Upper Mantle Convection Beneath the Central Rio Grande Rift Imaged by P and S Wave Tomography

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Abstract

We present models for upper mantle P and S velocity structure beneath a southwestern United States transect extending from near the center of the Colorado Plateau across the Rio Grande rift to the Great Plains. The models were derived from travel times of compressional and shear seismic phases recorded by the La Ristra passive seismic array deployed from July 1999 to May 2001. Large variations in seismic velocity (up to 8% in S and 5% in P) are seen across the transect in the upper 200 km of the mantle. Seismically slow mantle underlies the Rio Grande rift and Jemez lineament and relatively slow mantle is seen beneath the Navajo volcanic field within the Colorado Plateau. The relative variations of P and S velocity imply that high temperatures are the primary cause of the slow mantle although a small amount of partial melting or hydration cannot be ruled out. Sharp boundaries in mantle seismic velocity are coincident with boundaries of Proterozoic structural trends implying that ancient lithospheric structure exerts a control on the tectonic and magmatic activity in the region. Weaker seismic variations are imaged below 200 km depth with three southeastward dipping structures in our images. Two of the structures have fast seismic anomalies, beneath the central Colorado Plateau and the Great Plains respectively, and the third has anomalously slow seismic wavespeed. We interpret the western deep seismic anomaly to be foundering Farallon slab and the slow anomaly just to the east as upwelling mantle possibly associated with volatile release from the sinking Farallon slab. Beneath the Great Plains, there is also downwelling in the upper mantle. The combination of upwelling in the west and downwelling in the east are likely part of an upper mantle convection cell that may include entrained lithosphere from beneath the rift.

Introduction

Since the late Cretaceous, the southwestern United States has passed through several profound tectonic events resulting in widespread magmatic activity, large scale uplift and crustal shortening, and more recent rifting and extension. The forces that caused this activity produced different surface manifestations in different regions. The result is that the present day southwestern United States can be divided into a number of distinct tectonic regions (Figure 1). The Great Plains have low relief and have not undergone significant deformation since Precambrian times. They form the western edge of the North American craton. The Colorado Plateau to the west has also been tectonically stable since Precambrian times (Morgan and Swanberg, 1985) but today lies at an average elevation of about 1.8 km. Between the Plateau and the Great Plains lies the Rio Grande rift, a north-south trending series of faulted basins extending from Colorado to Chihuahua, Mexico and the Big Bend region of Texas. The rift has exhibited extensive deformation and volcanism during the Cenozoic. The most recent voluminous volcanism, however, has occurred to the west of the rift along a northeast trend called the Jemez lineament (Baldridge et al., 1991). In south-central New Mexico the Jemez lineament lies in what has been termed a transition zone between the Colorado Plateau and the Rio Grande rift. Surrounding the Colorado Plateau to the south and west lies the Basin and Range province that has undergone widespread extension since mid-Tertiary times and several episodes of deformation since the Mesozoic.

The Colorado P**LA**teau/ **RI**o Grande Rift **S**eismic **TRA**nsect Experiment (La Ristra) is a passive seismic experiment that was run along a transect from the Great Plains, across the Rio

Grande rift and Jemez lineament, to the middle of the Colorado Plateau. The experiment consisted of 54 broad band seismic instruments deployed in a northwest oriented linear array shown in Figure1. The array was deployed from July 1999 to May 2001 with continuous recording at 20 samples per second using Reftek 24-bit acquisition systems supplied by PASSCAL. The goal of the project was to seismically image the crust and mantle beneath the three tectonic provinces in order to better understand the different tectonic behavior of the provinces as well as to understand the forces driving recent activity in the region. A more refined understanding of the present subsurface should also result in a better understanding of the diverse history of the region. In this paper, we present two dimensional models of P and S velocity variations within the mantle beneath La Ristra. The models extend from the Moho to 600 km depth with resolving power on the order of 50 km. The models were determined using back projection tomography of compressional and shear wave teleseismic delay times. In the remainder of the introduction we briefly review the geologic and tectonic history of the region covered by the La Ristra array and previous geophysical work done in this area.

