United Nations Educational Scientific and Cultural Organization
and
International Atomic Energy Agency

THE ABDUS SALAM INTERNATIONAL CENTRE FOR THEORETICAL PHYSICS

UPPER MANTLE FLOW IN THE WESTERN MEDITERRANEAN

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MIRAMARE – TRIESTE

July 2006

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Abstract

Two cross-sections of the western Mediterranean Neogene-to-present backarc basin are presented, in which geological and geophysical data of the Transmed project are tied to a new shear-wave tomography. Major results are i) the presence of a well stratified upper mantle beneath the older African continent, with a marked low-velocity layer between 130-200 km of depth; ii) the dilution of this layer within the younger western Mediterranean backarc basin to the north, and iii) the easterly raising of a shallower low-velocity layer from about 140 km to about 30 km in the Tyrrhenian active part of the backarc basin. These findings suggest upper mantle circulation in the western Mediterranean backarc basin, mostly easterly-directed and affecting the boundary between upper asthenosphere (LVZ) and lower asthenosphere, which undulates between about 180 km and 280 km.

1. Introduction

Usually there is a gap between surface geology and mantle tomography. Since the Mediterranean is one of the most studied areas in the world, we try to integrate the deep structure of the Mediterranean along two Transmed geotraverses (Cavazza et al., 2004), i.e., II and III (Roca et al., 2004; Carminati et al., 2004) (Fig. 1). The Mediterranean has been shaped by a number of subduction zones since the Cretaceous, where segments of the Tethyan oceanic or thinned continental lithosphere have been recycled into the mantle. The present geodynamic setting is characterized by the widespread Neogene-to-Present backarc basin in the western Mediterranean in the hanging-wall of the westerly-directed Apennines-Maghrebides subduction, and the northeasterly directed Dinaric-Hellenic subduction in the eastern Mediterranean. In this study we focus mainly on the western side of the basin, comparing superficial geological and geophysical constraints (Figs. 2a and 3a) with a new shear-wave tomography of the area (Figs. 2b and 3b).

The backarc spreading in the western Mediterranean developed since 30 Ma and gradually shifted from west to east, moving from the Provennçal, Valencia and Alboran basins, to the Algerian and Tyrrhenian basins (e.g. Gueguen et al., 1998; Carminati and Doglioni, 2004). The rifting accompanied the "eastward" retreat of the subduction zone, and the slab retreat kinematically requires a contemporaneous "eastward" mantle flow, regardless this is the cause or a consequence of the retreat (Doglioni et al., 1999). However a number of controversies still exist in the geodynamic reconstruction of this area.

We contribute to this debate with a set of shear-wave velocity (V_s) models of the lithosphere-asthenosphere system along the two geotraverses. The two sections run respectively from south France, through Pyrenees, Valencia basin, to north Algeria into the Sahara platform (II), and from central France, through the Gulf of Lion, Sardinia, southern Apennines, Albania, Balkans and the Moesian platform in Bulgaria (III) (Figs. 2a and 3a).

The V_s structure of the very upper mantle, in the central western Mediterranean, is obtained by non-linear inversion of surface waves dispersion curves obtained from a tomographic analysis of regional and global dispersion data. The local dispersion curves are assembled in the period range 5-150 s, combining regional group velocity measurements and published Rayleigh wave dispersion data. The resolution of the tomography data is improved using a priori information about the shallow crustal velocity structure. Local Smoothness Optimization (LSO) is applied to select the representative cellular structures and the three-dimensional model. The lithosphere-asthenosphere velocity structure is reliably reconstructed to depths of about 300 km and highlights some new features along the geotraverses, which largely fit the superficial information.

The comparison of the superficial geological and geophysical constraints with the new V_s tomography of the area results is a new model of the mantle flow in a backarc setting.

2. Evolution of the Western Central Mediterranean and of the Balkan area

Although a detailed analysis of the tectonic and geodynamic evolution of the area is beyond the scope of this paper, a brief description of the evolution sketched in Fig. 1 will be provided. For a more complete discussion the reader is referred to Gueguen et al. (1998), Carminati et al. (1998), Dercourt et al. (2000), Stampfli et al. (2001), Roca et al. (2004), Carminati et al. (2004).

Traces of pre-Cenozoic deformations are widespread in the Western-Central Mediterranean area and adjoining regions. However, the present crustal and lithospheric geometry of the region developed since the Cenozoic. For this reason the description of the evolution will range from Eocene to Recent times.

Paleomagnetic studies (Alvarez et al., 1974; Vigliotti and Langenheim, 1995; Speranza et al., 2002) evidenced an anticlockwise rotation of the Corsica-Sardinia block between 19 Ma and 16 Ma (Burdigalian). In the Middle Eocene (45 Ma), after the restoration of the Corsica-Sardinia block at their position prior to rotation, the Alps were probably linked to the Betics through Alpine Corsica and the Balearic promontory to form a double vergent belt related to a south-eastward subduction of Neotethyan oceanic and European continental lithosphere underneath the African plate (Doglioni et al., 1999). Contemporaneous shortening occurred in the Pyrenees (until ~24.7 Ma according to a magnetostratigraphic study; Meigs et al., 1996) driving to the complete inversion of a basin intervening between Iberia and Eurasia.

At about 30 Ma ago, the west directed Apennines-Maghrebides subduction started, nucleating along the Alps-Betics retrobelt and possibly triggered by the occurrence, in the foreland east of the Alpine belt, of oceanic or thinned continental lithosphere, as proposed by Doglioni et al. (1999).

