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Seismic discontinuities beneath the southwestern United States from S receiver functions

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S-receiver functions along the Colorado Plateau-Rio Grande Rift-Great Plains Transect known as LA RISTRA in the southwestern United States have been utilized to map seismic discontinuities beneath this tectonically active region. Individual receiver functions were stacked according to ray piercing points with moveout corrections in order to improve the signal-to-noise ratio of the converted S-to-P phases. A mantle discontinuity, which is interpreted as the lithosphere-asthenosphere boundary (LAB), is observed along the profile with depth ranging from 80 km beneath the Rio Grande Rift (RGR) to 100 km beneath the Great Plains (GP) and 120–180 km beneath the Colorado Plateau (CP). The shallow LAB beneath the Rio Grande Rift is indicative of lithosphere extension and asthenosphere upwarp. The LAB deepens sharply at the RGR-CP and RGR-GP boundaries, providing evidence for edge-driven, small-scale mantle convection beneath LA RISTRA. Two local discontinuities beneath the southeastern Colorado Plateau are imaged at ~250 km and ~300 km and could be the top and base of the eroded lithosphere, respectively. The S receiver function images suggest that edge-driven, small-scale convection is probably the mantle source for recent extension and uplift in the Rio Grande Rift and the Colorado Plateau.

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 buoyancy 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

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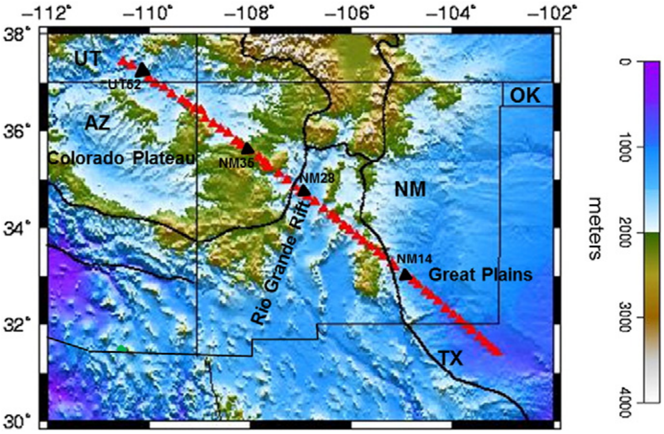

Fig. 1. Map of the RISTRA array and topography of the study area. Red triangles represent the locations of 54 stations; black triangles are stations shown in Fig. 4.

the Great Plains (Bailey et al., 2012; Bashir et al., 2011; Gao et al., 2004; Gilbert et al., 2007; Kumar et al., 2012a, b; Levander et al., 2011; Levander and Miller, 2012; Li et al., 2007; Liu et al., 2012; Schmandt and Humphreys, 2010; Shen et al., 2013; Sheehan et al., 1995; West et al., 2004a; West et al., 2004b; Wilson et al., 2003; 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

$$\mathrm{P(t) = I(T) * S(t) * ESv(t)} \tag{i}$$

$$\mathrm{Sv(t) = I(t) * S(t)} \tag{ii}$$

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

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

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

$$\mathrm{F}(\omega) = \max\{\mathrm{Sv}(\omega)\mathrm{Sv} * (\omega), \mathrm{cSv}(\omega)\mathrm{Sv} * (\omega)\} \tag{v}$$

where Sv*(ω) represents the complex conjugate of Sv(ω); The G(ω) in Eqs. (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 Eq. (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 (Ma, 2010) and a value of 0.02 was chosen for this research. After calculating ESv(ω) from Eq. (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 >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 Mb > 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

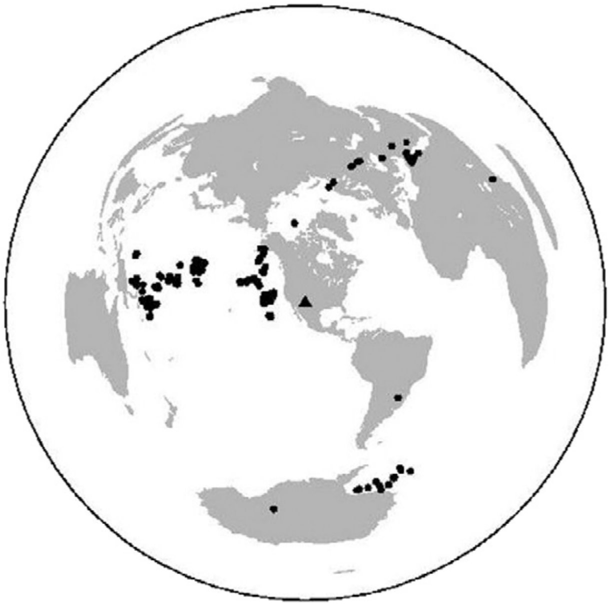

Fig. 2. The distribution of used earthquakes (black dots) with Mb > 5.8 recorded during the RISTRA deployment. The study area is represented by a black triangle as the center of the projection.