Alps (~40°; Piromallo and Morelli, 2003), Betics (~45°; Morales et al., 1999) and Dinarides–Hellenides (~25°; Bennett et al. 2008; Christova and Nikolova, 1993; Papazachos et al., 2005) and steep slabs below the Apennines (~70°, Frepoli et al., 1996) and the Carpathians (~75°; Oncescu and Bonjer, 1997; Oncescu and Trifu, 1987).

Finally, it is emphasized that the evolution of the Mediterranean was controlled by strongly non-cylindric 3D processes, whose representation in 2D cross sections is not always straightforward, since out-of-section motions (of tectonics blocks or magma) cannot be represented.

2.2. The Adriatic plate problem

The presence of the Adriatic plate represents a critical aspect for the formation and the evolution of the entire Central–Western Mediterranean. The Adriatic plate represents a continental plate where essentially shallow to deep marine carbonates were deposited almost continuously during Mesozoic and Early Tertiary (e.g., Bosellini, 2002; Vezzani et al., 2010; Vlahovic et al., 2005, and references therein). Originally it was about 1200 km by 400 km wide, elongated in NW–SE direction. The presence of a single continental plate (Adriatic plate) or a collage of more continental plates (e.g., Adriatic, Friuli, Kruja, Gavrovo–Tripolitsa, Menderes, Apulia, Campano–Lucana and Laziale–Abruzzese carbonate platforms), interrupted by deep basins, remains an unresolved aspect (e.g., Korbar, 2010 and references therein). Adria is alternatively considered to be in crustal continuity with the African mainland, at least during Cenozoic times (e.g., Muttoni et al., 2001; Schettino and Turco, 2011) or separated from the latter by an oceanic realm (called in literature Ionian, Mesogean, Ligurian, Eastern Alpine Tethys or Lucanian Ocean; Fig. 3; e.g., Catalano et al., 2001; Doglioni, 1991; Guerrera et al., 2005, and references therein), part of the Neotethys. In the following we will use the term Ionian Ocean for the oceanic realm comprising the Ionian and Maghrebian Basins. However, it must be emphasized that the Ionian lithosphere has been also interpreted as transitional-to-continental (e.g., Calcagnile et al., 1982; Cernobori et al., 1996). In this case, Adriatic and African plates should not be considered as two different plates, although some relative motions between the two sub-plates could have occurred.

Our model of evolution assumes the presence of an oceanic lithosphere completely separating the Adriatic plate from Africa (Ionian

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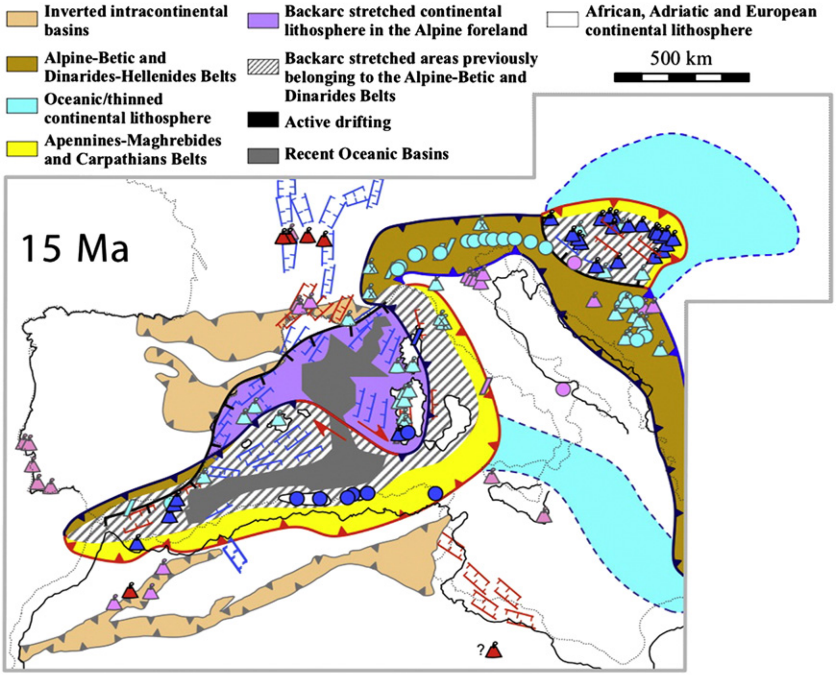

Fig. 8. Geodynamic reconstruction at 15 Ma.

Ocean), but, at the same time, we hypothesize the Adriatic plate as a lithospheric element kinematically coherent with main Africa after the (Mesozoic?) cessation of the Ionian spreading. It is worth noting, however, that present GPS data show the Adriatic plate slowly moving away from Africa (Devoti et al., 2008). According to other reconstructions (e.g., Handy et al., 2010) the Adriatic plate was in crustal continuity with Africa but, at the beginning of the Cenozoic, it was completely separated by it, being bounded along the western and southern margins by oceanic lithosphere. The kinematic coupling between Africa and Adria during the last 50 Myr strictly depends on the age of the Ionian oceanic lithosphere. Based on indirect arguments, the proposed ages vary from Cretaceous (Catalano et al., 2001; Dercourt et al., 1986) to Triassic or even Permian (Stampfli and Borel, 2002), i.e., older than the time span considered in our model. This is consistent with our assumption of kinematic coupling between Africa and Adria.

In our reconstruction, the Neotethys (Ionian Ocean) is assumed to be continuous from the Ionian Basin to the Maghrebian Basin (Fig. 3a). Alternative paleogeographic reconstructions suggest the occurrence of a continental corridor, where shallow water sedimentation occurred and that permitted the circulation of dinosaurs from Africa to Adria (Bosellini, 2002; Conti et al., 2005; Frizon de Lamotte et al., 2011; Muttoni et al., 2001; Rosenbaum et al., 2002a; Schettino and Turco, 2011; Zarcone et al., 2010). We adopt a single Ionian

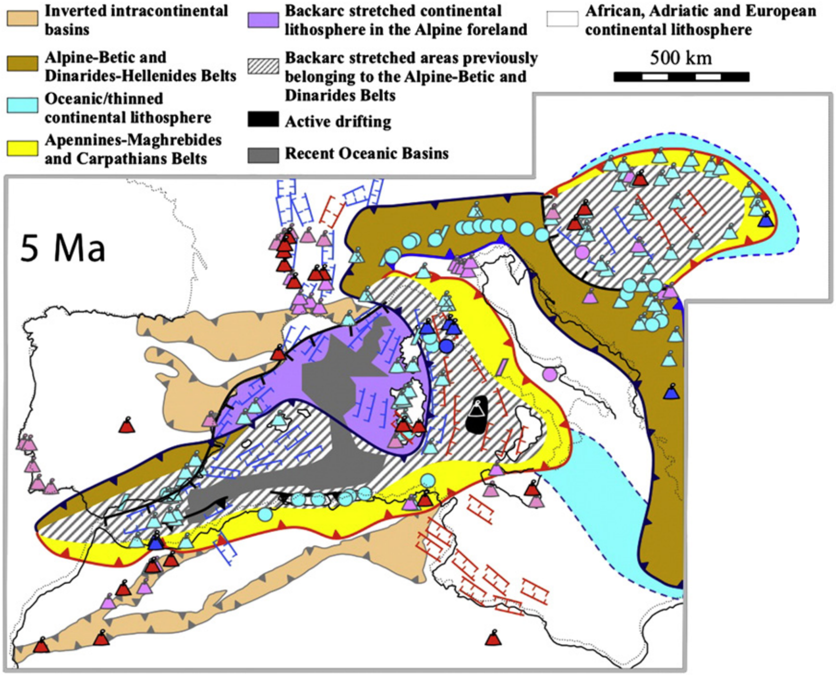

Fig. 9. Geodynamic reconstruction at 5 Ma.

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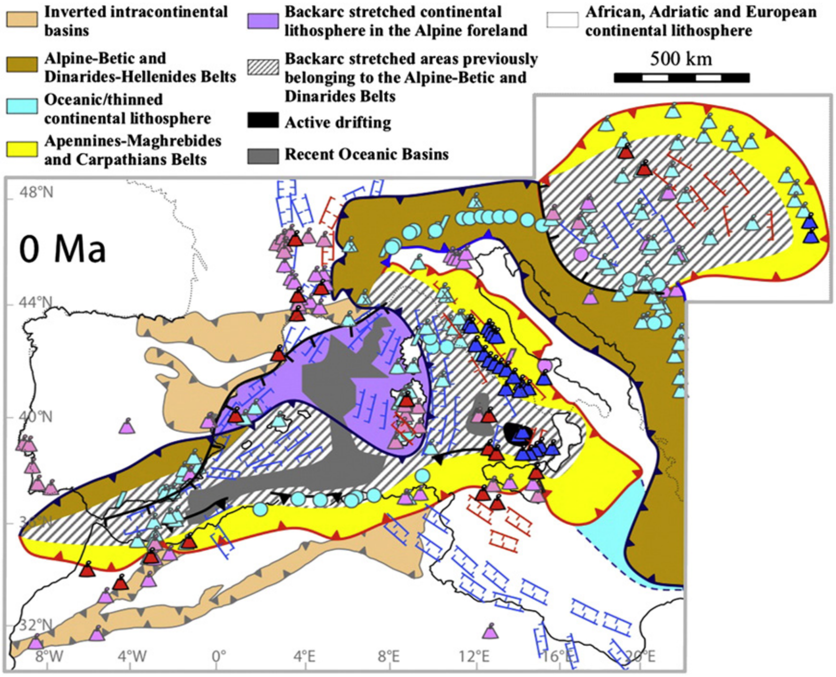

Fig. 10. Present-day geodynamic scenario for the Central–Western Mediterranean.