The Apennines-Maghrebides subduction zone was characterised, since the beginning, by fast radial eastward rollback, as evidenced by the migration of the subduction related calkalcaline volcanism and of the compressional front which induced widespread extensional tectonics accompanied by alkaline volcanism in the backarc (Gulf of Lions and Provençal Basin, in the Catalan Coastal Ranges area, in Sardinia, still attached to Iberia, in the Valencia trough and in the Algerian Basin). In the Provençal and in the Algerian Basins, continental crust stretching evolved into oceanization in the Lower-Middle (?) Miocene, coeval with the counter-clockwise rotation of the Corsica-Sardinia block. Large portions of the Alps-Betics orogen located backarc area were disarticulated and spread-out into the western Mediterranean (e.g., the metamorphic slices of Kabylie in northern Algeria, and Calabria in southern Italy). The Apennines and Maghrebides fold-and-thrust belt developed on top of the retreating subduction and the deformation front migrated to the east in the Apennines and to the South in the Maghrebides, following the slab roll-back (Malinverno and Ryan, 1986; Patacca et al., 1990; Doglioni, 1991). It has been proposed that, in the Langhian, continental collision was followed by slab breakoff along the northern African margin, as suggested by tomographic models, by the occurrence of bimodal volcanism and by uplift along the African margin (Carminati et al., 1998; Maury et al., 2000; Coulon et al., 2002). This hypothesis was accepted by Roca et al. (2004) and no subducting slab is imaged beneath northern Africa in the TRANSMED II section.

A shift of active extension from west to east of Sardinia occurred in the Langhian (ca. 15 Ma) as testified by the Middle Miocene to present opening of the Tyrrhenian Basin.

The extension affecting most of the western Mediterranean in the Tertiary developed in a context of relative convergence between Africa and Europe. The maximum amount of N-S Africa/Europe relative motion in the last 23 Ma was about 135 km at the Tunisia longitude, whereas the eastward migration of the Apennines arc exceeded 700 km during the same time span (Fig. 1). As a consequence the eastward migration of the Apennines-Maghrebides arc cannot be considered as a consequence of the relative N-S Africa/Europe convergence but it is rather a consequence of the Apennines-Maghrebides subduction rollback.

The Balkan area is characterized by the occurrence of a polyphased orogenic belt named differently in different areas: Dinarides, Hellenides and Taurides. This double vergence orogen is the result of at least two or three subduction zones since Mesozoic times, as testified by the occurrence of two distinct oceanic sutures (Vardar and Sub-Pelagonian ophiolites), representing one or two (contrasting reconstructions) branches of the Mesozoic Tethyan ocean, and the present oceanic subduction of the Ionian Sea. In the Vardar ocean, the northeast-directed subduction is dated Jurassic-Early Cretaceous and deformation did not significantly involve the shelf margin of the Adriatic plate. The east-dipping subduction of the Adriatic lithosphere beneath the Dinarides started in the Mesozoic and in the Tertiary became the main process acting in the area. The main structuration of the Dinarides occurred progressively from the latest Maastrichtian, with climax between the Oligocene and earliest Miocene.

3. The geological cross-sections

3.1 Sources of information

A detailed description of sources of information used to constrain the geological sections of Figs. 2a and 3a is in Roca et al. (2004) and Carminati et al. (2004); we summarize in the following the data used to constrain only the geometry of the base of crust and lithosphere, since these are the main features that can be compared with tomographic data.

3.1.1 TRANSMED II section

In the Aquitaine Basin-Pyrenees and in the Ebro basin-Catalan Coastal Ranges, the crustal geometry was constrained by the ECORS-Central Pyrenees deep seismic reflection profile (Roure et al., 1989) and by the ESCI-Catalanides deep seismic reflection profile (Gallart et al., 1994) respectively and by refraction and gravimetric data (e.g. Banda et al., 1983; Torné et al., 1989; Dañobeitia et al., 1992; Suriñach et al., 1993), by wide-angle reflection data (Vidal et al., 1995), by magnetotelluric modelling (e.g. Ledo et al., 2000) as well as by seismic tomography data (Souriau and Pauchet, 1998). The lithosphere-asthenosphere boundary was drawn from integrated lithospheric models combining topography, gravity, and heat flow (Zeyen and Fernàndez, 1994).

In the València Trough and Balearic promontory, the Moho depth was based on the interpretation of the ESCI-València trough deep seismic reflection profile (e.g. Santanach, 1997; Vidal et al., 1998) but regional refraction and gravimetric data (Gallart et al., 1990; Dañobeitia et al., 1992; Gallart et al., 1994) were also taken in account. The lithospheric mantle thickness was derived from both gravimetric and geoid anomaly models (Ayala et al., 2003) and from the integrated model of Zeyen and Fernàndez (1994) and from an integrated model by Roca et al. (2004).

The oceanic nature of the Algerian Basin crust and its thickness were constrained by seismic refraction results (Hinz, 1972, 1973) and by aeromagnetic data (Galdeano and Rossignol, 1977). The lithosphere-asthenosphere boundary was calculated specifically for the TRANSMED Project (Roca et al., 2004) by integrated lithospheric modeling combining thermal, gravity, and local isostasy analysis.

In the Kabylies-Tell-Atlas and in the Saharan portions of the transect, the Moho depth was drawn according to the results of gravity models (Mickus and Jallouli, 1999) whereas the lithospheric thickness and geometry was based on an original integrated model (Roca et al., 2004).

