Recent tectonic reorganization of the Nubia-Eurasia convergent boundary heading for the closure of the western Mediterranean

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Abstract: In the western Mediterranean area, after a long period (late Paleogene-Neogene) of Nubian (W-Africa) northward subduction beneath Eurasia, subduction is almost ceased as well as convergence accommodation in the subduction zone. With the progression of Nubia-Eurasia convergence, a tectonic reorganization is therefore necessary to accommodate future contraction. Previously-published tectonic, seismological, geodetic, tomographic, and seismic reflection data (integrated by some new GPS velocity data) are reviewed to understand the reorganization of the convergent boundary in the western Mediterranean. Between northern Morocco, to the west, and northern Sicily, to the east, contractional deformation has shifted from the former subduction zone to the margins of the two backarc oceanic basins (Algerian-Liguro-Provençal and Tyrrhenian basins) and it is now mainly active in the south-Tyrrhenian (northern Sicily), northern Liguro-Provençal, Algerian, and Alboran (partly) margins. Onset of compression and basin inversion has propagated in a scissor-like manner from the Alboran (c. 8 Ma) to the Tyrrhenian (younger than c. 2 Ma) basins following a similar propagation of the subduction cessation, i.e., older to the west and younger to the east. It follows that basin inversion is rather advanced in the Algerian margin, where a new southward subduction seems to be in its very infant stage, while it has still to properly start in the Tyrrhenian margin, where contraction has resumed at the rear of the fold-thrust belt and may soon invert the Marsili oceanic basin. Part of the contractional deformation may have shifted toward the north in the Liguro-Provençal basin possibly because of its weak rheological properties compared with those ones of the area between Tunisia and Sardinia, where no oceanic crust occurs and seismic deformation is absent or limited. The tectonic reorganization of the Nubia-Eurasia boundary in the study area is still strongly controlled by the inherited tectonic fabric and rheological attributes, which are strongly heterogeneous along the boundary. These features prevent, at present, the development of long and continuous thrust faults. In an extreme and approximate synthesis, the evolution of the western Mediterranean is inferred as being similar to a Wilson Cycle (at a small scale) in the following main steps: (1) northward Nubian subduction with Mediterranean backarc extension (since ~35 Ma); (2) progressive cessation, from west to east, of Nubian main subduction (since ~15 Ma); (3) progressive onset of compression, from west to east, in the former backarc domain and consequent basin inversion (since ~8-10 Ma); (4) possible future subduction of former backarc basins.

Key words: western Mediterranean, convergent boundary, tectonic reorganization, subduction, backarc basin, basin inversion.

Réorganisation tectonique récente du domaine de convergence de plaque Nubie-Eurasie en Méditerranée occidentale.

Résumé: En Méditerranée occidentale, après une longue période (Paléogène supérieur et Néogène) de subduction de la plaque nubienne sous l’Eurasie, la subduction s’arrête et la convergence doit alors être accommodée par d’autres processus géodynamiques impliquant une réorganisation tectonique de la méditerranée occidentale.

Une synthèse des données tectoniques, séismologiques, géodésiques, tomographiques, et de sismique réflexion complétée par de nouvelles mesures géodésiques, nous permet de proposer un modèle de cette réorganisation tectonique intégré à l'échelle de la Méditerranée occidentale. Entre le Nord du Maroc et le Nord de la Sicile, la déformation compressive résultant de la convergence s’est déplacée de la zone de subduction vers les marges des bassins océaniques d’arrière arc, que sont les bassins Algéro-Liguro-Provençal et Tyrrhenien. Les marges Tyrrhénienne (Nord de la Sicile), Liguro Provençale (SE de la France), Algérienne et de l’Alboran sont par ailleurs toujours actives. La compression, ainsi que l’inversion tectonique associée, se sont propagées, du bassin d’Alboran (c. 8 Ma) vers le domaine Tyrrhénien (< c.2Ma), parallèlement à la rupture du slab due à l’arrêt de la subduction. Ensuite l’inversion s’est propagée vers la marge algérienne soumise à une subduction embryonnaire vers le Sud. La compression le long de la marge Tyrrhénienne semble alors accommodée en arrière par une ceinture de plis et chevauchements actifs. Une partie de la compression resultant de la convergence s’est déplacée vers le Nord, sur la marge Liguro-Provençale, du fait de conditions rhéologiques plus favorables que dans le domaine entre la Tunisie et la Sardaigne où il n’y a pas de croûte océanique. Ce domaine est d’ailleurs caractérisé par une très faible activité sismique par rapport aux autres domaines frontaliers des plaques Nubie et Eurasie.

Globalement, la réorganisation géodynamique aux limites de plaques Nubie – Eurasie est fortement contrôlée par l’héritage tectonique et les conditions rhéologiques qui varient notablement, et qui , d’autre part, ne sont pas cylindrique le long de cette zone de frontière. Ces conditions aux limites empêchent le développement de systèmes de chevauchement longs et continus.

En résumé, l’évolution fini- cénozoïque suit un cycle de Wilson avec quatre étapes majeures: 1- une subduction vers le nord de la plaque nubienne qui produit une extension «classique» d’arrière arc depuis 35 Ma; 2- à partir de 15 Ma, un arrêt de la subduction qui se propage de l’Ouest vers l’Est, le long de la frontière de plaque; 3- à environ 8-10 Ma, on assiste à la mise en place d’une déformation compressive, de l’Ouest vers l’Est, en domaine d’arrière arc qui induit une inversion tectonique; 4- enfin, la possibilité de mise en place de nouvelles zones de subduction dans ces zones arrière arc.

Mots clefs: Méditerranée occidentale, frontière de plaque en convergence, réorganisation tectonique, subduction, bassin d’arrière arc, inversion tectonique.

1. INTRODUCTION

Both recent and old suture zones between continental plates usually consist of a complex juxtaposition of highly-deformed tectonic slices deriving from pristine paleogeographic and geodynamic domains as diverse as oceanic basins, sedimentary prisms, subduction complexes, volcanic arcs, seamounts, isolated continental blocks, and backarc basins [e.g., Cloetingh *et al*., 1982; Burg and Chen, 1984; Cloos, 1993; van der Voo *et al*., 1999; Murphy and Yin, 2003]. In several cases, evidence of multiple subduction events either contemporary or succeeding with even opposite polarities are documented or hypothesized in tectonic sutures [e.g., Teng *et al*., 2000; Michard *et al*., 2002; Tyson *et al*., 2002; White *et al*., 2002; Kapp *et al*., 2003; Doglioni *et al.*, 2007; Regard *et al*., 2008]. The juxtaposition of heterogeneous tectonic slices shows that the process occurring along destructive tectonic boundaries and anticipating the continental collision or the final suture often does not simply consist of an oceanic lithosphere steadily and cylindrically subducting beneath a continental one. Large and deep backarc oceanic basins may, for instance, develop during long periods of oceanic subduction. With the progression of plate convergence on the way to tectonic suture, these extensional structures may be inverted and even closed through the subduction of the oceanic lithosphere in the former backarc domains [e.g., Escayola *et al*., 2007; Vignaroli, Faccenna *et al*., 2008].

One ideal region to study the tectonic processes anticipating the final suture between continents is the Mediterranean region (Fig. 1), where the Nubian (i.e., the African plate to the west of the eastern rift) and Eurasian plates meet and interact within the general framework of active convergence [Dercourt *et al*., 1986; Dewey *et al*., 1989; Ricou *et al*., 1986; Doglioni, 1991; Cloetingh and Kooi, 1992; Jolivet and Faccenna, 2000; Jolivet *et al*., 2003; Rosenbaum *et al*., 2002; Stich *et al*., 2003; Allen *et al*., 2004; Nocquet and Calais, 2004; Nocquet *et al*., 2006; Fernández-Ibáñez *et al*., 2007; Mauffret, 2007; Serpelloni *et al*., 2007; D'Agostino *et al*., 2008; Zitellini *et al*., 2009; Carminati *et al*., 2010; Lustrino *et al*., 2011]. The western Mediterranean (i.e., from Calabria, southern Italy, to Algeria and Gibraltar), in particular, is a tectonically complex area where two small oceanic basins (Tyrrhenian and Liguro-Provençal backarc basins) occur along the Nubia-Eurasia convergent margin and are separated by the Corsica-Sardinia rigid continental block [Doglioni *et al*., 1997; Gueguen *et al*., 1997; Jolivet and Faccenna, 2000; Faccenna *et al*., 2001, 2002; Mascle *et al*., 2004; Jolivet *et al*., 2008]. Such a complex setting imposes a segmentation and reorganization of the convergent boundary as well as a complex (i.e., non-cylindrical) distribution of the zones accommodating the contractional deformation. The particular and favorable tectonic setting of the western Mediterranean may allow researchers to capture some snapshots of the very early processes of tectonic inversion eventually leading to the closure of backarc basins and final suture between continents. Assuming that the closure of the western Mediterranean basin will occur, at least in part, through subduction of the two backarc oceanic basins (Tyrrhenian and Liguro-Provençal), studying and monitoring them may provide insights into subduction inception [e.g., Cloetingh *et al*., 1982; Souriau, 1984; Toth and Gurnis, 1998; Faccenna *et al*., 1999; House *et al*., 2002; Strzerzynski *et al*., 2010].

In this paper, we address the present and recent tectonic reorganization along the Nubia-Eurasia convergent boundary in the central-western Mediterranean. To do so, we review previously-published datasets (integrated by some new GPS velocity data) constraining the present tectonic architecture and regime. We analyze, in particular, structural, geodetic, seismological, seismic reflection, and tomographic data. Based on these data, we discuss possible tectonic models explaining the reported datasets and eventually provide some implications for future tectonic scenarios of the studied tectonic boundary. As most strong earthquakes and related destructive tsunamis are generated at convergent margins and subduction zones [McCaffrey, 2008], this paper bears implications into earthquake and tsunami hazard in the Mediterranean even though this subject is beyond our aims.

2. EVOLUTION OF THE WESTERN MEDITERRANEAN SUBDUCTION ZONE

The late Paleogene-Neogene tectonic evolution of the western Mediterranean has been drawn in a series of studies based on geological, geophysical, and geochemical constraints supported, in some cases, by numerical and physical modeling [Malinverno and Ryan, 1986; Kastens *et al*., 1988; Dewey *et al*., 1989; Doglioni, 1991; Patacca *et al*., 1992; Carminati *et al*., 1998; Carminati, Wortel, Spakman, and Sabadini 1998; Jolivet and Faccenna, 2000; Faccenna *et al*., 2004, 2005; Rosenbaum and Lister, 2004; Pepe *et al*., 2005, 2010; Minelli and Faccenna, 2010]. Most of these studies use a slab retreat model [e.g., Malinverno and Ryan, 1986; Royden, 1993; Lallemand *et al*., 2008; Jolivet *et al*., 2009] to explain the present tectonic setting of the study area that includes fold-thrust belts and backarc basins along the Nubia-Eurasia convergent boundary, which is also known as the western Mediterranean subduction zone [Faccenna *et al*., 2004]. Three main tectonic domains can be recognized along this zone: the inner and outer orogenic domains, and the extensional backarc domain (Fig. 1).

The inner orogenic domain mostly consists of stacked slices of continental crystalline Variscan basement with its metasedimentary cover. Ophiolitic units are exposed in Calabria and norrthern Apennines, whereas the continental crystalline and metasedimentary formations are extensively exposed in the northern Apennines, Calabria, Kabylie, and Rif-Betic belts [Dercourt *et al*., 1986; Monié *et al*., 1988; Platt *et al*., 1998; Caby *et al*., 2001; Vignaroli *et al*., 2009; Rossetti *et al*., 2001, 2004, 2010].

The outer orogenic domain is overthrust by the inner domain and is mostly constituted by a stack of Meso-Cenozoic thrust sheets of sedimentary rocks deriving from the deformation of the African, Iberian, and Adriatic continental margins. This domain mostly developed during Neogene time with lateral heterogeneities in the geometry of structures and direction of tectonic transport (i.e., see the Calabria and Gibraltar arcs; Roure *et al*., [1991]; Outtami *et al*., [1995]; Gomez *et al*., [1998]; Frizon de Lamotte *et al*., [2000]; Thomas *et al*., [2010]).

The extensional backarc domain includes two main triangular backarc basins partly floored by oceanic crust: the Tyrrhenian basin, to the east, and the larger and older Liguro-Provençal basin, to the west, which elongates toward Gibraltar, where the Algerian and Alboran basins occur [Kastens *et al*., 1988; Kastens and Mascle, 1990; Sartori, 1990; Gorini *et al*., 1994; Faccenna *et al*., 1997]. The Tyrrhenian and Liguro-Provençal basins are separated by the Corsica-Sardinia continental block [Mascle *et al*., 2004].

The above-mentioned structures are obviously correlated with the Moho depth in the western Mediterranean area [Tesauro *et al*., 2008; Fig. 1b]. The Moho is shallow in the Tyrrhenian and Liguro-Provençal basins up to a minimum of only about 5 km and tends to deepen toward the surrounding continental areas and fold-thrust belts (Alps, Apennines, Betics, Pyrenees, and Rif-Atlas) up to a maximum of about 55 km in the northeastern Alps. In several cases, the backarc basins developed in crustal areas thickened during previous phases of mountain belt building. Afterward, backarc extension led to the exhumation of high-pressure metamorphic and plutonic rocks that are well exposed, for instance, in Corsica, and peri-Tyrrhenian, Alboran, and Aegean areas [Jolivet *et al*., 1990; Gautier and Brun, 1994; Platt *et al*., 1998; Jolivet and Faccenna, 2000; Rossetti *et al*., 2001, 2004].

Fig. 2 shows a schematic evolutionary model of the western Mediterranean subduction zone since about 35 Ma [Rehault *et al*., 1984; Carminati *et al*., 1998; Frizon de Lamotte *et al*., 2000; Faccenna *et al*., 2004]. This model is useful to understand the formation of backarc basins and their recent inversion.

At about 35 Ma, the subduction zone ran almost continuously from Gibraltar to Liguria (i.e., in a present-day geographic perspective) possibly along a NE-SW direction. This boundary was consuming a Mesozoic ocean that subducted toward the northwest with a decreasing velocity toward the west (Gibraltar), where continental collision had almost occurred.

