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Seismic Array Study of the Western Mediterranean and the United States Great Plains: Insight into the Modification and Evolution of Continental Lithosphere

by

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ABSTRACT

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Numerous tectonic processes are responsible for the modification and evolution of continental lithosphere. The continents, however, are generally resilient through geologic time and keep a record of Earth’s tectonic activity, both past and present. The focus of this work is to better understand the modification and evolution of continental lithosphere associated with continent-continent collisions. We do this by studying two orogenic systems: the Alpine Orogeny, associated with the ongoing collision between the African and Eurasian plates, and the Trans-Hudson Orogeny, associated with the initial formation of the North American craton during the Precambrian. This research focuses on the westernmost edge of the Alpine system in the western Mediterranean, where subduction and slab rollback have caused significant extension and Africa-Iberia convergence has caused simultaneous contraction. Here we calculate Pds receiver functions to constrain the discontinuity structure. Additionally, we jointly invert Pds receiver functions and Rayleigh wave phase velocity dispersion data to create a 3-D shear velocity model.
These results show a deep Moho around the western portion of the Gibraltar Arc. Below this deep Moho we see the Alboran Slab extending down to ~250 km. In the eastern Gibraltar Arc, there is a very shallow Moho where the slab has detached from the surface and removed continental lithosphere. In the Trans-Hudson Orogen we use receiver functions and gravity data to determine the discontinuity and density structure of the shallow lithosphere. This analysis reveals crustal-scale thrusting associated with the Wyoming-Superior suture zone. We also find a relatively low Moho density contrast throughout the Trans-Hudson and northern Yavapai Province. This low Moho density contrast is associated with a deep Moho (>50 km) and is interpreted to be evidence of a dense lower crustal layer resulting from mafic underplating. Finally, we investigate the contribution that this dense thick crust may have played in the isostatic stabilization of the North American craton as well as other cratons around the world. We find that the lithospheric mantle must provide a negative component to cratonic lithospheric buoyancy in order to account for the low elevations observed along with thick crust in the cratons.
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Chapter 1

Introduction

The Wilson cycle describes the periodic opening and closing of ocean basins, where dense oceanic lithosphere is continually created and destroyed while buoyant continental lithosphere is left preserved through geologic time (Wilson, 1965). Continental lithosphere, however, is still affected by Earth’s continuous tectonic activity. Numerous tectonic processes are responsible for the modification, destruction, and evolution of continental lithosphere and the crust-mantle boundary within it. These processes include shearing and thickening associated with collision, extension and thinning associated with rifting and orogenic collapse, melting and crustal growth associated with magmatism, and removal and recycling due to secondary convection processes. The continents, however, are resilient, preserving a record of these processes both past and present and providing great insight into the modern and ancient tectonics responsible for the formation and evolution of Earth’s lithosphere.
In this thesis, I present the results of research motivated by improving our understanding of continental lithospheric modification and evolution in orogenic systems associated with continent-continent collisions. Such collisions can involve a variety of the tectonic processes mentioned above, operating at various stages of collision. This research is focused on two major orogenic systems: the Alpine Orogeny and the Trans-Hudson Orogeny, both of which have resulted from continent-continent collision.

The Alpine orogenic system, extending from the western Mediterranean to the Zagros Mountains in Iran, marks the collision zone between the African and Eurasian plates, which were separated by the Tethys Ocean throughout the Mesozoic (Stow 2010). The Alpine system resulted from closure of this ocean, which initiated with subduction of the intervening oceanic crust and culminated with the collision between the African and the western part of the Eurasian plates (Coward and Dietrich, 1989). The research presented in this thesis focuses on the westernmost portion of the Alpine orogenic system in the western Mediterranean (Figure 1.1), where Africa-Iberia convergence began ~50 Ma. During this time, the western portion of the Tethys Ocean still separated Africa and Iberia. Around 35 Ma a northwest dipping subduction system developed at the southern margin of western Eurasia (Rosenbaum, 2000). As a result of subducting old (>110 Ma), dense Tethyian oceanic lithosphere and/or the slowing of Africa-Eurasia convergence, slab rollback began between ~30-25 Ma (Rosenbaum, 2000). Rollback initiated widespread extension in the overriding continental lithosphere and eventually led to the break up and drifting of extended continental lithospheric terranes across the
western Mediterranean (Rosenbaum, 2000). The Betic and Rif Mountains comprise the Gibraltar arc orogenic system, which formed when a portion of this extended continental terrain, known as the Alboran Domain, was thrust onto the passive margins of Iberia and Africa.

![Figure 1.1](image)

**Figure 1.1** – Map showing the extent of the Alpine orogenic system. The red square outlines the western Mediterranean study region discussed in this thesis.

Although uplift of the Betic-Rif orogen continues, accretion of the Alboran Domain was complete by ~10 Ma (Rosenbaum et al., 2000; Duggen et al., 2005). Africa and Iberia continue to converge at a relatively slow rate of ~4-6 mm/yr in a NNW-SSE direction (Stich et al., 2006; Perouse et al., 2010). A portion of this convergence (~2 mm/yr) is accommodated by the Atlas Mountains, which run NE-SW from the Rif Mountains to the eastern margin of Morocco.
The complexity of the western Mediterranean tectonic history has generated debate regarding the coeval extensional and compressional processes involved in the subduction and rollback of oceanic lithosphere, widespread extension in the overriding continental lithosphere, the transport and distribution of continental terrains, and the uplift of the Betic, Rif, and Atlas Mountains. Numerous tectonic models, involving slab rollback (Royden et al., 1993; Lonergan and White, 1997; Rosenbaum et al., 2000), slab break-off (Zeck, 1997), and lithospheric removal (Platt and Vissers, 1989; Seber et al., 1996; Platt et al., 2003), have been developed; although, no consensus has been reached. Recently, a number of broadband seismic networks, including the PICASSO (Program to Investigate Convective Alboran Sea System Overturn) network and the IberArray/TopoIberia network, have been deployed with a view toward better understanding the complex geodynamics of the Alboran system. In this thesis I present research results from two complementary seismic techniques applied to data from these recently deployed seismic stations.

The Trans-Hudson Orogen is a Paleoproterozoic collisional belt that extends south from the Hudson Bay through Canada and into the central U.S. (Figure 1.2). This collisional belt resulted from the closure of the Manikewan Ocean during the formation of the Archean core of Laurentia (North American craton) between ~2.0 and 1.8 Ga (Hammer et al., 2010; Whitmeyer and Karlstrom, 2007). The majority of this Archean core is located in Canada, although some portions extend into Montana, Wyoming, North Dakota and South Dakota. The northern portion of the Trans-Hudson Orogen formed when the Archean Hearne Province in the west collided with the Archean Superior Province in the east, while the southern portion formed
from the collision between the Archean Wyoming Province in the west and the Superior Province in the east. Ocean closure occurred from north to south, with the Wyoming – Superior collision trailing the Hearne – Superior collision by ~ 50-60 Ma (Dahl, 1999).

Figure 1.2 – Map showing the extent of the Trans-Hudson Orogen (THO) with the background outline of the North American states and provinces. The THO is indicated by the purple shaded section extending from the Hudson Bay to South Dakota. The grey shaded regions indicate Archean Provinces. The dark green, blue, and light green NE trending regions are the arc terrains that were accreted onto the southern margin of Laurentia during the Proterozoic. (Modified from Whitmeyer and Karlstrom, 2007)
The remaining portion of Laurentia is composed of several NE-SW trending Proterozoic provinces (Yavapai, Mazatzal, Granite/Rhyolite, Grenville) formed during a series of island arc accretion events that occurred between ~1.8 and .95 Ga (Whitmeyer and Karlstrom, 2007). These accretion events are associated with the intrusion of granitoids, which is believed to have helped “stitch” together these distinct provinces (Whitmeyer and Karlstrom, 2007). It is also believed that a major magmatic phase occurred ~1.4 Ga during the process of crustal accretion. Keller et al. (2005) suggest that this magmatic phase caused thickening of the crust (up to ~45 km) through the production of a mafic restite layer. The break-up of Rodinia began ~700 Ma and was marked by widespread normal faulting, dike intrusion, and failed rifting within Laurentia. It is thought, however, that magmatism and extension during this time was primarily concentrated in the Proterozoic terrains and had little effect on the Archean lithosphere (Karlstrom and Humphreys, 1998). In general, the Trans-Hudson Orogen and the adjacent Archean (Hearne, Wyoming, Superior) and Proterozoic (Yavapai) terranes have been unaffected by subsequent tectonism since the Precambrian.

The Canadian portion of the Trans-Hudson Orogen was studied during the Lithoprobe program, which consisted of numerous seismic reflection and wide-angle seismic refraction surveys throughout Canada (Hajnal, 1997; Nemeth and Hajnal, 1998; Corrigan et al., 2005; Nemeth et al., 2005; White et al., 2005). The U.S. component of the orogen was investigated with a series of COCORP seismic reflection profiles crossing the Trans-Hudson Orogen just south of the U.S.–Canada
border (Latham et al., 1988; Baird et al., 1996; Nelson et al., 1993; Klasner and King, 1990). A more extensive seismic investigation of the entire U.S. Trans-Hudson Orogen has been made possible by the USArray Transportable Array, which consists of a regular grid of ~400 temporary broadband seismic stations installed with ~70 km station spacing in a sweeping deployment from west to east across the contiguous United States. In this thesis, I present research results from a geophysical investigation of the Trans-Hudson Orogen and surrounding region, utilizing data from the USArray and four other seismic networks.

Together, the Alpine Orogeny and Trans-Hudson Orogeny provide an excellent opportunity to study both modern and Precambrian collisional tectonics. Additionally, they provide insight into the continental lithospheric modification processes associated with the various stages of collision: from the final stages of oceanic subduction and the transition to continent-continent collision to post-collisional evolution. The research presented in this thesis utilizes seismic, gravity, and geochemical data to characterize the continental lithospheric structure in both regions as well as provide geophysical constraints on the tectonic processes responsible for its modification.

The second chapter of this thesis describes the analysis of teleseismic Ps receiver functions, thermobarometric measurements of post-Oligocene basalts, and recent body wave and surface wave tomography results from the western Mediterranean. Receiver functions were calculated for 239 broadband seismic stations located in southern Iberia and northern Morocco between 33.5° and 40° North and -9° and 0° West. The thermobarometric analysis was performed on 19
volcanic samples distributed throughout the region. The receiver function images indicate a highly variable Moho depth, ranging from ~25 to ~55 km around the Gibraltar Arc. Additionally, they show a strong positive, sub-Moho horizon between ~45 and ~80 km depth beneath the central Betic and Rif Mountains. This horizon is interpreted as the top of the Alboran Slab. Thermobarometric constraints indicate mantle melting depths between 40 and 60 km in the eastern Betic and Rif Mountains where the crust is significantly thinned, suggesting little to no lithosphere exists in these regions. The receiver function and thermobarometric data, interpreted in conjunction with previous teleseismic tomography images, suggest ongoing and recent slab detachment and continental lithospheric delamination initiated by slab rollback and vertical slab-pull forces acting on the base of the Gibraltar Arc continental lithosphere.

In the third chapter, I discuss the results from the joint inversion of Ps receiver functions with ambient noise and ballistic Rayleigh wave phase velocity data. In this study a 3-D shear velocity model was created using 242 broadband seismic stations deployed from central Spain, across the Gibraltar Arc, and Atlas Mountains, and into the Sahara Desert. The joint inversion of these datasets provides better constraints on the Moho depth, the shallow slab structure, and the extent of lithospheric delamination described in Chapter 2 and in previous tomography and geochemical studies (Duggen et al., 2004; Bezada et al., 2013; Palomeras et al., 2014). The resulting shear velocity model indicates a deep Moho in the western Gibraltar Arc. Below this, the high velocity Alboran Slab extends from the base of the crust to > 250 km depth. The region with a deep Moho coincides with
where the slab remains attached to the base of the Gibraltar Arc lithosphere. Where the Gibraltar Arc crust abruptly thins towards the east, the Alboran Slab appears to descend into the mantle, removing continental lithosphere in the process. The Sierra Nevada represent the highest elevations within the Betic Mountains, despite being underlain by a relatively shallow Moho (~35 km). Their location coincides with where the slab has become detached from the base of the Gibraltar Arc lithosphere and asthenosphere has upwelled to fill the intervening space, evidenced by low shear velocities in this region. These results suggest that the uplift of the Sierra Nevada was an isostatic response to the emplacement of buoyant asthenosphere directly beneath the crust. The joint inversion data suggest a similar story in the Atlas Mountains, a portion of which is underlain by a relatively thin crust (~33-38 km) and lithosphere (~40-60 km). Beneath this thin lithosphere are low shear velocities of ~4.1-4.2 km/s, indicative of buoyant asthenosphere which may support much of the topography in the eastern Middle and High Atlas.

In the fourth chapter of this thesis I discuss the results of a Ps teleseismic receiver function analysis of the Trans-Hudson Orogen and the surrounding Great Plains/Midcontinent region. In this study 0.5 Hz receiver functions were calculated for over 800 broadband seismic stations. These receiver functions were used to investigate the crustal structure associated with the Wyoming-Superior suturing event that produced the Trans-Hudson Orogen as well as the crustal structure associated with the accretion events that occurred along the southeastern margin of North America between ~1.71 and ~1.68 Ga. These results show evidence of crustal scale thrusting of the Wyoming province in the west over the Superior province in
the east. Additionally, they reveal a relic subduction zone associated with the Yavapai-Superior boundary. The receiver functions were also used to conduct a density analysis of the region using the 2p1s and 0p1s receiver function phases at 233 stations. These results reveal a relatively low Moho density contrast throughout the Trans-Hudson Orogen and northern Yavapai Province. The results of gravity modeling along latitude profiles across the Trans-Hudson Orogen are consistent with this low Moho density contrast. The region characterized by a dense lower crust coincides with a region of thickened crust (>50 km) and is interpreted to be evidence of mafic underplating.

In chapter five I extend my discussion of continental lithospheric modification to the cratons, into the realm of post collisional evolution. Here I present the results of an analysis utilizing the Crust1.0 global crustal model (Laske et al., 2013), which provides global crustal thickness and velocity data on a 1° x 1° grid. This model is used to characterize the crust and lithospheric structure of cratons globally with the goal of providing insight into the evolution of continental lithosphere. This study is motivated by the thick, dense crust observed in the Trans-Hudson Orogen, described in Chapter 4. The Crust1.0 model allows for the global characterization of cratons as regions with thick crust and low elevations. In this study, I investigate the role that a dense, thick crust may have played in the isostatic stabilization of the cratons. A plot of crustal thickness vs. elevation for cratons worldwide shows that these regions do not behave according to the Airy or Pratt isostatic hypotheses. That is, the observed elevations cannot be the result of crustal thickness and/or density variations alone. The continental mantle lithosphere,
therefore, must contribute to the low elevations observed in cratons. This study uses the discrepancy between observed elevation and the calculated crustal component of elevation, as well as cratonic lithospheric thickness estimates from previous studies, to determine the density of the continental mantle lithosphere in various cratons. These results show that cratonic lithospheric mantle is not necessarily neutrally buoyant, but rather ~1-2% denser than the underlying asthenosphere. Further work can be done to assess the relative thermal and compositional contributions to this density in order to provide insight into the long-term stability of continental lithosphere.

Chapters two, three, and four of this thesis consist of three reformatted papers that have been published or submitted during my thesis work. The receiver functions in the western Mediterranean (Chapter 2) are based on a paper published in Geochemistry, Geophysics, Geosystems (Thurner et al., 2014). The joint inversion study from the western Mediterranean (Chapter 3) has been submitted to Earth and Planetary Science Letters (EPSL). These chapters provide insight into the complex tectonic system of the western Mediterranean and the continental lithospheric delamination that appears to accompany the final stages of subduction. The Trans-Hudson Orogen study (Chapter 4) has also been submitted to EPSL, while Chapter five represents a working manuscript to be submitted in the future. These chapters provide insight into continent-continent suturing processes as well as continental evolution associated with craton formation and stabilization. Together, the chapters in this thesis improve our understanding of continental lithospheric modification processes that are ongoing and have been ongoing since the Precambrian.
Chapter 2

Ongoing Lithospheric Removal in the Western Mediterranean: Evidence from Ps Receiver Functions and Thermobarometry of Neogene Basalts (PICASSO Project)

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Abstract

The western Mediterranean tectonic system consists of the Betic Mountains in southern Spain and the Rif Mountains in northern Morocco curved around the
back-arc extensional Alboran basin. Multiple tectonic models have been developed to explain the coeval compressional and extensional tectonic processes that have affected the western Mediterranean since the Oligocene. In order to provide constraints on these evolutionary models, we use Ps teleseismic receiver functions (RF), thermobarometric analyses of post-Oligocene basalts, and previous teleseismic tomography images to investigate the lithospheric structure of the region. Ps RFs were calculated using seismic data from 239 broadband seismic stations in southern Iberia and northern Morocco and thermobarometric analysis was performed on 19 volcanic samples distributed throughout the region. The RF images reveal a highly variable Moho depth (~25 km to ~55 km), as well as a strong positive, sub-Moho horizon between ~45 and ~80 km depth beneath the central Betic and Rif Mountains, which we interpret to be the top of the previously imaged Alboran Sea slab. Thermobarometric constraints from magmas in the eastern Betics and Rif indicate mantle melting depths between 40 and 60 km, typical of melting depths beneath mid-oceanic ridges where little to no lithosphere exists. Together, the RF and thermobarometric data suggest ongoing and recent slab detachment resulting from delamination of the continental lithosphere.

