Imaging Canary Island hotspot material beneath the lithosphere of Morocco and southern Spain

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Abstract

The westernmost Mediterranean has developed into its present day tectonic configuration as a result of complex interactions between late stage subduction of the Neo-Tethys Ocean, continental collision of Africa and Eurasia, and the Canary Island mantle plume. This study utilizes S receiver functions (SRFs) from over 360 broadband seismic stations to seismically image the lithosphere and uppermost mantle from southern Spain through Morocco and the Canary Islands. The lithospheric thickness ranges from ~65 km beneath the Atlas Mountains and the active volcanic islands to over ~210 km beneath the cratonic lithosphere in southern Morocco. The common conversion point (CCP) volume of the SRFs indicate that thinned lithosphere extends from beneath the Canary Islands offshore southwestern Morocco, to beneath the continental lithosphere of the Atlas Mountains, and

then thickens abruptly at the West African craton. Beneath thin lithosphere between the Canary hot spot and southern Spain, including below the Atlas Mountains and the Alboran Sea, there are distinct pockets of low velocity material, as inferred from high amplitude positive, sub-lithospheric conversions in the SRFs. These regions of low seismic velocity at the base of the lithosphere extend beneath the areas of Pliocene-Quaternary magmatism, which has been linked to a Canary hotspot source via geochemical signatures. However, we find that this volume of low velocity material is discontinuous along strike and occurs only in areas of recent volcanism and where asthenospheric mantle flow is identified with shear wave splitting analyses. We propose that the low velocity structure beneath the lithosphere is material flowing sub-horizontally northeastward beneath Morocco from the tilted Canary Island plume, and the small, localized volcanoes are the result of small-scale upwellings from this material.

Keywords

Lithosphere-asthenosphere boundary, mantle upwelling, receiver functions, western Mediterranean

1. Introduction

The tectonics of Northwest Africa have largely been dominated by late stage convergence of Africa with Eurasia during the Cenozoic. The last piece of Mesozoic oceanic lithosphere from the Tethys Ocean is being consumed across the Mediterranean, forming a series of arcuate shaped subduction zones [Royden, 1993; Rosenbaum *et al.*, 2002; Faccenna *et al.*, 2004; Govers and Wortel, 2005]. The tightly curved Gibraltar arc is one of these fragments, occupying the westernmost Mediterranean, bounded by the Rif Mountains along the

southern cusp and the Betic Mountains along the northern cusp. These interior units of the Gibraltar arc are part of a larger domain where the plate boundary between Africa and Iberia is distributed as a wide zone of deformation [e.g. *Koulali et al.*, 2011]. This arcuate region of rapidly uplifting topography in the Rif and Betics has fueled competing hypotheses of lithospheric-scale modification: mechanical delamination associated with adjacent slab rollback or convective-removal due to density contrasts induced by thickening or a combination of both mechanisms [e.g. *Duggen et al.*, 2005; *Platt et al.*, 2013; *Levander et al.*, 2014], but all are based upon the premise that the tectonics are controlled by the collision of Africa with Eurasia.

The lithosphere of northern and central Morocco is actively deforming over a 300-400 km wide belt that is bounded to the north by the Rif Mountains. The southern extent of the deformation zone is the High Atlas Mountains, a NE trending intercontinental mountain belt that formed by reactivation of Triassic-Jurassic aged rift structures that were inverted during the Cenozoic convergence of Africa with Eurasia [e.g. *Brede et al.*, 1992; *Gomez et al.*, 2000; *Pique et al.*, 2002; *Arboleya et al.*, 2004; *Missenard et al.*, 2006]. The high elevations of the Atlas are superimposed on a broad topographic swell that consists of the west Moroccan Meseta, the eastern High Plateau, and the Anti-Atlas Mountains in the southwest - none of which have undergone recent significant deformation (Figure 1A).