Ocean (Fig. 3a), rather than two distinct oceans (Ionian and Maghrebian Oceans; Fig. 3b), because of the in-depth continuity of seismicity in the Benioff zone beneath the Calabrian Arc (Catalano et al., 2001).

2.3. Closure of the Neotethys and Alpine Tethys oceans and continental collision

The Alps–Betics and Dinarides belts are collisional orogens that were preceded by the earlier Andean-style subduction of several branches of the Neotethys and Alpine Tethys Oceans (Fig. 3). Such oceanic subduction stages are testified by the occurrence of sparse ophiolite remnants, in many cases affected by blue-schist- and eclogite-facies metamorphism (Berger and Bousquet, 2008; Bousquet et al., 2008; Dal Piaz et al., 1972; Schmid et al., 2008; Spalla et al., 1996; Tortorici et al., 2009; Vitale et al., 2011) and associated with deposition of flyschoid sequences (e.g., Polino et al., 1990). No magmatism synchronous with the oceanic subduction phase is known along the north-western margin of the Adria plate (present-day Alps). The description of the complex (and still disputed) paleogeography of the Neotethys is beyond the scope of this work, which is mostly concerned with the post-collisional evolution of the Central and Western Mediterranean. The reader is referred to the works (among others) of Dercourt et al. (2000), Handy et al. (2010), Polino et al. (1990), Schmid et al. (2008), Scotese (1991), Stampfli and Hochard (2009) and Stampfli et al. (2002) for a broader discussion on this subject.

In the Dinarides, the consumption of the Neotethys realm occurred in Early Cretaceous time, when the Tisza block and the eastern Margin of the Adriatic plate collided (Schmid et al., 2008; Tari, 2002). The Meliata–Maliac–Vardar Ocean, originally separating the Adria plate from Eastern Europe, was, indeed, recycled with NE-directed polarity of subduction. This ocean was the westernmost branch of a larger Triassic–Jurassic Neotethys Ocean (e.g., Schmid et al., 2008, and references therein). The Late Jurassic to Early Cretaceous Vardar ophiolitic melange represents one of the suture zones of this oceanic mass. The collision of the Adriatic plate with Eastern Europe, coupled with important dextral shear zones, led to the formation of the Dinarides orogen, running NW–SE along the eastern margin of the Adriatic Sea (Korbar, 2010; Schmid et al., 2008, and references therein). The Dinarides are characterized by SW-directed thrusts and NE-directed back-thrusts in the Balkans retro-belt, with the Adriatic Sea representing the main foreland, and the Moesian Platform as the retro-foreland. Abundant magmatism, characterized by calcalkaline to ultrapotassic composition, is recorded in this area (e.g., Kovacs et al., 2007, and references therein).

The Alpine Tethys (alternatively referred to as Ligurian–Piedmontese, Liguride, Ligurian Piedmontese, Piemontese, Valais or Penninic Ocean, or western branch of the Neotethys), that separated the NW Adriatic plate from western Europe, was kinematically connected with the sea-floor spreading of the North Atlantic Ocean started from Early Jurassic times. The Alpine Tethys was subducted beneath the Adriatic plate since at least Early Cretaceous time (Dal Piaz, 2010; Handy et al., 2010; Rosenbaum and Lister, 2005; Schmid et al., 2008, and references therein). In the Alps, the continent–continent collision between the northern margin of the Adriatic plate and the southern passive margin of Central Europe occurred diachronously mostly during Eocene (~55–35 Ma), but commenced as early as Early Cretaceous (~110 Ma) as indicated by the radiometric age of high pressure metamorphic minerals (e.g., Dal Piaz et al., 2001; Nagel, 2008; Rosenbaum and Lister, 2005, and references therein). The collision of Africa (or an intervening continental micro-plate; Doglioni, 1992; Guerrera et al., 1993) with westernmost Europe (Iberia plate) led to the formation of the Betics in SE Spain. This aspect is known in less detail and the occurrence of oceanic subduction in this realm has been also questioned (see Michard et al., 2002 for a discussion). According to our view (Carminati and Doglioni, 2005; Carminati et al., 2010; Doglioni et al., 1999a), the Betics and the Alps were originally connected, forming a single NE–SW-trending fold-and-thrust chain with top-to-NW tectonic transport and top-to SE retro-belt. The continuity of the belt can be inferred restoring to their original position the Corsica, Calabria, Peloritani Mts. and Kabilies terranes, which display clear evidence of Alpine deformation in terms of age and tectonic transport (Heymes et al., 2010; Michard et al., 2006; Molli, 2008; Vignaroli et al., 2009, and references therein). The dismembering and dispersal of the terranes that formerly formed the Alpine belt between present-day Alps and Betics, occurred in the hangingwall of the back-arc basins related to the Apennines subduction. Despite the continental collision between Africa and

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Europe, remnants of likely Mesozoic oceanic branches still occur in the Central and Eastern Mediterranean and are at present consumed in the Calabrian and Hellenides subduction zones (Fig. 10). The impact of the occurrence of this remnant oceanic lithosphere (Ionian basin) on the geodynamic evolution (Fig. 3–10) of the Mediterranean is amply discussed in the following.

2.4. Initiation of W-directed subduction zones

The origin of W-directed subduction zones in the Central–Western Mediterranean is still debated. In this area two are the main competing models: 1) the Cenozoic evolution was characterized by a single (W-directed) permanent subduction polarity since Late Cretaceous time (e.g., Faccenna et al., 2001; Jolivet et al., 1998; Tortorici et al., 2009); 2) the proto-Central-Western Mediterranean was the place of an earlier Cretaceous E–SE-directed Alpine subduction, with a later (late Eocene?) flip into the W-directed Apennine–Maghrebide subduction, and related back-arc stretching (e.g., Doglioni et al., 1999a; Gueguen et al., 1997; Handy et al., 2010; Michard et al., 2002; Molli and Malavieille, 2011). The occurrence of a continuous Alpine belt before the initiation of the Apennines subduction is here proposed (Fig. 4) for the following reasons: 1) the rocks of the Betics and of Alpine Corsica (Molli, 2008), as well as the metamorphic relics dredged in the Tyrrhenian sea-floor and the basement outcrops of Tuscany (Argentario, Giglio, Monti Leoni, Gorgona; Rossetti et al., 1999, 2001) have P–T–t history which requires deep-seated, crustal scale, low-T decollements and thrust ramps which are kinematically required in Alpine settings, typical of E- or NE-directed subduction zones, and absent in the opposite W-directed subduction zones (e.g., Doglioni et al., 2007); 2) the back-arc extension in the Provençal Basin and Valencia Trough obliquely cross-cuts the contractional structures of the Betic cordillera and also occurs in the Alpine foreland (Fig. 6 and 7), indicating its independent origin with respect to the Alpine orogen (Doglioni et al., 1997); 3) when restoring the Corsica–Sardinia micro-continent to its Oligocene pre-rotation position (i.e., close to France and Iberia; Fig. 3–7; Alvarez, 1976; Vigliotti and Kent, 1990; Westphal et al., 1973), the Alpine nappes of Corsica appear as the natural morphologic and structural prolongation of the Western Alps (e.g., Molli, 2008, and references therein); 4) the Alps were double-verging since their early stages, implying a subduction hinge converging with the upper plate: this kinematics occurs primarily on "E"-directed subduction zones (Doglioni et al., 2007); 5) the Alps have an average width of 200–250 km, although they were possibly smaller during the early stage of accretion; this implies that from the front of the Alps to the leading edge of their retro-belt, there should have been a volume and width of the lithosphere affected by the Alpine contraction larger than 100 km (minimum thickness of the lithosphere) which should be now stretched into the Apennine back-arc basin, as in fact is reported (e.g., Kastens et al., 1988; Sartori, 1986).

