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From underplating to delamination-retreat in the northern Apennines

C. Chiarabba [a,*], G. Giacomuzzi [a], I. Bianchi [b], N.P. Agostinetti [c], J. Park [d]

[a] Istituto Nazionale di Geofisica e Vulcanologia, Roma, Italy
[b] Universität Wien, Institut für Meteorologie und Geophysik, Vienna, Austria
[c] Geophysics Section, School of Cosmic Physics, Dublin Institute for Advanced Studies, Dublin, Ireland
[d] Yale University, Department of Geology and Geophysics, New Haven, CT, USA

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ABSTRACT

Recordings of teleseismic earthquakes from a dense set of temporary and permanent broadband seismic stations reveal the lithospheric structure of the northern Apennines and support the scenario of a retreating detachment within the mid-crust. Lithospheric delamination appears crucial to the formation and evolution of the Apennines orogen. Receiver-function (RF) stacks outline a continuous west-dipping Ps converted phase from a positive velocity jump that we interpret as the top of the lower crust and mantle of the Adria continental lithosphere, which is descending into the shallow mantle. The correlation of seismicity with two RF profiles across the northern Apennines suggests distinct stages of lithospheric delamination. Active penetration of the detachment into the Adria lithosphere seems evident in the south/east, with induced shallow-mantle flow facilitated by slab dehydration. Penetration of the detachment in the north/west seems to have arrested, and is possibly marked by crustal underplating. This layer atop the Apennines slab is visible only down to 80 km depth and suspends above an oppositely-dipping paired positive/negative Ps converted phase in stacked receiver functions. The break in the west-dipping Adria lithosphere conflicts with a westward-subduction scenario continuous from the Oligocene. Lateral changes of deep structure and seismicity along the northern Apennines suggest that underplating of crustal material and delamination-retreat are distinct mechanisms active today in the western and eastern sectors, respectively, of the northern Apennines. Negative Ps-pulses at 100–120 km depth help to define a seismic lithosphere–asthenosphere boundary (LAB), but cross-cut a volume of high-velocity mantle rock, as inferred from tomographic models. We hypothesize that this seismic LAB is a rheological discontinuity that affects the frequency band of seismic body waves, but not the long-term viscous response that governs the evolution and eventual detachment of the continental slab.

1. Introduction

The central Mediterranean region has been shaped by subduction of Mesozoic oceans associated with the Alpine and Apennines orogenic events, as part of the collision between the European and African plates (Dewey et al., 1989). In most plate reconstructions, several continental terranes, intercalated by fully-developed, but poorly-defined, oceans, are introduced to explain ophiolite fragments and continental units accreted within the mountain belts (Stampfli and Borel, 2002; Handy et al., 2010; Carminati et al., 2012).

The northern Apennines is part of this puzzle (Fig. 1). Its evolution is usually described by the eastward migration of paired compression- and extension-tectonics related to retreating-slab models (Elter et al., 1975; Bally et al., 1986; Jolivet et al., 1998; Royden, 1988; Malinverno and Ryan, 1986; Doglioni, 1991; Faccenna et al., 2001). Although this simplification is attractive, recent experiments and studies argue that a simple process of subduction by a retreating slab cannot be applied to the northern Apennines belt, either spatially or temporally. Apatite fission tracks outline an orogen-parallel segmentation of wedge kinematics and a non-synchronicity of tectonic processes (Thomson et al., 2010). The asthenospheric flow identified by SKS birefringence is not consistent with a typical retreating slab (Plomerova et al., 2006; Salimbeni et al., 2007). Slab tears and slab windows have been recognized by seismic tomography suggesting a disruption of the subduction process (Lucente et al., 1999; Wortel and Spakman, 2000; Piromallo and Morelli, 2003; Chiarabba et al., 2008; Benoit et al., 2011; Giacomuzzi et al., 2012).

The distribution of seismicity, along with GPS velocities (Serpelloni et al., 2007; D'Agostino et al., 2009), helps to describe the active processes of the Apennines (Fig. 1). We observe a

* Corresponding author. Tel.: +39 0651860408; fax: +39 0651860507.
E-mail address: claudio.chiarabba@ingv.it (C. Chiarabba).

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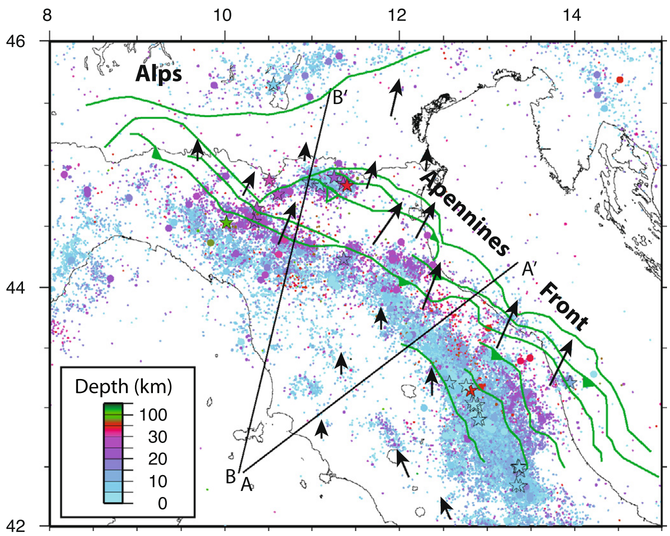

Fig. 1. Hypocentral map of the northern Apennines. Seismicity of the past 20 years is plotted with color code representing hypocentral depth. Stars show locations of events with $M < 5.0$. Shallow seismicity (light blue dots) occurs in the region currently undergoing extension along the Apennines belt, while compression is restricted to the Apennines front. Arrows indicate GPS velocities with respect to fixed Eurasia for some reference points taken from Serpelloni et al. (2007) and D'Agostino et al. (2009). The trace of the vertical section reported in Fig. 2 is plotted.

strip of shallow earthquakes that accommodate about 3–5 mm/yr of extension along the mountain range, while compressional earthquakes occur along the eastern Adriatic front, accommodating a few mm/yr of compression (Pondrelli et al., 2006; Bennett et al., 2012). Sub-crustal earthquakes are located on a west-dipping plane, down to a depth of 70–80 km (e.g. De Luca et al., 2009; Carannante et al., 2013).

