Seismic Discontinuities beneath the southwestern United States from S receiver functions

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1. INTRODUCTION

The region of the Colorado Plateau-Rio Grande Rift-Great Plains (Fig. 1) was assembled in a series of continental building events where the Proterozoic terranes with assemblages of Island arcs, oceanic plateau and marginal basin units were accreted to the continent (Bowring and Karlstrom, 1990; Condie, 1982; Condie and Silverstone, 1999; Frey et al., 2001). The Great Plains (GP) has low relief and has not undergone significant deformation since Precambrian times, and it represents the western edge of the North America craton (Gao et al., 2004). The Colorado Plateau (CP) that has also been tectonically stable since the Precambrian (Morgan et al., 1986) lies to the west at an average elevation of about 1.8 km. The high-standing Plateau is a relatively undeformed crustal block while surrounded by the deformed Rocky Mountains and the Basin and Range Province (BRP) (Foos, 1999). The region between the Colorado Plateau and the Great Plains (Fig. 1) is the Rio Grande Rift (RGR) that consists of a series of north-south trending faulted basins. The area has been a product of series of compressional and extensional forces that have swept this area from Mesozoic time through the Cenozoic and the last 10 Ma (Sheehan et al., 2004).

The southwestern United States has undergone varied tectonic extension along with much of the interior western North America in the past 30 Ma (Baldridge et al., 1991; Olsen et al., 1987; Wilson et al., 2005). The extension has resulted in the creation of the Rio-Grande Rift but has left the Colorado Plateau largely undeformed. The RGR has been formed in at least two stages, the initial stage at about 20-30 Ma and the later stage at about 10 Ma. Increased volcanism in the past 5 Ma has occurred along the Jemez Lineament including the Mount Taylor region and Jemez Mountains. The Jemez lineament lies in the southeastern Colorado Plateau and has been referred to as a transition zone between the CP and RGR.

Although the Colorado Plateau has been apparently undeformed during the Cenozoic extension in most of the southwestern United States, it has experienced continuous slow uplift between 25 Ma and 5 Ma up to ~800 m, while the main rapid uplift occurred in the last 5 Ma with up to ~1100 m displacement (Sahagian et al., 2002). The margins are marked by major volcanic accumulations that resulted in the bowl-shaped topography of the Plateau (Foos, 1999). Geochemical and geophysical studies indicate that the lithosphere of the plateau has been thermally and chemically eroded at the edge (Crow et al., 2011; Levander et al., 2011; Liu et al., 2012; Schmandt and Humphreys, 2010). Several hypotheses have been proposed to explain the general uplift of the plateau, which include large-scale mantle upwelling (Moucha et al., 2009), compositional buoyance in the lithosphere (Karlstrom et al., 2012; Roy et al., 2009), the removal of the Farallon slab and subsequent upwellings (Liu and Gurnis, 2010); and small-scale edge-driven convection (van Wijk et al., 2010).

Previous seismic experiments and the recent USArray in the southwestern US have resulted in numerous constraints on the crust and mantle structure beneath the Colorado Plateau, Rio Grande Rift, and the Great Plains (Bailey et al., 2012; Bashir et al., 2011; Gao et al., 2004; Gilbert et al., 2007; Kumar et al., 2012; Levander et al., 2011; Levander and Miller, 2012; Li et al., 2007; Liu et al., 2012; Schmandt and Humphreys, 2010; Shen et al., 2013; West et al., 2004a; West et al., 2004b; Wilson et al., 2005). The RGR is characterized by a thin crust and lithosphere with significantly low velocities in the upper mantle to 200-300 km depth. The Colorado Plateau and the Great Plains have a thick crust (~45 km) and relatively fast mantle above 200 km.

Despite many studies were conducted from the RISTRA experiment, the lithosphere thickness has not been fully investigated in previous studies. In this study, we processed S

receiver functions from the RISTRA array and mapped discontinuities in the underneath crust and lithosphere. The high density of the stations allows us to observe sharp depressions of the LAB at the RGR-CP and RGR-GP boundaries. The high-resolution image of crustal and mantle discontinuities sheds light on the dynamic process in the upper mantle beneath the RISTRA array and provides new evidence for small-scale mantle convection that is largely responsible for the recent rifting and uplifting in the area.

