Imaging the Mediterranean upper mantle by *P*-wave travel time tomography

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Abstract
Travel times of *P*-waves in the Euro-Mediterranean region show strong and consistent lateral variations, which can be associated to structural heterogeneity in the underlying crust and mantle. We analyze regional and teleseismic data from the International Seismological Centre data base to construct a three-dimensional velocity model of the upper mantle. We parameterize the model by a 3D grid of nodes – with approximately 50 km spacing – with a linear interpolation law, which constitutes a three-dimensional continuous representation of *P*-wave velocity. We construct summary travel time residuals between pairs of cells of the Earth's surface, both inside our study area and – with a broader spacing – on the whole globe. We account for lower mantle heterogeneity outside the modeled region by using empirical corrections to teleseismic travel times. The tomographic images show general agreement with other seismological studies of this area, with apparently higher detail attained in some locations. The signature of past and present lithospheric subduction, connected to Euro-African convergence, is a prominent feature. Active subduction under the Tyrrhenian and Hellenic arcs is clearly imaged as high-velocity bodies spanning the whole upper mantle. A clear variation of the lithospheric structure beneath the Northern and Southern Apennines is observed, with the boundary running in correspondence of the Ortona-Roccamonfina tectonic lineament. The western section of the Alps appears to have better developed roots than the eastern, possibly reflecting a difference in past subduction of the Tethyan lithosphere and subsequent continental collision.

Key words *seismology – tomography – travel time — Mediterranean region*

1. Introduction

The Alpine-Mediterranean region is the area of interaction between African and Eurasian plates, and is marked by high tectonic activity. Closure of the Tethys and still active convergence (*e.g.*, Dercourt *et al.*, 1986; Dewey *et al.*, 1989) are responsible for a complex geologic environment, still far from being completely understood. A basic step for improving our insight into the kinematics and dynamics of the region is the knowledge of the structure of the lithosphere and mantle, which bear the signature of active and recent tectonic evolution. An effective way to retrieve deep Earth structure, in a wide range of spatial scales, consists of imaging seismic wave velocity variations in three dimension with the methods of seismic tomography. This technique has already been applied with success to the Alpine-Mediterranean domain by various authors – among others, see, for instance, Babuska *et al.* (1984, 1990); Granet and Trampert (1989); Amato *et al.* (1993); Spakman *et al.* (1993). The last study, in particular, showed that regional dis-

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tance arrivals bring better resolution than teleseismic rays and can effectively be employed to derive significant pictures of three dimensional variation of *P*-wave velocity. Regional wave propagation is complicated by the presence of discontinuities in the upper mantle. Teleseismic rays alone are simpler to model but lack sufficient vertical resolution.

We follow therefore the suggestion by Spakman *et al.* (1993) and analyze *P* travel time data from local, regional and teleseismic distance reported by the Bulletins of the International Seismological Centre. The first motivation for our study is to provide an independent – although methodologically similar – analysis of travel time data to compare to previous results, in this way contributing to the assessment of the confidence of seismological models. In fact, reliability of tomographic models is always difficult to ascertain. A main reason lies in the fact that travel time data are affected by significant signals which cannot be modeled, usually regarded as random noise but, to some extent, systematic. This is a possible source of bias for tomographic models (Spakman *et al.*, 1993). Comparison among results of different studies is perhaps the most convincing – although empirical – test of their validity.

As another goal of our study we attempt to reach higher resolution than previously attained. Having access to an increased data set we employ a more detailed representation of the model. Our parameterization is inherently smooth, thus avoiding the need for image filtering, often done to ease interpretation of tomograms.

In the following, we briefly describe the method and the resulting tomographic model, and comment on the differences with previously published work. The main objective for this study is the assessment of the appropriateness of our tomographic inversion scheme. The resulting model may therefore be considered a first result, which we regard as highly significant but possibly still open to future improvement. For this reason, in the discussion we restrict ourselves to briefly commenting on the most prominent features of the model.

2. Data and method

We extract *P*-wave travel time data from the Bulletins of the International Seismological Centre (ISC) published for the years between 1964 and 1989. We only use earthquakes with focal depth less than 50 km, because they have a more uniform geographical distribution. Subcrustal earthquakes may add important information in areas of high interest, where lithospheric subduction takes place, but would considerably complicate the scheme we follow, and we plan to add them only in a future stage of our work.

A preliminary quality selection is based on several requirements, and is meant to isolate well recorded earthquakes. We select earthquakes for which a minimum of 30 *P* arrivals are reported worldwide, covering at least 6 out of 8 azimuthal sectors. We extract from the Bulletins both teleseismic and regional first-arrival travel times, in the distance range $0° \div 90°$, and use them firstly to locate all the events using the one-dimensional reference model SP6 (Morelli and Dziewonski, 1993), standard corrections for ellipticity (Dziewonski and Gilbert, 1976) and station elevation. Earthquake location is carried out using an approximation to the uniform reduction scheme (Buland, 1986). All available first arrival *P* phases, at teleseismic and regional distance worldwide, are used to locate the events, and all rays to or from the Mediterranean region (either source or receiver in the study area) are stored for subsequent use.

After these preliminary steps, we proceed to compute the *summary rays* which are the input to the inversion stage. Each summary ray is representative of all individual rays connecting the same pair of cells in which the Earth's surface is divided. On a global scale, the cells are trapezoids limited by parallels and meridians, approximately equi-areal and corresponding to $5° \times 5°$ at the equator. In the Mediterranean area – defined for our purposes as the region with latitude between 30° and 50°, and longitude between $-15°$ and $40°$ – cells are $0.5° \times 0.5°$ in size. A summary ray includes all rays connecting the same pair of cells, regard-less of which of the two cells contains the source and which the receiver. To each summary ray a residual is associated, defined as the average of all individual residuals composing it. In our analyses we take into account only individual residuals smaller than 7 s. Three main advantages descend from the use of summary rays: it reduces biases due to regions with a favourable (clustered) distribution of stations or events, it filters out the effects of small-scale heterogeneity and it reduces the redundancy of data in the inversions (Morelli and Dziewonski, 1991; Vasco *et al.*, 1994; Robertson and Woodhouse, 1995). This procedure isolates 25784 events and 2784228 rays, resulting in 92426 regional summary rays and 82325 teleseismic summary rays.