3.1.2 TRANSMED III section

In the Massif Central the Moho was drawn from refraction data (Sapin and Hirn, 1974), while lithospheric geometries were based on tomography results (Granet et al., 1995a, 1995b) and on the results of mantle thermal modelling (Sobolev et al., 1996). The structure of the Moho below the Gulf of Lion platform and slope was based on deep seismic reflection data (Pascal et al., 1993) and gravity modelling (Chamot-Rooke et al., 1997).

In the Provençal Basin and in the western continental shelf of Sardinia, the Moho depth was mainly based on the ECORS and CROP deep seismic reflection profiles (De Voogd et al., 1991) and by seismic refraction results (Egger et al., 1988; Pascal et al., 1993). The base of the lithosphere was drawn from the gravity models of Cella et al. (1998) and Yegorova et al. (1997).

Below Sardinia, Moho depth and lithospheric thickness were drawn according to the results of Nicolich and Dal Piaz (1990) and Panza et al. (1990) and on the results of the European Geotraverse project (Ansorge et al., 1992). The asymmetric lithosphere boundary under Sardinia was based on the gravity modelling of Cella et al. (1998).

In the Tyrrhenian Sea, the depth of the Moho was derived from Nicolich and Dal Piaz (1990) partly modified by Carrara (2002), while the geometry of the lithosphere-asthenosphere boundary was based on the models of Panza et al. (1990) and Pontevivo and Panza (2002).

In the continental Italy and Adriatic portions of the transect, running through the southern Apennines and the Apulian foreland, the geometry of the Moho and of the lithosphere-asthenosphere boundary was based on the studies of Nicolich and Dal Piaz (1990), Panza et al. (1990), Scarascia et al. (1994), Lucente et al. (1999) and Pontevivo and Panza (2002). The location of the Apenninic subducting slab (subcrustal seismicity is rare), constrained by the studies of De Gori et al. (2001), Chimera et al. (2003) and Panza et al. (2003), it is consistent with the occurrence of positive Bouguer anomalies (up to 120 mGal or more; Bigi et al., 1990) and very high heat flow values (up to 140 mW/m^2 or more; Della Vedova et al. 2001) along the Tyrrhenian margin and in the adjacent Tyrrhenian Sea. Moreover,

the occurrence of hot asthenospheric material at relatively shallow depth below the western portion of the Southern Apennines is consistent with shear waves attenuation (Mele et al., 1997) and geochemical studies (Italiano et al., 2000).

Below the Albanian Dinarides, the Moho depth was constrained by gravimetric data (Frasheri et al., 1996). The geometry of the slab subducting beneath Albania (subcrustal seismicity is also absent) was constrained by tomographic studies showing a clear a fast velocity body (interpreted as the evidence of the active subduction process) under the Albanian Dinarides (Piromallo and Morelli, 2003).

Below the Macedonian Balkanides, the depth of Moho discontinuity was both based on two deep seismic profiles and on magnetic (Stojkovic et al., 1976) and gravity (Bilibajkic et al., 1979) models. More to the east, in the Bulgarian Balkanides and in the Moesian platform, the Moho discontinuity and the base of the lithosphere were deduced from gravimetric and seismic data (Babuska et al., 1986; Volvovsky et al., 1987; Volvovsky and Starostenko, 1996; Boykova 1999).

3.2 Description of the geological cross-sections

3.2.1 TRANSMED II section

In its northernmost part, the transect crosses the northern foreland basin of the Pyrenees (the Aquitaine basin), which rests on a 35 km thick crust and 110 km thick lithosphere. More to the south, the Pyrenees, an orogenic belt which developed between the late Senonian (Late Cretaceous) and the mid-Oligocene times, characterized by a doubling of the crust which reaches depths of 70 km, whereas the base of the lithosphere is at 150 km. To the south, the crust and the lithospheric mantle (LID) gradually thins reaching below the southern foreland basin of the Pyrenees (the Ebro basin) thicknesses comparable to those of the Aquitaine basin.

The continental crust and the LID continue to thin below the Catalan Coastal Ranges, reaching minimum values (of 15-20 and 55 km respectively) below the Valencia trough, an extensional basin developed during the Oligocene-Lower Miocene. The Valencia trough is bordered to the south by the Balearic Promontory, characterized in the Tertiary by both compressional and extensional tectonics (Doglioni et al., 1997) and underlined by relatively thicker continental crust (25 km) and lithosphere (90 km).

South of the Balearic Islands the lithosphere thins quite abruptly, leading to the Algerian basin (Mauffret et al., 2004), a Miocene basin floored by 10 km thick oceanic crust and 50 km thick lithosphere. The crust and lithosphere thicken again rather continuously below northern Africa where the transect crosses the Tell fold-and-thrust belt and reach maximum thicknesses of 40 km and 170-180 km, respectively, below the Sahara Atlas (an intraplate fold-and-thrust belt) and the Saharan platform.

3.2.2 TRANSMED III section

At its northwestern end the section crosses the French Massif Central, where 30 km continental crust deformed during the Variscan orogeny outcrop. The lithospheric base, generally around 80-90 km

deep, is interpreted to be much shallower (around 50 km) below the Cenozoic Massif Central alkaline volcanic province. Since the crust is not thinned accordingly, the volcanism and the asthenosphere upwelling are interpreted in literature as a thermal effect due to the upwelling of a mantle plume (Granet et al., 1995a, 1995b).