Geological History

The basement rock of the southwestern United States consists of several Proterozoic terranes that were successively added to the North American craton during collisional tectonic events. The 1.8 Ga Yavapai province is exposed in Arizona and is juxtaposed to the Matzatzal province (1.6 to 1.7 Ga) to the southeast (Karlstrom and Bowring, 1988). The Grenville orogeny (~1.1 Ga) completed the assembly of the southwestern part of the craton during the collision of Laurentia with an uncertain land mass to the southeast (Mosher, 1998). The sutures connecting the different Proterozoic terranes tend to have a northeast-southwest trend through a region extending from Arizona to west Texas. Through much of the Paleozoic, the southwestern United States was stable at or below sea level. An exception to this was a major orogenic event that occurred about 300 Ma resulting in the Ancestral Rocky mountains that extended from west Texas through New Mexico to central Wyoming. By the beginning of the Cretaceous, however, the region extending from western Arizona to the central United States was below sea level. Ericksen and Slingerland (1990) show smoothly varying contours of paleobathymetry for the Cretaceous Interior Seaway indicating that lithospheric (crust and mantle) buoyancy was similar for a region extending from western Arizona to the present day Great Plains in the central United States at roughly 100 Ma. The subsidence causing the seaway is thought to be due to tilting of the North American continent caused by mantle flow resulting from subduction of the Farallon slab beneath the west coast of North America (Mitrovica et al., 1989).

From roughly 80 to 40 Ma, major uplift of the southwestern United States occurred during the Laramide orogeny. The uplifted region extended to central New Mexico and Colorado in a region not far from the Ancestral Rockies. Coincident with the Laramide uplifts far inland from the subduction zone to the west, was an eastward migration of volcanism from the Sierra Nevada volcanic arc (Coney and Reynolds, 1977) to New Mexico and Colorado. This has led to the hypothesis that a flattening of the subducting Farallon slab was the cause of the Laramide orogeny (Dickinson and Snyder, 1978) as well as the accompanying volcanism. Deformation during the Laramide was not uniform. The Colorado Plateau was little deformed during the Laramide whereas central Colorado and New Mexico underwent extensive high angle reverse faulting and crustal shortening. From 40 to 20 Ma, a westward migration of calc-alkaline ignimbrite eruptions occurred that is thought to be related to the foundering of the flat Farallon slab (Humphreys, 1995). During the same period extension began in New Mexico within the Rio Grande rift that runs north-south through the center of the state. Extension also began south and west of the Colorado Plateau throughout the Basin and Range province.

At present, the southwestern United States consists of several distinct tectonic blocks (Figure1). The 1.8 km high Colorado Plateau is an uplifted region that has experienced little deformation since Precambrian times (Morgan and Swanberg, 1985). The timing of the uplift is uncertain with estimates ranging from the Plateau attaining present elevations at the end of the Laramide (~40 Ma) to having 1 km of uplift within the last 5 Ma (Sahagian et al., 2002; Pederson et al., 2002; Chase et al. 2002). The Plateau is in isostatic equilibrium (Thompson and Zoback, 1979) so that the cause of the high topography is due to an increase in buoyancy of the crust or mantle beneath the Plateau since mid-Cretaceous time. Bird (1988) proposed that flat subduction of the Farallon plate during the Laramide orogeny displaced the lithosphere beneath the Plateau eastward. Subsequent foundering of the slab led to warm asthenosphere rising beneath the Plateau to cause the uplift. Isotopic evidence from xenoliths, however, shows that Proterozoic mantle rock still remains beneath the Plateau (Livaccari and Perry, 1993). This has led Spencer (1996) to propose that just the deeper lithosphere (below 80 km or so) was removed during the Laramide to account for the increase in elevation of the Plateau. Riter and Smith (1996), however, conclude from analyses of xenoliths erupted in the Navajo field that the mantle beneath the central Colorado Plateau was cold to at least 140 km depth 25 Ma. McQuarrie and Chase (2000) and Chase et al. (2002) propose that uplift of the Plateau was due to thickening of the crust during the Laramide. Selverstone et al. (1999) argue for a boundary in the central Colorado Plateau dividing different Proterozoic lower crustal xenolith populations arguing against significant lower crustal flow prior to 25 Ma or so. Humphreys et al. (2003) propose that hydration and heating of the lithosphere has been the main source of uplift throughout the western United States. Although it is clear the Plateau has been uplifted at some time during the past 40 Ma, the cause is still uncertain.