Only summary rays with residuals less than 3 s, and composed by at least 3 individual observations, are used in the inversion. The inversion domain is approximately 4000 km in E-W direction, 2200 km in N-S direction, and 700 km in depth. The model is parameterized as a perturbation of the *P*-wave slowness field represented by a grid of nodes, with about 50 km spacing both horizontally and vertically, and a tri-linear interpolation law. If compared, for instance, to a model involving blocks, the representation using linear interpolation between grid nodes gives a more realistic description of the velocity field, using the same number of parameters (Thurber, 1983; Nolet, 1996). Moreover, no graphical smoothing is required – as often done on block-type models to ease interpretation which could, to some extent, degrade the fit achieved by the optimization procedure. Rays are always traced in the one-dimensional background model SP6. This is a legitimate approximation if velocity gradients are sufficiently small. This condition may not be met in regions of strongest anomalies, where 3D ray tracing produces sharper images (Papazachos and Nolet, 1997) but does not alter the over-all picture of anomalies. The inversion is performed using the intrinsically damped LSQR algorithm (Paige and Saunders, 1982; Nolet, 1987) and an additional explicit smoothing factor introduced through an isotropic gradient minimization condition.

3. The model

The most prominent feature shown by the ray density maps of fig. 1 is the rather irregular path coverage of the study area. Ray density maps – described by a logarithmic color-coded scale – represent a quantity similar in meaning to the cell hit count usually considered for block parameterizations. In our case, the ray density is defined as the cumulative value of the partial derivative of slowness with respect to travel time. It is, in fact, the amount of information on the velocity (slowness) field that a given ray configuration can achieve. In our model representation, this is a continuous field because of the linear interpolation.

