

Upper mantle flow in the western Mediterranean

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Received 20 July 2006; received in revised form 18 January 2007; accepted 18 February 2007

Available online 27 February 2007

Editor: R.D. van der Hilst

Abstract

Two cross-sections of the western Mediterranean Neogene-to-present back-arc 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 back-arc 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 back-arc basin. These findings suggest upper mantle circulation in the western Mediterranean back-arc basin, mostly easterly-directed and affecting the boundary between upper asthenosphere (LVZ) and lower asthenosphere, which undulates between about 180 km and 280 km.

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Keywords: shear-wave velocity model; back-arc basin; mantle flow; Mediterranean geodynamics; mantle stratification

1. Introduction

There usually 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 sections [1], i.e., II [2] and III [3] (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 back-arc 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. 4a and 5a) with a new shear-wave tomography of the area (Figs. 4b and 5b).

The back-arc spreading in the western Mediterranean developed since 30 Ma and gradually shifted from west to east, moving from the Provençal, Valencia and Alboran basins, to the Algerian and Tyrrhenian basins,

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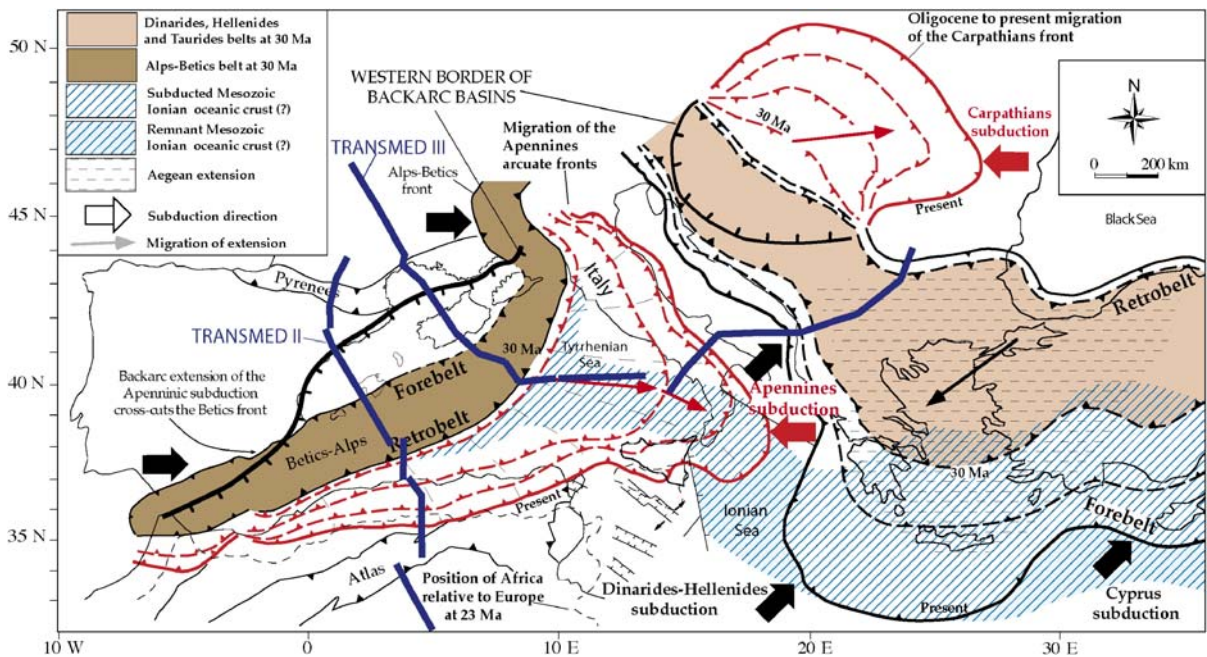


Fig. 1. Simplified tectonic map of the western–central Mediterranean and adjoining regions (after [5]). The positions through time of the subduction zones active in the last 45 Ma are shown. The position of the geological and tomographic transects discussed in the paper are shown as well.

[e.g., 4,5]. 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 [6]. 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 sections. 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. 4a and 5a).

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 TRANSMED sections, 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 in a new model of the mantle flow in a back-arc 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 [2,3,4,7].

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 mostly since the Cenozoic. For this reason the description of the evolution will range from Eocene to Recent times.

Paleomagnetic studies [8] 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 an intervening microplate between the European and African plates [6]. Contemporaneous shortening occurred in the Pyrenees (until ~ 24.7 Ma according to a magnetostratigraphic study [9]) 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 [6].