Much of the knowledge of crustal and lithospheric structure of northwestern Africa, until recently, had been obtained by geophysical modeling based on gravity, heat flow measurements, and seismic refraction experiments [*Teixell et al.*, 2003; *Ayarza et al.*, 2005;

Teixell et al., 2005; Zeyen et al., 2005; Missenard et al., 2006; Fullea et al., 2010; Jimenez-Munt et al., 2011; Ayarza et al., 2014]. These studies found a large range in lithospheric thicknesses across Morocco, with the thickest regions beneath the West African craton at ~180 km and thinnest lithosphere beneath the Atlas Mountains (~65 km). P-wave and surface wave tomography studies have also shown a thin lithosphere and a shallow low velocity zone beneath the Atlas [Seber et al., 1996; Calvert et al., 2000; Bezada et al., 2014; Bonnín et al., 2014; Levander et al., 2014; Palomeras et al., 2014]. Recent regional receiver function studies have also found that the lithosphere is thin beneath the westernmost Atlas and Anti-Atlas [Spieker et al., 2014], the central Atlas [Miller and Becker, 2014], and the eastern Rif [Dündar et al., 2011; Mancilla et al., 2012; Thurner et al., 2014].

An intriguing linear trend through northern Morocco towards southeastern Spain hosts Pliocene to Quaternary volcanism (see red polygons, Figure 1A) [Hoernle et al., 1995; Duggen et al., 2004; Duggen et al., 2009 and references therein]. This post-collisional magmatism in the region can generally be categorized into two groups: (1) alkali basalts that have geochemical similarities to intra-plate volcanics - possibly associated with plume-contaminated sub-lithospheric mantle, and (2) calc-alkaline, near the Mediterranean coast in both the Rif and the Betics that are linked to subduction beneath the Alboran [e.g. Duggen et al., 2005; Bosch et al., 2014]. This unusual linear trend of anorogenic Cenozoic magmatism, along with the thinned lithosphere beneath Morocco and uplifted topography, has been described by Frizon de Lamotte et al. [2009] as the “Moroccan Hot Line”. The thinning of the lithosphere and correlation with Cenozoic volcanism is suggested to be attributed to edge-driven convection [Ramdani, 1998; Missenard and Cadoux, 2012;

Kaislaniemi and van Hunen, 2014], delamination of the root of the Atlas Mountains [Ramdani, 1998; Bezada et al., 2014; Levander et al., 2014], or linked to the Canary Island hotspot [Hoernle et al., 1995; Oyarzun et al., 1997; Anguita and Hernan, 2000; Zeyen et al., 2005; Duggen et al., 2009; Miller and Becker, 2014; Mériaux et al., 2015].

The Canary Islands are an intra-plate island chain that is located ~100 km off the Atlantic coast of Morocco that are built upon Jurassic-age oceanic lithosphere (Figure 1A). The seven major islands are part of a ~600 km long chain of volcanoes that have been active since the early Miocene [Schmincke, 1982]. The age of the volcanoes decrease from east to west, which has been interpreted as being from a hot spot origin [e.g. Morgan, 1972]. The Canary Islands have been extensively studied yet the crustal, lithospheric, and uppermost mantle structure is still debated. However, recent seismic studies are consistent with a deep mantle source, revealing a thin crust, a low velocity layer just beneath the lithosphere [e.g. van der Meijde et al., 2003; Lodge et al., 2012; Martinez-Arevalo et al., 2013; Bonnín et al., 2014], and a depressed 410 km discontinuity [Saki et al., 2015].

Here, we present seismological evidence from S receiver functions (SRFs) to show that thinned continental lithosphere extends across the inferred “Moroccan Hot Line” beneath the Atlas Mountains. Along this corridor a large volume of low seismic velocity material has pooled beneath the lithosphere, as inferred from high amplitude positive, sub-lithospheric conversions that extend from the Atlas, across the Alboran and into the southern Iberia Peninsula. This low velocity feature is coincident with the areas of Pliocene-Quaternary magmatism, which are consistent with a Canary hotspot source, not as a clearly defined

sub-horizontal, sub-lithospheric channel, but as a discontinuous structure along strike of low velocity material that is inferred to be hot and perhaps melt bearing.

2. Data and Methods

Data were collected at 360 broadband stations deployed in the westernmost Mediterranean and the Canary Islands between 2007 and May 2013. The 102 XB network stations (in yellow triangles in Figure 1B) were part of the NSF-funded PICASSO experiment that extended between central Morocco and central Spain. The twenty-one 3D and ZI network stations (in green in Figure 1B) were deployed by the University of Münster and the University of Bristol, respectively [Spieker *et al.*, 2014]. One hundred thirty-four stations (shown in blue in Figure 1B) were part of the Topolberia-Iberarray experiment that extends across the Iberian Peninsula to southern Morocco. The rest are a combination of permanent stations, including the extensive ES network, and deployments with broadband data available from the IRIS DMC.