The maps in fig. 1 bear the consequences of the irregular distribution of seismicity and seismographic stations in this area. Black indicates no sampling or extremely poor illumination by ray paths. Earthquakes in the Hellenic and Aegean regions, recorded by European stations, yield a good coverage of the middle-eastern part of the model (brighter colours). Ray illumination within this highly sampled region is expected to be anisotropic, with most rays travelling along a southeast-northwest direction. This strong irregularity is especially present at shallower depth – see the map at 50 km depth, mostly influenced by P_n paths – and attenuates with increasing depth.

Another evident feature in ray density maps is the higher sampling of the interior of the domain with respect to the periphery. This is due to discarding regional-distance rays ($\Delta < 25°$) which do not entirely lie within the model. We operate such a selection to prevent mapping of lateral heterogeneity from outside the inversion domain into our model. This disparity of sampling attenuates with depth, as a consequence of the lateral broadening of coverage provided by teleseismic rays. We take lower mantle heterogeneity into account, when considering teleseismic rays, by using average teleseismic summary residuals as correction. These corrections are computed for all cells located at teleseismic distance. For each of these cells the correction is obtained by the average of all summary residuals connecting it to Mediterranean cells. Teleseismic corrections are as-

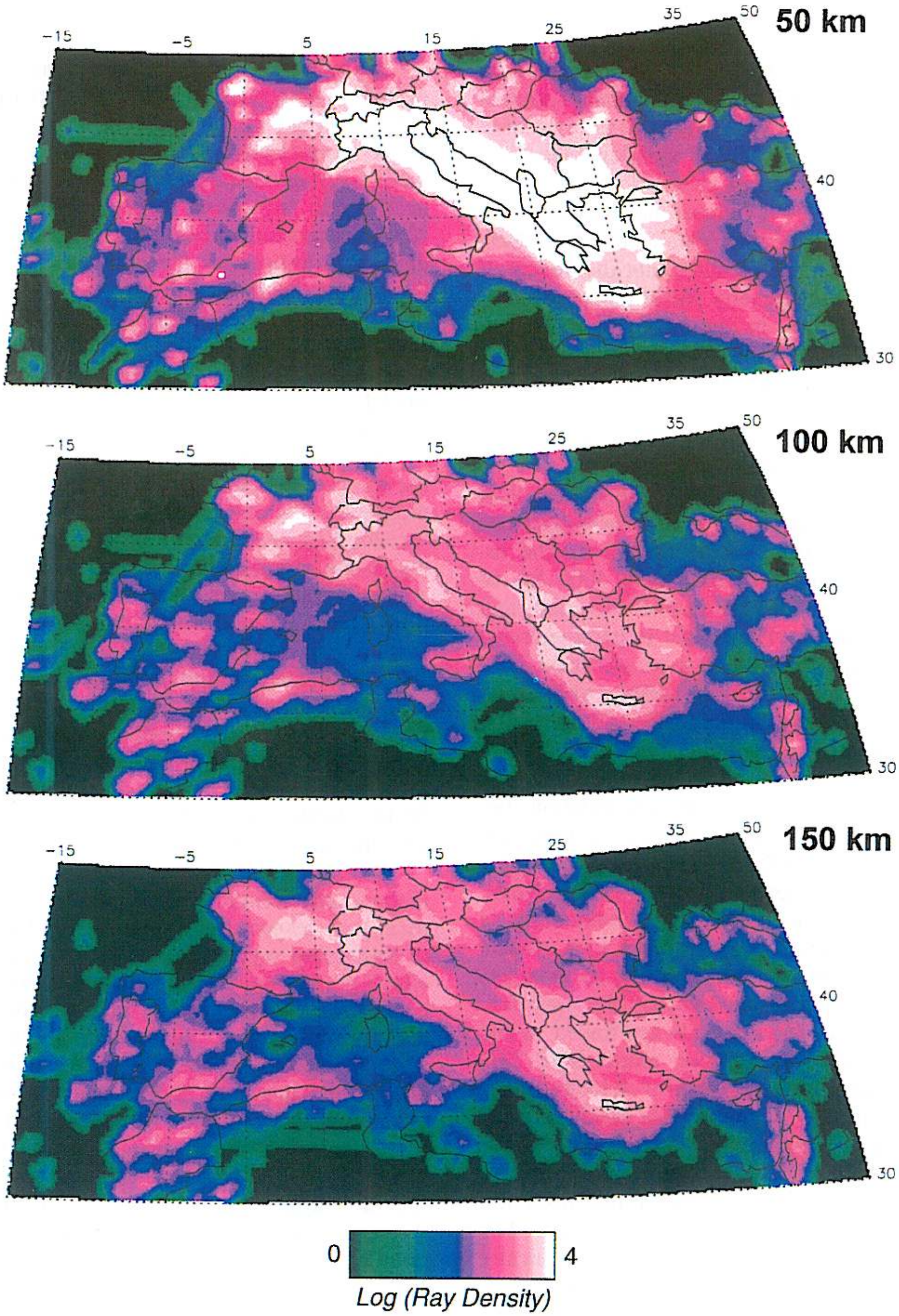
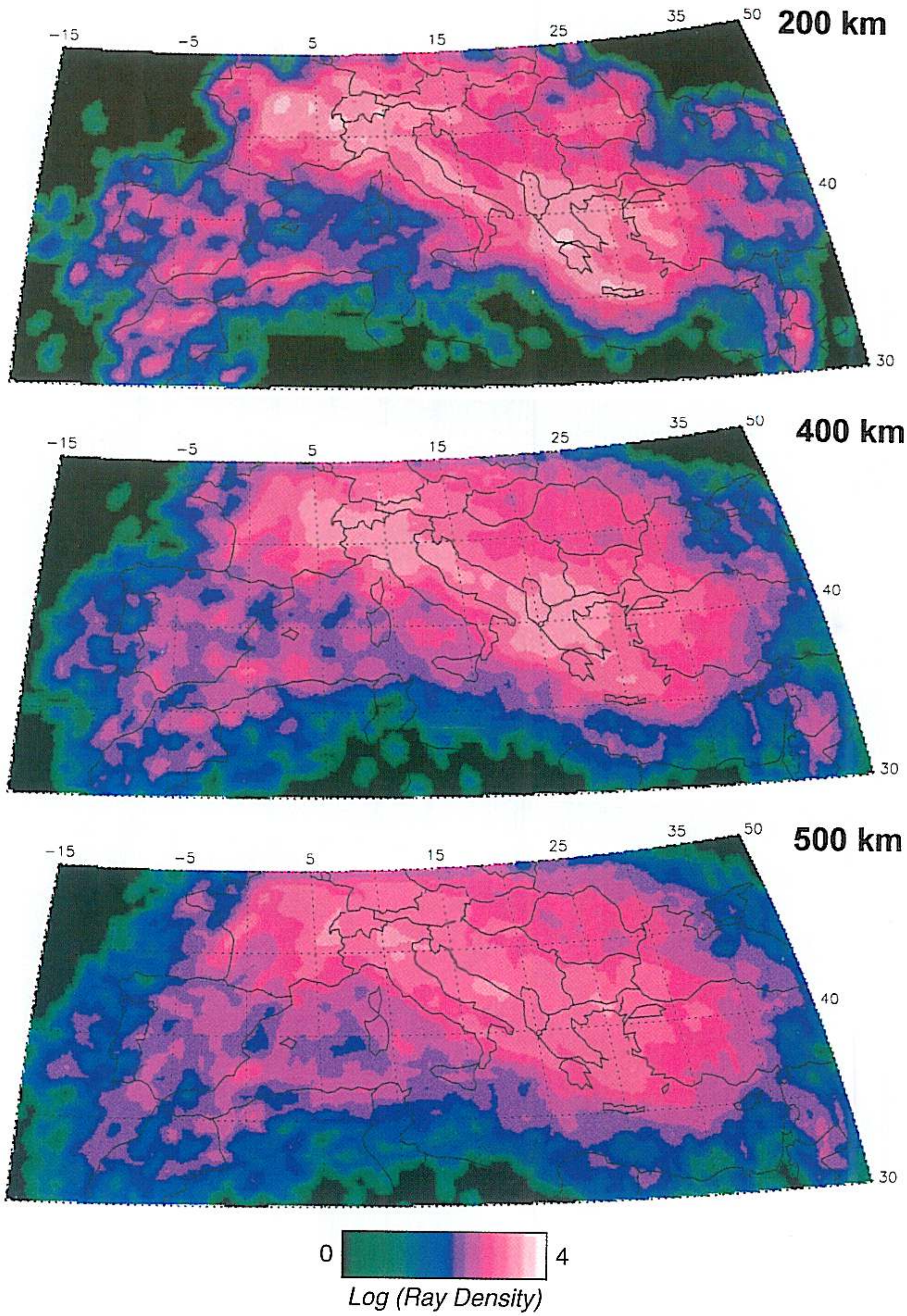