In order to unravel the geodynamic evolution of W-directed subduction zones, the Atlantic subductions are used as case history. The Atlantic W-directed subduction zones (Barbados and Sandwich arcs) appear to have nucleated along the retro-belt of the former E-directed Cordillera subduction, where oceanic or thinned continental lithosphere was present in the foreland of the E-verging retro-belt (Doglioni et al., 1999a). In fact, those two subductions developed along the front of the E-verging cordillera retro-belt (e.g., Rocky Mountains, Sub Andean thrust belt) only where Atlantic oceanic lithosphere in the foreland to the east of the retro-belt was available. The W-directed subductions did not develop where to the East of the retro-belts of the E-directed subduction there was thick continental lithosphere, like in the western interior of North America or in the western South America. The two Atlantic subduction zone case studies developed during the Cenozoic, and they have arcuate shapes and a length of ~2000 km. This seems to indicate that the development of W-directed subductions is enhanced by the presence of oceanic or thinned continental lithosphere in the foreland of the retro-belt of earlier E-directed subduction zones (Doglioni et al., 1999a).

We propose that similar geodynamic conditions determined the origin of the Cenozoic Apennines and Carpathians subduction zones. The former formed along the retro-belt of the Alps, while the latter developed along the retro-belt of the Dinarides (Figs. 4 and 7), in correspondence with the occurrence of the Ionian oceanic lithosphere and of the Dacian Basin (also known as Carpathian Embayment, Transylvanian Basin or Pieniny–Magura Ocean; Golonka, 2004; Gyorfi et al., 1999; Kovacs et al., 2007; Royden, 1988; Schmid et al., 2008; Fig. 3).

2.5. Evolution of W-directed subduction zones

As their Caribbean and New Scotia equivalents in the Atlantic, the Apennine and Carpathian subductions were soon characterized by trench retreat. The slab-retreat induced: 1) extension in their back-arc realms (Provençal–Algerian–Tyrrhenian basins and Pannonian basin for the Apennines–Maghrebides and Carpathians, respectively), 2) a continuous "eastward" migration of the front of the fold-and-thrust belt (Boccaletti et al., 1990; Patacca and Scandone, 1989; Vai and Martini, 2001), and 3) a continuous "eastward" migration of the related foredeeps (Ricci-Lucchi, 1986). As consequence of the subduction hinge roll-back, the NE and SW portions of the Apennines subduction trench impacted soon the continental lithosphere, respectively the NW margin of Adria and the NW margin of Africa, slowing down the subduction rate. Under these circumstances, the presence of more easily subductable oceanic lithosphere in correspondence with the central sector of the Apennine–Maghrebide subduction system (the Ionian oceanic corridor) forced an increase of curvature of the trench. The Apennine–Maghrebide subduction system was initially NE–SW-oriented, but in Early Oligocene time, the front of the Apennine trench was already convex in shape.

The differential "eastward" retreat was possibly associated with a lateral segmentation of the slab, particularly along lateral variation of lithospheric thickness and composition. The Malta escarpment, for example, separates the Sicilian continental lithosphere from the Ionian oceanic lithosphere. The inherited passive continental margin provided a tear of the subduction, allowing larger retreat in the Ionian side, associated with the magmatism in correspondence of the slab window (e.g., Doglioni et al., 2001; Faccenna et al., 2011; Gvirtzman and Nur, 1999; Schellart, 2010; Tonarini et al., 2001). Another tear in the slab likely occurred along the northern side of the Apulian platform (the SE portion of the larger Adriatic plate), separating an area of slow roll-back and uplift in the foreland to the south (Southern Adriatic Sea and Apulia) from an area of faster roll-back and subsidence in the foreland in the central-northern Adriatic Sea. Also in this case a volcanic activity developed in correspondence of the slab tear with the formation of Mt. Vulture volcano (e.g., Bianchini et al., 2008; D'Orazio et al., 2007; Rosenbaum et al., 2008).

Remnants of the former Alpine orogens were boudinaged and passively incorporated into the internal parts of the Apennine–Maghrebide accretionary wedges (Alvarez et al., 1974). Relics of metamorphic rocks emplaced by Alpine thrusts have been dredged in the Tyrrhenian (Kastens et al., 1988) and are scattered all around the back-arc basins (e.g., the Kabylie in northern Africa, and Calabria and Peloritani Mts. in southern Italy). Similarly, boudinage of the pre-existing Alpine–Dinaric orogens occurred in the Pannonian basin (e.g., the Apuseni Mts. which separate the Pannonian Basin s.s. from the Transylvanian Basin to the East; Schmid et al., 2008). These basins represent the Oligocene–Pleistocene back-arc basin related to the coeval W-directed Carpathians subduction zone, which retreated eastward during the Miocene and Pliocene (Horváth, 1993; Linzer, 1996; Oncescu, 1984; Tomek, 1993; Tomek and Hall, 1993).

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2.6. The formation of the Central–Western Mediterranean Sea

The beginning of the W-directed Apennine–Maghrebide subduction is not known in detail, the various hypotheses ranging from Late Cretaceous (~80 Ma) to Early Oligocene (~33 Ma; Lustrino et al., 2009, and references therein). The fast radial roll-back of the Adriatic slab along the northern sector of the Apennine–Maghrebide subduction system stretching of the upper plate and upwelling of asthenosphere, responsible for the high heat flow values (>100 mW/m^2) measured in the Central Mediterranean (e.g., Zito et al., 2003).

The Central–Western Mediterranean consists of a series of V-shaped sub-basins, developed from Oligocene time onward in the context of back-arc extension contemporaneous to the eastward-to-southward roll-back of the originally W-directed Apennine–Maghrebide subduction zone (e.g., Auzende et al., 1973; Boccaletti and Guazzone, 1974; Burrus, 1989; Carminati et al., 1998a, 1998b, 2010; Gueguen et al., 1998; Malinverno and Ryan, 1986; Mauffret et al., 1995; Rehault et al., 1984, 1985; Scandone, 1980). The discontinuous thinning process increases toward the E, SE and S from a pivotal area located around the present-day Provençal coast (southern France), leading to relevant lateral thickness variations (Banda and Santanach, 1992; Blundell et al., 1992; Calcagnile and Panza, 1980; Fernandez et al., 1995; Panza et al., 2007; Scarascia et al., 1994; Torné et al., 1992). The isolation of ribbons, during the east-to-south migration of the Apennine–Maghrebide thrust front, indicates discontinuous extensional processes in the back-arc area. From the Langhian onward, active extension shifted from West to East of Corsica and Sardinia, leading to the structuration of the Tyrrhenian basin (e.g., Sartori et al., 2001; Trincardi and Zitellini, 1987). The Sardinia–Corsica continental block represents the largest lithospheric ribbon of the Central–Western Mediterranean. The boudinage arrived to complete thinning of the continental lithosphere with the likely formation of new oceanic crust in the Ligurian–Provençal (~20–15 Ma), Algerian (~17–10 Ma), Vavilov (~7–3.5 Ma) and Marsili (~2 Ma–Present) basins (Beccaluva et al., 1990; Galdeano and Rossignol, 1977; Mauffret et al., 1995, 2004; Serri et al., 2001). Only for the Vavilov and Marsili basins a true oceanic nature of the crust has been ascertained.

During back-arc spreading, blocks moved radially, from northeastward to southward, and rotated both clockwise (southern arm) and counter-clockwise (northern arm). About 60° counter-clockwise rotation affected the Sardinia–Corsica continental block (Gattacceca et al., 2007; Montigny et al., 1981; Speranza et al., 2002; Vigliotti and Kent, 1990), while the Balearic promontory rotated ~20° clockwise (Martin, 2006; Parés et al., 1992). The retreat of the slab was accommodated by left-lateral transtension in the southern Tyrrhenian basin and by right-lateral transpression in the north-Africa accretionary wedge. On the other side, the Ligurian–Provençal Basin and northern Tyrrhenian Sea were controlled by a diffuse right-lateral transtension while the frontal Apennine accretionary prism formed in a regime of left-lateral transpression (e.g., Doglioni, 1991).

The recent stages of the evolution of the Central Mediterranean region are complicated by the diachronous contraction in the southern Algerian basin (starting ~8 Ma; Strzerzynsky et al., 2010) and in the southern Tyrrhenian Sea (since ~2 Ma; Billi et al., 2011). In our view, this tectonic inversion can be ascribed to the continuing Africa–Europe convergence, which is also responsible for the deformation of the southern, E–W trending, margin of the belt associated with the Apennine subduction, from Sicily to the Maghrebides. The Sicily Channel was characterized from Pliocene to Present by rifting trending NW–SE (stretching in NE–SW direction) that drove to the formation of the Malta, Pantelleria and Lampedusa grabens. Toward the NW, this rift possibly propagated into the Pliocene Campidano graben in SE Sardinia (Corti et al., 2006). The stretching also affects the Pelagian shelf offshore, eastern Tunisia and the Tarabulus and Jiffara troughs offshore Libya (Capitanio et al., 2011; Finetti, 1984). The rift-related faults were active during the accretion of the Apennines–Maghrebides prism. Therefore, in the Sicily Channel, NW–SE convergence coexists with NE–SW extension. This is an example of two geodynamic settings working together, supporting the possible interpretation that plate boundaries are passive features (Corti et al., 2006).

