Evidence for slab rollback in westernmost Mediterranean from improved upper mantle imaging

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d A R T I C L E  I N F O

Article history:
Received 15 November 2012
Received in revised form 28 January 2013
Accepted 19 February 2013
Editor: P. Shearer
Available online 29 March 2013

1. Introduction

The prevailing idea for the geologic evolution of the western Mediterranean in the Cenozoic is based on north dipping subduction near the gulf of Lyon undergoing vigorous rollback since ~30 Ma as a consequence of slow Africa–Eurasia convergence across the ~130 My old Alpine Tethys oceanic lithosphere (Lonergan and White, 1997; Malinverno and Ryan, 1986; Royden, 1993). As rollback propagated south, the subduction zone broke into three segments that continued to retreat independently to the east, south and west (Faccenna et al., 2004; Lonergan and White, 1997; Spakman and Wortel, 2004). The model for the evolution of the eastern slab is well established and generally accepted. That slab is now found under the Calabrian and Apenninic arcs and its roll-back led to the rotation of the Corsica–Sardinia block and the stretching of the crust in a back-arc setting producing the Liguria–Provençal and Thyrrhenian basins (Fig. 1) (Faccenna et al., 2004; Rosenbaum and Lister, 2004; Spakman and Wortel, 2004). The southern and western subduction segments are thought to have retreated to the south towards Africa and west towards Gibraltar, leading to the opening the Algerian basin, and extension in the Alborán terrane (Fig. 1) (Faccenna et al., 2004; Lonergan and White, 1997; Spakman and Wortel, 2004).

The latter part of this accounting is controversial, especially with regard to the geologic history of the Alborán domain and the associated Betic and Rif mountain range uplift (Fig. 1). From 27 to 5 Ma the Alborán continental crust underwent an episode of high-temperature metamorphism, was then uplifted, and transported a few hundred km into the westernmost Mediterranean where it was thrust onto the continental margins of Africa and Iberia as it gravitationally collapsed (Platt and Whitehouse, 1999; Platt et al., 2003, 1998). The cause of this activity and its association with the subducting ocean lithosphere is where the controversy lies. Several authors attribute this series of events fundamentally to a delamination (Calvert et al., 2000; Docherty and Banda, 1995; Platt et al., 1998; Seber et al., 1996) or Rayleigh–Taylor foundering (Houseman, 1996; Platt and Vissers, 1989) of Alborán lithosphere, whereas others make a direct link to slab rollback (Faccenna et al., 2004) or slab break-off (Zeck, 1996). These models appear to be incompatible with each other.

Tomographic imaging of mantle structure could help discern between the competing hypotheses. While both hypotheses predict the existence of a fast anomaly under the Alborán domain, these different hypothetical origins predict a significantly different morphology of the high-velocity structure. Extensive slab rollback predicts a vertically continuous anomaly reaching depths in excess of 600 km. Horizontal slices through this anomaly should show a well-defined linear or curvilinear shape, as the width of the subducted lithosphere should be much greater than its thickness. This would be true both for models that involve only slab rollback (Spakman and Wortel, 2004) or those that incorporate some Alborán mantle lithosphere delamination (Faccenna et al., 2004). Lithospheric delamination alone could produce an anomaly similar in shape to that caused by slab rollback but it should have a much shorter length, since the fast material found in the mantle would correspond only to the lithosphere directly under and to the west of the original position of the Alborán terrane. Foundering of a Rayleigh–Taylor stability would produce a much different anomaly. The high-velocity anomaly should in this case be detached from the surface and have a more rounded shape that is often described as blob-like.

