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Crustal velocity structure of the Apennines (Italy) from *P*-wave travel time tomography

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Abstract
In this paper we provide *P*-wave velocity images of the crust underneath the Apennines (Italy), focusing on the lower crustal structure and the Moho topography. We inverted *P*-wave arrival times of earthquakes which occurred from 1986 to 1993 within the Apenninic area. To overcome inversion instabilities due to noisy data (we used bulletin data) we decided to resolve a minimum number of velocity parameters, inverting for only two layers in the crust and one in the uppermost mantle underneath the Moho. A partial inversion of only 55% of the overall dataset yields velocity images similar to those obtained with the whole data set, indicating that the depicted tomograms are stable and fairly insensitive to the number of data used. We find a low-velocity anomaly in the lower crust extending underneath the whole Apenninic belt. This feature is segmented by a relative high-velocity zone in correspondence with the Ortona-Roccamonfina line, that separates the northern from the southern Apenninic arcs. The Moho has a variable depth in the study area, and is deeper (more than 37 km) in the Adriatic side of the Northern Apennines with respect to the Tyrrhenian side, where it is found in the depth interval 22-34 km.

Key words *P-wave tomography – deep structure – Apennines*

1. Introduction

The aim of this paper is to improve the knowledge on the deep crustal structure of the Apennines using *P*-wave velocity tomography. Previous information on the crustal structure of Apennines comes from gravity studies (Giese and Morelli, 1975), deep seismic profiling (Cassinis *et al.*, 1979; Mueller, 1982; Nicolich, 1989; Roeder, 1990) and surface waves dispersion (Calcagnile *et al.*, 1979). These studies reveal the presence of a variable geometry of the Moho (Geiss, 1987). Beneath the Tyrrhenian Sea the crust is as thin as 12 km. Moho depth increases eastward, towards the Adriatic foreland, where it is modelled by gravimetric study at about 38 km (Breda *et al.*, 1994). Suggestions for a velocity reversal in the lower crust derive from seismic profiling (Wigger, 1984).

Previous tomographic studies using teleseisms (Amato *et al.*, 1993) and including regional events (Spakman *et al.*, 1993) are characterized by a low resolution within the crust, imaged with a unique layer. Recent inversions of *P*-wave velocity, and P_n velocities and attenuation reveal the existence of strong low velocity and high attenuation zones beneath the Apenninic belt, whereas the Adriatic region exhibits high velocity and low attenuation

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(Alessandrini *et al.*, 1995; Mele *al.*, 1995, 1996).

In this study, we ride seismic tomography to invert local *P*-wave arrival times at seismic stations located within the Apennines. Since we used bulletin data, our attention has been focused on optimizing the information achievable by a data set that is affected by a high level of noise, resolving only for three layers (centered at 8, 22, and 37 km depth) with a horizontal grid-spacing of 0.5 degree in latitude and longitude. In fact, when inverting noisy data, an overparameterization of the model may result in an undesirable increase in instability and non-linearity of the inverse problem. We tried to mitigate these effects inverting for a not overly complex and fine model, at the expense of loosing structural details. Thus, our results are a first approximation of the real structure – the gross structure really resolvable by the actual dataset. Future directions of this work will be a comparison with results obtained using fewer but precisely repicked data.

7227 *P*-wave arrival times from 600 selected earthquakes have been inverted using the technique of Zhao *et al.* (1992) where the velocity model is parameterized with both a three-dimensional grid of nodes (the velocity is continuously defined within the grid) and velocity discontinuities. A three-dimensional robust ray-tracer traces rays travelling at regional distance. We believe that this technique works reasonably well in such regional studies (Zhao *et al.*, 1992) and helps to retrieve information on the Moho topography.

2. Geologic outline

The Apennines are an orogenic belt built by the convergence between the Eurasian and African plates during the Cenozoic (Scandone, 1979; Reutter, 1981; Malinverno and Ryan, 1986; Royden *et al.*, 1987). This thrust and fold belt consists of several rootless crustal units thrusted towards the Adriatic foreland (Bally *et al.*, 1986; Mostardini and Merlini, 1986; Patacca and Scandone 1987). Compressional tectonics developed since the middle Cretaceous and Oligocene with continental collision and has been followed by a rift process of the Tyrrhenian area which started in the Tortonian, and, contemporarely, by a second compressive phase responsible for the present-day structure of the Apennines (Patacca and Scandone, 1987; Patacca *et al.*, 1990). The compressional tectonics migrated in time from west to east, thrusting progressively the whole edifice toward the Adriatic foreland.

Both gravimetric anomalies and structural data (see Patacca and Scandone, 1987 and references therein) reveal the presence of two distinct arcs, convex toward the foreland, and separated by the Ortona-Roccamonfina (O-R) line (Locardi, 1982; Patacca and Scandone, 1987). These two arcs, namely the northern Apenninic arc and the Southern Apenninic arc (see fig. 1), experienced different amounts of shortening and rotation since late Tortonian (Patacca *et al.*, 1990). The O-R line is thought to be a major lithospheric discontinuity with a dextral lateral motion of the northern arc, driven by a larger lithospheric sinking of this area (see Patacca and Scandone, 1987). A sinking of the foreland lithosphere underneath the Apennines has been proposed to explain the present deformation of the belt (Malinverno and Ryan, 1986; Patacca *et al.*, 1990). Recently, earthquakes down to 80-100 km depth and a high-velocity body in the upper mantle have been recognized underneath the Northern Apennines (Selvaggi and Amato, 1992; Amato *et al.*, 1993; Spakman *et al.*, 1993; Amato *et al.*, 1994), suggesting that the subduction processes may still be active.

Thus, the present-day structure of the Apennines is composed of three distinct belts (Malinverno and Ryan, 1986; Patacca *et al.*, 1990). The Tyrrhenian area characterized by Plio-Quaternary extension and rifting related to the opening of the back arc Tyrrhenian basin (Wezel, 1985; Sartori, 1990). This extensional tectonics stretched the lithosphere, thinned the crust and is responsible for intrusions and volcanism all along the Tyrrhenian sea from Tuscany to Campania (Civetta *et al.*, 1978; Beccaluva *et al.*, 1989). Moving eastward, we find the belt with several thrust units characterized by a main extensional tectonics with normal

