Mapping Moho depth variations in central Italy from Ps_{Moho}-P delay times: evidence of an E-W transition in the Adriatic Moho at 42° N latitude

Giuliana Mele[1,*], Emiliano Di Luzio[2], Cristina Di Salvo[3]

[1] Istituto Nazionale di Geofisica e Vulcanologia, Via di Vigna Murata, 605, 00143 Roma, Italy; tel: +39 06 51860416; email: giuliana.mele@ingv.it

[2] Consiglio Nazionale delle Ricerche, Area della Ricerca Roma 1 - Via Salaria km. 29,300 - 00016 Monterotondo stazione (Roma), Italy; tel. +39 06 90672722; email: emiliano.diluzio@itabc.cnr.it

[3] Istituto di Geologia Ambientale e Geoingegneria, CNR, Area della Ricerca Roma 1 - Via Salaria km. 29,300 - 00016 Monterotondo stazione (Roma), Italy; tel. +39 06 90672740; email: cristina.disalvo@igag.cnr.it

* Corresponding author: G. Mele, Istituto Nazionale di Geofisica e Vulcanologia, Via di Vigna Murata, 605, 00143 Roma, Italy; tel: +39 06 51860416 (email: giuliana.mele@ingv.it)

Abstract

Along the Italian peninsula adjoin two crustal domains, peri-Tyrrhenian and Adriatic, whose boundary is not univocal in central Italy. In this area, we attempt to map the extent of the Moho in the two terrains from variations of the travel time difference between the direct P wave and the P-to-S wave converted at the crust-mantle boundary. We use teleseismic receiver functions computed at 43 broad-band stations in this and previous studies, and assign each of the recording sites to the Adriatic or peri-Tyrrhenian terrains based on station location, geologic and geophysical data and interpretation, and consistency of delays with the regional Moho trend. The results of the present study show that the Ps_{Moho} arrival time varies from 2.3 s to 4.1 s in the peri-Tyrrhenian domain and from 3.7 to 5.5 s in the Adriatic domain. As expected, the lowest time difference is observed along the Tyrrhenian coastline and the largest values are observed in the axial zone of the Apennine chain. A key new result of this study is a sharp E-W boundary in the Adriatic domain that separates a deeper Moho north of about 42° N latitude from a shallower Moho to the south. This feature is constrained for a length of about 40 km by the observations available in this study. The E-W boundary requires a revision of prior mapping of the Moho in central Italy and supports previous hypotheses of lithosphere segmentation.

1. Introduction

Peninsular Italy extends in the Mediterranean Sea from 38° to 46° latitude North and 8° to 18° longitude East. Its geologic setting is dominated by the Apennine chain that extends along the whole peninsula. This chain built up mostly during the Neogene and early Pleistocene following the deformation of the African continental margin of the Tethyan ocean [e.g., *Malinverno and Ryan*, 1986; *Albarello et al.*, 1995; *Vezzani et al.*, 2010].

In peninsular Italy, the topography of the Moho discontinuity, that is the object of this study, has been investigated through active seismic profiles collected during the DSS experiments in the 1960's-1990's [*Cassinis et al.,* 2003, and references therein] and the CROP Project in the 1980's-1990's [*Scrocca et al.*, 2003, and references therein], and passive seismology methods such as tomography and teleseismic receiver functions [e.g., *Piana Agostinetti et al.*, 2002; *Mele and Sandvol*, 2003; *Mele et al.*, 2006; *Di Luzio et al.*, 2009; *Di Stefano et al.*, 2009; *Piana Agostinetti and Amato*, 2009]. Active and passive seismic data have been combined in *Di Stefano et al.* [2011].