The continental crust and lithosphere thin first progressively below the Gulf of Lions continental margin, and then abruptly below the Provençal Basin, floored by Neogene oceanic crust, where they reach thicknesses of 10-15 km and 25 km, respectively. A rather progressive increase of crustal and lithospheric thickness occurs below the thinned continental lithosphere of the western Sardinia margin and below the continental swell of the Corsica-Sardinia block, which was structured during the Variscan and older orogenic cycles, and was later dissected by Neogene-Quaternary extensional tectonics. The Sardinia crust and lithosphere have maximum thicknesses of 30 and 70 km, respectively. Gravity modelling (Cella et al., 1998, 2006) suggests an asymmetric morphology of the lithospheric roots across Sardinia. Maximum lithospheric thicknesses are shifted to the east with respect to highest topography. Below the western Sardinian margin the increase of the lithospheric thickness toward the centre of the Island is less abrupt than below the eastern Sardinia margin.

East of Sardinia the lithosphere thins again progressively, reaching its minimum thicknesses (less than 20 km) in the Tyrrhenian Sea, a basin partly floored by oceanic crust formed mainly from the Tortonian to the Present. Farther east (crossing the Campania continental margin) the crust and lithosphere gradually thicken, reaching thicknesses of 30 km and 40 km, respectively, below the Southern Apennines, a Neogene fold-and-thrust belt, dissected by Late Neogene-Present extensional tectonics. This belt developed in the hanging-wall of a west-directed subduction zone, where the, up to 80-100 km thick, continental lithosphere of the Adria microplate steeply sinks to the west.

To the east, the Apennines foreland (partly in continental Italy and partly in the southern Adriatic Sea) rests on the about 100 km thick continental lithosphere of the Adria microplate, thinned during the Mesozoic rifting. The crustal thickness is around 30 km. In the eastern southern Adriatic Sea, the transect crosses the foreland basin of the Albanian Dinarides, an orogen associated with the northeastward subduction of the Adriatic lithosphere. Due to subduction related flexure, the base of crust and lithosphere reach depths of 40 km and 120 km below the foreland basin and even deeper depths below the Dinarides. Further to the east, the complex multistage Dinarides-Hellenides orogen is crossed together with its conjugate retrobelt, i.e., the Balkans. The transect ends in the Moesian platform, which is the undeformed foreland of the Balkans. The whole region was affected by Neogene-to-Present extensional tectonics. In these areas the lithosphere is continental and generally thickened (about 40 km thick crust and 120-130 km thick lithosphere), but local thinning occurs in correspondence of the Sofia graben.

4. The seismic cross-sections

4.1 Seismic data processing and methods

This study is part of a series of regional tomography investigations made in the Mediterranean area (Panza et al., 2002, 2003, 2006; Raykova and Panza, 2002, 2004, 2006; Raykova et al., 2004; Farina, 2006). The regional group velocity measurements are obtained by frequency-time analysis (Levshin et al., 1989 and references therein) of records from seismic stations and events located in Mediterranean region. Additionally published phase velocity measurements for Rayleigh wave are collected and included in the following processing. The surface-wave tomography method by Yanovskaya and Ditmar (1990) and Yanovskaya (1997) is applied to estimate lateral variations in group- and phase-velocities at properly chosen set of periods (5–80 s for group velocities and 15–150 s for phase velocities) and the spatial resolution of the used data set. The local values of group and phase velocity are calculated on a predetermined grid of 1°×1°, according to the lateral resolution of our data set and a priori independent constraints deduced from existing literature. The local dispersion curves are assembled at each grid point and group velocity dispersion curves are extended to 150 s, using data from global studies (Ritzwoller and Levshin, 1998; Ritzwoller et al., 2002). The cellular dispersion curve is calculated as the average of the local curves at the four corners of each cell and the standard deviation at selected periods is estimated. The period range of the local dispersion curve varies according to data coverage and resolution at the specified period.

We make an inversion of the dispersion data for 15 cells along Geotraverse II and 28 cells along Geotraverse III.

From the cellular dispersion curves, V_s models are retrieved employing the non-linear inversion method known as "hedgehog" (Panza, 1981 and references therein). One of the key parameters in the inversion is the absolute value of the single point error at each specified period and the r.m.s. value for the whole dispersion curve, estimated as 65 % of average single error determined for all selected periods (Raykova and Panza, 2006). In order to optimize the horizontal resolution of the data, the physical properties of the layers down to depths of about 6-8 km are fixed according to a priori and independent information, wherever it is available, like seismic, geophysical and geological data, derived from previous studies in the region. This information is used also in the parameterization of the velocity structures in the depth range from 6-8 km to 350 km, in agreement with the vertical sensitivity of our data set, estimated from the depth distribution of the partial derivative (Urban et al., 1993) of the group and phase velocity curves with respect to V_s. The deeper structure is fixed accordingly with the model published in Du et al. (1998).

For each cell, because of the well-known non-uniqueness of the inverse problem, a set of models fits the dispersion data, all with similar levels of reliability. The tested model is accepted as a solution if the differences of the measured and theoretical velocity values are less than the single error at specified periods and the r.m.s. value for the whole dispersion curve. The inversion procedure tests about 30 000 models per cell and, on average, 10 structures are accepted as the cellular solutions.

An optimized smoothing method (Farina, 2006) is used to define the representative cellular model by a formalized criterion, based on operational research theory (Bryson and Ho, 1975; De Groot-Hedlin and Constable, 1990). LSO fixes the cellular model as the one that has minimal divergence in velocity between neighbouring cells (Farina, 2006; Raykova and Panza, 2006).