2.1 Introduction

The Africa-Eurasia collision has resulted in a diffuse deformation zone within and around the Mediterranean Sea, where multiple narrow arcs and back-arc extensional basins have formed since the Oligocene [Rosenbaum, 2002]. The
western Mediterranean, comprising the westernmost range of the Alpine-Himalayan system, includes the Betic and Rif orogens curved around the Alboran back-arc extensional basin (Figure 2.1a). Beginning in the Miocene, the region extending from the Betic Mountains in southern Spain to the Atlas Mountains in Morocco, has been affected not only by collisional processes resulting from Africa-Eurasia convergence, but also by extensional processes initiated by slab rollback [Royden, 1993; Lonergan and White, 1997; Rosenbaum, 2002]. Multiple continental lithospheric removal events have also been proposed to explain the extension that has occurred throughout region [Platt and Vissers, 1989; Duggen et al., 2005]. Additionally, the western Mediterranean has been affected by widespread volcanism. The region is marked by a ~300 km northeast-southwest trending corridor of Miocene to Pleistocene igneous activity. This corridor extends from Spain through the Alboran Sea and into Morocco [Figure 2.1,a; Duggen et al., 2004]. The spatial and temporal variability of volcanic rocks found within this corridor suggests a transition from post-collisional subduction related magmatism to intraplate magmatism [Duggen et al., 2005].

Although multiple geodynamic models have been proposed to explain western Mediterranean tectonics and volcanic activity, a consensus has not been reached thus far. With a view toward better understanding the complex geodynamics of the Alboran system, several temporary and permanent broadband seismic arrays have been deployed. Together, the PICASSO array (Program to Investigate Convective Alboran Sea System Overturn), the IberArray, the Portuguese National Seismograph Network, the Spanish Digital Seismic Network, the University
of Lisbon Seismic Network, and the Western Mediterranean Seismic Network provide dense broadband station coverage of southern Iberia and Morocco (Figure 2.1b).

**Figure 2.1:** a) Map showing the elevation and major geologic features in the western Mediterranean. b) Map showing the locations of the 239 stations used in this study.

Using data from a total of 239 broadband seismic stations and 19 volcanic samples, we present a teleseismic Ps (P-wave to S-wave) receiver function (RF) and thermobarometric analysis of the western Mediterranean. The goal of this research
is to provide crustal and upper mantle constraints on the evolutionary tectonic models that have been proposed for this region. Our analysis, interpreted in conjunction with previous teleseismic tomography images, provides evidence for ongoing lithospheric delamination, initiated by slab rollback and vertical slab-pull forces acting on the base of the Gibraltar Arc continental lithosphere.

### 2.2 Tectonic Framework

The Gibraltar Arc, consisting of the Betic and Rif mountain systems, formed as the result of a series of tectonic events beginning with Africa-Iberia convergence ~50 Ma. At this time, units along the eastern Iberian margin were stacked to form the Alboran Domain with continental crust thickened to ~50-60 km [Vissers et al., 1995]. Around 35 Ma, a northwest dipping subduction system developed along the southeastern margin of Iberia [Rosenbaum, 2002] where rapid slab rollback to the southeast was initiated by the gravitational instability of old (> 110 Ma) subducting lithosphere and/or the slowing of Africa-Eurasia convergence [Rosenbaum, 2002]. Slab rollback resulted in significant extension within the Alboran Domain, the transport of the Alboran Domain southwestward across the Mediterranean, and the development of an eastward dipping subduction system between Iberia and Africa. Fragments of the westward migrating Alboran continental crust were eventually thrust onto the passive margins of Africa and Iberia, forming the Betic and Rif orogens [Rosenbaum, 2002]. These orogens now consist of External Domains, consisting of shortened passive continental margin rocks, and Internal Domains,
consisting of the previously deformed Paleozoic to Early Miocene Alboran rocks that were thrust onto the two continental margins [Figure 2.1a; Platt and Vissers, 1989; Rosenbaum, 2002; Platt et al., 2013]. Shortening occurred within the External zones, forming foreland fold and thrust belts, while extension, crustal thinning, and subsidence occurred within the Internal zones and the Alboran Basin [Platt and Vissers, 1989; Vissers et al., 1995; Lonergan and White, 1997; Torne et al., 2000; Faccenna et al., 2004]. Accretion of the Betic-Rif Internal zones was complete by ~10 Ma [Rosenbaum, 2002; Duggen et al., 2005]. Since ~7 Ma, however, the Betic-Rif orogen has been progressively uplifted [Vissers et al., 1995]. Duggen et al. [2005] hypothesized this to be the result of the removal of subcontinental lithosphere and its replacement by more buoyant asthenosphere.

The northeast-southwest trending Atlas mountain belt is the southern edge of the diffuse Africa-Eurasia plate boundary zone. The Middle and High Atlas Mountains have been uplifted since ~15 - 20 Ma [Ramadani et al., 1998; Missenard et al., 2006]. The Atlas Mountains now accommodate NNW-SSE Africa-Iberia convergence at a rate of ~2 mm/yr. [Missenard et al., 2006; Stich et al., 2006].

The widespread Cenozoic volcanic activity in the western Mediterranean provides constraints on the tectonic events described above. The Rif Mountains, Alboran Sea, and Betic Mountains have been affected by igneous activity of variable composition since the Middle Miocene. During the Middle to Upper Miocene subduction related tholeiitic and calc-alkaline volcanism was prevalent throughout the region [Duggen et al. 2005]. During the Upper Miocene to Lower Pliocene and Quaternary, igneous activity in southern Iberia and north Africa transitioned to
volcanism consisting of alkali basalts, suggesting delamination of continental
lithosphere and asthenospheric upwelling beneath southern Iberia and
northwestern Africa [Duggen et al., 2004, 2005]. The most recent magmatism in the
Atlas Mountains includes the eruption of Paleogene to Quaternary alkali basalts
within the Middle Atlas.

2.3 Tectonic Models

Numerous tectonic models, summarized by Calvert et al. [2000], have been
proposed to explain the western Mediterranean tectonic history. These models can
be grouped into three broad categories: 1) subduction and slab rollback with or
without slab break off [Royden, 1993; Lonergan and White, 1997], 2) lithospheric
convective removal or delamination [Platt and Vissers, 1989; Seber et al., 1996; Platt
et al., 2003] and 3) subduction and slab rollback accompanied by lithospheric
removal processes [Duggen et al., 2004, 2005; Palomeras et al., 2014].

The first group of models is supported by multiple tomography studies,
which have imaged the mantle structure beneath the Alboran Sea and Gibraltar Arc.
These studies revealed high velocities at depths between ~60 to 700 km, which
have been interpreted as an oceanic slab present beneath the Alboran Sea and
Gibraltar Arc [Blanco and Spakman, 1993; Zeck, 1996; Calvert et al., 2000; Gutscher
et al., 2002; Bezada et al., 2013; Palomeras et al., 2014]. The slab rollback
accompanied by slab break-off model is supported by the observation that high
velocity slab material beneath the Alboran Basin did not appear to extend to the
Slab detachment was used to explain the uplift of the Betic lithosphere as well as magmatism and anatexis within the Betic Mountains [Blanco and Spakman, 1993; Duggen et al., 2005]. The most recent P-wave and Rayleigh wave tomography results from this region, derived from PICASSO and IberArray data, support the slab interpretation [Bezada et al., 2013; Palomeras et al., 2014]. The evidence for complete slab break-off, however, appears weak as Bezada et al., [2013] image a high velocity anomaly beneath the Alboran Sea and southern Spain, extending continuously from ~50 km depth to the transition zone and Palomeras et al. [2014] use Rayleigh wave tomography data to show the same high velocity anomaly extending toward the surface.

The second group of models does not involve subduction, but rather removal of gravitationally unstable lithosphere from beneath the Alboran Basin and Gibraltar Arc by either delamination or convective processes [Seber et al., 1996; Calvert et al., 2000; Duggen et al., 2005; Valera et al., 2008]. Delamination is as an asymmetric process, analogous to “peeling off” the continental lithosphere causing extension in the overlying crust to migrate in one direction. In contrast, convective removal is supported by the simultaneous, symmetric extension that occurred throughout the entire Alboran region [Platt and Vissers, 1989; Housman and Molnar, 1997; Platt et al., 2003].

The third group of models combines subduction and lithospheric removal processes. Duggen et al. [2005] and Palomeras et al. [2014] used geochemical analysis and surface wave tomography, respectively, to suggest that slab rollback has initiated delamination of the continental mantle lithosphere beneath the
Gibraltar Arc. Our research seeks to evaluate these models through a detailed analysis of the lithospheric structure throughout the western Mediterranean.

2.4 Data and Methodology

2.4.1 Receiver Function Method

A Ps (P wave to S wave, PdS) receiver function (RF) is an approximation of the S-wave Green's function resulting from S conversions from an incident teleseismic P-wave on the receiver side of the P-wave path [Bostock, 2004; Rondenay, 2009]. The individual RF arrivals are P to S-wave (Ps) conversions from seismic impedance discontinuities such as the Moho, the lithosphere-asthenosphere boundary, and the transition zone discontinuities. The positive amplitude arrivals correspond to an increase in seismic impedance with depth (e.g. Moho) and negative amplitudes correspond to a decrease in seismic impedance with depth (e.g. lithosphere-asthenosphere boundary, LAB) [Langston, 1979; Ammon, 1991]. Deconvolving the vertical component from the horizontal components (radial and tangential), or the longitudinal component from the two orthogonal components, approximately removes the source and receiver response functions from the Ps conversion response. In this study, Ps RFs were calculated using both water-level frequency domain deconvolution [Burdick and Langston, 1978] and time-domain iterative deconvolution [Ligorria and Ammon, 1999]. We show results from the latter, which uses an iteratively updated spike train as the predicted RF. At each
iteration, an estimate of the radial component seismogram is calculated by convolving the predicted RF with the vertical component seismogram. The RF is determined by minimizing the difference between the estimated radial component seismogram and the observed radial seismogram until a predetermined upper error bound is reached.

**Figure 2.2:** a) Map showing the azimuthal distribution of the 243 Magnitude > 6.0, teleseismic events used in this study. b) Examples of 0.5, 1, and 2 Hz water level (blue) and iterative deconvolution (red) receiver functions used in this study.

### 2.4.2 Data

Ps RFs were calculated from 243 teleseismic earthquakes recorded by 239 broadband seismograph stations from five temporary and permanent seismic networks in southern Iberia and northern Africa, including the PICASSO (Program to Investigate Convective Alboran Sea System Overturn) and IberArray networks.
Station spacing in these arrays ranges from ~15 – 50 km. Receiver functions were calculated for earthquakes with magnitudes greater than 6.0, which were at source-receiver distances of 30° to 90° (Figure 2.2a).

2.4.3 Processing

Each receiver function was calculated using three separate Gaussian shaping filters corresponding to 0.5 Hz, 1.0 Hz, and 2.0 Hz (Figure 2.2b). At 2 Hz, assuming an average lower crustal Vs of ~4 km/s and a vertical resolution of ~0.25λ, we are able to resolve lower crustal layering on the ~0.5 km scale and at 0.5 Hz layering is resolvable at ~2km. Receiver functions that did not achieve an 80% variance reduction were eliminated, as were those that did not have their highest amplitude close to time zero, corresponding to the direct P arrival. Additionally, all of the RFs were manually inspected for bad traces. After editing, we were left with 6,000 high quality receiver functions. These RFs were common conversion point (CCP) stacked [Dueker and Sheehan, 1997], to create a 3D image volume using the depth conversion and spatial repositioning described in Levander and Miller [2012]. The image volume bin size chosen for this study was spaced 0.125° X 0.125° laterally and 1 km in depth. Conversion horizons were picked from the image volume along latitude and longitude CCP profiles taken every 0.25 degrees.

The depth conversion and spatial repositioning used the linear tomography assumption: The IASP91 1D velocity model was used to calculate P and S wave travel timetables, with 3-D travel time corrections determined from a hybrid 3D
velocity model, which combines the 3D finite-frequency body wave tomography model of Bezada et al. [2013] for mantle structure below 80 km, and the 3D Rayleigh wave model of Palomeras et al. [2014] for structure above 80 km.
Manto.

Figure 2.3: (a) Map showing Moho depth (km) determined from receiver functions. Bold black lines outline regions used to calculate average crustal velocities shown in (b-i). 1: Central Spain; 2: Eastern Betic; 3: Western Betic; 4: Western Rif; 5: Central Alboran Sea; 6: Western Alboran; 7: Eastern Betic. 1-2: Used to calculate average crustal velocities shown in (b-i). 1: Central Spain; 2: Eastern Betic; 3: Western Betic; 4: Western Rif; 5: Central Alboran Sea; 6: Western Alboran; 7: Eastern Betic.
P-wave and S-wave velocities were obtained from each tomography model by assuming IASP91 Vp/Vs ratios. The average crustal velocities of this 3D hybrid model range from Vp = 6.43 km/s, Vs = 3.65 km/s in the western Betic Mountains to Vp = 5.44 km/s, Vs = 3.15 km/s in the Alboran Sea (Figure 2.3).

**Figure 2.4:** a) Map showing the location of CCP profile at 35° latitude, marked by the black line. b-d) CCP profiles at 35° latitude using the final 3D model for depth conversion (b), an alternative 3D model incorporating only body wave tomography results and a constant Vp/Vs of 1.73 (c), and the IASP91 1D model (d). The final moho pick (solid black line), mid crustal discontinuity (dashed black line), the active source moho depth (grey line) and the top of the slab event (dashed white line) is plotted at the same depth in each profile. A comparison between the depths of these features and the corresponding events in each profile shows the difference in the depth between each velocity model. The exact depths of the Moho and “top of slab” at -6° and -4.5° longitude are shown in each profile.
Before the final analysis, we compared the results of the CCP stack with multiple velocity models and Vp/Vs ratios. These models included the 3D hybrid model discussed above, the IASP91 1D model with no 3D correction, and two 3D models derived from the body wave and surface wave models individually assuming a constant Vp/Vs of 1.73. While the shapes of the major events seen on the resulting CCP profiles remain very consistent between velocity models, the depth to these events is variable. Between the final 3D hybrid model and the other tested models, there is an average of ~1-3 km difference at ~30-45 km depth and a ~3-6 km difference between ~50-80 km depth (Figure 2.4).

Hit count maps (Figure 2.5) show the number of traces recorded at each grid point in the CCP volume at 45 km and 75 km depth for the 1 Hz RFs. These are the approximate depths of the Moho and deeper lithospheric structure that will be discussed in the following sections. From these maps we can see good data coverage around the Gibraltar Arc.

**Figure 2.5:** Hit count maps for the 1 Hz data showing the number of traces at each grid point at a) 45 km depth and b) 75 km depth, the depth at which we see the deepest strong, positive RF event throughout the study region.
2.4.4 Thermobarometry

We also made thermobarometric calculations on 19 volcanic samples distributed throughout Spain and Morocco (Figure 2.1a). Primitive magma compositions (basalts to basaltic andesites) were compiled from the literature [El Azzouzi et al., 1999; Turner et al., 1999; Duggen et al., 2004; Duggen et al., 2005; Duggen et al., 2008a; Duggen et al., 2008b]. The magmas represent Neogene magmatic centers spanning from the Atlas Mountains in Morocco, northeastward across the Alboran Basin and into southeastern Spain. For thermobarometric analysis [Lee et al., 2009], we selected only samples with MgO>8 wt. % and SiO$_2$ >45 wt. %. Samples with whole-rock atomic Mg/(Mg+FeT)<0.55 were then excluded. Because the magma thermobarometers are applicable to magmas derived from peridotites, Zn/FeT systematics [Le Roux et al., 2010a; Le Roux et al., 2010b] were examined to exclude those that show extensive clinopyroxene fractionation. Magma compositions were then corrected for olivine fractionation by back-addition of equilibrium olivine increments until the magma was in equilibrium with a mantle having an olivine Mg/(Mg+FeT) of 0.90. We assumed a Fe$^{3+}$/FeT ratio of 0.1. The composition of the primary magma was then inserted into a SiO$_2$-based barometer and MgO-based thermometer to obtain the pressure and temperature of the mantle source of the magma [Lee et al., 2009].
2.5 Thermobarometry Results and Interpretations

Magma thermobarometric calculations provide constraints on the temperature and pressures of the magma source region in the mantle throughout the western Mediterranean. We find that calculated temperatures decrease from 1500° to 1270° C towards the north, as do the pressures of equilibrium (Figure 2.6a-d). These estimated temperatures and depths of equilibration most likely represent an average of polybaric melting columns and thus represent maximum bounds on the thickness of the lithosphere. They are consistent with lithospheric thinning beneath the Rif Mountains and an already extremely thinned Alboran Sea. The hottest and deepest magmas occur beneath the Atlas range (Figure 2.6a-d). Average mantle potential temperature beneath mid-ocean ridges is between 1300°-1400° C [Lee et al., 2009], hence the 1400°-1500° C temperatures beneath the Atlas are anomalously hot. High equilibration pressures suggest deep melting (60-100 km). This is consistent with suggestions that hot plume material may underlie the Atlas Mountains [Duggen et al., 2008a]. Beneath the Alboran basin itself, the estimated temperatures and depths of equilibration are ~1350° C and 20 km, not unlike what is seen beneath typical mid-ocean ridges, implying a highly thinned lithosphere beneath the Alboran. Temperatures and depths of equilibration rise again in southeastern Spain. Depths of equilibration are 40-60 km, consistent with thinned continental lithosphere in southeastern Spain. We note that the magmas from this volcanic field are K-rich and have low Nb/La ratios (Figure 2.6e), suggesting involvement of water in the melting process. Because we have no constraint on the
initial water content of these magmas, it is possible that our estimated temperatures and pressures are maximum bounds for the southeastern Spain magma source.