The Apennines–Maghrebides subduction zone was characterized, since the beginning, by fast radial eastward rollback, as evidenced by the migration of the subduction related calcalkaline 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 in the back-arc 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 [10]. 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 [7]. This hypothesis was accepted by [2] and no subducting slab is imaged beneath northern Africa in the TRANSMED II section. However, along this section, the slab could alternatively be missing because stretched and shifted eastward during the rollback.

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.

However, 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 [6].

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, plus the oceanic part of the present day 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-northeast dipping subduction of the Adriatic lithosphere beneath the Dinarides started in the Mesozoic and in the Cenozoic it 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. 4a and 5a is in [2,3]; 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. Extended references are given in [2,3].

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 [11] and by the ESCI–Catalanides deep seismic reflection profile [12] respectively and by refraction and gravimetric data, e.g. [13], by wide-angle reflection data [14], by magnetotelluric modelling [15] as well as by seismic tomography data [16]. The lithosphere–asthenosphere boundary was

drawn from integrated lithospheric models combining topography, gravity, and heat flow [17].

In the València Trough and Balearic promontory, the Moho depth was based on the interpretation of the ESCI-València trough deep seismic reflection profile [14] but regional refraction and gravimetric data [12,13] were also taken in account. The lithospheric mantle thickness was derived from both gravimetric and geoid anomaly models [18] and from the integrated model of [17] and from an integrated model by [2].

The oceanic nature of the Algerian Basin crust and its thickness were constrained by seismic refraction results [19] and by aeromagnetic data [20]. The lithosphere–asthenosphere boundary was calculated specifically for the TRANSMED Project [2] 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 [21] whereas the lithospheric thickness and geometry was based on an original integrated model [2].

3.1.2. TRANSMED III section

In the Massif Central, the Moho was drawn from refraction data [22], while lithospheric geometries were based on tomography results [23] and on the results of mantle thermal modelling [24]. The structure of the Moho below the Gulf of Lion platform and slope was based on deep seismic reflection data [25] and gravity modelling [26].

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 [27] and by seismic refraction results [25]. The base of the lithosphere was drawn from the gravity models of [28].

Below Sardinia, Moho depth and lithospheric thickness were drawn according to the results of [29,30] and on the results of the European Geotraverse project [31]. The asymmetric lithosphere boundary under Sardinia was based on the gravity modelling of [28].

In the Tyrrhenian Sea, the depth of the Moho was derived from [29] partly modified by Carrara (2002, unpublished results), while the geometry of the lithosphere–asthenosphere boundary was based on the models of [30,32].

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 [29,30,32]. The location of the Apenninic

subducting slab (subcrustal seismicity is rare), constrained by the studies of [33,34], is consistent with the occurrence of positive Bouguer anomalies (up to 120 mGal or more [35]) and very high heat flow values (up to 140 mW/m² or more [36]) 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 [37] and geochemical studies [38].

Below the Albanian Dinarides, the Moho depth was constrained by gravimetric data [39]. The geometry of the slab subducting beneath Albania (subcrustal seismicity is also absent) was constrained by tomographic studies showing a clear and fast velocity body (interpreted as the evidence of the active subduction process) under the Albanian Dinarides [40].

Below the Macedonian Balkanides, the depth of Moho discontinuity was both based on two deep seismic profiles and on magnetic [41] and gravity [42] 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 [43,44].

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 thin 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 [45] 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 [46], 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–Maghrebides fold-and-thrust belt and reach maximum thicknesses of 40 km and 170–180 km, respectively, below the Sahara Atlas (an intraplate right-lateral transpressive inversion structure) 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 [23], but its deep origin is still under debate.

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 [28,47] 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 to 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 Adriatic 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 Adriatic 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 east-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 Eocene and 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 a part of a series of regional tomography investigations made in the Mediterranean area [34,48,49,50] (unpublished results of Raykova, 2004 and Farina, 2006). We use surface waves since, in the period range considered (about 5–150 s), they sample the elastic properties of the crust and upper mantle and this kind of waves is particularly suitable to investigate regions where a velocity inversion (decrement of velocity with increasing depth) can be present. The regionalization of the dispersion of surface waves by 2D tomography and the inversion of local dispersion curves by non-linear “hedgehog” method, taking into account the detailed information available about the structure of the upper layers of the crust, is a very perspective procedure to obtain high-resolution shear-wave velocity models, and their uncertainties, of the crust and upper mantle. Our model supplies a due revision of the results by [51] who have produced a tomographic S-wave model for the same region that is biased by a methodological problem, discussed in detail by [52].