The Moho map proposed by *Cassinis et al.* [2003] had the merit, unlike the majority of the maps derived from other studies, of distinguishing the crustal domains that characterize Italy and surrounding areas: continental crust in the European and African/Adriatic domains; oceanic/suboceanic crust in the Ligurian and Tyrrhenian Seas; transitional crust in the peri-Tyrrhenian side of peninsular Italy and northern Sicily. The boundary between the Adriatic and peri-Tyrrhenian crusts runs along peninsular Italy and northern Sicily (Figure 1). Recently, *Di Stefano et al.* [2009, 2011] have proposed two boundaries that differ from each other and from that of *Cassinis et al.* [2003] in central Italy, as shown in Figure 1. In this area, where the three boundaries deviate one from the other and one of them is partially unconstrained, we attempt to reconstruct the extent of the Adriatic and peri-Tyrrhenian crust.

To map the Adriatic and peri-Tyrrhenian Moho, we use the teleseismic receiver functions method that is based on the identification of the P wave converted to S at the Moho discontinuity (called Ps_{Moho} in the following). The delay time of the Ps_{Moho} with respect to the direct P arrival is affected primarily by Moho depth: the larger/smaller the delay, the deeper/shallower the Moho beneath the recording site; therefore, we interpret variations in the Ps_{Moho} time in terms of variations of Moho depth. We integrate the new data with previous receiver functions computed by *Mele et al.* [2006] and *Di Luzio et al.* [2009].

The 43 recording stations used in central Italy are assigned to one or the other crustal domain based on location, geologic and geophysical data, and consistency with the regional trend of the Moho. The Ps_{Moho}-P times are interpolated with the Ordinary Kriging statistical method to map the extent and the lateral variations of the Adriatic and peri-Tyrrhenian Moho.

2. Geologic setting

In peninsular Italy, the peri-Tyrrhenian area is characterized by a stretched transitional crust with positive Bouguer anomalies [e.g., *Morelli,* 1981], high heat flow [e.g., *Della Vedova et al.*, 2001] and relatively low uppermost mantle velocities [e.g., *Mele et al.,* 1998]. On the contrary, the Adriatic domain is a more stable area with low heat flow, low-to-moderate positive Bouguer anomalies and normal-to-high uppermost mantle velocities. Since *Mele and Sandvol* [2003], the Adriatic Moho was inferred to deepen to about 50 km beneath the Apennine chain.

Central Italy is characterized by Meso-Cenozoic platform and basin units of the Apennine chain verging NE-ward above the Bradanic foredeep and the Adriatic/Apulian foreland (Figure 2). To the west, Plio-Quaternary marine-to-continental deposits and Pleistocene volcanics cover large sectors of the internal Apennines that were downthrown by extensional faults since the late Miocene [e.g., *Patacca et al.*, 1990].

In the study area, the foreland sequence outcrops in the Gargano promontory and Tremiti Islands (Figure 2), mainly characterized by the carbonate units of the Apulian Platform (AP). Part of the Apulian Platform was involved in the Apennine deformation during the Pliocene-Early Pleistocene; it is exposed in the Maiella Massif and surroundings (Apennine external units of Figure 2) [*Bally et al.*, 1986; *Mostardini and Merlini*, 1986; *Cipollari and Cosentino*, 1995; *Patacca et al.*, 2008; *Cosentino et al.*, 2010].

The Apulian Platform Top (APT) is a regional key-horizon distinctive of the Adriatic crust; it was used to follow the westward dipping of the foreland monocline beneath the foredeep and the Apennines [*Mariotti and Doglioni*, 2000]. This horizon, made of Miocene limestones and/or evaporites, is reached at depths ranging from about 1 km in the peri-Adriatic region to about 3 km in the axial zone of the Apennines by the exploration wells plotted in Figure 2. In the CROP11 profile, a high-amplitude pair of reflectors interpreted as the APT horizon is followed from the Adriatic coast to the Fucino basin [*Scrocca et al.*, 2003; *Patacca et al.*, 2008]; west of the Fucino basin the CROP11 profile is not interpreted. In this work, the APT will be used to constrain the extent of the Adriatic crust.