2.7. The Alboran Sea problem

The Alboran basin and the arcuate belt surrounding it (the Betic–Rif arc) are subject to different interpretations. On one hand the area is conceived as a westerly migrating micro-plate, being the arc the result of this single mechanism (e.g., Frizon de Lamotte et al., 1991; Lonergan and White, 1997; Rosenbaum et al., 2002a). There are GPS and paleomagnetic evidences of active radial motion of the Betic and Riff nappes (Cifelli et al., 2008; Pérouse et al., 2010) and the tomography would also suggest an E-ward dipping slab (Gutscher et al., 2002; Spakman and Wortel, 2004). Indeed there is robust evidence for an “E-ward” directed slab, and the consequent oblique Betics orogen, where nappe stacking occurred since Cretaceous times in a right-lateral transpressional setting. However, there are a number of complications to this model. The following issues should be considered: the Alboran rifting is much younger (Oligocene–Miocene) than the Betic Belt and, even more important, the basin is shaped by NE-trending normal faults which cross-cut obliquely the Betics and continue into their foreland, opening the Valencia trough (Doglioni et al., 1997). Therefore Alboran rift and Betics are unrelated. Moreover, the Betics propagated westward, whereas the Alboran rift and subsidence propagated eastward (Docherty and Banda, 1995). The Alboran Basin is physically continuing into the Valencia and Provençal basins, which are the westernmost and oldest basins of the western Mediterranean Sea. Therefore the Alboran Sea should be considered as an integral part of the western Mediterranean, easterly migrating, back-arc rifting in the hangingwall of the Apennines–Maghrebides subduction zone. Moreover, the assumption that the Betic arc is a single-mechanism feature is debatable because 1) the Betics to the north of the Alboran basin, and the Rif to the south, have very different characters, which recall the asymmetries between Alps and Apennines (e.g., thick versus thin skinned, high versus low topography, deep versus shallow rocks involved, etc.); 2) the foredeep of the Betics (like the Alps) has a shallow foreland regional monocline dip (2–4°) and very slow subsidence rates, whereas the Rif (like the Apennines) has a steeper monocline (6–9°) and 3–4 times faster subsidence rates.

As a consequence, in our model the Betic–Rif arc represents the coalescence of the lateral western termination of two independent, opposed, adjacent and coeval (since at least the Oligocene) subduction zones (Doglioni et al., 1998). In this view, the Alboran Sea is the natural oblique back-arc basin of the Maghrebides. The Apennines–Maghrebides system merges together with the Alps–Betics system also at the northeastern termination of the Apennines. There, the different polarity of the accretionary prisms makes their distinction clearer. Therefore in the westernmost Mediterranean four geodynamic processes coexisted after the Cretaceous counter clock-wise rotation of Iberia, i.e., 1) the E- to SE-ward dipping Betic subduction, 2) the oblique W- to NW-dipping Riff–Maghrebides subduction (as the western termination of the Apennines–Maghrebides subduction), 3) the related Alboran back-arc basin and 4) the slower NNW relative contraction between Africa and Iberia (Europe).

2.8. The magmatic response to kinematic evolution: focus on the Italian area

A review of the complexity and variability of magmatism in the Central–Western Mediterranean is beyond the scope of this work. Interested readers are referred, among others, to the works of Wilson and Bianchini (1999), Lustrino and Wilson (2007); Lustrino

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et al. (2011) and the monographies edited by Beccaluva et al. (2007) and Beltrando et al. (2010). The Italian area will be taken as an example for such complexity.

The complex tectonic evolution of Italy is mirrored by the not yet fully understood origin and evolution of the Cenozoic igneous activity. Despite the wealth of geological, volcanological, petrographic, mineralogical, geochemical, isotopic and geochronological data, indeed, a clear consensus on the geodynamic significance of the igneous activity of Italy is not yet defined. The essential geological–petrological features of Cenozoic Italian magmatic rocks can be resumed as follows:

- The Cenozoic igneous activity developed in Italy intermittently in specific geographic districts and with long quiescence periods (Alagna et al., 2010; Carminati et al., 2010; Conticelli et al., 2010; Lustrino and Wilson, 2007; Lustrino et al., 2011; Peccerillo, 2005; Serri et al., 1993, and references therein). Remnants of Cenozoic igneous rocks crop out along the Alpine Chain in the Piedmont, Lombardy and Veneto regions, and mostly along the Tyrrhenian coast of the Italian peninsula in the Tuscany, Latium and Campania regions. Volcanic rocks are particularly abundant throughout the Island of Sardinia, and along the eastern margin of Sicily. Minor igneous centers are present in the Basilicata and Umbria regions, in the Sicily Channel and in the southern Tyrrhenian Sea.
- Many Italian islands are igneous in compositions [the Tuscan archipelago in Northern Tyrrhenian Sea, the Aeolian Islands in SE Tyrrhenian Sea, Pantelleria and Linosa islands in the Sicily Channel, several islands of the Gulf of Naples (Ischia, Procida, Vivara and Ventotene) and the Island of Ustica NW of Sicily].
- No igneous rocks are present along the Adriatic Sea margin.
- Several non in-situ volcanic rocks are present as volcaniclastic to pyroclastic successions in nearly all the regions of Italy. With few exceptions (e.g., Aveto–Petrignacola Formation, N Apennines), these volcaniclastic to pyroclastic successions are essentially made up by arenitic to pelitic fractions, implying long transport distance from the source areas.
- Three are the peaks of igneous activity in Italy, one around Early Oligocene (~32–28 Ma, essentially along the Alpine Chain), the second during Early–Middle Miocene (~22–18 Ma; essentially in Central–Northern Sardinia) and the third during the Pleistocene (~1–0 Ma, essentially in Latium and Campania and in Eastern Sicily).
- With very rare exceptions (Sulcis, SW Sardinia; Pietre Nere, Apulia; Mt. Queglia, Abruzzi) the oldest igneous products crop out in Central Alps (Adamello Pluton; Fig. 5). Here ~42–27 Myr-old plutonic rocks (mostly tonalites, trondjhemites and granodiorites; Tiepolo et al., 2011) crop out in connection with two large-scale traspressional lineaments, the Giudicarie Lineament to the SE and the Insubric Line to the NW (Fig. 1).
- During Eocene–Oligocene times, the igneous activity of Italy is essentially concentrated along the Alpine Chain, where plutonic (plus volumetrically insignificant volcanic) rocks are aligned along the sinistral cataclastic to mylonitic Insubric Line. The climax of radiogenic age determinations of these plutonic rocks (essentially diorites–tonalites) is concentrated around 32–28 Ma. Minor Eocene–Oligocene basic to acid sodic alkaline volcanic rocks crop out near the contact between the front of the Southern Alps (the retro-belt of the Alps) and the Adriatic foreland in the Veneto region.
- With only one exception (Mt. Queglia lamprophyre, Abruzzi region), during Paleocene–Oligocene no igneous activity is recorded along the present-day Apennines Chain.
- Few Myr before the Alpine Chain magmatism climax, a diffuse igneous activity started in the Island of Sardinia, whose peak of productivity developed essentially during the Aquitanian–Burdigalian (Early Miocene; ~22–18 Ma; Guarino et al., 2011; Lustrino et al., 2009, and references therein). Volumetrically the most abundant products are rhyolites, followed by dacites and andesites, with very rare basaltic and gabbroic products, all showing calcalkaline to high-K calcalkaline affinity.
- The Late Miocene–Pleistocene period is the most interesting and complex from a petrological point of view, because of the presence of a large range of chemical compositions spread over large distances. A Late Miocene–Late Pleistocene volcanic activity developed throughout the Island of Sardinia, in the forming Northern Tyrrhenian Sea and the Western Tuscany margin. Interestingly, the Sardinian products are petrographically, geochemically and isotopically very different from the Tyrrhenian–Tuscany products. Late Miocene–Pleistocene volcanic rocks of Sardinia are essentially basic to intermediate mildly sodic to tholeiitic rocks (hawaiites, mugearites alkali basalts, basaltic andesites) with much rarer evolved compositions (trachytes, phonolites and rhyolites). On the other hand, the Northern Tyrrhenian Sea igneous products are essentially calcalkaline plutonic rocks with granitoid compositions (Elba, Giglio and Montecristo Islands plus Vercelli Seamount) with minor potassic to ultrapotassic volcanic rocks at Capraia Island. Onshore Tuscany rocks are essentially SiO_2-strongly undersaturated to SiO_2-saturated potassic to ultrapotassic volcanic rocks, with minor buried plutonic bodies and crustal anatectic rhyolites (e.g., Alagna et al., 2010; Conticelli et al., 2010; Lustrino et al., 2011; Poli et al., 2003).
- The climax of the volcanic activity of Italy developed during the last Myr essentially along the Tyrrhenian margin of the Italian peninsula, with emplacement of a nearly continuous line of volcanoes (from North to South: Vulsini Mts., Cimini Mts.-Vico, Sabatini Mts., Alban Hills, Ernici Mts., Roccamonfina, Pontine Islands, Phlegrean Fields, Ischia Island and Somma-Vesuvius) from the southern Tuscany to Latium and Campania regions in central-southern Italy (Avanzinelli et al., 2009; Conticelli et al., 2010; Peccerillo, 2005). Essentially SiO_2-saturated potassic to SiO2-undersaturated ultrapotassic compositions are present (shoshonites to leucitites). Rare Early Pliocene calcalkaline to high-K calcalkaline compositions (andesites to rhyolites) are found only as deep drilling cores in the Campania plain and cropping out in the Pontine Islands.
- Volumetrically insignificant – but petrologically very interesting and peculiar – volcanic rocks are present also within the Apennines thrust and belt (the so-called Umbria–Latium ultra-alkaline province; e.g., Lavecchia et al., 2006) with strongly SiO_2-undersaturated ultrapotassic rocks (kamafugites to leucitites), carbonatites and silico-carbonatites compositions. Carbonatites and mildly potassic strongly alkaline compositions (tephrities to phonolites) are present also along the Apennines trust front in the Basilicata region (e.g., D'Orazio et al., 2007; Giannandrea et al., 2004).
- Jumping the Calabria region, characterized by the absence of in-situ Cenozoic igneous rock outcrops, volcanic rocks become abundant in the E–NE sectors of Sicily (Hyblean Mts. and Mt. Etna; e.g., Di Grande et al., 2002; Lustrino and Wilson, 2007; Viccaro and Cristofolini, 2008, and references therein), where strongly to mildly sodic alkaline (nephelinites to alkali basalts) and tholeiitic basalts to basaltic andesites crop out, and in SE-most Tyrrhenian Sea (Aeolian Islands), where arc-tholeiitic, calcalkaline, high-K-calcalkaline, potassic and ultrapotassic compositions are present (e.g., Francalanci et al., 2007; Peccerillo, 2005, and references therein).
- The central and south-eastern Tyrrhenian Sea is punctually characterized by the presence of oceanic basaltic floor and huge volcanoes mostly concentrated in the Vavilov and Marsili sub-basins. Here, and in neighboring areas, volcanic rocks with a wide spectrum of chemical compositions, ranging, in order of abundance, from calcalkaline (essentially andesites and basaltic andesites), Transitional- and Enriched-MORBs (mostly basalts and basaltic