2. METHODOLOGY AND DATA ANALYSIS

Receiver function methods utilize converted phases of P-to-S or S-to-P from a discontinuity and possess a great capability to resolve seismic boundaries in the crust and mantle. The S receiver function method is powerful in identifying interfaces in the lithosphere and upper mantle because it is free of contamination by multiple reverberations (Li et al., 2007). An S-receiver function is obtained by deconvolving the Sv component from P-component using the frequency domain deconvolution approach (Ammon, 1991; Langston, 1979). The method utilizes the equalization of the source and the path effects such that the S-receiver function images the underlying seismic structure beneath individual stations. In the time domain, the ground motion in the S wave window can be expressed as

$$P(t) = I(t) * S(t) * ESv(t) \dots\dots\dots(i)$$

$$Sv(t) = I(t) * S(t) \dots\dots\dots(ii)$$

Where P(t) and Sv(t) represent the time series of the P and Sv component, respectively; I(t) is the instrument response; S(t) is the earthquake source effect; and ESv(t) is the S receiver function, consisting of S-to-P conversions from underlying discontinuities.

In the frequency domain, the S receiver function is

$$ESv(\omega) = \frac{P(\omega)Sv^*(\omega)}{F(\omega)} G(\omega) \dots\dots\dots(iii)$$

$$G(\omega) = e^{-\frac{\omega^2}{4a^2}} \dots\dots\dots (iv)$$

$$F(\omega) = \max \{Sv(\omega)Sv^*(\omega), cSv(\omega)Sv^*(\omega)\} \dots\dots\dots(v)$$

Where $Sv^*(\omega)$ represents the complex conjugate of $Sv(\omega)$; The $G(\omega)$ in equation (iii) and (iv) represents a Gaussian function that acts as a low pass filter based on the value of “a”. A value of 0.4 is used in the study so that the Gaussian function matches the frequency contents of the receiver functions. In addition, “c” in equation (v) represents the water level factor that is used to avoid instability caused by the division of small values. This factor is controlled by signal to noise ratio of the original seismograms and a value of 0.02 was chosen for this research. After calculating $ESv(\omega)$ from equation (iii), we can obtain the S receiver function, $ESv(t)$, by applying the inverse Fast Fourier transform.

The dataset used in this study is from the Colorado Plateau-Rio Grande Rift-Great Plains Seismic Transect (LA RISTRA), which consisted of 57 broadband seismometers that were deployed in July 1999 and removed from the field in May 2001 (Fig. 1). The RISTRA seismic array was oriented approximately in NW-SE direction, parallel to the regional Proterozoic accretionary gradient (Bowring and Karlstrom, 1990; Wilson et al., 2005). It extended approximately along a great circle between Lake Powell, Utah and Pecos, Texas with an inter-station spacing about 18 km (Wilson et al., 2005). The experiment was designed to take advantage of the geometry of the geologic basement and back azimuths to prolific teleseismic source zones in the western South America and the northern Pacific.

For S receiver function study, the useful epicentral distance range is 55° - 85° for S wave and greater than 85° for SKS wave (Yuan et al., 2006). The distance range for S-P conversions depends mainly on the depth of the discontinuity and the conversions do not occur over critical angles. In this study, we requested all teleseismic events with $M_b > 5.8$ in a distance range of 60° - 110° from the IRIS data management center (Fig. 2) for both S and SKS receiver function analysis. The number of events at each station varies from 30 to 55 with an average of 45 events per station.

We filtered three-component seismograms for selected events with a frequency band of 0.05 – 0.12 Hz to remove high frequency noises and kept the waveforms with high signal-to-noise ratios for S and SKS phases. The selected seismograms were rotated from Z, N, and E components to P, SV, and SH components to separate the S wave and its converted P wave signals. Then individual receiver functions are obtained using the method above. Time is reversed on S receiver functions so that Sp converted phases are at a positive time because they actually arrived earlier than the direct S or SKS waves. The amplitude of S receiver function is also flipped relative to zero in order to be consistent with P receiver functions on which a positive converted phase corresponds to an interface with a positive velocity gradient. One criterion we applied to select useful receiver function is a clear S to P converted phase from the Moho. Individual receiver functions are inspected carefully and obvious bad traces are removed.