The availability of seismological broadband data collected during the past decade yields a strong improvement in the definition of the bulk lithospheric structure of the Apennines, mostly achieved by seismic tomography (Benoit et al., 2011; Piccinini et al., 2010; Giacomuzzi et al., 2011, 2012), receiver-function analyses (Bianchi et al., 2010; Agostinetti et al., 2011; Miller and Agostinetti, 2012) and the modeling of Moho geometry (Agostinetti and Amato, 2009; Di Stefano et al., 2011; Spada et al., 2013).

In this study, we present high-resolution Ps receiver-function (RF) profiles on 2D swaths that cross the northern Apennines along two distinct sectors, sampling the orogeny-parallel variation suggested by apatite fission-track data. We apply an RF-migration procedure to data collected by broadband seismic stations, from both the permanent network managed by INGV and the temporary RETREAT project (Margheriti et al., 2006). We use the procedure developed by Bianchi et al. (2010) that is similar to those applied in other subduction zones or collisional belts (Hetenyi et al., 2009; Kawakatsu and Watada, 2007; Rondenay et al., 2008). Depth-migration targeted at 40 and 100 km depths focus on the topmost mantle, in a volume where the resolution of tomographic models is usually weak (Di Stefano et al., 2009; Giacomuzzi et al., 2012). Enhancing the resolution in this topmost part of the mantle is crucial for better understanding of active-tectonic processes in the region and how they formerly connected to older processes whose remnants are now buried at depth in the mantle.

Ps-converted phases that are implied by our receiver-function profiles relate to velocity interfaces in the upper mantle, or to sharp anisotropic gradients. We compare upper-mantle structural features consistent with the RF profiles with seismicity recorded over the past 20 years to formulate an evolution scenario of active processes in the Apennines.

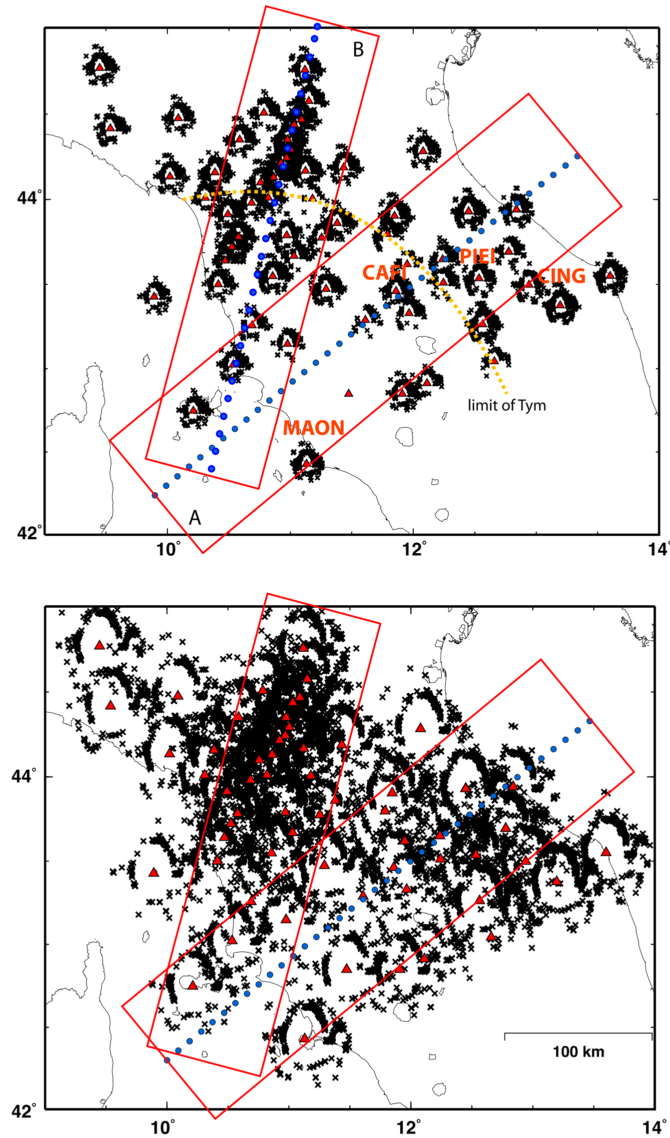

Fig. 2. Geometry of the seismic network used in this study. Red triangles show locations of seismic stations; black crosses represent piercing points at 40 (top) and 100 (bottom) km depth, for each teleseismic ray recorded at the stations. Blue dots are the discrete nodes along the profiles (spacing 10 km). Boxes A and B are relative to the profiles shown in Fig. 4 and 5. The yellow dashed line traces the extent of the high velocity slab present in the mantle at depth >160 km (from Giacomuzzi et al., 2012). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

2. Data and RF stacking

We selected high signal-to-noise P-wave coda of teleseismic events in the 30°–100° epicentral-distance range, and computed Receiver Functions (RF) from seismic records, after LQT rotation into ray-based coordinates, via frequency-domain multi-taper correlation estimates (Park and Levin, 2000). We used both permanent and temporary broadband seismic stations, collecting a variable number of teleseismic records with high signal/noise ratio for each station. For temporary networks, we obtained an average of 60 high-quality RFs for each station, while, for permanent stations, we selected an average of 100 high-quality RFs for each station. Our data show a good back-azimuthal coverage (see the distribution of piercing points in Fig. 2). The RFs were depth-migrated using a common-conversion-point technique (Dueker and Sheehan, 1998), to enhance the converted signals around a chosen depth (Bianchi et al., 2010). The migration accounts for different relative arrival