The models selected by LSO are appraised according to known geophysical constraints, since the non-linear inversion and the smoothing algorithm give us only a mathematical solution. The results of the non-linear inversion and LSO are shown in Figs. 2b and 3b as a mosaic of the chosen models in each cell: the value of V_s is colour coded for easier visualization of the velocity structure. The range of variability of the interfaces (the boundaries between layers can well be transition zones in their own right) and the velocity and its range of variability is given for most of the parameterized layers (some values are omitted for graphical reasons).

The lateral resolution of our group-velocity tomography varies between 100 km (periods from 10 s to 25 s) to about 400 km (period 80 s) and the phase velocity tomography resolution is about 400 km. The resolution of dispersion measurements is, in fact, only indirectly connected with the lateral variations of the structural models and areas with similar dispersion properties do not necessarily have laterally homogeneous structures as it is shown by numerical tests in Panza and Pontevivo (2004) and Pontevivo and Panza (2006). The introduction of a priori independent information about the crustal parameters significantly improves the resolving power of the tomography data at greater depths, even if some smearing cannot be excluded due to the used methods. In other words the lateral gradients we map are the smallest consistent with our data.

4.2 Description of the tomographic cross sections

The V_s structure along geotraverses II and III is shown in Figs. 2b and 3b, where the regional seismicity (ISC, 2004: on-line bulletin from 1904 to 2004, M≥ 3.0) is plotted as dots. The main features of the cross sections can be summarized as follows.

4.2.1 TRANSMED II section

Cells from 1 to 5 have a typical continental structure with the average velocity in the mantle about 4.35-4.40 km/s; the structure of cell 6 corresponds to the rift zone in the region, the velocity in the mantle varies slowly down to about 200 km of depth and it is relatively low (4.25-4.35 km/s); the cell 10 (central Algerian basin) with thin lid (15-20 km) and low velocity asthenosphere (4.10-4.15 km/s) has an oceanic signature; cells 7 in the Valencia trough and 11 in the southern Algerian basin have transitional character, suggesting to be thin continental lithosphere; the lid under Balearic islands is thin and fast (10-15 km and velocity 4.70-4.80 km/s) the lithosphere beneath north Africa is faster than in the Mediterranean (4.40-4.55 km/s), and its thickness is about 120 km; there is a large contrast in lithospheric thickness (between cells 11 and 12) in north Africa; in the cells 12 to 15 the thick lid is overlying a very low velocity zone that evidences the presence of a well developed asthenospheric

channel; there is no striking evidence for a continuous slab below northern Africa for depths greater than 100-120 km.

4.2.2 TRANSMED III section

Cells 1 and 2 have thick lithosphere with respect to the next cells: about 200 km with V_s velocity 4.40-4.50 km/s; the cells from 1 to 4 have higher velocities in the mantle with respect to the following cells and represent the NW continental part of the geotraverse; it is clearly visible a low velocity channel under all the Mediterranean from cell 5 to cell 18, at depths from 20 km to about 150 km; in the cell 16 a low velocity hot mantle reservoir is well visible, that can be interpreted as the shallow asthenosphere in the backarc basin, and sourcing the oceanic crust flooring the Tyrrhenian basin; there is no evidence for deep hot mantle plume under the Tyrrhenian sea; the mantle, consistent with the presence of a continental lithosphere under cells 20, 21, and 22, is quite different from that of North Africa, as seen along TRANSMED II in cells from 11 to 15 (Fig. 2); in cell 23, more or less coincident with a local maximum of the heat flow (85 kW/m^2 and more in Hurtig et al., 1992), there is a relatively low velocity layer centred at about 100 km of depth, very likely a remnant of the Mesozoic rifting; in cells from 24 to 28 the velocity inversion in the upper mantle is very gentle if not absent and they are consistent with a continental structure, where the lithosphere is quite thick. Seismicity and high velocities correlate quite well in the Apennines and in the Dinaric subduction zone.

5. Discussion

The new tomography highlights several important issues: i) the African lithosphere, shows a well developed, about 50-70 km thick, sandwiched low-velocity layer (4.0-4.3 km/s) between 120-200 km of depth (Fig. 2); ii) this layer, that seems to be a general feature of the North-Central Pan-African Orogenic block (Hazler et al., 2001), is missing or diluted within the backarc basin to the north, where the fast African lid (4.4-4.6 km/s, cells 13 to 15, Fig. 2) is missing, and V_s spreads in the range 4.1-4.3 km/s in the Algerian basin (cells 10-11) and iii)the recent easternmost backarc basin, i.e., the Tyrrhenian sea, is characterized by sub-Moho V_s as low as 3.85 km/s (Fig. 3); this low velocity layer raises from west to east from a depth of about 60 km - 100 km (cells 5 to 13, Fig. 3) to about 25 km - 30 km (cell 16, Fig. 3), where it matches with the northern prolongation of the Marsili basin, which is one of the easternmost active part of the backarc spreading and it is characterized by even lower velocities (Panza et al., 2003, 2006). These anomalously low velocities can be even smaller in the north-south direction if the east-west anisotropy inferred by Margheriti et al. (2003) and Barruol et al. (2004) is taken into account.

In cell 16 (Ischia-Naples), the crust-mantle transition is rather complex and seems to be consistent with a lithospheric doubling where the deeper Moho is at about 20 km and sits on soft mantle, or with the presence of a shallow layer of consolidated magma reservoir. Alternatively the crust-mantle transition can be very shallow, with a ~7 km thick crust overlying a ~5 km thick mantle layer, with V_s

around 4.40 km/sec, sitting on a ~10 km thick very low-velocity mantle, with V_s around 3.50 km/sec, or a thin continental crust is intruded by a 5 km thick layer of consolidated magma. The former interpretation turns out to be consistent with long wavelength Bouguer anomaly modelling (Panza et al., 2006).