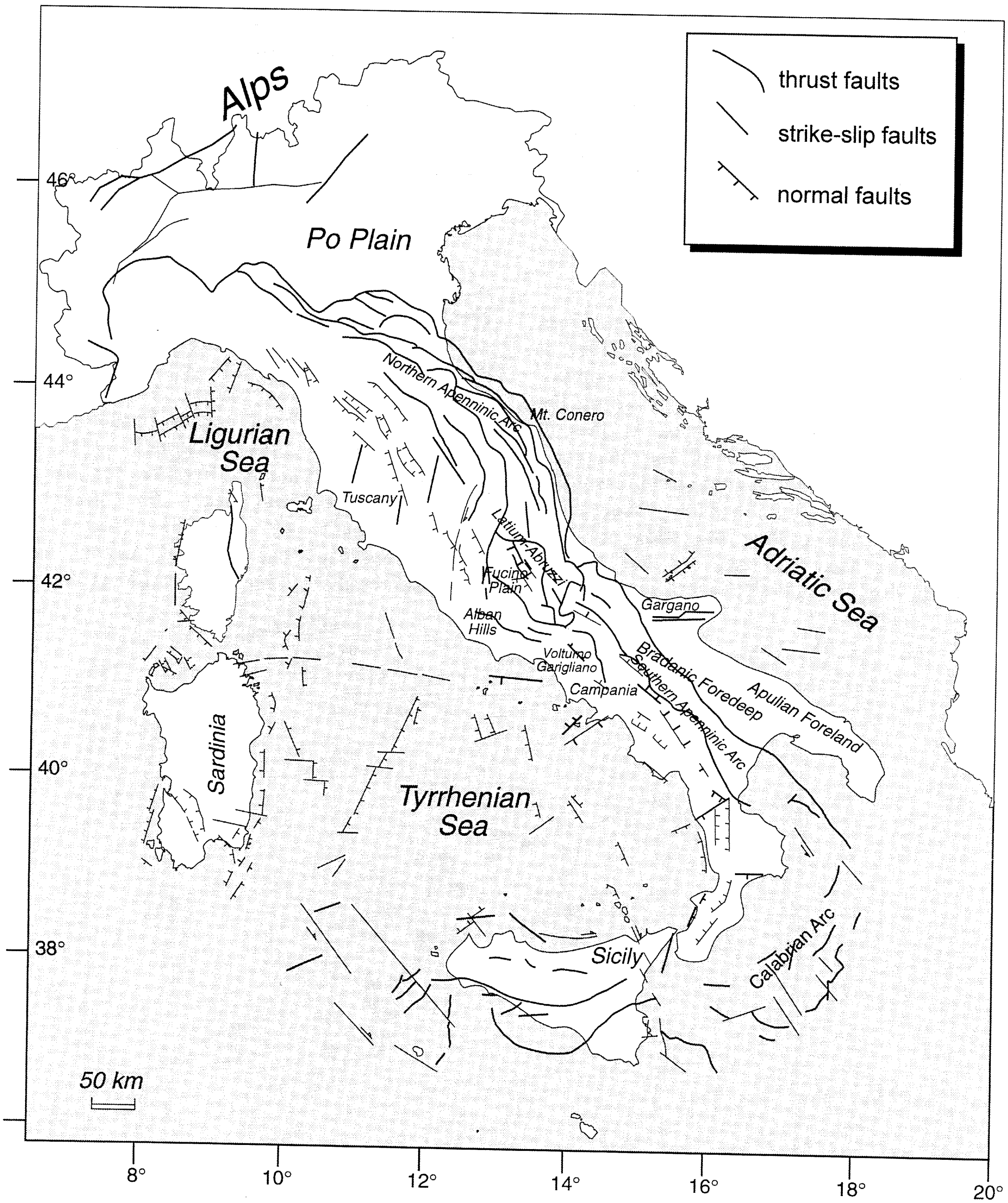


Fig. 1. Structural sketch of the Apennines, major structural elements are represented, redrawn from Patacca and Scandone (1989) with the courtesy of M. Mattei.

faulting earthquakes (Anderson and Jackson, 1987). Finally, compressional tectonics is presumably active along the outer margin of the Northern Apennines (Philip, 1987; Westaway, 1992; Frepoli and Amato, 1996), where the belt-foredeep-foreland system was migrating during the Quaternary.

3. Data and method

In this study, we used *P*-wave arrival times of local and regional Italian events recorded at 51 seismic stations (mostly vertical component) operating from 1986 to 1993 within the Northern and Southern Apennines. At present, the need for a great number of seismic phase arrival times for regional tomographic studies requires the use of bulletin data, whose accuracy suffers from routine operation. We selected earthquakes collected by the Istituto Nazionale di Geofisica (ING) seismic network and by other stations operating in the area (see fig. 2). We first located the overall dataset using Hypoinverse (Klein, 1989) and then selected 600 earthquakes with more than 12 *P*-wave arrival times at nearby stations, and hypocentral errors under 5 km both horizontally and in depth (fig. 3).

Two different kinds of errors affect the data: phase reading errors (random) and phase misidentification (systematic), and inadequacy of model parameterization and ray tracing (the complexity of the model that our procedure cannot resolve). Since reading errors strongly affect seismic tomography results (especially systematic errors), we compared arrival times for some events of our dataset (bulletin data) with the arrival times accurately re-picked (fig. 4). We found small differences for stations close to the epicentral area (~ 60 km), while for longer paths we observed larger differences, especially close to the P_n arrivals. Figure 4 shows that about 80% of the errors are

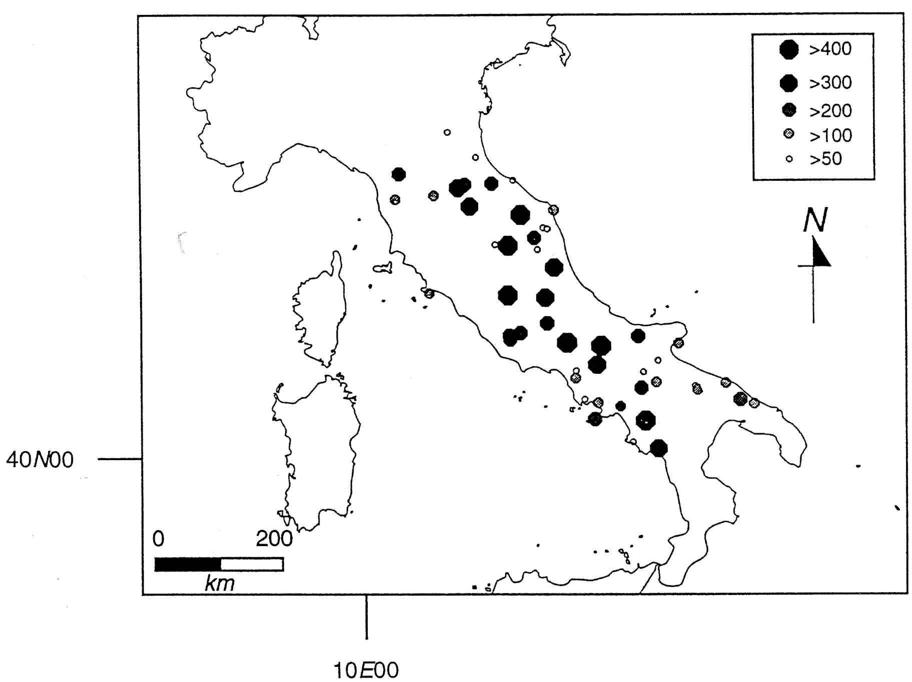


Fig. 2. Map showing the number of *P*-phases used for each station.