To the east of the Colorado Plateau lies the Rio Grande rift that separates the Plateau from the Great Plains. The region of the Rio Grande rift was faulted and uplifted during the Laramide orogeny. The rift also marks the eastern limit of Eocene arc volcanism (Lawton and Mcmillan, 1999). Extension within the rift has not been a uniform process (Morgan et al., 1986). A first period of extension occurred from about 30 to 20 Ma and a second period occurred beginning 10 Ma until about 3 Ma. Magmatism accompanied both periods of rapid extension while the time between rapid extension events was accompanied by a lull in magmatic activity (Morgan et al., 1986; Baldridge et al., 1991). The source of volcanics during the 30 to 20 Ma extension event appears to be mantle lithosphere whereas the more recent volcanism in the central and southern part of the rift has an asthenospheric isotopic signature (Perry et al., 1987; McMillan et al., 2000) indicating that the lithosphere has been thinned to something on the order of 50 km thickness over time. Volcanism today is far more active to the west of the rift along a northeast trend known as the Jemez lineament (Baldridge et al., 1991). The region of the Rio Grande rift is also topographically high today. Eaton (1987) noticed that the topography of the southern Rocky Mountains resembles that observed for oceanic ridges and proposed that the Rio Grande rift is the central graben of a north-south trending continental ridge. As with the Colorado Plateau, there is still uncertainty surrounding the timing of the uplift of the Rio Grande rift region. Gregory and Chase (1992) argue from paleobotanical evidence that the southern Rockies attained their topographic highs by the end of the Laramide (about 40 Ma) whereas Axelrod and Bailey (1976) and Heller et al. (2003) argue for significant uplift (over 1 km) during the past 10 Ma. As with the Colorado Plateau, there is reasonable agreement over the tectonic events that have shaped the Rio Grande rift region but the timing of the events and the underlying cause is still uncertain.

The Great Plains, to the east of the Rio Grande rift, has been tectonically stable since Precambrian times but has experienced uplift during and since the Cretaceous. The uplift has resulted in a tilt of the plains from a height of 1.5 km above sea level in eastern most New Mexico to elevations of about 200 m in eastern Texas and Oklahoma (Mitrovica et al., 1989). The cause of the tilt may be due to dynamical effects caused by the subducting Farallon plate (Mitrovica et al., 1989) or an increase in crustal thickness (Bird, 1984).

In summary, the southwestern United States has several well defined tectonic/physiographic provinces. The history of magmatism and tectonic activity is reasonably well documented although the timing of topographic changes is still uncertain. More uncertain are the underlying driving forces for the geologic activity in the region over time. Also not well understood is the reason for the different behavior of different tectonic provinces within the region.

Previous Geophysical Work

Many geophysical studies have been conducted in the region covered by the La Ristra array. Refraction and surface wave studies within the rift indicate a thinned crust and slow (presumably hot) shallow mantle (Keller et al., 1990; Sinno and Keller, 1986; Keller and Baldridge, 1999). Wilson et al. (2003) have analyzed receiver functions using data from the La Ristra array and confirm the thinning of the crust beneath the Rio Grande rift. They find crustal thicknesses of about 35 km beneath the rift as opposed to 42-45 km beneath the Colorado Plateau and 45-50 km beneath the Great Plains. Heat flow is high within the rift relative to the Great Plains to the east and the Colorado Plateau to the west (Lachenbruch and Sass, 1977). Pn velocities have been measured as low as 7.6 km/sec beneath the Rio Grande rift (Olsen et al., 1979) whereas Pn velocities beneath the Colorado Plateau are found to be about 8.1 km/sec (Beghoul et al., 1993). Lastowka et al. (2001) used surface wave dispersion measurements to find that the Colorado Plateau has a thin shear wave high velocity lid with low velocities beneath. Less work on the deep crust and mantle has been conducted to the east of the rift in the Great Plains. Large scale studies of upper mantle seismic velocity, however, show a strong contrast in upper mantle shear velocity to about 200 km depth between the Great Plains just to the east of the rift, and the Rio Grande rift - Colorado Plateau region (van der Lee and Nolet, 1997; Grand, 1994).