At the discontinuity between the upper (low-velocity layer) and the lower asthenosphere, the velocity rapidly increases from about 4.4 km/s to about 4.7 km/s. It is undulated boundary, ranging between 180 km - 250 km in section II (Fig. 2), and slightly deeper, between 220 km - 300 km in section III (Fig. 3). The deepest depression of this discontinuity partly coincides with the overlying Tyrrhenian basin (cells 13 to 16, Fig. 3). The swell of this boundary occurs beneath the Algerian basin (cells 9 to 11, Fig. 2), beneath the Provençal basin (cells 5 to 9, Fig. 3) and along the Apennines and Dinaric subduction zones (cells 17 to 20 and 23-27, Fig. 3).

No evidences of low velocity "fingers" traceable to those reported by Ritter et al. (2001) in the quality factor are seen beneath the Massif Central (cell 1-2, Fig. 3), but this is not surprising since their size is far below the resolving power of our data and the lack of uncertainty estimates about their size makes any comparison very difficult.

6. Conclusions

Beneath the tectonically stable and old Africa plate, the mantle shows a stratification which may represent chemical and physical variations such as fluids concentration and partial melting in the low-velocity zone (LVZ) in the upper asthenosphere (Fig. 4), or phase transition (e.g. Matsukage et al., 2005) or variation of the Mg/Fe ratio in the lower asthenosphere and upper mantle in general (e.g. Scalera et al., 1981), where the velocity increases below about 200 km -250 km. Moving into the younger and active backarc basin of the western Mediterranean, the mantle appears more chaotic, and the 150-200 km deep LVZ is less evident (Fig. 2). There is a much shallower layer, between 30-100 km deep, raising from west to east, right beneath the central part of the easternmost backarc basin (Fig. 3, Tyrrhenian Sea). The presence of such a layer suggests sizeable shallow partial melting, which is feeding the new oceanic crust in the Marsili basin and the related volcano south of the section. No evidence of a deeply rooted plume occurs beneath the backarc basin. The retreat of a slab implies that the upper mantle fills the space left by the removed lithosphere along W-directed subduction zones (Doglioni, 1991). This is in agreement with reconstructions of the mantle structure in the Mediterranean (Panza and Mueller, 1978; Calcagnile and Panza, 1981; Mele et al., 1997; Panza et al., 2003) showing a very thin lithospheric mantle in the hanging-wall of the subduction zone, low Q factor, low seismic velocities and high heat flow (Zito et al., 2003, Cella et al., 2006). The mantle wedge above the slab is inferred to be asthenosphere replacing the subducted lithosphere. Therefore, the "eastward" retreat of the slab should predict a cogenetic flow in the mantle along the same trend (Fig. 4), regardless this flow is generating the retreat or it is a consequence of it (Doglioni et al., 1999). Therefore, according to the kinematic reconstruction and the new tomography shown here, the upper asthenosphere (LVZ), which is well stratified and confined beneath the old northern Africa

continental lithosphere at about 130-200 km, it is rather dispersed in a thicker section between 40-200 km. This setting possibly generated a dilution of the LVZ, and the slower velocities are confined at the top of the mantle in the active backarc (Fig. 3). All these evidences are in favour of shallow upper mantle convection/circulation (e.g. Faccenna et al., 2003) and confirm the "west to east" flow of the mantle relative to the lithosphere, as suggested by the Apennines slab eastward retreat (e.g. Gueguen et al., 1998). This flow is consistent with shear-wave splitting analysis that mostly supports an E-W upper mantle anisotropy (Margheriti et al., 2003; Barruol et al., 2004), indicating olivine crystals preferred orientation induced by mantle flow.

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Fig. 1. Simplified tectonic map of the western-central Mediterranean and adjoining regions (after Carminati and Doglioni, 2004). The positions through time of the subduction zones active in the last 45 Ma are shown. The position of the geological and tomographic transects discussed in the paper are shown as well.

Fig. 2. (a) Lithospheric scale cross-sections simplified and redrawn from the Transmed II geotraverse (Roca et al., 2004). (b) Lithosphere-asthenosphere system along the trace of the TRANSMED II geotraverse: the tomographic cross section was obtained from selected solutions and related seismicity (body waves magnitude greater or equal to 3.0). The chosen shear velocity and its range of variability in km/s are printed on each layer. When the velocity ranges of vertically adjacent layers do not overlap, a hatched rectangle outlines the range of variability of their thicknesses. Numbers in *Italic* denote the velocities in the crustal layers. The hypocenters are denoted by dots. (c) Overlap between the TRANSMED II regional cross-sections and the V_s tomography; only a few average representative values of V_s are reported to avoid overcrowding of characters; the dashed line reproduces the lithosphere-asthenosphere boundary given in part (a), while the new geometry of the base of the lithosphere (blue) and of the limit between upper and lower asthenosphere (red) are shown by continuous lines. A well-developed low velocity layer under Africa is visible to the right.