Fig. 1. Ray density maps in the inversion domain. Ray density is the cumulative value of the partial derivative of slowness with respect to travel time, and represents the density of information given by the used



data set. Color scale is proportional to the logarithm of actual density. Brighter colors represent higher information. Maps refer to different depths: 50, 100, 150, 200, 400 and 500 km.

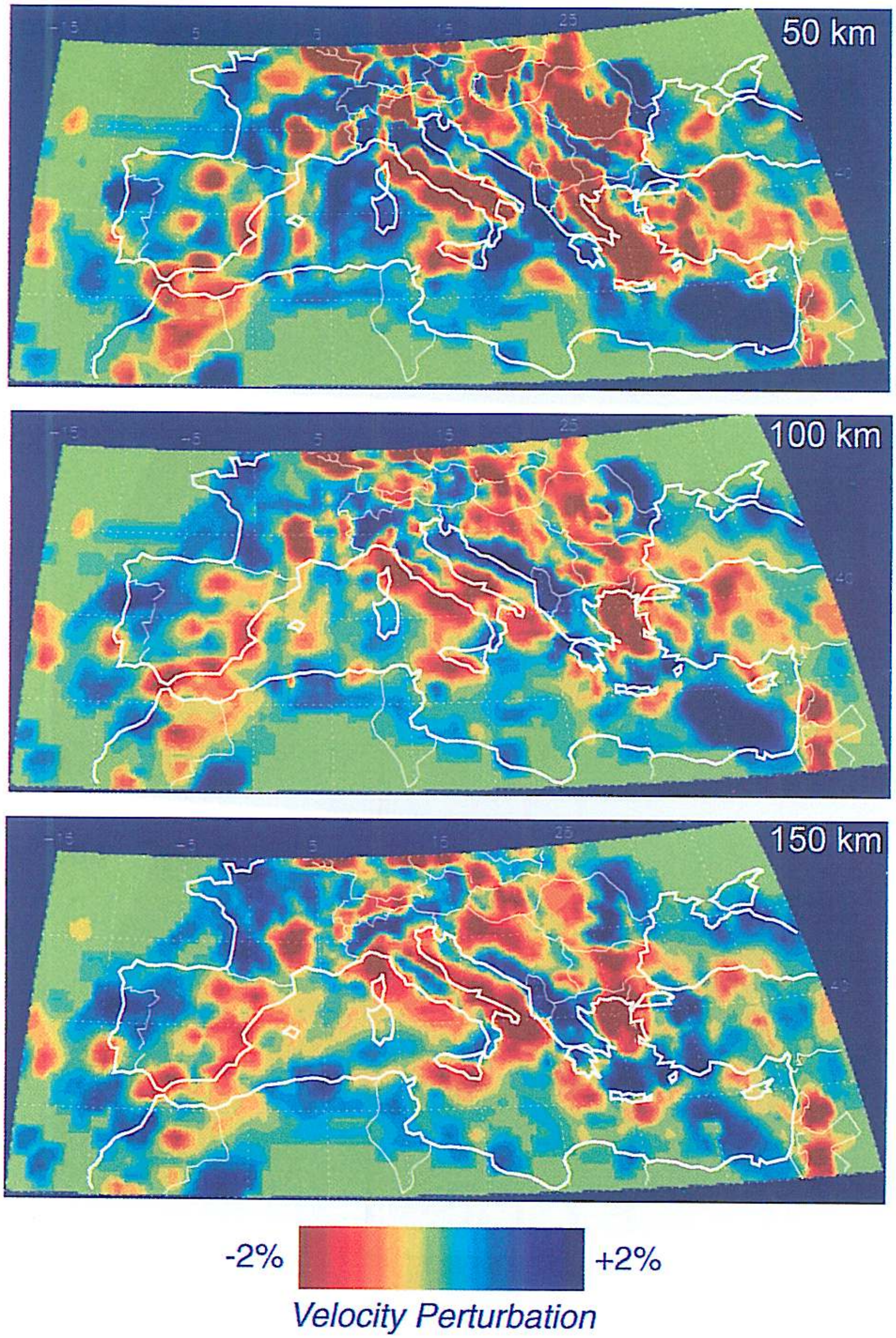
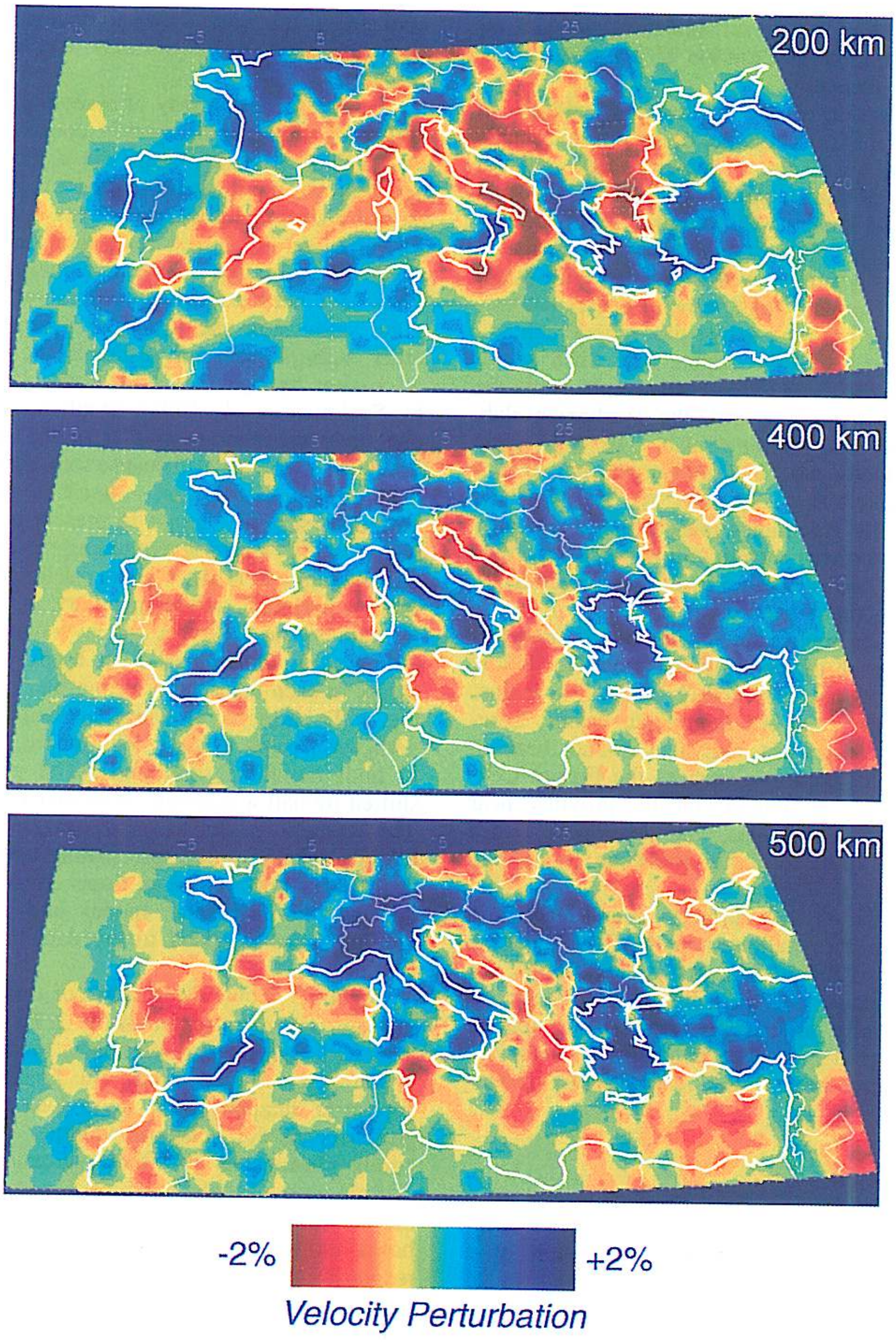


Fig. 2. Three-dimensional model of lateral heterogeneity in *P*-wave velocity, represented as percentage perturbation with respect to the reference, radial model. In the color code, as usual, red represents slower than



average, while blue is faster. Maps refer to different depths: 50, 100, 150, 200, 400 and 500 km. The color scale (covering ± 2%) is the same for all depths.

sumed to represent an empirical account of large-scale lateral variation of lower mantle *P* velocity structure.

The result of the inversion is plotted in fig. 2 as maps, at different depths, of percentage variation of *P*-wave velocity with respect to our reference model SP6. The resulting 3D model gives 20% variance reduction for summary residuals. Larger-scale anomalies in fig. 2 are generally well correlated with major tectonic features. High-velocity bodies are present along past and active lithospheric subduction zones along the Alpine belt. Such examples can be found, for instance, beneath the Maghrebides, along with the Tyrrhenian slab, Dinarides and Hellenic slab. The Alps have high-velocity roots and the Carpathians show a deep seated fast anomaly. At 100 km and deeper, the Apennines present a fast anomaly in their northern section, while there is no evidence of high velocities in the shallower layers of the southern part.

Another significant feature of the model is the presence of low-velocity anomalies below the extensional basins, such as the Tyrrhenian, the Aegean, and the Pannonian basins. Low-velocity anomalies are also found beneath some volcanic areas, such as the French Central Massif, the Italian volcanic province near the Tyrrhenian coast, the Moroccan Atlas and the Strait of Sicily (between Sicily and Tunisia).

Correlation with surface tectonic features is better for the shallowest part of the model – down to 200-250 km. At greater depth, the model appears to lose detail, both vertically and laterally, and mainly shows a fast feature concentrated on a central, belt-like, anomaly meandering around the Africa-Eurasia boundary.

4. Discussion

A tomographic model is the result of a rather long procedure involving a number of critical steps, such as data selection, pre-processing, removal of known sources of contamination of the signal, and, finally, inversion of a large, under-constrained matrix built over inaccurate data. It is crucial then to understand how trustworthy the result of such a complicated course is.