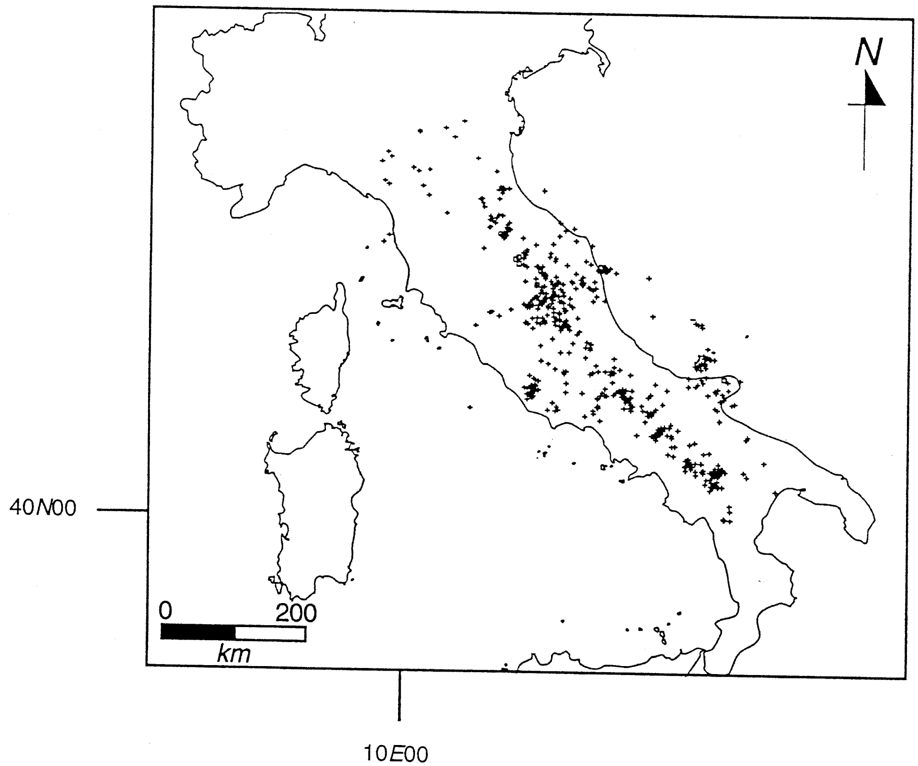


Fig. 3. Map of the 600 earthquakes selected for the three-dimensional inversion.

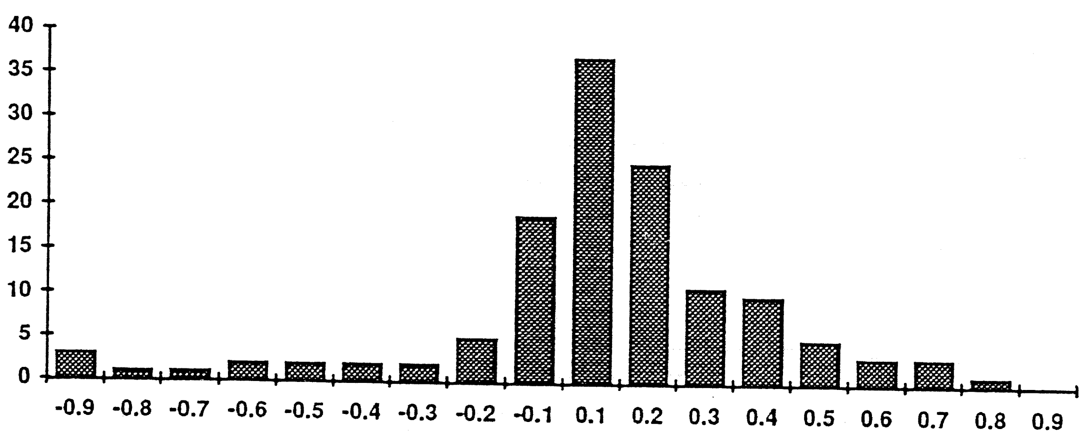


Fig. 4. Histogram of the difference between bulletin and re-picked data. Note the systematic, fictitious delays that affect the bulletin data.

smaller than 0.3 s, but we also observed up to 0.8 s of difference. We note a trend of *P*-wave delays, due to the fact that an emergent P_n arrival can easily be missed during routine bulletin analysis, as it is hidden in the background noise at the station.

The presence of large errors in ordinary bulletin data needs to be taken into account in seismic tomography, identifying the threshold above which it is better to terminate the inversion. In our case, the presence of some errors as high as 0.7-0.8 s (although the majority of errors is less than 0.3 s) is severe, and may justify the high final r.m.s. reached by the inversion (see the next section). This high data misfit strongly limits the definition and resolution of the image obtained, allowing only an approximate velocity model to be resolved. We believe that a finer model parameterization (for instance, a variable Moho geometry), that could reduce the intrinsic errors of the inversion procedure, may be applied in the future but only together with finely re-picked data. Our results reveal how much information about the earth's structure is contained in the bulletin data, a usual source for tomographic studies.

We used 11091 *P*-wave arrivals from the 600 selected events, whose ray paths sample the crustal volume underneath the Apennines. We excluded from the inversion long distance P_n phases (paths greater than 230 km) for which large reading errors are expected, with ray paths too badly influenced by the Moho topography. This selection reduces the inverted data to 7227 phases, mostly consisting of *P*-waves travelling within the Apenninic region. With this selection, we also aimed to exclude seismic rays travelling through complex heterogeneous zones outside the Apennines, as the Po plain, the Alps, the Tyrrhenian sea, removing the influence of objects located around the modelled volume. Figure 2 shows the number of data used for each station.

We performed a simultaneous inversion for hypocenters and velocity parameters first introduced by Thurber (1983) for local tomography (*i.e.*, rays shorter than ~ 50 km). We adopted the technique of Zhao *et al.* (1992), since it is more suited for regional tomographic studies, due to the robust three-dimensional ray tracer that better reproduces long (200-300 km) raypaths. We refer to these previous works for a detailed description of the procedure (see Zhao *et al.*, 1992, 1994, and references therein). The velocity model is parameterized with a three-dimensional grid of nodes (unknown parameters) and sharp velocity discontinuities (fixed). In our case, the velocity at any point is calculated in the horizontal plane using a linear interpolation between the adjacent nodes of each layer, and vertically using a layered model with velocity discontinuities (see fig. 5). The model parameterization (nodes of a three-dimensional grid, plus fixed vertical discontinuities) was chosen limiting the number of inverted parameters, and using three layers (see the one-dimensional model of fig. 5). Nodes are spaced 0.5 degrees in latitude and longitude (about 55 and 40 km, respectively). In all, we used a grid of $14\times16\times3$ nodes. The inversion of model parameters was solved using LSQR (Paige and Saunders, 1982).