The deep seismicity in the area provides additional and complementary information on mantle structure. A set of intermediate-depth earthquakes, defines a N–S trend of seismicity...
that occurs to the east of the strait of Gibraltar at depths above 160 km (Buforn et al., 2004) (Fig. 1). Because the Alborán domain was transported west a few hundred km (Platt et al., 2003), potentially over oceanic lithosphere, all models for the deeper high-velocity anomaly are consistent with this structure. An additional cluster of deep-focus seismicity occurs under southern Spain at ~640 km depth (Buforn et al., 2004) (Fig. 1). The occurrence of these earthquakes is consistent with a calculated low slab thermal parameter of 2800 km (Gorbatov and Kostoglodov, 1997) (using a lithospheric age of 130 m.y. and a time of initial subduction of 30 Ma). If this deep-focus seismicity were not occurring in old subducted ocean lithosphere, it would be the only known exception. An alternative interpretation consistent with the delamination hypothesis is that intermediate depth and the deep-focus seismicity occur within different bodies that are separated from each other: a shallow slab and a sunken piece of delaminated lithosphere (Calvert et al., 2000).

There are two questions we must answer in order to discriminate between the possible origins of the high velocity anomaly under the Alborán: Is there one continuous high-velocity body extending from the shallow subsurface to the bottom of the transition zone? And, does this body have a well-defined slab shape or is it a more massive, blob-like body?

Mantle imaging thus far has been ambiguous enough as to leave these two questions open to debate. Calvert et al. (2000) interpreted their model as showing two separate bodies, thus supporting a delamination model. While Blanco and Spakman (1993) interpret the imaged high-velocity structure as a detached slab, Spakman and Wortel (2004) used improved images from Bijwaard and Spakman (2000) to suggest a continuous slab. Gutscher et al. (2002) used the model from Bijwaard and Spakman (2000) to make the case for the presence of a slab, although Platt and Houseman (2003) interpreted the same tomographic images as supporting the delamination model. The divergent interpretations are possible because the anomaly imaged in Spakman and Wortel (2004) acquires a blob-like quality at depths greater than 400 km. Piromallo and Morelli (2003) and Faccenna et al. (2004) present mantle images that they interpret as resulting from slab-rollback followed by delamination, however the geometry of the anomaly in their images also makes it subject to re-interpretation as a large lithospheric drip.

More recently, additional seismic studies have supported the rollback hypothesis through the estimation of mantle strain based on SKS splitting (Bokelmann et al., 2011; Díaz et al., 2010) and analyses of seismic wave interaction with the high-velocity mantle structure (Bokelmann and Maufroy, 2007; Bokelmann et al., 2011). However, a consensus has not yet been reached, mostly because the paleo-position and geologic history of the Alborán domain, as well as the timing of under-thrusting and continental uplift in the Nevado–Filabride complex of southern Spain seem inconsistent with the evolution of the trench position predicted by the rollback model. Therefore, the mantle imaging and the geologic constraints appear to be at odds with one another and the mantle imaging has thus far not been conclusive enough to settle the debate.

In this study, we take advantage of large, coordinated seismograph deployments in Spain and northern Morocco (Fig. 1) to build a dense, high-quality dataset. We also use an improved seismic tomography procedure that combines frequency-dependent sensitivity kernels and iterative 3-D ray tracing to obtain an image of mantle structure with significantly improved resolution. The resulting velocity model shows a vertically continuous high-velocity anomaly under the Alborán domain and southern Spain that has a well-defined shape and is most consistent with the subducted slab hypothesis. With the new images, we propose a geodynamic interpretation that reconciles the rollback hypothesis with the geological observations.

2. Method

Classical (linear, ray-based) teleseismic tomography ignores ray-path dependence on velocity structure as well as the effects on travel times of velocity anomalies located off the geometric ray. The first limitation is addressed with iterative ray tracing, where the model is progressively updated and new ray paths are found at each iteration (e.g. Bijwaard and Spakman, 2000; Papazachos and Nolet, 1997; Sambridge, 1990; Zhao et al., 1992) and the second limitation has been addressed by using finite-frequency sensitivity kernels, most commonly the Born or banana-doughnut kernels of
Dahlen et al. (2000). Both of these limitations can be overcome more completely by using methods in which the sensitivities of the travel time delays to all model parameters are calculated for each waveform (e.g. Tromp et al., 2005). Of course, with each step away from simplification the computational burden rises. In this paper we present an approach of intermediate sophistication that combines iterative ray tracing with the use of approximated Born kernels, thereby addressing the two limitations mentioned above with a relatively modest computational cost.