All relevant group velocity measurements available in the literature for different study areas, like Spain [53], Italy [32,49], Greece [54], Alps [55], and Balkan

Peninsula [50,56], have been considered. In addition many pertinent seismic records from VBB and BB stations available on-line at data centres as IRIS (www.iris.washington.edu), ORFEUS (orfeus.knmi.nl), GEOFON (www.gfz-potsdam.de/geofon), GEOSCOPE (www.geoscope.ipgp.jussieu.fr), MEDNET (mednet.ingv.it), GRNS (www.szgrf.brg.de), and CRSN (www.ig.cas.cz) have been processed. The hypocentre information for the analysed events is taken from on-line bulletins of NEIC (neic.usgs.gov) and ISC (www.isc.ac.uk). The location of stations and earthquake epicentres are selected in order to have as homogenous as possible path coverage of the area under investigation. The length of paths-considered does not exceed 2500 km in order to have as much as possible usable measurements of group velocity at short periods. The regional group velocity measurements of Rayleigh wave fundamental mode are obtained by frequency–time analysis (FTAN) [57] (and references therein). More than 1000 records are analysed to assemble the data set of about 800 dispersion curves, which ensures, in the period range from 5–10 to 50–80 s, a relatively good coverage of the study region. One example of such measurements by FTAN is given in Fig. 2.

To increase the penetration depth of the inverted data set we consider published phase velocity measurements (extended references can be found in [49,53,58]) for

Rayleigh waves with periods from about 15–20 s to about 150 s.

A surface-wave tomography method [59,60] 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). The method also evaluates the resolution of the data, defined as the average size of the equivalent smoothing area and its elongation [60] and hence it is not necessary to perform check board or similar tests. The results of tomography (path, group-velocity and resolution distributions) for Rayleigh wave fundamental mode, at a period of 20 s, is shown in Fig. 3. The lateral resolution of our tomography varies between 120 km (periods from 15 to 25 s) to about 400 km (periods from 50 to 80 s) for group velocities and between about 200 km (periods from 20–35 s) to about 600 km for phase velocities (periods from 100 to 150 s). The resolution of dispersion measurements is only indirectly connected with the lateral variations of the structural models and areas with similar dispersion properties do not necessarily have laterally homogeneous structures. Therefore the introduction of a priori independent information about the crustal parameters [58,61] significantly improves the resolving power of the tomography data also at mantle depths, even if some smearing cannot be excluded at the greatest depths.