Over 160 Mw 6.0+ earthquakes between 55-85 degrees away from the broadband stations were analyzed to produce ~5300 individual S receiver functions (Figure 1B inset). The data for these final ~5300 SRFs were selected by first hand-editing the seismograms for each event-station combination, then producing and hand-editing each receiver function multiple times for each individual RF. Correction for residual statics using a cross-correlation method was applied to account for near surface velocity irregularities beneath the stations. The resulting time-domain SRFs were then mapped from time to depth, stacked at each station (receiver gathers) or mapped station by station, for each earthquake, by

summing all the individual event images (Common Conversion Point imaging). The process of quality controlling the receiver functions involves no *a priori* knowledge of event parameters.

The S receiver function technique [Langston, 1977; Vinnik, 1977] utilizes steeply inclined teleseismic shear waves (S) and the partial conversion of energy into a compressional wave (Sp) at velocity boundaries beneath a seismic station. The three-component broadband seismic data were rotated from ZNE, into ZRT, and then into the LQT coordinate system [Vinnik, 1977] from the incidence angles predicted by the 1D velocity reference model ak135 [Kennett *et al.*, 1995]. Then the Q (S phase) motion was deconvolved from the L (Sp phase) motion using frequency domain water level conditioning in order to isolate the Sp phase(s) and produce 0.1 Hz receiver functions. The 10s Gaussian was chosen based on previous work using synthetics and data from other regions that utilized this code [Miller and Eaton, 2010; Levander and Miller, 2012]. These previous works found that 0.1Hz accurately position the conversion from both the crust-mantle boundary and the lithosphere-asthenosphere boundary (See Supplementary Figure 2 in Levander and Miller, 2012 for example). Finally, the SRFs are reversed in polarity and time domains to compare directly with more commonly known P receiver functions, so that red denotes a positive impedance contrast, such as the Moho conversion.

Due to the high spatial density of the temporary stations in the region common conversion point (CCP) stacks were computed [Dueker and Sheehan, 1997]. This methodology allows for more accurate mapping of the signals to depth and lateral space at their approximated conversion points by taking advantage of overlapping conversion points from neighboring

stations. This technique maps individual SRFs to depth based on ray-theory pathways that are a function of three-dimensional velocity structure within the imaging volume. A merged model taken from *Bezada et al.* [2013] and *Palomeras et al.* [2014], using a constant V_p/V_s ratio of 1.78, is used for velocity-depth mapping. As seismic waves are sensitive to a finite volume surrounding each raypath, the individual SRFs are mapped to depth along the 3D volume to account for the spatial sensitivity of the propagating energy. At any particular depth the amplitude is spread over the first Fresnel zone using a 2D tapered function having unit amplitude at the conversion point and decreasing linearly to zero at the edge of the Fresnel zone function. The construction of these Fresnel zone sensitivities naturally defines a minimum horizontal wavelength of observation, ~ 200 km, that scales to the 10 s central frequency of the parent S and daughter S_p waves. Since SRFs are made with lower frequency waves (compared to P-wave receiver functions) and have more oblique incidence angles, calculation of a CCP stack substantially reduces the spatial averaging that occurs in the receiver gather approach. No smoothing of the CCP stacks other than the Fresnel zone weighting was applied. For a more comprehensive description of the CCP imaging method utilized, see *Levander and Miller* [2012] and *Thurner et al.* [2014].

3. Results

Upon quality control for candidate SRFs, the initial data set was culled to ~ 5300 receiver functions, producing a well-sampled volume that extends from -20°W , 27°N to 42°N , 2°E . Supplementary Figure S1 shows a hit count map for the model volume. Individual receiver functions are shown in Figure 2 for a series of stations from the Betics (XB.PS04), Rif (XB.PM11), Middle Atlas (XB.PM24), High Plateaux (XB.PM30), Anti-Atlas (3D.MM07), and

