

Teleseismic tomography of the southern Tyrrhenian subduction zone: New results from seafloor and land recordings

Caterina Montuori,¹ Giovanni Battista Cimini,¹ and Paolo Favali^{1,2}

Received 21 October 2005; revised 16 June 2006; accepted 28 August 2006; published 31 March 2007.

[1] Teleseismic traveltime data, recorded by temporary ocean bottom seismographs deployed in Tyrrhenian Sea around the Aeolian Islands (Tyrrhenian Deep-sea Experiment (TYDE)), have been used for the first time in Italy to refine the 3-D model for the deep *P* wave velocity structure of the southern Tyrrhenian subduction zone. The arrival times of 35 teleseisms have been combined with those recorded by the Italian National Network. In order to obtain a more complete azimuthal coverage of teleseismic rays, 80 events recorded by land stations from 1990 to 2002 have been included in the data set. In total, 2904 *P* and 314 PKPdf phases, 1300 recorded by ocean bottom instruments, have been collected. The upper mantle structure is reconstructed down to 500 km by a nonlinear inversion of the relative residuals computed with respect to the reference 1-D velocity model ak135. The obtained tomographic model has a higher resolution than those previously published thanks to the recordings of TYDE seafloor stations. Tomographic results confirm the presence of the Tyrrhenian slab imaged as a high-velocity body extending from the uppermost mantle down to the bottom velocity model with dip 70–75° NW. The model better defines the geometry of the seismogenic part of the slab. Its lateral extension is about 200 km in the depth interval 150–300 km, where most of the deep seismicity is concentrated. At uppermost mantle depths the fast structure has smaller lateral dimensions (about 100 km). The inversion also points out a wide well-resolved low-velocity zone completely surrounding the steeply dipping fast structure from the lower crust down to about 300 km. This feature suggests the presence of a three-dimensional circulation of asthenospheric flow around the Ionian slab caused by retreat and roll-back of the slab. Our results are in agreement with recent laboratory experiments, mantle anisotropy studies, geochemical and isotopic analyses, and modeling based on residual topography.

Citation: Montuori, C., G. B. Cimini, and P. Favali (2007), Teleseismic tomography of the southern Tyrrhenian subduction zone: New results from seafloor and land recordings, *J. Geophys. Res.*, 112, B03311, doi:10.1029/2005JB004114.

1. Introduction

[2] The lithosphere and asthenosphere structure of the Tyrrhenian basin have been analysed by several authors using different approaches, such as gravimetric methods [Morelli, 1970, 1997], seismic exploration [Finetti and Del Ben, 1986; Pascucci *et al.*, 1999], and dragging and volcanological studies [e.g., Kastens *et al.*, 1988; Francalanci *et al.*, 1993]. Most of the published studies pointed out a slab beneath the Marsili Basin, subducting the Ionian lithosphere below the Calabrian Arc [e.g., Anderson and Jackson, 1987; Giardini and Velonà, 1991; Selvaggi and Chiarabba, 1995]. The presence of this slab is supported by the occurrence of intermediate and deep earthquakes, which define a continuous, well-documented Wadati-Benioff zone from the surface to at least 500 km depth [e.g., Caputo *et al.*, 1970,

1972; Anderson and Jackson, 1987; Giardini and Velonà, 1991; Selvaggi and Chiarabba, 1995; Cimini and Marchetti, 2006]. The oceanic nature of the subducted lithosphere is derived by heat flow measurements [e.g., Della Vedova *et al.*, 1984], calc-alkaline magmatism of the Aeolian Islands [e.g., Barberi *et al.*, 1973], and crustal thickness and seismic reflection profiles in the Ionian Sea [e.g., De Voogd *et al.*, 1992; Catalano *et al.*, 2001].

[3] In the last decades, many tomographic studies have been performed to define the seismic velocity field in the crust and upper mantle beneath the Tyrrhenian-Apennine system in order to provide constraints to the geodynamic evolution of the western Mediterranean Sea [e.g., Scarpa, 1982; Spakman, 1990; Amato *et al.*, 1993a, 1993b; Cimini and Amato, 1993; Piromallo and Morelli, 1997; Lucente *et al.*, 1999; Cimini, 1999; Di Stefano *et al.*, 1999; Cimini and De Gori, 2001; Piromallo and Morelli, 2003; Spakman and Wortel, 2004]. All tomographic models, using high-quality teleseismic delay times [e.g., Amato *et al.*, 1993b] or both regional and teleseismic events reported by International bulletins [e.g., Spakman *et al.*, 1993], highlighted a contin-

¹Istituto Nazionale di Geofisica e Vulcanologia, Roma, Italy.

²Also at Università "La Sapienza", Roma, Italy.

uous subvertical cold oceanic slab underneath Calabrian Arc dipping 70–75° NW down to 400 km depth. Below this depth, a subhorizontal deflection of the slab is observed jointly with the northwestward shift of the deep seismicity. The subcrustal seismicity, located between 40 and 300 km depth, correlates well with the strong high-velocity anomaly imaged by tomography.

[4] However, the lateral resolution of these tomographic models is limited by the absence of offshore seismic stations. Recently, some authors [e.g., *Zhao et al.*, 1997; *Ito et al.*, 2000] obtained a highly reliable tomographic reconstruction of subducting systems, in which the lithosphere dips downward toward offshore areas, integrating land based and ocean bottom seismic stations.

[5] In this paper we present a new 3-D model for the deep *P* wave velocity structure of the southern Tyrrhenian Sea obtained by adding teleseismic waveforms recorded by ocean bottom instrumentation during the Tyrrhenian Deep-sea Experiment (TYDE) [*Dahm et al.*, 2002] to a high-quality data set of events recorded by the permanent stations of the Italian National Network (RSNC) during the last fifteen years. The seafloor temporary seismic stations (December 2000 to May 2001) provided a unique opportunity to better image the three-dimensional shape and size of the slab in the first 200 km depth, and the mantle wedge velocity structure, which has been scarcely resolved by previous investigations. Moreover, we find and interpret a clear pattern of low-velocity anomalies completely surrounding the high-velocity body.

2. Tectonic Setting

[6] The present-day tectonic setting of the Tyrrhenian-Apennine system is the result of the complex and slow convergence (0.5–0.8 cm/yr) between Eurasian and Africa plates, that was active since at least 65 Ma [*Dercourt et al.*, 1986; *Dewey et al.*, 1989; *Patacca et al.*, 1992]. After the Alpine orogenesis (Eocene-Oligocene), the geodynamic evolution of the Tyrrhenian-Apennine system was driven by the eastward migration of the subduction hinge [*Malinverno and Ryan*, 1986]. The Adriatic lithosphere, subducting westward beneath the European plate, underwent a gravitative roll-back which generated both a shortening of the Adriatic prism accretionary wedge (Apennine Orogen) and rifting in the back-arc region (Tyrrhenian Basin).

[7] During the late Oligocene–early Miocene (~26 Ma) began a phase of rifting that led to the opening of the Ligurian-Provençal Basin as result of the roll-back toward southeast of an older northwest dipping subduction zone [e.g., *Faccenna et al.*, 1996; *Rosenbaum and Lister*, 2004]. The opening of this basin was accompanied by a counter-clockwise rotation of the Corsica-Sardinia together with the former western margin of Adria (Apulia) and gave rise to the formation of the Apennines [*Patacca and Scandone*, 1989; *Patacca et al.*, 1992]. During this tectonic phase the Apennine accretionary wedge underwent shortening. The roll-back of the subducting Ionian slab caused the opening of the southern Tyrrhenian Basin since Tortonian (10 Ma). At that time the lithospheric rifting separated the Calabria block from the Sardinia basement. This event led to the formation of new oceanic crust generating westward the

Vavilov Basin (4.3–2.6 Ma, *Sartori* [1989]; 8.5–4.5 Ma, *Argnani*, [2000]) and then southeastward the Marsili Basin (1.6 Ma, *Kastens et al.* [1988]; 2.0–1.7 Ma, *Argnani* [2000]). The highest rate of opening of the southern Tyrrhenian Basin was estimated in 6 cm/yr by *Faccenna et al.* [2001]. This evolution was accompanied by calc-alkaline volcanism that migrated from west to southeast, from Sardinia (32–13 Ma) to the currently active Aeolian Island Arc [*Marani and Trua*, 2002]. At the present day, most of the oceanic lithosphere in the Tyrrhenian-Apennine system has been consumed with exception of the Calabrian Arc sector where there is still subducting oceanic Ionian lithosphere, confined to the southwest by the Malta Escarpment (Figure 1).

3. Teleseismic Data

[8] During the 6 month deployment of the TYDE array, 50 teleseismic events with $m_b \geq 5.0$ were recorded, 35 selected for this study. Figures 2a–2b shows the 13 (out of 14, one was lost) ocean bottom instruments deployed with an average spacing of 25 km in an area of $100 \times 150 \text{ km}^2$ between the Tyrrhenian Calabrian coast and Marsili volcano. Six stations were equipped with a three-component broadband seismometer (three with 0.3 Hz natural frequency and three with 0.03 Hz, 50 Hz sampling rate for both types) and an ocean bottom seismometer hydrophone (OBS/H). The other eight had only a hydrophone (OBH) [*Dahm et al.*, 2002; *Sgroi et al.*, 2006]. Also the bathymetry is shown (Figure 2b). Each seafloor instrument operated with an internal quartz clock and used analog continuous recording on cassettes. The internal clocks were synchronized with GPS signal before of deployment and the same operation was repeated after the station retrieval. The clock drift of each station, considering the total time of the mission (about 6 months), never exceeded $\pm 0.6 \text{ s}$. It means that for each day we applied a correction of 0.2 ms. Being the drift of the OBS clock signal linear, the correction for this time delay preceded the data processing [*Sgroi et al.*, 2006]. This approach is justified by the fact that the ambient temperature of the internal clock is constant (see Web sites <http://www.iris.edu> and <http://www.ifm-geomar.de>).

[9] The teleseismic waveforms collected by the seafloor stations were combined with those from 60 short-period land stations of RSNC located in southern Italy (Figure 2a), and with the waveforms collected by a hydrophone hosted by GEOSTAR during its contemporary deep sea experiment near Ustica Island (2000 m w.d., September 2000 to April 2001) [*Favali et al.*, 2002, 2006]. The *P* and *PKP* arrival times were picked selecting the first break on the vertical seismograms. Moreover, because of the different instrument characteristics, all the broadband seismograms were convolved with the short-period RSNC instrument response to verify that the selected times led to a correct alignment of the first 2–3 cycles of seismic traces. The differences in the arrival times picked using both methods are generally limited to a few samples (0.02–0.04 s).