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andesites) and sodic alkaline (essentially alkali basalts and hawaiites) have been dredged and cored (e.g., Trua et al., 2007, and references therein).

- In a completely isolated position along the NW margin of the Island of Sicily is located the Ustica Island, made up of Quaternary volcanic rocks, essentially represented by alkali basalts and mugearites plus rarer more evolved types (trachytes). This island is surrounded by several small seamounts (e.g., Anchise and Prometeo) with very variable compositions ranging from island-arc basalts to mugearites (e.g., Trua et al., 2004, 2007).
- Finally, in the Sicily Channel an important volcanic activity developed during the last Myr producing several seamounts (e.g., Graham and Nameless Banks) and volcanic islands (Pantelleria and Linosa). Here alkali basaltic to rhyolitic volcanic rocks have been sampled, all showing mildly alkaline sodic compositions (e.g., Avanzinelli et al., 2004; Civetta et al., 1998; Ferla and Meli, 2006; Rotolo et al., 2007).

3. How far can magmatism constrain kinematic reconstructions?

Syracuse and Abers (2006) reviewed the location of >800 volcanic arc centers along ~33,000 km-long subducting plates, essentially located along the Pacific ring of fire. These authors investigated the depth of the top of the slab (H) beneath each volcano and found that average H values over 500 km-long arcs range from ~72 to ~173, with a global average of ~105 km. This value agrees with previous estimates of the depth of the top of the slab below front arc volcanoes, hypothesized in the range of ~95–150 km (Gill, 1981; Schmidt and Poli, 1998; Stern, 2002; Tatsumi, 1986; Tatsumi and Eggins, 1995). Kimura et al. (2009, 2010) modeled thermodynamically and geologically the metamorphic and metasomatic processes during subduction of oceanic plates and the interaction with the mantle wedge, reaching similar conclusions. In other words, it is possible to propose a depth of ~110 ± 20 km as the place where partial melting typically occurs in the supra-subduction mantle wedge. The chemical composition of the produced liquid depends on several parameters among which the most important are the nature of the subducted components (pelitic, terrigenous, calcareous or cherty sediments; altered vs. pristine oceanic crust), the physical state of the metasomatizing components derived from the slab (melts, fluids aqueous melts or silicate/carbonatite melts), the temperature and the thickness of the slab, the contribution of the slab portion to the chemistry of the metasomatizing agent compared with the oceanic slab portion, the pressure and temperature of zone refining and chromatographic processes in the mantle wedge, and the original composition of the unmetasomatized mantle wedge.

About forty years of experimental petrology studies indicate that the system peridotite + H_2O stabilizes a wide range of hydrous minerals at depths up to 10 GPa (Fig. 11a). Some of these phases may host high amount of water (up to ~12–13 wt.% H_2O in serpentine and chlorite; Schmidt and Poli, 1998, and references therein) but are not stable at relatively high temperatures typically recorded at the top of the downgoing slab (~900–100 °C; Syracuse et al., 2010, and references therein; Fig. 11a). The most important H_2O carrier at typical temperatures recorded in the mantle wedge is amphibole, with a relatively low water budget (~3 wt.%). In natural cases, if amphibole is stable in a peridotitic mantle, it is difficult to reach water-saturation conditions. This means that if H_2O is released from the downgoing slab it is essentially retained in the amphibole lattice. Water is effectively available (i.e., an H_2O-oversaturated condition is reached) only after the stability limit of the amphibole is reached. This occurs in a depth range between ~2 and ~3 GPa (~70–110 km depth; Fig. 11a, b; Green et al., 2010; Niida and Green, 1999; Wallace and Green, 1991). When H_2O-oversaturated conditions are reached,