3. Method of analysis, seismologic data, and observations

Since the first observations in the 1950's [*Cook et al.*, 1962, and references therein], teleseismic P waves converted to S at major velocity discontinuities of the Earth were used to infer the gross seismic structure under a recording station. The Ps_{Moho} is often the highest-energy signal in the coda of the direct P arrival due to the large velocity contrast between the crust and the mantle, and is used to build regional Moho maps [e.g., *Priestley et al.*, 1988; *Kind et al.*, 1995; *Jones and Phinney*, 1998; *Al-Damegh et al.*, 2005; *Lloyd et al.*, 2010]. Data usable for these studies are three-component, possibly broad-band recordings of teleseismic events with epicentral distance of 30° to 90°.

The Ps_{Moho} phase arrives few seconds after the direct P and most of the times it is hard to observe in the seismogram. The method used to identify the Ps_{Moho} consists in deconvolving the vertical component of the ground motion from the horizontal component rotated into the radial direction (source-to-receiver path) where Ps conversions have the largest amplitude [*Langston*, 1979]. Deconvolution filters out most of the common features such as source, travel path effects, and instrumental response, producing a simpler time series called receiver function. This last is composed by the first positive P pulse followed by Ps conversions and reverberations. Deconvolution also enables to compare receiver functions from various seismic sources that are stacked together to enhance the coherent signals.

The time delay between Ps_{Moho} and P (t_{Ps} hereinafter) can be used to estimate the depth of the Moho (H) for given bulk crustal velocities Vp and Vs and P-wave incidence angle (expressed through the ray parameter p):

$$H = \frac{t_{Ps}}{\sqrt{(1/Vs^2 - p^2)} - \sqrt{(1/Vp^2 - p^2)}} \qquad (1)$$

In this work, we have collected teleseisms with minimum magnitude Mw 5.5 recorded in the 2004-2009 period by 29 permanent stations of the Italian Seismic Network. The epicentral distance is computed from the center of the study area. Given the abundance of seismic sources in the distance range 80°±10°, we selected these events because steeper incidence angles yield larger energy of the incoming P wave.

After a selection of the recordings in terms of the signal-to-noise ratio, we cut a window of 30 s from the seismograms of 148 events (Figure 3a), starting 5 s before the P onset. To compute receiver functions, we applied the time-domain deconvolution technique of *Ligorria and Ammon* [1999]; a Gaussian low-pass filter with width parameter α=2.0 was used to remove the high-frequency noise.

In the receiver functions of 24 stations, a positive peak arriving 2.3 to 5.2 s after P was interpreted as the Ps wave converted at the Moho discontinuity; 5 stations were discarded due to noisy or inconsistent observations.

We also used the t_{Ps} computed by *Mele et al.* [2006] at the permanent station AQU and 12 temporary stations installed for few months in 1995 (0-4C, 6-9C, 11C, 12C, 14C), and by *Di Luzio et al.* [2009] at the permanent station FRES (Figure 3b). For most of these stations, only events from the north-east and 80°±10° distance were available [see Figure 5 of *Mele et al.*, 2006].

In the present study, most of the observations are naturally clustered between 330° and 100° backazimuth (Figure 3a) and this prevented to analyze the crustal response as a continuous function of azimuth. For this reason, and for consistency with previous works, we stacked the receiver functions of events occurred in the NE quadrant. This ensures also to sample the same Moho structure beneath each station.

Depending on the working state and quality of the recording site, the number of receiver functions varies from 4 (4C, GUAR, CIGN) to 56 (INTR). In Figure 3b are shown the stacks of 5 stations arranged along a SW-NE profile that crosses the boundaries between the peri-Tyrrhenian and Adriatic crusts.

4. Mapping the peri-Tyrrhenian and Adriatic Mohos

In order to estimate Moho depth from the Ps_{Moho} delay, a bulk crustal velocity must be provided for all stations. However, previous works propose conflicting models in the study area, especially at mid-crustal depth. As an example, we show in Figure 4 two seismic tomography sections where high-velocity anomalies are imaged on both sides of the Fucino basin [*Chiarabba et al.*, 2010] and two crustal sections interpreted from active seismic data where low velocity is inferred in the same area [*Cassinis et al.*, 2003; *Patacca et al.,* 2008].