At about 30 Ma, backarc extension started in the northern part of the Liguro-Provençal basin and then propagated toward the south in the Valencian and Alboran basin. This process was accompanied by widespread calc-alkaline volcanism in regions such as Sardinia, Valencian, and Alboran, and was induced by the slab retrograde fast motion.

Between about 23 and 15 Ma the opening of the Liguro-Provençal basin was mostly accomplished, but extension was still active in the Alboran basin at about 15 Ma. At that time, in fact, the western portion of the slab broke off and the subduction zone became discontinuous in the Mellilla-Oranie region, where NE-SW left-lateral tectonics occurred (Table 1). This process left the western portion of the slab free to rapidly retreat and drive backarc extension in the Alboran region. Also at this stage, the entire subduction zone was characterized by the presence of calc-alkaline volcanism, whose melt was possibly generated through the crustal contamination of a mantle metasomatized melt [e.g., Fourcade *et al*., 2001; Coulon *et al*., 2002].

Afterward, extension shifted in the Algerian basin and then in the Tyrrhenian region, where synrift sedimentation started at about 10-12 Ma [Kastens and Mascle, 1990], and basaltic volcanism occurred at 4-5 Ma in the Vavilov basin and at 2 Ma in the Marsili basin [Sartori, 1990; Nicolosi *et al*., 2006]. Also in this case, episodes of intense backarc extension are possibly related with episodes of lateral slab tearing (Table 1) and associated subduction zone segmentation that facilitated the retrograde motion of a progressively narrower slab in the eastern sector of the subduction zone. In particular, the Nefza and Mogodos, Tunisia, (Na)-alkaline basalts (circa 8-6 Ma) are interpreted as due to the lateral segmentation of the slab and consequent mantle return flows around the ruptured slab, which was then free to retreat toward the southeast in the Tyrrhenian region. Analogously, the sodic-alkaline magmatism of Ustica and Prometeo in the south-western Tyrrhenian probably marks a further episode of slab breakoff and enhanced retreat of the subduction zone during Messinian-Pliocene time (circa 4-5 Ma) [Faccenna *et al*., 2005].

In Fig. 2, backarc compression and inversion start, from the west, prior than 5 Ma and then propagate toward the east [Wortel and Spakman, 2000; Faccenna *et al*., 2004] (Table 1). This evolution is the main subject of this paper and it is addressed in the following sections.

3. FROM SUBDUCTION TO BACKARC COMPRESSION AND INVERSION

In the western Mediterranean area, seismological data (see next section) and other geophysical and geological evidence show that some strands of the backarc basin margins have recently undergone and are presently undergoing compressional tectonics [e.g., Serpelloni *et al.*, 2007]. This tectonic regime, in particular, is active, from west to east, along the east-Alboran, Algerian, and south-Tyrrhenian margins. Seismicity between Tunisia, northwestern Sicily, and Sardinia is weaker and less frequent than off Algeria and northern Sicily, whereas on the opposite side of the western Mediterranean (toward the north), in the northern Liguro-Provençal margin, in Provence (France), and in Liguria (Italy), a seismicity slightly stronger than that of the area between Tunisia and Sardinia occurs. Also a portion of southern Spain (the northeastern margin of the Alboran basin) is undergoing contraction as attested by compressional earthquakes, geological-geophysical evidence, and GPS data [Fernádez-Ibáñez *et al.*, 2007; Serpelloni *et al.*, 2007].

Alboran margin - The present tectonic setting of the Alboran Sea area (Fig. 1) is the result of a complex orogenic evolution occurred mostly during Cenozoic time within the framework of Eurasian-Nubian plate convergence [Michard *et al.*, 2002; Platt *et al.*, 2003; Duggen *et al.*, 2004, 2005 Rossetti *et al.*, 2010]. The area presently includes the Rif and Betic fold-thrust belts in northern Morocco and southern Spain, respectively, and the Gibraltar Arc that connects these structures in the Atlantic area (Gulf of Cadiz), where an eastward narrow subducting slab is supposed to be still active and seismogenic [Gutscher *et al.*, 2002, 2009]. The inner (Mediterranean) sector of the curved fold-thrust belt includes the Alboran narrow basin that is articulated in sub-basins and troughs, and is characterized by a thinned transitional crust (circa 10-12 km in the eastern sector) compared with the surrounding belt (up to circa 30-35 km) [Torne *et al.*, 2000]. Extension in the Alboran basin developed mostly during Burdigalian-Langhian times [Bourgois *et al.*, 1992; Mauffret *et al.*, 1992] possibly because of slab rollback [Royden, 1993; Lonergan and White, 1997; Faccenna et al., 2004].

Compression resumed in the Alboran area since about 8 Ma (Table 1), producing reverse and strike-slip faulting and related folding, which have involved, since then, the entire basin (i.e., diffuse deformation) and not only its margins [Bourgois *et al*., 1992; Campos *et al*., 1992; Comas *et al*., 1992, 1999; Mauffret *et al*., 1992; Morel and Meghraoui, 1996]. NNW-SSE compression resumed at the end of Tortonian time (c. 8 Ma) also along the northern margin of the Alboran Basin (e.g., Granada basin, southern Spain) after a period, during Miocene time, of extension and exhumation of metamorphic complexes [Martínez- Martínez *et al*., 2002; Rodríguez-Fernández and Sanz de Galdeano, 2006]. Toward the west, off southwestern Portugal, the Gorringe Bank developed by NW-verging subcrustal thrusting at c. 8 Ma or slightly earlier (10.5 Ma) [Jiménez-Munt *et al*., 2010].

The present tectonic activity in the Alboran and surrounding areas is proved by instrumental and historical earthquakes [Buforn *et al*., 1995, 2004; Morel and Meghraoui, 1996; López Casado *et al*., 2001; Gràcia *et al*., 2006]. Instrumental earthquakes are predominantly low magnitude (M < 5), but M > 5 earthquakes occurred both in historical and instrumental times. For instance, the Al Hoceima area (located on the Moroccan coast in front of the central Alboran Sea) experienced several destructive earthquakes. Significant events or swarms dated 1522, 1624, 1791, and 1800-1802 have been reported by El Mrabet [2005]. Afterward, a M 5.9 earthquake was generated in this area by a left-lateral strike-slip buried fault [Calvert *et al*., 1997], and, eventually, the strong (M 6.3) Al Hoceima earthquake occurred on 24 February 2004 generated by a NNE-striking, right-lateral, strike-slip fault [Stich *et al*., 2005]. Earthquakes in the Alboran basin occurred in response to a complex stress regime resulting from the general NW-SE Eurasia-Nubia compression combined with local stress sources, which induce anticlockwise rotation of the maximum compression up to almost 80° [Fernández-Ibáñez *et al*., 2007]. The resulting tectonic regime is, on average, compressive in the eastern portion of the Alboran basin and off Gibraltar in the Atlantic Ocean, and transtensional in the central-western portion of the Alboran basin [Stich *et al*., 2006; Serpelloni *et al*., 2007].

Algerian margin - The Algerian basin is located to the east of the Alboran basin, between the Sardinia and Balearic blocks (Fig. 1). The crust thins from about 7 km in the Alboran basin to about 4 km in the Algerian basin and changes its nature from transitional to oceanic [Comas *et al*., 1997; Catalano *et al*., 2000; Tesauro *et al*., 2008].

Northern Algeria includes the Atlas-Tell fold-thrust belt, which developed during Cenozoic times in response to the collision between the Kabylian block (European affinity) and the Nubian plate. Along this boundary, the Tethyan oceanic lithosphere subducted toward the north since Eocene or Oligocene times [Rosenbaum *et al*., 2002] with a significant dextral oblique component [Saadallah *et al*., 1996]. Because of slab rollback [Jolivet and Faccenna, 2000], the Algerian basin formed in the backarc area [Roca *et al*., 2004] with stretching starting possibly since late Oligocene-early Miocene times [Dewey *et al*., 1989; Rosenbaum and Lister, 2004]. Ductile extension dated at c. 25 Ma in the Grand Kabylie [Monié *et al*., 1984; Saadallah and Caby, 1996] supports the age of stretching onset. The opening of the Algerian basin involved the formation of oceanic crust as recently shown by wide-angle seismic surveys [Pesquer *et al*., 2008]. Magnetic anomalies of the Algerian basin suggest compartmentalization of this basin and related strike-slip faulting during its growth [Maillard and Mauffret 1999; Schettino and Turco, 2006].

With continental collision and subsequent slab breakoff (Table 1), spreading of the Algerian basin ceased during Burdigalian-Langhian time (c. 15 Ma) as also supported by the chemical change of volcanism along the Algerian margin [Maury *et al*., 2000]. Although extensional displacements may have persisted until even 6 Ma, main extension in the Algerian basin endured until about 8 Ma when the Tyrrhenian basin formation had already started [Mauffret *et al*., 2004]. Compressional inversion of the Algerian basin started afterward during late Miocene time (c. 5-7 Ma) and then progressed during Pliocene-Quaternary times [Auzende *et al*., 1975; Meghraoui *et al*., 1986; Déverchère *et al*., 2005; Domzig *et al*., 2006; Mauffret, 2007] when extension and oceanic crust emplacement were active toward the east in the Tyrrhenian basin (Table 1). In particular, N-dipping S-verging reverse structures developed or were reactivated in the onshore portion of the Algerian margin, whereas S-dipping N-verging faults developed mainly as newly-originated structures at the transition between continental and oceanic domains off Algeria [Déverchère *et al*., 2005; Domzig *et al*., 2006; Mauffret, 2007; Yelles *et al*., 2009; Kherroubi *et al*., 2009], thus suggesting a southward subduction inception of the Algerian basin [Strzerzynski *et al*., 2010]. The active margin off Algeria is, however, still poorly organized (laterally) and consists of several fault strands, which are unconnected or connected through perpendicular or oblique tear faults. This pattern is at least in part inherited from pre-existing structures [Strzerzynski *et al*., 2010]. A similar pattern, although with opposite vergence and dip of reverse faults, and a younger age, is observed along the south-Tyrrhenian margin [Pepe *et al*., 2005; Billi, Presti *et al*., 2007].

Unlike the Alboran basin, compressional structures and inversion tectonics of the Algerian margin are rather concentrated as also shown by reflection seismics. Figs. 3(a) and 3(b) show two high-resolution seismic profiles acquired across the Algerian margin [Déverchère *et al*., 2005; Domzig *et al*., 2006; Kherroubi *et al*., 2009], displaying some evidence or clues of basin margin inversion. Fig. 3(a), in particular, shows a profile from the eastern Algeria (Annaba area), running from the continental margin to the deep basin along a track perpendicular to the main structures of the area. The profile interpretation includes S-dipping, N-verging reverse faults at the foot of the continental platform. Although no historical earthquakes are reported for this area, these fault segments could have been responsible for large past events. Moreover, the profile displays a well stratified Plio-Quaternary unit on top of Messinian unit (UE) folded in asymmetrical anticlines with gently-dipping, relatively-flat, long, landward backlimbs, and steeper and shorter seaward limbs. The upper part of the Plio-Quaternary unit shows growth strata onlapping the landward limb of folds, thus depicting an active or recently-active fault-related fold system [Kherroubi *et al*., 2009].

Fig. 3(b) shows a high-resolution seismic profile acquired during the Maradja Cruise (2003) across the lower slope and deep basin off Algiers [Déverchère *et al*., 2005; Domzig *et al*., 2006]. The profile shows a series of N-verging fault-propagation folds characterized by slope breaks and curved scarps. In the profile, the presence of a wedged piggyback basin with active growth strata developed above a thrust ramp rooted below the Messinian salt-layer is evident. The tilting of strata in the basin begun during Pliocene time and is still active. Below the basinal deposits, typical salt deformations are present (see S in Fig 3b).

The compressional tectonics presently acting along the Algerian margin is also proved by strong compressional earthquakes recorded in recent (instrumental) and historical times. These earthquakes are consistent with other evidence such as GPS data (see next sections and Serpelloni *et al.*, 2007). Based on historical, instrumental, and tectonic evidence, the maximum potential magnitude attributed to earthquakes in the Algerian margin is c. 7.3 [Rothé, 1950; Benouar, 2004; Strzerzynski *et al.*, 2010]. Three recent moderate-to-strong earthquakes, i.e., the 1980, M 7.1 El Asnam, the 1989, M 5.7 Tipaza, and the 2003, M 6.9 Boumerdès earthquakes, are representative of the present compressional tectonics of the Algerian margin. The Boumerdès earthquake occurred on a 55 km long, ENE-striking, S-dipping, reverse fault along the Algerian coast and caused a 0.7 m coastal uplift. The aftershock sequence occurred on reverse but also strike-slip faults [Meghraoui *et al.*, 2004; Ayadi *et al.*, 2008, 2010; Déverchère *et al.*, 2010]. The Tipaza earthquake occurred on an ENE-striking, S-dipping, reverse, onshore fault located near the Algerian coast [Meghraoui, 1991; Bounif *et al.*, 2003]. In contrast, the strongest known earthquake of the Algerian margin, i.e., the 1980, M 7.1 El Asnam earthquake, was generated by a NW-dipping, NE-striking reverse fault located near El Asnam (Chlef), central-western Algeria [Philip and Meghraoui, 1983].