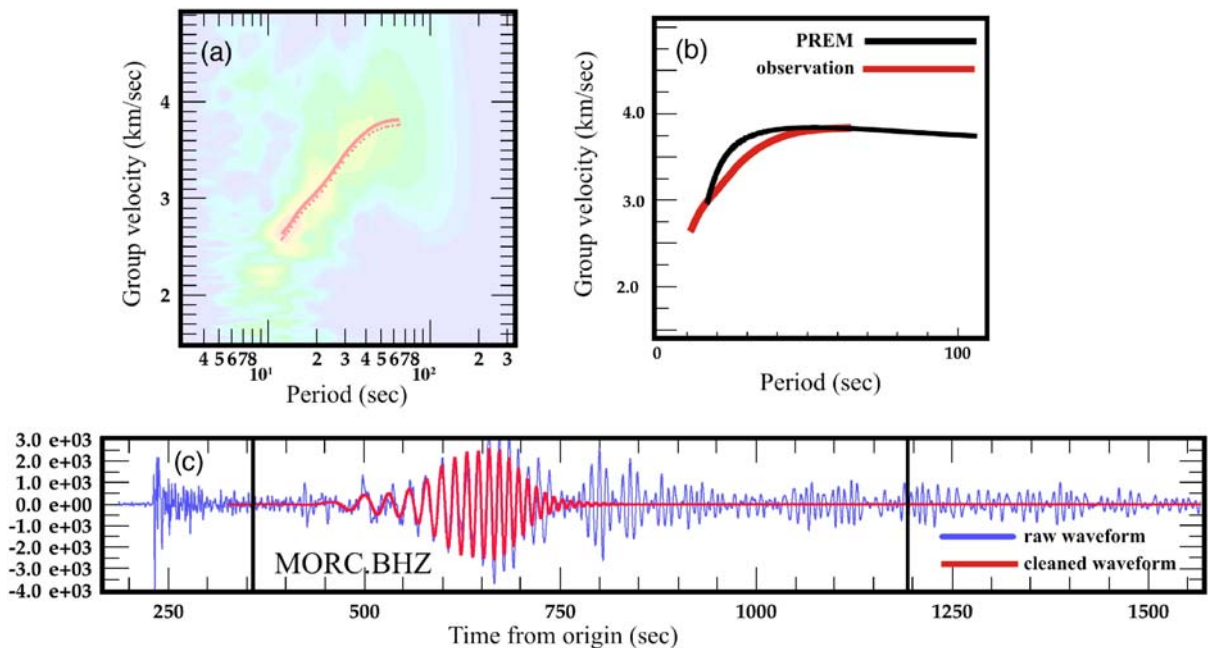


Fig. 2. Example of FTAN analysis of a seismogram recorded at station MORC (CZ) for an earthquake that occurred in Algeria (36.6° N 4.77° E10, November 2000, mb magnitude 5.8 by ISC): (a) FTAN diagram of the "raw" waveform; (b) comparison between observed (red line) and predicted by PREM model (black line) group velocity dispersion curve of the Rayleigh wave; (c) comparison between raw waveform (blue line) and extracted waveform of fundamental Rayleigh wave (red line), which corresponds to the dispersion curve that will be used in the 2D tomography.

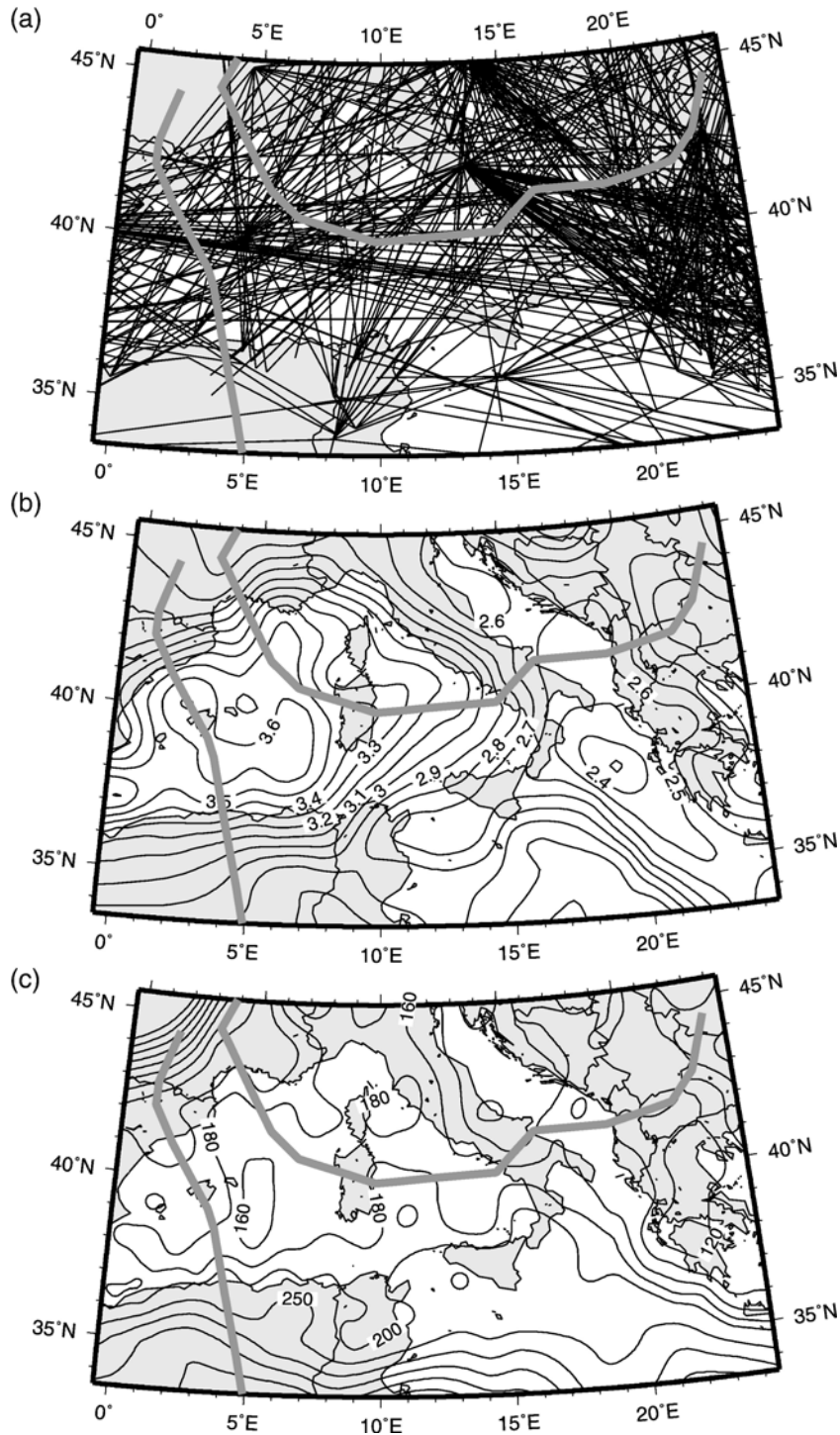


Fig. 3. Results of Rayleigh waves tomography for period 20 s: (a) station-to-epicentre paths; (b) group velocity distribution in km/s; (c) data resolution distribution as mean size of averaging area in km. TRANSMED II and TRANSMED III are denoted by thick grey lines.