A rather common way of assessing the reliability of tomographic models consists of so-called synthetic tests, where the ability to image a known input model – often a regular pattern of alternating positive and negative velocity variations – is tested. Such tests mainly address the resolution of the inversion scheme. They have been extensively considered, for the same study area and a similar procedure, by other authors (see, for instance, the detailed discussion by Spakman and Nolet, 1988, and by Spakman *et al.*, 1993). Synthetic tests for our study result quite similar to the study of Spakman *et al.* (1993) because they mainly depend on data distribution – much the same for both analyses. We choose to present here only one such test, aimed at evaluating the ability of our technique to retrieve greater detail. We perform a harmonic model recovery test, where we address the ability of our inversion scheme to image an input pattern given by the superposition, in each layer, of two sine functions giving a 2D harmonic slowness anomaly pattern, with wavelength of four cells (about 220 km) in both horizontal directions. The pattern is shifted by half a wavelength in adjacent layers. The synthetic model has ± 4% maximum amplitudes relative to the surrounding upper mantle velocity profile. Travel times are computed through this synthetic model and gaussian random noise is added to the data before inversion.

Figure 3 shows the results for this test for the same layers presented in fig. 2. By comparison with fig. 1 the resolution is much higher in the best sampled areas. The amplitude response is damped in the areas located in the periphery of the inversion volume which are barely illuminated by either regional or teleseismic rays. Poorly sampled areas, such as the central part of the Western Mediterranean basin, show clear evidence of scarce amplitude response in the shallower layers and smearing between adjacent nodes. However, the largest part of the adequately sampled inversion volume exhibits good resolution of these intermediate-scale structures.

It should be stressed, however, that sources of uncertainty exist at every step of the complex process leading to a tomographic model. The synthetic tests mentioned and shown cannot therefore entirely answer the question of how dependable a tomographic model is, as they are designed to specifically address inversion resolution for specific features (Leveque *et al.*, 1993). This is why, in the comparison of this study with EUR89B, the synthetic test responses are much more similar than the actual models. Every single step or assumption should instead be tested independently, but this is a formidable effort – much larger than derivation of a model – which has never been tackled so far. We think that an alternative, empirical but very significant, assessment of model reliability is however represented by an *a posteriori* comparison with results obtained by other authors with independent analyses, and by a discussion in view of information derived from tectonic research. Good agreement with other studies is convincing evidence in favour of the reliability of tomographic images derived from different works. Our technique has many differences with respect to that adopted, for instance, by Spakman *et al.* (1993) for EUR89B – the most similar study. The main points are that: we use more stringent criteria for data selection; we use a different starting reference model, presenting discontinuities in the upper mantle, but no low velocity zone; we perform a preliminary relocation of all events using the same reference model employed throughout the whole linearization procedure; we use a continuous representation rather than a discontinuous one (blocks) to describe the model and to compute partial derivatives; we do not smooth the model prior to plotting; we compute (symmetric) summary rays, rather than *composite* rays; we do not perform simultaneous relocation; we do not simultaneously compute station corrections, but rather use average empirical teleseismic corrections: we have a finer parameterization. These differences are such that we may expect that contaminating signal biases the two models in different ways. We will briefly outline here the main issues stemming from comparison of our model with EUR89B and other studies of this region.

The Africa-Eurasia plate kinematics for at least the last 150 My (*e.g.*, Dercourt *et al.* 1986; Dewey *et al.*, 1989) implies consumption of lithospheric material connected to the closure of the Tethys. Signature of past and present subduction is therefore expected in the upper mantle beneath the Mediterranean. This shows up with rather clear evidence, and was noted already in model EUR89B by Spakman *et al.* (1993). In our model this signature is even more clear. De Jonge *et al.* (1994) used published reconstruction of the late Mesozoic and Cenozoic tectonic evolution of the region to calculate the resulting *P*-wave velocity field, which we may regard as a possible expectation for a tomographic model. The positive velocity anomaly due to the subducted lithosphere is very clear at 200 km depth in our model, in agreement with theirs, inferred from tectonic reconstruction. At this depth, a high-velocity belt runs from the North African coast, beneath the Maghrebides, to the Southern Tyrrhenian, beneath the Alps, along the Dinarides and on to the Hellenic subduction. The comparison is excellent. As we go deeper, this belt becomes even more clearly the dominant feature of our model. The now extinct subduction of oceanic lithosphere beneath the Carpathians (Oncescu *et al.*, 1984), in the Vrancea region, shows only in layers at 150 km and below. Another deep-seated positive anomaly is imaged below the Betic-Alboran Sea, starting at about 200 km depth and extending downwards, in agreement with the slab structure shown by the detailed tomographic study by Blanco and Spakman (1993).

The Apennines present a distinct difference between the northern and the southern sections. This boundary coincides with a well-known tectonic lineament, the Ortona-Roccamonfina line (Patacca *et al.*, 1990). In the shallower 150 km, where slow material is present beneath the Southern Apennines, a fast anomaly appears below the northern section. The presence of a high-velocity body beneath the Northern Apennines is consistent with subcrustal seismicity observed beneath the Northern Apennines by Selvaggi and Amato (1992) who interpreted it