Individual receiver functions are stacked by ray piercing point to improve the signal-to-noise ratio of the S to P converted phases. The piercing points of S-to-P converted phases for both S and SKS phase are calculated from the IASP91 model (Kennett and Engdahl, 1991) at a conversion depth of 90 km (Fig. 3). The conversion points of Sp phases could be 100 km away from the associated station, which makes it necessary to stack receiver functions by binning the

138 piercing points instead of by station. 54 bins are produced with the bin dimension represented by
139 consecutive parallel lines covering a distance of ~44 km (Fig. 3); and the center of each bin
140 coincides with the station location. The width of each bin is designed to contain at least 10
141 piercing points. We used a bin width of 44 km, which provides a coherent image of the stacked
142 receiver functions, after trying several values (11, 33, 44, and 55 km). The number of piercing
143 points within each bin varies from 15 to 25. Given that the station spacing is about 18 km, the
144 neighboring two bins overlap about 26 km along the station profile.

145 Both S and SKS receiver functions are stacked together provided their piercing points are
146 located in the same bin. Moveout for distance is corrected with a reference ray parameter 6.4
147 s/deg using the IASP91 model (Kennett and Engdahl, 1991) as performed by Li et al. (2007).
148 The moveout value represents the time difference between the travel time difference of the S
149 phase and Sp phase for a given ray with respect to the travel time difference of the reference ray.
150 The moveout correction aligns all responses from different ray paths from the same discontinuity
151 at the same time. Therefore, stacking receiver functions with the moveout corrections improves
152 the amplitude of the converted phase while suppresses random noises that are canceled out by
153 destructive interferences. The stacked receiver functions are migrated from time to depth domain
154 based on the IASP91 model (Kennett and Engdahl, 1991). The migration produces a depth image
155 of the earth's internal discontinuities and helps to better view the subsurface structure.

157 3. RESULTS OF RECEIVER FUNCTION IMAGING

158 We generated S receiver functions from S and SKS waveform data for 54 stations of the
159 RISTRA array (Fig. 1). The average number of individual receiver functions varies from 40 to

60 at one station, but many were discarded due to large noise or lack of clear S to P conversion from the Moho. Fig. 4 displays S receiver functions in four bins corresponding to four stations that are located in the GP, RGR, CP-RGR boundary and CP, respectively. A summation trace is produced for each bin by stacking individual S and SKS receiver functions whose piercing points are within the same bin. It is noticeable that the energy level after 20 s is significantly reduced on the summed traces, suggesting that much of that energy on individual S receiver functions is caused by random noise. Therefore, we only focus on coherent signals across the profile, which are presumably from true structures. The stack traces at all stations (Fig. 5) are migrated to depth using the IASP91 model (Kennett and Engdahl, 1991) and plotted with color-coded amplitude across the RISTRA profile (Fig. 6a).

3.1 The Moho discontinuity

The most obvious and coherent converted phases are around 4-6 s beneath all the stations, which are the S to P conversions from the Moho. The Moho depth varies from 30 to 40 km beneath the RGR and from 35 to 45 km beneath the Great Plains and the Colorado Plateau. The Moho beneath the Great Plains is imaged by a sharp Sp conversion from stations TX01-NM14 with a depth of about 40 km. A slightly deeper Moho at ~45 km is observed beneath stations NM16-NM18, corresponding to the Ruidoso volcanic field known as the Pajarito Mountain (Bowsher, 1991; Roberts et al., 1991). Similarly, the Moho depth at Mount Taylor, a major volcanic field of the Jemez lineament is about 45 km beneath stations NM32-NM34. The Moho depth under stations NM20-NM32 ranges from 30 – 40 km, roughly coinciding with the surface morphology of the rift. The Moho beneath the Colorado Plateau west of station NM34

ranges from 40-45 km. The variations of the Moho depth imaged from S receiver functions from this study are largely consistent with those constrained from P receiver functions (Wilson et al., 2005). In addition, other P and S receiver function studies also show that the Moho in the CP further away from the RGR boundary is at 40-45 km (Bashir et al., 2011; Gilbert et al., 2007; Hopper et al., 2014; Lekic and Fischer, 2014). The agreement in the Moho depth between S and P receiver functions gives us confidence about the S receiver function image for upper mantle discontinuities beneath the RISTRA array.