Liguro-Provençal margin - The Liguro-Provençal (Fig. 1) is a segmented oceanic basin generated mostly during Miocene (i.e., older than the Tyrrhenian basin) in the backarc region of the southeastward retreating western Mediterranean subduction zone (Fig. 2), at the rear of the counter-clockwise rotating Corsica-Sardinia microplate [Gueguen *et al.*, 1998; Rollet *et al.*, 2002; Speranza *et al.*, 2002; Gattacceca *et al.*, 2007; Bache *et al.*, 2010]. Extension and connected rifting in the Liguro-Provençal basin started at about 25-30 Ma. Tectonics and related sedimentation study of the

Oligocene Provence basins provides evidence for a complex Oligocene extension history due to the special situation between the western European rift and the Liguro-Provençal basin. Indeed, extensional tectonics during Oligocene time up to the Chattian is related to the western European rifting, whereas normal faulting that begun in the Late Chattian is influenced by the Liguro-Provençal basin opening [Hippolyte *et al*., 1993]. The 17.5-Ma-old (Ar-Ar age by Aguilar *et al.*, [1996]) Beaulieu basalts (exposed about 40 km to the north of Marseille), which contain numerous peridotite xenoliths, are contemporaneous with the last increments of the Oligocene to late Early Miocene rifting episode that affected Provence [e.g., Hippolyte *et al*., 1993] and that subsequently led to the formation of the Ligurian-Provençal oceanic basin [e.g., Cheval *et al.*, 1989]. During the Miocene, the western Alps contractional phases and structuration were still ongoing. In particular, the southwestward Castellane and southward Nice compressional arcs developed during Miocene and Mio-Pliocene times, respectively. In Provence, the compressional tectonics and the related south-verging thrusting seemingly propagated southward from the Miocene to the Present. The 1909 Mw 6 earthquake represents the last reactivation of a southern Provence thrust [Trevaresse fault, e.g., Lacassin *et al*., 2002; Chardon and Bellier, 2003; Chardon *et al*. 2005].

Oceanic crust emplacement in the Liguro-Provençal basin occurred between about 21 and 16 Ma [Le Pichon *et al*., 1971; Gueguen *et al*., 1998; Speranza *et al*., 2002]. Related lithosphere thinning was as large as c. 50 km relative to the stable European lithosphere, which makes this basin the site in the western Mediterranean with the highest acclivity of the Moho topography and with an analogous topographic acclivity, i.e., from about 3000 m a.s.l. on the Argentera Massif to about 2500 m u.s.l. in the oceanic floor of the Ligurian Sea [Chamot-Rooke *et al*., 1999; Rollet *et al*., 2002; Tesauro *et al*., 2008].

A weak but frequent seismicity has been recorded in instrumental time along the Liguro-Provençal margin and in southeastern France. Historical database as well as paleoseismological studies provide evidence for earthquakes with Mw 6.0-6.5 [Ferrari, 1991; Sébrier *et al*., 1997; Larroque *et al*., 2001; Baroux *et al*., 2003; Nguyen et al., 2005; Chardon *et al*., 2005; Cushing *et al*., 2008; and SISFRANCE, e.g., Lambert *et al*., 1996]. This seismic activity results from a N- to NNW-trending compression as shown by focal mechanism solutions [Baroux *et al*., 2001; Cushing *et al*., 2008, Larroque *et al*., 2009], geodetic measurements [Calais *et al*., 2002], and tectonic field evidence [Hippolyte and Dumont, 2000; Champion *et al*., 2002; Dutour *et al*., 2002; Lacassin *et al*., 2002; Chardon and Bellier, 2003; Guignard *et al*., 2005; Chardon *et al*., 2005].

Unambiguous evidence of basin inversion along the Liguro-Provençal basin margin is still not available in the geological literature. One of the most suitable evidence for basin inversion is the MA31 seismic reflection profile (Fig. 3c). This is a multichannel profile acquired in 1995 by Ifremer (Malis Cruise) in the Ligurian basin [Bigot-Cormier *et al*., 2004]. The profile extends from the thinned continental margin of southeastern France (Nice) to the deep portion of the Liguro-Provençal basin. A set of NE-dipping reflections (see the dashed red line in Fig. 3c) are interpreted as a late Pliocene-Quaternary blind thrust inverting the margin of the Liguro-Provençal basin since about 3.5 Ma [Bigot-Cormier *et al*., 2004] (Table 1). This interpretation is also based on further evidence including a set of offshore seismic reflection profiles showing vertical deformation and southward tilting of Pliocene-Quaternary strata [Bigot-Cormier *et al*., 2004], and fission track thermochronology data suggesting a general uplift at ~3.5 Ma of the Argentera Massif [Bigot-Cormier *et al*., 2000]. Moreover, further N-dipping reverse faults are signaled onshore to the east of Provence (at Capo Mele in Liguria), where these faults involve Messinian and lower Pliocene sediments [Réhault, 1981; see also Sanchez *et al*., 2010]. At the regional scale, the above-mentioned Pliocene compressional deformation of the Liguro-Provençal margin is interpreted as a foreland thrust propagation (i.e., the Nice arc) and Alpine front migration toward the south [Bigot-Cormier *et al*., 2004].

Toward the north, in the western Alpine domain, the stress regime is complicated by gravitational body forces connected with the high topography and thickened crust that produce forces in competition with the horizontal boundary forces, resulting in a general orogen-perpendicular extension of the western Alps to the north of the Argentera Massif [Sue *et al*., 1999; Delacou *et al*., 2008].

South-Tyrrhenian margin - In southern Italy, the Neogene complex convergence (and associated subduction) between Eurasia and Nubia resulted in the NW-trending Apennine and W-trending Maghrebian fold-thrust belts in peninsular Italy and Sicily, respectively [Malinverno and Ryan, 1986; Dewey *et al*., 1989; Patacca *et al*., 1992]. The two belts are connected through the Calabrian arc [Minelli and Faccenna, 2010], below which a narrow remnant of the former subducting slab is still active, but close to cessation [Piromallo and Morelli, 2003; Mattei *et al*., 2007; Neri *et al*., 2009]. Recent marine surveys have shown that the Calabrian outer accretionary wedge is still active in the Ionian offshore [Polonia *et al*., 2008; see also Minelli and Faccenna, 2010]. In Sicily and southern Tyrrhenian, the Maghrebian fold-thrust belt includes two main units made up of several thrust sheets mostly accreted toward the south over the Nubian foreland, locally named Hyblean foreland. The two orogenic units are the innermost and structurally highest Calabrian unit (mainly crystalline basement rocks) and the outermost and lowest Sicilian unit (mainly basinal sequences with Tethyan ocean affinity) [Vignaroli, Rossetti *et al*., 2008; Corrado *et al*., 2009; Ghisetti *et al*., 2009]. The youngest and outermost (southernmost) thrust sheet is the curved Gela Nappe, which is mostly buried beneath the foredeep infilling and the Mediterranean Sea (Sicily Channel). The youngest phases of inner or basal contraction and displacements are dated back to the end of early Pleistocene time [Lickorish *et al*., 1999; Ghisetti *et al*., 2009]. This age is consistent with the hypothesized end of subduction beneath Sicily and slab breakoff (Table 1), which is presumably and roughly dated with the onset of Etna’s volcanism (circa 0.5 Ma), whose origin is possibly connected with important discontinuities (slab windows) through the subducting slab [Gvirtzman and Nur, 1999; Doglioni *et al*., 2001; Faccenna *et al*., 2005, 2011]. It should be also noted that contraction in the Sicilian Maghrebides may still be weakly active. For instance, the

1968 M 6.4 Belice earthquake (central-western Sicily) has been interpreted as a compressive event within the orogenic wedge [Monaco *et al.*, 1996].

Contraction in Sicily is, at present, mainly accommodated at the rear of the fold-thrust belt in the southern Tyrrhenian (i.e., southern margin of the Tyrrhenian basin) where a series of contractional earthquakes recorded during the last decades define a W-trending seismic belt [Goes *et al.*, 2004; Pondrelli, Piromallo, and Serpelloni, 2004]. In particular, the epicentral and hypocentral analysis of single seismic sequences pointed out that the seismically active contractional structures are high-angle, N-dipping, poorly-connected, short, reverse faults (10-20 km in length; Billi, Presti *et al.*, 2007). The steep attitude of these faults is explained by invoking the reactivation of inner thrusts that were progressively steepened by the growth, at their footwall, of external thrusts during the orogenic wedge forward accretion (i.e., piggy-back sequence). Toward the east, the compressional seismic belt is delimited by the seismically-active Tindari Fault, to the east of which, both earthquakes and GPS data provide evidence for an ongoing extensional tectonics possibly connected with the residual subduction beneath the Calabrian arc and related backarc extension [Hollenstein *et al.*, 2003; D'Agostino and Selvaggi, 2004; Pondrelli, Piromallo, and Serpelloni, 2004; Govers and Wortel, 2005; Billi *et al.*, 2006]. The age for the onset of the ongoing contraction in the south-Tyrrhenian margin is unknown, but the cessation of volcanism at Ustica (i.e., a volcanic island located along the south-Tyrrhenian contractional belt) during middle-late Pleistocene time may be connected with the onset of contractional tectonics in this area. This age corresponds with or is a little younger than the cessation of contractional displacements along the outermost Gela Nappe in southern Sicily [Lickorish *et al.*, 1999; Ghisetti *et al.*, 2009]. It should also be considered that, in the south-Tyrrhenian region, compressional events older than the one that possibly started since middle-late Pleistocene time are documented [e.g. Ghisetti, 1979; Pepe *et al.*, 2000, 2005]. These events may be interpreted as prior rejuvenation phases of the inner orogenic wedge to reestablish the taper subcriticality during the progressive continental collision in this sector of the Mediterranean.

The high-penetration multichannel seismic reflection profile CROP M6A [Scrocca *et al.*, 2003; Pepe *et al.*, 2005], NNE-SSW oriented, is located across the continental margin of northern Sicily (southern Tyrrhenian Sea), perpendicular to the normal listric faults that bound the Cefalù basin (Fig. 4). The tectonic evolution of this margin is very complex due to the compressional-to-transpressional and extensional deformation phases that took place from the early Miocene to recent times. Moving from north to south, the CROP M6A profile shows the overthrusting of the KCU (crystalline rocks of the Calabrian unit) on the African margin (SMU, Sicilian unit) occurred in Oligocene-early Miocene time along the south-verging Drepano thrust system. In between the Sicilian unit, a southeast-vergent tectonic stack occurs consisting of Meso-Cenozoic basin and platform carbonate rocks and Miocene flysch of African pertinence. Following the development of the northern Maghrebian belt [Pepe *et al.*, 2000, 2005 and references therein], the inner northern portion of this belt, corresponding to the present-day north-Sicilian margin, was affected by the onset of back-arc extensional tectonics. As an example, the Cefalù basin, recognizable at the southern end of the CROP M6A profile (Fig. 4), developed on top of the accretionary complex since late Tortonian-early Messinian time. Unpublished seismic reflection profiles document the presence of a widespread tectonic reactivation or positive inversion of previously generated fault systems since late Pliocene and throughout Quaternary times, as documented by growth strata, tilted onlaps, and anomalous thickness of the Plio-Pleistocene units associated to several structural highs [Scrocca *et al.*, 2006].

4. EARTHQUAKES

To define the main seismically-active sectors of the western Nubia-Eurasia convergent margin, in Fig. 5(a) we show the map of crustal seismicity (epicenters of earthquakes with depth ≤ 35 km and magnitude ≥ 4.0) recorded between 1962 and 2009. Earthquakes are mainly located along the northern African margin (northern Morocco, Alboran Sea, and Algeria), in southern

Spain, in Sicily and southern Tyrrhenian Sea, and along the Apennine fold-thrust belt. Seismicity becomes sparser and prevalently of small-to-moderate magnitude in Tunisia and Sicily Channel, thus interrupting the continuity of the seismic belt running from the north-African margin to the southern Tyrrhenian region. Sectors characterized by an almost absent seismicity are the Balearic Basin and the central Tyrrhenian Sea. The Liguro-Provençal Basin is characterized by small-magnitude earthquakes [Eva *et al*., 2001; Larroque *et al*., 2009] that are not shown in Fig. 5(a) due to the magnitude threshold ($M \geq 4.0$). In the Ligurian section (east), indeed, the seismic record is significantly richer in $M \geq 4.0$ earthquakes than the Provençal section (Fig. 5a). In the Italian peninsula, crustal seismicity developed along a continuous belt including the Apennines, Calabrian Arc, and south-Tyrrhenian margin. The seismic regime, however, changes radically from the Apennines and Calabrian Arc (mainly extensional earthquakes; Chiarabba *et al*., 2005) to the south-Tyrrhenian margin (mainly compressional earthquakes to the west of the Calabrian subduction zone; Pondrelli, Piromallo, and Serpelloni, 2004; Billi *et al*., 2006, 2007) (Fig. 6).

Intermediate and deep seismicity (depth > 35 km) is mapped in Fig. 5(b). Subcrustal earthquakes are concentrated in two main regions, which are known for active or recently active subduction processes, namely the Calabrian and Gibraltar arcs [Faccenna *et al*., 2004]. To the northwest of the Calabrian Arc, in particular, intermediate and deep earthquakes are clustered and aligned along a narrow (less than 200 km) and steep (~70°) Wadati-Benioff zone striking NE-SW and dipping toward northwest down to 500 km of depth [Piromallo and Morelli, 2003; Neri *et al*., 2009]. The subcrustal earthquakes of the Gibraltar area are interpreted as related to a relic litospheric slab dipping toward the east in the mantle beneath southern Iberia [Calvert *et al*., 2000; Gutscher *et al*., 2002; Faccenna *et al*., 2004].

Fig. 5(c) shows epicentral locations collected within the framework of the EUROSEISMOS Project (http://storing.ingv.it/es_web/) for the 1900-1961 period and provides a general overview of the seismicity during the last century before the advent of modern seismic networks. Fig. 5(c) confirms the seismic activity, rather energetic in some instances, of southern Iberia, Algeria, and

Italy. No earthquakes are reported for northern Morocco, Tunisia, and southern Tyrrhenian, but this evidence is possibly due to the poor coverage of seismic networks [Buforn *et al.*, 1988; Giardini *et al.*, 2002]. In contrast to what observed in Fig. 5(a), during 1900-1961, M $\geq$ 4 earthquakes were recorded also in the Liguro-Provençal area, southern France (Fig. 5c).