Several passive seismic experiments span the Rio Grande rift and sample the margins of the Colorado Plateau and Great Plains (Spence and Gross, 1990; Parker et al., 1984; Davis et al., 1993). Slack et al. (1996) combined all the available data from previous experiments to determine a three-dimensional model of P velocity beneath the rift and surrounding regions. Their results show slow P wavespeeds from 35 to 145 km depth along a NE-SW trend associated with the Jemez lineament. Relative to the Great Plains the velocities are 7-8% slower within this feature. They do not find very slow mantle velocities beneath the Rio Grande rift and conclude that any thermal signature associated with Miocene rifting of the rift has greatly diminished. Examination of the Slack et al. data, however, shows a clear distinction between rift and the Jemez lineament only at the southern edge of their model where resolution of structure may not be very good. Also, below 145 km depth, P velocity appears slower beneath the rift than beneath the Jemez. In contrast to the Slack et al. results, West et al. (2003), from the phase velocities of Rayleigh waves recorded by the La Ristra array, find shallow mantle shear velocity slower beneath the rift than beneath the Jemez lineament. The study of West et al. also finds a fast seismic lid beneath the Colorado Plateau to 120-150 km depth underlain by much slower shear velocities. This is similar to the results of Lastowka et al. (2001) except that their seismic lid only extended to 65-70 km depth.

Geophysical studies of the Colorado Plateau, Rio Grande rift, and Great Plains show that both the crust and mantle have been modified beneath the rift and Jemez lineament transition zone, with thinner crust and slower mantle beneath these regions relative to the east and west. It also appears that the mantle at some depth beneath the Colorado Plateau is seismically slower than the mantle beneath the Great Plains. The detailed geophysical results, however, show many differences. Part of the differences may be due to the fact that each individual study only samples a portion of a particular tectonic province and there may be variability in mantle properties within each province. Resolution issues may also be important. Issues that are still unresolved include at what depth and how has Colorado Plateau lithosphere been modified to give the added buoyancy needed to uplift it 2 km? What has controlled the location of magmatism and extension in the southwest? What role does deep mantle flow play in the rifting and magmatism observed today?

Inversion

The La Ristra experiment consisted of a deployment of 54 three component broadband seismometers in a linear array extending from the Great Plains of west Texas across the Rio

Grande rift and the Jemez Lineament and ending in the center of the Colorado Plateau (Figure 1). The seismic line extended about 950 km with an average station spacing of 18 km. The orientation of the line was chosen such that the line lies close to the great circle path connecting the array to seismically active regions in Alaska, the Kuriles, Japan and Central and South America. The array operated from July 1999 to May 2001 with continuous recording. The combination of the long length and the relatively tight station spacing, make the La Ristra array a unique data set for examining the seismic properties of the mantle beneath the southwestern United States.