Because of the uncertainty in the regional velocity structure, in the present study we use the Ps_{Moho} delays as indicative of Moho depth variations. The delay of the Moho conversion is read from the stack trace of each station and mapped in Figure 5a. Ps_{Moho} delays span from 2.3 to 5.5 s, and the conversion points at the Moho occur NE of the stations, at an average distance of 10 km. In this map, we attributed each station (i.e. observation points of t_{Ps}) to the Adriatic or peri-Tyrrhenian terrain based on location with respect to the proposed boundaries; where the boundaries deviate from each other, the attribution is based on surface and shallow geology (well logs) or on the consistency of t_{Ps} with the Moho trend defined by the

Tyrrhenian stations 0-4C and the Adriatic stations 6-14C and FRES [*Mele et al.,* 2006; *Di Luzio et al.*, 2009].

Stations located west of the three boundaries are assigned to the peri-Tyrrhenian crust (from north to south: MAON, LATE, CESX, MNS, TOLF, MTCE, ROM9, RDP, CERT, GUAR, GIUL), while stations located east of the boundaries are assigned to the Adriatic crust (TERO, CAMP, CAFR, LPEL, CIGN, SGRT, MSAG). Other stations can be attributed to the Adriatic crust based on the following aspects: i) MIDA and CERA, located close to two explorations wells that reached the APT horizon at about 3 km of depth, and to outcrops of the deformed Apulian domain (see area framed in Figure 2); ii) INTR, located along the segment of the CROP11 profile where the reflection package interpreted as the Apulian Platform Top is recognized beneath the Apennine units [*Patacca et al.*, 2008]; iii) CAMP and FAGN, where the relatively large t_{Ps} (5.0 and 5.2 s) is consistent with the westward deepening Adriatic Moho.

The attribution of stations matches the boundaries of *Cassinis et al.* [2003] and *Di Stefano et al.* [2009], while it is inconsistent with the boundary proposed by *Di Stefano et al.* [2011] (Figure 5a). Stations FIAM, VVLD, and POFI are uncertain because their location is not constrained by geologic evidence and the t_{Ps} matches the trend of the Moho in both crustal domains.

Figure 5b displays a contouring of t_{Ps} obtained with ArcGIS® Geostatistical Wizard [*ESRI*, 2009]. We used the Ordinary Kriging prediction method [*Matheron*, 1970] to model the spatial trend of a single variable; to avoid a-priori bias, data were interpolated without using barrier polylines between the Mohos. The basic assumption, when using statistics to handle heterogeneity in Earth systems, is that properties are not random, but have some spatial continuity or are correlated over some distance. The Geostatistical Analyst Extension module of ArcGIS® examines the distribution of the data to create a semivariogram model that allows to compute the parameter value in unsampled locations. The Kriging model generates the predicted surface after selecting the best suitable model based on regression statistics. Observed vs simulated t_{Ps} resulting from the cross-validation procedure are plotted in the inset of Figure 5b. In the map of Figure 5b, smaller differential times occur in the western sector of the peninsula, characterized by brown colors (t_{Ps} between 3.3 and 3.8 s), matching the attribution of most of the 16 peri-Tyrrhenian stations. As to the Adriatic stations, the contouring highlights two regions with different t_{Ps} that define a sharp transition of the Moho surface along the 42° N latitude: t_{Ps} changes from 4.6-4.7 s to the north to 3.7-3.8 s to the south. The receiver function stacks of the 5 stations straddling the Moho transition are shown in Figure 5b.

5. Discussion

In central Italy, we have distinguished stations located in the peri-Tyrrhenian and in the Adriatic terrains to reconstruct the variations of the Moho in these crustal domains.