Fig. 6 shows the focal mechanisms of the earthquakes with magnitude $\geq$ 4.5 available for the study region since 1976. The chosen magnitude threshold allows us to obtain insights into regional scale geodynamic processes rather than local ones. Data of Fig. 6(a) are taken from the Harvard Centroid-Moment Tensors (CMT) catalog (http://www.globalcmt.org/CMTsearch.html) and provide robust, stable, and reliable seismic source mechanisms based on the fitting of long period seismic waveforms recorded at the global scale [Dziewonski *et al.*, 1981, 2000]. The CMT data have been integrated with data from the European-Mediterranean Regional Centroid Moment Tensor (RCMT) catalog (http://www.bo.ingv.it/RCMT/) for the 1997-2004 period (Fig. 6b). The RCMT catalog is based on the fitting of surface waves with intermediate and long period recorded at regional distance and its quality evaluation processes ensure high-reliability of data [Pondrelli *et al.*, 2002, 2004, 2006, 2007]. The CMT catalog provides moment tensors for earthquakes with M > 5, whereas the RCMT procedure involves data from earthquakes with magnitude as small as 4.2. Different colors for focal mechanisms (Fig. 6) indicate different types of mechanisms according to the Zoback's classification adopted for the World Stress Map (http://dc-app3-14.gfz-potsdam.de/; Zoback, 1992). Figs. 6(a) and 6(b) show that reverse faulting is the main style of seismic deformation along the Nubia-Eurasia margin between Gibraltar and Sicily. The focal mechanisms available for the southern Tyrrhenian area indicate reverse displacements as the main or solely seismic mechanism active in this area, whereas, moving toward the Algerian margin, earthquakes with thrust mechanisms are associated to several transpressional and strike-slip earthquakes, which become dominant in the Betics, northern Morocco, and Alboran Sea [Lammali *et al.*, 1997; Bezzeghoud and Buforn, 1999; Henares *et al.*, 2003; Vannucci *et al.*, 2004; Billi, Presti *et al.*, 2007]. The above-depicted focal features are synthesized in the polar plots of P- and T-axes selected by source areas (Fig. 6c). Seismic activity in the south Tyrrhenian belt occurred in response to a NNW-SSE oriented compressive stress, which is also predominant in northern Algeria together with some evidence of WSW-ENE extension. Toward the west (Betics, northern Morocco, and Alboran Sea), in contrast, a WSW-ENE extension becomes predominant. Evidence of seismic reverse faulting is also present in Tunisia (E-W compression), off eastern Sardinia (E-W compression), and southeastern France (N-S compression). To the east of the study area, focal mechanisms and plots of P- and T-axes (Fig. 6) from the Apennines and Adriatic Sea show the post-orogenic extension active in the Apennines in response to a regional NE-SW extension [Montone *et al*., 2004; Pondrelli *et al*., 2006), and the compressional seismic displacements active mainly in the eastern side of the Adriatic block (in response to a NNE-SSW regional compression) but also in the mid-Adriatic area and Gargano promontory, where compressional-to-transpressional earthquakes are also recorded [Montone *et al*., 2004; Billi *et al*., 2007].

5. KINEMATICS FROM GPS DATA

The precise measurement of plates and microplates kinematics through the use of modern GPS networks has significantly influenced most recently proposed interpretations concerning the tectonics and geodynamics of the Nubia-Eurasia plate boundary in the Mediterranean area. Although the number of GPS sites in the western Mediterranean region has remained quite limited until a few years ago, the number of available GPS networks has significantly increased in the last five years in most European countries, particularly in Italy, thanks to the development of new networks devoted to both geophysical and topographical issues. Unfortunately, the same progress is not true for the northern African countries, where the number of GPS data and measurements is still sparse or absent.

Although the number of GPS networks specifically designed for geophysical purposes is still limited compared with the number of networks developed for topographic goals, the optimal combination of all available data has recently provided an increasing number of details concerning the kinematics of plates and microplates in the central and western Mediterranean [Hollenstein *et al*., 2003; D'Agostino and Selvaggi, 2004; Serpelloni *et al*., 2005; Stich *et al*., 2006; Serpelloni *et al*., 2007; D'Agostino *et al*., 2008].

Here we present a horizontal velocity field at the scale of the western Mediterranean obtained from the combination of published and original GPS velocities (Fig. 7). We used the GAMIT/GLOBK [Herring *et al*., 2006] software to process data from continuously operating GPS (CGPS) networks in Italy and surrounding regions, following standard procedures for regional networks [Serpelloni *et al*., 2006, 2007]. By analyzing position time series originally defined in the IGS realization of the ITRF05 reference frame [Altamimi *et al*., 2007], we estimated velocities together with seasonal (i.e., annual and semiannual) signals and offsets due to instrumental changes. We used high quality CGPS sites in central Europe (characterized by longer time-series and low position scatters) to define a fixed Eurasian reference frame (located at Longitude -98.85±0.24°E, Latitude 54.74±0.30°N and with rotation rate of 0.257±0.001°/My).

In Fig. 7(a), we present only horizontal velocities obtained from the time-series modeling of high quality CGPS networks (i.e., originally developed for geophysical or geodetic studies), of which the backbone network is represented by the INGV-RING stations [Avallone *et al*., 2010], integrated by other regional geodetic networks, including the ASI, EUREF and FredNet (see Serpelloni *et al*., [2006]; and Avallone *et al*., [2010] for more details). Velocity uncertainties were estimated adopting a white+flicker noise model [Williams *et al*., 2004]. We used, however, only CGPS stations presenting more than three years of measurements. It is worth noting that the distribution of CGPS sites in the western Mediterranean basin is largely heterogeneous, making the analysis of crustal strain-rates at the plate boundary scale quite challenging. To improve the spatial resolution of our velocity field, for northern Africa and southern Iberia, we used also previously-published GPS velocities [Stich *et al*., 2006; Serpelloni *et al*., 2007; Tahayt *et al*., 2008; Peñaa *et al*., 2010], which are rigorously aligned to our Eurasia-fixed reference frame. To compute the 6-

parameters (3 rotations and 3 translations) Helmert transformation and align the published velocity fields to our Eurasian-fixed frame realization, we used GPS sites that are common to our solution and to the published ones (mostly belonging to the EUREF of IGS networks).

Fig. 7(a) shows the horizontal velocities given with respect to Eurasia, together with plate motion vectors predicted by the estimated Nubia-Eurasia relative rotation pole, at points in northern Africa for which one can assume a purely rigid behavior. Clearly, the number of available GPS velocity vectors in Italy is significantly larger than the remaining area, so only long wavelength features of the crustal strain rate field can be reasonably investigated at the western-Mediterranean scale.

We used the approach described in Shen *et al*. [1996], which accounts for velocity uncertainties, network geometry, and inter-station distances, to estimate the velocity gradient field. Horizontal strain-rate tensors at points of a regular grid (0.25° x 0.25° spacing), extending between longitude 8.0E°-14.5E° and latitude 42.0N°-46.0N°, are estimated from the velocity data through weighted least squares. Velocities are re-weighted by a Gaussian function $\exp(-\Delta R2/D2)$, where ΔR is the distance between a geodetic station and the grid point being evaluated and D is a smoothing distance that is optimally determined in this algorithm (between a priori defined lower and upper bound) through balancing the trade-off between the formal strain rate uncertainty estimate and the total weight assigned to the data [Shen *et al*., 2007]. In this way, measurements made closer to a grid point contribute more to the strain estimate at that point, and the smoothing is applied according to the station distribution and density. The re-weighting determines the degree of smoothing around a given spot and the uncertainties of the strain estimates, while the optimally determined D value can be considered as an indicator of how “locally” or “regionally” determined is the strain-rate tensor inverted at each grid point. Given the largely heterogeneous distribution of GPS stations in the investigated area, we aim at illuminating the long-wavelength features of the velocity gradient field using a starting D value of 80 km (i.e., features of wave-length lower than 80 km are filtered out by spatial smoothing), obtaining values ranging between 80 and 400 km, and an average value of 142 km.

Fig. 7(a) shows the horizontal velocities (with 95% uncertainties) obtained from the combination of original data (in the central Mediterranean) and published velocities (in the western Mediterranean), together with the velocities interpolated over the regular grid, as obtained from our least-squares estimates of the velocity gradient field. Fig. 7(b) shows the horizontal strain-rate field, i.e., the maximum and minimum eigenvectors of the strain-rate tensors.

The most evident kinematic features is the motion toward the northwest of the Nubian plate, with respect to Eurasia, and the progressive clockwise rotation of the velocity vectors toward the central Mediterranean, i.e., from northwestward in the western Mediterranean (Morocco) to northeastward in the central Mediterranean (Calabria). Further “local” deviations from this kinematic pattern are observed in northern Morocco and Alboran basin, and also in northeastern Sicily and northern Calabria, where fast deformation rates occur. In particular, if SW-NE oriented shortening is observed along the Moroccan Rif, extension, mainly E-W oriented, characterizes the western Alboran basin and southern Iberia. In northern Algeria and Tunisia, the lack of a good coverage of GPS sites prevents any detailed estimate of the contemporary strain-rate field. However, a few sites along the coast of northern Algeria suggest that this segment of the Nubia-Eurasia plate boundary accommodates about 2-4 mm/yr of SE-NW convergence [Serpelloni *et al.*, 2007]. Moving toward the east, while N-S shortening characterizes the Sardinia Channel, a few available sites make this features purely interpolated and representative of broad geodynamic features. Fast N-S shortening characterizes the southern Tyrrhenian basin, where the fastest deformation rates are observed in the central Aeolian area (southeastern Tyrrhenian Sea). In Sicily, the northward velocities suggest a SW-NE extension between mainland Sicily and the Nubian plate. This extension, which is of the order of 1.5÷2 mm/yr, is likely accommodated across the Pantelleria Rift system in the Sicily Channel. Large extensional deformation rates are observed in northeastern Sicily and southern Calabria, where a sudden change in the velocity trends, which change from northward to northeastward, are accommodated mainly across the Messina straits and also across the Cefalù-Etna seismic belt in Sicily [Pondrelli, Piromallo, and Serpelloni, 2004; Billi *et al*., 2010].

Along the Italian peninsula, the velocity field is mainly characterized by two distinct trends. Sites located on the Tyrrhenian side of the Apennine chain move toward the northwest, whereas sites located on the Adriatic side of the chain move toward the north-northeast. This differential motion results in the SW-NE oriented extensional deformation that characterizes the Apennines chain [Montone *et al*., 2004].

Across the Liguro-Provençal basin, GPS data show no significant deformation rates. In particular, no active extension is observed between the Corsica-Sardinia block and the coasts of southern France and Spain. Only a very limited, but statistically not significant, NW-SE shortening is observed north of Barcelona.

6. DISCUSSION

We cannot predict the future evolution and final setting of what will be the Nubia-Eurasia suture in the western Mediterranean with the progression of plate convergence. A glance at the present setting of this area (Fig. 1), in fact, points out the strong heterogeneity and non-cylindricity of physiography and tectonic structures along this boundary, whose evolution will be therefore highly non-cylindrical as it has been so far since at least Paleogene time [Faccenna *et al*., 2004]. We can provide, however, some significant insights into the recent evolution of this boundary to attempt understanding what will be its progression in the near future.

The spatio-temporal evolution of the studied segment of the Nubia-Eurasia boundary indicates that its past evolution has been obviously influenced by the Nubian subduction (Fig. 2). To understand a possible future progression of this boundary it is therefore necessary to know the present state of the Nubian subduction beneath Eurasia. To do so, we reconsidered a previously published tomographic model of the western Mediterranean [Piromallo and Morelli, 2003].

Fig. 8 shows five horizontal layers (150-to-500 km deep) extracted from the tomographic model PM0.5, whose thorough description is provided by Piromallo and Morelli [2003] and Faccenna *et al*., [2004]. The model, which encompasses the upper mantle beneath the western Mediterranean region, is obtained by inversion of regional and teleseismic P wave residuals from the International Seismological Centre Bulletin. In the upper layers, we observe a high seismic velocity (i.e., positive) anomaly running discontinuously from the northern Apennine, turning around the Calabrian Arc, and heading toward the Gibraltar Arc. We interpret this high velocity volume as the cold lithosphere sunk (subducted) in the mantle along the convergent boundary between Nubia and Eurasia in the western Mediterranean [Wortel and Spakman, 2000; Faccenna *et al*., 2004]. At 150 km depth (Fig. 8), the high-velocity anomaly is located beneath the northern-central Apennines, Calabria, northern Algeria, and the Gibraltar Arc. Evident lateral interruptions of this structure occur in three regions, namely beneath the southern Apennines, the Sicily Channel and the Oranie-Melilla region, where low velocity anomalies occur [Faccenna *et al*., 2004]. The first two gaps (Apennines and Sicily Channel) close at larger depths. At 250 km depth, the Calabria high-velocity anomaly merges, toward the north-east, with the Apennines and, toward the west, with the Algerian anomalies (see the 250 km layer in Fig. 8). The westernmost interruption (Oranie-Melilla), conversely, persists down to about 400-450 km depth (see deep layers in Fig. 8). The high-velocity anomaly belt detected by tomography (Fig. 8) is therefore laterally fragmented beneath the southern flank of the Gibraltar and Calabrian arcs, where deep slab breaks line up with the main discontinuities in the geological trends, i.e., the Sicily Channel and the Oranie-Melilla region. At about 500 km depth (Fig. 8), the high-velocity anomaly spreads horizontally over the whole western Mediterranean area, whereas no coherent trace of high-velocity anomaly is present below 670 km depth. This evidence suggests that the material pertaining to different subduction zones is ponding at the upper/lower mantle discontinuity [Wortel and Spakman, 2000; Piromallo *et al*., 2001; Piromallo and Faccenna, 2004].

The tomographic model (Figs. 8 and 9) shows that the Nubian subduction beneath Eurasia in the western Mediterranean is, after millions of years of efficiency and accommodation of plate convergence, now largely discontinuous. Several studies agree that subduction is here progressively decaying, thus becoming poorly- or non-operative [e.g., Wortel and Spakman, 2000; Faccenna *et al*., 2004; Neri *et al*., 2009]. As plate convergence is nonetheless still operative at rates between about 1 and 5 mm/y [Nocquet and Calais, 2004; Serpelloni *et al*., 2007], a new tectonic reorganization is presently in progress in the western Mediterranean to accommodate such a convergence. The original and previously-published evidence presented in this paper help us understanding such a tectonic reorganization.