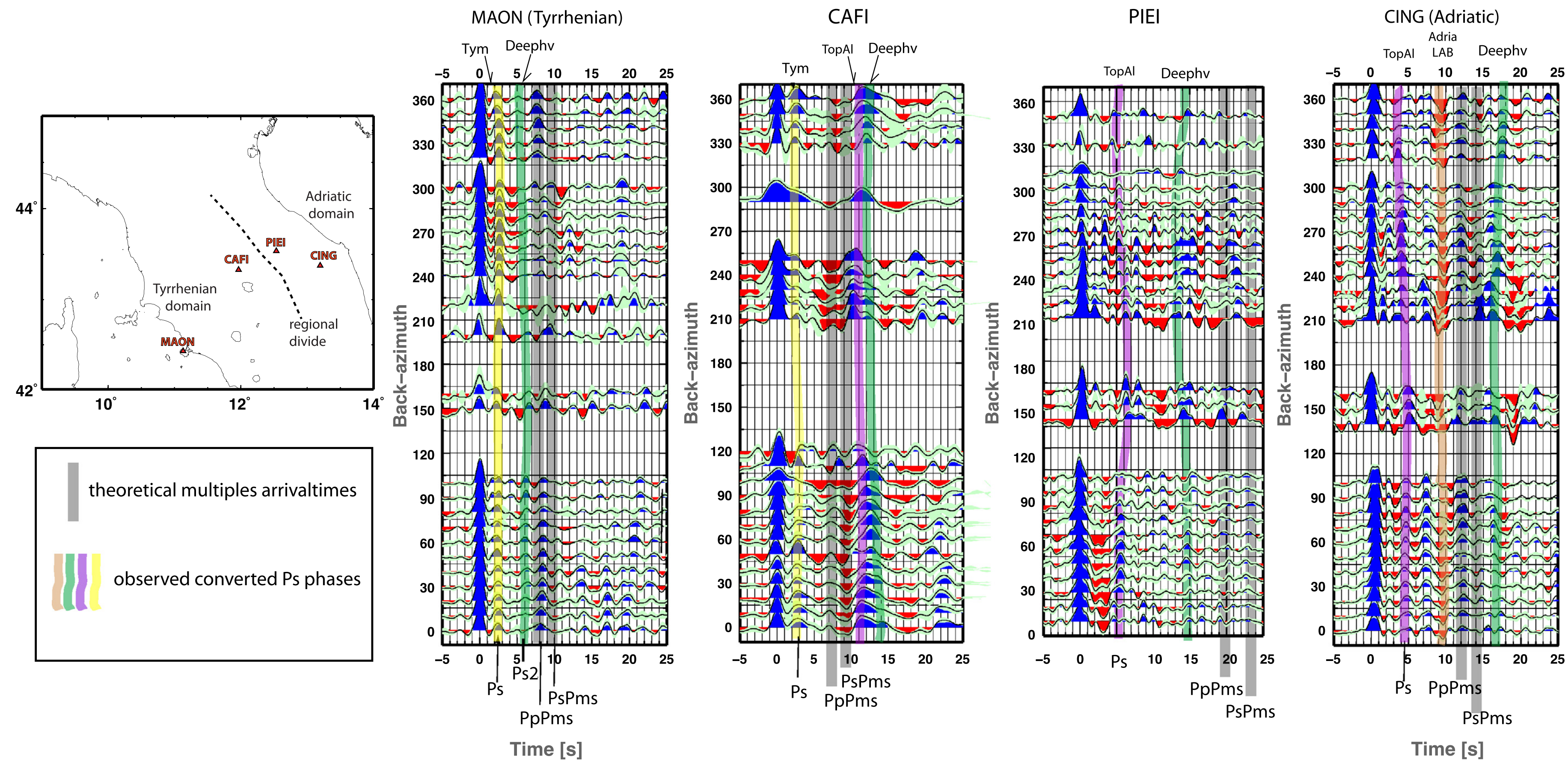

Fig. 3. Binned RFs at four stations along profile A–A′ (see Fig. 2 for location), showing the converted phases at the main interpreted interfaces and the multiples (PpPms, PsPms) of the converted phase at the Moho (Ps). Note that the interpreted deep features (yellow and purple lines) are clearly visible out of the Moho multiples. Abbreviations: Tym = Tyrrhenian Moho; Deephv = deep high velocity; TopAl = top of the Adria lithosphere.