Obviously, any linearized inversion is strongly sensitive to the starting model. Thus, we tried different reference one-dimensional models considering the available – but scant – geologic and geophysical information on the deep structure of the Apennines (AA.VV., 1991a,b). We decided to invert for two crustal

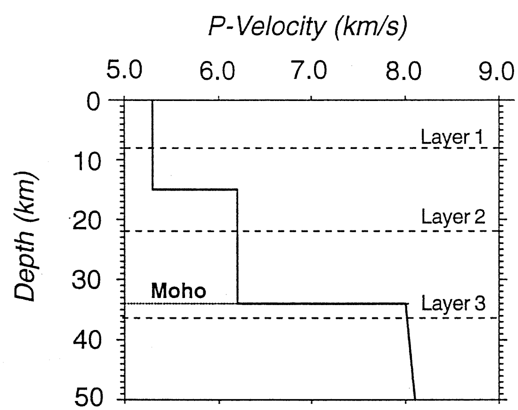


Fig. 5. 1D velocity model used in the inversion. The grid nodes are located at 8, 22 and 37 km depth. The starting Moho is located at 34 km depth.

layers (see fig. 5), one in the upper crust (8 km depth) and one for the lower crust (22 km depth). These two layers are separated by a sharp discontinuity located at 15 km depth. The deepest layer is located at 37 km depth, below the Moho (34 km depth). A Moho with variable depth would be more realistic to represent the Apenninic structure, but the scarce information available on its geometry precludes this representation. Since detailed studies are developing to define the Moho geometry in the Apenninic area («CROP» Projects, see AA.VV., 1991a,b; and teleseismic transects, see Amato *et al.*, 1994), we will include a Moho with variable geometry in a future work.

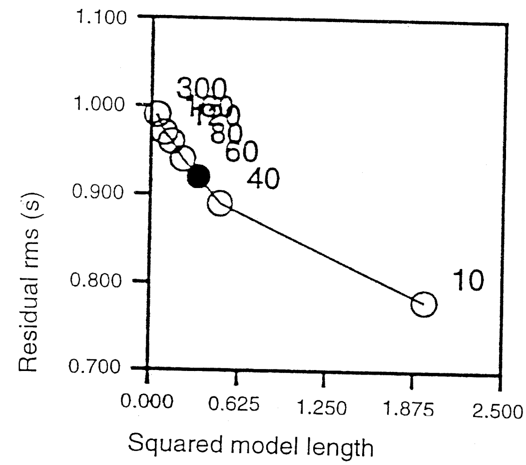


Fig. 6. Residual variance *vs.* squared model length for different damping parameters. The values are relative to the first iteration step. The selected damping is indicated by the black dot. We preferred a slightly more conservative damping parameter than the optimum one.

3.1. *Three dimensional inversions*

We inverted 7227 arrival times from 600 events with a velocity damping parameter of 60 $(km/s)^2$ (see fig. 6), selected optimizing data variance reduction and model length. Earthquake hypocenters, updated after each iteration, do not move more than 5 km from the starting location. A total of 262 velocity parameters were modelled. Final r.m.s. is 0.83 s, with a variance improvement of 53% achieved after 4 iterations. Most of the variance reduction (41%) was achieved after the first iteration. We inverted the data also using a damping parameter of 40 $(km/s)^2$ (the optimum choice in the trade off of fig. 6) achieving a final r.m.s. = 0.83 s, and obtaining a final three-dimensional model similar to that computed with the lower damping. We preferred the model obtained with the higher damping, because, both results remaining the same, it is more conservative.

We also tried a model with one more layer in the crust and the same grid-spacing in latitude and longitude, finding a variance improvement of only 15% after four iterations, smaller than that obtained with three layers, and a final r.m.s. of 0.75 s. We found in this inversion that the r.m.s. increases between the second and the third iteration steps, suggesting that the inversion is not stable, and the finer velocity model is badly constrained by the data. We then rejected these results. The over-parameterization of this model increased the non-linearity of the problem beyond the limit, and we decided that the previous model with three layers was the maximum detail resolvable with the present dataset.

3.2. *Results*

Figure 7 shows the final *P*-wave velocity model, as depicted by the inversion of the 600 earthquakes dataset. The poor illumination of deep structures underneath the Tyrrhenian side of Tuscany is due to the scarce number of stations operating in this area. Local microseismic arrays that operate in the geothermal areas of Tuscany did not send arrival time data to be included in the bulletin. Thus, we warn the readers not to consider our results for that area.

In layer 1, 8 km in depth, the pattern of high and low velocities reflects the strong lateral heterogeneities of the sedimentary cover. Since horizontal resolution is limited by the parameterization to about 50 km, this image is an