[10] Then an a priori weight at each reading has been assigned on the basis of the observed onset quality which represents the uncertainty on the picked arrival time. The weights (0, 1, 2, and 3) correspond to an estimated

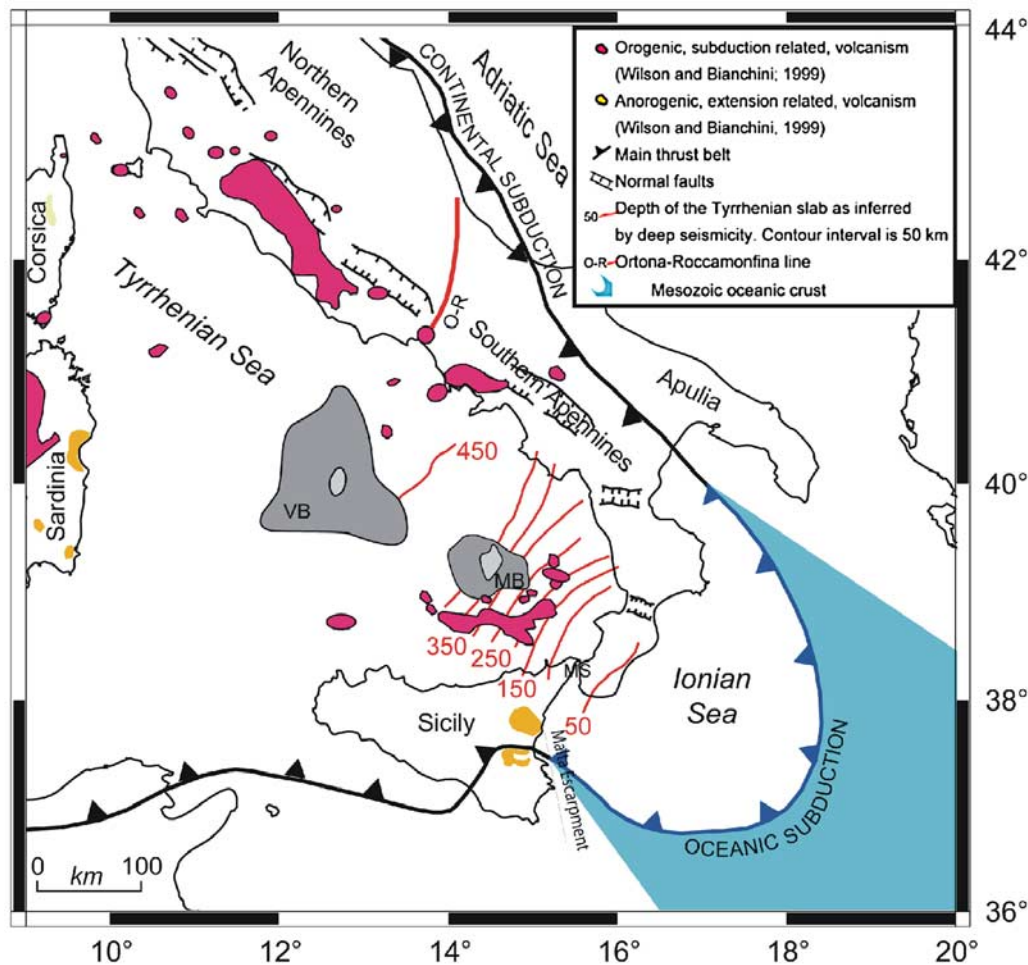


Figure 1. Simplified tectonic map of the southern Tyrrhenian zone (redrawn from *Cimini* [2004]). Offshore, the approximate extent of the Vavilov and Marsili basins is traced in dark grey; the light grey areas identify the homonymous central volcanoes of the basins [Marani and Trua, 2002]. MS: Messina Strait.

picking accuracy of 0.05, 0.2, 0.35 and 0.5 s respectively. The OBS/Hs, having both hydrophone and seismometer, produce for each event a couple of recordings on the vertical component. We picked the teleseismic phases on the best quality seismogram.

[11] Figure 3 shows an example of teleseismic waveforms for an event which occurred in Japan at epicentral distance of about 90° and back-azimuth of 50° (arrow in Figure 4). Usually, we used the reading on seismometer when both waveforms had similar quality (Figure 3c); moreover, the P arrival times on hydrophone and seismometer were coincident, being the time differences lower than the smallest reading error (0.05 s) (Figure 3, top).

[12] We also included 80 events recorded by the land stations during 1990–2002 to fill the sectors characterized by a low level of seismicity, obtaining a more homogenous distribution of ray paths. Figure 4 shows the azimuthal distribution of the 115 selected earthquakes occurring at epicentral distances of $30\text{--}100^\circ$ (P phases, indicated with circles) and $>110^\circ$ (PKPdf phases, indicated with triangles). The PKP events have been used to improve the sampling of the target volume to near-vertical incidence angles. In total, retaining only the events with at least 10 recordings, we

collected for the inversion 2904 P phases and 314 PKPdf phases.

4. Traveltime Residual Analysis

[13] First of all, we studied the pattern of traveltime residuals because they provide the primary evidence on the degree of heterogeneity in the Earth's structure inside the target volume. Their azimuthal variation allows the depth of seismic velocity anomalies to be qualitatively checked. Specifically, we considered relative residuals, calculated by subtracting a weighted residual mean for each earthquake from the absolute residual computed for the given observation. Relative residuals has been chosen to minimize errors coming from ray paths outside the target volume, from hypocentral mislocation, and from origin time determination. We computed the theoretical traveltimes for each teleseismic ray using the ak135 global velocity model [Kennett et al., 1995]. Since the reference velocity model assumes the Earth as a sphere, residuals are corrected for the ellipticity [Dziewonski and Gilbert, 1976] and for station elevation (negative for the seafloor instruments). Negative

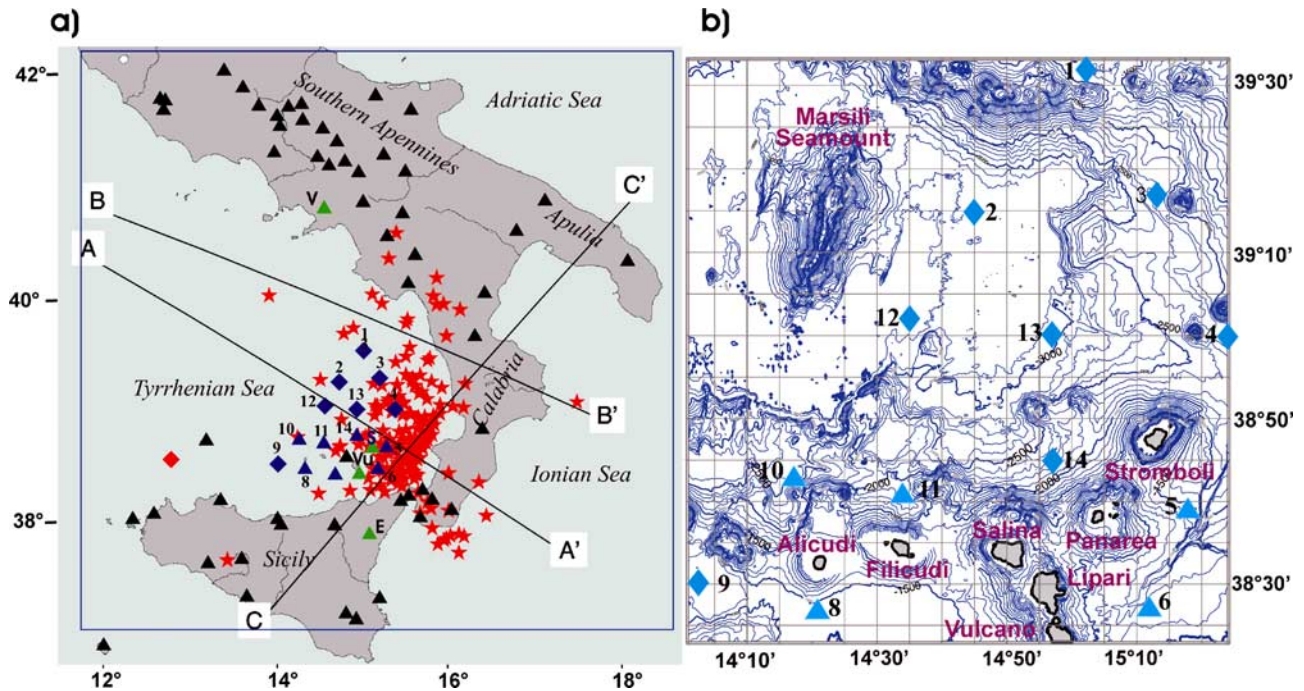


Figure 2. (a) Distribution of seismic stations used in the inversion. Blue triangles and diamonds indicate ocean bottom seismometer hydrophone (OBS/H) and hydrophone (OBH), respectively, of the Tyrrhenian Deep-sea Experiment (TYDE). Black triangles are the permanent stations of the Italian network. Red diamond indicates OBH of the GEOSTAR observatory. Green triangles are the active volcanoes (from north to south, V = Vesuvio; S = Stromboli; Vu = Vulcano; E = Etna). Red stars indicate the subcrustal seismicity which occurred in the south Tyrrhenian Basin [Cimini and Marchetti, 2006]. Black AA', BB' and CC' lines are the strikes of tomographic sections shown in Figures 7, 8, and 9; (b) Bathymetric map of the southern Tyrrhenian Sea (kindly provided by Istituto di Scienze Marine (ISMAR-CNR); Marani *et al.* [2004]) together with the deployment sites of the temporary seafloor seismic stations.

residuals imply the presence of high-velocity anomalies, whereas positive indicate low-velocity structures.

[14] Figure 5 shows the distribution of the traveltime relative residuals, observed at stations close (± 50 km) to profile AA' (Figure 2a), for two *P* wave fronts originating from point source at 81° and 87° in epicentral distance and arriving from the NNW (Alaska; back-azimuth 354°) and SSE (South Africa; back-azimuth 168°) respectively. The main feature of the residual plot is the opposite trend of spatial variation for the two directions. Moving northwestward from Calabria to the Aeolian Islands, the stations show progressively earlier arrival time (negative residuals) for the rays from SE and later arrival times for the rays from NW (positive residuals). This pattern is better displayed in a 2-D map that shows the relative residuals at the Earth's surface for the two above-mentioned teleseismic events (Figure 6). In the map of the South Africa earthquake (Figure 6a) it is evident the separation between negative (most of the seafloor stations) and positive residuals in Sicily and Calabria. This limit, extending toward onshore, crosses the OBH 04 and a station installed on Salina Island (SA1) with residuals $+0.3$ and $+0.5$ s. The highest positive residuals, as large as $+1.4$ s, are observed at stations located in NE Sicily, probably related to the deep thermal anomaly beneath Etna volcano. Considering the average spacing (~ 25 km) and the average incidence angles of teleseismic rays ($\sim 20^\circ$), the pattern of residuals, observed in Figure 6a, suggests the

presence of a high-velocity body in the upper mantle (above 70 km depth) beneath the seafloor stations with an extension of about 200 km and a thickness of about 80 km. It is a clear expression of the southern Tyrrhenian subduction zone whose fast structure is projected toward the basin. The map for the Alaska event (Figure 6b) shows an opposite trend with strong positive residuals observable at the seafloor stations, suggesting that a region of low-velocity anomaly is located at shallow depth beneath the Aeolian Arc.