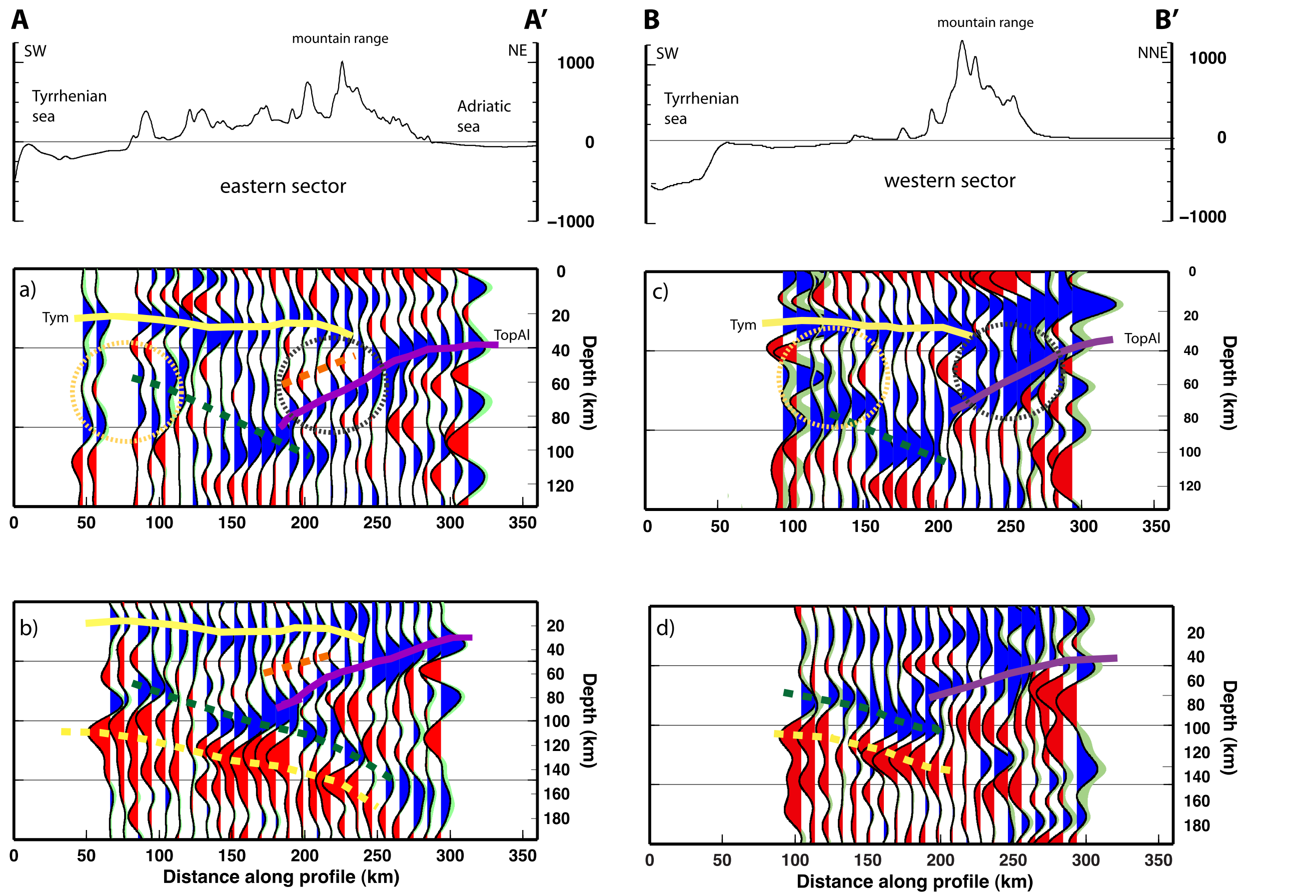

Fig. 4. Vertical sections of stacked receiver functions migrated at 40 (middle panels) and 100 km (bottom panels) depth for the eastern section (A–A′, left column) and western section (B–B′, right column) of northern Apennines. Top panels show topographic heights. Each profile shows major features: a), b), c) the positive (blue) phases relative to the sinking Adria Moho (purple line) and the flat Tyrrhenian Moho (yellow line); a), b) orange dashed line marks the top of a low velocity layer atop the sinking Adria Moho. b), d) NE-dipping layer marked by positive (earlier) and negative (later) phases (green and yellow dashed lines respectively).

times between the P-direct and P-to-S converted waves, due to differing locations of the conversion points along the same interface for different seismic paths (Bianchi et al., 2010). Migrated RFs were then grouped according to the surface projection of the conversion point and associated to the nearest node along a discretized profile (nodes are spaced at 10 km, see Fig. 2). For each node along the profile, the ensemble of Q-RFs is stacked together. The standard deviation of the stacked RFs was estimated using a bootstrap method with re-sampling of the entire RF dataset. The procedure allows focusing to the desired depth, while reducing the signal associated with multiples generated by shallow discontinuities (e.g. the Moho). The analysis of the RF data-set is given along two profiles that strike almost normal to the local axis of the orogen. One profile (box A in Fig. 2) transects the Apennines in the area where a number of geophysical observations (e.g. SKS splitting, Margheriti et al., 1996) suggest the presence of a retreating slab at depth. The second profile (box B in Fig. 2) crosses the Apennines where a crustal kinematics dominated by vertical motion has been recently proposed (Thompson et al., 2010).

Examples of the high-quality data used in this study are shown in Fig. 3 for four stations that best represent the Tyrrhenian and Adria lithospheric domains and the interposed Apennines orogen. Signals associated with the Moho discontinuities are clear and immediately point to a substantial depth difference of the Moho interface over the area. Multiples associated with the Moho interfaces are marked to show that the most prominent signals outlined by our study do not overlap their theoretical arrivals. The main features include:

- The shallowing and progressive disappearance of the Tyrrhenian Moho signal (Tym), eastward of the Apennines divide (visible at stations MAON - PIEI).
- The progressive westward deepening of the Adria lithosphere (TopAl). We refer to the part of the Adria microplate that plunges into the mantle, either of crustal or mantle origin, suggesting a peeling-back of the units stacked in the belt.
- The presence of a deep positive Ps converted phase (Deephv), slightly east-dipping beneath the entire region, paired with a deeper negative Ps converted phase (arriving at 15–20 s).

3. Results

Fig. 4 shows stacked RF profiles for two migration depths (namely 40 and 100 km) for profile A–A′ (Figs. 4a and 4b) and profile B–B′ (Figs. 4c and 4d). Comparison of the two migration depths demonstrates that certain features come into better focus after an appropriate moveout correction. The stacked Q-RF in the two profiles can be interpreted structurally, in terms of major S-wave velocity contrasts in the upper mantle, or else as anisotropic gradients related to anisotropy. Our stacks use constant weighting with back azimuth, so anisotropic signals should be suppressed. We interpret positive (negative) amplitudes that are shown in blue (red) to represent positive (negative) velocity contrasts with depth.