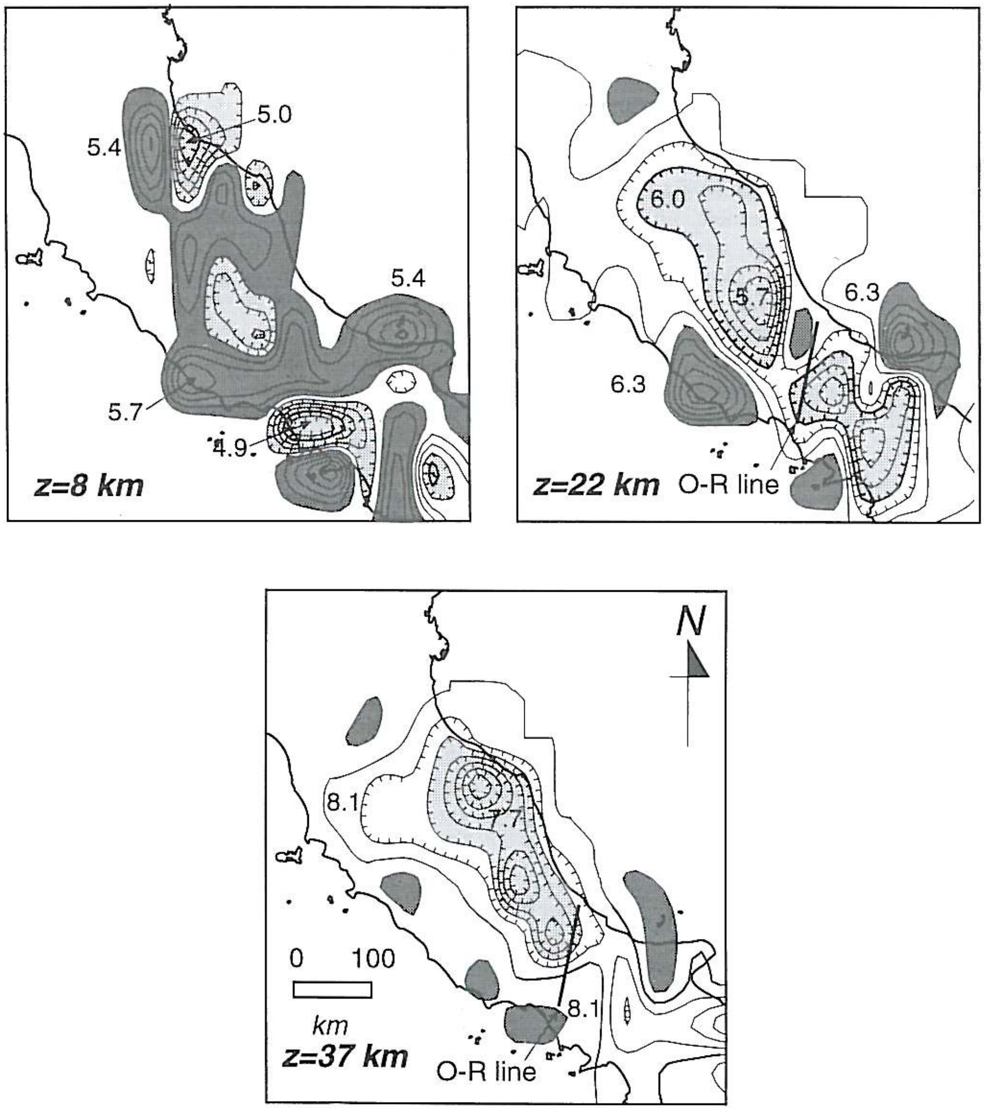


Fig. 7. Final *P*-wave velocity model in the three crustal lays. Darker shaded areas represent high velocity zones, whereas lightly shaded areas indicate low velocity anomalies. For reference, the O-R line is reported on the deep layers.

aliased picture of the lithologic heterogeneities related to the different shallow geologic units. We noted a low velocity area from Mt. Conero towards the Po plain that corresponds to a region in the foredeep where Plio-Quaternary sediments are thicker (up to 7 km thick, Patacca *et al.*, 1990). The Apenninic belt was characterized by relatively high velocity anomalies. A central low velocity zone was present approximately underneath the Latium-Abruzzi limestone platform, ringed by high velocities, suggesting the presence of a different crust type. Two prominent low velocity anomalies were found beneath the southern part of the Bradanic foredeep, and the Volturno/Garigliano plain. These anomalies may be related to thick Plio-Quaternary sediments, the former filling the foredeep (Patacca *et al.*, 1990), whereas the latter lies in a strongly subsided area of the Tyrrhenian coastal region.

In layer 2, 22 km depth, a continuous low velocity anomaly was present in the lower crust underneath the northern and the central Apenninic belt. This anomaly was separated by a second low velocity anomaly centered beneath the Southern Apenninic belt, in correspondence with the O-R line. Positive velocity anomalies were present beneath the Tyrrhenian side of the Apennines from the Alban Hills (Latium) to the south (but we have no information north of the Alban Hills for the Tyrrhenian coast).

Although a systematic pattern of fictitious delays was recognized (see fig. 3), their small amplitude (0.2-0.3 s) cannot justify the large velocity decrease revealed in the lower crust.

In the lower crust, low-velocity zones were observed underneath the entire Apenninic belt, while high velocities were found beneath the Tyrrhenian side. This feature is consistent with the recognized thinned crust of the Tyrrhenian area that becomes thicker under the Apennines. We lack evidence of low-velocity bodies underneath the western areas where Quaternary volcanism occurred, suggesting the absence of large crustal magma chambers (but our spatial resolution is limited to about 50 km). Magma chambers smaller than (45×50) km^2 may remain undetected by this study. Northern and Southern Apennines seem to be separated by a major N-S trending discontinuity – the O-R line (see Patacca *et al.*, 1990 for details about the role played by this lithospheric discontinuity).

A less obvious consideration regards the velocity values obtained for the *P*-waves underneath the belt. We found values as low as 5.6 km/s in the lower crust, extremely low for *P*-waves. *P*-wave velocities of mid-crustal rocks are normally in the range of 6.1 to 6.4 km/s (Christensen, 1982), some 10% higher than what we observed. V_p values for lower crust rocks are usually larger than 6.2 km/s (Kern, 1982). One plausible explanation for the observed low velocities (~ 5.6 km/s) is the existence of high temperatures in the lower crust that decrease the *P*-wave velocity. In this case, more than 700°C are needed to obtain the observed values for a granitic rock type (see Kern, 1982). Therefore, the lower crust underneath the Apenninic belt may be close to the melting temperature, opening a tantalizing debate about the state of the lower crust beneath this region. Thus, an intriguing issue is that we did not image low velocities beneath the areas where Quaternary magmatic activity developed, but under the orogenic belt. Evidence of low velocities beneath the Apenninic belt has recently been found by Alessandrini *et al.* (1995) and Mele *et al.* (1995). This latter study, focused on regional P_n waves, identified a strong attenuation and a pervasive low-velocity anomaly underneath the Central Apennines. Di Bona and Selvaggi (personal communication), studying teleseismic receiver functions along a transect crossing the Apennines (Amato *et al.*, 1994), disclosed the presence of a diffuse scattering in the lower crust beneath the central part of the belt, with not efficient *P*-to-*S* conversions at the Moho.

In layer 3, 37 km depth, the main anomaly was a low velocity area underneath the Adriatic side of the Northern Apennines. We found velocity values of about 7.6-7.7 km/s that indicate a greater depth for the Moho in this region. Thus, the data force at depths greater than 37 km the *a priori* constant Moho depth imposed in the starting model at 34 km depth. This observation is in agreement with recent works on the Moho geometry, among which gravity anomalies (Breda *et al.*, 1994), receiver function analysis (Amato *et al.*, 1994) and P_n tomography (Mele *et al.*, 1995). The Moho was deeper than 37 km in the Adriatic area and beneath the belt of the Northern Apennines towards the O-R line. In the Tyrrhenian area and in the Southern Apennines we found velocity values in agreement with the presence of upper mantle rocks. Thus, the depth of the Moho here lies between layer 2 (22 km depth) and the *a priori* constant Moho (34 km depth).