We use a hybrid ray-tracing scheme that combines 1-D travel times to the outside of the model volume (the “box”) and 3-D travel times inside the box. Outside-the-box times are found using a \(\tau-p\) calculator, while inside-the-box times are found using a graph-theory method (Hammond and Toomey, 2003; Toomey et al., 1994). For each event–station pair, in each iteration we find the ray of minimum total travel time. To build the sensitivity matrix, we use the approximated Born kernels of Schmandt and Humphreys (2010), which are conformed to the ray paths found through the hybrid ray-tracing technique.

It should be noted that our inversion of relative travel-time delays is not very sensitive to the mean velocity at any given depth, and deviations from our 1-D reference model (AK-135, Kennett et al., 1995) would not be resolved. However, deviations in the 1-D structure will influence raypath locations and introduce uncertainties that, while not large, are difficult to predict. Additional uncertainty is introduced by the fact that we ignore 3-D structure outside of our model domain, which could affect the location of the entry point into the box, particularly for events coming from strongly heterogeneous regions like the Andean subduction zone. Despite these misgivings, it is reasonable to assume that accounting for the perturbational 3D structure inside the box yields ray paths that are closer to the real paths than those traced through a radial earth model.

Damping and smoothing regularization constraints are applied in a jumping manner, minimizing the whole model norm and roughness in each iteration. The dataset is composed of \(\sim 43,000\) delay times resulting from the cross-correlation of seismograms from 222 events (Fig. 2) in three frequency bands (center frequencies of 1.0, 0.5 and 0.3 Hz) recorded by 338 stations including 6 OBS instruments (Fig. 1). After each iteration, the delays are re-calculated using the travel times found by the hybrid ray-tracing procedure.

An important source of uncertainty in teleseismic tomography is the poor constraint that teleseismic arrival times place on depths shallower than 1–2 times the average station spacing. In addition to poor resolution near the top of the model, this leads to the possibility of un-imaged shallow anomalies being erroneously mapped into the deeper parts of the model. The problem is most commonly addressed with the use of station terms and crustal corrections. One can improve the constraint on those shallow layers by including local earthquakes in the inversion or by imposing constraints from other sources of information. In this paper we use a velocity model obtained from surface wave analyses to help constrain the shallow layers of the model and we also include delay time observations from 15 events located inside the box. The inside-the-box events are re-located after each iteration using a grid-search approach. The surface wave model we use to constrain the shallow layers (Palomeras et al., in preparation) is the result of the analysis of teleseismic earthquakes using the two plane wave approximation of Forsyth and Li (2005), covers \(\sim 35\%\) of our study area (Fig. 2) and we use it as a constraint for depths shallower than 80 km. The S-wave velocities from the surface wave model were converted to P-wave velocities using a constant \(V_p/V_s\) ratio of 1.8 and the mean velocity value at each depth is removed to produce anomalies compatible with the inversion of relative time delays. We use this model as starting point for our inversion and apply damping to the model updates in the affected nodes. Given that lateral resolution provided by the body wave data is better than that of the surface wave inversions, damping is applied in a manner that allows the model to evolve in the upper layers.

Of the local earthquakes we use for shallow structure constraint, most are located in the Gulf of Cadiz, five are of intermediate depth (>35 km), two occur at depths of 50–90 km within the subducted slab (Fig. 1), and one is the deep-focus earthquake of April 2010 beneath southern Spain (Bezada and Humphreys, 2012; Buforn et al., 2011), which provides an especially useful illumination of the structures.