178 Canary Islands (ES.CRAJ). Individual receiver functions computed at the station are sorted
179 and colored by backazimuth, as represented in the rose diagram inset in Figure 2. The
180 resulting stack (a direct, non-weighted summation) is shown to the right as a thin black line
181 with peak amplitudes filled with red and/or blue to match the coloring convention
182 presented in the CCP stacks. We compute the standard deviation of individual station
183 receiver gathers (light-green area) using a bootstrap algorithm that is applied 1000 times
184 to produce 100 new stacks from which we compute an estimate of the standard deviation
185 of the original stack according to the number of individual receiver functions (n). The
186 shallowest positive (red fill) amplitude phase (always above 50 km depth) is
187 representative of the Moho in each of the stacks. As discussed in more detail below,
188 negative-amplitude phases (blue fill) at greater depths are often interpreted as the
189 lithosphere-asthenosphere boundary (LAB), and in some cases there is a deeper (110-175
190 km) positive and/or negative (150-220 km) conversion that are interpreted as additional
191 velocity interfaces.

192 SRFs processed at 10s period are not able to resolve internal crustal structure nor fine
193 details in crustal thickness as commonly observed from P receiver functions (PRFs) that
194 are more commonly computed at 0.5-1s [e.g. *Levander and Miller, 2012; Thurner et al.,*
195 *2014*]. They provide a broad estimate of the Moho topography while effectively imaging
196 the lithosphere-asthenosphere boundary and deeper structure. Station XB.PM11 in the Rif
197 (Figure 2C) shows a thickened Moho beneath the Rif, similar to previous results [*Dündar et*
198 *al., 2011; Mancilla et al., 2012; Thurner et al., 2014*]. Beneath the Atlas (Figure 2D) the
199 Moho is approximately ~35-40 km deep, consistent with previous results [*Sandvol et al.,*
200 *1998; Miller and Becker, 2014; Spieker et al., 2014; Thurner et al., 2014*]. Beneath the

201 Canary Islands (Figure 2A) the crust is the thinnest (<25 km), which is also similar to
202 previous PRF results that were able to image detailed crustal boundary layers [*van der*
203 *Meijde et al.*, 2003; *Lodge et al.*, 2012; *Martinez-Arevalo et al.*, 2013].

204 The station beneath the Betics (XB.PS04, Figure 2B) illustrates the presence of two negative
205 conversions at ~80 and ~135 km depth. This is also seen in cross section B-B' through the
206 CCP volume in Figure 3 that is perpendicular to the strike of the Rif and Betics, where the
207 labeled LAB and sLAB negative conversions almost blur together and in C-C' shown in
208 Figure 4. Figure 4 illustrates this deeper negative conversion beneath southern Spain and
209 the northern portion of the Alboran Sea with a cross section that runs E-W through
210 Gibraltar. This deeper negative conversion is interpreted to be the base of the subducted
211 slab beneath southern Spain, which correlates with the position of the slab from P-wave
212 tomography with fast seismic velocity perturbations [*Bezada et al.*, 2013].

213 Similar to previous SRF receiver gather results that estimated the depth to the LAB beneath
214 the Middle Atlas [*Miller and Becker*, 2014], the LAB is shallow (~65 km) and deepens on
215 either side of the mountain range (Figure 2D). To the southeast of the Atlas, the
216 lithosphere beneath the High Plateaux is >125 km thick, and the LAB is similarly thick
217 beneath the Moroccan Mesetas to the northwest of the high topography. Beneath southern
218 Morocco in both the Anti-Atlas and the High Plateaux the SRFs have a negative conversion
219 near 150 km depth with a deeper negative conversion at 220 km (Figures 2E&F and 5). We
220 interpret the shallower negative signal to be a mid-lithosphere discontinuity (MLD), similar
221 to what was first described from S receiver functions from the middle of North America by
222 *Abt et al.* [2010] and subsequently described as a global feature of cratons [*Cooper and*

Miller, 2014; Selway et al., 2015]. Signals from the MLD have previously been observed in the West African craton in Morocco, but also across its entirety to its southern boundary [Cooper and Miller, 2014]. The deeper negative conversion is likely to be the base of the cratonic lithosphere.

The discontinuity structure beneath the Canary Islands is unique in comparison to that observed on the African continent, due to their location on the oceanic lithosphere. The inferred LAB depth is shallow (~80 km) and there is a positive conversion beneath the western Canary Islands that located at approximately 100-125 km depth (Figure 2A, 5 and 6). The second, deeper positive phase caused by a low velocity structure does not extend along the entire archipelago, but is centered beneath the western islands.