A key finding of this study is a sharp variation of t_{Ps} in the Adriatic domain, at about 42° N latitude: from north to south, t_{Ps} changes from 4.6-4.7 s at stations 9C, 11C and 12C to 3.7 - 3.8 s at stations INTR and LPEL, within a distance of 15 km (Figures 5a,b). At stations 9C, 11C, and 12C, *Di Luzio et al.* [2009] have estimated a Moho depth of 38±1 km using a local bulk crustal Vp of 6.3 km/s derived from the interpretation of the CROP 11 profile. This is a good crustal average commonly used in literature. Adopting such Vp value in equation (1), we estimate a Moho depth of 30 and 31 km beneath stations LPEL and INTR, respectively, i.e. the Adriatic Moho is ~ 8 km shallower south of the 42° N parallel. The E-W Moho transition can be constrained for about 40 km with the observations available for this study (Figure 5b). It is worth to underline that the Ps_{Moho} delays of the Adriatic stations are consistent on either side of the Moho transition: 4.6 to 5.5 s are observed at all stations located north of 9C-12C while 3.8 to 4.2 s are observed at all stations located south of INTR and LPEL (Figures 5a,b).

The E-W step of the Adriatic Moho supports previous ideas of lithosphere segmentation in central Italy [*Royden et al.,* 1987; *Doglioni et al.*, 1994]. *Royden et al.* [1987] based their model on the morphology of the Apennine foredeep basin (correlated with Bouguer gravity anomalies) and of the outermost thrust of the chain; both show differential offsets from north to south reflecting a different amount of lithosphere retreat (Figure 5c). *Doglioni et al.* [1994] hypothesized that a differential lithosphere rollback occurs between the central Adriatic and the Puglia region, caused by the difference in the lithospheric thickness inherited from the Mesozoic rifting: the downgoing of the 40-km thicker Puglia lithosphere slowed down since the middle Pleistocene favouring the uplift of the foreland and the Moho in the Gargano promontory (Figure 5d). The present study results confirm that the Moho is shallower over the whole sector below 42° N latitude, not only beneath the Gargano promontory, and is rather flat: stations MSAG and SGRT show the same t_{Ps} of the nearby stations, including the one located in the Tremiti islands where *Mele et al.* [2006] estimated a Moho depth of 33 km.

The step of the Moho in central Italy is not displayed in the Moho map of *Piana-Agostinetti and Amato* [2009], obtained with the receiver functions stacking technique of *Zhu and Kanamori* [2000]. The reason could be that this map is a smoothed image of Moho depth variations with less than 1/3 high-quality stations (class 1-2 defined by the authors). Additionally, the temporary stations 0-14C and the permanent station LPEL, i.e., 4 of the 5 stations that constrained the E-W Moho step, are not used by these authors; this produces a low-resolution image of the Adriatic Moho around the 42° N parallel. It is worth noting that INTR, that is the only station shared by the two studies around the Moho step, has the same average Moho depth (Table 1).

The Moho depths estimated by *Piana-Agostinetti and Amato* [2009] are used by *Di Stefano et al.* [2011] to integrate active seismic data and reconstruct the Moho topography in Italy. In central Italy, the Tyrrhenian/Adriatic boundary of *Di Stefano et al.* [2011] is in contrast with the Ps_{Moho} delays: several stations located west of this boundary have t_{Ps} of 5.0 s and more (Figure 5a), corresponding to Moho depths larger than 40 km, that cannot be associated with the peri-Tyrrhenian Moho.

6. Conclusions

We have presented a revised mapping of the peri-Tyrrhenian and Adriatic Moho in central Italy supplementing previous receiver function studies (14 stations) with results obtained from 24 additional stations. We have compared the cumulative receiver function results with constraints from well data and active source imaging to assign each station to either crustal domain. The new result of the present study is evidence for a sharp E-W transition in the Adriatic Moho that rises of ~ 8 km south of ~42° N parallel. This feature can be constrained for a length of ~ 40 km with the data available in this study. The E-W transition requires a major revision to prior mapping of crustal domains and supports previously hypothesized lithosphere segmentation.

References

Albarello, D., E. Mantovani, D. Babbucci, and C. Tamburelli (1995), Africa–Eurasia kinematics: main constraints and uncertainties, Tectonophysics, 243, 25–36.

Al-Damegh, K., E. Sandvol, and M. Barazangi (2005), Crustal structure of the Arabian plate: New constraints from the analysis of teleseismic receiver functions, Earth Planet. Sci. Lett., 231, 177–196.