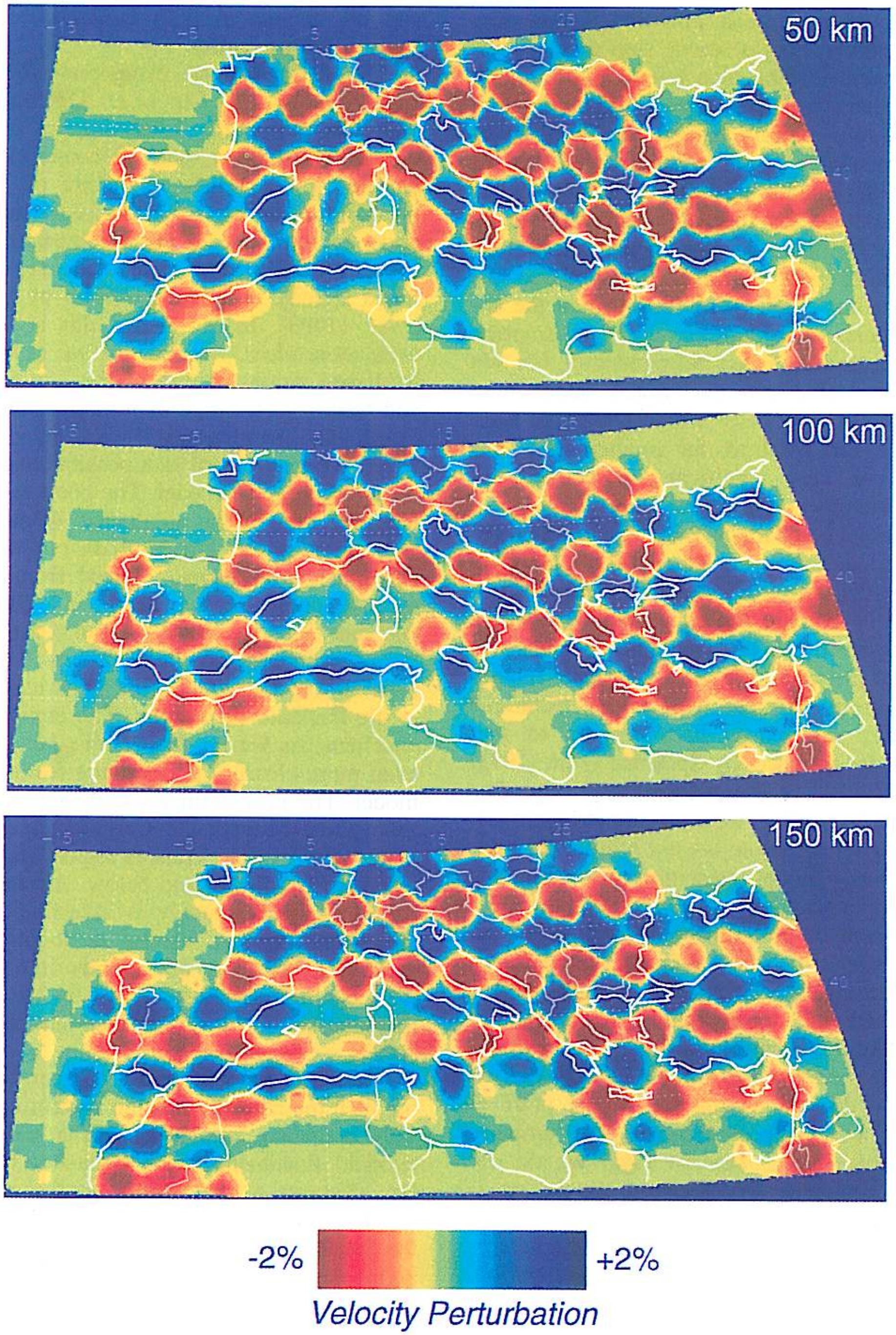
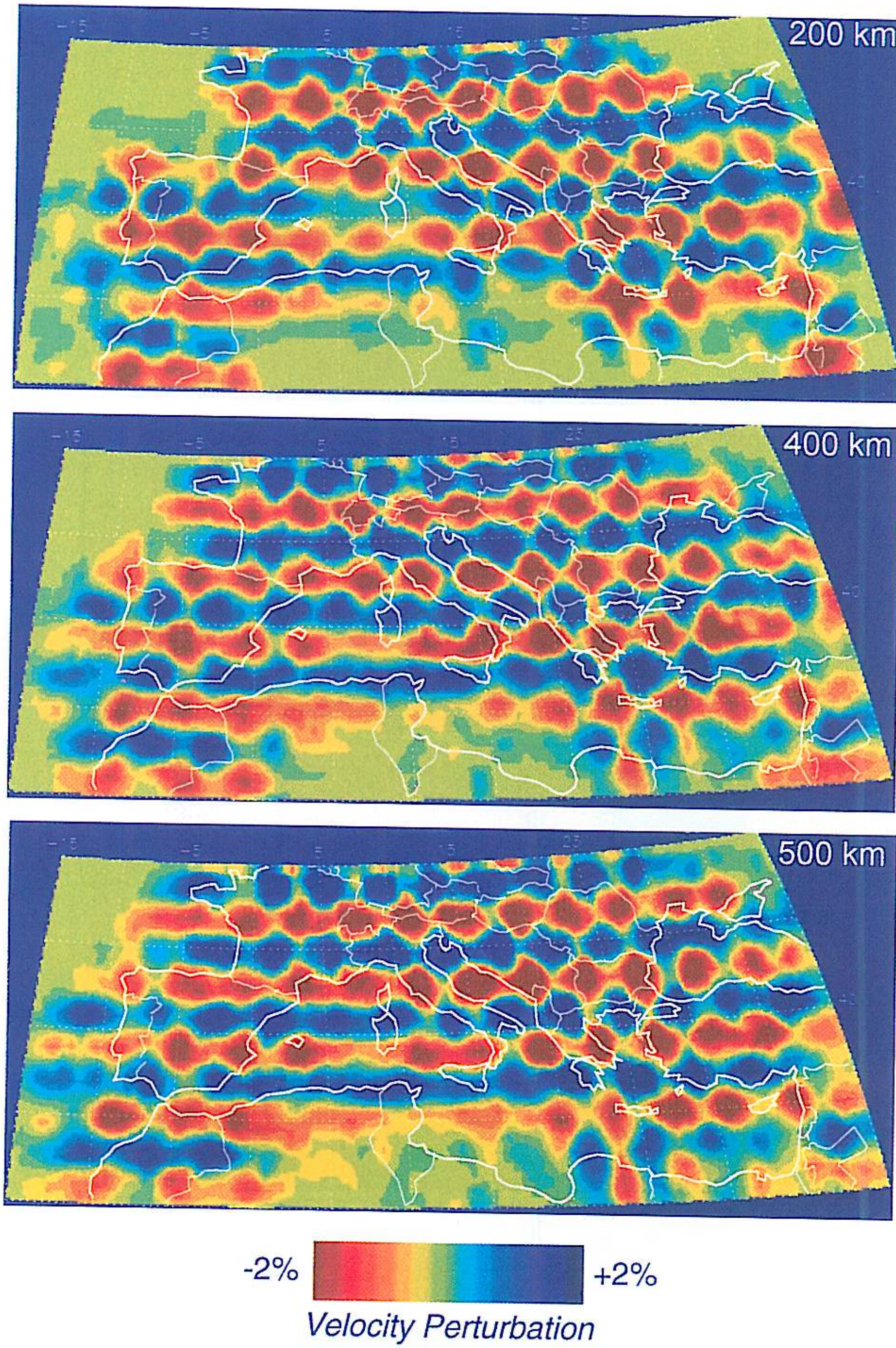


Fig. 3. The inversion response to the harmonic model test with four cells wavelength (~ 220 km) and noisy data, represented as percentage velocity perturbation with respect to the reference model.



Maps refer to different depths: 50, 100, 150, 200, 400 and 500 km. The color scale is the same as in fig. 2.