Although rather diffuse, the contractional deformation in the western Mediterranean (as inferred from seismicity, Figs. 5 and 6) tends to concentrate at the margins of the oceanic backarc basins (i.e., Algeria and southern Spain, Liguria-Provence, and southern Tyrrhenian), where the rheological contrast between adjacent continental and oceanic domains as well as other factors such as the geometric and thermomechanical properties of the transitional crust controls and favors the localization of contractional deformation [e.g., Béthoux *et al*., 2008]. The Alboran basin, in contrast, where no oceanic crust occurs (at least in the western section), has undergone a diffuse inversion tectonics since about 8 Ma. Hypothesizing a future subduction of this basin seems, therefore, irrational. Concerning the two oceanic backarc basins, whether the final suture between Nubia and Eurasia will be reached through their closure and subduction is highly debatable, among other reasons, because of the young age (i.e., temperature and density poorly appropriate for subduction) and reduced dimensions of these basins [Erickson and Arkani-Hamed, 1993; Cloos, 1993]. A thermomechanical modeling of these basins to infer their potential aptitude to being subducted is beyond our scopes. As, however, some evidence suggests subduction inception across segments of the basin margins (e.g., off Algeria; Strzerzynski *et al*., [2010]), we assume subduction will be the process toward which plate tectonics is heading to in the western Mediterranean, regardless of whether a complete subduction of the Algerian-Liguro-Provençal and Tyrrhenian oceanic basins will ever be accomplished. Upon this assumption, the geological and geophysical evidence presented in this paper provides insights into the subduction inception, which is, in general, a process still poorly known for the substantial absence of instructive instances on the Earth [Cloething *et al*., 1982, 1990; Shemenda, 1992; Toth and Gurnis, 1998; Faccenna *et al*., 1999; House *et al*., 2002].

Backarc basin inversion in the western Mediterranean started at about 8 Ma from the west (Alboran and Algerian margins) following a substantial cessation of subduction (Table 1), and then propagated toward the east up to the present time, also in this case as the consequence of a substantial cessation of subduction beneath Sicily (Faccenna *et al*., 2004; Fig. 9). This spatio-temporal migration (from west to east since about 8 Ma) allows us to provide insights into the process of subduction inception at different temporal stages (i.e., disregarding tectonic differences between inverted margins). Starting from the youngest margin (Tyrrhenian), where pre-existing and presumably-weak reverse faults occur, these faults are possibly reactivated before inversion and then subduction of the Tyrrhenian margin will start. The seismic activity in the south-Tyrrhenian margin occurs, in fact, along inner S-verging reverse faults (i.e., vergence opposite to that expected for the hypothesized subduction of the Tyrrhenian basin) of the Maghrebian belt, which is therefore being rejuvenated at its rear [Billi, Presti *et al*., 2007]. Also in the Algerian margin, where basin inversion (*sensu stricto*) and possibly subduction inception have already started with newly generated reverse faults verging toward the north, reverse faults verging toward the south and possibly inherited from the Paleogene-Neogene contractional phases are still active with associated seismic release [e.g., Mauffret, 2007]. It follows that subduction inception (i.e., the formation of a new plate boundary) is possibly an energetically-consuming process [Toth and Gurnis, 1998], which is substantially anticipated by less consuming processes such as the reactivation of inner thrusts up to the ultimate locking of the orogenic wedge and subsequent transfer of contraction to the backarc basin margin heading for a subduction inception. A viable model of subduction inception (supported by the diachronous age of backarc compression onset) may be the lateral propagation, in a scissor-like fashion, of the new plate boundary where subduction is progressively initiated (Fig. 10). Moreover, our GPS data show that strain rates across the studied boundary are laterally very heterogeneous. In particular, contractional strain rates computed for the south-Tyrrhenian region are significantly larger than those for the Algerian margin. This evidence is difficult to interpret. One explanation may simply be connected with the poor coverage of GPS stations in northern Africa. An alternative explanation may be that contraction accommodation through reactivation of pre-existing reverse faults (south-Tyrrhenian; Pepe *et al*., [2005]; Billi, Presti *et al*., [2007]) is more efficient than subduction inception and basin inversion along newly-generated reverse faults (Algeria; Strzerzynski *et al*., [2010]). The heterogeneous strain rates and GPS velocities implies that, probably, the Sicilian domain is moving independently by the Nubian domain (Algeria) and that a transcurrent (right-lateral)-to-extensional decoupling zone in the Sicily Channel area should enable such differential movements [Serpelloni *et al*., 2007].

Assuming that, from west to east, since about 8 Ma, the Nubia-Eurasia convergence has been mainly accommodated through backarc basin inversion (and perhaps subduction inception off Algeria), we expect the following horizontal displacements (i.e., heaves normal to the new plate boundary), which are calculated from the displacement trajectories of Africa with respect to stable Eurasia [Dewey *et al*., 1989; Faccenna *et al*., 2004]: c. 25 km across the Alboran basin since about 8 Ma and c. 15 km across the Algerian basin since about 5 Ma. As above pointed out, the contractional displacement in the Alboran basin is non-localized, whereas the one off Algeria is rather localized and may have partly contributed to subduction inception. From the above-estimated displacements, we obtain displacement rates of about 3.1-3.2 mm/y, which are consistent with the present GPS velocities of Nubia (Fig. 7). For the south-Tyrrhenian margin, the expected horizontal contractional displacement since 0.5 Ma has already been assessed as about 2.5 km [Billi, Presti *et al*., 2007] by assuming the present GPS velocity of Nubia with respect to fixed Eurasia (c. 5 mm/y for the south-Tyrrhenian-Sicilian region) as constant during the last 500 ky and the motion of Nubia during this period as entirely accommodated across the south-Tyrrhenian deformation zone.

The Liguro-Provençal northern margin is different from the Algerian and Tyrrhenian margins for several reasons. The Liguro-Provençal is, geodynamically, an oceanic backarc basin developed at the rear of the south- and southeast-verging Apennine-Maghrebian belt and anticlockwise rotating Corsica-Sardinia block, but, at the same time, it can be considered as the foreland of the south-verging southwestern Alpine front (i.e., the Nice arc). Therefore, unlike the active tectonics of the south-Tyrrhenian margin, the ongoing compression along the Liguro-Provençal margin can be considered as a resumption of the Alpine thrust propagation toward the foreland and not a rejuvenation at the rear of the belt (either Alpine or Apennine) as it happens in the Tyrrhenian and, partly, in the Algerian margins. The reason why the Liguro-Provençal basin is presently undergoing compression (as evidenced by a seismicity significantly larger than that in the surrounding areas) is still matter of debate [Béthoux *et al*., 1992; Sue *et al*., 1999; Larroque *et al*., 2009], but the hypothesis of a thermomechanical weakness of this region is a viable one [Béthoux *et al*., 2008]. Assuming this hypothesis as true, part of the contractional deformation would have been transferred from the Sicily Channel (i.e., the area between northern Tunisia and northern Sicily where seismicity is weaker than that in the adjacent segments of the Nubia-Eurasia boundary), where no thinned oceanic crust occurs, to the north in the weaker oceanic domain of the Liguro-Provençal basin [Billi, Presti *et al*., 2007; Serpelloni *et al*., 2007; Larroque *et al*., 2009; Billi *et al*., 2010]. Due to the largely inhomogeneous distribution of GPS stations along the studied boundary and hence to the choice of analyzing only the most significant longer wavelengths strain-rate features, our GPS data for the Liguro-Provençal margin show no significant deformation rates (Fig. 7). A local geodetic study, however, pointed out a weak (c. 1 mm/y) contraction in the Provençal region [Calais *et al*., 2002; Nocquet and Calais, 2004] consistently with the recorded compressional earthquakes [Baroux *et al*., 2001; Larroque *et al*., 2009] and with a recent seismological analysis indicating a strain rate of c. 3 x 10^{-9} y^{-1} for the Ligurian Sea (i.e., note that not all deformation is seismic; Barani *et al*. 2010). It should also be considered that the onset of basin inversion (c. 3.5 Ma) in the Liguro-Provençal margin is younger than the same type of tectonics in the Algerian margin (c. 6-7 Ma), but older than the compression resumption in the south-Tyrrhenian margin (younger than c. 2 Ma). This evidence suggests that basin inversion in the Liguro-Provençal margin is in the wake of the spatio-temporal tectonic reorganization of the western Mediterranean after the substantial cessation of Nubian subduction and, therefore, is part of this new reorganization heading for the inversion and closure of the western Mediterranean.

Alternative, viable interpretations for the contractional tectonics in the northern Liguro-Provençal margin involve isostasy/buoyancy forces rather than Nubia-Eurasia plate tectonic collision, and anticlockwise rotation of the Apulian plate [Sue *et al*., 1999; Sue and Tricart, 2003; D'Agostino *et al*., 2008; Delacou *et al*., 2008]. We propend for our model (i.e., accommodation of Nubia-Eurasia convergence also in the Liguro-Provençal basin) on the basis of the spatio-temporal evolution of the compressional wave from west to east (i.e., basin inversion; Fig. 9) and on the basis of substantial absence of compressional tectonics between Tunisia and northwestern Sicily (northwestern Sicily Channel), this latter evidence suggesting that the compression has to be somehow redistributed, for instance shifting it toward the north in the Liguro-Provençal basin. We acknowledge, however, that the debate on the cause of the Liguro-Provençal compressional tectonics is still open and the evidence still weak to unambiguously support a single model.

It should eventually be noted that part of the contractional displacements are transferred to the north not only in the Liguro-Provençal region, but also in the Alboran-Algerian basin. The portion of southern Spain facing the Alboran-Algerian basin between about Alicante and Malaga, in fact, is presently undergoing compression [Fernádez-Ibáñez *et al*., 2007; Serpelloni *et al*., 2007].

The above-discussed tectonic reorganization of the western Mediterranean together with some local tectonic evidence [e.g., Pepe *et al*., 2005; Billi, Presti *et al*., 2007; Strzerzynski *et al*., 2010] points out the segmentation of the new incipient boundary, where inherited structures and lateral crustal heterogeneities prevent, at the present stage, the development of a continuous long convergent boundary consisting of thrust faults and an incipient subduction zone. If compression, basin inversion, and subsequent inception of subduction will continue in the presently active margins (i.e., east Alboran, Algerian, Liguro-Provençal, and south-Tyrrhenian; Fig. 9), we may hypothesize, in about 1500 km of convergent boundary (from east-Alboran to Tyrrhenian), a slab dip reversal (southward in the Algerian and south-Tyrrhenian margins and northward in the Liguro-Provençal margin) and a subduction zone spatial shift of about 600 km (the N-S distance between the Algerian and Liguro-Provençal margins).

In synthesis, the above-discussed past evolution and hypothesized future scenario for the western Mediterranean outlines a process similar to the Wilson Cycle (at a small scale), i.e., the opening and closing of ocean basins [Wilson, 1963]: (1) northward Nubian subduction with Mediterranean backarc extension (since ~35 Ma); (2) progressive cessation, from west to east, of Nubian main subduction (since ~15 Ma); (3) progressive onset of compression, from west to east, in the former backarc domain and consequent basin inversion (since ~8-10 Ma); (4) possible future subduction of former backarc basins.

7. CONCLUSIONS

Basin inversion, subduction, and thrusting on ocean-scale convergent boundaries, which may be at the origin of strong earthquakes and related tsunamis, are tectonic processes rather well known from several geologic and geophysical data. The nucleation modes and infant stages of these processes and related structures, however, are far less known for the paucity of evidence and uncertainty on where, on the Earth, these processes and structures are going to nucleate soon.

The geologic and geophysical evidence presented in this paper indicate that the western Mediterranean is a suitable region to study the onset of basin inversion that may possibly lead to subduction inception and to the final suture between Nubia and Eurasia in the study region. The same evidence reveal also some insights into the recent and present processes such as the lateral (scissor-like) migration of backarc inversion from west to east following a similar migration of subduction cessation, and the transfer of inversion tectonics toward the north in a weaker strand (i.e., Provence and Liguria) of the basin margin (Table 1). These insights will be useful to understand future tectonic scenarios in the western Mediterranean. An improvement of the geodetic and seismic networks (e.g., marine stations, see Dessa *et al*., [2011]) is, however, mandatory to better understand the kinematics and deformation rates of the studied boundary, thus possibly contributing to the knowledge of earthquake and tsunami hazard and to the assessment of their effects.

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FIGURE CAPTIONS

Figure 1. (**a**) Tectonic setting of the western Mediterranean area [Cadet and Funiciello, 2004]. (**b**) Moho depth (below sea level) in the western Mediterranean area [Tesauro *et al*., 2008]. Numbers are Moho depth in km. Abbreviations are as follows: A, Apennines; AB, Alboran Basin; AP, Adriatic Promontory; BC, Betic Cordillera; BS, Black Sea; CM, Cantabrian Mts; D, Dinarides; FB, Focsani Basin; MC, Massif Central; P, Pyrenees; PB, Pannonian Basin; TS, Tyrrhenian Sea; URG, Upper Rhine Graben; VT, Valencia Trough.

Figure 2. Schematic tectonic evolution of the western Mediterranean subduction zone between Nubia and Eurasia since 35 Ma [Faccenna *et al*., 2004].

Figure 3. Seismic cross-sections across the Liguro-Provençal and Algerian margins. (**a**) High-resolution MDJS43 seismic reflection profile acquired across the eastern Algerian margin, showing anticlinal deformations and inferred S-dipping, N-verging thrusts [Kherroubi *et al*., 2009]. (**b**) High-resolution A (Maradja cruise) seismic reflection profile acquired across the central Algerian margin, showing anticlinal deformations and inferred S-dipping, N-verging thrusts [Domzig *et al*., 2006]. S = salt. (**c**) High-resolution MA31 seismic reflection profile acquired across the Liguro-Provençal basin, showing a NE-dipping reflection dubiously interpreted as a late Pliocene-Quaternary S-verging thrust inverting the basin margin since about 3.5 Ma [Bigot-Cormier *et al*., 2004]. S = salt.