We can gauge the effect of including 3-D ray tracing into the tomography workflow by comparing the final iteration results with those of the first iteration (which uses rays traced through the reference 1-D velocity model). The most significant change is its effect on the high-velocity slab-like anomaly. With iteration, the arcuate shape of this anomaly is more evident, and its thickness and lateral extent are reduced, giving the impression of a more focused image (Fig. 3). Synthetic tests confirm that this is the expected effect of a fast structure with a geometry similar to that imaged. When the geometry of the input anomaly is known, the last iteration more accurately reproduces the
prescribed geometry and lateral extent of the anomaly, especially at depths greater than ∼200 km (Fig. S2, Supplementary Material). This improvement in the imaging is of critical importance in our study area, given the necessity to accurately image the shape of the anomaly in order to distinguish between a subducted slab and a lithospheric drip.

Our final model adequately accounts for much of the observed delay times. The RMS of the observed delays has a value of 0.57 s, whereas the RMS of the residuals after the last iteration has a value of 0.23 s, corresponding to a total variance reduction of 84%. For details about the evolution of the model and the data misfit with iteration see the Supplementary material.

3. Model description and comparison with previous models

Our model extends to the Earth’s surface but the shallower structure is constrained mainly by the surface wave model of Palomeras et al. (in preparation). We focus our discussion on anomalies deeper than 75 km, where the body wave constraints are most important. The most prominent imaged feature is in the structure we interpret to be the subducted Alpine Tethys slab. This arcuate, vertically continuous high-velocity anomaly is more than 600 km long and is located beneath the western Alborán Sea and southern Spain (Fig. 4). Additional anomalies of interest include low-velocity anomalies beneath the Middle Atlas and NE Morocco and a small high-velocity velocity anomaly ∼450 km beneath the Middle Atlas.

The Alborán slab anomaly dips steeply to the east, extending continuously from 50 km to depths greater than 600 km. Below 350 km it is nearly vertical. Calvert et al. (2000) suggested that the uneven distribution of teleseismic raypaths in the westernmost Mediterranean predisposes tomographic models towards continuous, steeply east dipping anomalies and advocates the presence of a gap in the high-velocity feature. Synthetic tests with gaps at different depths show that a gap in the slab can be resolved by our experimental geometry and method (Fig. 5), strongly suggesting that the continuity of the anomaly in our model reflects the real geometry of the slab and is not an artifact of ray distribution.

At its greatest depth of ∼640 km, the anomaly is found mostly under southeastern Spain, elongated to the northeast. Above 150 km, the anomaly is composed of 2 arms that are orientated at right angles to each other. The arm of greater amplitude anomaly is located east of the strait of Gibraltar, under the Alborán Sea at a longitude of ∼4.5°W and trending N–S. The E–W trending arm is imaged under southern Spain at a latitude of ∼37.5°N. The Spanish arm of the anomaly appears not to reach the surface.

Fig. 3. Comparison of horizontal slices through the resulting model at different depths (as indicated on the figure) for the first iteration (1-D ray-tracing, top row) and last iteration (iterative 3-D ray-tracing, bottom row). Note that ray tracing reduces the spatial extent of the high-velocity anomaly and largely eliminates the “halo” of low velocities present near the slab in the 1-D case.
(as noted by Spakman and Wortel (2004) and Garcia-Castellanos and Villaseñor (2011)). Specifically, the anomaly under the Alborán is present at depths as shallow as 50 km (i.e. at depths below the surface plate) with amplitudes greater than 4%, while the anomaly under southern Spain is small and weak at a depth of 75 km, and does not attain the magnitude of the Alborán arm until depths of 160 km. The two arms appear to be separated by a narrow vertical gap at depths shallower than 125 km. At this depth, the Alborán arm of the anomaly has a horizontal length of ~200 km and a width of 75–100 km, and the Spanish arm is ~150 km long and ~60 km wide. Below 150 km depth the two arms join to form a single, laterally continuous anomaly.