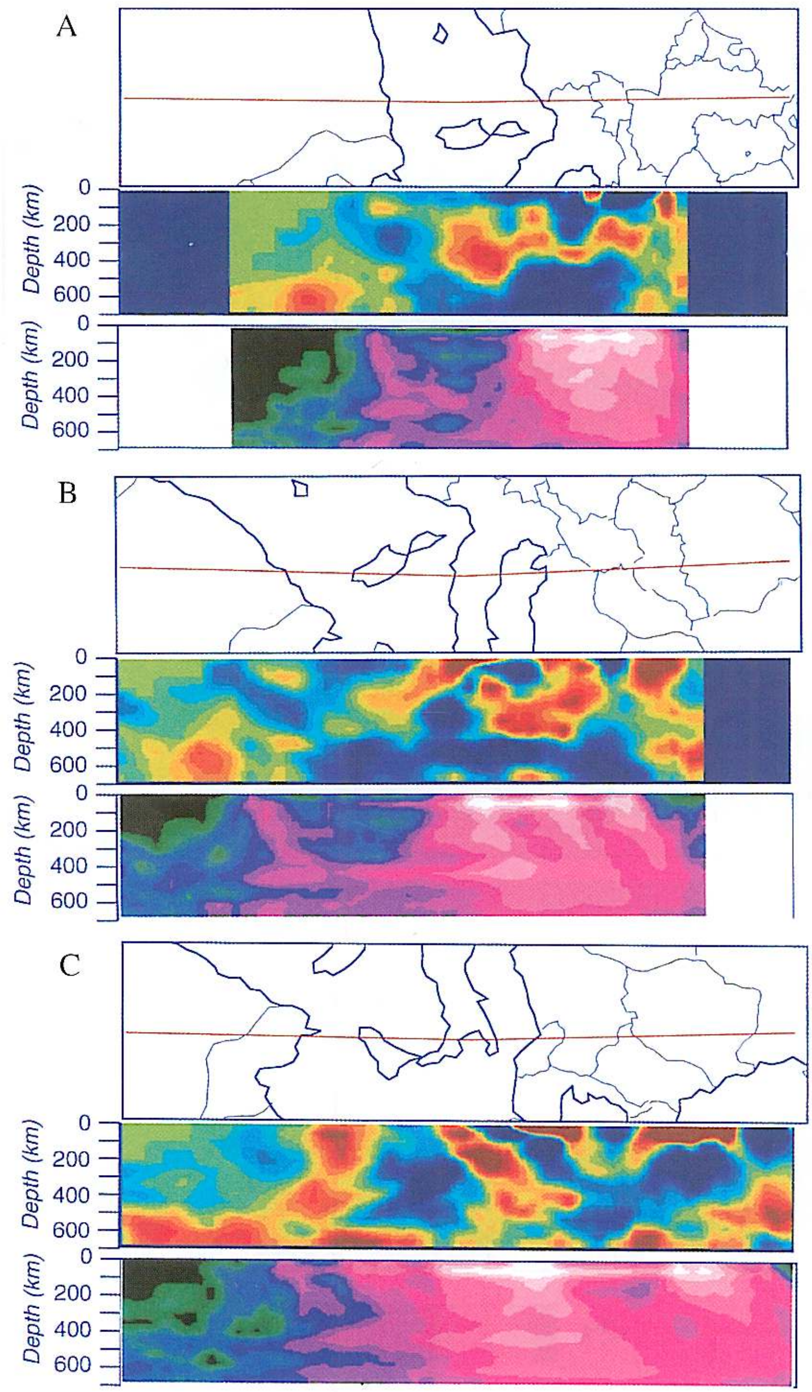
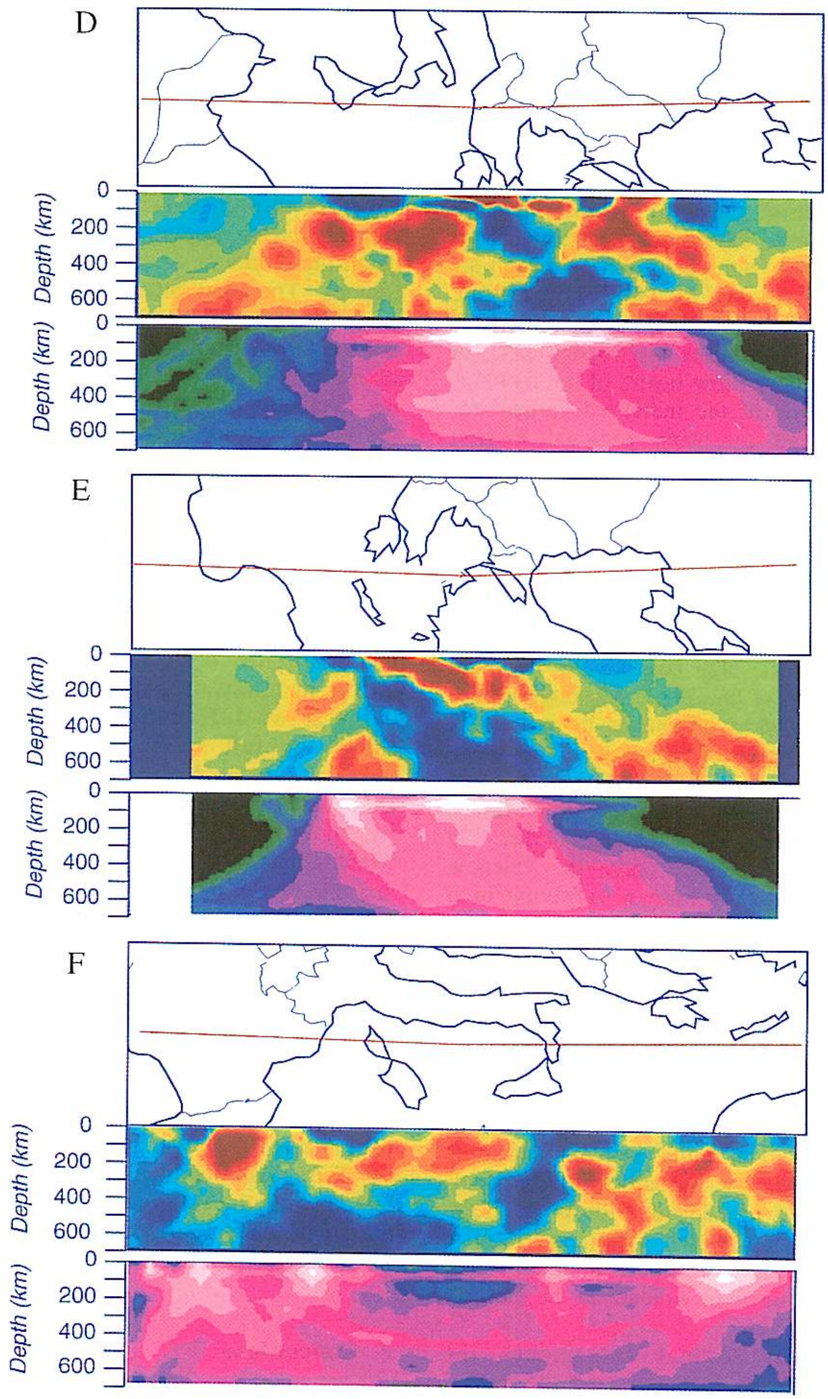


Fig. 4. Vertical cross-sections cut through the tomographic model. In each panel, the top subpanel shows the location of the surface trace of the section; the middle shows the model; and the bottom subpanel shows ray density. The maps on the top sometimes extend outside the



study area, and this is shown by the model not encompassing the whole rectangle. Horizontal and vertical scales are the same for all cross sections. The color codes and plotting conventions are the same as in fig. 1 and 2.

as due to still active subduction. Moreover, recent results provide independent tomographic images of this high velocity body confined to the northern part of the Apenninic chain (Lucente *et al.*, 1996). It is also noteworthy that the slow shallow (50 km) anomaly in the Tyrrhenian side of the whole Apenninic chain of our model is in agreement with the high attenuation zone detected by Mele *et al.* (1996), using P_n arrivals. Agreement with both teleseismic and P_n studies – respectively, in the deep and the shallow layers of the model – is a confirmation of the increased vertical resolution attained by using regional P first arrivals.