Similar to the signal produced by the base of a low velocity discontinuity beneath the Canary Islands, there is another deep positive conversion that lies between ~110-160 km depth beneath parts of the Anti-Atlas, Atlas, and Alboran Sea (Figure 6 labeled as LVZ). It is not a continuous discontinuity across the entire CCP volume, but appears to be localized beneath these regions and beneath the Canary Islands. This positive signal must be from a conversion at the base of a sub-lithospheric low velocity zone.

4. Discussion

The variable lithospheric thickness beneath Morocco and Southern Spain was initially established through geophysical modeling of topography, heat flow, gravity and geoid anomalies [e.g. Teixell et al., 2005; Zeyen et al., 2005; Missenard et al., 2006]. One of the striking features of these results was the very thin lithosphere (~65 km) beneath the High and Middle Atlas, where there is significant (~4000 m) topography and recent alkaline

volcanism, and a lack of a deep, crustal root. Receiver functions have confirmed these results both using receiver gather analysis [Sandvol *et al.*, 1998; Miller and Becker, 2014] and CCP stacks of P receiver functions [Thurner *et al.*, 2014]. Here the SRF CCP volume agrees with these prior results and illustrates the extent of the shallow LAB along the axis of the Atlas (Figures 2 and 6). Figure 7 summarizes the various negative conversions fields observed in our model. The lithosphere-asthenosphere boundary (LAB) is auto-picked at the maximum negative amplitude for each depth column of the CCP volume. In some regions there are deeper negative conversions that require hand-picking because they are often lower amplitude than shallower discontinuities. For example, the shallow LAB observed across much of the region is distinct from secondary deeper conversions off the Alboran slab and from the northern edge of the West African Craton, shown in Figures 2F and Figure 5.

Beneath the southern part of the study region, the lithospheric structure is very different. Between 130 and 170 km depth a negative polarity conversion is present along with a deeper negative conversion at ~210 km (Figure 5). The shallower signal is interpreted to be a mid-lithospheric discontinuity (MLD) that is similar to signals identified with SRFs in the North America craton [Abt *et al.*, 2010; Miller and Eaton, 2010], cratonic Australia [Ford *et al.*, 2010], South Africa [Wittlinger and Farra, 2007; Hansen *et al.*, 2009], and the central and southern portions of the West African craton [Cooper and Miller, 2014]. The origin of the MLD is still a debated issue [Selway *et al.*, 2015], but it is becoming increasingly recognized as a feature common to Paleoproterozoic and Archean cratons and can be identified with SRFs. The presence of the MLDs with the deeper LAB signal in Figure 5 indicates that the West African craton extends as far north as 29°N at 11°W, which is

consistent with estimates from the 3SMAC model of *Nataf and Ricard* [1996] and from geodynamic modeling of mantle flow in the region [*Alpert et al.*, 2013].

Due to the numerous seismic stations on the Canary Islands as well as the oblique incident angle of incoming S waves the SRFs are able to image both the upper mantle structure beneath the islands themselves, and to a lesser extent, the region between the islands and the coast (Figure SUP1). The CCP volume indicates a slightly shallower lithosphere-asthenosphere boundary beneath the western islands and then remains nearly constant towards the continental margin. Petrological studies of ultramafic xenoliths from the eastern Canary Islands indicate a thinner lithosphere than beneath the western islands [*Neumann et al.*, 1995], but recent P receiver function studies do not image this thinning towards the east [*Martinez-Arevalo et al.*, 2013].

In addition to the negative velocity conversion that we identify as the LAB, there are spatially variable discontinuity structures beneath the generally thin lithosphere. The SRFs detect an upper mantle positive velocity conversion that we interpret as the base of low velocity structures that span from the Canary Islands to southeastern Spain (Figure 8). Slow upper mantle velocities in the region have been observed in previous studies. Larger scale images from tomographic methods [*Bezada et al.*, 2014; *Palomeras et al.*, 2014], previous regional receiver function studies of the Anti-Atlas [*Spieker et al.*, 2014], SRF receiver gather transect across the Middle and High Atlas [*Miller and Becker*, 2014], and P receiver functions imaging beneath the Canary Islands [*Lodge et al.*, 2012; *Martinez-Arevalo et al.*, 2013] all indicate a broader presence of low velocity upper mantle. The new SRF CCP images imply localized conversions from low velocity zones beneath the Atlas, northern-

most Anti-Atlas, Alboran Sea, and the western Canary Islands (Figure 8) at depths ranging from 110-160 km. However, due to the nature of receiver function imaging, both in this work and generally, these conversions are only produced by nearly horizontal structures.