5. Inversion Method

[15] We inverted the *P* and PKPdf relative residuals using a nonlinear teleseismic tomography technique [Weiland *et al.*, 1995; Cimini and De Gori, 2001] that incorporates a robust minimum traveltime ray tracing [Steck and Prothero, 1991] for accurate ray paths and traveltimes computation in 3-D media. The algorithm performs a sequence of linearized inversion within a standard iterative scheme. A weighted damped least squares solution of the linearized equations, relating the observed relative traveltime residuals to the slowness perturbations from a background velocity model, is computed at each iteration. In this study, the ak135 spherical earth model [Kennett *et al.*, 1995] was used as the initial reference *P* wave speed model for the inversion. Note, however, that though a known background model is needed for the forward calculation of the theoretical trav-

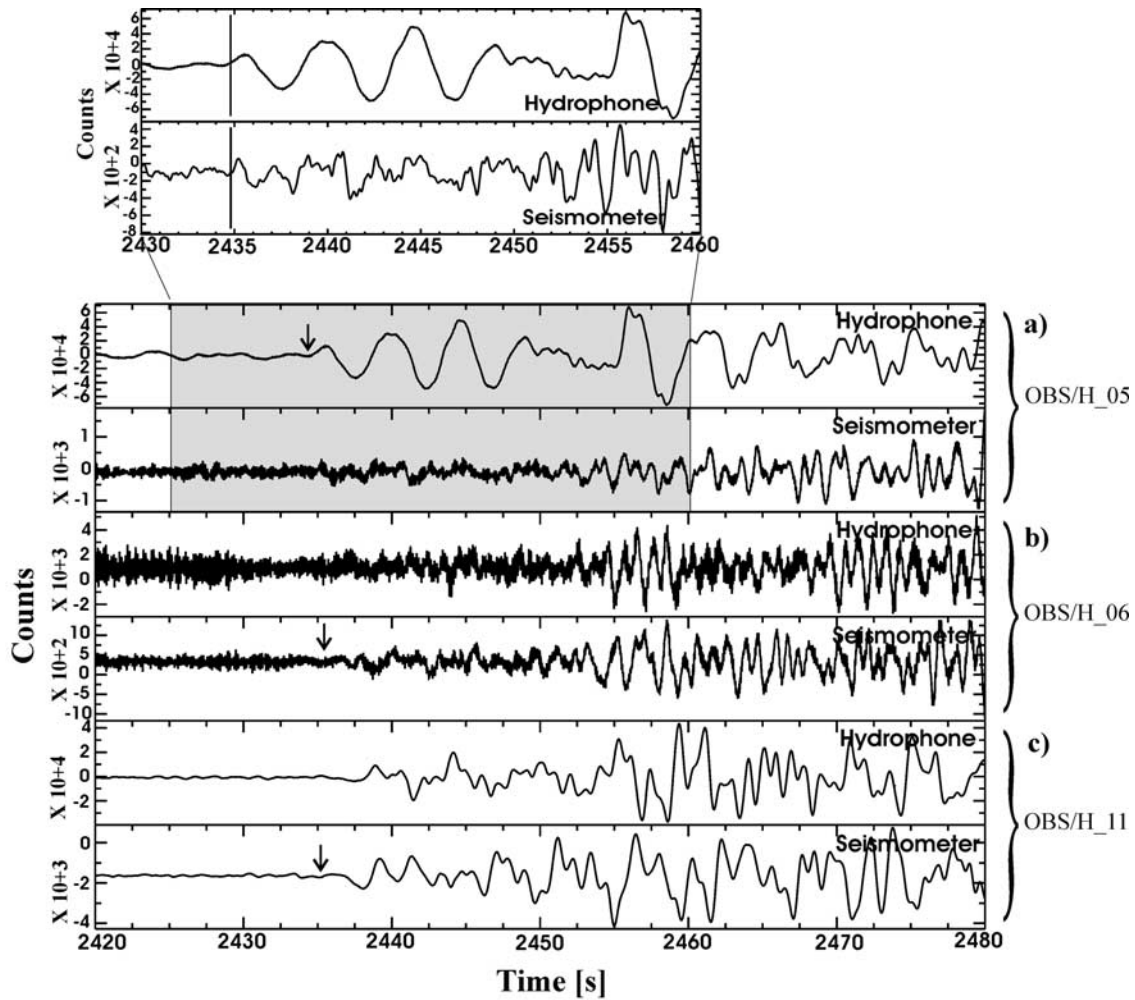


Figure 3. Teleseismic waveforms recorded at three seafloor stations of the m_b 6.8 Japan earthquake which occurred on 24 March 2001 0627 UTC (arrow in Figure 4). On the top is a zoom of OBS/H_05 showing the arrival time of the P wave. The seismometer waveform has been filtered with a 5 Hz low-pass filter to point out the simultaneous P wave arrival time.

eltimes, the resulting velocity perturbations cannot be related to this background model, because of inversion of relative arrival time residuals. As a consequence, the absolute 3-D wave speeds obtained through iterations by superimposing the wave speed perturbations are uncertain without independent constraints on the depth variation of the horizontally averaged wave speeds. As illustrated by Koch [1985] testing the influence of the approximations in the ACH method [Aki *et al.*, 1977], there is no way to remove this inherent nonuniqueness completely in the inverse solution. We can, to some extent, only diminish its degree by iterating with 3-D ray tracing. Indeed, the numerical computations performed by Koch [1985] show how the reconstruction of the velocity models can be improved by nonlinear iterative refinement of the tomographic solution, allowing to better retrieve the magnitude of lateral heterogeneities, generally underestimated by linear arrival time inversions, and to achieve larger data misfit reduction. Moreover, if the target volume is large enough, we may assume that the layer average velocities are close to some commonly accepted 1-D reference model [Leveque and Masson, 1999].

[16] The basic equations of the iterative procedure we have applied are exposed by Cimini and De Gori [2001]. We used the singular value decomposition (SVD) method [Menke, 1989] to calculate the inverse of the matrix kernel. The size of our inverse problem (matrix G of partial derivatives is $\sim 3500 \times 1500$) is not prohibitive for using this approach that, although more computational expensive than other methods used to solve the tomographic equations, circumvents problems present in approximate techniques and also allows a direct estimate of the model resolution. The damping parameter, needed to stabilize the inversion by reducing the high-frequency component of the solution, was chosen through trade-off analysis between resolution and standard errors. We selected a damping value of 300 s^2 which provides a good average resolution (diagonal element >0.5 on average) and limited model standard errors ($<1\%$).

[17] The starting model we adopted for the inversion is a 3-D grid of nodes consisting of the station locations and nine layers of nodes which spacing increase from $35 \times 35 \text{ km}$, for the crustal layer at 30-km depth, to $80 \times 80 \text{ km}$ for the bottom layer at 600-km depth (Table 1). Each layer is

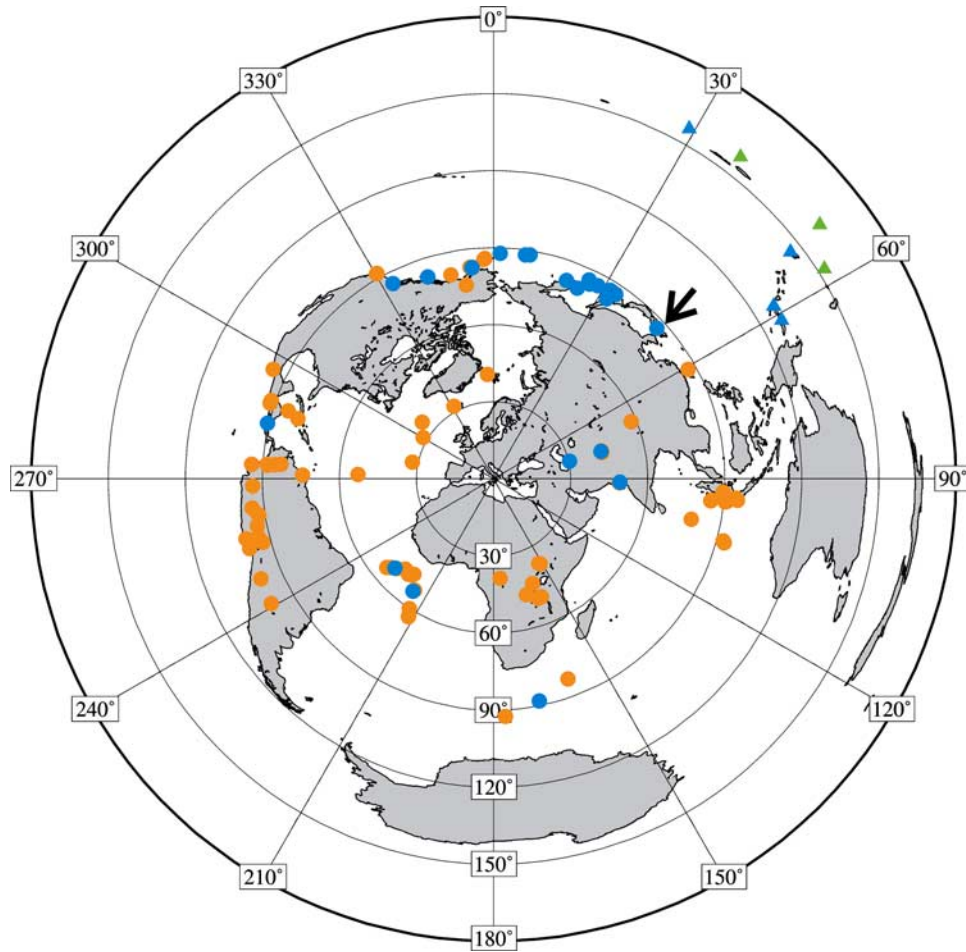


Figure 4. Geographical distribution of the teleseismic events used in this study. Blue circles indicate P arrivals, while blue triangles indicate the *PKP* phases, both recorded by seafloor stations. Orange circles and green triangles represent the *P* and *PKP* phases recorded by the Italian National Network (RSNC) from 1990 to 2002. The arrow indicates the location of the event shown in Figure 3.