4. Stability and resolution analysis

We performed a partial inversion, using the same one-dimensional model as the previous inversion but decreasing the number of events. We inverted 4171 phases from 350 events, randomly selected, with a damping of 60 $(km/s)^2$, reaching a final r.m.s. of 0.85 s, with a variance improvement of 30% obtained after 4 iterations. These values are similar to those of the previous inversion (fig. 8). We observed that the velocity images were smoother than but similar to those obtained with the whole dataset. Thus, we believe that our results are

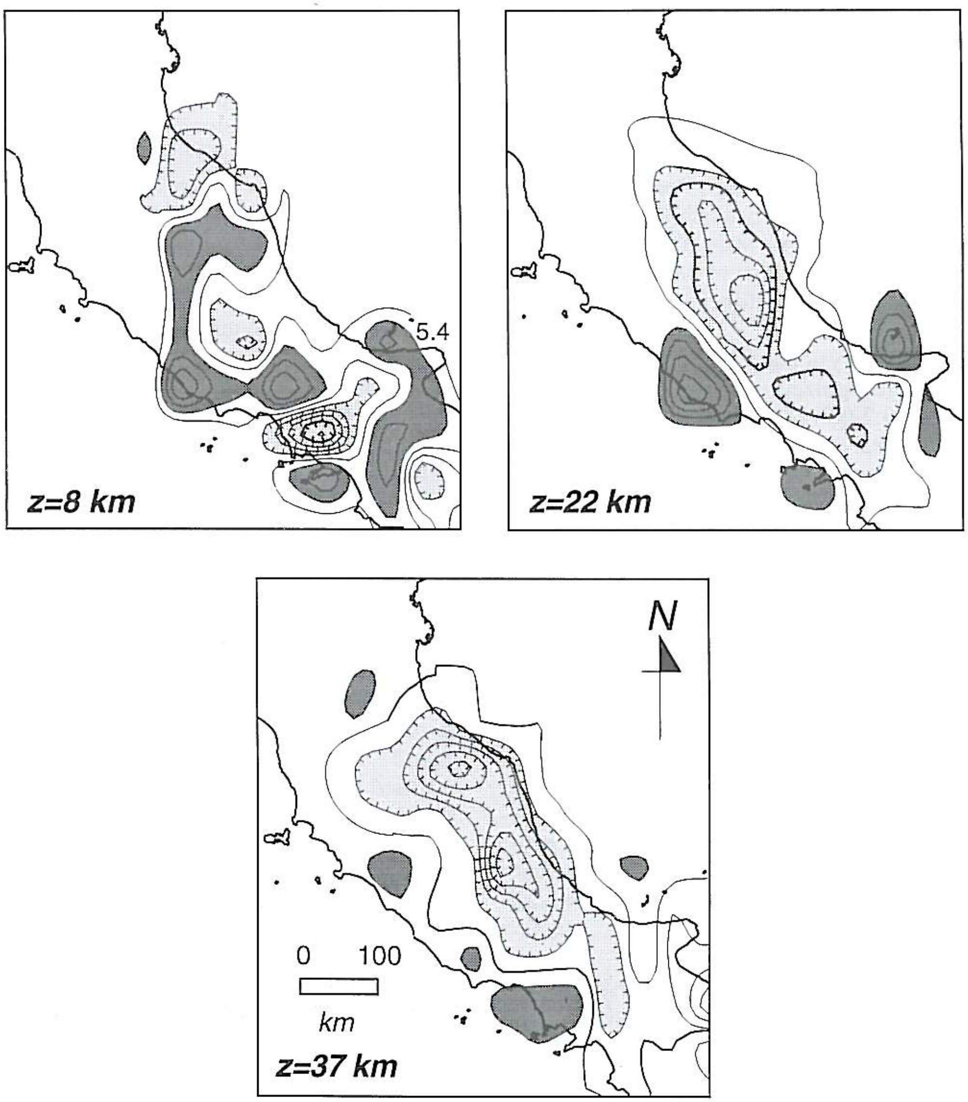


Fig. 8. *P*-wave velocity model computed by the partial inversion. The same representation as fig. 7 is used.

scarcely dependent on the number of events used (in our case, a full dataset of 600 earthquakes does not contain much more information than if only 350 were used). This demonstrates that the results are mostly determined by the station and earthquake distribution and reading errors rather than by the amount of data. Since we fitted the data above a high noise level, further improvement of the images can only come from accurately re-picked arrival times rather than from a larger, still poor dataset. In any case, this test indicates that our results are stable.

Verifying the reliability of the computed images is the «longer suffering» of any tomographic study. Resolution depends on several factors and is mainly limited by station spacing and earthquake distribution that may result in a scarce criss-crossing of raypaths. Different analyses have been proposed in the literature to visualize in which way the velocity is averaged throughout the medium. For instance, the analysis of the whole resolution matrix, as defined by Menke (1989) and Thurber (1993), is very effective for damped least squares inversions. Chiarabba *et al.* (1995) compared the whole resolution matrix with results of synthetic tests, finding that the two approaches similarly define the well resolved volumes of the model. In our study, the data were inverted with an LSQR alghorithm and we decided to use synthetic tests rather than calculate the resolution matrix. We performed a «checkerboard» test, where high and low velocities alternate in the three space directions. Although the meaning of such test has been criticized by Leveque *et al.* (1993) considering that checkerboard tests are somehow not realistic, we still believe that this test can be meaningful, at least at a first approximation (see Chiarabba *et al.*, 1995).

Figure 9a,b shows the results of the test computed with lateral heterogeneities of $\pm 2\%$. Earthquake hypocenters were fixed, and a random error equal to 0.05 s was added to the synthetic travel time dataset. The computation of synthetic travel times through a checkerboard model with lateral heterogeneities similar to those obtained in the inversion (*i.e.*, $\pm 8\%$) failed mostly for raypaths travelling in the periphery of the model. Thus, we were forced to use a small amplitude for the checkerboard, and accordingly reduce the variance of the random noise added to the data.

We found that the central volume beneath the Apennines was sufficiently well-resolved in the three layers. Resolution was poor beneath the Tyrrhenian side of Tuscany, and in the peripheral Adriatic area. Here, we found velocity anomalies larger than the starting synthetic ones, indicating that the velocity computed for these nodes were ill-constrained by the data (poor station coverage). Moreover, we performed a second test in which the synthetic travel times were computed in a laterally homogeneous model and subsequently inverted. We found no changes in the central well-resolved volume, whereas the peripheral Adriatic nodes were strongly perturbed. Thus, we concluded that the noise in the inversion yields undesirable velocity perturbations in the poorly resolved peripheral nodes.

5. Relocated seismicity

Figure 10a,b shows the earthquakes relocated using the heterogeneous model. We observed that hypocentral locations did not change dramatically (less that few km), with a general improvement for the hypocentral r.m.s. residual of about 30% with respect to the locations computed using the lateral homogeneous model.