3.2 Discontinuities in the upper mantle

Several coherent Sp converted phases are imaged in the upper mantle beneath the RISTRA profile. The first one is a negative phase at 8-12 s (blue line in Fig. 5 and Fig. 6a) beneath all three provinces. This phase has a sharp peak and large amplitude beneath the RGR and the interior of the CP and GP, but is broad and weak at the boundaries of RGR-CP and RGR-GP. This phase shares a similar variation pattern along the profile as the Moho converted phase; indicate that it could be partially attributed to the side-lobe of the strong Moho conversion. However, the side lobe of the Moho conversion is usually much weaker (Lew, 2010) than some observed phases in Fig. 5, suggesting possible contributions from a seismic discontinuity with a negative velocity gradient. This is especially true beneath the RGR, where the phase at ~10 s has a large amplitude and no other coherent phase is imaged in the upper mantle. It is possible that the Sp phase from the lithosphere-asthenosphere boundary beneath the RGR coincides with the side lobe of the Moho conversion. If so, the LAB beneath the RGR is mapped at the depth of ~80 km. Beneath the interior of CP and GP away from the RGR, if this phase is not fully due to the Moho side lobe but also correlates with a seismic discontinuity, the corresponding depth of

the discontinuity is about 80-90 km and ~70 km, respectively. This discontinuity is too shallow to be the LAB for these two provinces. We therefore interpret it as the mid-lithosphere discontinuity that might be caused by anisotropy variation within the lithosphere as found in other stable continental region (Rychert and Shearer, 2009; Ford et al., 2010).

Another weak but consistent negative phase is observed at ~16 s beneath the CP-RGR boundary (station 34-41) and gradually deepens to ~22 s towards the interior of the CP (Fig. 5). This phase is also present at the RGR-GP boundary (station 08-14) at ~12 s. We interpret that this weak and later phase beneath the CP and GP as S to P conversions from the lithosphere-asthenosphere boundary. The depth of the LAB is ~100 km beneath the Great Plains and at 120-180 km beneath the Colorado Plateau. The LAB depth increases sharply from ~80 km beneath the RGR to 120 km beneath the CP and 100 km beneath the GP in less than 50 km distance on the surface, evidence for sudden variations of the lithosphere structure across the boundaries of tectonic provinces.

The increase of the lithosphere thickness from the RGR to the interior of the Colorado Plateau observed in this study is similar to previous results from receiver function images using the USArray data (Levander et al., 2011; Levander and Miller, 2012; Kumar et al., 2012; Kind et al., 2012; Lekic and Fischer, 2014), while the absolute lithosphere thickness beneath the CP is about 40-60 km thicker than the previous values, which could be caused by different frequencies used in the data. In addition, the sharp increase of the lithosphere thickness at the RGR-CP and the RGR-GP boundary is a unique feature from this dataset owing to the much higher station density.

We also observe a strong, coherent positive phase at ~32 s and a negative phase at ~37 s at stations of 41-49 in the Colorado Plateau, corresponding to a discontinuity at the depth of ~250 km and ~300 km, respectively. These discontinuities must be a local feature as no similar phases appear in other places of the profile. Seismic velocity increases across the shallower discontinuity but decreases across the deeper one, which can be explained as a segment of the eroded lithosphere whose velocity is higher than the surrounding asthenosphere.

4. DISCUSSION

4.1 Uncertainties of the depths of discontinuities

The depths of imaged discontinuities from S receiver functions depend on the accuracy in picking the peak time of an S-P phase, the velocity model used in depth migration, and the dominant frequency in the data. The picking error is related to the amplitude and sharpness of the converted phase that reflects the velocity gradient across the discontinuity. A larger velocity gradient across a boundary tends to produce a sharper conversion (Kind et al., 2012). The errors of our phase picks range from 0.25 s to 0.5 s, corresponding to 2-4 km in depth. The velocity model that we used in depth migration is a global reference model, IASP91, which is faster than the regional velocity model beneath the RGR and comparable with the model beneath the CP (Liu et al., 2011). An average 5% slower model would map the Moho and LAB about 2 km and 5 km shallower, respectively. If this correction is applied to the RGR area, the LAB variation across the RGR-CP boundary will be even more significant.