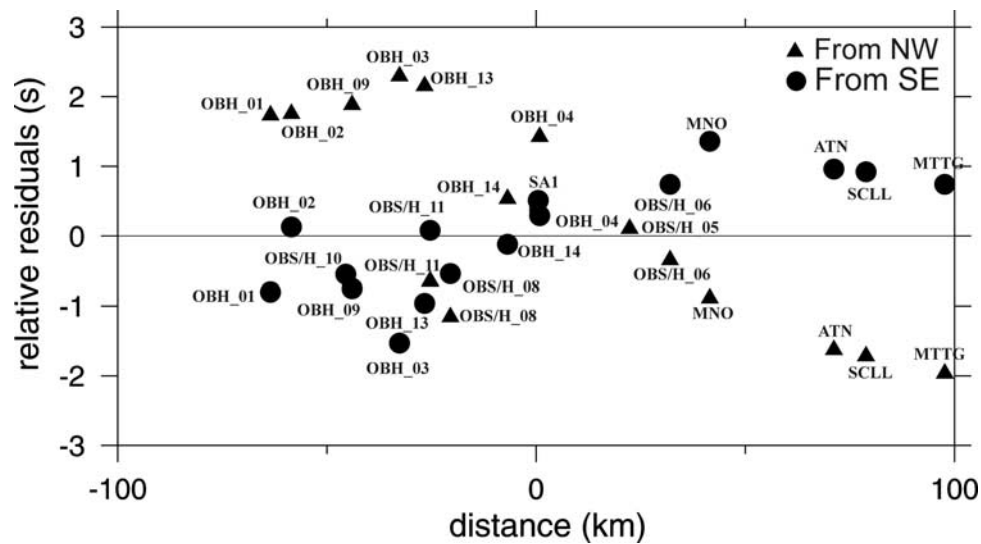


Figure 5. Relative residuals observed at some stations close to the profile AA' (see Figure 2) for rays coming from NW (triangles) and SE (circles). Observed trends may be justified if a high-velocity body is located approximately at 100 km depth and a low-velocity anomaly is above the fast body in the first 100 km of the target volume.

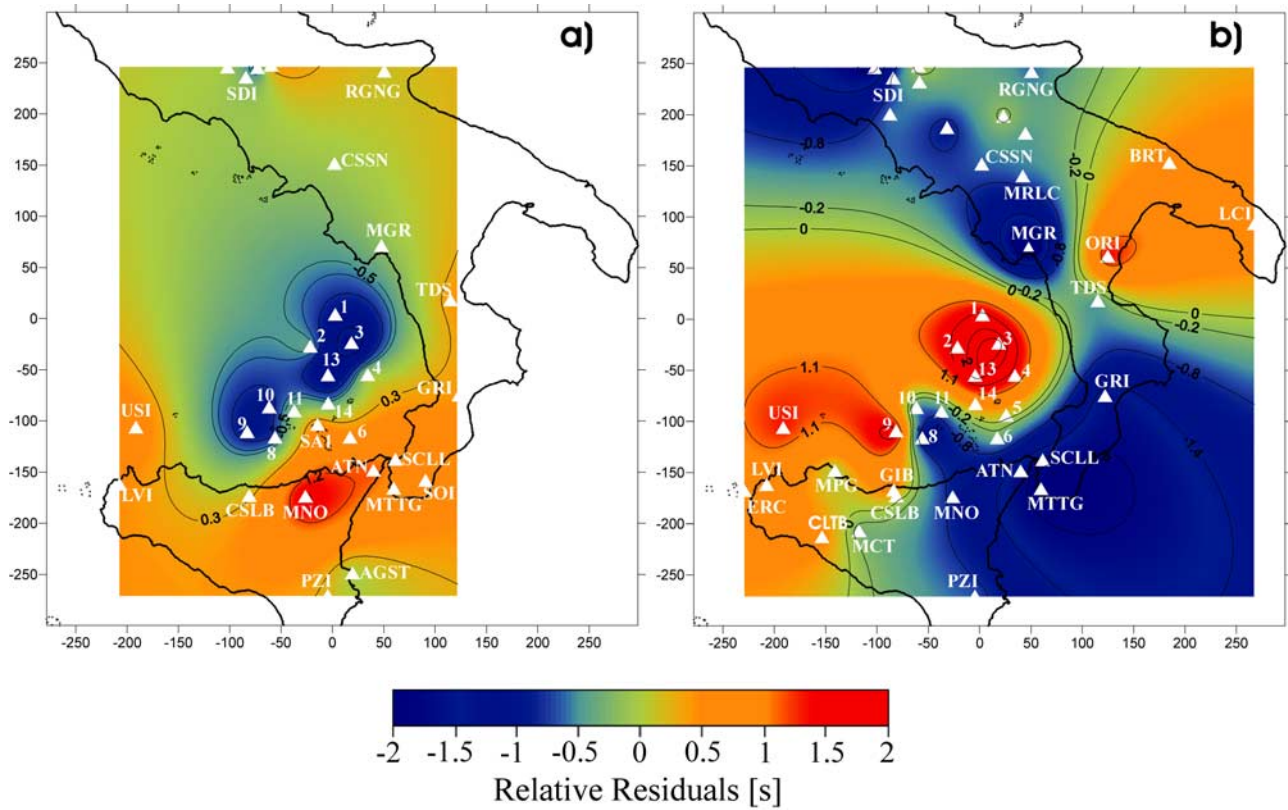


Figure 6. Map of the relative residuals for two teleseismic events: (a) m_b 6.0 South Africa event which occurred on 22 January 2001 1509 UTC; (b) m_b 6.0 Alaska event which occurred on 20 February 2001 0548 UTC. White triangles indicate the stations used in the calculation of the relative residuals for each event. Black thin lines contour the main pattern of residuals.

composed by a bidimensional mesh of 19×19 nodes with initial P wave velocity assigned according to the ak135 model. The centre of the grid is at 39.5° N, 15.0° E, and, for ray paths computation, the bottom layer (600-km depth) was not inverted. This model parameterization, which varies the length scale of the upper mantle structures that can be recovered by the present data sampling, attempts to better fit the distribution of the teleseismic rays at depth.

6. Tomographic Images

[18] In this section we present the results of the inversion through 2-D horizontal (layers, Figure 7) and vertical (cross sections, Figure 8) slices of the P wave velocity pattern. The seismic structure is displayed as velocity perturbations, ranging $\pm 4\%$, from the ak135 starting model [Kennett *et al.*, 1995]. Thus red and blue regions are, respectively, regions with lower and higher P wave velocity with respect to an unknown background model from the ak135 starting model. As mentioned above, the layer average velocities of the correct model remain unknown because we have inverted relative residuals [Aki *et al.*, 1977]. The nonlinear tomography was carried out through a maximum of five iterations to check the stability and convergence of the optimization process. We show the velocity changes recovered after three iterations, as only minor adjustment to the model parameters appear on further iterations. The data variance reduction that is obtained after these steps amounts to 72%. In total, our model consists of 3218 P observations

and an average of 1228 inverted nodes (the grid nodes sampled by at least five rays) which lead to an over determination factor of about 2.6.

[19] For each model parameter we also provided an estimation of the spatial resolution, obtained by investigating the resolution matrix after the first iteration (the linear solution). The resolution analysis has been carried out by computing the resolution spread, as defined in Menke [1989]. The resolution spread is a simple but effective way to assess the reliability of the velocity anomalies based on the size of the off-diagonal elements. In particular we adopted the measure based on the L_2 norm of the difference between the resolution matrix $\mathbf{R}$ and the identity matrix $\mathbf{I}$,

Table 1. Grid Model and Initial Reference P Wave Velocities Used for the Iterative Inversion

Layer	Depth, km	Nodes in x	Nodes in y	Δx , km	Δy , km	V_p , km/s
1	30	19	19	35	35	6.50
2	65	19	19	40	40	8.04
3	105	19	19	45	45	8.05
4	150	19	19	50	50	8.13
5	200	19	19	55	55	8.27
6	260	19	19	60	60	8.48
7	330	19	19	65	65	8.74
8	410	19	19	70	70	9.19
9	500	19	19	75	75	9.66
10 ^a	600	19	19	80	80	10.00

^aNot inverted.

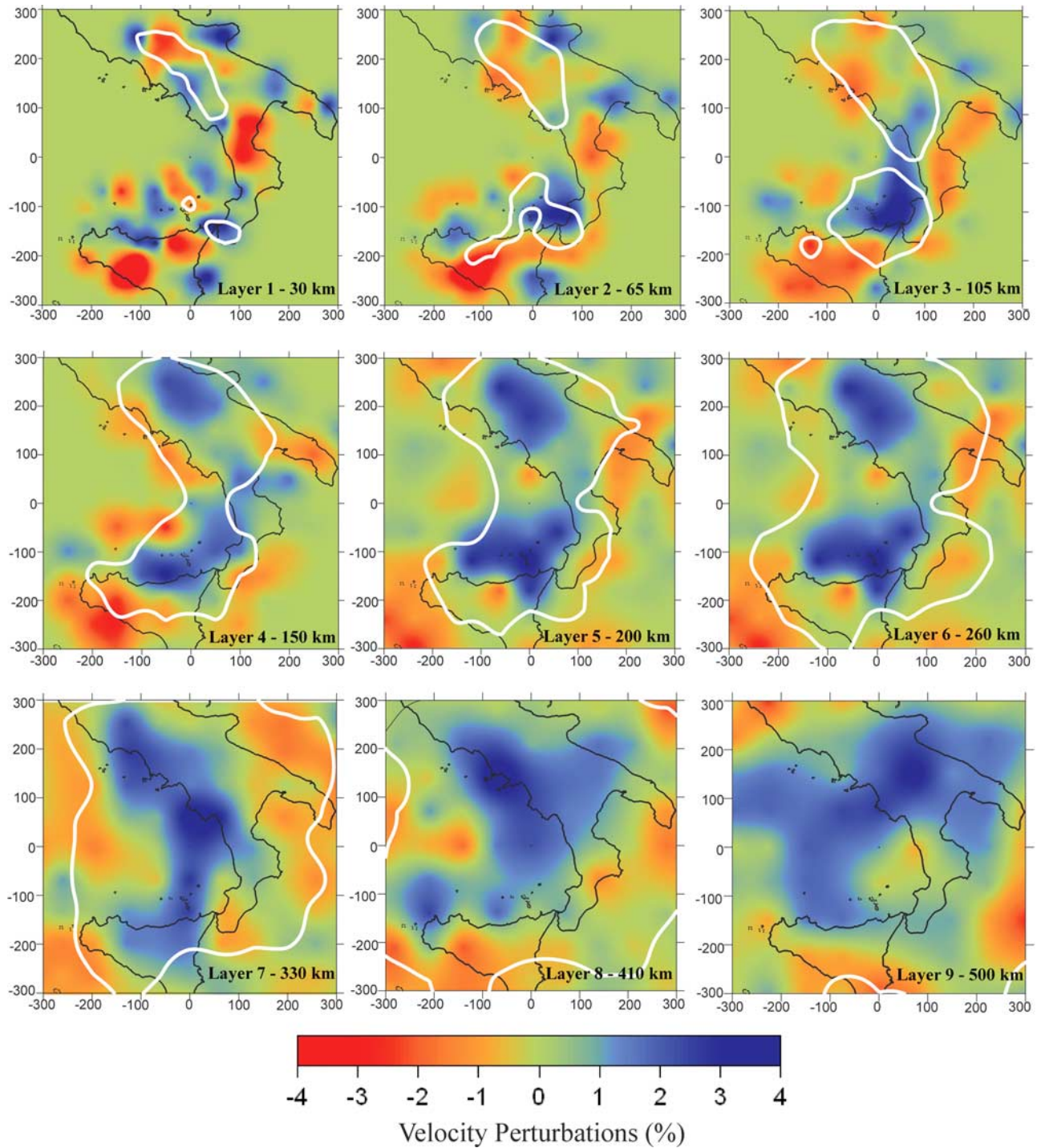


Figure 7. Maps of the P velocity perturbations for nine layers (30 to 500 km), obtained from the nonlinear inversion after three iterations. The white line contours the well-resolved part of the model having resolution spread ≤ 0.4 .