High resolution seismic imaging from local and teleseismic waveforms also resolve a low velocity zone beneath the Middle and High Atlas that is coincident with the recent volcanism [Sun *et al.*, 2014]. The edges and depth extent of this low velocity structure are constrained by waveform modeling and geodynamic modeling of a hot, buoyant source that extends down to 200 km [Sun *et al.*, 2014]. The sub-horizontal structural extent originally observed with SRFs [Miller and Becker, 2014] was imaged with more detail and the presence of a steeply dipping conduit was also defined by this method. This conduit of low velocity material was robustly detected along a 15 station linear array that was perpendicular to the strike of the Atlas, but the along-strike constraint on its geometry is not clear from this study. Our development of a 3D receiver function CCP volume using 360 stations (Figure 1A) allows for a broader view of the sub-horizontal component of this particular low velocity feature and possible relationship to adjacent sub-lithospheric low velocity zones.

The presence of the sub-horizontal extent of the inferred low velocities spatially correlate with the locations of recent (Quaternary to Pliocene) volcanic activity (Figure 8). Any near vertical or steeply dipping structures, such as that imaged by Sun *et al.* [2014] beneath the Middle Atlas, would not be possible with SRF methodology, however, SRF studies of other regions have observed conversions from low velocity structures beneath active regions of active or recent volcanism [Vinnik *et al.*, 2004; Vinnik *et al.*, 2005; Rychert *et al.*, 2012;

312 *Rychert et al., 2013; Hopper et al., 2014; Rychert et al., 2014*]. Beneath continental
313 lithosphere of the Afar region, low velocities have been interpreted as the onset of melting
314 [*Vinnik et al., 2004; Rychert et al., 2012*] or more broadly as possibly due to a change in
315 volatile content, melt traction, or grain size. Similar observations have been made beneath
316 the High Lava Plains of the western United States [*Hopper et al., 2014*]. Beneath oceanic
317 lithosphere and active hotspot volcanoes of the Galapagos [*Rychert et al., 2014*], Iceland
318 [*Vinnik et al., 2005*] and Hawaii, the low velocity zones inferred from receiver functions
319 have been interpreted to be from the onset of melting, but appear to offset from active
320 magmatic centers [*Rychert et al., 2013*]. The depth of the low velocity zones beneath all of
321 these other active volcanic regions are between ~110-150 km, similar to our observations
322 beneath the Canary Islands, the “Moroccan Hot Line”, and southern Spain.

323 *Duggan et al. [2009]* jointly interpreted geochemical composition of lavas and existing
324 geophysical models to link the Canary Island hotspot volcanoes and the intraplate
325 volcanism in northwest Africa. The presence of a basal lithospheric corridor beneath the
326 Atlas was proposed to allow mantle plume material to flow laterally from the hotspot
327 beneath the African continent. The surface manifestation of this flow is posited to be a trail
328 of mafic lavas derived from low-degree partial melting of a sub-lithospheric mantle. This
329 model has three components: deflection of the mantle plume along the base of the
330 continental lithosphere; delamination of the mantle lithosphere beneath the Atlas; and
331 suction of plume material towards southern Spain due to rollback of the subducted slab at
332 the Gibraltar arc [*Duggan et al., 2009*]. Analogue modeling of slab-plume interaction
333 applied to the Gibraltar slab and Canary Island setting suggests that the rollback induced

334 mantle flow could have dragged plume material towards the Atlas corridor [*Mériaux et al.*,
335 2015].