Figure 4. High-penetration seismic profile across the south-Tyrrhenian margin [Pepe *et al*., 2005]. See the cross-section track in Fig. 3. The hypocenter of the 2002 "Palermo" earthquake (M_w = 5.6, compressional fault plane solution) is approximately located on a N-dipping, S-verging reverse fault below the crystalline thrust sheet (KCU). The yellow bar is the error bar for the hypocenter depth.

Figure 5. (**a**) Crustal (depth ≤ 35km) earthquakes occurred between 1962 and 2009 in the western Mediterranean region. Data are from the ISC Bulletin (http://www.isc.ac.uk/) and include earthquakes with magnitude ≥ 4. (**b**) Same as (a) for intermediate and deep (depth > 35km) earthquakes. (**c**) Epicentral map of M ≥ 4 earthquakes occurred between 1900 and 1961 (data are from the EUROSEISMOS catalog available online at http://storing.ingv.it/es_web/).

Figure 6. Maps of moment tensor solutions for crustal, M ≥ 4.5 earthquakes. (**a**) Data are from the Harvard CMT catalog (1976-2009, http://www.globalcmt.org/CMTsearch.html). Grey background solutions indicate data also shown in (b). (**b**) Data are from the RCMT catalog (1997-2004, http://www.bo.ingv.it/RCMT/yearly_files.html). Red = normal faulting, blue = thrust faulting, green = strike-slip faulting, and black = unknown stress regime (see text for details). (**c**) Polar plots of P- and T-axes for earthquakes grouped in sectors of significant seismic relevance (red = CMT and blue = RCMT). Full and empty dots correspond to P- and T-axes, respectively. Dashed line indicates the Adria lithosphere according to Chiarabba *et al.* [2005]. For the 1997-2004 time interval, P- and T-axes values included in (c) are taken from the RCMT catalog, whereas data pertaining to the grey background solutions plotted in (a) are not considered.

Figure 7. (a) Horizontal GPS velocities given with respect to a fixed Eurasian frame (see text). Red arrows show observed velocities, with 95% error ellipses; white arrows show the velocities predicted by the Nubia-Eurasia relative rotation pole; grey arrows show the interpolated velocity field, obtained from inversion of the horizontal velocity gradient field (see text). **(b)** Horizontal strain-rate field obtained from least-squares interpolation of the horizontal velocity field presented in (a). Red diverging arrows show extensional strain-rates, whereas blue converging arrows show contractional.

Figure 8. Velocity anomalies in the upper mantle below the western Mediterranean area after the tomographic model of Piromallo and Morelli [2003]. P-velocity perturbation is displayed with respect to reference model *sp6* [Morelli and Dziewonski, 1993].

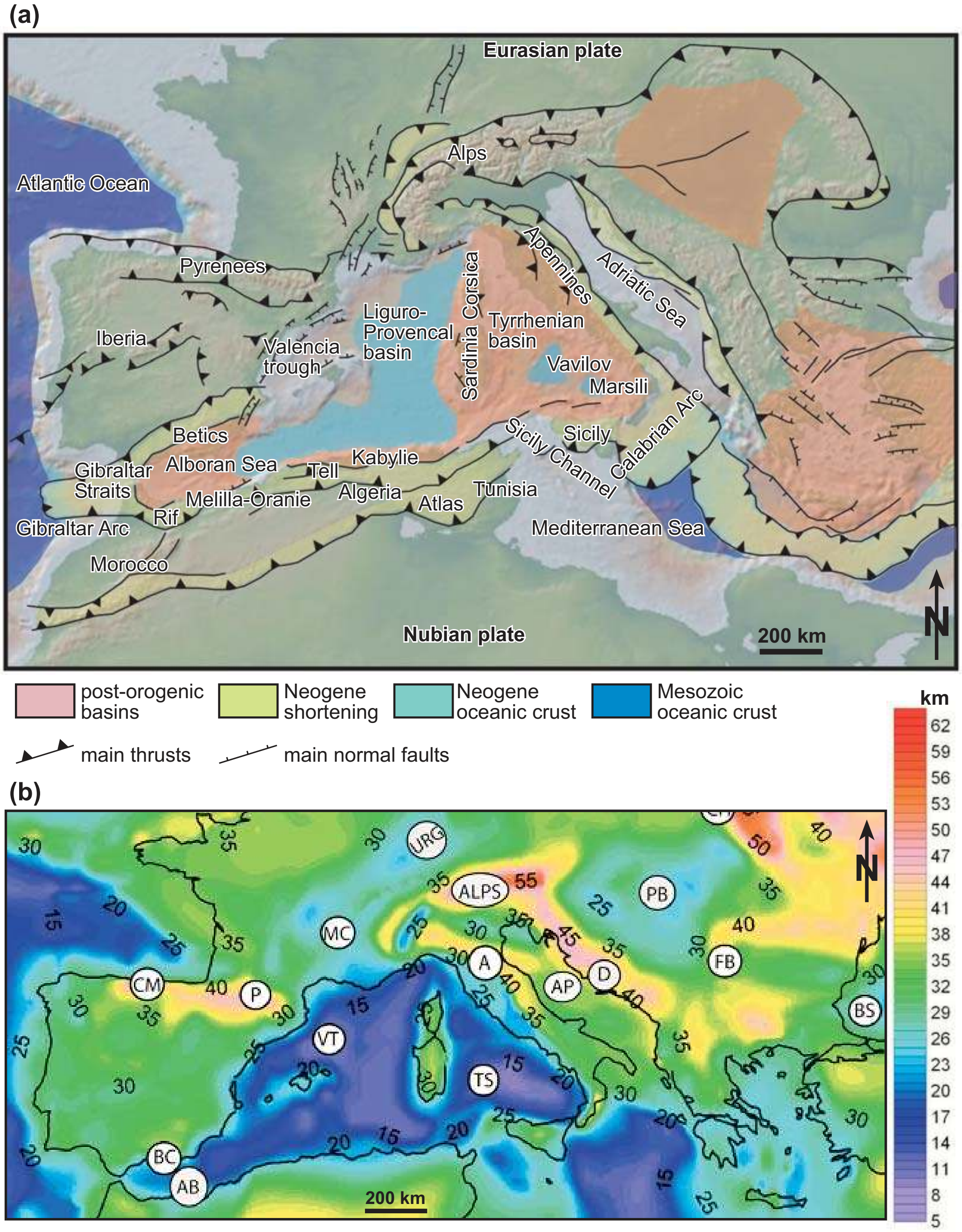
Figure 9. Three-dimensional tomographic-topographic model (tomography after Piromallo and Morelli, 2003) of the western Mediterranean region. The isosurface encloses a volume characterized by P-wave velocity anomalies larger than 0.6% relative to the reference model. Hypothetical future segments of the Nubia-Eurasia convergent boundary (indicated as incipient convergent boundary) are inferred from the original and previously-published evidence presented in this paper. Basin inversion has propagated in a scissor-like manner from the Alboran basin (c. 8 Ma) to the south-Tyrrhenian domain (younger than c. 2 Ma) following a similar propagation of the subduction cessation, i.e., older to the west and younger to the east (Table 1). Color code of the incipient convergent boundary indicates progressively younger (lighter color) inversion or compression from west to east (see age numerical indications for the onset of compression-inversion). The possibility of a new convergent boundary in the Liguro-Provençal region is still debated (see the text). Dashed segments of the incipient boundary are between Tunisia and Sicily, where no oceanic crust occurs and the occurrence of a future convergent boundary is therefore unlikely, and in the southern Iberian margin, where basin inversion is not as advanced as in the opposite Algerian margin. The track of the former subduction zone (trench) can be inferred from relics of the Nubian slab to the east of Sicily (Calabrian arc) and in the Algerian inland.

Figure 10. Cartoon showing the proposed model of subduction inception. Subduction inception is influenced by friction along the plate interface and therefore by the size of the interface [Toth and Gurnis, 1998]. The lateral propagation, in a scissor-like fashion, of the new plate boundary may result less friction-resistant than other models of subduction inception where subduction is initiated all at once along longer segments of the new plate boundary. In the case of the western Mediterranean, the area of subduction inception corresponds with the Algerian margin [Strzerzynski *et al.*, 2010] and the propagation of subduction is toward the east. The Alboran basin is excluded by this process (subduction) for the substantial absence of oceanic crust.

Table 1. Tentative ages for slab window formation and onset of backarc compression in the western Mediterranean. These episodes are inferred from the available literature and help to understand the spatio-temporal evolution of the tectonic reorganization in the study area, i.e., from subduction cessation to backarc compression and inversion perhaps leading to subduction of former backarc basins. Proposed references are only a selection of the vast literature on the subject (see text).

slab window formation			**backarc compression**		
geographic location	**age of slab window**	**Ref.**	**geographic location**	**age of backarc compression onset**	**Ref.**
Oranie-Melilla (Morocco-Algeria)	~ 15 Ma	(1), (2)	Alboran margin	~ 8 Ma	(5), (6)
Nefza and Mogodos (Tunisia)	~ 10-8 Ma	(2), (3)	Algerian margin	~ 7-5 Ma	(7)
			Provençal margin	~ 3.5 Ma	(8)
Prometeo and Ustica (S-Tyrrhenian)	~ 4 Ma	(2), (4)	S-Tyrrhenian margin	younger than ~ 2 Ma	(9), (10)

(1) Coulon *et al.*, [2000]; (2) Faccenna *et al.*, [2004]; (3) Maury *et al.*, [2000]; (4) Faccenna *et al.*, [2005]; (5) Bourgois *et al.*, [1992]; (6) Comas *et al.*, [1999]; (7) Mauffret, [2007]; (8) Bigot-Cormier *et al.*, [2000]; (9) Goes *et al.*, [2004]; (10) Billi *et al.*, [2007].



(a)
Eurasian plate
Alps
Atlantic Ocean
Pyrenees
Apennines
Adriatic Sea
Liguro-Provencal basin
Corsica
Sardinia
Tyrrhenian basin
Iberia
Valencia trough
Vavilov
Marsili
Calabrian Arc
Sicily
Betics
Sicily Channel
Gibraltar Straits
Alboran Sea
Tell
Kabylie
Melilla-Oranie
Algeria
Tunisia
Rif
Atlas
Gibraltar Arc
Mediterranean Sea
Morocco
Nubian plate
200 km
N
post-orogenic basins
Neogene shortening
Neogene oceanic crust
Mesozoic oceanic crust
main thrusts
main normal faults
(b)
km
URG
ALPS
PB
MC
A
AP
D
FB
CM
P
VT
TS
BS
BC
AB
200 km
N

Figure 1

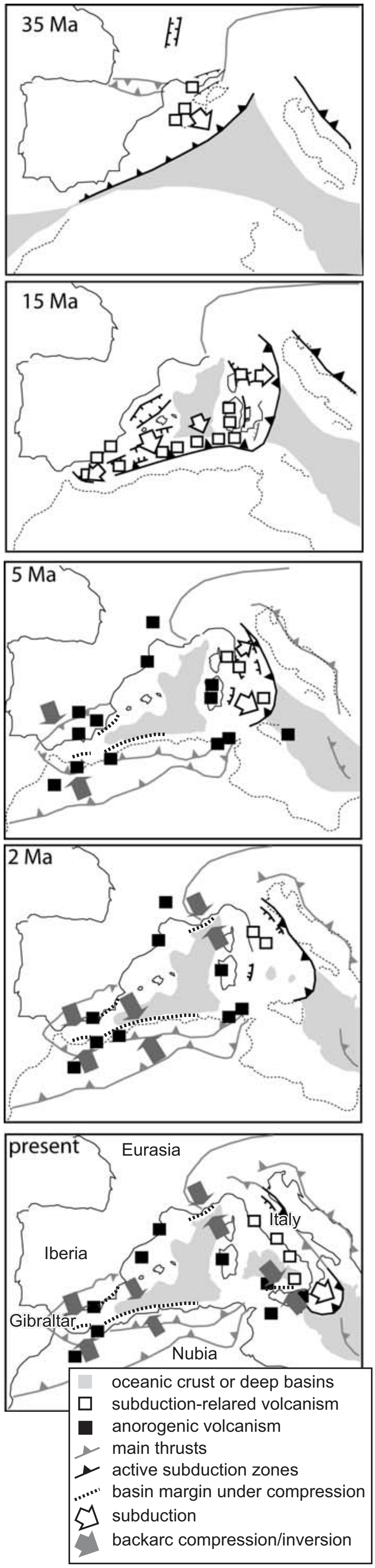
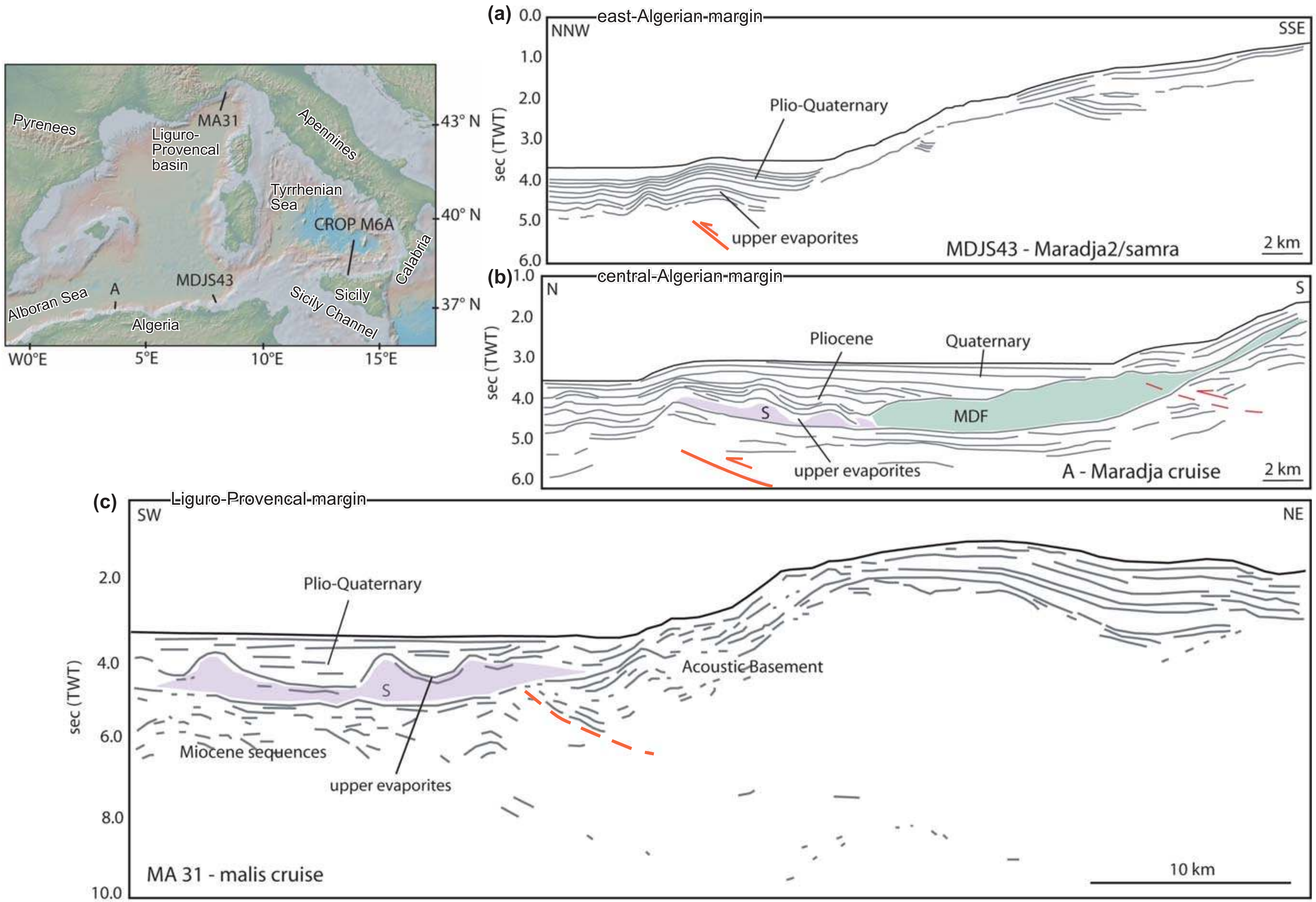