Figure 2.6: a) Map and graph showing decreasing equilibration pressures of estimated primary magmas towards the north. c-d) Map and graph showing decreasing equilibration temperatures to the north. e) Map showing low Nb/La ratios in the Alboran Sea and southern Spain. f) Graph showing temperature and depths (converted from pressure) of equilibration relative to a typical mantle adiabat for mid-ocean ridge mantle. High temperatures towards the south are hotter than typical mid-ocean ridge mantle.
2.6 Receiver Function Observations

The Moho depth map derived from the CCP image volume (Figure 2.7) shows a highly variable Moho beneath the Gibraltar Arc region with crustal thicknesses ranging from ~25 km in the eastern Betic and Rif Mountains to over 50 km in the western Betic and Rif Mountains as well as the Gibraltar Strait. The Moho map shows an arcuate shaped crustal root (+45 km) around the Gibraltar Arc. We suggest this crustal root may be the cause of the arcuate shaped negative Bouguer anomaly also seen in the region [Fuella et al., 2010]. The regions of maximum crustal thickness in the central Betics and Rif (~53-55 km) correspond to the largest negative Bouguer anomalies of -120 mGal. Mancilla et al. [2012] obtained similar crustal thicknesses with a more limited data set confined to the Betic and Rif Mountains. In the following sections we provide a detailed description of the crustal and lithospheric structure observed in the CCP volume from each part of the study area.
2.6.1 Central Spain/Iberian Massif

In central Spain between latitudes 38° and 40°, we identify the Moho as the strong, positive (red) event between ~27 to 35 km depth below sea level (Figures 2.7 and 2.8). These depths are consistent with previous active source estimates of ~30 - 35 km in the Iberian Massif of central Spain [Simancas et al., 2003; Díaz and
We also observe a series of strong negative (blue) events at ~50-100 km depth between -4.5° and -2° longitude, near latitude 38° (Figure 2.8), which suggest a strong negative velocity gradient under this region. Rayleigh wave tomography along the same profile [Palomeras et al., 2014] shows shear velocities decreasing from 4.35 km/s to 4.2 km/s between ~50 and 125 km, consistent with the RF image (Figure 2.8c). The low velocities are likely associated with the Calatrava Volcanic Province (Figure 2.1a), an intraplate basaltic volcanic field active from the Late Miocene to the Quaternary, that consists of a series of vents, lava flows, and pyroclastic flows distributed over an area of ~4,000 km² within the Iberian Massif [Lopez-Ruiz et al., 1993; Cebriá and López-Ruiz, 1995].

**Figure 2.8:** a) Map showing the location of the CCP and surface wave profiles at 38° latitude (black line). b) 0.5 Hz CCP profile showing consistent crustal thicknesses of ~27-35 km (black line) and strong negative events associated with the Calatrava Volcanic field (white lines). c) Surface wave profile showing low shear velocities associated with the Calatrava Volcanic field (white circle, surface wave data from Palomeras et al., 2014).
Further west and north the CCP profiles show a series of strong positive and weak negative sub-Moho events between ~60 and ~100 km depth beneath the Iberian Massif, suggesting strong mantle heterogeneity between -4.5° to -7°W and 39° to 40° N. In an active source seismic study, the *ILIHA DSS Group* [1993] also found upper mantle heterogeneity in the same location, where they observed multiple upper mantle high and low velocity layers between 40 and 90 km depth.

**Figure 2.9:** a) Map showing the location of CCP profiles at 39.25° and -6°. b-c) 1 and 2 Hz CCP profile at 39.25°. d-e) 1 and 2 Hz CCP profiles at -6°. The solid black line marks the Moho depth in both profiles. The white dashed line indicates the shallower positive arrival coinciding with the layer described by *Ayarza et al.* [2010]. The black dashed line indicates the deeper positive arrival, suggesting layering in the mantle down to ~100 km.
Ayarza et al. [2010] also find evidence for mantle heterogeneity in this region. They suggest an 11 km thick gradational zone with P-wave velocity change from 8.2 km/s to 8.3 km/s between ~61 and 72 km depth. They attribute this zone to the spinel-lherzolite to garnet-lherzolite phase transition (Hales interface). It is possible that the positive event seen in Figure 2.9 at ~65-75 km depth corresponds to this phase transition zone.

### 2.6.2 Betic Mountains

The CCP images of the crust beneath the Betic Mountains (Figure 2.10b-d) are more complicated than those from the Iberian Massif (Figures 2.8 and 2.9). Beneath the Guadalquivir Basin in the west, the RFs show a fairly strong flat Moho signal at ~30 km, consistent with active source results [Carbonell et al., 1998; Torne et al., 2000; Díaz and Gallart, 2009]. Beneath the western Betics the Moho signal weakens and dips eastward to ~45 km depth over a distance of ~50 km. Under the central and eastern Betics, the Moho signal strengthens and dips slightly eastward to ~53 km. Below the Moho in the central Betics, a second positive event extends from ~80 km depth in the west up to ~53 km depth in the east, where it intersects the shallower Moho event. Based on surface wave tomography we correlate the deeper positive event with the top of the Alboran slab (Figure 2.10e). This pair of intersecting RF horizons (Moho and top of slab) is visible under the central Betics between 36.75° and 37.25° latitude (Figure 2.10b-d). The CCP profiles show,
however, that the events appear closer together and the “top of slab” event extends laterally over a smaller distance towards the north.

In the eastern Betic Mountains, where the two events intersect, the combined positive event becomes weak and discontinuous as it shallows from ~53-55 km to ~25-35 km. On both the east and west sides of the deep RF event pair, we observe strong negative horizons at ~50 and ~80 km depth, directly below the shallowing Moho signal beneath the western and eastern Betics (Figure 2.10b-e).

### 2.6.3 Gibraltar Strait/Alboran Sea

In 1 Hz and 2 Hz CCP profiles across the Strait of Gibraltar (Figure 2.11) we identify a variable strength mid-crustal event at 25-30 km depth, extending from the Iberian Massif across the Gibraltar Strait into the Rif Mountains. The Moho is discontinuous: We identify the strong, flat Moho signal beneath the Iberian Massif at 30-35 km (Figure 2.9), and the strong Moho event under the western Betics at ~50 km (Figure 2.10) which we can trace in the 2 Hz CCP profile under the strait of Gibraltar and into the Rif. The Moho is not imaged under the Guadalquivir Basin. The positive event at ~80 km depth under the Betics, which we associate with the Alboran slab, is clearly visible as is a similar event at ~70 km beneath the Rif. We note, the absence of a strong positive event at ~70-80 km depth beneath the Strait of Gibraltar.
The surface waves, velocity slab and the strong negative events in the west and east coincide with the tops of low velocity zones seen in tomography results along latitude 37°. We can see that the deep positive P” event coincides with the top of the high events suggest a negative velocity gradient (h) shows the Hz receiver function and the rayleigh wave.

The white diamonds mark earthquake hypocenter locations. AV indicates the locations where strong negative velocity CCP profiles at latitude 36.75°. The black dashed line marks a strong mid crustal event. The black solid line marks the continental Moho. The deepest positive event, interpreted as the top of the slab, is marked by the white dashed line.

Figure 2.10: A map showing the location of CCP profiles at 37.25°, 37°, and 36.75° latitude, marked by the black.
**Figure 2.11:** a) Map showing the location of the north-south CCP profile extending from the Iberian Massif, through the Betic Mountains, across the Gibraltar Strait, and into the Rif Mountains. (white line) b) 1 Hz CCP profile. C) 2 Hz CCP profile. The black dashed line indicates the shallow mid-crustal arrival. The Moho is marked by the solid black line. The white dashed line indicates the signal from the top of the slab. The white circle outlines the region where we see a negative RF event rather than a strong positive arrival between ~ 70 and 80 km depth.

The crustal thickness beneath the central Alboran Sea was determined using data from the seismic station ES.EALB on Alboran island (35.9°N, -3.0°W; Figure 2.1b). Receiver functions in each frequency band indicate a very shallow Moho, ~15 km depth, in this region (Figure 2.7). Thermobarometric analysis of magmas from this region indicates that beneath the Alboran basin the estimated temperatures and depths of equilibration are ~1350°C and 20 km (Figure 2.6,a-d), implying a highly thinned lithosphere in this region.
2.6.4 Rif Mountains

The complicated crustal structure of the southern Betics and the Strait of Gibraltar extends into northern Rif Mountains of Morocco where a shallow mid-crustal event is seen near ~25 km depth beneath much of the western and central Rif. We identify the Moho as the strong positive event at ~45-50 km depth beneath the Rharb Basin and western Rif Mountains. As under the Betics, the Moho beneath the central Rif dips and intersects a deeper positive event with a maximum depth of about ~70 km. Again, we use surface wave tomography to correlate the deeper positive event with the top of the Alboran slab (Figure 2.12e). This pair of RF events is visible as far south as 34.25° between -5.5° and -3.75° longitude (Figure 2.12b-d). In the eastern Rif, the deep event disappears while the Moho weakens, in places is discontinuous, and shallows from ~50 km to ~25-30 km.

The Moho depths beneath the western and eastern Rif, agree well with preliminary estimates of crustal thickness from the PICASSO-related refraction profiles RIFSIS [Gil et al., 2013]. Figure 2.12 shows close correlation between the Moho depth determined from refraction plotted (bold grey line) and the Moho on the CCP profiles (black lines).

Lastly we see a strong negative event at 60-80 km depth beneath the thinned crust of the eastern Rif between 4.25°W and 3°W, which lies beneath the Guilliz volcanic field. Thermobarometric analysis of basalts from this field indicate mantle melting depths between 40-60 km, suggesting a very thin lithosphere in the region.
marks the Moho depth taken from active source results. The black centered locations mark the melting depths of basaltic samples from the region. The gray bold line
positive event is interpreted as the top of the slab is marked by the white dashed line. The white diamonds mark earthquake
marks the Moho, the black long-dashed line marks the weak discontinuous moho seen in the western part, the black solid
show velocity zones seen in the surface wave. The black short-dashed line marks a strong mid crustal event. The black solid
Hc CCP profiles at latitude 35° (d-e) The CCP profiles at latitude 34°7.5° (f) Shows the
ranges indicate locations of Cenozoic volcanic fields from which samples were used for geochemical analysis. b-c) 1 and 2
Figure 2.12. a) Map showing the location of CCP profiles at 35° 34°7.5° and 34°5° latitude, marked by the black lines. White
2.7 Discussion

2.7.1 Comparison with Surface Wave Tomography

Here we discuss the RF results in comparison to a recent Rayleigh wave tomography velocity model [Palomeras et al., 2014]. As we noted above, the deepest positive RF events beneath the Betics and Rif are correlated with the top of the high velocity Alboran slab imaged by both surface wave tomography and body wave tomography. These studies indicate that at depth this slab has an arcuate shape, and extends from the northern Rif around Gibraltar under southern Spain, extending into the transition zone [Bezada et al., 2013]. Figures 2.10e and 2.12e compare CCP profiles across the Betic and Rif Mountains with the corresponding Rayleigh wave tomography profiles [Palomearas et al., 2014]. In addition to the positive signal from the top of the slab, the strong negative signals in the western and eastern Betics coincide with the tops of low velocity zones seen on both sides of the Alboran slab. Figure 2.13a shows a slice at 75 km depth from the surface wave results, where we see a strong, high velocity slab anomaly beneath the entire Gibraltar Arc and western Alboran Sea. On top of this velocity depth slice we show Moho depth contours of the thickened crust as seen in the Moho map (Figure 2.7), demonstrating that the thickened crust and the slab anomaly coincide.
Figure 2.13b is a 3-D image showing the Moho and subcrustal (~70-80 km depth) positive RF arrival beneath the Gibraltar Arc as a red isosurface at the 50% level, and the Alboran slab as a blue isosurface (at 4.5 km/s). The subcrustal RF signal coincides with the high velocities that define the top of the Alboran slab. These images provide compelling evidence that the deep positive RF events observed beneath the Betics, the Gibraltar Strait, and the central Rif are all conversions from the top of the Alboran slab.

**Figure 2.13:** a) Depth slice at 75 km from the Rayleigh wave tomography model, showing the high velocity body beneath the Gibraltar Arc. White lines show contours of moho depth for the region of thickened crust as seen in Figure 2.7. b) 3-D image of the deep receiver function signal (red) and the high velocity slab material imaged by the surface wave tomography (blue). It is clear that the deepest extent of the receiver function feature coincides with the shallowest extent of the high velocity slab.
2.7.2 Delamination of Continental Lithosphere

As described in the previous sections, we observe a single RF event at ~30-45 km depth beneath the western Rif Mountains (Figure 2.12). We interpret this RF event as the continental Moho. In both the central Rif and Betic Mountains we observe two intersecting positive events, the shallower of which corresponds to the Moho, and the deeper of which corresponds to the top of the Alboran slab. We interpret these intersecting RF events as evidence of ongoing lithospheric detachment. The unusually thick crust (~45-55 km, Figure 2.7, 2.10 and 2.12) results from a depression of the continental lithosphere caused by the load of the descending slab, which Valera et al. [2008] suggest may be enough to initiate delamination of the lower continental crust and mantle lithosphere.

Hypocenter locations for earthquakes (M > 4.5) occurring within the Betic and Rif Mountains are shown in Figures 2.10 and 2.12. In the Betic Mountains, seismicity appears concentrated along 36.75°N between the two positive RF events (Figure 2.10d), which we interpret as deformation associated with lithospheric detachment. In the Rif Mountains, along 35°N (Figure 12b), a concentration of earthquake activity coincides with the location where the Trans Alboran Shear Zone (TASZ) intersects the northern coast of Morocco near 4°W, where the deeper RF event disappears. Focal mechanisms indicate that much of this seismic activity can be attributed to nearly vertical motion initiated by a slab pull [Buforn et al., 1995; Seber et al., 1996; Morales et al., 1997; Wortel and Spakman, 2000].
In the eastern Rif and the eastern and western Betic Mountains the deep RF event pairs transition to a single weak RF event as the crust thins from ~50 km to ~30 km depth. Thermobarometric analysis of basalts from the eastern Rif indicate shallow melting depths between ~40 and ~80 km (Figure 2.12d), directly below the thinning crust, coincident with the strong negative events observed at ~50-60 km depth beneath the eastern Rif and at either end of the Betics. We suggest that the volcanism is evidence of recently removed lithosphere, and that the strong negative events represent asthenosphere that has replaced the delaminated slab. The deep, top of the slab event, therefore, represents the impedance contrast between the high velocity slab lithosphere and the low velocity asthenosphere that has filled the delaminated region.

We have interpreted the weak, shallow RF event directly above the low velocity zones as the continental Moho. The sharp decrease in crustal thickness suggests possible removal of a portion of the lower crust along with the continental lithospheric mantle. The weak character of this signal may be caused by the dip of the interface since the common-conversion point stacking method assumes that PdS conversions are generated by horizontal discontinuities and does not account for scattering of the PdS conversion that may occur at dipping interfaces. However, this shallow signal is weak even where flat in the easternmost Rif and on both sides of the Betics. This may be caused by a decrease in impedance contrast at the Moho, resulting from the emplacement of asthenosphere at or near the base of the crust where continental lithosphere has been removed with accompanying basaltic dike emplacement in the lowermost crust. The surface wave tomography results indicate
a small velocity contrast from the crust (depth slice 25 km) where $V_s \approx 3.8 \text{ km/s}$ to the upper mantle (depth slices 35 and 50 km) where $V_s \approx 4.1-4.2 \text{ km/s}$ [Palomeras et al., 2014]. A joint inversion of the RF and surface wave datasets will better constrain the velocity contrast associated with the shallow Moho boundary, and will be discussed in the following chapter.

We suggest that delamination of the continental margin lithosphere was initiated when oceanic lithosphere was consumed by subduction and slab rollback. The load on the continental margin resulting from the Alboran slab is depressing the Moho beneath the Rif and Betic Mountains, initiating lithospheric detachment, and permitting asthenosphere to replace lithosphere at shallow depths (Figure 2.14). The presence of asthenospheric material at subcrustal depths may be responsible for the recent uplift of the Betic Mountains.

This model is similar to that proposed by Duggen et al. [2005] for the western Mediterranean. Additionally, Garcia-Castellanos and Villasenor [2011] use a lithospheric delamination and slab tear model to explain the Messinian salinity crisis and the cyclicity observed in early salt deposits in the western Mediterranean. In North America, Levander et al. [2011] noted that a similar set of observations show crustal detachment at the top of a lithospheric drip beneath the Colorado Plateau.
delamination of the continental lithosphere.

continental lithosphere and where it has detached in cartoon form. A profile across the Rif Mountains showing east to west surface shows the surface projection of the slab. The color contours indicate where the slab remains attached to the surface and where it has detached. The dotted line along the surface projection of the slab beneath the Gibraltar Arc. The dotted line along the surface projection of the slab beneath the Gibraltar Arc.

**Figure 2.14:** (a) 3-D cartoon showing the geometry of the slab beneath the Gibraltar Arc. The dotted line along the surface projection of the slab beneath the Gibraltar Arc. The dotted line along the surface projection of the slab beneath the Gibraltar Arc.
2.7.3 Attached vs. Detached Slab

The deep RF events seen in the central Betics appear to merge with the Moho between -4.5°W and -3.5°W along the northernmost profile at 37.25° N (Figure 2.10b). In the central profile (Figure 2.10c) the two positive events merged over an even shorter distance near -3.5° W, and in the southernmost profile at 36.75° N the two events appear nearly completely separated (Figure 2.10d). We suggest that the slab is still attached to the crust beneath a portion of the central Betic Mountains, likely where the Moho reaches its maximum depth of ~53 km, and that the seismicity at and below the Moho reflects the ongoing detachment. In the Gibraltar Strait we do not observe a deep top of the slab RF event in the CCP profile. Instead we see negative events below the Moho (Figure 2.11), in agreement with both the P wave and Rayleigh wave tomography images that indicate the subducting slab is not as shallow beneath the Gibraltar Strait as along the adjacent continental margins [Bezada et al, 2013; Palomeras et al., 2014].