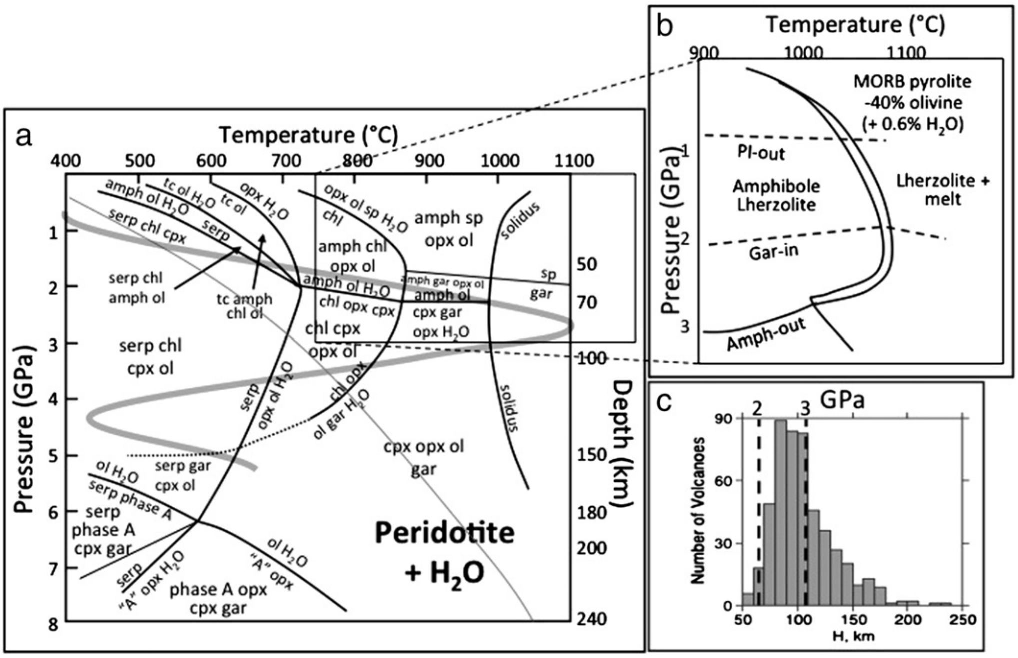

Fig. 11. a) Phase diagram for H_2O-saturated average mantle peridotite (modified after Schmidt and Poli, 1998). *A* = phase A; *amph* = amphibole; *chl* = chlorite; *cpx* = clinopyroxene; *gar* = garnet; *ol* = olivine; *opx* = orthopyroxene; *serp* = serpentine; *sp* = spinel; *tc* = talc. The amphibole stability limit is here shown at a pressure of ~2.2 GPa (corresponding to a depth of ~70 km). The thick gray curvilinear line is a qualitative estimate of the vertical temperature gradient measured in correspondence of the front-arc volcanoes as estimated by Stern (2002). With increasing depth the temperature in the mantle wedge increases up to ~1100 °C. Nearing the top of the cold slab, the temperature of the mantle wedge starts to decrease, reaching very low values (around 400–500 °C) in correspondence of the top of the slab. With further increase of depth, the temperature raises again going through the inner portion of the slab, reaching thermal equilibrium with ambient mantle at higher depths (>15 GPa). b) Stability field of pargasitic amphibole in a model mantle composition (MORB pyrolite) experimentally determined by Niida and Green (1999). This experimental petrology study suggests that the stability for a particular bulk H_2O content is mostly controlled by alkali content of the lherzolite composition. The temperature stability limit of pargasitic amphibole coincides with the water-undersaturated solidus (amphibole-dehydration solidus) at pressures below 3 GPa. c) Histograms of H, distance from trench for front-most arc volcanoes (modified after Syracuse and Abers, 2006). The peak of the number of the volcanoes falls very close to the amphibole stability limits shown in a) and b).

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the solidus of a peridotite drops down and is lowered up to 400–500 °C. The depth of formation of arc magmas estimated by Syracuse and Abers (2006) and previous authors falls just in the limit range of amphibole stability (Fig. 11c). In other words, partial melting would develop essentially when the amphibole is no longer stable in the peridotitic assemblage, i.e., at depths around 2–3 GPa. At relatively high temperatures (i.e., >850 °C), where no other hydrous phases like chlorite are present, the amphibole-out reaction is described by Niida and Green (1999) as: amphibole + orthopyroxene = clinopyroxene + garnet + olivine + H_2O.

The presence of K_2O in the subducting geochemical budget increases the complexity of this simple scheme, becoming other hydrous mineral phase (phlogopitic mica and K-richteritic amphibole) abundant in the K-doped natural and synthetic peridotitic experimental assemblages. The field of stability of these phase largely exceeds that of pargasitic amphibole (~3 GPa), reaching depths as high as 9–10 GPa for phlogopite and up to 15 GPa for K-richterite (e.g., Harlow and Davies, 2004; Konzett et al., 1997; Kushiro et al., 1968; Sudo and Tatsumi, 1990; Trønnes, 2002). For example, the occurrence of crust-derived subduction fluids at 200 km depth has been inferred by Scambelluri et al. (2008).

According to a simple model without K_2O excess, it is possible to hypothesize the paleo-depth of the top of a slab when subduction-related igneous activity is recognized in a given area. This top-of-the-slab depth/arc magmatism correlation effectively works in simple steady-state systems like those occurring in the Pacific ring of fire, but cannot be adopted in much more geologically complex area such as the Central–Western Mediterranean. Here, active arc magmatism is developing in the Aeolian Archipelago where the top of the near-vertical Ionian slab has been seismologically identified at a depth ranging between ~150 and ~250 km (e.g., Chiarabba et al., 2008). In addition, subduction-related magmatism (in the form of calcalkaline andesites) is recorded in the Vavilov seamount (e.g., Trua et al., 2007). The arc magmatism developed here with H parameter (the depth of the top of the slab) in the order of 300–400 km (Chiarabba et al., 2008; Piromallo and Morelli, 2003; Fig. 10).

The rule of having arc magmatism when the slab reaches depths in the order of ~70–110 km is not respected also in many other areas considered in this review. Examples are the Late Eocene and Early Miocene igneous activity in Sardinia, developed when the slab was, respectively, much shallower and much deeper compared with the ~70–120 km depth. Also the Carpathian arc magmatism developed when the Carpathian subduction system was at the first stages of development or very close to the subduction hinge (and, therefore, where the slab was very shallow; Figs. 5 and 7). Another, even more complex, example is the presence of huge amounts of arc volcanic rocks (mostly andesites to dacites) in the Northern Apennine foreland or within the accretionary wedge but very close to the thrust front (Aveto–Petrignacola Formation; Fig. 6; Mattioli et al., in press). In that case subduction-related rocks were produced in a subduction-unrelated tectonic setting (i.e., in the Apennine foreland).

It is stressed, as a note of caution, that the third dimension (i.e., depth) is always the least constrained in past reconstructions. In principle, the above discussed differences between our reconstructions and Syracuse and Aber's (2006) model predictions could be due to errors in our paleogeographic maps and cross-sections. However, our reconstructions were built to account all available stratigraphic and tectonic data. Different geometrical reconstructions would lead to inconsistencies with geological data. We believe that these anomalies with respect to the classical "Pacific-style" arc magmatism are simply related to the existence of ancient modifications recorded in the uppermost mantle. The presence of old suture zones (e.g., the Hercynian Chain in Sardinia or the Dinarides for the Carpathians) or the presence of "fossil" slabs (e.g., the S–SE immerging Alpine Tethys slab beneath the northern Apennines) can be the main responsible for acquiring the "subduction-related" geochemical characteristics (e.g., relative mobility of LILEs, low HFSEs, high LILE/HFSE ratios, high Pb and variable but generally high K content, plus $^{87}Sr/^{86}Sr$ higher than Bulk Silicate Earth estimate, $^{143}Nd/^{144}Nd$ lower than chondritic Uniform Reservoir estimate, high $^{207}Pb/^{204}Pb$ and $\Delta 7/4$ values) of the subduction-related rocks (the blue symbols in the figures and the movie). Similarly, the subduction-related characteristics of the Plio–Pleistocene Tuscan lamproites (Northern Italy) have not been related with coeval subduction but, rather, to ancient modifications of their mantle sources (e.g., Peccerillo, 2005, and references therein).

The Alpine E-directed subduction had effects essentially on the Alpine Chain igneous rocks and on the Aveto–Petrignacola volcaniclastic succession in northern Apennines. Not shown in our reconstruction is also the metasomatism in the lid of the upper plate generated by the fluids released by the Alpine slab. The subduction-related geochemical characteristics of all the Alpine Chain igneous rocks are related to volcanoes (now represented essentially by their plutonic roots) emplaced when subduction had already ceased by several Myr (e.g., Alagna et al., 2010; Lustrino et al., 2011 and references therein). In other words, the "subduction-related" geochemical features of these rocks are not mirrored by a "subduction-related" tectonic setting. It has not yet understood why igneous activity is completely lacking during the long-lasting subduction phase of the Alpine Tethys subduction before the Adria–Europe collision.

The complex tectonic history recorded by the area now occupied by the Mediterranean Sea and neighboring realms had profound effects on the composition of the upper mantle, resulting in depletion processes, related to the formation of normal sea-floor MORBs and back-arc basalts, and enrichment processes, essentially related to the recycling of crustal lithologies along subduction zones. As a consequence, the classical approach, working well in other Earth's more simple geological cases, cannot be adopted in complex systems like the Central–Western Mediterranean. The principle itself of actualism (i.e., invoking the same geological causes to explain a given phenomenology happened in the past) should be re-thought in complex settings such as the Mediterranean. The presence of arc-magmatism in a given time interval cannot be considered a proof for the existence of coeval subduction-related tectonics.

To complicate this scenario is the existence of "anorogenic" or "intraplate" volcanic activity in areas very close to subduction systems or even in hinterland positions, such as the Veneto magmatic Province in Northern Italy and several seamounts in Southern Tyrrhenian Sea. In the first case, a very long-lasting Na-alkaline volcanic activity developed before and during the "subduction-related" igneous activity along the Alpine Chain. The Veneto Volcanic Province rocks developed on the Adriatic retro-foreland, on the Adriatic Plate that was overriding the Alpine Tethys and the Southern European paleo-margin (Figs. 3–7). Also in the Southern Tyrrhenian Sea several seamounts (Magnaghi, Vavilov, Quirra, Prometeo), essentially characterized by Na-alkaline chemical compositions with subduction-unrelated geochemical, mineralogical and petrographic characteristics were active more or less contemporaneously with the other subduction-related volcanic products (Cornacya, Marsili, Anchise seamounts; for a location of these seamounts see Peccerillo, 2005 and Carminati et al., 2010). The presence of active volcanism with completely different chemical composition in a very short distance (e.g., Mt. Etna and Aeolian Archipelago) testifies both how little we have really understood on mantle dynamics, chemical composition and inherited heterogeneities and how complex is the tectonic and geodynamic setting. Only models that mix geological observations of a given area with geochemical data of igneous rocks can be considered sound. Very often geodynamic models (e.g., existence of mantle plumes or existence/absence of subduction processes) have been proposed considering geochemical data only or not considering at all the presence of igneous activity. Only a synergic cooperation of different

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Earth Sciences disciplines can be considered as a true success for the developing of winning models.