Bally, A. W., L. Burbi, C. Cooper, and R. Ghelardoni (1986), Balanced sections and seismic reflection profiles across the central Apennines, Mem. Soc. Geol. Ital., 35, 257–310.

Cassinis, R., S. Scarascia, and A. Lozej (2003), The deep crustal structure of Italy and surroundings areas from seismic refraction data. A new synthesis, Boll. Soc. Geol. It., 122, 365–376.

Chiarabba, C., S. Bagh, I. Bianchi, P. De Gori, and M. Barchi (2010), Deep structural heterogeneities and the tectonic evolution of the Abruzzi region (Central Apennines, Italy) revealed by microseismicity, seismic tomography, and teleseismic receiver functions, Earth Planet. Sci. Lett., 295, 462–476.

Cipollari, P., and D. Cosentino (1995), Miocene unconformities in the central Apennines: Geodynamic significance and sedimentary basin evolution, Tectonophysics, 252, 375–389.

Cook, K. L., S. T. Algermissen, and J. L. Costain (1962), The status of Ps converted waves in crustal studies, J. Geophys. Res., 67, 4769–4778.

Cosentino, D., P. Cipollari, P. Marsili, and D. Scrocca (2010), Geology of the central Apennines: a regional review. In The Geology of Italy: tectonics and life along plate margins, edited by M. Beltrando, A. Peccerillo, M. Mattei, S. Conticelli, and C. Doglioni, Journal of the Virtual Explorer, Electronic Edition, ISSN 1441-8142, vol. 6, paper 12, doi:10.3809/jvirtex.2010.00223.

Della Vedova, B., S. Bellani, G. Pellis, and P. Squarci (2001), Deep temperatures and surface heat flow distribution. In: G.B. Vai (eds), Anatomy of an Orogen: The Apennines and adjacent Mediterranean basins, Kluwer academic publishers, 65-76.

Di Luzio, E., G. Mele, M. M. Tiberti, G. P. Cavinato, and M. Parotto (2009), Moho deepening and shallow upper crustal delamination beneath the central Apennines, Earth Planet. Sci. Lett., 280(3–4), 1–12.

Di Stefano, R., E. Kissling, C. Chiarabba, A. Amato, and D. Giardini (2009), Shallow subduction beneath Italy: Three-dimensional images of the Adriatic European-Tyrrhenian lithosphere system based on high-quality P wave arrival times, J. Geophys. Res., 114, B05305, doi:10.1029/2008JB005641.

Di Stefano, R., I. Bianchi, M. G. Ciaccio, G. Carrara, and E. Kissling (2011), Three–dimensional Moho topography in Italy: New constraints from receiver functions and controlled source seismology, Geochem. Geophys. Geosyst., 12, Q09006, doi:10.1029/2011GC003649.

Doglioni, C., F. Mongelli, and P. Pieri (1994), The Puglia uplift (SE Italy): an anomaly in the foreland of the Appenine subduction due to buckling of a thick continental lithosphere, Tectonics, 13, 1309–1321.

ESRI (Environmental Systems Research Institute) (2009), ArcMap 9.3.1 Redlands, CA, USA.

Jones, C. H., and R. A. Phinney (1998), Seismic structure of the lithosphere from teleseismic converted arrivals observed at small arrays in the southern Sierra Nevada and vicinity, California, J. Geophys. Res., 103, 10065–10090.

Kind, R., G. L. Kosarev, and N. V. Petersen (1995), Receiver functions at the stations of the German Regional Seismic Network (GRSN), Geophys. J. Int., 121, 191–202, doi:10.1111/j.1365-246X.1995.tb03520.x.

Langston, C. A. (1979), Structure under Mount Rainier, Washington, inferred from teleseismic body waves, J. Geophys. Res., 84 (B9), 4749–4762.

Ligorria, J. P., and C. J. Ammon (1999), Iterative deconvolution and receiver function estimation, Bull. Seismol. Soc. Am., 89, 1395–1400.