The earthquakes were mainly confined underneath the Apenninic belt, and in two small areas. One is the Alban Hills Quaternary volcano, and the second is the Gargano-Tremiti area, within the Apulian foreland. The main seismic release occurred within the belt in correspondence with the highest elevations characterized by Quaternary uplift (see Westaway, 1992).

The depth distribution of events (fig. 10a,b) revealed that seismicity concentrated between the surface and 12 km depth, both in the Northern and Southern Apennines. Thus, the seismogenic zone was confined above the low velocity of the lower crust (see fig. 7). How-

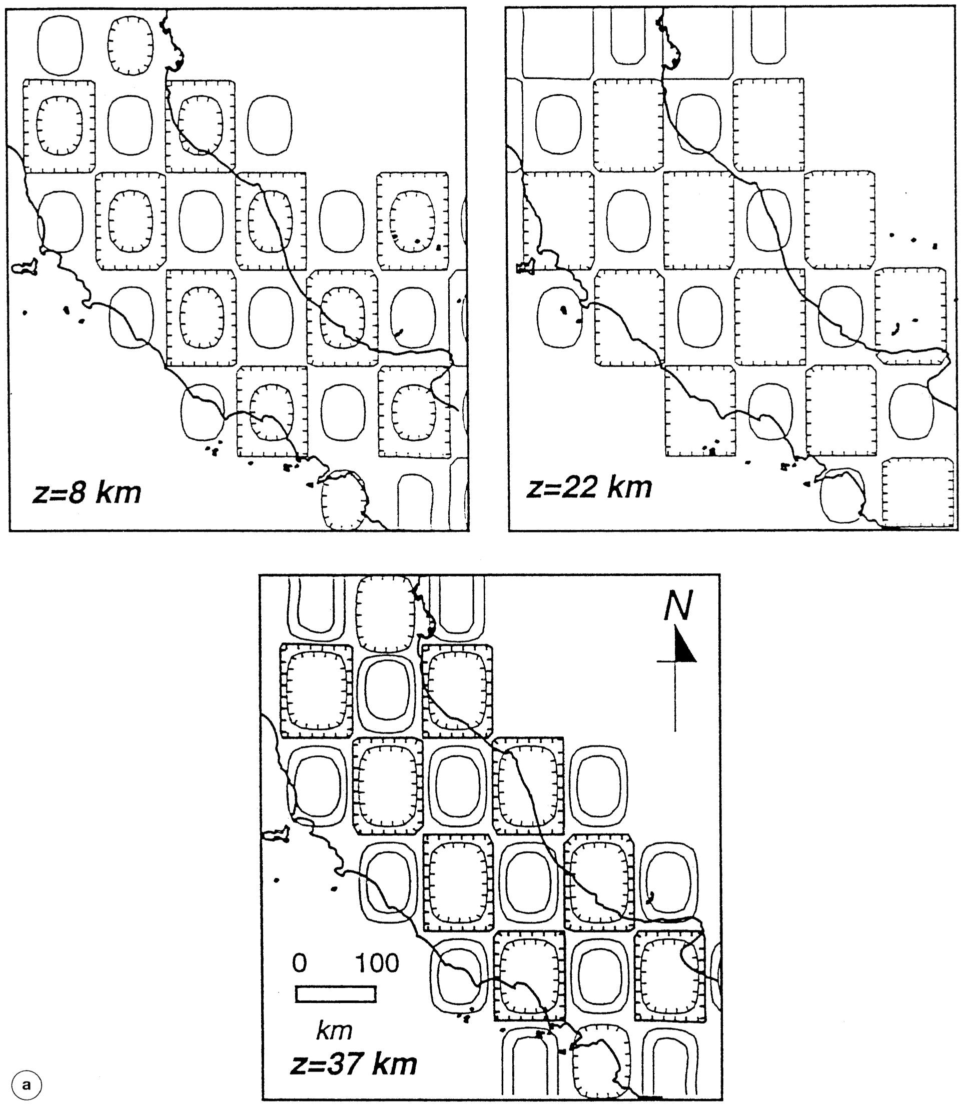
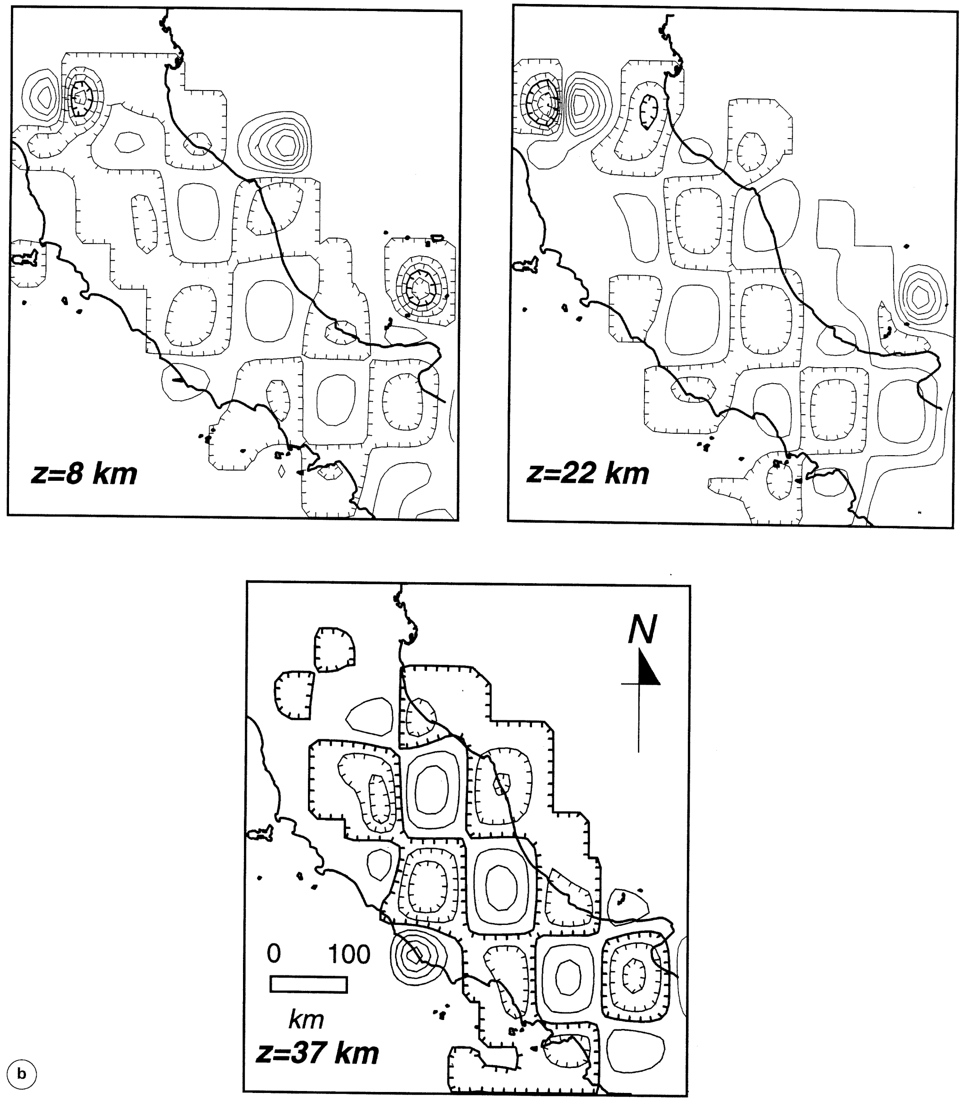