4. The Alps–Apennines subduction flip: a close-up

A close-up of the subduction flip at the latitude of Sardinia is shown in Figs. 3–7. In Fig. 5 the first subduction-related magmatic activity (Calabona; 38.28 ± 0.26 Ma, feldspar $^{40}Ar/^{39}Ar$ step heating age; Lustrino et al., 2009) of the embryonic Western Mediterranean occurred. The presence of a coeval dyke swarm in the Malaga district (Southern Spain) with calcalkaline compositions (37.6 ± 0.9 Ma, whole-rock $^{40}Ar/^{39}Ar$ step heating age; Turner et al., 1999) is interpreted by us as an evidence for partial melting processes along the Apennine–Maghrebide NE–SW-oriented subduction system, although, in principle, for the Malaga activity an origin related with the Alps–Betics subduction cannot excluded.

The presence of two Late Eocene subduction-related small volcanic districts suggests a change of the tectonic stresses in the forming Western Mediterranean. This makes Eocene time one of the key periods in the geodynamic reconstruction of the Central Mediterranean is, indeed, the Eocene. During this period, the consumption of oceanic lithosphere along the E- to SE-directed Alpine subduction system had almost completely ceased (Figs. 3 and 4). Following the continental collision of the Southern European paleo-margin with the Mesomediterranean Terrane (see below), a new NW-directed Apennine–Maghrebide subduction system started, as already discussed, in the Central–Western Mediterranean. The Islands of Sardinia and Corsica were part of the European foreland in the Alpine–Betic subduction system (from Early Cretaceous to Middle Miocene). Once this Alpine Tethys oceanic subduction ceased, the two Islands became part of the hinterland of the Apennine–Maghrebide subduction system (from Middle–Late Eocene–Present). During this second phase, Corsica and Sardinia were affected by stretching, with the development of huge volcanic activity during the formation of the Ligurian–Provençal back-arc basin. The origin and evolution of this subduction flip (from roughly SE-directed (Alpine–Betic) subduction to roughly NW-directed (Apennine–Maghrebide) subduction) can only be speculated. Doglioni (1991) and Doglioni et al. (1998) proposed the presence of a continental terrane (defined as Mesomediterranean Terrane) located in between the Ligurian–Piedmontese (western Alpine Tethyan, to not be confused with the much more recent Ligurian–Provençal basin) and of the Ionian oceanic lithosphere as the necessary ingredients for the onset of the flip. Other authors (e.g., Dercourt et al., 1986; Guerrera et al., 1993, 2005; Michard et al., 2002, 2006; Molli, 2008; Belayouni et al., 2010) reached similar conclusions, proposing the existence of a continental terrane separating the two branches of the Alpine Tethys and called it AlKaPeCa Block (where Al = Alboran Terrane, now present in the westernmost Mediterranean Sea; Ka = Kabylies, now present along the northern Algerian margin; Pe = Peloritani Mts., in north-easternmost Sicily and Ca = Calabria, the tip of the Italian boot). Handy et al. (2010) proposed the existence of a short-lived Jurassic–Early Cretaceous micro-plate that comprised both the AlKaPeCa continental fragment and adjacent Liguria oceanic lithosphere and called it Alkapacia. The terrigenous record of the areas surrounding the Mediterranean testifies the occurrence of a micro-continent (Guerrera et al., 1993, 2005) incorporated into the Betics and in the south-western prolongation of the Alps.

This view is forwarded in our model with an important difference with respect of AlKaPeCa: we propose that the present-day Alboran area did not belong to the Mesomediterranean continental slice bounded by two oceanic branches, the Neotethyan Ligurian–Piedmontese Ocean or western Tethyan Ocean branch, sensu Handy et al., 2010) and the Ionian Ocean (or the eastern Tethyan Ocean branch, sensu Handy et al., 2010; Fig. 5). Once the Mesomediterranean Terrane was completely involved in the structuration of the belt of the Alpine–Betics subduction, the continuing convergence of Africa (and Africa-derived Adriatic plate) against Europe was accommodated by a new subduction system with opposite polarity. In this case the eastern branch of the Neotethys (Ionian Basin), started to be subducted towards the NW (Figs. 4 and 5). This is the most critical stage of the entire Central-Western Mediterranean area evolution, marking the transition between Alpine s.s. subduction to Apennine tectonics. To resume, a subduction flip is here imaged, marking the transition between the Alpine–Betic subduction to the Apennine–Maghrebide subduction stage. During the flip, the Alpine slab was truncated by the new Apennines slab, which immediately started to retreat radially. However, owing to the continued convergence, the Alpine collisional system likely continued to accommodate part of the convergence along the northern margin of the Adriatic plate.

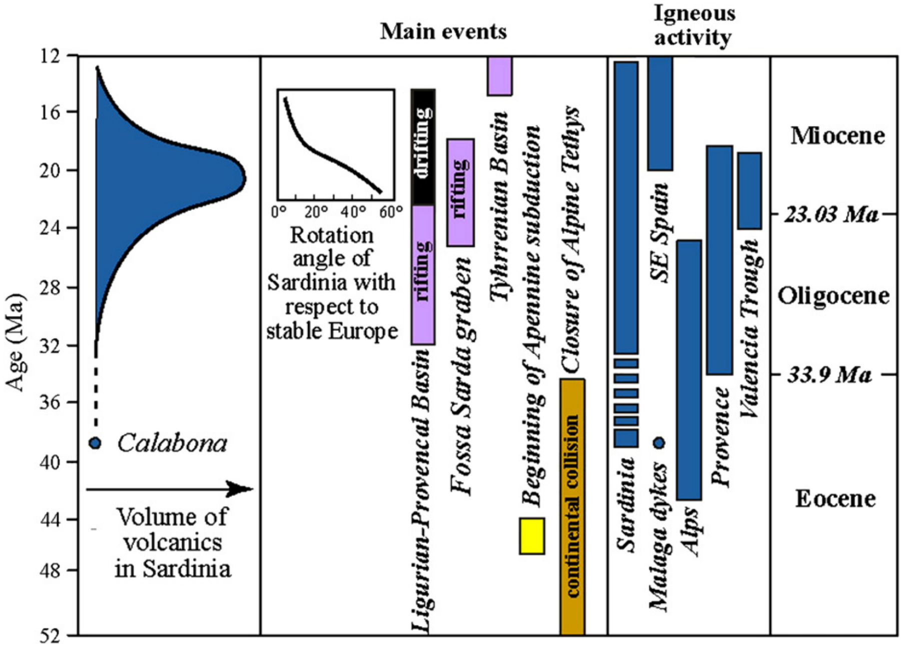

Fig. 12. Chrono-diagram for the 52–12 Ma period, showing the approximate volume of igneous rocks of Sardinia, the angular velocity of rotation for the Sardinia–Corsica block (Gattacceca et al., 2007) and the main tectonic and magmatic events in the Central–Western Mediterranean.

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As written above, the beginning of the Apennine–Maghrebide subduction has been alternatively dated as back as to Late Cretaceous (~80 Ma; Jolivet et al., 1998) or as young as to Early Oligocene (~33 Ma; Gueguen et al., 1998). According to our model, the flip of subduction occurred around Middle Eocene time (~45 Ma), based on the age of the first subduction-related magmatic products cropping out in Sardinia (Lustrino et al., 2009) and southern Spain (Turner et al., 1999) and on fission track analyses on the Tuscan (i.e., Western Adriatic) metamorphic complex in Northern Apennines (Balestrieri et al., 2011). As already discussed, the Mesomediterranean/AlKaPeCa terrane is now dismembered as consequence of the centrifugal forces associated with the opening of the Apennines–Maghrebides back-arc basins (Ligurian–Provençal Basin, Valencia Through, North Algeria Basin, Alboran Sea and Tyrrhenian Sea).

The volume (Fig. 12) and, partially, the composition of the partial melts of the supra-subduction mantle wedge can be used to reinforce our model. As already evidenced (e.g., Lustrino et al., 2004, 2009), the "subduction-related" igneous activity in Sardinia ranged from ~38 to ~13 Ma, but the volume follows a Gaussian distribution, peaking around Aquitanian–Burdigalian time (~22–18 Ma; Guarino et al., 2011, and references therein). During this Early Miocene interval, the Sardinia–Corsica block recorded the fastest counter-clockwise angular rotation (e.g., Speranza et al., 2002, Gattacceca et al., 2007, and references therein) associated with the production of the MgO-rich melts (e.g., Mattioli et al., 2000; Morra et al., 1997). High-MgO igneous melts are very rare in continental settings, because they are denser than typical crustal rocks. These melts can reach the Earth's surface only during major extensional tectonics when pathways connecting Moho to the surface may locally form. During the ~22–18 Myr interval, huge pyroclastic deposits (with an estimated volume of ~1000–2500 km^3, only in the continental sector of the Island; Guarino et al., 2011), covered the central-northern part of the Sardinian Trough causing also strong chemical modification in the marine water composition of the entire Central–Western Mediterranean (e.g., Brandano et al., 2010).