Lloyd, S., S. van der Lee, G. S. França, M. Assumpção, and M. Feng (2010), Moho map of South America from receiver functions and surface waves, J. Geophys. Res., 115, B11315, doi:10.1029/2009JB006829.

Malinverno, A., and W. B. F. Ryan (1986), Extension in the Tyrrhenian Sea and shortening in the Apennines as a result of arc migration driven by sinking of the lithosphere, Tectonics, 5, 227–245.

Matheron, G (1970), The Theory of Regionalized Variables and its Applications, Les Cahiers du Centre de Morphologie mathématique, Fascicule V, Ecole des Mines de Paris, 211 pp.

Mariotti, G., and C. Doglioni (2000), The dip of the foreland monocline in the Alps and Apennines, Earth Planet. Sci. Lett., 181(1-2), 191-202, doi: http://dx.doi.org/10.1016/S0012-821X(00)00192-8.

Mele G., A. Rovelli, D. Seber, T.M. Hearn, and M. Barazangi (1998), Compressional velocity structure and anisotropy in the uppermost mantle beneath Italy and surrounding regions, J. Geophys. Res., 103(B6), 12,529-12,543.

Mele, G., and E. Sandvol (2003), Deep crustal roots beneath the northern Apennines inferred from teleseismic receiver functions, Earth Planet. Sci. Lett., 211, 69–78.

Mele, G., E. Sandvol, and G. P. Cavinato (2006), Evidence of crustal thickening beneath the central Apennines (Italy) from teleseismic receiver functions, Earth Planet. Sci. Lett., 249(3-4), 425–435.

Morelli, C. (1981), Gravity anomalies and crustal structures connected with the Mediterranean margins. In: Sedimentary Basins of Mediterranean Margins (F.C. Wezel, ed.), pp. 33–53. C.N.R. Italian Project of Oceanography, Tecnoprint, Bologna.

Mostardini, F. and S. Merlini (1986), Appennino centro meridionale. Sezioni geologiche e proposta di modello strutturale, Mem. Soc. Geol. It., 35, 177–202.

Patacca, E., R. Sartori, and P. Scandone (1990), Tyrrhenian basin and Apenninic arc: Kinematic relations since Late Tortonian times, Mem. Soc. Geol. Ital., 45, 425–451.

Patacca, E., P. Scandone, E. Di Luzio, G. P. Cavinato, and M. Parotto (2008), Structural architecture of the central Apennines: interpretation of the CROP 11 seismic profile from the Adriatic coast to the orographic divide, Tectonics, 27, TC3006, doi:10.1029/2005TC001917.

Piana Agostinetti, N., F. P. Lucente, G. Selvaggi, and M. Di Bona (2002), Crustal structure and Moho geometry beneath the Northern Apennines (Italy), Geophys. Res. Lett., 29(20), doi:10.1029/2002GL015109.

Piana Agostinetti, N., and A. Amato (2009), Moho depth and Vp/Vs ratio in peninsular Italy from teleseismic receiver functions, J. Geophys. Res., 114, B06303, doi:10.1029/2008JB005899.

Priestley, K., G. Zandt, and G. Randall (1988), Crustal structure in eastern Kazakh, U.S.S.R. from teleseismic receiver functions, Geophys. Res. Lett., 15, 613–616.

Royden L., E. Patacca, and P. Scandone (1987), Segmentation and configuration of subducted lithosphere in Italy: an important control on thrust-belt and foredeep-basin evolution, Geology, 15, 714-717.

Scrocca, D., C. Doglioni, F. Innocenti, P. Manetti, A. Mazzotti, L. Bertelli, L. Burbi, and S. D'Offizi (Eds.) (2003), CROP Atlas: seismic reflection profiles of the Italian crust, Mem. Descr. Carta Geol. Ital., 62, pp. 194.

Vezzani, L., A. Festa, and F. C. Ghisetti (2010), Geology and tectonic evolution of the Central-Southern Apennines, Italy, Geol. Soc. of America Special Paper, 469, 1–58.