when $\mathbf{R} = \mathbf{I}$ the model parameters are uniquely determined and spread $\mathbf{R} = 0$. In the velocity anomaly maps of Figures 7 and 8, the white curves delimit regions with resolution spread lower than 0.4. We consider this value as the upper limit for well-resolved model parameters, as we have found by examining single rows of the resolution matrix (averaging vectors) spatially distributed throughout the sampled volume. Nodes within the 0.4 contour are characterized by negligible values of the off-diagonal elements associated

with the nodes distant more than $\sim 1-2$ grid spacing, thus indicating smearing effects and volume averaging of the velocity estimates limited in the neighbourhood of the given node.

6.1. Maps

[20] Figure 7 displays the P wave velocity anomalies in the nine layers within the adopted 3-D grid (Table 1). The first layer of the model (layer 1, 30 km) is an attempt to

reconstruct the velocity structure in the lower crust. A low-velocity anomaly (LVA) beneath the Aeolian Arc and a high-velocity anomaly (HVA) (>4%) beneath Messina Strait are clearly evident. As in previous investigations using teleseismic data, the station coverage is still inadequate

for a high-resolution sampling and reliable imaging of the lateral heterogeneities. At this depth, the well-resolved parts of the model are limited in extension (the areas contoured by the white line), hence only some qualitative observations can be drawn. The southern Apennines are dominated by a

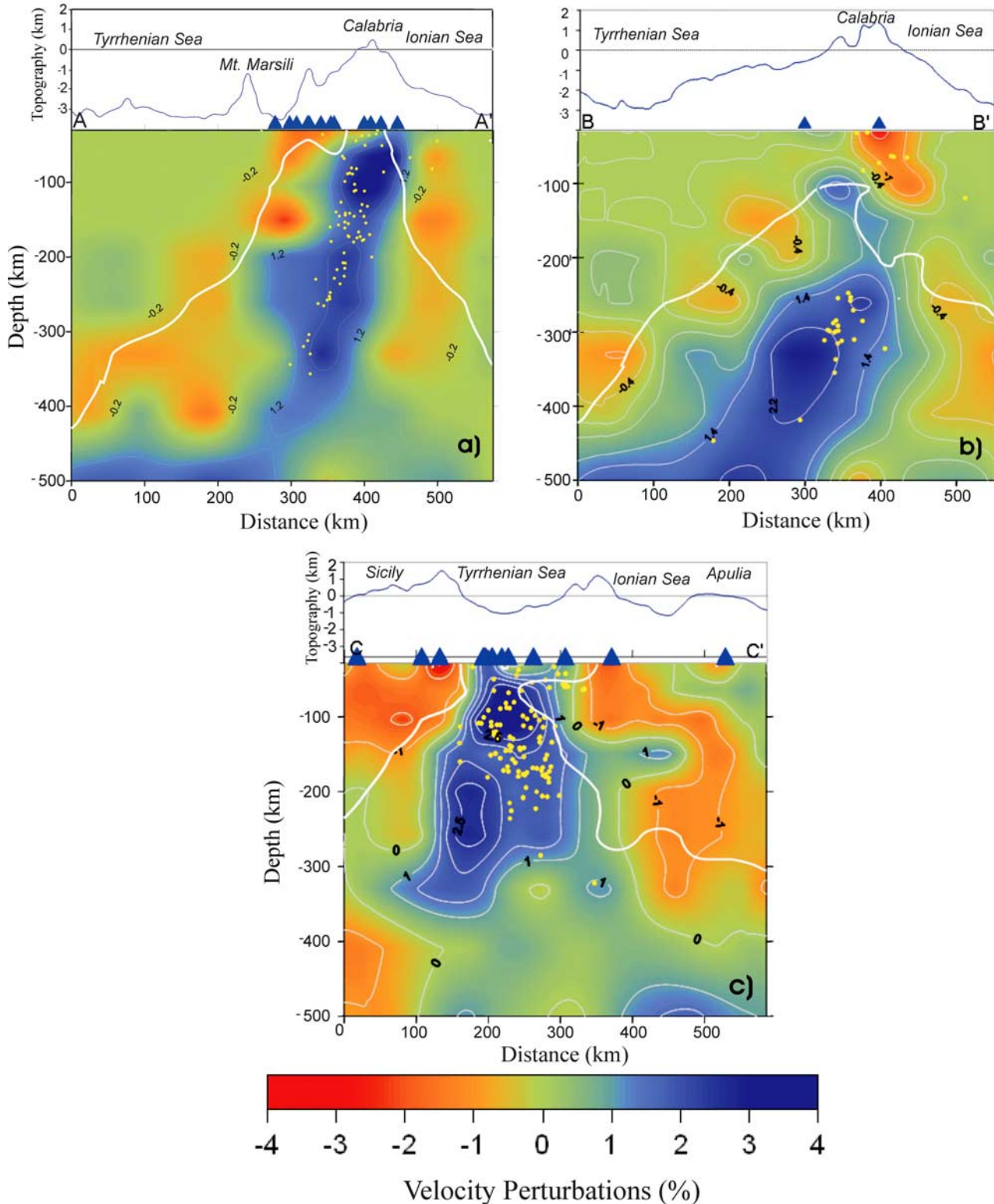


Figure 8

wide LVA that extends from NW of Vesuvio area to the Adriatic coast, in the region of the separation between the northern and southern Apennine arcs [e.g., *Amato et al.*, 1993b; *Cimini and De Gori*, 2001].

[21] In the upper mantle, at 65 and 105 km depths (layers 2 and 3 in Figure 7), the velocity structure is characterized by HVA beneath the Aeolian Islands and the southern Tyrrhenian Basin with perturbations above 4%. This fast body is arc shaped with concavity toward NW following the Calabria–Sicily coastline. It extends toward the center of the basin at increasing depths. Low-velocity zones, visible down to 150 km depth, are reconstructed in the back-arc basin (bathyal plain) where a regional high heat flow is observed [*Della Vedova et al.*, 2001; *Zito et al.*, 2003]. Negative velocity perturbations as large as -4% are also visible beneath central Sicily and the Ionian Sea. Although these features are only partially included in the well-resolved part of the model, it is interesting to note the lateral extension of this LVA from Sicily to Calabria, reaching the Apulian foreland and bordering east with the Tyrrhenian Basin HVA. The Tyrrhenian fast structure is thus completely surrounded by LVA areas.

[22] In the upper mantle, between 150 and 260-km depth (layers 4 to 6 in Figure 7), two prominent high-velocity anomalies separated by a ~ 200 km wide area of weak positive perturbations ($\Delta V < 1\%$) below the southern Apennines, are observed. The first anomaly is located beneath the southern Tyrrhenian Basin. Its lateral extent reaching about 200 km along the NNE–SSW direction, while its shape broadens along NW–SE direction at increasing depths. The second positive perturbation, found beneath the central southern Apennines, presents a marked vertical continuity, as already described in a previous nonlinear tomography of the region [*Cimini and De Gori*, 2001]. These two fast structures, not apparently in continuity within this depth interval, define two different styles of lithospheric sinking in the Apennine subduction [*Cimini*, 2004, *Cimini and Marchetti*, 2006]. The tomogram at 150-km depth also shows broad areas of negative velocity perturbations NW of the Aeolian Islands, western Sicily, the Sicily channel, and the Apulia foreland.

[23] In layers 7 and 8 at 330 and 410 km (Figure 7), the two strong positive anomalies join to form a continuous fast structure that extends laterally for about 500 km, from eastern Sicily to central Italy. At these depths, the low-velocity anomaly beneath the Tyrrhenian Sea still persists.

[24] The bottom layer of our model (layer 9, 500-km depth) images the aspherical P wave velocity structure within the upper mantle transition zone (410–660 km).

The velocity pattern is marked by a broad HVA, extending from the Tyrrhenian basin to the southern Apennines.

6.2. Cross Sections

[25] Figure 8 shows the resolved P wave velocity structure beneath southern Italy as displayed by three cross sections. Section AA' (Figure 8a) runs from the Tyrrhenian to the Ionian Sea, striking NW–SE (Figure 2a). The tomogram displays a continuous high-velocity structure (ΔV up to $+5\%$) from the lower crust down to the bottom of the present model. It suggests a ~ 100 km thick lithospheric slab steeply dipping NW from the uppermost mantle continuously to at least 400 km depth. This continuity was already observed by previous tomographic studies [*Cimini*, 1999; *Piomallo and Morelli*, 2003]. The deep seismicity is enclosed within the fast body. At greater depth, the sub-horizontal HVA may represent the oldest portion of the lithosphere which is presently stagnant in the transition zone [*Cimini*, 1999; *Lucente et al.*, 1999]. The teleseismic image also reveals two continuous low-velocity areas flanking the slab structure. The maximum negative perturbations ($\Delta V_p < -2\%$) are defined beneath the Aeolian Islands at uppermost mantle depths, where the resolution is fairly good because of the presence of the seafloor stations.

[26] Section BB' (Figure 8b) is shifted ~ 70 km northeastward respect to profile AA', attempting to define the transition between the fast structures pointed out in the maps of Figure 7 (layers 4 to 6). This image depicts a continuous high-velocity body ($>4\%$) below 200 km depth reaching a thickness of 200 km between 300 and 400 km depth. At shallower depths (30–200 km), the main feature is represented by a strong LVA located beneath Calabria, on the top of the HVA, even if it is not inside the well-resolved zone.