Along profile A–A′, the positive pulses of the crust–mantle interfaces are clearly visible (Fig. 4a). West-dipping positive contrasts outline the west-dipping Adria lithosphere (TopAl) that ends abruptly at roughly 80 km depth. In the Tyrrhenian side, the discontinuity (Tym) is almost flat at 20–24-km depth. Further Ps converted phases are prevalent at depths beneath the Tyrrhenian Moho, consistent with positive velocity contrasts (yellow dashed circle). Negative amplitudes are present in the volume between the two Moho interfaces in Profile A (orange dashed line), indicating the presence of a low-velocity zone beneath the mountain range. At greater depth, positive pulses (green dashed line, Fig. 4b) suggest an eastward-deepening velocity contrast, detectable from 60–70-km depth and dipping eastward down to 120 km. Deeper negative Ps conversions are present along the entire profile (yellow dashed line Fig. 4b) with a dip that parallels the overlying positive Ps conversion. The west-dipping Adria lithosphere lies above these east-dipping seismic features.

Along profile B–B′, the same overall features are visible indicating a first-order similarity of structures and tectonics along the strike of the northern Apennines. The principal difference resides in the shallow structure beneath the mountain range at the connection between the two Moho interfaces (grey dashed circle). In profile B–B′, the negative Ps converted phases, present in profile A–A′, do not extend beneath the orogen. This suggests that the northernmost segment of the orogen is not underlain by a low-velocity mantle nose.

4. Seismicity distribution

Seismicity in the Apennines is abundant. Historical (Stucchi et al., 2013) and digital-era catalogues (Chiarabba et al., 2005; Pondrelli et al., 2006) outline a main NW-trending strip of extensional earthquakes along the mountain range (Chiaraluce et al., 2004), a diffuse external belt of compressional earthquakes in the Adriatic foreland (De Luca et al., 2009; Carannante et al., 2013) and "sub-crustal" earthquakes along a broadly west-dipping plane (Amato and Selvaggi, 1993). The complete instrumental catalog yields a significantly improved characterization of tectonics and reveals some interesting features relevant to our RF stacks (Fig. 5). Shallow earthquakes (<100-km depth) dominate directly beneath the Apennines orogen. The seismicity cutoff is shallow (15 km) on the Tyrrhenian side, and numerous deep crustal earthquakes (20–30 km depth) are common on the Adriatic side. This seismicity has been ascribed to scraping process within the crust (Chiarabba et al., 2009; Carannante et al., 2013).

The distribution of intermediate-depth seismicity (depth greater than 30–40 km) distinguishes profiles A–A′ and B–B′ across the northern Apennines. Intermediate-depth seismicity is distributed beneath the orogen in profile B–B′, while it only occurs within the Adria lithosphere that plunges in the mantle in profile A–A′ (Fig. 5). This distribution is consistent with the difference in the "mantle nose" structure implied by the Ps converted-wave polarity. In the southern segment of the Northern Apennines, a low-velocity mantle nose is paired with a lack of co-located seismicity. In the northern segment, the mantle nose does not exhibit a velocity inversion, and is brittle enough for earthquakes.

5. Discussion

The similarity in the structure of the upper mantle observed in the two profiles indicates that a common process of delamination occurred at the expense of the Adria microplate. We distinguish delamination from subduction for the northern Apennines, based on the lack of evidence that the upper 20 km of crustal rock is carried into the mantle. The type-locality for continental subduction, the India–Eurasia plate convergence at the edge of the Tibetan plateau, has abundant evidence that the surface rocks of the Indian plate underthrust the Himalaya (Pandey et al., 1999; Ding et al., 2003; Leech et al., 2005; Kumar et al., 2006; Nabelek et al., 2009). By contrast, the uppermost sedimentary units in the Po river foredeep at the margin of the "downgoing" Adria microplate can be traced into the northern Apennines orogen across a deeply-buried blind thrust (Picotti and Pazzaglia, 2008; Bruno et al., 2011), suggesting that only the lower portion of the Adria microplate descends into the mantle (Chiarabba et al., 2009: Carminati et al., 2012). Ps receiver-function transects across the northern Apennines (Bianchi et al., 2010; Agostinetti et al., 2011; this study) and seismic tomography (Chiarabba et al., 2009) strongly suggest that the Tyrrhenian Moho represents a detachment surface with its leading edge beneath the Apennines crest, beneath which the lower crust and continental lithosphere of the Adria microplate separate from the upper crust.