The local values of group and phase velocity are calculated on a predetermined grid of $1^\circ \times 1^\circ$, 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 [62]. The cellular dispersion curve is calculated as the average of the local curves at the four corner grid points of each cell, and the standard deviation at selected periods is estimated. The period range of the cellular 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 TRANSMED II and 28 cells along TRANSMED III.

From the cellular dispersion curves, V_s models are retrieved employing the non-linear inversion method known as “hedgehog” [63] (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 [48,49]. The value of the group velocity at 80 s is calculated as an average between our tomography results and data from published global tomography [62] with resulting average standard deviation of 0.03 km/s. 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. When the cell is located in a sea region, the thickness of the water layer is defined (data from NGDC, www.ngdc.noaa.gov) as a weighted average of all bathymetric measurements inside the area of the cell. The thickness of the sediments is defined, in a similar way, from the map of [64] or the data in [65]. The properties of the remaining upper crustal layers are taken from published information and some extended references are given in [32,48,49,50,53,56]. This information is used also in the parameterization of the velocity structures (such as range of the Moho depth, range of the shear-wave velocity in crustal and mantle layers) 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 [66] of the group and phase velocity curves with respect to V_s . The deeper structure is fixed accordingly with the model published in [67].

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 [48,49] (Farina, 2006, unpublished results) is used to define the representative cellular model by a formalized criterion, based on operational research theory [68,69]. LSO fixes the cellular model as the one that has minimal divergence in velocity between neighbouring cells [48].

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. 4b and 5b 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).

4.2. Description of the tomographic cross-sections

The V_s structure along TRANSMED sections II and III is shown in Figs. 4b and 5b, where the regional seismicity (International Seismological Centre, on-line bulletin from 1904 to 2004, $M \geq 3.0$) 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