These uncertainties from the picking errors and velocity model together cannot explain the difference between the 160-180 km thick lithosphere imaged beneath the Colorado Plateau from this study and ~120 km found in previous studies (Levander and Miller, 2012; Lekic and Fischer, 2014). We suspect that the difference in dominant frequencies among the studies (0.2

Hz in (Levander and Miller, 2012); 0.03-0.25 Hz in (Lekic and Fischer, 2014); and 0.05-0.12 Hz in our study) is probably a key reason for the discrepancy of the LAB depth. Kind et al. (2012) have shown that S receiver functions of different frequency have varying resolution for a smoothing discontinuity with gradational velocity variation. S receiver functions at a period of ~13 s can still image a gradational velocity drop (~0.5 km/s over 100 km depth), which cannot be detected by S receiver functions at shorter periods (4 s and 7 s). The discrepancy between our results and previous ones indicates that the LAB beneath the CP has varying velocity gradients, a sharper gradient above ~140 km, which is observed from short period data (Levander and Miller, 2012; Lekic and Fischer, 2014), and a smoother gradient underneath, which is imaged from long period data (this study). Such velocity gradient variation with depth is also consistent with velocity models constrained from surface waves (West et al., 2004; Liu et al., 2011).

4.2 Lithosphere extension and mantle upwelling beneath the RGR

The estimated lithospheric depth beneath the Rio Grande Rift ranges from 70 to 80 km, much thinner than beneath the adjacent Colorado Plateau (120-180 km) and the Great Plains (~100 km). The observation of an elevated LAB is consistent with many previous studies in the RGR. Lithosphere thinning and asthenosphere upwarp beneath the RGR has been evident from P wave residual and gravity modeling (Parker et al., 1984), from chemical and isotopic constraints (Perry et al., 1988), and from body and surface wave tomography models using data from the RISTRA experiment and the USArray (Gao et al., 2004; West et al., 2004; Levander et al., 2011; Liu et al., 2012; Schmandt and Humphreys, 2010; Shen et al., 2013). Besides the shallow LAB, asthenosphere upwelling beneath the RGR is evident by the strong Sp phases,

which indicates a sharp velocity drop across the LAB and a presumably hot asthenosphere. Tomography models have indeed imaged significantly low velocity in the upper mantle of RGR (Gao et al., 2004; West et al., 2004; Liu et al., 2011).

We estimate a stretching factor of 1.25-1.43 under the RGR assuming a uniform extension of the lithosphere (McKenzie, 1978) and an initial lithosphere thickness of 100 km although the actual lithosphere stretching is probably depth dependent and more complicated than McKenzie's model. Hot asthenosphere upwelling, which is a passive consequence of lithosphere stretching in McKenzie's rifting model, is probably produced by the removal of flat Farallon slab and enhanced by edge-driven convection (Liu and Gurnis, 2010; van Wijk et al., 2010) and is likely the cause of lithosphere extension in the Rio Grande Rift.

4.3 Mantle dynamics under the RISTRA

The lithospheric thickness varies from 80 km beneath the RGR to 100 km beneath the GP and 120-180 km beneath the CP. The LAB variation imaged from S receiver functions is consistent with a negative velocity gradient beneath the RISTRA in the S velocity model by West et al. (2004) (Fig. 6b). An interesting observation is the sharp lithosphere step at the RGR-CP boundary, where the LAB deepens by ~40 km in less than 50 km horizontal distance. This lithosphere step, which is at a scale smaller than the station spacing in the USArray network, was not revealed in regional LAB maps from the USArray data (Kumar et al., 2012; Levander et al., 2011; Levander and Miller, 2012; Lekic and Fischer, 2014). The lithosphere step at the RGR-CP boundary corresponds to the rapid change of surface elevation and probably marks the boundary of the two tectonic provinces at depth.