The surface wave tomography shows low S-wave velocities (~4.1 km/s) down to ~60 km depth [Palomeras et al., 2014], suggesting that the subducting slab has already detached from the crust and has been replaced by asthenosphere.

In contrast to the Betics, the deep RF events in the central Rif Mountains are very close together everywhere but near their eastern end (~4°W), suggesting that the slab remains attached beneath a large portion of the central Rif. This is in agreement with geodetic modeling of GPS data [Perouse et al., 2010] that suggests that the Alboran slab is still attached to the Rif crust. We suggest that the maximum
Moho depths (~53 km) beneath the central Betic and Rif Mountains correspond to places where the slab remains attached to the base of the continental lithosphere.

2.8 Conclusions

Ps receiver functions and thermobarometric constraints from late Cenozoic basalts in the western Mediterranean provide evidence for ongoing lithospheric delamination beneath the Gibraltar Arc, with the detachment surface being the continental Moho. Beneath the central Betic and Rif Mountains, we see both a strong Moho and an intersecting dipping strong, positive mantle event which we interpret as the top of the Alboran Sea slab imaged previously by surface wave tomography [Palomeras et al., 2014]. We interpret the intersection of the Moho and upper mantle event from the top of the Alboran slab as indicative of ongoing lithospheric delamination. The RF images reveal a highly variable Moho with depths ranging from ~53 km in the western Betic and Rif Mountains where the slab is still attached, to ~20-25 km in the eastern Betic and Rif Mountains and the central Alboran Sea where the slab appears to have already detached. Detachment is being driven by the slab-pull force due to the load of the sinking Alboran Sea slab.

In addition to thinner crust, in the eastern Rif and the eastern and western Betics, we observe negative polarity mantle RF events, which we interpret as a shallow, recently formed, post-detachment LAB. Thermobarometric analysis of basalts indicate shallow mantle melting depths of 40-60 km in the eastern Rif, the eastern Betics, and the central Alboran Sea, supporting the interpretation of recent
lithospheric thinning and removal. The surface volcanics and shallow LAB are indicative of decompression melting that has accompanied the upwelling asthenosphere that replaces the Alboran slab.

We conclude that at the end of oceanic subduction rollback, the edges of the Alboran slab began detaching from adjacent continental margins. The Alboran slab has detached from the base of the crust beneath the eastern Betic and Rif Mountains, is still attached beneath the central Betic and Rif mountains, and has already detached from the western Betic Mountains, and the Strait of Gibraltar.
Chapter 3

Imaging the Alboran Slab and Western Mediterranean with the Joint Inversion of Ps Receiver Functions and Rayleigh Wave Phase Velocities

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Abstract

The Africa-Eurasia collision has created a diffuse zone of deformation within and around the Alboran Sea located at the westernmost end of the Mediterranean where the Iberian Peninsula meets North Africa. This region has a complex tectonic history involving subduction, slab rollback, and lithospheric delamination. Here coeval contractional and extensional processes are responsible for the uplift of the Betic, Rif, and Atlas Mountains as well as the formation of the Alboran basin. We
present a 3-D shear wave crust and upper mantle velocity model developed from
the joint inversion of Ps receiver functions (1 and 2 Hz) and Rayleigh wave phase
velocity data (4-167s) gathered from 242 broadband seismic stations deployed
throughout central Spain and Morocco. The velocity model reveals a thickened crust
throughout the western Gibraltar Arc and a nearly vertical high velocity Alboran
Slab directly beneath this. The crust thins abruptly to the east under both the Betic
and Rif Mountains where the top of the Alboran Slab descends into the mantle and
low velocities (~4.2) are present above the slab directly beneath the crust. These
results suggest that detachment of the Alboran Slab has resulted in delamination of
the continental lithosphere and upwelling of asthenosphere. We suggest that the
emplacement of hot buoyant asthenosphere at shallow depths supports the high
elevations of the Sierra Nevada in southern Spain. We also observe a thin crust
(~33-38 km) and a NE-SW trending corridor of thin lithosphere (~40-60 km)
beneath the Atlas Mountains. These results are consistent with previous studies and
suggest that the high elevations of the Atlas Mountains are also supported by
buoyant asthenosphere.

3.1 Introduction

The Africa-Eurasia collision has resulted in a diffuse, actively deforming plate
boundary in the western Mediterranean between the Iberian Massif (Eurasian
plate) and the Sahara Craton (African Plate). Within the plate boundary, the Betic
and Rif Mountains form the Gibraltar Arc orogenic system and surround the
extensional basin known as the Alboran Sea. The Atlas Mountains form an
intracontinental mountain chain running NE-SW from the western margin of
Morocco to the Rif Mountains. They accommodate ~2 mm/yr of ongoing Africa-
Iberia convergence (Ramdani, 1998; Missenard et al., 2006; Stich et al., 2006; Figure
3.1), which is occurring at a relatively slow rate of ~4-6 mm/yr in a WNW-ESE
direction (Argus et al., 1989; McClusky et al., 2003; Perouse et al., 2010). Despite
this slow NW directed convergence, GPS data show that both the Betic and Rif
Mountains currently exhibit southward motion (up to ~5 mm/yr), relative to Africa,
suggesting more complicated sub-crustal or sub-lithospheric processes are affecting
the Gibraltar Arc (Perouse et al., 2010). Numerous tectonic models have been
proposed to explain the processes responsible for Cenozoic compression within the
Gibraltar Arc, simultaneous extension in the Alboran Sea, and uplift of the Atlas
Mountains. These models invoke slab rollback (Royden et al. 1993; Lonergan and
White, 1997; Rosenbaum et al., 2000; van Hinsbergen et al., 2014), slab break-off
(Zeck, 1997), and lithospheric removal (Platt and Vissers, 1989; Seber et al., 1996;
Platt et al., 2003) and combinations thereof. Although no consensus has been
reached, there is general agreement upon a tectonic history driven by subduction
and slab rollback.
Figure 3.1: a) Map showing the elevation and major geologic features in the western Mediterranean. The black dashed line indicates the boundary between exposed Paleozoic basement rock (Variscan Spain) and Mesozoic/Cenozoic basin deposits and mountain ranges (Alpine Spain). B) Map showing the locations of the 242 stations used in this study. Black lines show the locations of profiles discussed in the text and shown in Figure 3.5.

Around 35 Ma, a northwest dipping subduction zone along southeastern Iberia began to retreat rapidly toward the south and east (Rosenbaum, 2002; Faccenna et al., 2004; van Hinsbergen et al., 2014). This south directed rollback caused significant extension in the overriding continental lithosphere. A portion of
this lithosphere, known as the Alboran Domain, was transported across the Mediterranean as the trench turned towards the southwest ~25 Ma and formed an eastward dipping subduction zone between Iberia and Africa (Rosenbaum, 2002). As the subducting Neo-Tethys plate continued to roll back towards the west, the Betic and Rif Mountains formed ~20-10 Ma when the edges of the extending Alboran Domain were thrust onto the passive margins of southern Iberia and North Africa. The Alboran Domain now forms the Betic and Rif Internal Zones (Figure 3.1a) with the passive margin sequences forming the External Zones (Figure 3.1a; Platt and Vissers, 1989; Rosenbaum, 2002; Platt et al., 2012). Two accompanying foreland basins, known as the Guadalquivir and Rharb, with the Betic and Rif thrust belts respectively.

Previous tomography studies have imaged a high velocity anomaly, referred to as the Alboran Slab, extending vertically from beneath the Gibraltar Arc lithosphere into the transition zone (Blanco and Spakman, 1993; Zeck, 1996; Calvert et al., 2000; Gutscher et al., 2002; Bezada et al., 2013; Palomeras et al., 2014). The most recent of these, along with recent receiver function and geochemical studies, support a tectonic model in which slab rollback initiated delamination of the continental mantle lithosphere beneath the Gibraltar Arc (Duggen et al., 2004; 2005; Palomeras et al., 2014; Thurner et al., 2014, Levander et al., 2014), and possibly helped trigger whole or partial delamination of mantle lithosphere from beneath the Atlas (Ramdani, 1998; Bezada et al., 2014).

Here we discuss a three-dimensional shear velocity model constructed from the joint inversion of Ps receiver functions and fundamental-mode Rayleigh wave
phase velocity dispersion data (4-167s). We use data from 8 seismic networks in the region (Figure 3.1b), including those of the US PICASSO project and the Spanish Iberarray/ Topolberia project (http://iberarray.ictja.csic.es/). We provide additional details on lithospheric seismic velocity structure of the Alboran Slab beneath the Gibraltar Arc, building on recent tomography studies (Bezada et al., 2013; Palomeras et al., 2014). The joint inversion of these datasets provides better constraints on the Moho depth and shallow slab structure, and provides additional evidence of delamination of the mantle lithosphere under the Betic and Rif Mountains. We find a highly variable Moho depth around the Gibraltar Arc where increased Moho depths (up to ~45-50 km) in the west are coincident with the region where the slab remains attached to the base of the Gibraltar Arc lithosphere and shallow Moho depths (~25 km) are located in the region of slab detachment where we suggest the Alboran Slab has delaminated the Gibraltar Arc mantle lithosphere and possibly the lower crust. Under the portion of the Atlas Mountains crossed by the PICASSO array, we find thin lithospheric mantle, which we attribute to previous mantle downwelling events as suggested by Ramdanai et al. (1998) and Bezada et al. (2014).

3.2 Data

We used data from 242 broadband seismic stations from 8 different networks in the region, including 107 stations from the IberArray network and 88 stations from the PICASSO (Program to Investigate Convective Alboran Sea System...
Overtorn) network. Station spacing for each array ranges from ~15 within the PICASSO network to ~50 km within the areal IberArray network. For this study, we used broadband surface wave dispersion curves (4-167 s) for each station determined from ambient noise and ballistic Rayleigh wave analysis (Palomeras et al., in prep). We inverted for the isotropic Vs structure using the joint inversion of these dispersion curves and Ps receiver functions made with data from the same stations (Thurner et al., 2014).

3.2.1 Rayleigh Wave Phase Velocities

The Rayleigh wave phase velocity dataset combines ambient noise and ballistic phase velocities for periods between 4 and 167 seconds (Palomeras et al., in prep). Ambient noise tomography approximates the surface wave Green’s functions for the path between two stations by computing the cross-correlation waveform of seismic noise between station pairs (Bensen et al., 2007; Shapiro et al., 2005; Sabra et al., 2005). The ambient noise observations provide shorter period dispersion data (4-40s), increasing the dataset sensitivity to crustal shear velocity. The ballistic Rayleigh wave phase velocities were calculated for 18 frequency bands from 20 to 167 s using 168 teleseismic events (M > 6.0) at epicentral distances between 30° and 120°. The phase velocities were calculated using the modified two-plane wave technique (Forsyth and Li, 2005) and finite-frequency kernels (Yang and Forsyth, 2006). Palomeras et al. (in prep) calculated broadband dispersion curves (4-167 s) by combining both ballistic and ambient noise phase velocities. We use
these combined dispersion curves to better constrain both crust and mantle velocities.

3.2.2 Ps Receiver Functions

The Ps receiver functions were calculated from 243 teleseismic events (M >6.0) at epicentral distances of 30° to 90° for center frequencies of 1 Hz and 2 Hz (Thurner et al., 2014). In order to improve the signal-to-noise ratio, the receiver functions at each station were divided into ray parameter groups of <0.05 s/km, 0.05-0.06 s/km, and > 0.06 s/km and stacked at the central ray parameters (0.045, 0.055, and 0.065 s/km) after applying a Ps move-out correction assuming the 1-D IASP91 mantle velocity model. Additionally, all receiver functions at a given station were collectively stacked at a ray parameter of 0.05 s/km. In order to avoid modeling multiple reflections in the joint inversion, we used a receiver function time window from -5 to 10 seconds, 10s being equivalent to approximately 100 km depth. This time window is appropriate as previous receiver function and tomography studies from this region have shown that the primary discontinuities of interest (Moho, top of Alboran Slab, LAB) are typically less than 100 km depth (Palomeras et al., 2014; Thurner et al., 2014).
3.3 Joint Inversion Method

In the joint inversion, the surface wave dispersion data provides constraints on the bulk shear velocity, while the receiver functions provide constraints on the shear velocity contrast and depths to discontinuities. The goal of the joint inversion of receiver function and phase velocity data is to reduce the non-uniqueness arising from the velocity-depth trade-off inherent in receiver function analysis alone and the depth distribution of Vs inherent in phase velocity inversion alone (Julia et al., 2000).

Using the algorithm of Herrmann and Ammon (Computer Programs in Seismology, 2002), we inverted for the 1-D shear velocity structure beneath 242 stations using the data described above and following a procedure similar to that described by Liu et al. (2012). The cost function that this inversion seeks to minimize is:

$$
1 - p \frac{1}{N_{rf}N_{pts}} \sum_{i=1}^{N_{rf}} \sum_{j=1}^{N_{pts}} \| \epsilon_{ij} \|^2 + \frac{p}{N_{disp}} \sum_{k=1}^{N_{disp}} \| \epsilon_{k}^{disp} \|^2
$$

(1)

where $\epsilon = (d^{OBS} - d^{PRE})/\sigma$ is the relative residual between the observed and predicted data and $N_{rf}$, $N_{pts}$, and $N_{disp}$ are the total number of receiver functions, total number of sampling points in each receiver function, and total number of surface wave dispersion frequencies, respectively. $p$ is the influence factor.
(0 \leq p \leq 1), which weights how each of the two datasets affects the minimization procedure. Thirty iterations were used for the inversion beginning with a 1-D mantle model based on ak135 with no a priori crustal structure. This initial model allows the data to change the model while avoiding artificial low-velocity zones and the input of sharp discontinuities that could persist throughout the inversion (Herrman and Ammon, 2002). Previous receiver function results were used to provide a Moho depth estimate at each station. The joint inversion allows for a significant change in the model velocity in the layers close to Moho depth by relaxing the smoothing regularization constraint.

The joint inversion was performed station by station and the resulting 1-D velocity profiles (e.g. Figure 3.2) were interpolated onto a regular grid to form a pseudo 3-D velocity volume. The horizontal resolution of this velocity volume is controlled by the station spacing, which varies throughout the region with the densest station distribution surrounding the Alboran Sea in the Gibraltar Arc. The ability of the dispersion data to constrain the shear velocity decays with depth, with sensitivity peaks at depths of \( \sim 0.3 \, \lambda \) (Yang and Forsyth, 2006). Given a receiver function's vertical resolution of \( \sim 1/4\lambda \), these data provide \( \sim 1\) km and \( \sim 0.5\) km vertical resolution for the 1 Hz and 2 Hz frequencies, respectively, assuming \( V_s \sim 4.0 \) km/s in the lower crust. The vertical resolution of the joint inversion, therefore, is controlled by that of the dispersion data, the receiver functions, and the influence factor used in the inversion at each station.
black circles and the predicted dispersion curve calculated from the final Vg model is shown in red.

determined from the joint inversion (c) Rayleigh wave phase velocity used for the joint inversion are shown by the

depth model (blue). The red line indicates the final Vg model.

parameter for each stacked trace. b) The starting Vg model for the joint inversion consists of a constant velocity taken

shows observed (blue) and predicted (red) receiver functions along with the percentage fit of the model.

Figure 3.2: An example of the receiver function and Rayleigh wave joint inversion results at station ES7BAD.

(a) Receiver functions

(b) Velocity Model

(c) Rayleigh Wave Dispersion Curve
We can evaluate the accuracy in the joint inversion procedure by calculating the percentage fit of the forward modeled receiver functions and the residual of the calculated dispersion curves. In general, we find our forward modeled receiver function fit to be ~90% and the average of the absolute value of the residual for the dispersion curves to be ~.01 km/s. The estimated errors for the resulting absolute Vs model are not easily computed, as they depend on the uncertainties in the input receiver function and dispersion data (Herrman and Ammon, 2002). However, Julia et al. (2000) evaluate the joint inversion procedure using synthetic data contaminated with non-Gaussian noise. In this case, the 95% confidence-intervals for the predicted model (estimated from the model covariance matrix) were frequently ~.1 km/s for an influence factor ($p$) of 0.5. They show similar results for a real dataset from the Arabian shield.

### 3.4 Results

#### 3.4.1 Moho and LAB Depth

To create a Moho map for the region we manually picked the Moho depth beneath each station from the 1-D profiles produced by the joint inversion. The LAB depth was determined for each station by picking the maximum negative Vs gradient from the 1-D profiles while applying a minimum velocity constraint of $Vs =$
4.4 km/s on the lithospheric mantle. Examples of station data, including 1-D profiles, receiver functions, dispersion curves, and the corresponding Moho, LAB, and “top of slab” picks for six different stations distributed throughout the region can be seen in Figure 3.3. Figure 3.4 shows the resulting Moho and LAB depth maps. Although the lithosphere thickens to ~110 km beneath a small portion of southwestern Spain and southern Portugal, these results show little variation in Moho (~28 - 31 km) and LAB depth (~80 km) beneath central Spain and the Iberian Massif. However, both boundaries shallow significantly towards the eastern margin of Spain, with depths of ~25 km and <50 km for the Moho and LAB (Figure 3.4a,b). Our Moho depth estimates in this region are consistent with those from previous receiver function studies (Thurner et al., 2014) and active source experiments (e.g. Carbonell et al., 1998; Ayarza et al., 2010) summarized by Diaz and Gallart (2009).