Travel times of P waves with distances from 24^0 to 90^0, PKP, PKIKP, and PcP were measured by visually aligning the wavelets from an individual earthquake, stacking them, and then using cross correlation of the first down (or up) swing in the data with the event wavelet determined by stacking. The P-wave data were band passed filtered from 1.0 Hz to 0.25 Hz. Only data with backazimuths between -20^0 and 20^0 of the azimuth of the seismic line were used to minimize the influence of seismic heterogeneity perpendicular to the array on the tomographic inversion. The events used for the inversion are shown in Figure 2. Using this process 5007 compressional wave travel times were measured. Shear waves were measured using a similar approach but we filtered the data from 0.03 Hz to 0.1 Hz. For S waves and ScS waves, the data were rotated to radial and tangential components and the travel times were measured from the tangential component. The travel times of SKS waves were measured from the radial component. A total of 2164 S, ScS, and SKS travel time measurements were made. Travel time residuals were calculated for the data with respect to the IASP91 model of Kennett and Engdahl (1991) and means were then removed for each earthquake source for compressional and shear waves separately. An example of P and S waves produced by a single event is shown in Figure 3 with predicted travel times using the IASP model marked on each seismogram. The effect of topographic variations, crustal thickness variations, and sedimentary thickness variations were removed from the residuals using the results of Wilson et al. (2003) who analyzed receiver functions for each of the La Ristra stations. We used their model for seismic velocity and crustal layering to correct each residual to a 35 km thick crust with a simple velocity structure measured at sea level. Figure 4 shows the average residual at each station for all compressional and shear phases. We show the original data as well as the data corrected for near surface structure. The near surface corrections reach a maximum of 0.25 s for compressional phases and 0.75 s for shear phases respectively. The corrected residuals should reflect variations in mantle velocity beneath the array. In the inversion we heavily damped the solution in the crust.

The corrected travel time residuals of all the phases are shown in Figure 5. In this figure the data are divided into two groups depending on whether the earthquakes producing the seismic waves were to the northwest of the array or the southeast, respectively. Note the difference in residuals for both shear and compressional waves depending on the azimuth of the waves. The clear difference of the residuals with azimuth indicates that significant heterogeneity exists within the mantle at depth. Note also the similarity in pattern between the shear and compressional waves although the amplitude is different with shear waves varying a little more than 3 s in time across the array and P waves varying by about 1.5 s in time. Individual events show larger variations in residuals. Finally, note that the longwavelength pattern in residuals is primarily due to slow arrivals near the center of the array, likely due to slow mantle seismic velocities beneath the rift, but that there are also clear short wavelength anomalies in the data such as fast residuals from earthquakes to the southeast recorded by stations 13-17.

The travel time residuals were inverted for a two-dimensional model of P and S velocity beneath the array. The mantle beneath the array was divided into a two dimensional grid of equal size cells 25 km in dimension. For each residual, a ray was traced through the grid using the IASP91 velocity model (Kennett and Engdahl, 1991). The travel time residuals are then related to velocity variations in the mantle by:

$$\delta t_r = \Sigma l_{rb} \delta s_b + \delta t_e \qquad (1)$$

where δt_r is the time residual of the r^{th} ray, δs_b is the difference in slowness of the b^{th} block with respect to the starting model, l_{rb} is ray path length of ray r through block b, and δt_e is a static time shift for earthquake e. The linear equations given by equation (1) were solved iteratively using the LSQR algorithm of Paige and Saunders (1982). Nolet (1985) discusses this technique in more detail in relation to seismic travel time tomography. Aside from the travel time equations, we also added the following equations to the inversion.

$$S_b - 0.5S_{b-1} - 0.5S_{b+1} = 0 \qquad (2)$$

S_b is the slowness in block b and S_{b-1} and S_{b+1} are the slownesses in horizontal adjacent blocks to block b. These equations impose a smoothness constraint on the model. The degree of smoothing can be controlled by the weight given to equations (2). The stronger the smoothing constraint is the lower the variance reduction of the travel times in the inversion. Various smoothing weights were used and we present a model that shows short wavelength structure with minimal incoherence in the model.

To estimate the resolution in the inversion we performed synthetic tests where we create a fictitious earth model, compute the residuals for the actual raysets that this model would produce, and then invert the synthetic data in the same manner as was done in the actual inversion. Figure 6 shows the results for one such test. The starting model is shown in the top frame. This model was used for both P and S velocity inversions. The alternating blocks in the test structure are 50 km in dimension. The lower two panels show the results of the P and S inversions respectively. The smoothing constraint degrades the inversion but blocks within the center of the array are still relatively well resolved to depths of around 600 km. Note that at the edges of the array, the input anomalies are not well resolved and show evidence of streaking due to the lack of crossing rays.