In the model of Piromallo and Morelli (2003, PM), the fast anomaly does not extend above ~200 km, and the area directly west of the strait of Gibraltar shows slow velocities instead. The model presented in Spakman and Wortel (2004, BS) does show a
Fig. 5. East–West cross-sections through input and recovered models for seven test cases and the real model for reference. Location of the cross sections slightly north of the strait of Gibraltar as in figure X. Test cases: A, continuous slab, no flat-lying segment; B, slab with a gap at 100–250 km depth, no flat lying segment; C, slab with a gap at 250–300 km depth, no flat lying segment; D, continuous slab with a short flat-lying segment; E, continuous slab with a long flat-lying segment; F, continuous slab with a detached flat-lying segment. G, Same as A, but DC shifted by 2%. Magnitude of input anomalies is 5%. Test case is indicated in each cross section, subscripts “i” and “r” indicated input and recovered models respectively. The tests show that gaps in the slab are resolvable but the data are not very sensitive to a flat-lying section of slab at the bottom of the transition zone and a DC shift in the anomalies cannot be recovered. Notwithstanding the lack of sensitivity to flat-lying slabs, subtle differences in the recovered model for test cases D–F suggest that we may be able to tentatively discern between these three scenarios.
fast velocity anomaly in this general area, although its shape is less sharply defined and generally shifted to the west with respect to our image. In both the PM and BS models, the fast anomaly attains a magnitude of \(-\sim 2\%\), in contrast to the \(-\sim 4\%\) in our model.

Between 200 and 350 km depth, our model shows a fast anomaly with a generally arcuate shape that is concave to the east or southeast. At this depth range, the anomaly we image is broadly similar in shape to that in the PM and BS models, although generally our anomaly is laterally smaller and has greater amplitude. The difference between our model and previous ones becomes strong at depths below \(-\sim 400\) km. Whereas the PM and BS models show blob-like anomalies at these depths, the anomaly in our model preserves a well-defined concave-east arcuate shape down to \(-\sim 600\) km depth (Fig. 6). At 635 km depth, near the bottom of the model, our anomaly is greater in magnitude than previous models (3–4% in contrast to 1.5–2%), and although some slow anomalies appear in areas surrounding the fast anomaly in our model, these are of smaller magnitude (\(-\sim 1\%)\) and different in form from the pervasive slow rim surrounding the fast anomaly in both the BS and PM models. As noted in the previous section, because we are not able to resolve any mean shift in velocity away from the reference 1-D model (Fig. 5), we are thus unable to resolve a relatively slow ambient mantle (as imaged with absolute travel times in the BS and PM models). Hence, the amplitude of the velocity anomaly in our model should be interpreted as the relative contrast between the slab and the surrounding mantle. If all velocities were made slower, the amplitude of our slab anomaly would then be closer to the values shown by BS and PM.

Both the PM and BS models image significant volumes of high velocity mantle at the bottom of the transition zone, that in both cases are interpreted as flat-lying sections of subducted slabs (Facenna et al., 2004; Spakman and Wortel, 2004). Our model does not show equivalent features, but there is a deep anomaly east of the arcuate slab anomaly (Fig. 4) that suggests the presence of fast mantle in the lower part of the transition zone in this poorly-sampled part of our model. Resolution tests were carried out to explore different scenarios and they show that our experimental geometry is not able to resolve a flat-lying piece of slab on the bottom of the transition zone (Fig. 5, Supplementary material). Furthermore, these tests suggest that the deep anomaly SE of the arcuate slab anomaly near the bottom of the model (Fig. 4, 580 km depth slice) may indeed be an artifact caused by incorrect mapping into the model of travel time advances occurring during transit through a flat piece of slab.

We conclude that our data is consistent with the presence of a flat-lying piece of subducted slab in the bottom of the transition zone east of the Alborán slab and suggest that this piece of slab is not attached to the Alborán slab (Fig. 7, Supplementary materials, Fig. S3).