Figure 3.3: Station data for 6 stations distributed throughout the study region. The corresponding Moho, LAB, and “top of slab” (TOS) picks are identified on the 1-D profiles. B and c show stations in the eastern Betics. In be we can see a gradational increase in velocity below the Moho (blue dashed box) indicating the slab is still attached in this region. In c we can see a sharper increase in velocity around 60 km depth (blue dashed box). This is the TOS arrival, suggesting the slab is detached from the crust beneath this station. D and e show data for stations in the Rif Mountains were we see evidence for lower crustal delamination. The low velocity regions are outlined with a green dashed box and the corresponding RF arrivals are identified with green arrows.
Figure 3.3
**Figure 3.4**: Maps showing the Moho depth (a) and LAB depth (b) estimated from the joint inversion results. The white triangles indicate Cenozoic volcanism and the grey shaded region in part b represents the Alboran Slab at 100 km depth. No LAB pick was made at stations located within this region.

Around the Gibraltar Arc, we observe a much deeper Moho, reaching ~50 km in the Rif and up to ~45 km in the central Betics (Figure 3.4a). The deeper Moho overlies the high velocity Alboran Slab, which has been imaged in this and previous studies (Bezada et al., 2013; Palomeras et al., 2014) and will be discussed further in the sections to follow. The LAB is undefined (grey areas in Figure 3.4b) where the Alboran slab extends from the base of the crust to the transition zone (Bezada et al., 2013). The LAB directly to the east of the Alboran Slab in both the eastern Betics
and Rif is shallow (~50-60 km, Figure 3.4b), consistent with that along the eastern margin of Spain. The Moho depth in this same region sharply decreases to ~22-25 km from west to east over a very short distance close to the intersection of the Trans Alboran Shear Zone and the eastern Betics and Rif (Figure 3.4a).

The Gibraltar Arc Moho depth in this study follows a similar pattern as that determined from previous receiver function studies, which also show increased Moho depths to the west and a sharp decrease in Moho depth to the east (Mancilla et al., 2013; Thurner et al., 2014). Our results in the Betic Mountains, however, are shallower than those shown by Thurner et al. (2014), who indicate Moho depths up to ~45 km in the eastern Betics were we observe Moho depths of ~35 km (Figure 3.5). This discrepancy exists in the eastern Betics, around the southern edge of the Alboran Slab where there has been recent volcanism and where we observe a very shallow LAB depth. In these locations there is a weak or discontinuous receiver function Moho signal, which Thurner et al. (2014) suggest may have been caused by the emplacement of low velocity asthenosphere at or near the base of the crust, resulting in a low impedance contrast. The joint inversion, utilizing dispersion and receiver function data, allows for the determination of absolute velocity, providing better constraints on the Moho depth and shallow mantle structure in regions where the impedance contrast may be small and the RF signal weak.
Figure 3.5: Map showing the difference between the moho depth determined by the receiver functions and the moho depth determined by the joint inversion. In general, the receiver function moho is deeper than the joint inversion moho. However, beneath much of Spain the two data sets generally agree within 3 km. The largest discrepancies are seen at the edges of the Alboran Slab (dashed line) where asthenosphere has upwelled to shallow depths decreasing the impedance contrast and producing a weak receiver function signal. In these locations we use the absolute velocities determined from the joint inversion to differentiate between crustal velocities and asthenospheric velocities in order to better constrain the moho depth. White triangles indicate the locations of Cenozoic volcanic fields.

The most obvious feature visible on the LAB map in the Gibraltar Arc and central Spain (Figure 3.4b) is the contrast between LAB depths of ~80 - 100 km in the west with LAB depths of ~60 km in the east. The west to east LAB shallowing coincides with a sharp west to east decrease in Moho depth around the Gibraltar
Arc. This transition occurs close to the onshore extensions of the Trans Alboran Shear Zone in Spain and Morocco. It also coincides with the NE-SW trend of Miocene to Pleistocene basaltic volcanism observed throughout the region. We discuss these Moho and LAB depth associations further below.

In the Atlas Mountains, we observe a variable Moho depth ranging from ~33 km in the Middle Atlas to ~40 km in the High Atlas (Figure 3.4a). These results agree well with those obtained by the RIFSIS and SIMA experiment (Ayarza et al., 2014; Gil et al., 2014), which consisted of two long offset wide-angle seismic reflection lines, that jointly extend from the northern Rif through the Middle and High Atlas Mountains. These seismic lines were coincident with the dense profile of PICASSO broadband seismic stations in Morocco (Figure 3.1). Along this seismic profile, Moho depths vary from ~40 km in the Rif Mountains, decreasing to ~33 km beneath the Middle Atlas Mountains, to ~40 km beneath the High Atlas. These Moho depths are shallow compared to those required (~45-50 km) for crustal compensation of the high elevations (3-4 km) observed in the Atlas Mountains (Figure 3.10a). A shallow LAB depth of ~40-65 km is observed throughout much of the Atlas Mountains. A region of thin to non-existent mantle lithosphere appears to extend from southwestern Morocco north to the eastern Rif (Figure 3.6b, 3.7d).
3.4.2 Central Iberian Peninsula Vs

In the upper crust (~15 km depth) of central Iberia we observe shear wave velocities of ~3.6 km/s in the west and a rather sharp decrease in Vs to ~3.5 km/s in the east (Figure 3.6a). The lower crust becomes more homogeneous with velocities ranging from ~3.7 to 3.8 km/s. In the upper mantle directly beneath the Moho (~30-45 km depth), we observe velocities of ~4.4 km/s in the west beneath the Iberian Massif. To the east, beneath the Calatrava Volcanic Province (CVP) in central Spain, the upper mantle velocities decrease to ~4.1-4.3 km/s and increase again to ~4.5 km/s along the eastern margin of Spain. Velocities in the central Iberian Peninsula become fairly homogeneous in the deeper mantle where we observe shear wave velocities of ~4.4 km/s at 80 km and ~4.3 km/s below this (Figure 3.6b).

The west to east transition from high to low velocity in the upper crust of the central Iberian Peninsula is likely associated with the transition from “Variscan Spain”, consisting of exposed Precambrian and Paleozoic basement, to “Alpine Spain”, consisting primarily of Mesozoic and Cenozoic sedimentary basins and young mountain belts (Gibbons and Moreno, 2002). The low velocities observed in central Spain between 35 and 45 km depth are likely caused by the magmatic source for the Calatrava Volcanic Province, an intraplate basaltic volcanic field that has been active since the Late Miocene (Lopez-Ruiz et al., 1993; Cebria and Lopez-Ruiz, 1995).
3.4.3 Gibraltar Arc Vs

In the Betic and Rif Mountains we observe an arcuate pattern of low crustal shear wave velocities down to ~50 km depth, which curve around the Alboran Sea along the Gibraltar Arc (Figure 3.6a), similar to that shown by Palomeras et al. (2014). In the shallow to mid crust (15-25 km) these velocities are lowest in the western Betic and Rif, increasing slightly under the central Betics. Between 25 and 50 km depth, we observe a sharp velocity increase from crustal velocities in the west to mantle shear wave velocities (up to ~4.2 km/s) in the east. The opposite pattern is observed in the mantle beneath the Gibraltar Arc, with relatively high velocities in the west and low velocities in the east (Figure 3.6b). We observe velocities between ~4.5-4.6 km/s extending from ~50km down to 100 km depth throughout the western Rif and the majority of the Betics. Although they decrease with depth to ~4.4-4.5 km/s at 150 km (Figure 3.6b), these relatively high velocities are observed as deep as 250 km and they are bounded to the east by relatively low (~4.2 km/s) mantle velocities.

The high mantle velocities observed around the Gibraltar Arc are consistent with recent body wave and surface wave tomography studies (Bezada et al., 2013; Palomeras et al, 2014) and are interpreted as the Alboran slab, which is no longer subducting, but rather hanging nearly vertically in the mantle. The strong correlation between the low crustal velocities and high mantle velocities suggests that the hanging Alboran slab is responsible for depressing the Moho around the Gibraltar Arc as proposed in previous studies (Palomeras et al., 2014; Thurner et al.,
Additionally, the presence of asthenospheric mantle velocities (~4.2 km/s) at shallow depths (~25-30 km/s) just east of the Alboran Slab suggests that lithospheric delamination has occurred beneath the eastern Betic and Rif Mountains, a topic to be discussed in greater detail in the following sections.

3.4.4 Atlas Mountains Vs

The first 15-25 km of the crust beneath the Atlas Mountains is fairly homogeneous, with velocities ranging from ~3.5 to 3.7 km/s (Figure 3.6a). By 35 km depth, however, we observe shear wave velocities as high as ~4.2 km/s beneath the eastern Middle and High Atlas, increasing to ~4.3 km/s by 45 km depth (Figure 3.6b). Velocities decrease slightly in the deeper mantle beneath the eastern Middle and High Atlas Mountains becoming as low as ~4.1 km/s between 80 and 100 km depth and increasing to ~4.2 km/s by ~120 km (Figure 3.6b). The presence of asthenospheric mantle velocities directly beneath the Moho (<45 km depth) suggests that very little to no mantle lithosphere exists beneath the eastern High and Middle Atlas (Figure 3.6b and 3.7d).
Figure 3.6: Depth slices showing shear velocities for the crust and upper mantle from 15-150 km depth.

The white triangles indicate the locations of Cenozoic volcanic fields. The Calatrava Volcanic Field (CVF) in central Spain is labeled in the crust and upper mantle depth slices. The black dashed line (a) indicates the boundary between exposed Paleozoic basement rock (Variscan Spain) and Mesozoic/Cenozoic basin deposits (b). Note that the color bar changes for each depth slice in the crust and upper mantle.
Figure 3.7

a) A'  
Latitude 38° N  
Iberian Massif  
Betic Mountains  
Moho  
LAB  
Depth (km)  
Longitude (Degrees)  
Vs (km/s)

b) B'  
Latitude 38° N  
Sierra Nevada  
Granada Basin  
Moho  
LAB  
Lithospheric Detachment  
Depth (km)  
Longitude (Degrees)  
Vs (km/s)

c) C'  
Latitude 35.25° N  
Rif Mountains  
Trans-Alboran Shear Zone  
Moho  
LAB  
Alboran Slab  
Lithospheric Detachment  
Depth (km)  
Longitude (Degrees)  
Vs (km/s)

d) D'  
Latitude 35.25° N  
Iberian Massif  
Betic Mountains  
Rif Mountains  
Atlas Mountains  
Moho  
LAB  
Alboran Slab  
Melting  
Thin Lithospheric Corridor  
Depth (km)  
Longitude (Degrees)  
Vs (km/s)
Figure 3.7: Profiles through the Vs velocity volume at 38° N (a), 37° N (b), 35.25° N (c), and along a N-S profile extending the entire length of the study region. The black triangles mark the station locations, the black dashed line marks the Moho along each profile, the white dashed line marks the LAB along each profile, the red dashed circles indicate regions of slab and lithospheric detachment, the white diamonds indicate earthquake hypocenters, and the black diamonds indicate melting depths of basalts sampled from the region (presented in Thurner et al., 2014).

3.5. Discussion

3.5.1 Lithospheric Detachment

By providing constraints on both the absolute shear velocities and the seismic discontinuities beneath each station, the joint inversion allows us to identify the top of the Alboran slab and determine where it is attached to and detached from the lithosphere beneath the Gibraltar Arc. Figure 3.8 shows the locations of all stations above the Alboran Slab; the blue squares indicate where the slab is attached to the Gibraltar Arc lithosphere and the red squares indicate where the slab is completely detached. The slab depth indicated in this figure is relative to the Moho depth at each station. This figure shows that the slab is attached to the base of the crust beneath most of the western Betics and Rif and becomes detached towards the east. There is also evidence for slab detachment towards the Alboran Sea as Figure 3.8 indicates detachment in the southwestern Betics (~36.5° N and ~4.75° W) in an area that is also the site of Cenozoic volcanism. Figure 3.9b suggests slab detachment beneath the Alboran Sea. Although receiver functions do not provide
constraints in this region, the dispersion data from Palomeras et al. (in prep) do. In this figure we see that the slab appears detached beneath the center of the Alboran Sea (white dashed circles).

**Figure 3.8:** Map showing stations located above the Alboran Slab with the symbol color indicated the depth of the slab beneath the Gibraltar Arc crust. TASZ – Trans Alboran Shear Zone.

Cenozoic volcanism has occurred along the eastern edges of the detachment on either side of the Alboran Sea (Figure 3.7b,c). Slab detachment in the Rif Mountains is found below the Trans Alboran Shear Zone and is associated with Cenozoic Volcanism (Figure 3.7c). In this region, we also observe low velocities
directly below mantle velocities, forming what appears to be a “double Moho”
associated with a high concentration of earthquakes (Figure 3.7c and 3.9a,b). Figure
3.9a,b shows a 3D view of this feature where we can see that it extends north from
the Rif Mountains into the Alboran Sea. Figure 3.3d,e shows the 1-D velocity profiles
at stations PM36 and PM37, located directly above this feature. We interpret these
low velocity bodies seen below the Moho as remnant lower crust that is
delaminating beneath the eastern Rif Mountains and the southern portion of the
Alboran Sea. The mantle lithosphere and the base of the crust that have delaminated
are replaced by upwelling asthenosphere, evidenced by the low velocities observed
directly beneath the eastern Rif and the ∼13 Ma volcanism at the surface (Figure
3.7c and 3.9a).

In the Betic Mountains, we do not see the delaminating lower crustal feature
discussed above (Figure 3.7b and 3.9). However, we do see thinned crust and
lithospheric mantle beneath the detached slab region in the eastern Betics (Figure
3.7b and 3.9). Therefore, we suggest that slab detachment and lithospheric
delamination occurred earlier in the Betic Mountains than in the Rif.
beneath the Sierra Nevada in the Betic mountains and the Alboran Sea.

Figure 3.9: a) 3D figure showing the high velocity Alboran slab (blue), drawn as a 4.45 km/s isovolume, and the delaminating crust beneath the Rif mountains at 35.25° latitude. Black dots represent melting depths for Cenozoic volcanic samples.

b) 3D figure viewed from the ESE with crossing velocity profiles at 4° longitude and 37° latitude. This figure is viewed from the SE. The velocity profile cross the Rif isovolume, and the delaminating crust beneath the Rif mountains (red) drawn as a 4.1-4.15 km/s isovolume.
3.5.2 Uplift of the Sierra Nevada

The Sierra Nevada Mountains, located near the southern coast of Spain, have been uplifted since the late Miocene to elevations greater than 3 km, making them the highest mountains in the Betic Range (Braga et al., 2003; Perez-Pena et al., 2010). The Granada Basin, just northwest of the Sierra Nevada, has also been uplifted since this time, evidenced by ~10 Ma marine sediments now located at ~1500 m elevation within the basin (Galindo-Zalvidar et al., 1999). Geological surface data indicate that this basin and the Sierra Nevada are still tectonically active, as uplift has continued into the Holocene and there is a high concentration of earthquakes occurring along extensional faults in both regions.

Figure 3.8 shows the location of the Sierra Nevada and Granada Basin relative to the Alboran Slab. Here we see that the Sierra Nevada are located entirely above the detached slab in the eastern Betic Mountains (Figure 3.7b, 3.8, 3.9). The shallow Moho (~35 km) beneath the Sierra Nevada is inconsistent with that expected (~45-50km) for crustal support of such high elevations (Figure 3.10a). Additionally, figures 3.7b and 3.9 show a concentration of low mantle velocities (~4.1-4.2 km/s) directly beneath the Sierra Nevada above the top of the Alboran Slab. These data suggest that, as beneath the Rif, the delaminated crust and lithospheric mantle were replaced by asthenosphere, providing support for the Sierra Nevada (Figure 3.10b).
Figure 3.10: a) Shows an isostatically balanced crustal column for elevations of 3 km observed in the Sierra Nevada and the Atlas Mountains. Using a reference crustal density of 3825 kg/m³ and a reference lithospheric mantle density of 3300 kg/m³ this figure shows that these elevations should be supported by a ~19 km crustal root. b) Shows an isostatically balanced crustal column beneath the Sierra Nevada where we observe a Moho depth of ~32 km. The average crustal density used for this calculation was determined using the velocities resulting from the joint inversion and the velocity-density relationships presented by Zoback and Mooney (2003) modified from Christensen and Mooney (1995). Figure 3.7b shows low velocities below the Sierra Nevada extending down to ~75 km depth. Assuming these velocities represent hot, buoyant asthenosphere supporting the Sierra Nevada elevation, then we calculate a sub-crustal /asthenospheric density of 3240 kg/m³ required to achieve compensation at 75 km depth.
The Granada Basin lies just west of the Sierra Nevada at the edge of the attached slab (Figure 3.7b and 3.8). This basin is known to have a high concentration of seismic activity along its eastern margin (Perez-Pena et al., 2010; Galindo-Zalvidar et al., 1999). These earthquakes occur along high-angle normal faults (Galindo-Zalvidar et al., 1999), which we speculate may be accommodating the nearly vertical, asthenospheric driven uplift of the Sierra Nevada relative to the Granada Basin. We suggest that the Alboran Slab is still attached and exerting a vertical slab-pull force beneath the western portion of the Granada Basin. Partial slab detachment beneath the basin could explain uplift of the basin since the late Miocene, but not to the extent seen in the adjacent Sierra Nevada where the slab is already completely detached.