The largest sources of error in our models are likely due to our neglect of anisotropy and 3D variations in seismic velocity. Gok et al. (2003) examined shear wave splitting across the La Ristra array. They found the fast direction for SKS waves to be relatively uniform across the array with a northeast orientation. The magnitude of the splits has some shortwavelength variation but based on SKS splitting measurements alone, we felt there is not enough constraint on anisotropy to include corrections in our inversion. Correcting our SKS travel time data for the measured splitting has negligible effect on our results. It is likely that including the effects of anisotropy on all the data would change the images in detail but given the relatively uniform fast direction of SKS splitting across the array, it is unlikely the major patterns of heterogeneity would change. Neglecting variations in velocity perpendicular to the seismic line is also a source of error that is difficult to quantify. Restricting the data to backazimuths within 20^0 of the azimuth of the seismic line should minimize these errors.

Model

The final compressional and shear velocity models of the mantle beneath the La Ristra array are shown in Figure 7. The upper 50 km of the model shows little lateral variation because the laterally variable crustal structure of Wilson et al. (2003) was used in the reference model and these depths were heavily damped in the tomographic inversion. The inverted model reduces the variance in the P wave data set by 83% and the variance in the S wave data by 79%. The largest amplitude variations in both P and S velocity are within the upper 200 km. Within the upper 200 km, the lateral variations in P and S wavespeeds are similar in pattern although in detail there are differences in relative amplitude. Both models show very slow wavespeeds within the center of the array extending from station NM20 to station NM35. Station NM35 was situated on Mount Taylor, a volcano that is part of the Jemez lineament and can be considered the western most point in the transition zone along our line between the Rio Grande rift and the Colorado Plateau. Station NM20 was situated near the town of Carrizozo, NM. Carrizozo is situated at the eastern limit of the Rio Grande rift along the La Ristra line and has also been the site of recent volcanism (Laughlin et al., 1994). Our results differ from the model of Slack et al. (1996) in that we find very slow seismic velocity in the mantle beneath the Rio Grande rift as well as the Jemez lineament although the slow anomaly extends deeper beneath the Mount Taylor region than beneath the center of the rift. Our results do, however, agree qualitatively with the surface wave study of West et al. (2003). Although mantle velocities are higher beneath the Great Plains (stations TX01-NM19) and the Colorado Plateau (stations NM36-UT54) relative to the rift, there are significant variations of mantle velocity within those two provinces. In particular, the eastern edge of the Plateau has high P and S velocity, but near the center of the

Plateau (near station AZ50), seismic velocity is significantly lower. The fast eastern part of the Colorado Plateau is beneath the San Juan basin and the slower central part is beneath the Navajo volcanic field that saw a number of volcanic eruptions 20-30 Ma (Riter and Smith, 1996). The eastern part of the model also shows lateral variation in velocity. Although the mantle above 120 km depth is fast beneath stations TX01-NM16 that cover the western Great Plains, it is particularly fast at depths greater than 120 km beneath stations TX05-NM10. This narrow feature extends deeper in the shear wave image than the compressional wave image. The variability within the Colorado Plateau and Great Plains makes it difficult to generalize differences in lithosphere thickness between the two provinces.

The mantle below 200 km depth shows less lateral variability in structure and more differences between the P and S velocity models. The most prominent structure in the deep mantle is a slow anomaly beneath stations NM40 to AZ45 from 250 to 500 km depth labeled B in Figure 7. The anomaly is clearly seen in the shear model but appears weaker in the P model. We tested whether the deep slow anomaly could be due to streaking of shallow structure to depth by performing "squeezing" experiments where by inversions are performed forcing all lateral variation in seismic wavespeeds above a certain depth and then gradually releasing constraints on structure to greater depths (Saltzer and Humphreys, 1997). We found that the slow deep feature remained even after every effort was made to fit the travel time data with structure only above 200 km depth. Near the edges of our model, two deep fast anomalies are seen. The western anomaly, near 500 km depth beneath AZ50, is at the edge of the model in a region where there are no crossing rays and therefore is probably not well resolved. On the other hand, the anomaly is dipping to the southeast which is the opposite dip direction than one would expect for streaking of rays (Figure 6). The eastern anomaly is seen in the shear wave model beneath station

TX05 extending from 200 to 600 km depth. The P model is also anomalous in the same region but is far stronger above 250 km depth. If the eastern fast anomaly is a downwelling, the difference in P and S may be due to anisotropy variations unaccounted for in the inversion. What is clear, however, is that there is anomalously fast mantle at depth over a relatively narrow horizontal distance beneath stations TX05-NM10.