Fig. 9a,b. Comparison between starting (a) and computed (b) models of the checkerboard test. Note the good

ever, in the Northern Apennines we also observed a sparse seismicity in the deepest crust, between 30 and 40 km depth, in agreement with results of Selvaggi and Amato (1992) that describe earthquakes underneath the Northern Apennines down to 90 km depth. The seismicity of the Apulian foreland was scattered in a larger interval of depths. Seismicity occurred there within a high velocity zone, in an area that recorded a strong Quaternary uplift, and was separated by the rest of the Apulian foreland.

Earthquakes are concentrated in the upper crust mainly within the high velocity bodies of



reproduction of the synthetic anomalies underneath the Apenninic belt. Amplitude scale of anomalies is ±2%.

fig. 7a,b. A paucity of earthquakes is recognizable close to the Fucino Plain imaged as a low velocity zone, probably characterized by a different crust. We did not note a continuous arcuate belt of seismicity in the Northern Apennines, as suggested by Cocco *et al.* (1993), but rather a broad NW-SE trending seismic region composed of different seismic areas. Major lithospheric discontinuities (like the O-R line, or other hypothesized lines cutting the Apenninic belt) were not affected by seismicity, probably suggesting that presently differential motion is negligible there, or occurs aseismically.

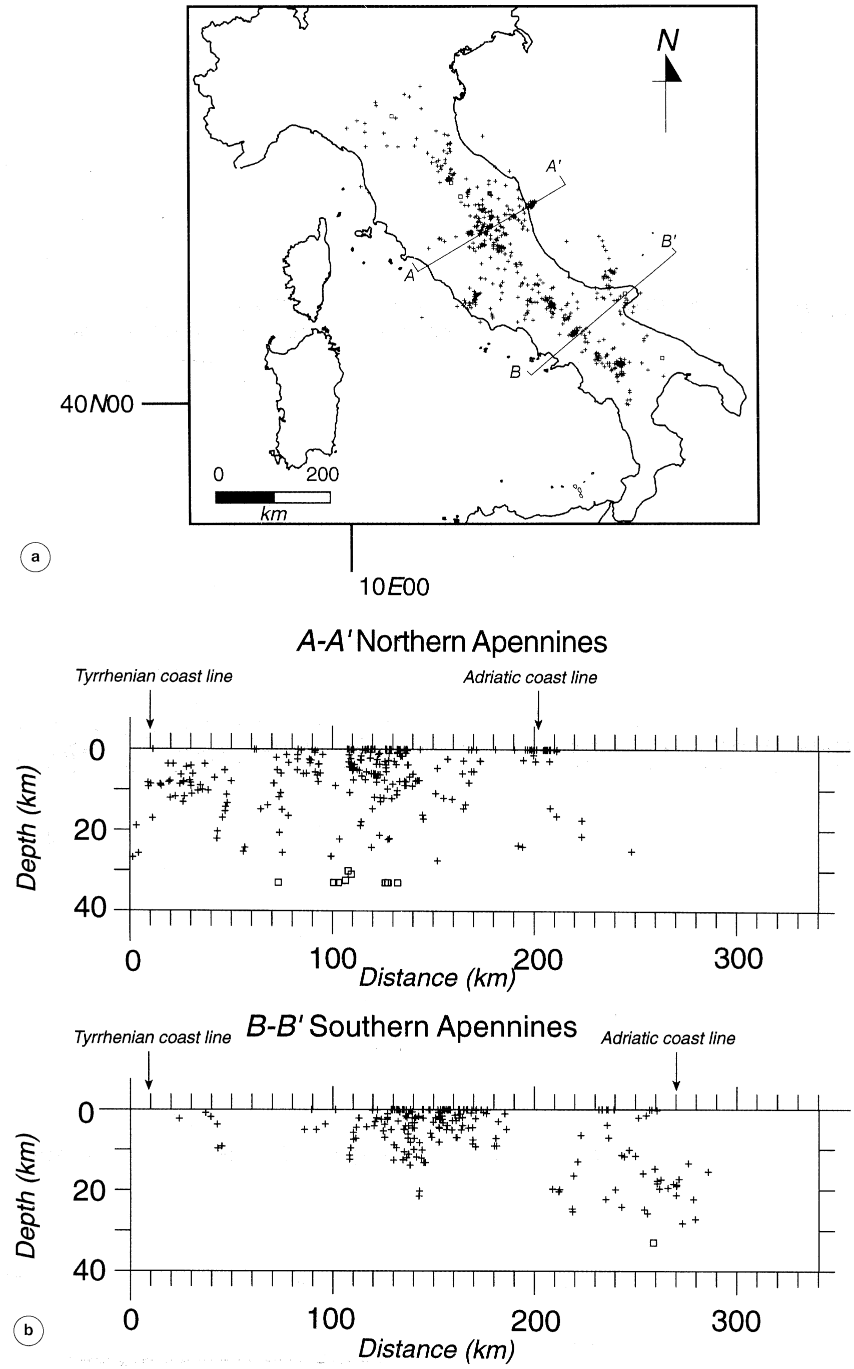


Fig. 10a,b. a) Map; b) SW-NE vertical sections (across Northern and Southern Apennines) of the relocated events.

6. Conclusions

In this paper, we present velocity images of the Apenninic crust obtained inverting bulletin data. We discussed the limits of this inversion due to the noisy dataset we used. We mitigated this problem by overdamping the solution and resolving for a simple structure – modeling few parameters only. With this choice, we lost details in our images, resolving only the gross crustal structure underneath the Apennines. However, we believe that to derive a more detailed crustal model requires a higher quality dataset. The main velocity heterogeneities recovered in this work were a low velocity beneath the Apenninic belt, and a variable depth of the Moho. The Moho was found at 22-34 km depth beneath the Tyrrhenian margin of the Apennines, and deepened down to 37 km toward NE. The unexpected low velocity values found in the lower crust may suggest remarkable thermal effects active beneath the chain, and open an intriguing debate on the state of the deep crust. Finally, velocity heterogeneities were in agreement with the model in which the Apennines are divided into two distinct arcs by the O-R line.

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