The volume of subduction-related magmatism in Sardinia remained substantially very scarce from ~38 to ~26 Ma (Lustrino et al., 2007, 2009; Fig. 12), at least from what we can observe from the present-day outcrops. We interpret this sequence of events (~38–26 Ma = scarce igneous activity; ~26–22 Ma = increasing igneous activity; ~22–18 Ma = peak of igneous activity; ~18–13 Ma = diminishing and end of igneous activity) as a consequence of the composition and the subducting velocity of the slab. In the first phase the new Apennine subduction system scraped off part of the lowermost continental crust and lithospheric mantle of Sardinia, pushing them to greater depths (Fig. 7). The lower crust of Sardinia (point #5 in Figs. 4 and 5), reworked and possibly hydrated during the Hercynian orogeny (Carmignani et al., 1994), was tectonically forced to follow the Alpine Tethys OCT (Oceanic–Continental Transition) zone. Following this approach, the first igneous phase of the Apennine subduction system in Sardinia is rather related to the recycling and dehydration of ancient (pre-Carboniferous) lithologies (Fig. 5), likely identified in the already subduction-modified Hercynian lower crust and uppermost mantle. Interestingly, close to the Calabona latitude, in NW Sardinia passes the so-called Posada–Asinara Line, a supposed tectonic line that would represent an ancient suture zone, where Gondwana and Laurussia plates collided during Late Paleozoic leading to the formation of the Pangea super-continent (e.g., Cappelli et al., 1992; Lustrino, 2000). The peak of subduction-related magmatic activity at 21 Ma is, in our view, rather related with the dehydration of Ionian oceanic crust at depth of around 120 km (point #3 in Fig. 7).

The geological evolution of Sardinia indicates that the local pre-Apennine subduction lithosphere was already metasomatized during ancient (possibly Hercynian or Caradocian) modifications. Bearing this in mind, the subduction-related geochemical characteristics of the first products (e.g., Calabona micro-diorite) could be, at least in part, related to activation of anciently modified mantle sources, rather than to contemporaneous subduction of the Ionian (or Eastern Tethys) Ocean. The absence of complete geochemical data for the first-phase igneous products (i.e., those generated in the ~38–26 Myr interval) limits the possibility of a more detailed petrological study aiming to validate this model. The only complete whole-rock analysis available (Calabona; Lustrino et al., 2009) indicates some chemical differences between the bulk of the subduction-related igneous rocks of Sardinia (e.g., Conte et al., 2010; Guarino et al., 2011; Lustrino et al., 2004, 2009, and references therein). In particular, the Calabona microdiorite shows roughly similar incompatible element patterns in primitive mantle-normalized diagrams, but shifted towards higher absolute concentrations (Fig. 5 in Lustrino et al., 2009). If these features indicate mantle sources modified by different subducted material or if they are simply related to a different degree of evolution of the partial melts it is an aspect that still remains to be solved. From an isotopic point of view, the recycling of ancient (pre-Carboniferous) material should be easily detectable only assuming a large amount of unconstrained parameters (e.g., the age and lithology of the subducted material plus its elemental content of Sr, Nd and Pb, and the style of inter-elemental fractionation between the Rb–Sr and Sm–Nd pairs and the U–Th–Pb triplet). A Sr–Nd–Pb isotopic study of the Calabona micro-diorite is still missing. For this reason such a study on the Calabona rocks and on the roughly coeval Esterel micro-diorites (S France, ~34 Ma) is currently under progress in order to shed light on the possible causes responsible for the subduction-related geochemical signature of these rocks.

We have speculated that chemical modifications produced by the Hercynian orogeny could have controlled the early stages of magmatism associated with the Apennines subduction. In addition, it can also be speculated that the kinematics (in particular the velocity of retreat) of the newly formed Apennines slab could be controlled by the occurrence of the lithospheric thickening of the Hercynian orogeny. In the Mediterranean region, the plates have been inferred to move westward with respect to the underlying mantle, that conversely moves eastward with respect of the plates (e.g., Doglioni, 1991; Doglioni et al., 2007). The eastward motion of the mantle is predicted to control and force the eastward rollback of the Adriatic and Dinaric slabs, which are encroached and are expected to behave as sails.

The mantle flow is expected to have been deviated downward by the Hercynian lithospheric thickening of Sardinia, as shown in Figs. 3 and 4. In the early stages of the Apennines subduction (Fig. 5) the area of the slab encroached by the deviated mantle flow should have been rather small, thus the rollback velocity was very small. At later stages (Figs. 6 and 7), the back-push associated with the mantle encroachment was likely much more effective and the roll-back velocity increased significantly. As an alternative, if slab dynamics is interpreted to be controlled by the negative buoyancy of slabs due to their denser nature with respect to the adjacent mantle, the increase of roll-back speed could be explained by an increase of the dimension of the slab and consequently by the increase of slab pull forces. However, this second hypothesis is at odds with the geophysical observation that present-day Apennine slab does not show positive density contrasts with respect to the mantle (Brandmayr et al., 2011).

5. Concluding remarks

The kinematic representation of the Central–Western Mediterranean geodynamics presented here is clearly biased by a number of still unconstrained boundary conditions. Regardless the missing accuracy, it shows a number of regional and general indications. The first robust result is that the subduction zones shaping the Cenozoic of the Mediterranean follow geometrically the inherited Mesozoic lateral variations of thickness and composition of the lithosphere.

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Since the wavelength of these lateral variations is relatively short, the Mediterranean basin seems to be more complicated than other areas of the world (e.g., Lustrino et al., 2011, and references therein). However, the subduction zones that characterize this region follow the same rules that have been recognized worldwide, being the geographic polarity one of the primary controlling factors (Carminati et al., 2004, 2010; Doglioni et al., 1999b). In this context, the subduction zones with the subduction hinge moving towards the upper plate (i.e., Alps–Betics and Dinarides–Hellenides) are characterized by Alpine-type belts, i.e., doubly vergent, higher elevation, deeper *decollements*, etc. On the other hand, the subduction zones where the subduction hinge migrates away from the upper plate (i.e., Apennines and Carpathians) have an Apennine-style, single vergence, low elevation, shallower *decollements*, and back-arc basin formation. Moreover, following the example of the Atlantic subduction zones, the Apennines are here interpreted as formed along the retro-belt of the Alps where oceanic or thinned continental lithosphere was occurring.

The movie associated to this article is the main result of the research, where, albeit a number of acknowledged limitations, we present our simplified interpretation of the geodynamic evolution of the Central–Western Mediterranean. It is shown that the evolution of this area is characterized by flips in subduction polarity that led to the development of the Apennines and Carpathian subductions. Geometry, kinematics and magmatism of the Apennine-related subduction flip were strongly influenced by the inherited tectonic history. The complex magmatic evolution of the Mediterranean area cannot easily reconciled with simple magmatological models proposed for the Pacific subductions. This is most probably due to the occurrence of several subduction zones in the area that strongly perturbed the chemical composition of the mantle in the Mediterranean region.

Acknowledgments

Thanks to Marco Cuffaro and Davide Scrocca for several discussions. The revision of D. Frizon de Lamotte (Cercy, France) and of an anonymous reviewer helped to improve the quality of the manuscript. The authors thank also the guest editors G. Molli (Pisa, Italy) and J. Malavieille (Montpellier, France). ML thanks, as usual, Enrica, Bianca and Laura for being so patient also during the writing of this manuscript. This work has been supported by research grants to EC and CD (TopoEurope-Topo4D) and to ML (AST 2008, 2009, Ateneo 2010, 2011 and MIUR-PRIN 2008 research grant 2008HMHYFP_005).

Appendix A. Supplementary data

Supplementary data to this article can be found online at doi:10.1016/j.tecto.2012.01.026.

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Please cite this article as: Carminati, E., et al., Geodynamic evolution of the central and western Mediterranean: Tectonics vs. igneous petrology constraints, Tectonophysics (2012), doi:10.1016/j.tecto.2012.01.026

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Please cite this article as: Carminati, E., et al., Geodynamic evolution of the central and western Mediterranean: Tectonics vs. igneous petrology constraints, Tectonophysics (2012), doi:10.1016/j.tecto.2012.01.026

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Please cite this article as: Carminati, E., et al., Geodynamic evolution of the central and western Mediterranean: Tectonics vs. igneous petrology constraints, Tectonophysics (2012), doi:10.1016/j.tecto.2012.01.026