3.5.3 LAB Beneath the Atlas Mountains

The Atlas Mountains currently accommodate NNW-SSE Africa-Iberia convergence by thrust reactivation of normal faults formed during a Mesozoic rifting event (Beauchamp et al., 1999). Previous studies have suggested that, despite its high topography, shortening within the Atlas is modest, ranging from 10-25% (Beauchamp et al., 1999; Teixell et al., 2003; Arboleya et al., 2003). Figures 3.4a and 3.6a indicate shallow Moho depths ranging from ~33-38 km beneath both the Middle and High Atlas, consistent with Moho depth estimates from active source results (Ayarza et al., 2014). This suggests that the Atlas Mountains are not
isostatically compensated by a crustal root to match their high elevations (Figure 3.10a). Crustal thickening, therefore, does not fully explain the high elevations of the Atlas, suggesting a mantle component of buoyancy has contributed to their uplift (Missenard et al., 2006; Teixell et al., 2003).

Previous studies have used elevation, geoid, surface heat flow, and seismic data to model the LAB beneath the Atlas (Fuella et al., 2010; Zeyen et al., 2001). These authors indicate LAB depths of < 100 km and suggest that a NE-SW trending corridor of thin lithosphere places buoyant asthenospheric material at shallow depths beneath the Atlas, supporting the high elevations. Gravity data are consistent with thin lithosphere beneath the Atlas as Free Air anomalies of + 100 mGal suggest the region is not isostatically balanced (International Gravimetric Bureau).

Our results show evidence for velocities as low as ~4.1 km/s present at subcrustal depths down to 120 km beneath the Middle and High Atlas (Figure 3.6). We interpret these low velocities as buoyant asthenosphere (Figure 3.7d). The LAB map (Figure 3.4b) shows thin lithosphere between -8° and -4° beneath the Middle and High Atlas. Our LAB depth estimates in this region range from ~40 to 60 km, with the thinnest lithosphere associated with a high concentration of Cenozoic volcanic activity in the Middle Atlas and the region directly north (Figure 3.4b and 3.7d). Our LAB depth estimates are shallower than those suggested in previous modeling studies (Fuella et al., 2010). However, they agree well with recent surface wave tomography results (Palomeras et al., 2014) that also indicate low velocities and thin lithosphere beneath the Middle and High Atlas, which Bezada et al. (2013) suggest have resulted from delamination of Moroccan lithosphere.
3.6. Conclusions

We performed a joint inversion of Ps receiver functions with ambient noise and ballistic Rayleigh wave phase velocity data to develop a quasi 3-D shear velocity model beneath 242 broadband seismic stations deployed from central Spain through Morocco to the Sahara Desert. The resulting shear velocity data show a thickened crust throughout the western Gibraltar Arc, with the nearly vertical high velocity Alboran Slab directly beneath and contiguous with the thickened crust. The crust thins abruptly to the east under both Iberia and North Africa where the top of the Alboran slab descends into the mantle, with the intervening space filled by low shear velocities indicative of asthenosphere. We interpret this point of thinning crust and deepening Alboran Slab as the westernmost point of lithospheric delamination, suggested in previous studies (Duggen et al., 2004; 2005; Palomeras et al., 2014; Thurner et al., 2014; Levander et al., 2014). The images in Figures 3.7c and 3.9 also suggest that the crust is involved in delamination beneath the Rif. We suggest that this slab detachment and simultaneous asthenospheric upwelling supports the high elevations of the Sierra Nevada in the central Betic Mountains. These results are summarized in the 3-D cartoon shown in Figure 3.11.

Beneath the Atlas Mountains we observe thin lithosphere (~40-60 km) thin crust (~33-38 km), and a weakly developed mantle lid, i.e., one having slightly higher than asthenospheric velocity (4.2-4.3 km/s). Beneath this is a well-developed asthenosphere, with velocities of 4.1-4.2 km/s. The thin crust and lithosphere are
consistent with previous studies, and suggests that here also, buoyant asthenosphere supports much of the topography.

**Figure 3.11:** 3-D cartoon showing the geometry of the slab beneath the Gibraltar Arc modified from Figure 2.14 and Thurner et al., 2014. The colored squares indicate where the slab is attached and detached (same as Figure 3.8) as determined from the joint inversion of both receiver function and dispersion datasets.
Chapter 4

Ps Receiver Function Evidence for Crustal Scale Thrusting, Relic Subduction, and Mafic Underplating in the Trans-Hudson Orogen and Yavapai Province

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Abstract

The Trans-Hudson orogen (THO) in the north central United States represents a major suturing event between the Wyoming and Superior Archean provinces. It is bounded to the south by the NE-SW striking Yavapai province, which was accreted along the southeastern margin of North America between ~1.71 and
~1.68 Ga and was one of a series of major collision events responsible for the assembly of Laurentia. In this study, Ps teleseismic receiver functions were used to investigate the deep crustal structure associated with these collisions. Using data from over 800 broadband seismic stations distributed throughout the Great Plains/Midcontinent region, we calculated 0.5 Hz receiver functions using 245 M > 6.0 teleseismic events. The receiver functions were then CCP (common conversion point) stacked to create a 3D image volume. Profiles through this image volume show evidence of crustal scale thrusting of the Wyoming province in the west over the Superior province in the east and a relic subduction zone associated with the Yavapai-Superior boundary. We also performed a density analysis of the region using the 2p1s and 0p1s receiver function phases from 233 stations. These data indicate a relatively low Moho density contrast throughout the THO and northern Yavapai Province associated with a region of thickened crust (> 50 km), which we interpret to be evidence of a dense lower crustal layer that is the result mafic underplating.

4.1 Introduction

The Trans-Hudson Orogen (THO) is a Paleoproterozoic collisional belt that extends from the Hudson Bay through the Canadian provinces of Manitoba and Saskatchewan and south into the U.S. through western Montana and North and South Dakota (Figure 4.1a). This orogen developed during the formation of the North American cratonic core between 2.0 and 1.8 Ga by a series of plate collisions
between Archean continents (Whitmeyer and Karlstrom, 2007). It contains juvenile Paleoproterozoic terranes caught within these collisions as well as the reworked Archean margins of the Superior, Hearne, and Wyoming provinces, which bound the THO to the east, northwest, and southwest respectively.

**Figure 4.1:** a) Geologic map showing the distribution of geologic terranes associated with the formation of the North American continent. The black square indicates the focused study region, the white bold line marks the location of the Lithoprobe profile across the Trans-Hudson Orogen. b) Topography and station distribution map showing the location of all broadband stations for which receiver functions were calculated.
The Trans-Hudson Orogen is bounded to the south by the Yavapai Province, which is primarily comprised of juvenile arc crust and was accreted along the southeastern margin of North America between 1.71 and 1.68 Ga (Whitmeyer and Karlstrom, 2007). This accretion event, known as the Yavapai orogeny, was one in a series of accretion events that progressively built the North American continent to the southeast between ~1.84 and ~0.95 Ga (Hoffman, 1988; Whitmeyer and Karlstrom, 2007).

The THO, comparable to the modern Himalayas in scale (Whitmeyer and Karlstrom, 2007), was an integral part of the tectonic evolution of North America. The modern lithospheric structure of this region, largely unaffected by subsequent tectonism, provides insight into the formation of a stable craton as well as Proterozoic plate tectonics. For these reasons the Canadian Trans-Hudson Orogen was extensively studied during the Lithoprobe program, which consisted of numerous seismic reflection and wide-angle seismic surveys (Hajnal, 1997; Nemeth and Hajnal, 1998; Corrigan et al. 2005; Nemeth et al., 2005; White et al., 2005). The U.S. component of the orogen, however, has not received as much attention, in part due to the lack of basement exposure within the U.S. where outcrops are restricted to the Black Hills (Dahl et al., 1999). Latham et al. (1988), Baird et al. (1996), Nelson et al. (1993) and Klasner and King (1990) have all discussed results from a series of COCORP seismic reflection profiles crossing the Trans-Hudson Orogen just south of the U.S. – Canada border. Recent and more extensive investigations of this region, however, do not exist. There have been, however, numerous studies to the west of the U.S. Trans-Hudson Orogen including the COCORP, Deep Probe, and CD-ROM
experiments, which reveal visible subduction zone scars across the WY-Proterozoic boundary (Cheyenne Belt) (Yuan and Dueker, 2005), an indistinct Moho west of the Trans-Hudson Orogen (Allmendinger et al., 1982; Baird et al., 1996; Braile et al., 1989; Brown et al., 1983; Nelson et al., 1993), possibly resulting from a gradational phase change (Smithson, 1989), and a high velocity lower crustal layer within the Wyoming province, which may be the result of Proterozoic underplating and metamorphism of the Archean crust (Braile, 1989; Gorman, 2002; Snelson et al., 2005; Karlstrom et al., 2005; Clowes et al., 2002).

Using data from the USArray Transportable Array, we present a Ps receiver function analysis of the crustal structure throughout the central U.S., including the southern extension of the Trans-Hudson Orogen and the Proterozoic accreted terranes to the south. At 70 km station spacing this network provides an excellent opportunity to image the deep crustal structure associated with continental suturing processes.

4.2 Geologic Setting

The northern Trans-Hudson Orogen is a suture between the Archean Hearne Province and the Archean Superior Province, which collided after the closure of the intervening Manikewan Ocean (Hammer et al., 2010). Ocean closure began ~1.92 Ga and the resulting subduction initiated the formation of island arc and ophiolite complexes within the ocean basin, which were accreted onto the Hearne passive
margin by both west and east dipping subduction processes (Corrigan et al., 2005).
By ~1.835 Ga the northward moving Sask craton, a much smaller Archean continent, collided with the Hearne craton and the accreted terranes along its margin (Hammer et al., 2010). The advancement of the Superior craton to the east closed the remaining ocean basin and the terminal collision between the Hearne craton, the Superior craton, and the accreted terranes in between was complete by ~1.65 Ga (Hammer et al., 2010; Corrigan et al., 2005).

The southern Trans-Hudson Orogen is a suture between the Archean Wyoming Province and the Superior Province. This collision began ~1.77-1.715 Ga, 50-60 My after the northern collision (Dahl, 1999). Unlike in the northern extension of the Trans-Hudson Orogen, there is little isotopic evidence for early Proterozoic juvenile volcanic terranes within the southern segment, as drill core information suggests a crust dominated by Archean age granulitic material (Baird et al., 1996). Some terrains in the southern portion, however, do resemble those in the north, including the reworked Archean margins (Klasner and King, 1990). Additionally, some studies have suggested the presence of an Archean micro-continent, similar to the Sask craton, in the northern segment (Baird et al., 1996; Klasner and King, 1990).

Following the formation of the Trans-Hudson Orogen, the North American craton (Laurentia) continued to grow through a series of accretion events occurring between ~1.8 and ~.95 Ga along the southeastern margin, forming the NE-SW trending Yavapai, Mazatzal, Granite-Rhyolite, and Grenville provinces. Both the Yavapai and Mazatzal accretion events were accompanied by the intrusion of
granitoids, which served to “stitch” these terranes to the older crust to the north (Whitmeyer and Karlstrom, 2007). Additionally, there was a major magmatic phase ~1.4 Ga, which Keller et al. (2005) suggest resulted in the thickening of the crust to ~45 km through the production of a mafic restite layer.

4.3 Data and Methods

4.3.1 Data

The USArray Transportable Array is a regular grid of temporary broadband seismic stations deployed across the United States with ~70 km station spacing. Beginning in 2004, ~400 stations were installed along the west coast. With a residence time of ~2 years, these stations were gradually moved west to east across the United States, and by 2013 have covered the entire contiguous United States. For this study, we collected data from ~800 broadband seismic stations located within the Great Plains and Midcontinent region between -110° and -90° longitude (Figure 4.1b). The majority of these stations were a part of the USArray TA deployment. However, we supplemented this dataset with data from four other networks in the region: the US permanent network, the New Madrid seismic network, the Oklahoma seismic network, and the Indiana University seismic network. We calculated Ps receiver functions for 245 magnitude greater than 6.0 events located between 30° and 90° from each station. We also analyzed Bouguer gravity data from the study
region. These data were collected from the Gravity Database of the U.S. operated by the Pan-American Center for Earth and Environmental Studies (PACES).

### 4.3.2 Methods

#### 4.3.2.1 Receiver Functions and CCP Stacking

A receiver function (RF) is a time series computed from a three-component seismogram, with arrivals corresponding to P-wave to S-wave (Ps) conversions generated by discontinuities in the earth beneath a station. It is an approximation of the S wave Green’s function resulting from S conversions from an incident teleseismic P wave (Bostock, 2004; Rondenay, 2009). The positive amplitude arrivals correspond to an increase in seismic impedance with depth (e.g., Moho) and negative amplitudes correspond to a decrease in seismic impedance with depth (e.g., lithosphere-asthenosphere boundary, LAB) (Langston, 1979; Ammon, 1991). This earth structure response can be isolated from the source and instrument response on a seismogram by deconvolving the vertical component from the horizontal component (radial and tangential). In this study, Ps RFs were calculated using both water-level frequency domain deconvolution (Burdick and Langston, 1977) and time-domain iterative deconvolution (Ligorria and Ammon, 1999). We show results from the Water Level method, which uses a simple division of the radial component by the vertical component in the frequency domain. However, noise can cause this division to become unstable when the vertical component
spectrum reaches very small values. The Water Level method employs a type of stabilizing regularization in order to perform the division. A Gaussian filter is also used to control the width of the pulse and the high corner frequency of the resulting RF (Ammon, 1991).

We calculated receiver functions with a high corner frequency of 0.5 Hz. Assuming an average lower crustal Vs of ~4 km/s and a vertical resolution of ~0.25\(\lambda\), this allows for the resolution of vertical structure at the ~2 km scale. All receiver functions were visually inspected and those with a low signal to noise ratio or otherwise flawed by dead traces or noise spikes were removed. In total, we were left with 41,737 high quality receiver functions, with an average of ~50 traces for each station. The common conversion point stacking technique (Dueker and Sheehan, 1997) was then applied to the data using the depth conversion and lateral migration as outlined in Levander and Miller (2012). A 3-D image volume spaced 0.25° X 0.25° laterally and 1 km vertically was created and conversion horizons were picked directly from this volume. The depth conversion was performed using the IASP91 1-D velocity model modified with the Crust2.0 model at shallow depths (Bassin et al., 2000).

### 4.3.2.2 Density Analysis

Receiver functions are commonly used to determine Moho depth from the P to S converted phase, which is most often the dominant phase on a Ps receiver function. Crustal reverberations, however, can be used to better constrain Moho
depth and Vp/Vs ratio, as well as the density contrast associated with the crust-mantle boundary. Niu and James (2002) show that the density contrast across the Moho can be estimated from the relative amplitude of the Moho conversion and reverberation phases. We employ this technique to determine the relative density contrast at the Moho for stations distributed throughout the entire Great Plains/Midcontinent region. Receiver functions at each station were stacked using an n-th root stacking method with n = 2 (Muirhead, 1968). Depending on the Moho depth, the depth to other discontinuities beneath a station, and the signal to noise ratio, the reverberated phases may not always be easily identified on a receiver function. We, therefore, analyzed the receiver function stacks for each station, eliminating those with a poor signal to noise ratio and those without a clear 2p1s reverberated phase. We then calculated the ratio of the 2p1s reverberated phase amplitude and the 0p1s Moho amplitude at 233 stations where a clear 2p1s phase is observed.

4.3.2.3 Gravity Modeling

Deep crustal and lithospheric density variations cause gravity anomalies with long to intermediate wavelengths: ~500 km for lithospheric mantle thickness and density variations, and greater than ~250 km for Moho depth and lower crustal density variations (Tiwari et al., 2013; Watts and Daly, 1981). Wavelength filtered gravity data can, therefore, provide valuable information about deep crustal and lithospheric structure. Additionally, the spectrum of potential field data has been
used in previous studies to determine the source depth of gravity and magnetic anomalies (Tiwari et al., 2013; Fedi et al., 1997; Bilim, 2007). A spectral analysis of the Bouguer gravity data in the Great Plains/Midcontinent study region reveals the deepest source depth to be ~140 km, roughly coincident with lithosphere-asthenosphere boundary depth estimates from surface wave tomography results in the same region (Margolis et al., submitted 2014). This source depth corresponds to a cut off gravity wave number of ~0.0025 km⁻¹ and wavelength of ~400 km. We filtered the Bouguer gravity data with a low pass filter, resulting in gravity anomalies with wavelengths of ~310 km that can be attributed to lower crust and uppermost mantle density and crustal thickness variations. We then used the GM-SYS Gravity/Magnetic Modeling Software to forward model the filtered gravity data along a W-E profile across the Trans-Hudson Orogen at 47° N (Figure 4.5).

**4.4 Receiver Function Observations**

After generating a 3D CCP image volume, the Moho depth was determined by averaging picks on latitude and longitude profiles throughout the region. The average Moho depth for the entire Great Plains/Midcontinent region is ~45 km. The Moho depth map in Figure 4.2, however, shows a much thicker crust (up to ~55 km) beneath the Trans-Hudson Orogen and northern Yavapai boundary east of the Rocky Mountains. The area outlined by the dashed line in Figure 4.2 represents a region where the Moho appears to shallow abruptly towards the east from ~50 km to ~30 km depth over ~150 km distance. This abrupt shallowing of the Moho seems
geologically unrealistic for a stable continental region that has not undergone
tectonic deformation since the Precambrian. However, a clear Moho signal below 30 km cannot be identified in the receiver functions. This region also coincides with high Rayleigh wave phase velocities (20-25s, Ge Jin, http://www.ldeo.columbia.edu/~ge.jin/projects/USarray.html) and high shear wave velocities determined from both Finite Frequency Rayleigh Wave Tomography and Ambient Noise Tomography. These studies indicate positive Vs anomalies of ~3-6% at 40 km depth (Margolis et al., Submitted 2014), elevated Vs within the crust at ~30-32 km depth, and Vs as high as 4.5 km/s at 38-40 km depth (Ryan Porter, personal communication).