ViDEPI Project, Visibility of petroleum exploration data in Italy, http://unmig.sviluppoeconomico.gov.it/videpi/en/pozzi/pozzi.asp.

Zhu, L., and H. Kanamori (2000), Moho depth variation in southern California from teleseismic receiver functions, J. Geophys. Res., 105 (B2), 2969–2980.

Figure Captions

Figure 1. Moho isobaths and crustal domains of Italy and adjacent areas after *Cassinis et al.* [2003]. The boundaries between Tyrrhenian and Adriatic plates at Moho depth proposed by *Di Stefano et al.* [2009] and [2011] are superimposed for comparison. White segments along the boundary of *Di Stefano et al.* [2009] are poorly constrained.

Figure 2. Geologic sketch of central Italy (Fb=Fucino basin; Mm: Maiella massif). The exploration wells that drilled the Apulian Platform Top (APT) and the trace of the CROP11 deep reflection profile are shown. The red square indicates the most internal part of the Apennine chain where the APT, distinctive of the Adriatic crust, is drilled [ViDEPI Project: http://unmig.sviluppoeconomico.gov.it/videpi/].

Figure 3. a) Azimuthal projection of the events used in the present study, centered in the study area. b) Topography map of central Italy showing the seismic stations used in this (29) and previous (14) studies; 5 stations were discarded because no clear identification of the Moho conversion could be made. The boundaries between Adriatic and peri-Tyrrhenian Moho proposed by *Cassinis et al.* [2003] (CA03) and by *Di Stefano et al.* [2009, 2011] (DS09, DS11) are also shown. Receiver function stacks of 5 stations projected along the profile A-A' and the position of the Adriatic/peri-Tyrrhenian boundaries are shown in the upper panel. In the receiver functions, arrows mark the P onset (time=0) and the Ps_{Moho} phase; n indicates the number of events used in the stack.

Figura 4. Upper panel: traces of active and passive seismic profiles in the study area. Lower panel: (left) Vp models obtained by *Chiarabba et al.* [2010] combining local earthquakes tomography and teleseismic receiver functions and (right) interpreted crustal sections along the "Latina-Pescara" DSS profile [after *Cassinis et al.,* 2003] and the CROP 11 profile [simplified after *Patacca et al.*, 2008]. Profiles 2-2 and CROP11 are parallel, such as the profiles 6-6 and DSS.

Figure 5. a) Delay times of Ps waves converted from the Moho discontinuity beneath 43 stations. The recording sites are tentatively assigned to the Adriatic or peri-Tyrrhenian crust. Ps_{Moho} delays range from 2.3 s along the Tyrrhenian coastline to 5.5 s in the Apennine region. b) Contouring of Ps_{Moho} delays interpolated with the Ordinary Kriging prediction method; the range of delays is divided into contour intervals assigned to different colors. From the seismic sources used in this study the Moho conversion occurs at 10±5 km from the station, depending on crustal thickness. The red segment indicates the offset of the Moho and the minimum extent that can be constrained with the data presented in this study; the receiver function stacks of the 5 closest stations are also shown. In the inset are plotted the observed vs simulated t_{Ps} resulting from the cross-validation of the predictive model (root mean square is 0.325 s, mean error is 0.024 s, average standard error is 0.409 s). c) Sketch of lithosphere segmentation after *Royden et al.* [1987] and d) *Doglioni et al.* [1994].

Table 1. Seismic stations used in this study listed in alphabetical order with Ps_{Moho} time delays (t_{Ps}), assigned crustal domain (AD: Adriatic; TR: peri-Tyrrhenian), and Moho depths computed in this study ($^{+}$), *Mele et al.* [2006] (^{x}), and *Di Luzio et al.* [2009] (^{xx}). In the last column are listed for comparison the Moho depths of *Piana-Agostinetti and Amato* [2009] (PA-A 2009); in parenthesis is the quality class of each station defined by these authors, decreasing from 1 to 5. The Adriatic stations located within 50 km from INTR are highlighted in boldface.