293 A sharp lithosphere thickness change can produce edge-driven convection in the upper
294 mantle as simulated by van Wijk et al. (2010). Our observation suggested such a step of the
295 LAB indeed exists at the RGR-CP and RGR-GP boundaries, which would promote small-scale
296 mantle convection, upwelling beneath the RGR and downwelling at its boundary, as shown in
297 the schematic model in Figure 7. Our image of two local discontinuities at ~250 km and ~300
298 km depth beneath the southeastern CP boundary could be the top and base of the removed
299 lithosphere, a consequence of the small-scale mantle convection.

300 Geophysical, geochemical and geological evidence all support that the extension and
301 uplift in southwestern US are driven by mantle processes (Crow et al., 2011; Karlstrom et al.,
302 2008; Karlstrom et al., 2012; Levander et al., 2011; Schmandt and Humphreys, 2010). Multiple
303 mechanisms must have played important roles for regional tectonic activities at different times.
304 Strong asthenosphere upwelling was expected following the descent of Farallon slab and largely
305 responsible for the extension and widespread magmatism in the RGR at 30-40 Ma. The
306 extension in the RGR significantly thinned the lithosphere while the strong lithosphere beneath
307 the CP remains largely undeformed. The lithosphere step then formed at the RGR-CP boundary,
308 as imaged in this study, and invoked edge-driven convection, which causes upwelling beneath
309 the CP and RGR and downwelling at the RGR-CP boundary as simulated by numerical modeling
310 (van Wijk et al., 2010). The edge-driven, small-scale mantle convection is likely the current
311 mechanism that is responsible for recent uplift and magmatism in the Rio Grande Rift and the
312 Colorado Plateau.

CONCLUSIONS

S receiver functions along the RISTRA transect clearly image the Moho and LAB. A shallower Moho and LAB beneath the RGR than the adjacent CP and GP indicate lithosphere extension and asthenosphere upwarp beneath the rift. The lithosphere under the CP deepens from ~120 km at the RGR-CP boundary to ~180 km towards the northwestern interior. An anomalous lithosphere step of ~40 km is observed at RGR-CP boundary, providing evidence for edge-driven convection. Two local discontinuities at ~250 km and ~300 km beneath the southeastern CP probably image the top and bottom boundary of the eroded lithosphere. The LAB variation along the RISTRA provides further evidence for small-scale, edge-driven mantle convection, which is likely the dominant mechanism for recent magmatism and uplift in the Rio Grande Rift and Colorado Plateau region.

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Figure 1. Map of the RISTRA array and topography of the study area. Red triangles represent the locations of 54 stations; black triangles are stations shown in Figure 4.

Figure 2. The distribution of used earthquakes (black dots) with $M_b > 5.8$ recorded during the RISTRA deployment. The study area is represented by a black triangle as the center of the projection.

Figure 3. Configuration of bins of common conversion piercing points used in stacking receiver functions. The piecing points (blue dots) are calculated at a depth of 90 km. Red triangles are stations and also the centers of each bin. Lines that are 44 km apart and perpendicular to the array show two bins centered at station 25 and 36, respectively.

Figure 4. Individual receiver functions and their CCP stacks (the most right trace) at four bins centered at station NM14, NM28, NM35, and UT52, which are marked as black triangles in Fig 1. Negative amplitudes are filled with black.

Figure 5. Profiles for S receiver function stacks at bins centered at each station. (A) Stacks of receiver functions in time. (B) Stacks converted to depth using the IASP91 model. The topography is plotted at the top to show tectonic provinces. Coherent phases are highlighted by dashed lines. Red dashed lines represent a positive phase from the Moho; black dashed lines represent a negative phase from the LAB; light blue dashed lines are largely side-lobes of the Moho conversion beneath the CP and GP or phases from a mid-lithosphere discontinuity. The green dashed lines mark local discontinuities with a positive and negative velocity gradient, respectively, probably the top and bottom of the eroded lithosphere.

Figure 6: (A) Same as Fig. 5b with color-coded amplitude. (B) Shear wave velocity model beneath the RISTRA transect constrained from surface waves (West et al., 2004). Black dashed line is the LAB from (A).

Figure 7. Schematic illustration for present day mantle dynamics beneath the Colorado Plateau, Rio Grande Rift, and Great Plains based on S-receiver function images in this study.