![Figure 4.2: Map showing Moho depth (km). The bold white lines show the locations of profiles shown in Figures 4.3 and 4.4. The thin black lines represent the terrane boundaries as shown in Figure 4.1. The black dashed line outlines the region discussed in section 4.](image)
4.4.1 Trans-Hudson Orogen

In order to determine the structure associated with the transition from the Trans-Hudson Orogen and the bounding Archean provinces, we examined latitude CCP profiles roughly perpendicular to the strike of the orogen. Figure 4.3 shows two of these profiles at 48.75° and 47° N. In the northern profile (Figure 4.3a) we see a strong positive (red) event at ~37-41 km depth beneath the western Trans-Hudson Orogen. This Moho event is shallower than that seen on average throughout the Great Plains. However, it is consistent with both the COCORP seismic profile (Baird et al., 1996) and Ps and Sp receiver functions (Keenan et al., 2012), which show a relatively shallow Moho (~40 km) in this same region. This Moho event deepens to ~50 km beneath the central part of Trans-Hudson Orogen. Beginning at ~102° W, we observe a second strong positive event within the crust, extending from ~102° to ~98° W and dipping east to west from ~25 km depth to ~50 km depth where it merges with the Moho signal. The deeper, Moho event is observed east of 101° W at ~48-50 km depth beneath the Superior Province. Figure 4.3b shows a CCP profile at latitude 47° N. Here we see a single positive event throughout the entire profile. This signal drastically decreases in depth in the center of the profile between 102° to 99°W where we observe the dipping event in the northern profile, suggesting that the two features are of the same origin.
**Figure 4.3:** Profiles through the CCP image volume showing crustal scale thrusting feature across the Trans-Hudson Orogen at 48.75° N (a) and 47° N (b). The blue lines indicate topography along each profile, the black lines indicate the Bouguer gravity anomaly along each profile, and the red triangles indicate stations with a low Moho density contrast as shown on the Moho map.

**4.4.2 Northern Yavapai Boundary**

We also examined the crustal structure associated with the boundary between the Yavapai province and the bounding Archean provinces to the north. Figure 4.4 shows two diagonal profiles perpendicular to this boundary. In the first profile, oriented NE-SW, we see a strong Moho signal at ~45 km depth to the northeast (Event A, Figure 4.4a). This signal shallows to the southwest to ~33 km depth in the central portion of the profile where we see two additional strong
positive events, one shallow between ~18 and 31 km depth dipping to the southwest (Event B, Figure 4.4a) and one deep between ~50 and 60 km depth dipping to the northeast (Event C, Figure 4.4a). In the southwestern portion of the profile we again see one positive Moho signal at ~ 43 km depth (Event D, Figure 4.4a).

**Figure 4.4:** Diagonal profiles through the CCP image volume showing evidence of relic subduction associated with Yavapai accretion. The dashed line indicates where obduction of the upper crust may have occurred and the dotted line indicates deep, subcrustal event that may be evidence of relic subduction and structural underplating of oceanic crust. The bold letters indicate the events discussed in section 4.2 of the text.

We see a very similar structure in the second profile oriented NW-SE. There is a strong shallow signal in the northwest at ~ 30 km depth (Event A, Figure 4.4b). Again, in the central portion of the profile we see multiple positive events between
~15 and 60 km depth (Events B and C, Figure 4.4b) and a single Moho event in the southeast at ~40 km depth (Event D, Figure 4.4b).

4.5. Discussion

4.5.1 WY-Superior Suture Zone

We obtain crustal depth estimates ranging from ~47-55 km beneath the Trans-Hudson Orogen. These estimates are slightly deeper than those based on COCORP reflection data by Baird et al (1996) who propose a crustal model across the Trans-Hudson Orogen with Moho depths extending from ~42 to 50 km depth. However, the deeper Moho depth estimates of this study are consistent with those from previous studies of the northern portion of the Trans-Hudson orogen. Hajnal et al (1984), for example, report crustal thicknesses greater than 50 km from COCRUST refraction profiles just north of the U.S. – Canada border across the Williston Basin and Trans-Hudson Orogen. Results from the Lithoprobe project also indicate Moho depths greater than 50 km beneath the western portion of the Trans-Hudson Orogen (White et al, 2005; Hammer et al, 2010).

We interpret the dipping crustal events, seen in Figure 4.3 as evidence of crustal scale overthrusting. Previous Lithoprobe and COCORP studies (White et al, 2005; Hammer et al, 2010; Baird et al., 1996; Nelson et al., 1993) also suggest crustal thrusting at the same scale. The Lithoprobe results to the north indicate an antiformal crustal culmination cored by the Sask micro-continent with strong
reflections dipping beneath both bounding Archean cratons (Lewry, 1994). Baird et al. (1996) and Nelson et al. (1993) present similar results from a COCORP seismic reflection transect just south of the U.S.-Canada border. In this study we observe strong, dipping signals in the crust consistent with thrusting of the Wyoming craton in the west over the Superior craton in the east.

It is likely that the strong, dipping crustal events represent the decollement or suture zone between the Superior craton and the overriding Wyoming craton. We suggest that this structure is analogous to the Kapuskasing uplift in the central Superior province. Thought to be a result of Trans-Hudson orogen compression, this structure consists of mid-to lower crustal rocks that have been brought to the surface (Percival, 1986; Clowes, 1992, Cook et al., 1994). Another analogous location may be in the Athabasca terrane to the north, in the western Churchill province of central Canada, where the lower crust of the Rae domain was thrust above the middle crustal rocks of the Hearne domain during the early stages of the Trans-Hudson Orogeny (Dummond, 2008; Williams and Hanmer, 2006).

### 4.5.2 Relict Yavapai Subduction

We interpret the deep, positive event in Figure 4.4 to be evidence of a relic subduction zone. This signal extends from Moho depth to >60 km and dips from south to north. This is indicative of north directed subduction of the Yavapai Province beneath the Superior Craton. We suggest that the deep positive event seen in Figure 4.4 is analogous to a “Type III reflection pattern near the Moho” described
by Cook et al. (2002) as “reflections that can be traced from the lower crust to beneath the reflection Moho”. A similar mantle signal can be seen in numerous studies, including the seismic reflection profiles of the Abitibi-Grenville segment of the Lithoprobe project, which crosses the Grenville front at the Kapuskasing structural zone (Cook et al., 1999; Cook et al., 2010; Calvert et al., 1995). Yuan and Dueker (2005) also present evidence for a high velocity, north dipping, ancient slab preserved beneath the Cheyenne belt along the southern boundary of the Wyoming Province. We suggest that this north directed Yavapai subduction zone was preserved as oceanic crust underthrust the Superior province and became partially eclogitized as described by Cook et al. (1999). The shallow positive arrival may be indicative of a large scale crocodile structure where separation of the upper and lower crust occurs when the upper layer is obducted (Event B, Figure 4.4) and the lower layer is subducted (Event C, Figure 4.4) as seen in previous studies of oceanic arcs (Lizarralde et al., 2002; Nakanishi et al., 2009, Thybo and Artemieva, 2013).

4.5.3 Gravity Modeling and Mafic Underplating

Figure 4.5 shows the results of the forward modeling of the long-intermediate wavelength filtered Bouguer gravity data where we assume an asthenospheric density of 3.25 g/cm³ and a lithospheric mantle density of 3.30 g/cm³. The lithosphere-asthenosphere boundary was taken from surface wave tomography results (Margolis et al., submitted 2014). We are able to model the observed Bouguer gravity data (RMS Error = 0.276) using a lower crustal density of
2.95 g/cm³ throughout the profile and a layer of 3.1 g/cm³ below the shallow positive event in the central and eastern portion of the profile. The absence of a consistent positive signal at the base of this layer, suggests that the density increase throughout the layer is gradational in nature.

![Diagram of gravity anomaly and CCP profiles](image)

**Figure 4.5:** Filtered Bouguer gravity anomaly and CCP profiles between 0 and 75 km depth. Along with density model used to forward model the intermediate (~300 km) wavelength Bouguer anomaly. The red dots indicate observed data and the black line indicates the forward modeled gravity anomaly. LAB depth taken from Margolis et al. (submitted 2014).

Also shown on this profile, directly above the dense layer, are the locations of two stations for which a relatively low Moho density contrast value was measured using the 2p1s/0p1s ratio. The gravity modeled dense layer, low Moho density contrast, and the seemingly transparent lower crust occur in the same region where
anomalously high S-wave velocities were found with ambient noise tomography (Ryan Porter, personal communication), providing compelling evidence for mafic underplating in this region. Thybo et al. (2013) discuss similar crustal structures seen at other Precambrian suture zones including the tectonic boundary between the Hearne and Wyoming provinces, imaged in the Lithoprobe seismic refraction profile SAREX (Clowes et al., 2002; Gorman et al., 2002), and the boundary in the Baltic Shield between the Archean Karelian craton and the adjacent Proterozoic mobile belt. In both cases, dense, high velocity lower crustal layers have been attributed to Proterozoic magmatic underplating, which produced a dense lower crust consisting of a mixture of mafic granulites, pyroxenites, and eclogites (Kuusisto et al., 2006; Thybo et al., 2013). This is consistent with Durrheim and Mooney (1991; 1994) who find that Proterozoic crust (~40-55 km thick) is generally thicker than Archean crust (~27-40 km thick) and contains a much thicker high-velocity (>7.0 km/s) lower crustal layer, which they attribute to basaltic underplating facilitated by a decline in mantle temperatures in the Proterozoic.

4.5.4 Preservation of a Dense Lower Crust

We calculated the 2p1s/0p1s ratio at 233 stations distributed throughout the study region (Figure 4.6) and found an average ratio of .4529. We used the calculated ratios to qualitatively assess the relative Moho density contrast between various tectonic and geologic terranes throughout the region. In the Rocky Mountains we calculate an average ratio of .4775, although we observe highly
variable ratios throughout. This is consistent with a recently deformed region where crustal structure is highly variable over short distances. We calculate a similar average value of .4561 in the Mazatzal and Granite-Rhyolite provinces, which have been relatively stable and undeformed since the Paleozoic. In the Rio Grande Rift we observe a high average moho density contrast of .6401, consistent with significant crustal thinning that has occurred in this region where a dense, high velocity lower crustal layer does not exist (Sinno et al., 1986). In contrast, we observe a relatively low ratio in the Yavapai province (.3904) and an even lower ratio in the Archean terranes (.3448). Additionally, we calculated the average 2p1s/0p1s ratio within the region of thickened crust shown in the Moho map (Figure 4.3) and outlined in Figure 4.6. Within this region we observe the lowest average ratio of .3093, indicating a low-density contrast at the Moho.

We, therefore, suggest the presence of a dense lower crustal layer throughout the Trans-Hudson region and the Yavapai province directly to the south. This layer may be the result of structural underplating and/or mafic underplating as described by Cook et al. (2010). The profiles in Figure 4.4 suggest structural underplating associated with Yavapai accretion.
**Figure 4.6:** Map showing the stations used for the Moho density contrast calculation and the regions for which average ratios were calculated. RM = Rocky Mountains, RGR = Rio Grand Rift, GR=Granite-Rhyolite.

<table>
<thead>
<tr>
<th>Location</th>
<th>2p1s/0p1s Ratio</th>
<th>Moho Density Contrast</th>
</tr>
</thead>
<tbody>
<tr>
<td>Thickened Crust</td>
<td>.3093</td>
<td>Low</td>
</tr>
<tr>
<td>Archean Terranes</td>
<td>.3448</td>
<td>Low</td>
</tr>
<tr>
<td>Yavapai Province</td>
<td>.3904</td>
<td>Average</td>
</tr>
<tr>
<td>Mazatzal/Granite-Rhyolite Province</td>
<td>.4561</td>
<td>Average</td>
</tr>
<tr>
<td>Rocky Mountains</td>
<td>.4775</td>
<td>Average</td>
</tr>
<tr>
<td>Rio Grand Rift</td>
<td>.6410</td>
<td>High</td>
</tr>
</tbody>
</table>

In this case, subducted oceanic crust may have been emplaced below the continental crust during subduction and may have subsequently undergone partial eclogitization (Cook et al., 2010). However, it is also likely that magmatic
underplating played a role in the formation of the dense lower crustal layer in the THO-Yavapai region. Accretion of the Proterozoic terranes along the southeastern margin of Laurentia was accompanied by plutonism and voluminous intrusion of granitoids (Whitmeyer and Karlstrom, 2007), which may have resulted in the emplacement of dense high-velocity basaltic intrusions into the crust as described by Keller et al. (2005). The relative abundance of this dense phase and the degree of eclogitization, however, may be variable throughout the region. We suggest the area with relatively thin crust to the east of the Trans-Hudson orogen underwent more complete eclogitization in order to account for the high shear wave velocities as well as the “unrealistic shallowing of the Moho”, two characteristics which Mjelde et al. (2012) and Thybo et al (2013) present as evidence for the presence of mafic underplating and eclogitization. The 2p1s/0p1s ratios suggest mafic underplating over a large region including much of the Williston Basin, both the Wyoming and Superior Archean terranes, and a portion of the Yavapai province. We speculate that the widespread emplacement of such a dense lower crustal layer along with tectonic quiescence since the Precambrian played a role in the stabilization of this portion of the craton.

4.6 Conclusions

The Trans-Hudson orogen in the north central United States formed during a major suturing event between the Wyoming and Superior Archean provinces, and was one of the major collisional events responsible for the assembly of Laurentia.
Receiver function CCP profiles show evidence of crustal scale thrusting of the Wyoming province in the west over the Superior province in the east. The THO is bounded to the south by the Yavapai province which was accreted to the southern boundary of the Archean provinces ~1.7 Ga. Profiles perpendicular to the strike of this boundary show evidence of structural underplating of what we interpret as oceanic crust, i.e. a frozen subduction zone. We also present evidence from receiver functions and gravity data suggesting the presence of a dense lower crustal layer throughout the Trans-Hudson region and Yavapai province. Density analysis using the 2p1s and 0p1s receiver function phases indicates a relatively low Moho density contrast throughout this area with respect to the surrounding Great Plains, Rocky Mountains, and Rio Grand Rift regions. These data suggest mafic underplating and the formation of a dense lower crust throughout the region, which may have contributed to the stabilization of this portion of the craton.
Chapter 5

Estimating Lithospheric Mantle Density in Cratons

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5.1 Introduction

Isostasy is a state of gravitational equilibrium between the earth’s lithosphere and asthenosphere where columns of material exert equal pressure at a compensation depth (Lillie, 1999). The original isostatic models considered loads at the earth’s surface to be compensated locally by variations in crustal thickness and density (Airy and Pratt). The research presented in this chapter applies these models to the earth’s continental lithosphere, where a lithospheric block, consisting of both crust and upper mantle, floats on the asthenosphere. Lachenbruch and Morgan (1990) proposed this
model as a means of determining continental lithospheric thickness. They show that observed elevation, along with estimates of crustal thickness, crustal density, asthenospheric density, and Moho temperature, can be used to estimate the thickness of a lithospheric column floating on asthenosphere. Zoback and Mooney (2003) apply this method to data from a global crustal structure database (Mooney et al., 1998; 2002) and produce a map of global lithospheric thickness estimates. Their method requires an estimate of lithospheric mantle density, derived from heat-flow values, to calculate lithospheric thickness. The method presented in this chapter, however, utilizes seismic data to constrain the lithospheric thickness, allowing for the calculation of lithospheric mantle density.

The availability of seismic and other geophysical data from both active and passive source experiments around the world make this approach possible. The recently released Crust1.0 model (Laske et al., 2013), for example, provides global crustal structure data in unprecedented detail, allowing for a more reliable estimation of crustal thickness, velocity, and density. The research presented here takes advantage of the large amount and variety of published seismic studies to determine estimates of average crustal thickness, crustal density, and lithospheric thickness in the earth’s cratons. The goal of this research is to estimate cratonic lithospheric mantle density and provide insight into the isostatic condition of cratons.
5.2 Isostatic Calculations

When considering a lithospheric column floating on asthenosphere, Lachenbruch and Morgan (1990) show that elevation ($\varepsilon$) is related to lithospheric thickness ($L$) by the relative density contrast between the asthenosphere and the lithosphere.

$$
\varepsilon = \left[ \left( \rho_a - \rho_L \right) \times L \right] - H_o \quad \varepsilon \geq 0 \quad (1)
$$

where $\varepsilon$ is elevation above sea level, $\rho_a$ is asthenospheric density, $\rho_L$ is lithospheric density, $L$ is lithospheric thickness, and $H_o$ is height of sea level above asthenosphere with no overlying lithosphere. Furthermore, they show that the total elevation ($\varepsilon$) can be separated into a crustal component ($\varepsilon_c$) and a lithospheric mantle component ($\varepsilon_m$):

$$
\varepsilon_c = \left[ \left( \rho_a - \rho_c \right) \times L_c \right] - H_o \quad (2)
$$

$$
\varepsilon_m = \left[ \left( \rho_a - \rho_m \right) \times L_m \right] \quad (3)
$$

where $L_c$ is crustal thickness, $\rho_c$ is crustal density, $L_m$ is lithospheric mantle thickness, and $\rho_m$ is lithospheric mantle density. Zoback and Mooney (2003) calculated $H_o$ to be 2.78 ± 0.35 km, based on crustal thickness, crustal density, asthenospheric density,
seawater density, and ridge elevation measured at five Mid-Ocean Ridge locations (Weiland and MacDonald, 1996; Stevenson and Hildenbrand, 1996; Madsen et al., 1984; Doin et al., 1996).