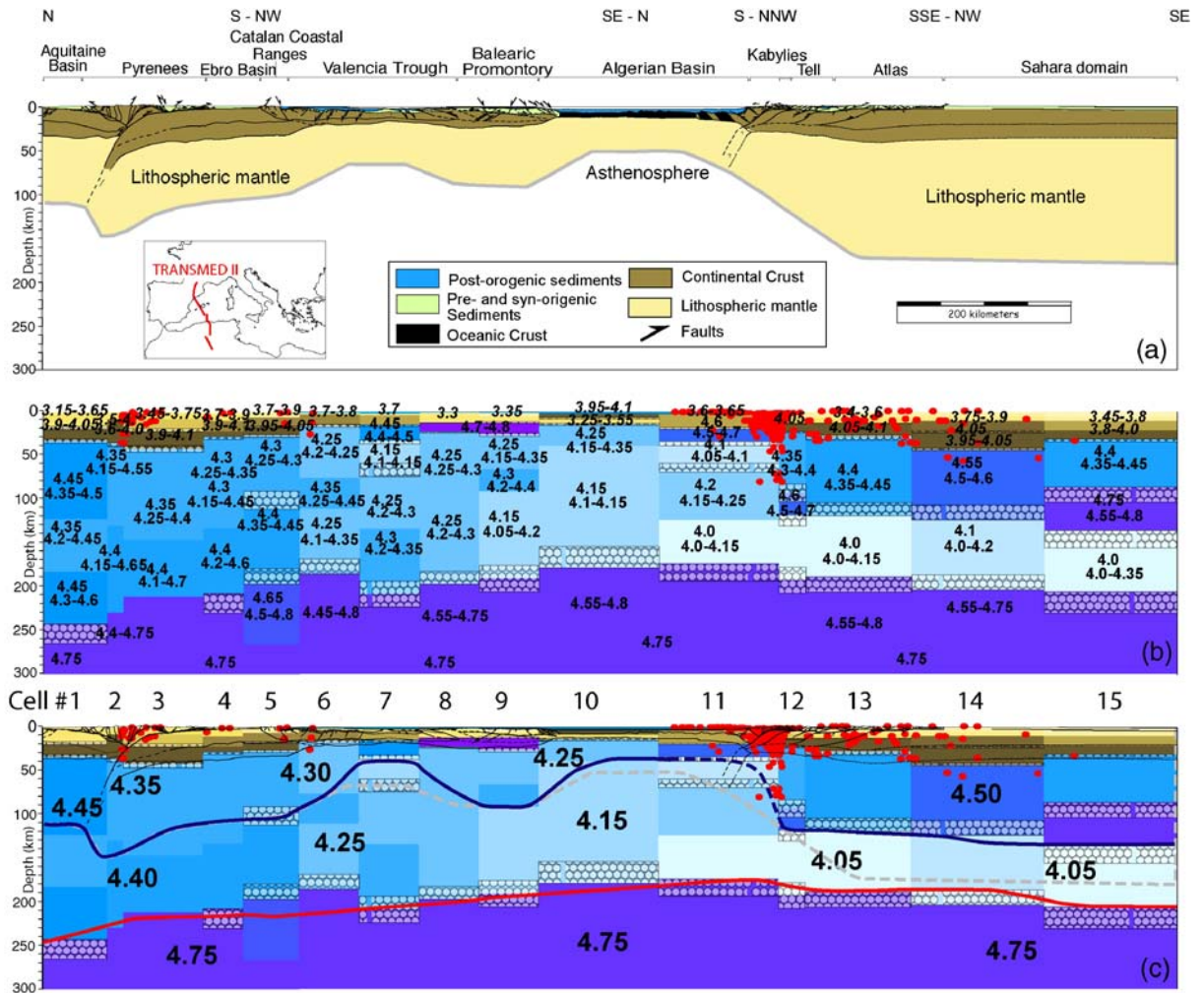


Fig. 4. (a) Lithospheric scale cross-sections simplified and redrawn from the TRANSMED II geotraverse [2]. (b) Lithosphere–asthenosphere system along the trace of the TRANSMED II geotraverse: the tomographic cross-section was obtained from selected solutions and related seismicity (body waves magnitude 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 hypocentres are denoted by red 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.

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 section; it is clearly visible in 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 back-arc 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. 4); in cell 23, more or less coincident with a local maximum of the heat flow (85 kW/m² and more in [70]), 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. 4); ii) this layer, that seems to be a general feature of the North-Central Pan–African Orogenic block [71], is missing or diluted within the back-arc basin to the north, where the fast African lid (4.4–4.6 km/s, cells 13 to 15, Fig. 4) 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 back-arc basin, i.e., the Tyrrhenian sea, is characterized by sub-Moho V_s as low as 3.85 km/s (Fig. 5); this low velocity layer raises from west to east from a depth of about 60 km–100 km (cells 5 to 13, Fig. 5) to about 25 km–30 km (cell 16, Fig. 5), where it matches with the northern prolongation of the Marsili basin, which is one of the easternmost active parts of the back-arc spreading and it is characterized by even lower velocities [34,49]. These anomalously low velocities can be even smaller in the north–south direction if the east–west anisotropy inferred by [72,73] 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/s, sitting on a ~ 10 km thick very low-velocity mantle, with V_s around 3.50 km/s, 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 [49].

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. The boundary is undulated, ranging between 180 km–250 km in section II (Fig. 4), and slightly deeper, between 220 km–300 km in section III (Fig. 5). The deepest depression of this discontinuity partly coincides with the overlying Tyrrhenian basin (cells 13 to 16, Fig. 5). The swell of this boundary occurs beneath the Algerian basin

(cells 9 to 11, Fig. 4), beneath the Provençal basin (cells 5 to 9, Fig. 5) and along the Apennines and Dinaric subduction zones (cells 17 to 20 and 23–27, Fig. 5).

No evidences of low velocity “fingers” traceable to those reported by [74] in the quality factor are seen beneath the Massif Central (cell 1–2, Fig. 5), 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. 6), or phase transition [75] or variation of the Mg/Fe ratio in the lower asthenosphere and upper mantle in general [76], where the velocity increases below about 200 km–250 km. Moving into the younger and active back-arc basin of the western Mediterranean, the mantle appears more chaotic, and the 150–200 km deep LVZ is less evident (Fig. 4). There is a much shallower layer, between 30–100 km deep, raising from west to east, right beneath the central part of the easternmost back-arc basin (Fig. 5, 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 [49]. No evidence of a deeply rooted plume occurs beneath the back-arc basin. The retreat of a slab implies that the upper mantle fills the space left by the removed lithosphere along W-directed subduction zones [10]. This is in agreement with reconstructions of the mantle structure in the Mediterranean [34,37,77,78] showing a very thin lithospheric mantle in the hanging-wall of the subduction zone, low Q factor, low seismic velocities and high heat flow [47,79]. 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 contemporaneous flow in the mantle along the same trend (Fig. 6), regardless this flow is generating the retreat or it is a consequence of it [6]. 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 the back-arc of the Apennines subduction, having a thicker section between 40–200 km. This setting possibly generated a dilution of the LVZ, and the slower