Fig. 3. (a) Lithospheric scale cross-sections simplified and redrawn from the TRANSMED III geotraverse (Carminati et al., 2004). (b) Lithosphere-asthenosphere system along the trace of the TRANSMED III geotraverse: the tomographic cross section was obtained from selected solutions and related seismicity (body waves magnitude greater or equal to 3.0). The chosen V_s and its range of variability in km/s are printed on each layer. When the velocity ranges of vertically adjacent layers do not overlap, a hatched rectangle outlines the range of variability of their thicknesses. Numbers in *Italic* denote the velocities in the crustal layers. The hypocenters are denoted by dots. (c) Overlap between the TRANSMED III regional cross-sections and the V_s tomography; only a few average representative values of V_s are reported to avoid overcrowding of characters; the dashed line reproduces the lithosphere-asthenosphere boundary given in part (a), while the new geometry of the base of the lithosphere (blue) and of the limit between upper and lower asthenosphere (red) are shown by continuous lines. The well-developed low velocity layer, visible under Africa in Fig. 2, is either missing or dispersed in the shallower asthenosphere, particularly in the centre of the back-arc basins.

Fig. 4. Hypothetical, schematic reconstruction of the upper mantle stratification (upper section) and later mobilization into the back-arc basin (lower section).
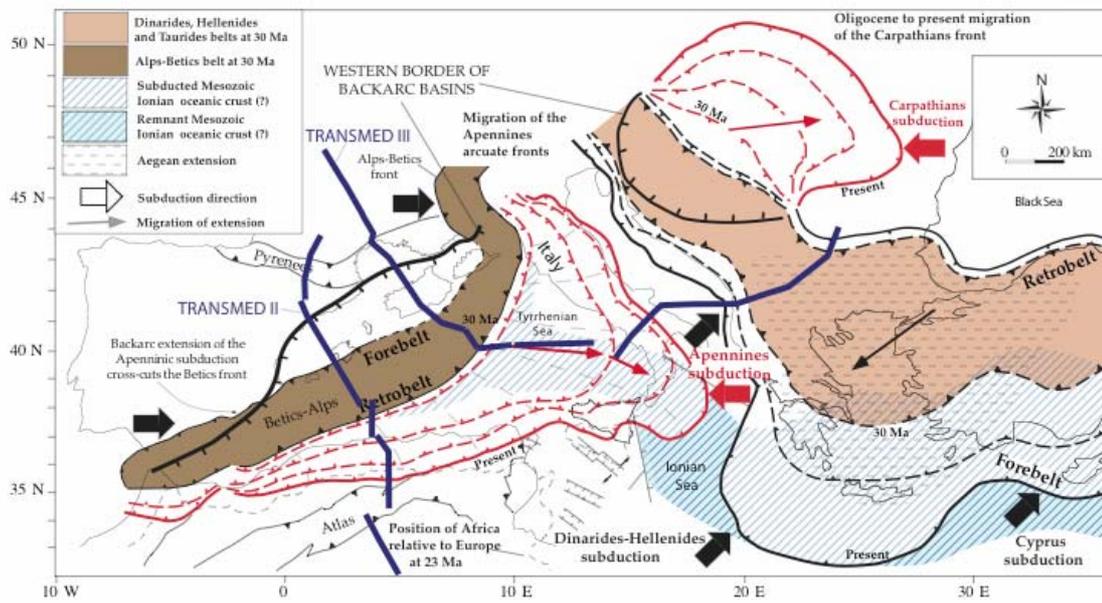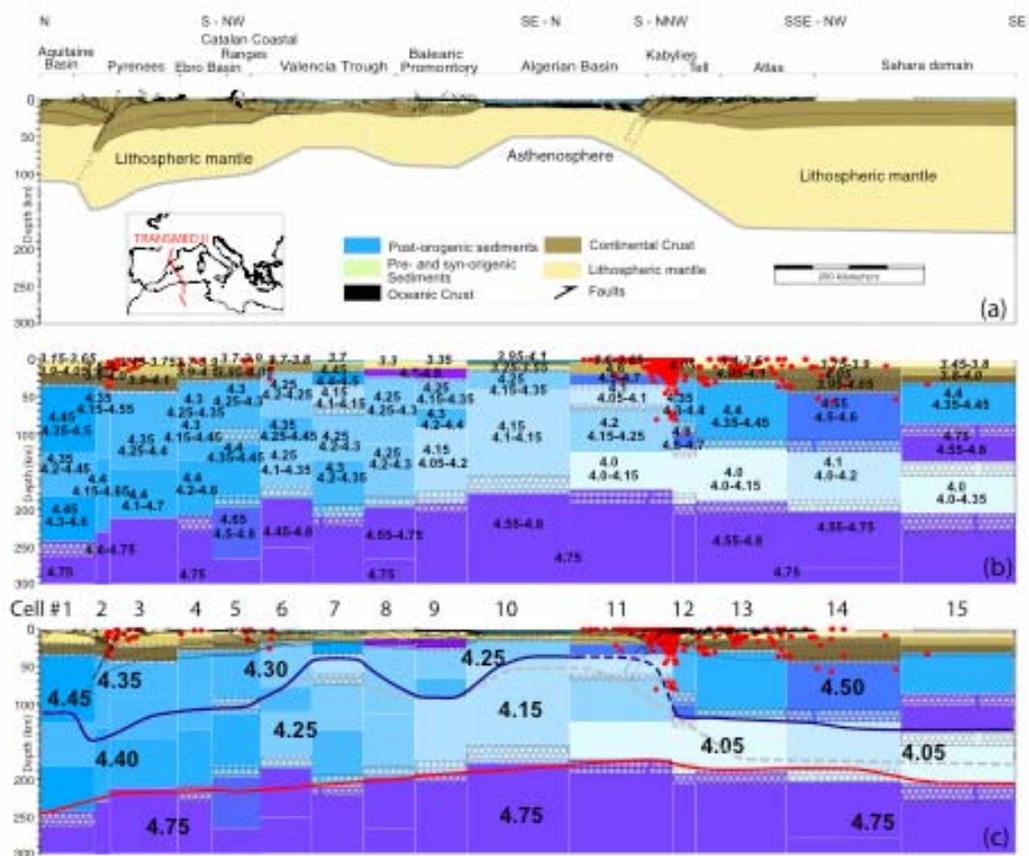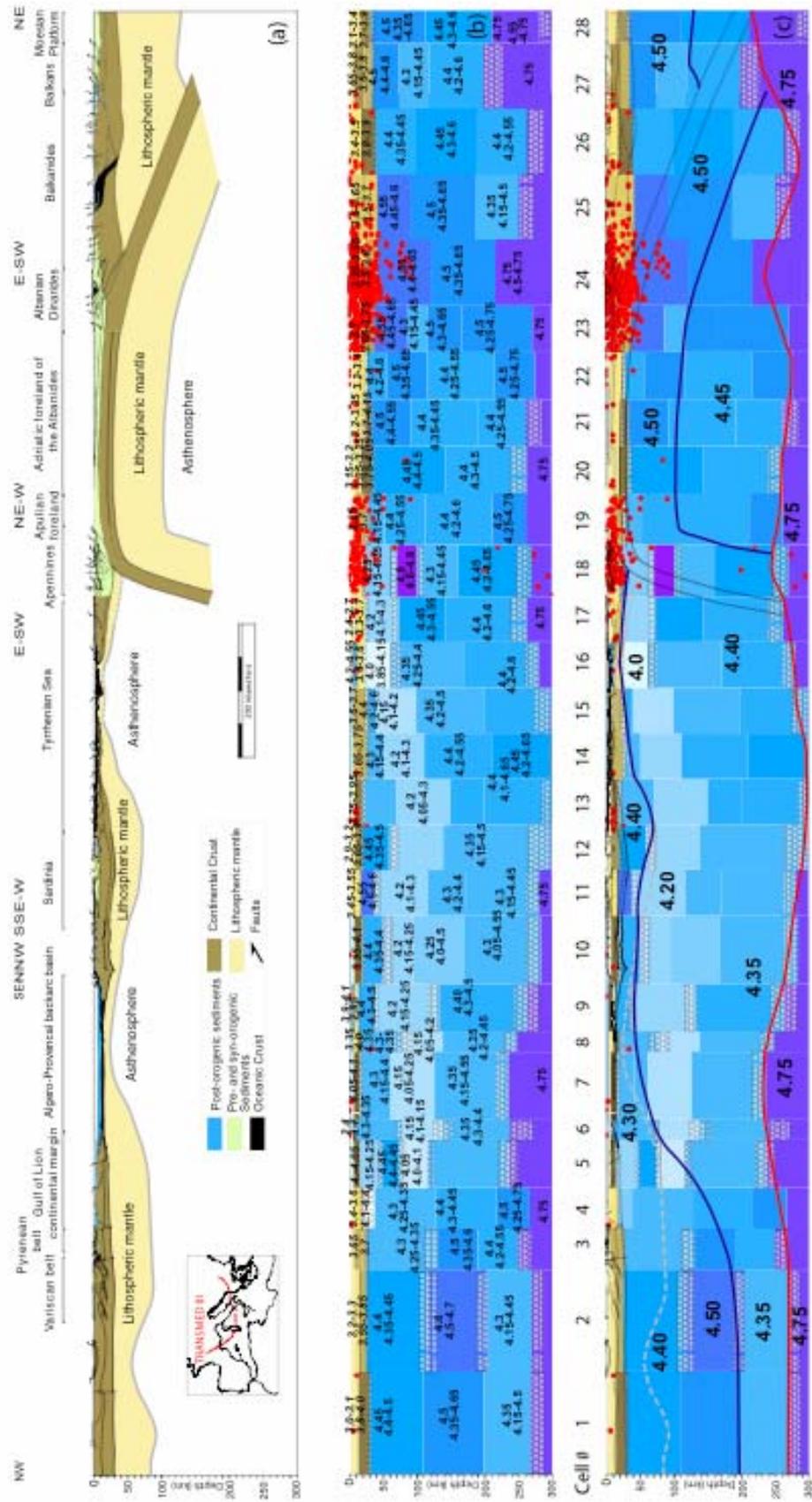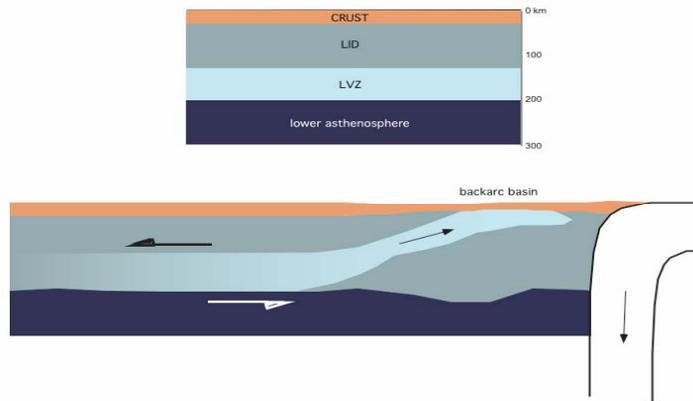
Fig.1

Fig. 2

Fig. 3

CRUST
LID
LVZ
lower asthenosphere

0 km
100
200
300

backarc basin

Fig. 4