336 Geodynamic modeling of mantle flow patterns driven by mantle density anomalies,
337 including the position of the slab and deep keel beneath the West African craton, is
338 consistent with the concept of guided flow of mantle below the north of Morocco [*Alpert et*
339 *al.*, 2013]. Observed seismic anisotropy from shear wave splitting analysis [*Diaz et al.*,
340 2010; *Miller et al.*, 2013; *Díaz and Gallart*, 2014] largely motivated the *Alpert et al.* (2013)
341 model, but does not support a continuous basal lithospheric channel between the Atlas
342 Mountains and Canary Islands. Averaged splitting measurements are oriented parallel to
343 the strike of the Atlas and show large delay times ($> 1.5\text{s}$), but the amplitude of the delay
344 times decrease ($< 0.5\text{s}$) significantly at stations in the westernmost Atlas and near the
345 Atlantic coast. This change in the anisotropy field and implied mantle flow direction does
346 not substantiate the model for a continuous channel directly from the Canary Islands, along
347 the thinned lithosphere beneath the Atlas, and towards the Alboran.

348 The SRF CCP volume presented here reveals the presence of multiple, localized sub-
349 lithospheric low velocity zones beneath the Canary Islands, Atlas, and the Alboran, rather
350 than one continuous, subhorizontal channel of low velocity material that extends from the
351 hotspot to southern Spain. We also do not image a continuous, localized thin lithosphere
352 along the “Morrocan Hot Line” coincident with the posited basal lithospheric channel of
353 *Duggen et al.* [2009]. Therefore, we offer a modification of the sub-lithospheric flow model
354 to explain the recent intracontinental volcanism. Many of the geodynamic models that link
355 the Canary Island plume and volcanism in northwest Africa [*Duggen et al.*, 2009; *Alpert et*

al., 2013; *Mériaux et al.*, 2015] imply that the plume has been deflected by the induced mantle flow from the sinking Alboran slab beneath the Gibraltar arc. This is in agreement with numerical models of large scale convection [*Steinberger and O'Connell*, 1998], and leads to the notion that the plume may be tilted away from its deep source from vertical to nearly horizontal in the upper mantle. Laboratory experiments demonstrate that when compositional plumes are subjected to shearing, small-scale gravitational instabilities are generated when the plumes were tilting towards the horizontal [*Kerr et al.*, 2008; *Mériaux et al.*, 2011 and references within]. Therefore, the localized small volcanoes distributed between the northern Anti-Atlas, Middle and High Atlas, eastern Rif, and eastern Betics may be the result of instabilities sourced from a highly tilted, nearly horizontal plume in the upper mantle as detected by the CCP images that is bounded in the south by the West African craton (Figure 8 inset), and is near vertical beneath the Canary Islands [*Saki et al.*, 2015].

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606

607 **Figure Captions**

608 **Figure 1:** A) Generalized map of the study region with plate convergence from Rosenbaum
 609 et al. (2008) shown with a black vector, and the political boundaries are indicated with thin
 610 black lines. B) Map of the broadband stations distribution, where the yellow triangles are
 611 the PICASSO stations, the blue triangles are IberArray stations, and the green triangles are
 612 from the Bristol-Münster array. The locations of the cross-sections and fence diagrams for

613 Figures 3-6 are shown with dark red lines. The globe inset shows the 180 earthquakes (in
614 orange) used in the analysis.

615 **Figure 2:** Individual (number listed as 'n') and stacked receiver gathers for a selection of
616 stations in different tectonic domains. A) ES.CRAJ on Tenerife in the Canary Islands; B)
617 example of station XB.PS11 in the Betics; C) XB.PM08 in the Rif, D) Example of station
618 XB.PM24 in the Middle Atlas; E) XB.PM30 in the High Atlas; and F) 3D.MM07 in the Anti
619 Atlas. To the right of each subfigure is the simple summation stack of all the receiver
620 functions (thin black line) for that station along with the standard deviation of the
621 individual stacks (light green outlines), which were computed using a bootstrap algorithm.
622 The positive amplitude Moho signal is marked with an "M", the negative amplitude signal
623 interpreted as the lithosphere-asthenosphere boundary (LAB) is marked with an "L", the
624 cratonic LAB is marked with "cL", and the "sL" indicates the interpreted base of the
625 subducted slab LAB. "MLD" marks a possible signal from a mid-lithospheric discontinuity
626 and "LVZ" marks the positive conversion from the base of the low velocity anomaly. A rose
627 diagram in upper right shows the distribution of events used in the analysis colored by
628 back-azimuth as are the individual receiver functions for each subfigure.