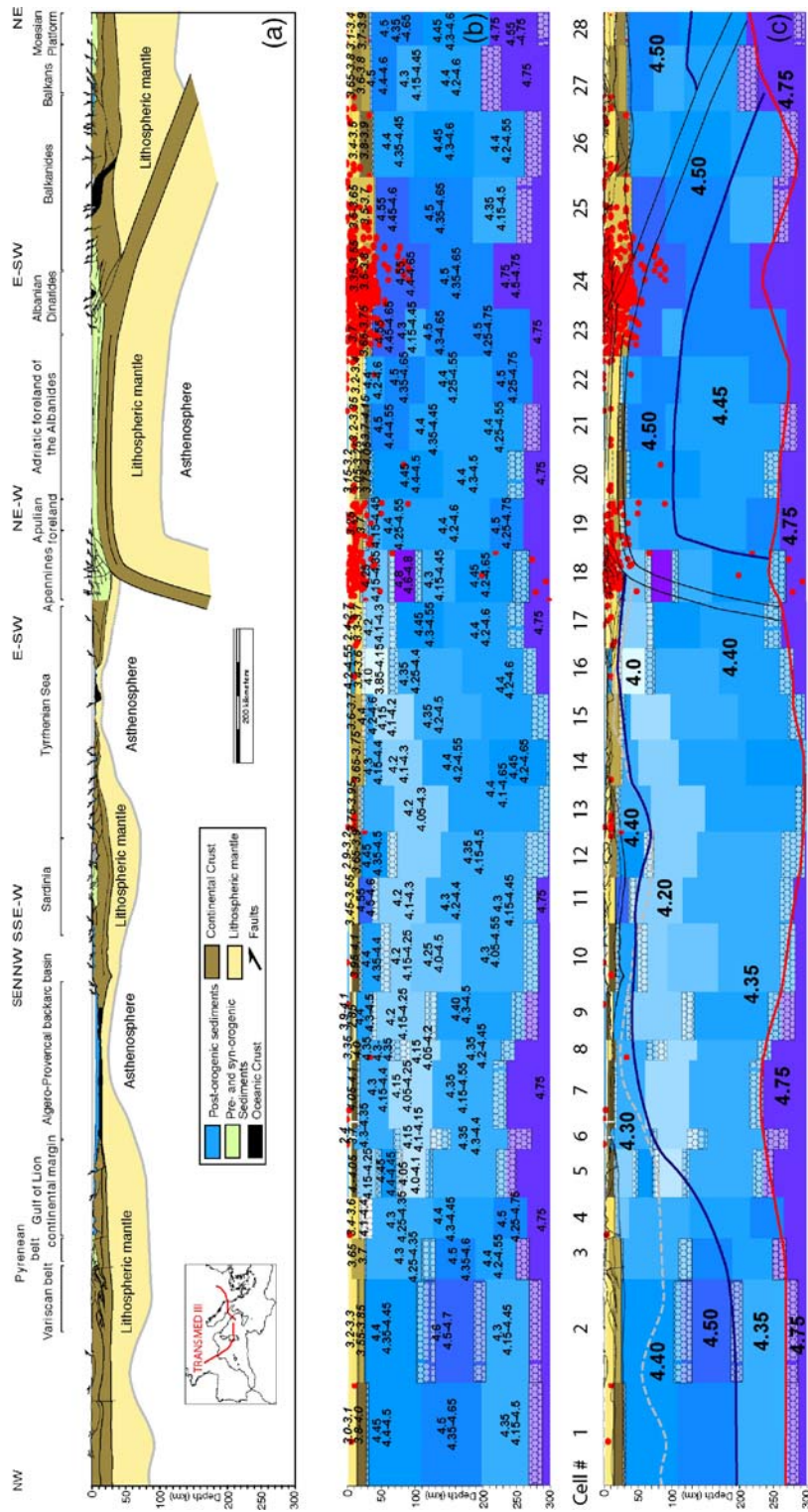


Fig. 5. (a) Lithospheric scale cross-sections simplified and redrawn from the TRANSMED III geotraverse [3]. (b) Lithosphere–asthenosphere system along the trace of the TRANSMED III geotraverse: the tomographic cross-section was obtained from selected solutions and related seismicity (body waves magnitude 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 hypocentres are denoted by red 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. 4, is either missing or dispersed in the shallower asthenosphere, particularly in the centre of the back-arc basins.