[27] The along-strike view of the NW-dipping high-velocity body is imaged in section CC' (Figure 8c), where the pattern of the P wave velocity anomalies is displayed on a SW–NE vertical section crossing the investigated area from Sicily to the Apulia foreland. In this vertical plane, the subducting lithosphere is characterized by a lateral extension of about 200 km in the depth interval 150–300 km. In the uppermost mantle, the fast structure has a smaller lateral dimension (about 100 km) and it is flanked to NE and SW by two pronounced low-velocity regions. The NE anomaly, located below the Ionian Sea and Apulia, is not a well resolved feature. However, as displayed in the maps of Figure 7, these low-velocity anomalies, that border the fast structure on the Ionian–Apulia “side”, are not completely outside the 0.4 resolution spread contour, indicating that

Figure 8. Cross sections of the 3-D velocity model (see traces in Figure 2). In each section the white line shows the well-resolved part of the model with resolution spread ≤ 0.4 , and the yellow dots represent the subcrustal seismicity that occurred in the south Tyrrhenian Basin [*Cimini and Marchetti*, 2006] (a) Profile AA': An almost continuous high-velocity anomaly is visible. The strong perturbations ($>4\%$) down to 500 km and the presence of deep seismicity suggest that the lithosphere is oceanic. Two prominent low-velocity anomalies flank the subducting slab indicating a strong correlation between subduction and mantle flow. It is noteworthy that the low-velocity anomaly located in the first 200 km, beneath the marine array, is well resolved; (b) Profile BB': The subducting slab appears as a strong and thick (about 150 km) positive perturbation below 200 km depth. From 100 to 200 km the positive V_p anomaly is about 50 km thick. In the first 100 km the high-velocity anomaly is absent; conversely a low-velocity anomaly, characterized by low resolution, is found beneath Calabria, indicating a laterally heterogeneous subducting lithosphere; (c) Profile CC': The lateral extension of the high-velocity body is clearly shown. It increases with depth, and it reaches about 200 km from 150 to 300 km. The oceanic slab is surrounded by low-velocity anomalies.

they are fairly reliable for the interpretation, at least as possible extension of better sampled heterogeneities.

[28] The deep seismicity, located within the high-velocity body down to 300 km, is absent at greater depths.

7. Discussion

[29] The addition of teleseismic waveforms collected from temporary seafloor stations deployed around the Aeolian Island to the RSNC data set allowed us to obtain more detailed and better resolved tomographic images of the lower crust and uppermost mantle structure beneath the southern Tyrrhenian subduction zone. The tomographic images clearly delineate an area of strongly positive perturbation immediately close to the Calabrian Arc. Such a fast region depicts a continuous body down to 500 km depth, shown by the images as a unique structure. As in the more recent literature [e.g., Hirahara and Haseami, 1993; Piromallo and Morelli, 1997, 2003; Spakman and Wortel, 2004], such positive anomaly is interpreted as the expression of the Ionian slab subducting toward NW under the Aeolian Arc. Compared to the most recent tomographic models developed for the southern Tyrrhenian and/or for the entire Italian peninsula [e.g., Cimini, 1999; Lucente et al., 1999], the results of our inversion define more clearly the three-dimensional shape and dimensions of the Ionian slab beneath the southern Tyrrhenian Basin. The lateral extension of the subducting lithosphere is 200 km between 150 and 260-km depths and 80–100 km in the layer at 65 km (Figure 8c). These dimensions are ascribed to the core of the slab corresponding to the seismogenic part of the subducting Ionian lithosphere. In the layer at 30 km depth, the HVA is limited to a small area beneath Messina Strait (Figure 7, layer 1). This increase of lateral extension with depth, observed also in the other tomographic models, has been interpreted as the expression of the Tyrrhenian slab during eastward migration of the back-arc system in the last 2–3 Ma, when the thick Apulian continental lithosphere reached the subduction hinge [Doglioni et al., 1994]. The narrow lateral extent of the HVA depicting the shallower part of the slab may reflect the narrow strip of oceanic lithosphere remaining after the continental collision took place at its eastern (southern Apennines) and western (Sicily) borders.

[30] Regarding the continuous lateral extension of the slab toward N, the uppermost mantle structure (from lower crust to 260-km depth) does not show a clear anomaly pattern (section BB', Figure 8b) probably because of the still insufficient station coverage. Despite the poor coverage, the apparent discontinuity in the subduction system may be related to the transition from the oceanic subduction beneath

Calabria and southern Tyrrhenian Sea to the continental collision/subduction below central southern Apennines [Cimini and Marchetti, 2006]. From 260 down to 450-km depth, the upper mantle structure is characterized by a clear SW dipping HVA which is correlated with the body observed in the section AA' (Figure 8a). Such a body represents the lithosphere subducted before the collision of continental lithosphere in the central southern Apennines occurred during the Messinian (6.5 Ma) [Speranza et al., 1997]. The lateral extension of the slab, from Sicily to Central Apennines (see maps of Figure 7) reaches about 500 km.

[31] One of the most relevant results of this tomographic model is the broad LVA at the uppermost mantle reconstructed behind and around the steeply dipping HVA, and inside the well-resolved zone. The negative perturbation reaches -3% with respect to the background model (section AA', Figure 8a). At increasing depths it is weaker and it follows the trend of the slab. In the previous tomographic reconstructions the low-velocity regions have been observed under 200-km depth and interpreted as thermal anomalies induced by slab retreat and the subsequent rising of the asthenosphere in the Tyrrhenian Basin [Cimini, 1999; Piromallo and Morelli, 2003]. Our interpretation of this pronounced LVA pattern is that the asthenospheric flow moves toward the wedge around the sinking Ionian slab, suggesting a three-dimensional mantle circulation. This mantle flow has been hypothesized by a recent anisotropy study [Civello and Margheriti, 2004], where the authors draw possible trajectories of mantle flow around the western slab edge beneath Tyrrhenian plate at 150-km depth. We suppose that this circulation of the mantle flow exists around the entire fast structure down to 300-km depth, on the basis of the low-velocity anomaly pattern observed in our imaging.

[32] Recent laboratory experiments [Kincaid and Griffiths, 2003] fully support our interpretation. The authors have shown fundamental differences in three-dimensional mantle circulation and temperature structure in response to subduction with and without a roll-back component. As the Ionian slab is characterized by a translational component of motion (roll-back), migration trench and the deep plate tip [Faccenna et al., 2001], we hypothesize that the wedge of the Ionian slab is filled by material from around the edges of the plate. The strong LVA above the central part of the Ionian subducting slab suggests that in this region of the mantle wedge high values of temperature are reached. Kincaid and Griffiths [2003] also demonstrated in their laboratory work that a slab with roll-back component is characterized by higher temperatures in the central part of the descending plate, where the asthenospheric material in the wedge flows with higher velocity. While the lower

Figure 9. 3-D schematic sketch of the Tyrrhenian subduction zone drawn on the basis of the tomographic images. The high- and low-velocity anomalies are outlined drawing the 1% and -1% P wave velocity perturbation contour lines, respectively. (a) SW view of the NW-SE vertical section (profile AA' of Figure 8a), where the continuity of the slab down to 400 km depth and its flattening at 500 km depth are visible; in the uppermost mantle the low-velocity anomalies around the slab are also imaged. (b) Illumination from NW of cross section CC' of Figure 8c, where the lateral extension of the seismogenic part of the Tyrrhenian slab down to 300-km depth is represented. The two low-velocity areas are located below Apulia and Sicily regions. The latter is better resolved and hence more reliable. The former is resolved only in the first 30-km depth because of poor distribution of seismic stations on the Apulia region and to the lack of OBS in the Ionian Sea (see Figure 2a). Both pictures include the layer at 500 km depth (layer 9 of Figure 7). The distribution of the low-velocity anomalies point to circulation of mantle flow around the slab (see arrows) down to about 300-km depth.

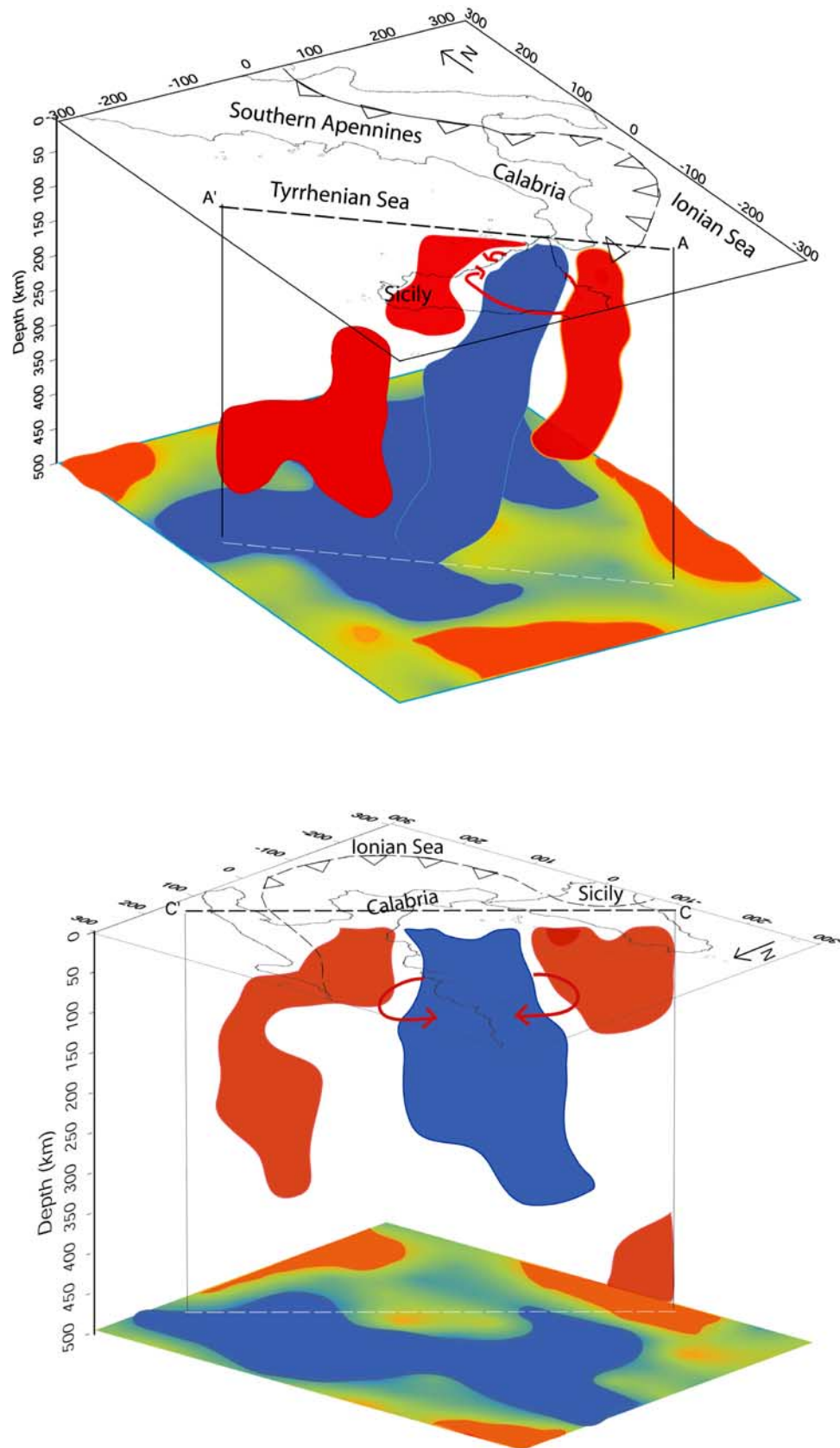


Figure 9

temperatures are displayed at the edges of the wedge. The area, modeled by *Kincaid and Griffiths* [2003] with higher temperatures and high-velocity flow, corresponds to our LVA region where it is possible to assume a melt production that leads to the Aeolian magmatism.