Beneath the Alps, a broad high-velocity anomaly is observed and related to a lithospheric root or slab. A clear difference is seen between the eastern and the western section, the latter having better developed fast roots, suggesting the presence of a south-vergent European lithospheric slab beneath the Po plain (see also Amato *et al.*, 1993; Kissling, 1993).

Some of the slow anomalies in our model correspond to active or recent volcanic areas, such as for the Central Italian Tyrrhenian-coast volcanic province, the French Central Massif, the Sicily Channel, the Moroccan Atlas. We also find a slow anomaly under the Valencia trough, a zone already known as having high attenuation values, low shear velocity and low ground motion amplitudes (Vila *et al.*, 1996). Extensional basins, such as the Tyrrhenian and the Aegean, also show slow anomalies at shallow depths, which correlate with the high surface heat flux (*e.g.*, Čermák, 1979; Hutchinson *et al.*, 1985; Fytikas *et al.*, 1985) and the large attenuation of seismic waves (Hashida *et al.*, 1988) observed. The region beneath Turkey is characterized by low velocities in the shallower layers, in agreement with both shear-wave (Zielhuis and Nolet, 1994) and P-wave (Spakman *et al.*, 1993) velocity studies, which are possibly related to volcanism in that region.

Vertical cross sections cut through the model may help us to understand the spatial distribution of velocity anomalies. Six noticeable cross sections are shown in fig. 4 – where the P velocity model is displayed along with ray density distribution. The first section (panel A) is a cut through the Africa-Eurasia convergence, showing northward and southward subduction marked by the two fast (blue) traces dipping toward the center of the figure. Panel B illustrates the high velocity of the Adriatic plate, dipping toward southwest, beneath the Northern Apennines, apparently continuous. The amplitude of the high velocity anomaly beneath the Sardinia-Corse block is not very well resolved in the shallower layers of the model, as evident by comparison with the synthetic test response. On the contrary, the slow volcanic province under the Italian Tyrrhenian coast is well resolved both in amplitude and shape.

Panels C and D are similar in orientation but slightly shifted in location. The Tyrrhenian slab is cut in a direction approximately parallel to the strike in C, and is too far to the northwest to show up in D, where only a shallow Ionian lithosphere is clearly visible. The Hellenic slab is manifest in both C and D, which slice into it at two different latitudes. Section C also shows the fast signal corresponding to the old Vrancea subduction zone, and the slow anomaly of the Sicily Channel. This low velocity, though clearly affected by smearing between adjacent nodes, is well resolved in sign and amplitude, especially at depth.

The two major active slabs of the Mediterranean are well depicted in panels E and F. Panel E shows the large signal associated with the Hellenic subduction (*e.g.*, Spakman *et al.*, 1988, 1993; Papazachos *et al.*, 1995), the low-velocity, extensional, Aegean basin and the high velocity lid of the Eurasian lithosphere in the Northern Aegean (also detected by Spakman *et al.*, 1993). Panel F shows the sub-vertical Tyrrhenian subduction under slow anomalies marking the extensional basin, the fast Sardinia-Corse block and the slow volcanic Central Massif (see also Granet and Cara, 1988).

5. Conclusions

In this paper we present a new model of three-dimensional variation of P-wave velocity in the upper mantle below the Mediterranean.

The most direct comparison for our model is EUR89B (Spakman *et al.*, 1993), which shares the same general approach. In fact, our first goal for this study is to provide an independent check of the validity of EUR89B, and to ascertain whether additional data and a different procedure could add detail to the picture. Good agreement between these models is in fact an empirical – yet significant – sign of reliability for both. Additional detail shown by our model is in good correspondence with inferences which can be derived from tectonics, and we think it is reliably resolved. We conclude that our study has confirmed the general features exhibited by previous research, and at the same time that higher resolution can be reached nowadays.

A main point of discrepancy with respect to EUR89B is the lack, in our model, of the low velocities found at depth between 170-240 km by Spakman *et al.* (1993) both underneath the Dinarides-Hellenides and the Apennines-Tyrrhenian regions. These slow anomalies have been interpreted as a gap in the subducted lithospheric slab, with major geodynamic implications for the evolution of the Mediterranean. At the present stage of our work, we find no evidence to support this hypothesis of «slab detachment».

Other model features do not directly relate to prior knowledge, and some of them are rather puzzling. One such example is represented by the strong fast, anomaly in the shallower 200 km south of Cyprus. This lineament is shared by EUR89B, and is connected to fast P_n velocity revealed by Middle Eastern seismographic stations recording Aegean earthquakes. Another interesting feature is the slow anomaly that we clearly image in the Strait of Sicily. It can be related to the presence of Neogene extensional structures (*e.g.*, Argnani, 1992), even if its extent is surprisingly continuous to the deepest layers in our model. These are aspects deserving additional attention, which we plan to examine in the future.

This work is by no means to be considered definitive, even within the general approach that we follow (travel time analysis of regional and teleseismic bulletin data). Future developments may for instance address the estimation of the interplay between hypocentral location and structure; an assessment of the effect of the one-dimensional reference model; the inclusion of intermediate-depth and deep events; the correction for crustal thickness. We may also add more data, as made possible by the 30-year data set now released by the International Seismological Centre. The enlargement of the study area to consider also data recorded by North-European seismographs, will expand the model so as to detect mantle features located in the continental European area, and, at the same time, it will also improve the data coverage of the area currently studied by including more regional-distance rays (epicentral distance smaller than 25°) entirely contained within this area. An extensive analysis of the reliability of the tomographic model also constitutes important additional information. A complete analysis should go beyond the standard synthetic tests, such as the one shown here – aimed at addressing inversion resolution, rather than model error. Every critical step and assumption in the whole procedure should be individually tested. However, the comparison among results of different studies is one of the most significant assessments of the validity of a model. This is why we think that it is also important to tackle the problem with many different approaches, using different data sets and techniques.

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