Another noteworthy aspect of our model is a pair of anomalies that occur under the Atlas Mountains. At 75 km depth we find prominent low-velocity anomalies (\(-\sim 3\%)\) located under the Middle Atlas range and near the coastline on the Morocco–Algeria border. Smaller low-velocity anomalies under the southwestern High Atlas range complete a SW–NE trending corridor of low-velocity anomalies that follow the general trend of the Atlas mountain chain. Although these anomalies can be traced to depths as great as 200 km, their magnitudes diminish substantially below 90 km. At depths between 390 and 530 km, under the Middle Atlas, we find a high-velocity anomaly with a diameter of \(-\sim 100\) km that is separate from the Alborán slab anomaly. These two anomalies are interesting because they provide evidence for a delamination event under the Middle Atlas. The slow anomaly could represent asthenosphere at shallow depth while the fast anomaly could correspond to the delaminated Atlas lithosphere.

4. Discussion

The high-velocity anomaly that we image under the westernmost Mediterranean has the characteristics of a steeply east-dipping subducted lithospheric slab. Our imaging shows this structure does not have a blob-like quality at any depth and instead has an arcuate shape (Fig. 4) that is similar in appearance to the Calabrian slab (e.g., Piromallo and Morelli, 2003; Spakman and Wortel, 2004), as expected for a subducting narrow slab (Funiciello et al., 2003; Schellart et al., 2007). The structure is continuous from the surface,

Fig. 6. A horizontal slice at 435 km through our velocity model and slices through recent models at comparable depths. PM: Piromallo and Morelli (2003). BS: Bijwaard and Spakman (2000) (reproduced from Spakman and Wortel (2004)). Note the arcuate shape of the anomaly in our model at this depth as opposed to the more blob-like shape of the anomaly in previous models.
east of Gibraltar, to the base of the transitions zone, and it is seismically active at intermediate and deep depths.

The geometry of the high-velocity anomaly is consistent with a slab origin. If this feature is restored to the surface by unfolding it without significant in-plane strain, it fills the area occupied by the western Alpine Tethys at ∼30 Ma (Fig. 8). That is to say, the anomaly we image has roughly the shape of the lithosphere that presumably subducted.

Attributing the high-velocity anomaly in the westernmost Mediterranean to subducted ocean lithosphere is not new (e.g. Blanco and Spakman, 1993; Faccenna et al., 2004; Gutscher et al., 2002; Piromallo and Morelli, 2003; Spakman and Wortel, 2004). However, this is not generally accepted. The primary objection comes from the apparent incompatibility of this model with the presence of the thinned continental crust of the Alborán domain, which lies between Africa and Iberia, where rollback would have occurred. Previously, authors have attempted to address this point by proposing that the Alborán domain migrated 500–700 km into its current position following the retreating trench from the Balearic margin, with extension in the backarc creating the Algerian Basin (Lonergan and White, 1997; Spakman and Wortel, 2004). The amount of Alborán motion required by this model is not consistent with palinspastic reconstructions based on balanced cross-sections of the Betic–Rif belts that constrain Alborán–Iberia convergence to ∼250 km since 20 Ma (Platt et al., 2003, 2013) and there is no geologic evidence for the presence of Alborán domain remnants that far east or the existence of a strike-slip fault required for this transport (Platt et al., 2003).

The paleo-position of the Alborán terrane obtained from geological constraints places it on the subducting plate, within the area involved with subduction (Fig. 8). We therefore prefer a model that has rollback beginning at the Balearic margin northeast of the Alborán domain and continuing under the Alborán crust, taking the Alborán mantle lithosphere with it. Such a model is consistent with the thermal history of the Alborán domain. Starting at ∼27 Ma the base of the Alborán crust experienced near-asthenospheric temperatures accompanied by uplift, suggesting that most of the lithospheric mantle was lost (Platt et al., 1998). This has been attributed to convective removal or delamination of an over-thickened lower lithosphere (Houseman, 1996; Platt and Vissers, 1989). We propose that the delamination event occurred as a result of subduction of the Alborán lithospheric mantle along with the larger slab that we find presently under the westernmost Mediterranean. The Alborán crust, having avoided subduction, would have been transferred or accreted to the overriding plate. Given the early history of high pressure–low temperature metamorphism of the Alborán crust (Azañón and Crespo-Blanc, 2000; Platt et al., 2005), it is probable that some degree of subduction and later exhumation of the crust preceded the delamination event (Bialas et al., 2011; Brun and Faccenna, 2008).