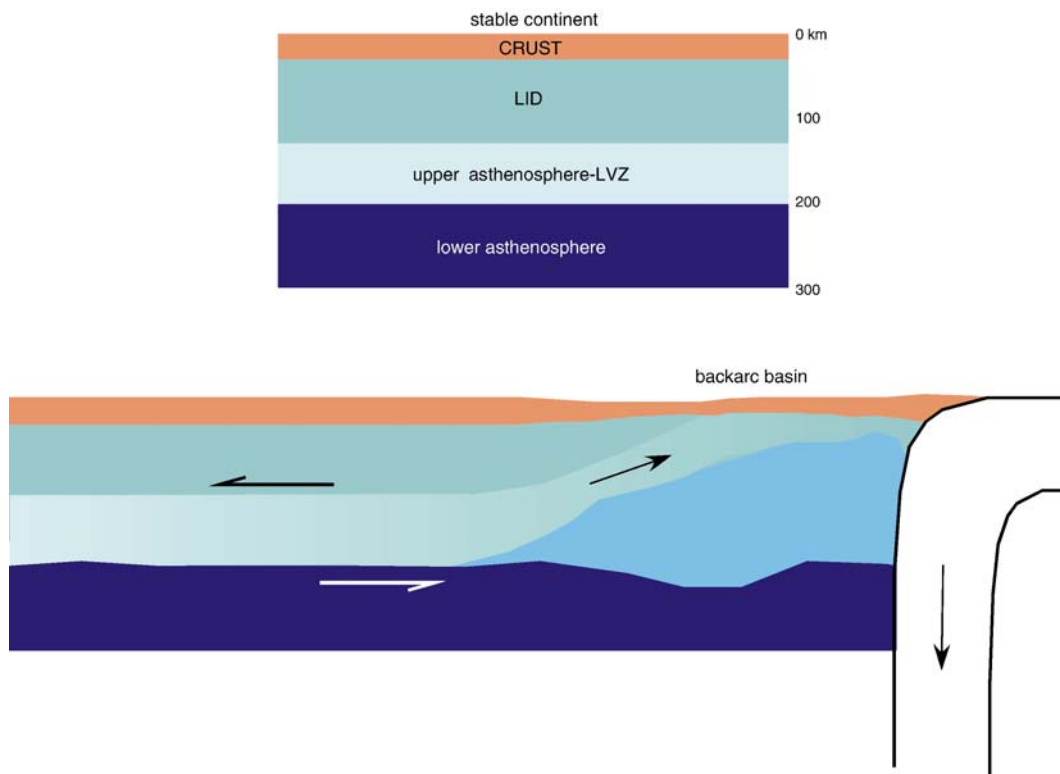


Fig. 6. Hypothetical, schematic reconstruction of the upper mantle stratification (upper section) and later mobilization and dispersion into the back-arc basin (lower section).

velocities are confined at the top of the mantle in the active backarc (Fig. 5). All these evidences are in favour of shallow upper mantle convection/circulation [80] and confirm the “west to east” flow of the mantle relative to the lithosphere, as suggested by the Apennines slab eastward retreat [4]. This flow is consistent with shear-wave splitting analysis that mostly supports an E–W upper mantle anisotropy [72,73], indicating olivine crystals preferred orientation induced by mantle flow.

Acknowledgements

We express our thanks to A. Levshin, who provided global group velocity measurements available at CIEI, University of Colorado at Boulder, to T. Yanovskaya, who provided the 2D tomography software from St. Petersburg State University. This research was sponsored by ICTP “Abdus Salam” Associates Scheme (R. Raykova) and by CEI University network project “Geodynamical model of central Europe for safe development of ground transportation systems”. The manuscript was made during the stay of R. Raykova at Istituto Nazionale di Geofisica e Vulcanologia, sezione di Bologna as a post-doc SPICE researcher, supported from the European Commission’s

Human Resources and Mobility Programme, Marie Curie Research Training Networks. Part of this research is funded by Italian MIUR PRIN 2004: Modelling of the mantle–crust system by petrological and geophysical methods; PNRA 2.7–2.8, 2004–2006 “Sismologia a banda larga nella regione del Mare di Scotia e suo utilizzo per lo studio della geodinamica della litosfera”; ALPS-GPSQUAKENET Project (I/II/3.3/02) of the EU Interreg III-B Alpine Space Programme.

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