35 Ma
15 Ma
5 Ma
2 Ma
present
Eurasia
Italy
Iberia
Gibraltar
Nubia
oceanic crust or deep basins
subduction-relared volcanism
anorogenic volcanism
main thrusts
active subduction zones
basin margin under compression
subduction
backarc compression/inversion

Figure 2



Pyrenees
MA31
Liguro-
Provencal
basin
Apennines
Tyrrhenian
Sea
CROP M6A
Calabria
Alboran Sea
A
MDJS43
Sicily
Sicily Channel
Algeria
43° N
40° N
37° N
W0°E
5°E
10°E
15°E
(a)
east-Algerian-margin
NNW
SSE
sec (TWT)
0.0
1.0
2.0
3.0
4.0
5.0
6.0
Plio-Quaternary
upper evaporites
MDJS43 - Maradja2/samra
2 km
(b)
central-Algerian-margin
N
S
1.0
2.0
3.0
4.0
5.0
6.0
Pliocene
Quaternary
S
MDF
upper evaporites
A - Maradja cruise
2 km
(c)
Liguro-Provencal-margin
SW
NE
sec (TWT)
2.0
4.0
6.0
8.0
10.0
Plio-Quaternary
Acoustic Basement
S
Miocene sequences
upper evaporites
MA 31 - malis cruise
10 km

Figure 3

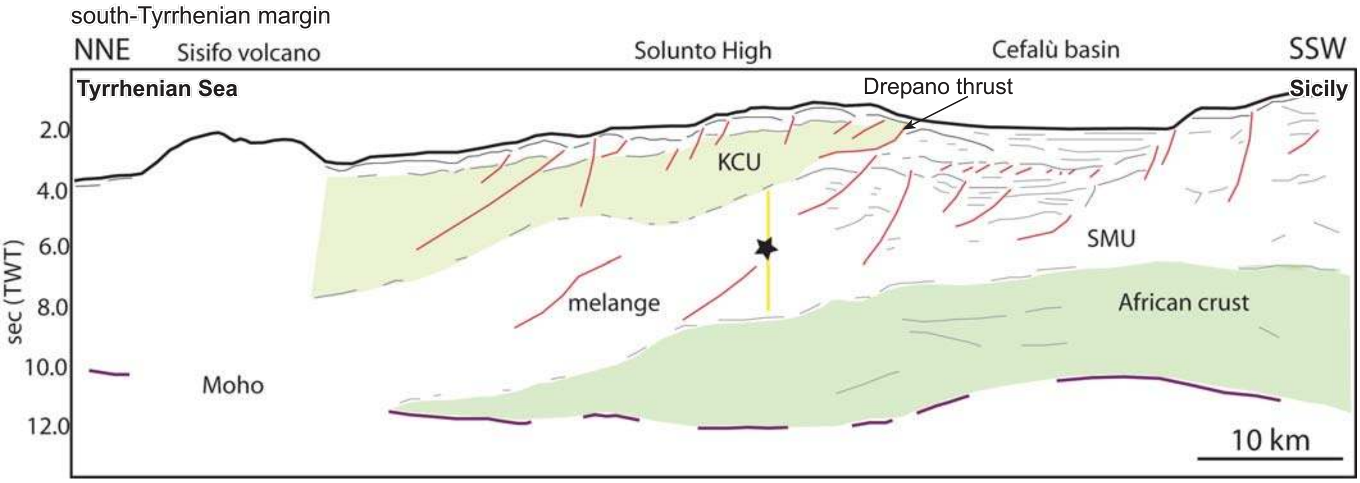

south-Tyrrhenian margin
NNE
Sisifo volcano
Solunto High
Cefalù basin
SSW
Tyrrhenian Sea
Drepano thrust
Sicily
2.0
4.0
6.0
8.0
10.0
12.0
sec (TWT)
KCU
SMU
melange
African crust
Moho
10 km

Figure 4

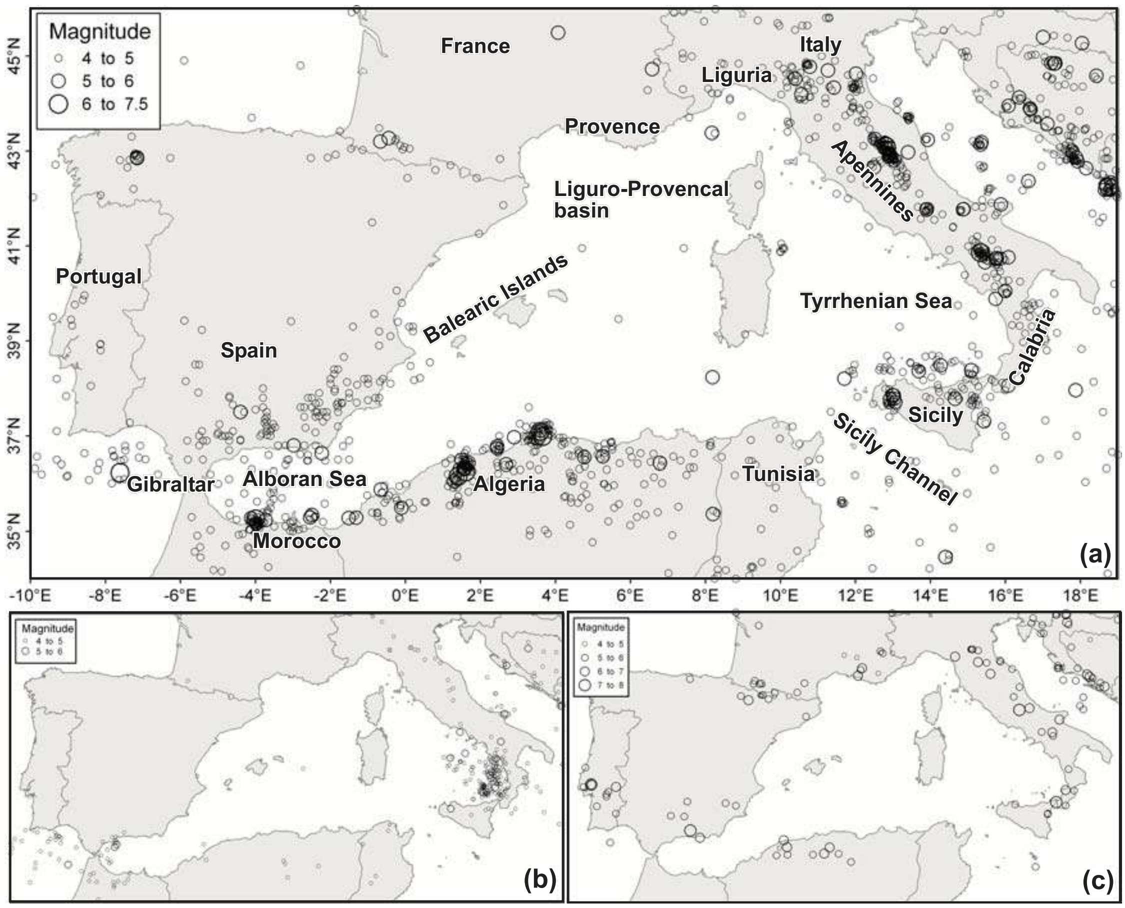

Magnitude
4 to 5
5 to 6
6 to 7.5
France
Italy
Liguria
Provence
Liguro-Provencal
basin
Apennines
Portugal
Spain
Balearic Islands
Tyrrhenian Sea
Calabria
Sicily
Sicily Channel
Tunisia
Algeria
Alboran Sea
Gibraltar
Morocco
(a)
45°N
43°N
41°N
39°N
37°N
35°N
-10°E
-8°E
-6°E
-4°E
-2°E
0°E
2°E
4°E
6°E
8°E
10°E
12°E
14°E
16°E
18°E
(b)
(c)

Figure 5

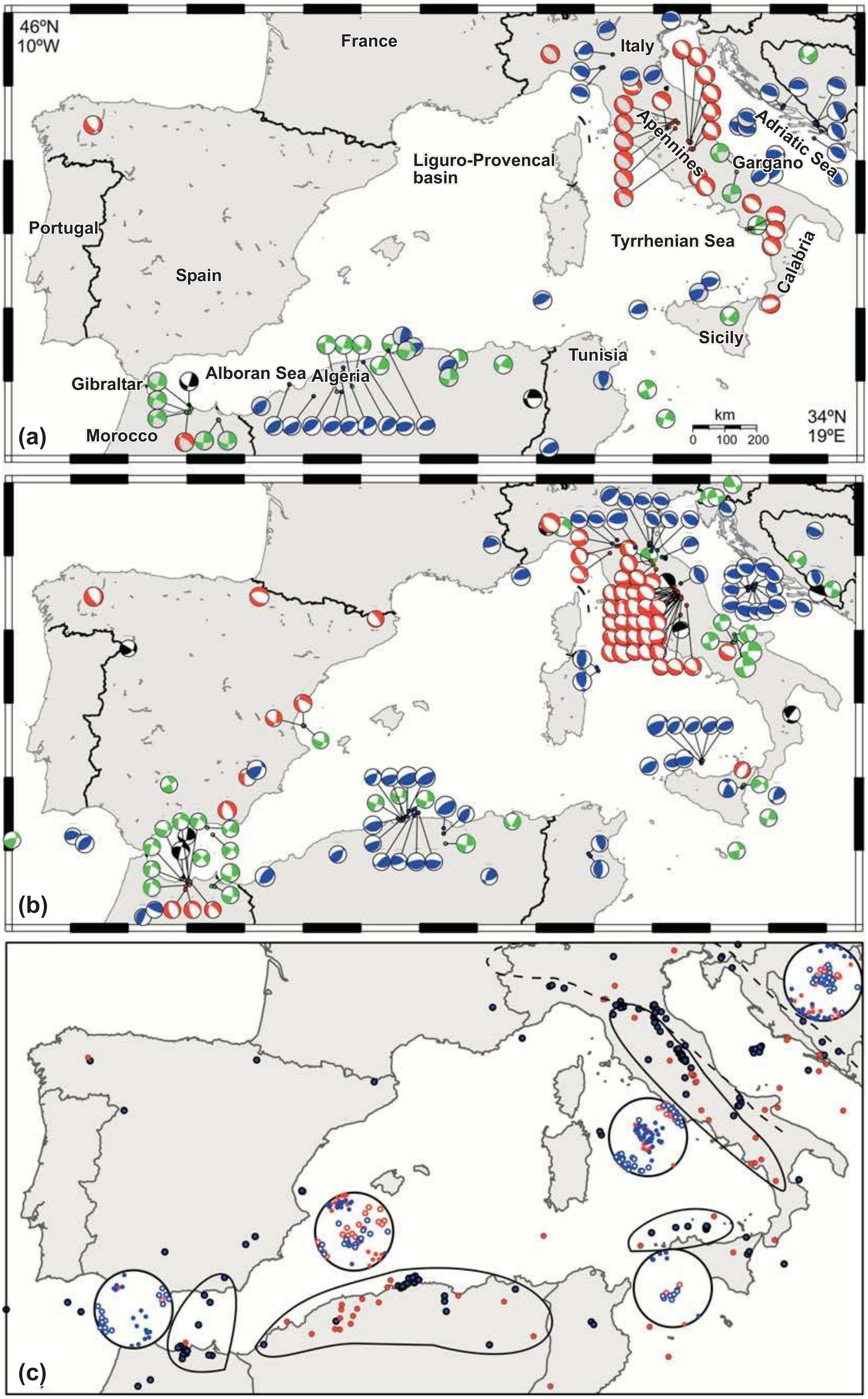
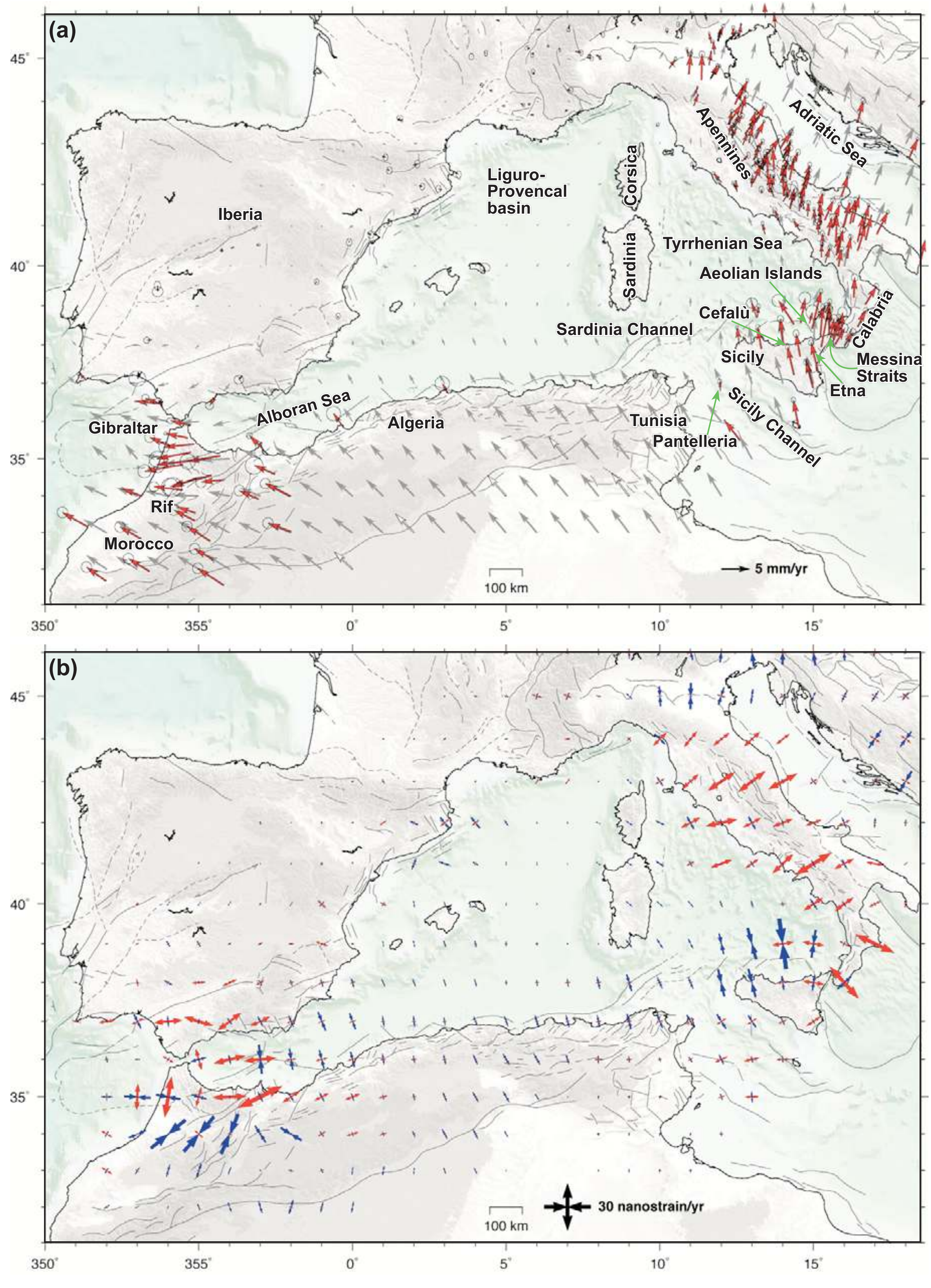