After the delamination–accretion event, rollback would have continued by consuming the lithosphere west of the Alborán, which we presume to be largely oceanic in nature (given the occurrence of intermediate depth seismicity that usually is linked to dehydration reactions in subducting oceanic lithosphere, Green...
and gravitational collapse of the Alborán crust and its translation following trench retreat would have driven emplacement of the Betic–Rif thrust belts, while subduction under the western Alborán margin would have caused the calc-alkaline magmatism that intruded the Alborán at this time (Duggen et al., 2004). In this, our model differs from the otherwise similar model of Faccenna et al. (2004). They interpret the absence of ophiolitic remains as evidence for a lack of oceanic subduction following the delamination of the Alborán lower crust implying that delamination was the final stage in the process. In our view, it is difficult to account for the ~200 km of westward translation undergone by the Alborán after the thermal event (Platt et al., 2003) as well as the calc-alkaline magmatism in the Alborán (Duggen et al., 2004) without subduction of ocean lithosphere in the area over which the Alborán crust moved.

The current position of the slab beneath southern Spain (Fig. 4), and its location when restored to the surface (Fig. 8b), suggests that relative motion between Iberia and Africa (Dewey et al., 1989; Jolivet and Faccenna, 2000; Rosenbaum et al., 2002a, 2002b) was accommodated by an underthrusting of ocean lithosphere beneath the Iberian margin.

Our model resembles the westward-directed rollback model of Spakman and Wortel (2004). The main difference lies in the initial location of the Alborán domain and the nature of its interaction with the retreating subduction zone. In our interpretation the Alborán lower lithosphere is a part of the subducted slab.

5. Conclusions

We image the western Mediterranean upper mantle with improved resolution through the use of a large teleseismic dataset, a crustal velocity structure derived from surface wave modeling (Palomeras et al., in preparation) iterative 3-D ray tracing and finite-frequency kernels. Our main conclusions are:

1. The dominant upper mantle feature is the 4–5% high-velocity, slab-like structure shown in Fig. 4. This structure is connected to the surface and extends to the bottom of the transition zone.
2. Its form, when restored to the surface, fills the area occupied by the Alpine Tethys ocean lithosphere and embedded continental Alborán domain at 30 Ma (Fig. 8b).
3. We interpret the high-velocity structure as a slab of lithosphere composed of subducted Alborán mantle lithosphere and the surrounding Alpine Tethys ocean lithosphere.
4. This lithosphere is thought to have been subducted during ~30 m.y. of generally west-directed subduction rollback, which continued beneath the Alborán domain by delaminating its mantle lithosphere.
5. A more drip-like delamination of the Middle Atlas mantle lithosphere may be evidenced by the irregular but pronounced zone of low velocities beneath these mountains at 50–100 km depth and an underlying high-velocity body at ~400–500 km depth.

Acknowledgments

We thank Claudio Faccenna and Wim Spakman for thoughtful and insightful reviews that helped improve the manuscript. Our understanding of the westernmost Mediterranean has greatly benefited from many conversations with John Platt. This work is funded by the PICASSO project (NSF grant EAR-0808939), The deployment and data processing for Spanish stations was funded by Consolider-Ingenio 2010 project TOPO-IBERIA (CSD2006-00041) as well as ALERT-ES (CGL2010-19803-C03-02). We thank Ingo Greve mayer for access to data from the TOPO-MED OBS deployment as well as Christine Thomas and James Wooley for access to data from Münster and Bristol stations in western Morocco.

Appendix A. Supporting information

Supplementary data associated with this article can be found in the online version at http://dx.doi.org/10.1016/j.epsl.2013.02.024.

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