[33] Figure 9 shows two schematic diagrams along the two vertical planes of Figures 8a and 8c, respectively, summarizing the results of the present tomographic study. The western flank of the low-velocity anomaly pattern is superimposed on the area including Ustica Island to NW and Etna volcano to SE, in which *Trua et al.* [2003] inferred a lateral north-northwestward flow of African mantle. This is based on the geochemical evidence of the significant similarity between the OIB-type magmatism of Ustica Island and Prometeo, a newly discovered submarine lava field [*Trua et al.*, 2003], and Etna and Hyblean plateau. The flow would be channeled along the southwestern edge of the Ionian oceanic plate in roll-back subduction. *Gvirtzman and Nur* [1999], proposed that the Ionian slab sunk deep enough to open a new route of mantle flow from under the African plate toward the tip of the asthenospheric wedge under Calabria. Such African mantle flow have been also invoked to explain the genesis of Etna [*Gvirtzman and Nur*, 1999; *Dogliani et al.*, 2001]. The clear separation between high velocity and low-velocity anomalies, showed in our model, could display the tear along with the African flow is driven.

[34] In a recent paper on the residual topography, lithospheric structure and sunken slabs located exactly in the same Tyrrhenian-Ionian zone of this study, *Gvirtzman and Nur* [2001] concluded that the Tyrrhenian SE edge (Calabria and northern Sicily) is particularly interesting because of its unusually high residual topography. There the sharp transition, from very low residual topography over the edge of the subducting Ionian plate to the unusually high residual topography over the edge of the overriding Tyrrhenian plate, is explained as evidence of absence of mantle lithosphere, associated to a rapid slab roll-back, a weak plate coupling, and high asthenospheric penetration. This framework is consistent with a pattern of strong gradient between high and low-velocity anomalies, these latter surrounding the fast structure also toward the crustal layer, as our tomographic images show. At shallower depth the mantle flow deeply penetrates between the subducting and overriding (the Calabria peninsula) plate (section AA', Figure 8a). On the other hand, according to *Nur et al.* [1991] and *Dvorkin et al.* [1993], the width of the slab controls the rate of the mantle flow into the corner between a subducting and overriding plates (Ionian and Calabria). When the slab is narrow, such as the Tyrrhenian slab (see Figure 8c), the mantle flow reduces the suction between the two plates.

8. Conclusions

[35] An improved 3-D *P* wave velocity model for the crust and upper mantle below southern Tyrrhenian zone has been computed. For the first time in Italy, teleseismic data recorded by ocean bottom instruments deployed in the southern Tyrrhenian Sea (TYDE project) and integrated with data from the land-based Italian seismic network, have been used.

[36] The new data set has allowed an increase in resolution of the model beneath the Aeolian Island in the first 200 km depth, thus improving the image of the lithosphere-

asthenosphere system of the region. The Tyrrhenian slab is characterized by a continuous high-velocity body, steeply dipping northwestward from the lower crust down to about 400 km. The lateral extension of the seismogenic part of downgoing lithosphere is well defined by the present tomography and is about 100 km in the uppermost mantle and 200 km down to about 300-km depth. At increasing depths the slab presents a northward lateral extension of about 500 km. At uppermost mantle depths the prominent low-velocity regions lie beneath the Aeolian Islands and are better resolved than in previous investigations. These anomalies suggest an asthenospheric flow around the Ionian slab, visible above 300-km depth. This new 3-D velocity model agrees with results from laboratory experiments, mantle anisotropy studies, geochemical and isotopic analyses, and residual topography modeling. The availability of reliable seafloor seismic recordings, in addition to the land-based data set, demonstrates their usefulness in improving the quality of seismological studies.

[37] **Acknowledgments.** The authors wish to thank the Centro Nazionale Terremoti section of INGV for providing the seismic data from land network. We are grateful to the Partners of TYDE project: INGV, ISMAR-CNR, IfM-GEOMAR and Hamburg University, and EC for its support. We also thank Stephen Monna for discussions on this topic and Francesca Di Laura for help in drawing. We especially thank Ulrich Achauer and the anonymous referee for their constructive and helpful reviews. First and second authors contributed equally.

References

- Aki, K., A. Christofferson, and E. Husebye (1977), Determination of the three-dimensional seismic structure of the lithosphere, *J. Geophys. Res.*, **82**, 277–296.
- Amato, A., B. Alessandrini, and G. B. Cimini (1993a), Teleseism wave tomography of Italy, in *Seismic tomography: Theory and Practise*, edited by H. M. Iyer and K. Hirahara, pp. 361–397, CRC Press, Boca Raton, Fla.
- Amato, A., B. Alessandrini, G. B. Cimini, A. Frepoli, and G. Selvaggi (1993b), Active and remnant subducted slabs beneath Italy: Evidence from seismic tomography and seismicity, *Annali di Geofisica*, **36**, 201–214.
- Anderson, H., and J. Jackson (1987), The deep seismicity of the Tyrrhenian Sea, *Geophys. J. R. Astron. Soc.*, **91**, 613–637.
- Argnani, A. (2000), The southern Tyrrhenian subduction system: Recent evolution and neotectonic implications, *Ann. Geophys.*, **43**, 585–607.
- Barberi, F., P. Gasparini, F. Innocenti, and L. Villari (1973), Volcanism of the southern Tyrrhenian Sea and its geodynamic implications, *J. Geophys. Res.*, **78**, 5221–5232.
- Catalano, R., C. Dogliani, and S. Merlini (2001), On the Mesozoic Ionian Basin, *Geophys. J. Int.*, **144**, 49–64.
- Caputo, M., G. F. Panza, and D. Postpischil (1970), Deep structure of the Mediterranean Basin, *J. Geophys. Res.*, **75**, 4919–4923.
- Caputo, M., G. F. Panza, and D. Postpischil (1972), New evidences about the deep structure of the Lipari Arc, *Tectonophysics*, **15**, 219–231.
- Cimini, G. B. (1999), P-wave deep velocity structure of the southern Tyrrhenian subduction zone from nonlinear teleseismic traveltime tomography, *Geophys. Res. Lett.*, **26**(24), 3709–3712.
- Cimini, G. B. (2004), Tomographic studies of the deep structure of the Tyrrhenian-Apennine system, in *From Seafloor to Deep Mantle: Architecture of the Tyrrhenian Backarc Basin*, *Mem. Descr. Carta Geol. d'Ital.*, vol. LXIV, edited by M. P. Marani, F. Gamberi, and E. Bonatti, pp. 15–28, APAT, Rome.
- Cimini, G. B., and A. Amato (1993), P-wave teleseismic tomography contribution to the delineation of the upper mantle structure of Italy, in *Recent Evolution and Seismicity of the Mediterranean Region*, edited by E. Boschi, E. Mantovani, and A. Morelli, pp. 313–331, Springer, New York.
- Cimini, G. B., and P. De Gori (2001), Nonlinear P-wave tomography of subducted lithosphere beneath central southern Apennines (Italy), *Geophys. Res. Lett.*, **28**(23), 4387–4390.
- Cimini, G. B., and A. Marchetti (2006), Deep structure of peninsular Italy from seismic tomography and subcrustal seismicity, *Ann. Geophys.*, **49**(1), 331–345.