46ºN
10ºW
France
Italy
Apennines
Adriatic Sea
Gargano
Liguro-Provencal
basin
Portugal
Spain
Tyrrhenian Sea
Calabria
Sicily
Tunisia
Gibraltar
Alboran Sea
Algeria
Morocco
km
0 100 200
34ºN
19ºE
(a)
(b)
(c)

Figure 6



(a)
45°
40°
35°
350°
355°
0°
5°
10°
15°
Iberia
Liguro-
Provencal
basin
Corsica
Sardinia
Tyrrhenian Sea
Apennines
Adriatic Sea
Aeolian Islands
Cefalù
Calabria
Sardinia Channel
Sicily
Messina
Straits
Etna
Sicily Channel
Tunisia
Pantelleria
Algeria
Alboran Sea
Gibraltar
Rif
Morocco
100 km
5 mm/yr
(b)
100 km
30 nanostrain/yr

Figure 7

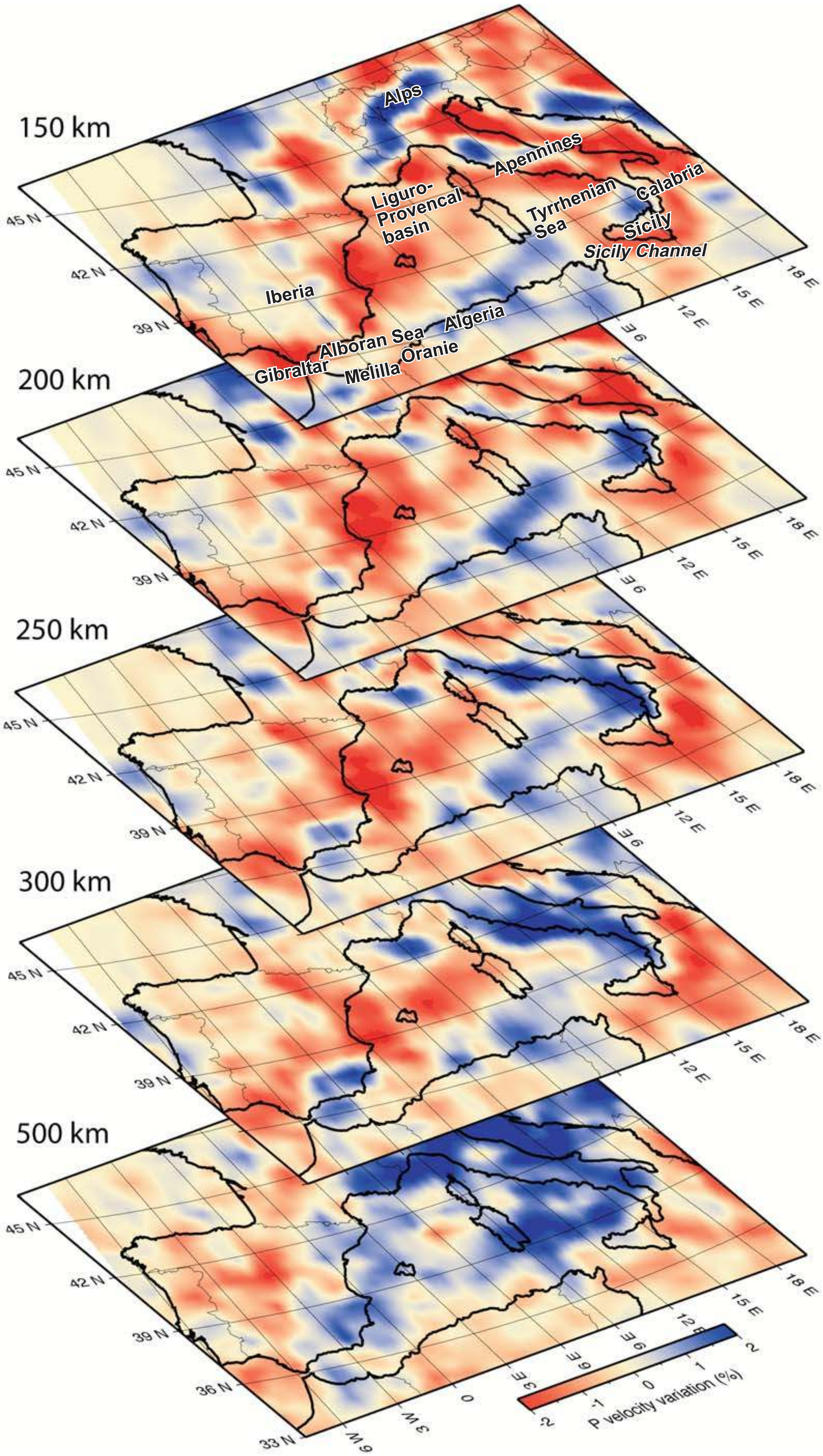
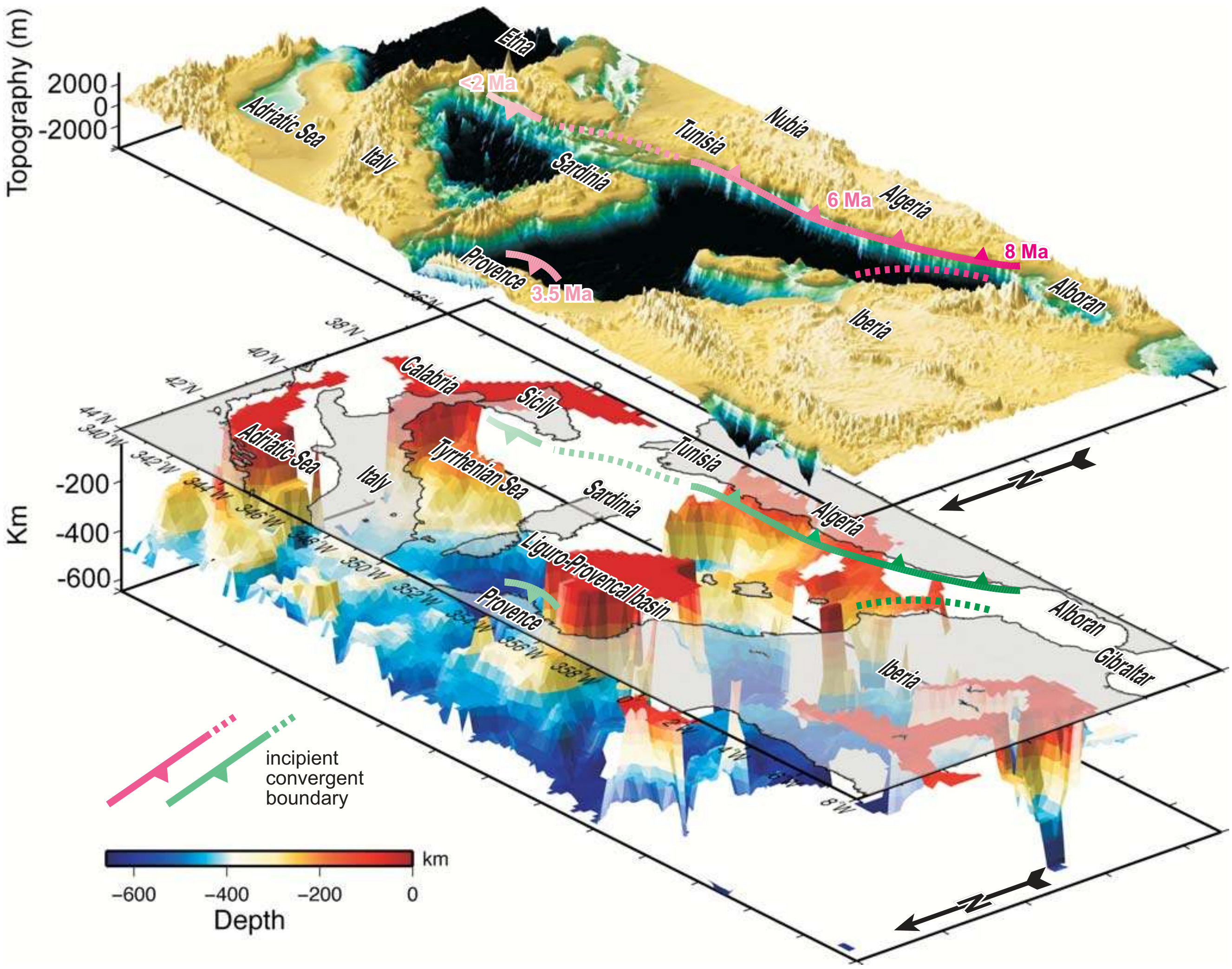

150 km
200 km
250 km
300 km
500 km
Alps
Apennines
Liguro-
Provencal
basin
Tyrrhenian
Sea
Calabria
Sicily
Sicily Channel
Iberia
Algeria
Alboran Sea
Oranie
Gibraltar
Melilla
45 N
42 N
39 N
36 N
33 N
9 W
6 W
3 W
0
3 E
6 E
9 E
12 E
15 E
18 E
-2
-1
0
1
2
P velocity variation (%)

Figure 8



Topography (m)
2000
0
-2000
Etna
<2 Ma
Adriatic Sea
Italy
Sardinia
Tunisia
Nubia
6 Ma
Algeria
8 Ma
Provence
3.5 Ma
Alboran
Iberia
Calabria
Sicily
Tyrrhenian Sea
Liguro-Provencal basin
Gibraltar
Km
-200
-400
-600
incipient
convergent
boundary
km
−600
−400
−200
0
Depth
N

Figure 9

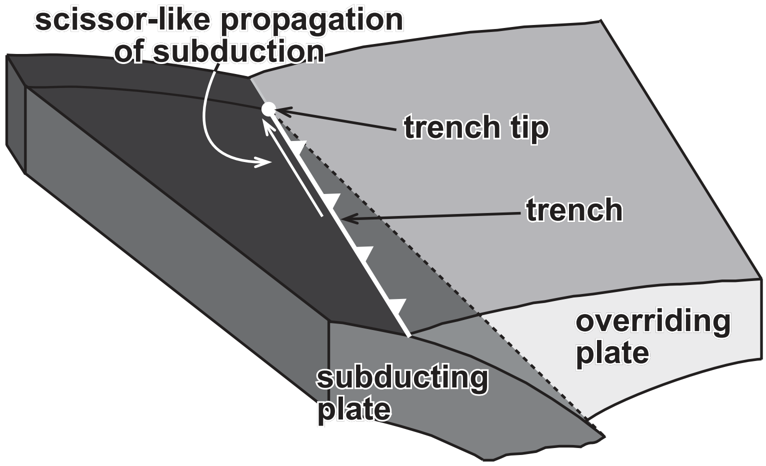

scissor-like propagation
of subduction
trench tip
trench
overriding
plate
subducting
plate

Figure 10