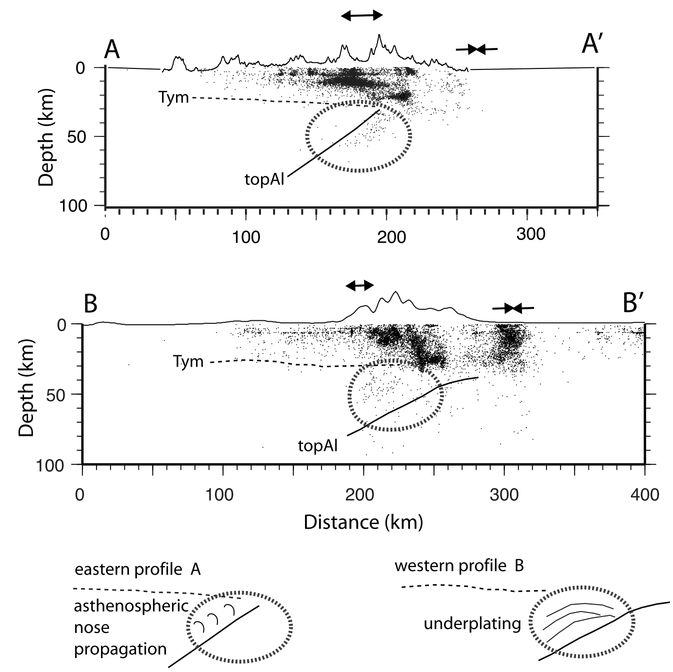

Fig. 5. Vertical sections of seismicity (CSI catalog, updated after Chiarabba et al., 2014) along profiles A–A′ and B–B′ (Fig. 1 for location). The geometry of the Tyrrhenian Moho and the top of delaminated Adria lithosphere are taken from Fig. 4. In the bottom of the figure, a cartoon shows the different processes active along the Apennines mountain range. In the western profile, deep underplating of crustal material prevails, while asthenosphere propagation prevails in the eastern profile (delamination-retreat).

The nature of the delamination process beneath the northern Apennines is not uniform along strike. For profile A–A′, the evidence suggests that delamination induces the flow of mantle rocks into the migrating detachment, in agreement with SKS fast directions that change from trench-normal to trench-parallel following the lithospheric delamination (Margheriti et al., 1996). The seismicity in the Adria-microplate lower crust that marks the terminus of the Tyrrhenian Moho is diffuse (Fig. 5), suggesting a broadly distributed deformation suggestive of a crack propagating into the Adria lithosphere (Reches and Lockner, 1994), even though most of the larger events have thrust mechanisms consistent with a simple blind fault (Pondrelli et al., 2006). The wedge between the Tyrrhenian Moho and the dipping surface of the descending Apennines slab is free of earthquakes, consistent with the ductile motion of asthenosphere. Piccinini et al. (2010) report high attenuation beneath the Tyrrhenian side of profile A–A′ consistent with hydrated rocks within this seismicity-free wedge.

The negative-polarity Ps conversions in the mantle wedge locate where Agostinetti et al. (2011) argued for dehydration metamorphism of the delaminated crust and upward fluid migration into the wedge beneath the mountain range. Hydration of the mantle wedge at the slab interface could bring peridotite closer to

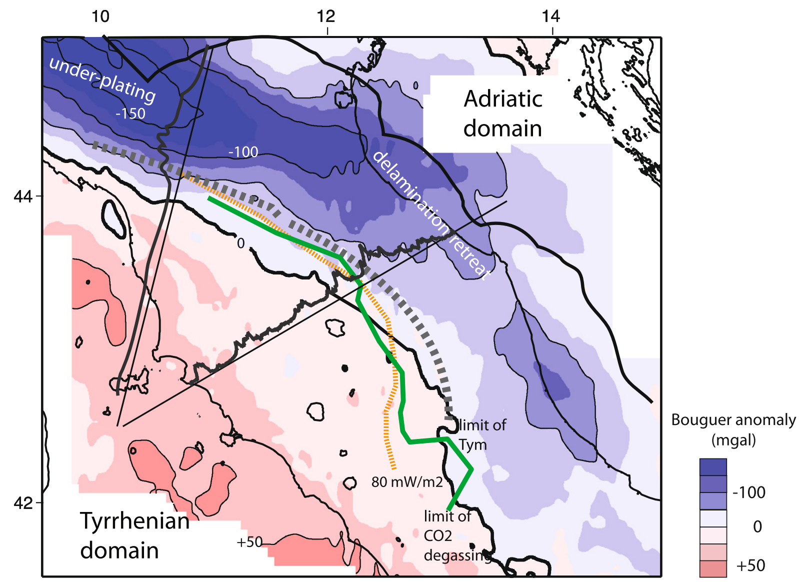

Fig. 6. Map of Bouguer anomalies (M.G. Ciaccio personal communication). Limits of the deep degassing region (green line) (from Chiodini et al., 2004), heat flow (orange dashed line) (from Della Vedova et al., 2008), and extent of the Tyrrhenian Moho (Tym, grey dashed line) inferred from this study are shown. The trace of the two profiles and the topography along the profile are plotted. Note the difference in topography across the Apennines between the two regions.

its solidus, decrease V_s, and induce a wavespeed inversion within the wedge, leading to negative-polarity Ps conversions. Geodynamic models (Gogus and Pysklywec, 2008) predict that a rapid increase in Moho temperature would migrate along the lithospheric gap, and continue to increase while sub-lithospheric flow is active. This scenario would help explain the anatectic magmatism of the region (Peccerillo, 2005).

Under profile B–B′, by contrast, small-event seismicity is present in the mantle wedge between the Tyrrhenian Moho and the top of the Apennines slab, and Ps converted phases in the wedge have positive polarity. We argue that hot ductile asthenospheric mantle is not sucked here into a growing lithospheric detachment. The seismicity suggests brittle rheology, and positive-polarity Ps argues against dehydration here in the downgoing slab. Piccinini et al. (2010) report lower t^* values for stations west of profile B–B′, suggesting lower attenuation at the northern extreme of the orogeny, though not consistently for stations on the profile itself. An explanation for the differing mantle-wedge properties beneath profiles A–A′ and B–B′ are suggested by the Bouguer anomalies (M.G. Ciaccio personal communication, Fig. 6). There is a prominent long-wavelength negative Bouguer anomaly in the western sector (−150 mgal), consistent with the low-density crustal material we infer beneath the mountain range. The negative Bouguer anomalies weaken southeast along the orogen, leading to a clear distinction between the cross-orogen profiles. Beneath profile B–B′, the absence of a velocity inversion beneath the mountain range, combined with the distributed pattern of sub-crustal seismicity (Fig. 5), suggest the accumulation of crustal material underplating the belt. This model is consistent with interpretation of recent tectonic uplift by Thomson et al. (2010), based on fission-track and (U–Th)/He ages, that the Apennines segment beneath profile B–B′ undergoes uniform uplift via crustal underplating, while the segment beneath profile A–A′ has undergone accretionary-wedge internal deformation.