- Civello, S., and L. Margheriti (2004), Toroidal mantle flow around the Calabrian slab (Italy) from SKS splitting, *Geophys. Res. Lett.*, **31**, L10601, doi:10.1029/2004GL019607.
- Dahm, T., M. Thorwart, E.R. Flueh, T. Braun, R. Herber, P. Favali, L. Beranzoli, G. D'Anna, F. Frugoni, and G. Smriglio (2002), First deployment of new ocean bottom broadband seismological instruments in the Tyrrhenian Sea reveals unknown centre of submarine activity, *Eos Trans. AGU*, **83**(29), 309–320.
- Della Vedova, B., G. Pellis, J.-P. Foucher, and P. Rehault (1984), Geothermal structure of the Tyrrhenian Sea, *Mar. Geol.*, **55**, 271–289.
- Della Vedova, B., S. Bellani, G. Pellis, and P. Squarci (2001), Deep temperatures and surface heat flow distribution, in *Anatomy of an Orogen: The Apennines and Adjacent Mediterranean Basin*, edited by G. B. Vai and I. P. Martini, pp. 65–76, Springer, New York.
- Dercourt, J., et al. (1986), Geological evolution of the Tethys Belt from the Atlantic to the Pamirs since the Lias, *Tectonophysics*, **123**, 241–315.
- De Voogd, B., C. Truffert, N. Chamot-Rooke, P. Huchon, S. Lallemand, and X. Le Pichon (1992), Two-ship deep seismic soundings in the basin of the eastern Mediterranean Sea (Pasiphae cruise), *Geophys. J. Int.*, **109**, 536–552.
- Dewey, J. F., M. L. Helman, E. Turco, D. W. H. Hutton, and S. D. Knott (1989), Kinematics of the western Mediterranean, in *Alpine Tectonics*, edited by M. P. Coward, D. Dietrich, and R. G. Park, *Spec. Publ. Geol. Soc. London*, **45**, 265–283.
- Di Stefano, R., C. Chiarabba, F. Lucente, and A. Amato (1999), Crustal and uppermost mantle structure in Italy from the inversion of P-wave arrival times: geodynamics implications, *Geophys. J. Int.*, **139**, 483–498.
- Doglioni, C., F. Mongelli, and P. Pieri (1994), The Puglia uplift (SE Italy): An anomaly in the foreland of the Apenninic subduction due to buckling of a thick continental lithosphere, *Tectonics*, **13**, 1309–1321.
- Doglioni, C., F. Innocenti, and G. Mariotti (2001), Why Mt. Etna?, *Terra Nova*, **13**, 25–31.
- Dvorkin, J., A. Nur, G. Mavko, and Z. Ben Avraham (1993), Narrow subduction slabs and the origin of back-arc basins, *Tectonophysics*, **227**, 63–79.
- Dziewonski, A. M., and F. Gilbert (1976), The effect of small, aspherical perturbations on travel times and a re-examination of the corrections for the ellipticity, *Geophys. J. R. Astron. Soc.*, **44**, 7–17.
- Faccenna, C., P. Davy, J. P. Brun, R. Funicello, D. Giardini, M. Mattei, and T. Nalpas (1996), The dynamics of back-arc extension: An experimental approach to the opening of the Tyrrhenian Sea, *Geophys. J. Int.*, **126**, 781–795.
- Faccenna, C., F. Funicello, D. Giardini, and P. Lucente (2001), Episodic back-arc extension during restricted mantle convection in the central Mediterranean, *Earth Planet. Sci. Lett.*, **187**, 105–116.
- Favali, P., et al. (2002), Towards a permanent deep sea observatory: the GEOSTAR European experiment, in *Science-Technology Synergy for Research the Marine Environment: Challenges for the XXI Century. Developments in Marine Technology*, vol. 12, edited by L. Beranzoli, P. Favali, and G. Smriglio, pp. 111–120, Elsevier, New York.
- Favali, P., et al. (2006), A fleet of multiparameter observatories for geophysical and environmental monitoring at seafloor, *Ann. Geophys.*, **49**(2/3), 659–680.
- Finetti, I., and A. Del Ben (1986), Geophysical study of the Tyrrhenian opening, *Bol. Geofis. Teor. Appl.*, **28**, 75–156.
- Francalanci, L., S. R. Taylor, M. T. McCulloch, and J. Woodhead (1993), Geochemical and isotopic variations in the calc-alkaline rocks of Aeolian Arc, southern Tyrrhenian Sea, Italy: Constraints on magma genesis, *Contrib. Mineral. Petrol.*, **113**, 300–313.
- Giardini, D., and M. Velonà (1991), The deep seismicity of the Tyrrhenian Sea, *Terra Nova*, **3**, 57–64.
- Gvirtzman, Z., and A. Nur (1999), The formation of Mount Etna as the consequence of slab rollback, *Nature*, **401**, 782–785.
- Gvirtzman, Z., and A. Nur (2001), Residual topography, lithospheric structure and sunken slabs in the central Mediterranean, *Earth Planet. Sci. Lett.*, **187**, 117–130.
- Hirahara, K., and A. Haseami (1993), Tomography of subduction zones using local and regional earthquakes and teleseisms, in *Seismic Tomography: Theory and Practice*, edited by H. M. Iyer and K. Hirahara, pp. 519–562, CRC Press, Boca Raton, Fla.
- Ito, S., R. Hino, S. Matsumoto, H. Shiobara, H. Shimamura, T. Kanazawa, T. Sato, J. Kasahara, and A. Hasegawa (2000), Deep seismic structure of the seismogenic plate boundary in the off-Sanriku region, north-eastern Japan, *Tectonophysics*, **319**, 261–274.
- Kastens, K., et al. (1988), ODP Leg 107 in the Tyrrhenian Sea: Insights into passive margin and back-arc basin evolution, *Geol. Soc. Am. Bull.*, **100**, 1140–1156.
- Kennett, B. L. N., E. R. Engdahl, and R. Buland (1995), Constraints on seismic velocities in the Earth from traveltimes, *Geophys. J. Int.*, **122**, 108–124.
- Kincaid, C., and R. W. Griffiths (2003), Laboratory models of the thermal evolution of the mantle during rollback subduction, *Nature*, **425**, 58–62.
- Koch, M. (1985), A numerical study on the determination of the 3-D structure of the lithosphere by linear and non-linear inversion of teleseismic travel times, *Geophys. J. R. Astron. Soc.*, **80**, 73–93.
- Leveque, J.-J., and F. Masson (1999), From ACH tomographic models to absolute velocity models, *Geophys. J. Int.*, **137**, 621–629.
- Lucente, F. P., C. Chiarabba, G. B. Cimini, and D. Giardini (1999), Tomographic constraints on the geodynamic evolution of the Italian region, *J. Geophys. Res.*, **104**(B9), 20,307–20,327.
- Marani, M. P., and T. Trua (2002), Thermal constriction and slab tearing at the origin of a super-inflated spreading ridge, Marsili Volcano (Tyrrhenian Sea), *J. Geophys. Res.*, **107**(B9), 2188, doi:10.1029/2001JB000285.
- Marani, M. P., F. Gamberi, and E. Bonatti (Eds.) (2004), From Seafloor to Deep Mantle: Architecture of the Tyrrhenian Backarc Basin, *Mem. Descr. Carta Geol. d'Ital.*, vol. LXIV, 195 pp.
- Malinverno, A., and W. B. F. Ryan (1986), Extension in the Tyrrhenian sea and shortening in the Apennines as result of arc migration driven by the sinking of the lithosphere, *Tectonics*, **5**, 227–245.
- Menke, W. (1989), *Geophysical Data Analysis: Discrete Inverse Theory*, *Int. Geophys. Ser.*, 289 pp., Elsevier, New York.
- Morelli, C. (1970), Physiography, gravity and magnetism of the Tyrrhenian Sea, *Bol. Geofis. Teor. Appl.*, **12**(48), 275–311.
- Morelli, C. (1997), Recent deeper geophysical results better account for the tectonics in the Italian area, *Ann. Geophys.*, **40**, 1345–1358.
- Nur, A., J. Dvorkin, G. Mavko, and Z. Ben Avraham (1991), Speculations on the origin and fate of back-arc basins, *Ann. Geophys.*, **36**, 155–163.
- Pascucci, V., S. Merlini, and P. Martini (1999), Seismic stratigraphy of the Miocene-Pleistocene sedimentary basin of the northern Tyrrhenian Sea and western Tuscany (Italy), *Basin. Res.*, **11**, 337–356.
- Patacca, E., and P. Scandone (1989), Post-Tortonian mountain building in the Apennines, the role of the passive sinking of a relic lithosphere, in *The Lithosphere in Italy: Advances in Earth Sciences Research*, vol. 80, edited by A. Boriani, et al., pp. 157–176, Acc. Naz. Lincei, Rome.
- Patacca, E., R. Sartori, and P. Scandone (1992), Tyrrhenian basin and Apenninic arcs: Kinematic relations since late Tortonian times, *Mem. Soc. Geol. Ital.*, **45**, 425–451.
- Piomallo, C., and A. Morelli (1997), Imaging the Mediterranean upper mantle by P-wave travel time tomography, *Ann. Geophys.*, **40**, 963–979.
- Piomallo, C., and A. Morelli (2003), P-wave tomography of the mantle under the Alpine-Mediterranean area, *J. Geophys. Res.*, **108**(B2), 2065, doi:10.1029/2002JB001757.
- Rosenbaum, G., and G. S. Lister (2004), Neogene and Quaternary rollback evolution of the Tyrrhenian Sea, the Apennines, and the Sicilian Maghrebides, *Tectonics*, **23**, TC1013, doi:10.1029/2003TC001518.
- Sartori, R. (1989), Evoluzione neogenico-recente del bacino tirrenico e i suoi rapporti con la geologia delle aree circostanti, *Giorn. Geol.*, **3**(51/2), 1–39.
- Scarpa, R. (1982), Travel-time residuals and three-dimensional velocity structure of Italy, *Pure Appl. Geophys.*, **120**, 583–606.
- Selvaggi, G., and C. Chiarabba (1995), Seismicity and P-wave velocity image of the southern Tyrrhenian subduction zone, *Geophys. J. Int.*, **121**, 818–826.
- Sgroi, T., T. Braun, T. Dahm, and F. Frugoni (2006), An improved seismicity picture of the southern Tyrrhenian area by the use of OBS and land-based networks: The TYDE experiment, *Ann. Geophys.*, **49**(2/3), 801–817.
- Spakman, W. (1990), Tomographic images of the upper mantle below central Europe and the Mediterranean, *Terra Nova*, **2**, 542–552.
- Spakman, W., and R. Wortel (2004), A tomographic view on western Mediterranean Geodynamics, in *The Transmed Atlas*, p. 31–52, Springer, New York.
- Spakman, W., S. Van der Lee, and R. der Hilst (1993), Travel-time tomography of the European-Mediterranean mantle down to 1400 km, *Phys. Earth Planet. Inter.*, **79**, 3–74.
- Speranza, F., L. Sagnotti, and M. Mattei (1997), Tectonics of the Umbria-Marche-Romagna Arc (central-northern Apennines, Italy): New paleomagnetic constraints, *J. Geophys. Res.*, **102**, 3153–3166.
- Steck, L. K., and W. A. Prothero (1991), A 3-D raytracer for teleseismic body-wave arrival times, *Bull. Seismol. Soc. Am.*, **81**, 1332–1339.
- Trua, T., G. Serri, and M. P. Marani (2003), Lateral flow of African mantle below the nearby Tyrrhenian plate: Geochemical evidence, *Terra Nova*, **15**, 433–440, doi:10.1046/j.1365-3121.2003.00509.
- Weiland, C. M., L. K. Steck, P. B. Dawson, and V. A. Korneev (1995), Nonlinear teleseismic tomography at Long Valley caldera, using three-dimensional minimum travel time ray tracing, *J. Geophys. Res.*, **100**, 20,379–20,390.

- Wilson, M., and G. Bianchini (1999), Tertiary-Quaternary magmatism within the Mediterranean and surrounding regions, in *The Mediterranean Basins: Tertiary Extension Within the Alpine Orogen*, edited by B. Durand et al., *Geol. Soc. Spec. Publ.*, 156, 141–168.
- Zhao, D., Y. Xu, D. A. Wiens, L. Dormoan, J. Hildebrand, and S. Webb (1997), Depth extent of the Lau Back-arc spreading centre and its relation to subduction processes, *Science*, 278, 254–257.
- Zito, G., F. Mongelli, S. De Lorenzo, and C. Doglioni (2003), Heat flow and geodynamics in the Tyrrhenian Sea, *Terra Nova*, 15, 425–432.

G. B. Cimini, P. Favali, and C. Montuori, INGV, Via di Vigna Murata, 605, 00143 Roma, Italy. (montuori@ingv.it)