5.3 Crustal Thickness vs. Elevation

5.3.1 Tibet - Andes - U.S. Rocky Mountains

In actively deforming, mountainous regions it is expected that the majority of the observed elevation be derived from the crustal buoyancy component, since significant crustal thickening occurs during orogenesis. If this assumption is correct, then equation 2 indicates that we should observe a linear relationship between crustal thickness and observed elevation, where the crustal derived elevation \( \varepsilon_c \) is approximately equal to the observed elevation \( \varepsilon \) (Figure 5.1).

Figure 5.2 shows a plot of crustal thickness vs. elevation data in Tibet, the Andes mountains, and the U.S. Rocky mountains (USRM), taken from the Crust1.0 model. This plot shows a clear linear relationship between crustal thickness and elevation for these recently deformed orogenic regions, confirming the previous assumption. Equation 2 shows that the slope of a line through these data is equal to \( \left( \frac{\rho_a - \rho_c}{\rho_a} \right) \).
**Figure 5.1:** Map showing global elevation taken from the Crust1.0 model. The bold black lines indicate the recently deformed high elevation regions (Tibet, Andes, USRM) discussed in the text. The bold colored lines indicate cratonic regions with Precambrian basement ages (3.5 to 1.7 Ga). Red – Superior craton, Cyan – Kola-Karelian Craton, Baltic Shield, and Russian Platform, Purple – Indian Shield, Green – Pilbara and Yilgarn Cratons.

We can, therefore, estimate the asthenospheric density ($\rho_a$) using the slope of the best-fit line shown in Figure 5.2 and the average crustal density for the three regions ($\rho_c=2.84$ g/cm$^3$). Densities were calculated using the velocities from Crust1.0 and the velocity-density equations shown in Zoback and Mooney (2003), which were modified from Christensen and Mooney (1995). This results in an estimated asthenospheric density of 3.2 g/cm$^3$ beneath Tibet, the Andes, and the U.S. Rocky Mountains. Using this estimated asthenospheric density, we can plot lines of various slopes determined by different assumed average crustal densities. Scattered data falling along these lines
would represent regions where average crustal density deviates from that observed in Tibet, the Andes mountains, and the U.S. Rocky Mountains.

![Graph showing crustal thickness vs. elevation for Tibet, the Andes mountains, and the USRM as outlined in figure 5.1.](image)

**Figure 5.2:** Plot of crustal thickness vs. elevation for Tibet, the Andes mountains, and the USRM as outlined in figure 5.1. Data is taken from the Crust1.0 model. Bold black line is a best fit line for the scattered data points with a y-intercept of -2.78 km as calculated by Zoback and Mooney (2003). Thin black lines represent various crustal thickness-elevation relationships assuming different average crustal densities.

### 5.3.2 Stable Cratons

The results from Chapter 4 show a thick crust (up to 50 km) associated with low elevations in a portion of the North American craton located in the central U.S. These
observations are consistent with those made in other cratonic regions (3.5-1.7 Ga) around the world, including those found in Scandinavia, southern India, and western Australia (Figure 5.1). Crustal thicknesses in these regions range from ~30 to ~ 55 km, while elevations are restricted to less than 1 km. The results from Chapter 4 also indicate that the lower crust in the Trans-Hudson Orogen and surrounding cratonic region is denser than that found in the younger regions to the south and west. It is conceivable that a dense lower crustal layer could account for the low elevations observed coincident with large crustal thickness estimates in the North American craton and other cratons around the world. Additionally, this would be consistent with the results from Kelly et al. (2003) whose geochemical analysis suggests that there must be dense layers within the cratonic crust or upper mantle in order to achieve neutral buoyancy in cratonic lithosphere. In this research, we use the Crust1.0 model to investigate the role that a thick, dense crust might play in producing the observed low elevation in cratons.

Figure 5.3 shows a plot of crustal thickness and elevation data for the four cratonic regions shown in Figure 5.1: central North America, Scandinavia, southern India, and western Australia. These data are shown by the red, cyan, purple, and bright green colors, respectively. This figure agrees well with a similar figure shown by Zoback and Mooney (2003). It indicates that there is no correlation between crustal thickness and observed elevation in the cratons. That is, the linear relationship that is clearly visible in the data from recently deformed mountain ranges does not hold true in the cratons.
**Figure 5.3:** Plot of crustal thickness vs. elevation for the four cratonic regions shown in Figure 5.1. Red – North America Superior craton, Purple – Indian Shield, Cyan – Scandinavia, Bright Green – western Australia. This figure shows that these data do not fall along any of the lines representing reasonable variations in crustal density. Also shown are data for the portion of the Wyoming province deformed during the Laramide (pink), the portion of the Wyoming province adjacent to the Rocky mountain deformation front (dark green), and the western portion of the Yavapai province adjacent to the Rocky mountain deformation front (yellow). These data show that recent deformation has affected the crustal thickness vs. elevation relationship in these regions.

Figure 5.3 also shows crustal thickness vs. elevation data for the Wyoming craton and the portion of the Yavapai province just west of the Superior craton. The pink data points represent the portion of the Wyoming province contained within the Rocky mountain deformation zone. The dark green and yellow data points represent the portions of the Wyoming and Yavapai provinces just east of the deformation zone.
These data show that, although the Wyoming province and western Yavapai province are of Precambrian age, they exhibit much different behavior than the other cratonic regions shown. These differences are likely the result of the more recent tectonic deformation that has affected both provinces. The Wyoming province was affected by the Laramide orogeny, responsible for the uplift of the Rocky Mountains. Additionally, Margolis et al. (submitted EPSL 2014) show that the Wyoming and western Yavapai provinces are underlain by a relatively thin lithosphere (~100-120 km). They suggest that asthenospheric expansion below this thin lithosphere may contribute up to ~1 km of elevation in these regions. On the other hand, the adjacent Superior craton has experienced very little tectonic deformation since the Precambrian. These differences suggest that the crustal thickness – elevation relationship, observed for the cratonic regions shown in Figure 5.3, is not necessarily derived from some inherent property of Precambrian cratonic crust.

5.4 Lithospheric Mantle Derived Elevation

Figure 5.3 shows that variations in crustal density and/or thickness cannot account for the observed elevations in cratons. The crustal component of elevation does not dominate in the cratons as it does in recently deformed, mountainous regions. The lithospheric mantle, therefore, must compensate for a portion of the observed elevation in these cratonic regions.
The crustal derived elevation in the cratons can be calculated using equation 2. Figure 5.4 shows this result, calculated using crustal densities derived from Crust1.0 velocities and a reference asthenospheric density of 3.3 g/cm³ (Lee et al. 2012).

**Figure 5.4:** Map showing crustal derived elevation calculated using calculated crustal densities and a reference asthenospheric density of 3.3 g/cm³

Figure 5.4 shows a consistent overestimation of elevation in the cratons from the crustal component alone. We conclude, therefore, that the lithospheric mantle must contribute a negatively buoyant component to the observed elevation. The amount of lithospheric mantle derived elevation can be calculated by subtracting the crustal component of elevation from the observed elevation. This result, shown in Figure 5.5,
indicates that the lithospheric mantle contributes up to -3 km to observed elevation in the cratons.

**Figure 5.5:** Map showing lithospheric mantle derived elevation calculated using a reference asthenospheric density of 3.3 g/cm³ and crustal densities derived from Crust1.0 velocities. This figure shows a negative contribution from the lithospheric mantle in the cratonic regions discussed above.

5.5 Determining Lithospheric Mantle Density

Seismic data can be used to provide constraints on lithospheric thickness beneath the cratons. Figure 5.6 shows a map of lithospheric thickness estimates for the four cratons of interest, derived from previous seismic studies. In the North America
Superior Craton, LAB depth estimates were taken from two studies. Darbyshire et al. (2010) carried out a two-station phase velocity analysis of teleseismic Rayleigh waves using broadband stations from the POLARIS/FedNor seismic network and the Canadian National Seismograph Network located in central Ontario and western Quebec. They measured dispersion curves for 100 two-station paths throughout the region providing phase velocity data for periods from ~25 to 170 s. They then inverted for shear velocity at 30 stations, providing LAB depth estimates defined by velocity perturbations within 1-D shear velocity models extending down to ~300 km. Other studies define the LAB as the depth at which the direction of seismic anisotropy changes from one associated with lithospheric deformation to one associated with the flow of asthenosphere in the direction of absolute plate motion (Yuan et al., 2011; Plomerova and Babuska, 2010).

Yuan et al. (2011) developed a 3-D upper mantle model of North America, which includes isotropic shear velocity as well as radial and azimuthal anisotropy. They define the LAB as described above and provide an LAB depth map for all of North America. We used these LAB depth estimates for the portion of the Superior craton not discussed by Darbyshire et al. (2010).

In the cratonic region of eastern Europe and Scandinavia, we use the LAB depth estimates from Plomerova et al. (2002, 2006, 2008) summarized by Jones et al. (2010). These LAB depth estimates are defined by a change in the direction of seismic anisotropy, as described above. The LAB depth estimates presented in a global study by Plomerova et al. (2002) were used in the Russian Platform where results from local seismic studies were not available. In the Indian Shield we used the LAB depth estimates provided by Kumar et al. (2013), who calculated Ps receiver functions at 59
broadband seismic stations deployed throughout southern India. In western Australia we use LAB depth estimates from Kennett and Blewett (2012) who summarize the results from a number of previous tomography studies in the region.

![Figure 5.6: Map of lithospheric thickness for the four cratonic regions of interest. LAB depth estimates were compiled from previous local and global seismic studies.](image)

With estimates of average lithospheric thickness for each cratonic region, the average lithospheric mantle density can be calculated using the average crustal density, derived from Crust1.0 velocities, and the average lithospheric mantle derived elevation (Figure 5.5). These results are summarized in the table below.
Table 5.1

<table>
<thead>
<tr>
<th>Location</th>
<th>$\rho_c$ (g/cm$^3$)</th>
<th>$L_c$ (km)</th>
<th>$L_m$ (km)</th>
<th>$\varepsilon_m$ (km)</th>
<th>$\rho_a$ (g/cm$^3$)</th>
<th>$\rho_m$ (g/cm$^3$)</th>
<th>$\rho_L$ (g/cm$^3$)</th>
<th>% Increase</th>
</tr>
</thead>
<tbody>
<tr>
<td>Superior Craton</td>
<td>2.90</td>
<td>41</td>
<td>159</td>
<td>-1.81</td>
<td>3.3</td>
<td>3.34</td>
<td>3.26</td>
<td>~1.2</td>
</tr>
<tr>
<td>Fennoscandia</td>
<td>2.89</td>
<td>43</td>
<td>148</td>
<td>-2.08</td>
<td>3.3</td>
<td>3.35</td>
<td>3.26</td>
<td>~1.49</td>
</tr>
<tr>
<td>S. India</td>
<td>2.89</td>
<td>38</td>
<td>69</td>
<td>-1.52</td>
<td>3.3</td>
<td>3.37</td>
<td>3.23</td>
<td>~2.08</td>
</tr>
<tr>
<td>W. Australia</td>
<td>2.89</td>
<td>37</td>
<td>157</td>
<td>-1.35</td>
<td>3.3</td>
<td>3.33</td>
<td>3.25</td>
<td>~0.9</td>
</tr>
</tbody>
</table>

Values derived from Crust1.0 Model
Values estimated from previous studies
Values calculated using equations 2 and 3

For each region $\rho_c$ is the average crustal density, $L_c$ is the average crustal thickness, $L_m$ is the average lithospheric mantle thickness, $\varepsilon_m$ is the average lithospheric mantle derived elevation, $\rho_a$ is the assumed asthenospheric density, $\rho_m$ is the calculated lithospheric mantle density, and $\rho_L$ is the calculated density of the entire lithospheric column. The last column represents the percent increase in density of the lithospheric mantle relative to the asthenosphere, thus providing information about lithospheric mantle buoyancy without depending on an assumed reference asthenospheric density.

There is uncertainty involved in this analysis primarily derived from the crust and lithospheric thickness estimations as well as the crustal density estimations. Previous studies have shown crustal thickness and velocity-derived density estimates to have an average uncertainty of 5% (Mooney, 1989; Christensen and Mooney, 1995; Zoback and Mooney, 2003), while seismically derived LAB depth estimates can contribute up to 50 km of uncertainty in lithospheric thickness, depending on the type
of analysis. This study, however, benefits from the quantity of available seismic data, which allows us to perform regional analyses. The goal of this research is to provide information about average lithospheric mantle buoyancy in cratons. By utilizing average properties (shown in Table 5.1), we reduce the effect of the uncertainty associated with the measurements mentioned above. A more detailed analysis of how these uncertainties affect our estimate lithospheric mantle densities should be investigated further by quantifying the errors in the individual seismic methods used to acquire our LAB depth estimates.

5.6 Conclusions

We use the Crust1.0 model and LAB depth estimates from previous seismic studies to estimate the continental lithospheric mantle density in four cratonic regions in central North America, Scandinavia, southern India, and western Australia. Our results show that the cratonic lithospheric mantle contributes a negative component to the total observed elevation in these cratons, where we estimate lithospheric mantle densities are ~1-2% higher than asthenospheric density (between 3.32 and 3.35 g/cm³, assuming a reference asthenospheric density of 3.3 g/cm³). This analysis suggests that cratonic lithospheric mantle is not necessarily neutrally buoyant as described by the isopycnic hypothesis (Jordan, 1988). These estimated cratonic lithospheric mantle densities can be used to explore the thermal and compositional contributions to mantle lid density with the goal of improving our understanding of the processes involved in the stabilization of the cratons.
Conclusions

This thesis investigates the lithospheric modification processes associated with the various stages of continent-continent collision and orogeny, from the final stages of oceanic subduction to the continent-continent suturing process, to post-collisional stabilization. Chapters two and three focus on the westernmost portion of the Alpine orogeny in the western Mediterranean. These chapters provide insight into the tectonic and lithospheric modification processes associated with the final stages of oceanic subduction and the seismic images presented here can be considered snapshots of a “dying” subduction zone. The tectonic history of this region has been the subject of intense debate as various models have been proposed to explain the uplift of the Betic, Rif, and Atlas Mountains and the simultaneous extension in the Alboran Sea. This research provides insight into this tectonic history by imaging the Alboran Slab, which has rolled back to its current position and is currently detaching from the base of the Gibraltar Arc lithosphere, and by providing evidence for continental lithospheric removal associated with slab detachment. Receiver functions from this region reveal a deep Moho (~45-50 km) around the western Gibraltar Arc, which shallows abruptly (~25 km) in the eastern Betic and Rif Mountains. The joint inversion of receiver functions and Rayleigh wave dispersion data produces a shear wave velocity model in which the high velocity Alboran Slab is seen extending vertically from the base of the crust in the western Betic and Rif Mountains down to ~250 km. In the eastern Betic and Rif Mountains,
the top of the Alboran slab does not reach the base of the crust; instead, it appears detached. Shallow low velocities above the slab suggest that this detachment has resulted in lithospheric removal from the base of the Gibraltar Arc. Subsequent upwelling of hot, buoyant asthenosphere may support the high elevations observed in the Sierra Nevada Mountains. Results from the western Mediterranean tectonic region illustrate that continent-continent collisions can produce diffuse zones of deformation, in this case, extending from the Atlas Mountains in Morocco to the Betic Mountains in southern Spain.

Chapter four provides insight into the continent-continent suturing process. Receiver functions from the Trans-Hudson orogen show evidence for crustal scale thrusting associated with the collision between the Archean Wyoming and Superior provinces, which occurred in the Precambrian during the formation of Laurentia. Receiver functions and gravity data also provide evidence for a dense lower crustal layer throughout the Trans-Hudson orogen and surrounding region. This layer may have been emplaced by mafic underplating during orogeny and/or after orogeny during a large scale magmatic event occurring ~1.4 Ga. The fifth chapter explores the effect that a dense lower crustal layer would have on the isostatic stabilization of the Superior craton and other cratons around the world. This chapter shows that there is no correlation between crustal thickness and observed elevation in the cratons, suggesting that the lithospheric mantle must contribute a negatively buoyant component to the observed elevations in these regions. Seismically derived LAB depths are used to estimate the density of cratonic lithospheric mantle. These
calculations show that the lithospheric mantle is not necessarily neutrally buoyant, but rather \( \sim 1-2\% \) denser than the underlying asthenosphere.

Figure C.1a shows a cartoon illustration of what may be considered an “idealized” picture of oceanic subduction and continent-continent collision. With the findings presented in this research this cartoon can be modified to illustrate a more detailed picture of the various tectonic processes associated with the closure of an ocean basin and the formation of a continent-continent suture (Figure C.1b). This figure provides a brief overview of the continental lithospheric modification processes discussed in this thesis.
Figure C.1: a) This is an idealized cartoon of oceanic subduction preceding continent-continent collision (right), analogous to the Western Mediterranean study region, and a continent-continent suture zone (left), analogous to the Trans-Hudson orogen study region. b) This is an updated cartoon representing the two study regions discussed in this dissertation. In the Trans-Hudson orogen (left), results suggest mafic underplating (both structural and magmatic), the preservation of a Precambrian subduction zone (relic subduction), and crustal scale thrusting associated with the formation of the WY-Superior suture zone. In the Western Mediterranean (right), subduction and slab rollback have led to the closure of the ocean basin between Africa and Iberia as well as delamination of continental lithosphere associated with detachment of the subducting oceanic plate from the continental margins.
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