The extent of the Tyrrhenian Moho matches the limit of the negative Bouguer anomalies and that of the strongly degassing, high-temperature flux region. This indicates that mantle substitution is complete underneath the Tyrrhenian side and part of the mountain range, responsible for high flux (Della Vedova et al., 2008) and deep-fluid degassing (Chiodini et al., 2004) related to a temperature increase at the base of the crust. We infer that a broad process of asthenospheric substitution is taking place beneath the mountain range only in the sector of the northern Apennines south and east of profile B–B′, roughly where high body-wave attenuation (Piccinini et al., 2010) overlaps the low-velocity anomaly in the uppermost mantle and the region of anomalous CO_2 flux (see also Chiarabba and Chiodini, 2013). This process of propagating mantle substitution is what we call delamination-retreat. We speculate that volatile release from the detaching lower crust (Agostinetti et al., 2011) facilitates delamination-retreat by lowering the viscosity of mantle peridotite in the mantle wedge, encouraging it to flow into the detachment. Where mantle hydration is lacking, delamination-retreat would stall, perhaps terminating the orogen, or else drawing crustal rock into the detachment as underplated material.

Our results might change the view of central Mediterranean tectonics. We propose that the evolution is better described as discontinuous processes of continental downwelling after delamination and buoyancy forces from the heterogeneous density distribution govern the present-day deformation of Central Italy, as modeled by Aoudia et al. (2007). This delamination process differentiates either in space and time along the belt, passing from crustal underplating to the northwest, to delamination-retreat to the southeast (Fig. 7). Despite the differentiation in the crustal and shallow-mantle structure, the continental sinks seems to be confined above a paired positive and negative velocity contrasts in migrated RF stacks at 100 km depth (Fig. 4).

The velocity jump has an apparent eastward dip toward the orogen and might be interpreted as a structural feature, in particular, as relict lithosphere from the earlier Alpine collision event. The negative-polarity Ps phase, that appears stronger than the positive Ps phase, could define the lithosphere–asthenosphere boundary (LAB) of the relic Europe lithosphere. Typically imaged via Sp receiver functions, Jones et al. (2010) report the LAB to lie at 96 ± 17 km in Phanerozoic Europe. Geissler et al. (2010) report the LAB to lie at 98 km beneath station VLC in Tuscany, from Sp receiver functions, which is largely corroborated by Miller and Agostinetti (2012) for RETREAT stations within the orogen. Both estimates are shallower than the negative Ps phase in our study, which lies at 100–140-km depth in the 100–175-km inter-

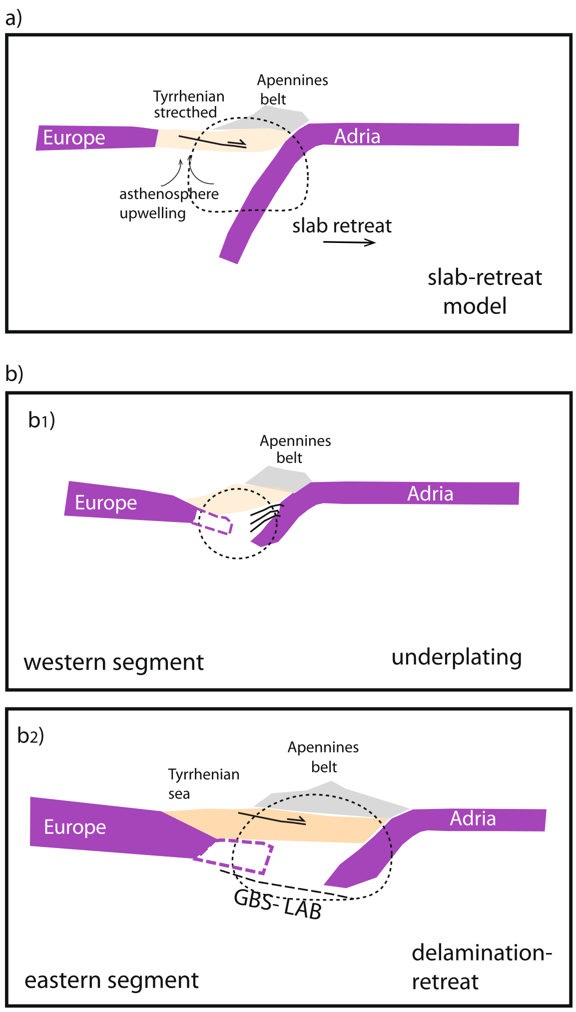

Fig. 7. Vertical sketches describing two different evolution models of the Apennines: a) subduction-retreat model, b) subduction via delamination model for the western (b_1) and eastern profile (b_2).

val within profiles A–A′ and B–B′. A better agreement is found with LAB estimates from shifts in anisotropic fabrics (Plomerova and Babuska, 2010), which average 133 ± 49 km in Phanerozoic Europe (Jones et al., 2010).

The placement of an east-dipping tabular body beneath the northern Apennines at this depth contradicts conventional interpretations of seismic tomography. When used to reconstruct the evolution of the Mediterranean mountain belts (see Faccenna et al., 2001; Wortel and Spakman, 2000; Rosenbaum and Lister, 2004; Faccenna and Becker, 2010; Handy et al., 2010), high-velocity mantle anomalies in tomographic models (Amato et al., 1998; Spakman et al., 1993; Lucente et al., 1999; Piromallo and Morelli, 2003; Giacomuzzi et al., 2011, 2012) have been interpreted as remnants of lithosphere subducted during the evolution of the Apennines system. Seismic tomography sustains a subduction-retreat model (discussed and reviewed by Faccenna et al., 2001; Lustrino et al., 2009, and Carminati et al., 2012). In these models, the evolution of the area, at least for the past 35 Myr, is dominated by the south–eastward retreat of a mixture of either oceanic or continental slab, usually referred as the Ionian/Adria slab.