629 **Figure 3:** N-S profile through the S receiver function CCP volume across the Alboran Sea,
630 the Atlas, to the Moroccan High Plateaux along B-B' as shown in Figure 1B and Figure 8.
631 The red colors indicate positive conversion of the Sp phase from fast to slow (upward
632 traveling wave) and the blue colors indicate a negative conversion of the Sp phase from
633 slow to fast (upward traveling wave). Labels for the Moho, lithosphere-asthenosphere
634 boundary (LAB), base of the subducted slab (sLAB), and positive signal from a low velocity

zone (LVZ) along with dashed lines marking the largest amplitude in the SRF, indicating the interpreted boundaries.

Figure 4: E-W profile through the S receiver function CCP volume through Gibraltar, parallel to the coastline, along C-C' shown in Figures 1B and Figure 8. The red colors indicate positive conversion of the Sp phase from fast to slow (upward traveling wave) and the blue colors indicate a negative conversion of the Sp phase from slow to fast (upward traveling wave). Labels for the Moho, lithosphere-asthenosphere boundary (LAB), base of the subducted slab (sLAB), and positive signal from a low velocity zone (LVZ) along with dashed lines marking the largest amplitude in the SRF, indicating the interpreted boundaries.

Figure 5: E-W profile the S receiver function CCP volume through the northernmost extent of the West African craton in southern Morocco and the Anti-Atlas along D-D' shown in Figures 1B and Figure 8. The red colors indicate positive conversion of the Sp phase from fast to slow (upward traveling wave) and the blue colors indicate a negative conversion of the Sp phase from slow to fast (upward traveling wave). Labels for the Moho, lithosphere-asthenosphere boundary (LAB), mid-lithospheric discontinuity (MLD) and base of the cratonic lithosphere (cLAB), and positive signal from a low velocity zone (LVZ) along with dashed lines marking the largest amplitude in the SRF, indicating the interpreted boundaries.

Figure 6: Fence diagram through the CCP volume extending from the Canary Islands to southern Spain parallel to the strike of the Atlas Mountains to illustrate the extent of the

low velocity anomaly structure along A-A' as shown in Figures 1B and Figure 8. Labels for the Moho, lithosphere-asthenosphere boundary (LAB), base of the subducted slab (sLAB), and positive signal from a low velocity zone (LVZ) along with dashed lines marking the largest amplitude in the SRF, indicating the interpreted boundaries.

Figure 7: Fields of negative conversion depths hand-picked from the CCP volume. Three surfaces represent the West African craton LAB (Blue, > 200 km), the oceanic and continental LAB (red-orange, < 100 km), and the Alboran slab LAB (orange-yellow, 120-130 km). Yellow dashed line indicates western boundary of thin lithosphere that underlies Morocco.

Figure 8: Interpretive depiction of upper mantle structure along with a map of the lateral extent of the sub-lithospheric low velocity zone pockets (in pink), locations of Quaternary and Pliocene volcanic outcrops (from [Duggen et al., 2004; Frizon de Lamotte et al., 2009; Fullea et al., 2010]). (black) Outline of West African craton extended from observations of LAB in Figure 7. (blue) Location of Alboran slab in upper 200 km of mantle, taken from Bezada et al. [2013]. (Pink) Positive velocity conversions interpreted to be the base of pockets of slow velocity bodies in upper mantle, identified as LVZ in Figures 3, 5, & 6. Orange polygon represents low velocity body identified by waveform modeling of Sun et al. [2014]. Yellow dashed line denotes western edge of thin lithosphere below Morocco, based on LAB depth field in Figure 7. White arrows represent mantle flow field induced by regional flow [Conrad and Behn, 2010] and slab-induced flow [Alpert et al., 2013]. (inset) Cross section along X-X' from Canary Islands to eastern Alboran sea showing our interpretation of slab-plume and lithosphere interaction. The Canary plume is deflected

679 eastward by mantle flow, giving rise to smaller-scale upwellings, such as the Atlas
680 upwelling [*Sun et al.*, 2014]. These trapped pockets of low velocity bodies are inferred to
681 give rise to localized basaltic volcanism that has been sourced from sub-lithospheric depths.
682 Dotted lines schematically indicate the LAB and the 410 discontinuity interpreted from the
683 CCP imaging results and those of Saki et al. [2015].

Figure 1a

Figure 1b

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Graphical Abstract