In the southern Tyrrhenian region, consistent traces for this prolonged Ionian/Adria subduction are clearly outlined by high velocity anomalies in the mantle that are continuous. Here, the deep structure is easily imaged by teleseismic data (see the more recent results by Giacomuzzi et al., 2011, 2012) and local-earthquake data (see Chiarabba et al., 2008) allow imaging the shallow structure in detail. Conversely, in the central-northern Apennines the resolution of tomographic images based only on teleseismic body waves has been too coarse to resolve the upper 100–150 km. In particular, the sub-vertical high-velocity body found in the upper mantle by teleseismic tomography is too steep to define the slab polarity independently (see the most recent papers by Benoit et al., 2011 and Giacomuzzi et al., 2011, 2012), and a near-horizontal high-velocity layer 20–30-km thick (hypothetically a relict Alpine lithosphere, see Fig. 7) could remain unresolved by tomography.

Alternatively, it is possible to consider the negative velocity jump as a rheological transition and reconcile it with its occurrence within the Adria slab. Karato (2012) proposes that the seismic LAB is a rheological feature that marks the onset of grain-boundary sliding (GBS) as a deformation mechanism. Olugboji et al. (2013) explores how the GBS dissipation frequency ω_{gbs} depends on pressure, temperature and water content, fitting a parameterized model to the laboratory data of Jackson and Faul (2010). Using the Karato (2012) hypothesis, a transition in the grain-boundary sliding rheology creates a frequency-dependent V_S inversion in the slab at 120–140-km depths by shifting the natural GBS dissipation from low to high frequencies within the seismic band. The GBS transition occurs without weakening the long-term strength of the slab, because long-period elasticity is unaffected by the shift in ω_{gbs}.

Although it can explain how a discontinuity in elastic properties might develop across the grain of a downwelling lithospheric slab, the grain-boundary-sliding hypothesis for an LAB within the slab is still speculative and leaves other LAB-like features in the stacked RFs enigmatic. The negative polarity Ps phase under the Tyrrhenian Sea (50–125 km within Profiles A and B) is shallower than that within the delaminated Adria lithosphere, consistent with the latter body being cooler. However, negative-Ps is found at depth >100 km in this interval, implying that the Tyrrhenian plate has thickness typical of continental lithosphere, with no trace for a recent thermal event. This conflicts with the conventional model of Apennines-slab rollback, in which asthenosphere replaced the Tyrrhenian uppermost mantle no earlier than 15 Ma. The GBS model for the seismic LAB is thusfar speculative, and was formulated for an oceanic-plate model, so its application to continental regions may involve additional factors that we are not yet taking into account. However, extrapolation of the models derived by Olugboji et al. (2013) predict that a transition to the grain-boundary-sliding rheology in cold, dry continental lithosphere should be deeper than in warm, hydrated continental lithosphere, and this is consistent with our observations.

6. Conclusions

Velocity contrasts in the upper mantle defined by RF modeling support an evolutionary model for the northern Apennines that is based on the delamination of the Adria continental lithosphere. Lateral changes along the belt, envisaged by our new seismological data and consistent with previous inferences, indicate that the process is diachronous and heterogeneous in space passing from lithospheric underplating to delamination-retreat from northwest to southeast. These two different processes also have differing signatures in the topography of the belt and expression in seismicity, surface evolution, heat flow and magmatism.

The dipping Ps phase that marks the top of the Apennines slab begins at the Tyrrhenian Moho in the mid-crust of the Adria lithosphere. A cluster of seismicity in the crust of southern Tuscany (profile A–A′) suggests strains caused by the further penetration of this detachment into the Adria lithosphere. An absence of seismicity in the mantle wedge suggests that upper-mantle rocks, perhaps hydrated by the release of fluids from blueschist-to-eclogite metamorphism within subducted Adria lower crust (Agostinetti et al., 2011), flow up the slab to fill the wedge-volume opened by the detachment. In southern Tuscany the negative-polarity Ps within the mantle wedge could be caused by hydration of the lower part of the wedge. In northern Tuscany the seismicity at the incipient Tyrrhenian–Moho detachment extends into the wedge-volume, suggesting that rocks sucked into the detachment are brittle and resistant to flow, probably not hydrated and perhaps associated with crustal underplating, rather than asthenospheric flow. Ps in the wedge has positive polarity, consistent with no fluid release from the downgoing Adria lower crust. Termination of the Apennines slab near this profile is consistent with this evidence that the Tyrrhenian–Moho detachment in the northwest experiences greater resistance to further penetration of the Adria lithosphere.

We associate deeper (100–140 km) negative-polarity Ps conversions in the RF stacks with either a relict piece of lithosphere delaminated during the early stage of the Apennines formation or to a rheologic transition that has been associated in other regions with the "seismic" lithosphere–asthenosphere boundary (LAB). A relict Alpine lithosphere, however, would need to cross a "fast" tomographic feature that has been associated with the downwelling Apennines slab. The rheologic transition, alternatively, could occur within the Apennines slab, and create a seismic feature without a requiring a compositional or structural layering. In either case, the deep Ps conversions underneath the Tyrrhenian side of the Apennines argue against large-scale asthenosphere uplift under southern Tuscany. We favor instead the local uplift of slab-hydrated mantle under southern Tuscany, in conflict with conventional subduction-retreat models for the Apennines convergence zone.

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