Determining the cause of seismic heterogeneity is always problematic. Anisotropy can play a role in P and S wavespeed variations in tomographic models as well as temperature variations, melting, and compositional variations. Shear wave splitting measurements have been made using the Ristra seismic stations (Gok et al., 2003). The splitting is generally uniform across the array with a northeast oriented fast direction and thus it is unlikely that our neglect of anisotropy will produce large scale artificial laterally varying structures beneath the array. Relative variations of P and S wavespeeds can also be diagnostic of the physical conditions that result in seismic heterogeneity. The P and S data sets have different sampling of the mantle and we feel direct comparison of P and S velocity from the models is likely to be uncertain. On the other hand, a direct comparison of P travel time residuals versus S travel time residuals involves less assumption in analyzing P versus S velocity variations in models. Figure 8 shows a plot of S versus P travel time residuals for data where both P and S waves were measured from common earthquakes at common stations. There is a clear correlation between the two residual data sets. The best fitting least squares line through the data has a slope of 2.9. From the results of Karato (1993), one would expect the slope to be 2.9 for purely thermal effects assuming a Q_s of 89 and a Q_p of 200. The variance in travel time residuals is mostly due to the large heterogeneity within the upper 200 km of the models. Thus, we conclude that P and S heterogeneity above 200 km depth is primarily due to lateral changes in shallow mantle temperature. This is also consistent with the lateral changes in heat flow in the region (Morgan et al., 1986). Again using the results of Karato (1993), the maximum 8% variation in shear wave speed in our model at 100 km depth implies a temperature variation of 470 K in temperature across the array. This is a large temperature variation but not unreasonable. Nataf and Ricard (1996) estimate a temperature difference at 100 km depth of a little over 500 K between asthenosphere and typical platform lithosphere. The Karato model assumes relatively strong anelastic effects on the temperature derivatives of seismic velocity and our calculations assume relatively low, but reasonable, Q values so that our estimate of the expected slope of the S versus P residual curve due to solid state thermal effects is likely on the high end. It is possible then, that a small amount of partial melt in the shallow mantle beneath the center of our line would be consistent with our seismic observations as this would increase the Poisson ratio further than a solid state temperature increase (Williams and Garnero, 1996).

Chemical and mineralogic variation can also affect seismic velocity. The intrinsic variation in seismic velocity between fertile and infertile peridotite is relatively small (Humphreys and Dueker, 1994) so that it is unlikely that such variations are directly the cause of the seismic heterogeneity we observe. Water content can also affect seismic velocity with hydrated minerals having slower velocity than a typical dry peridotitic mantle (Tyburczy et al., 1991; Christensen, 1996). We find a narrow zone of low P and S velocity beneath station AZ50 near the Four Corners region. This area is associated with the Navajo volcanic field. Smith et al. (2002) have proposed, based on examination of mantle xenoliths, that a wedge of mantle beneath the Navajo volcanic field was hydrated from 70 to 35 Ma. They also propose that the hydration occurred preferentially within an old zone of weakness separating two Proterozoic terrains (Figure 1). The Four Corners region is near the southwestern end of the Colorado mineral belt.

To the northeast a similar shortwavelength slow seismic anomaly has been observed in the mantle beneath Aspen, Colorado in the Colorado mineral belt (Dueker et al., 2001). Given the thick lithosphere beneath the Navajo field (West et al., 2003; Riter and Smith, 1996) and the relatively narrow width of the slow velocity anomaly seen beneath station AZ50, it is possible the Four Corners seismic anomaly we see is due to the effect of hydration or chemical variation within the lithosphere of the Colorado Plateau. Serpentine has been found to have a very high Poisson ratio (Abers, 2000). Since the Four Corners anomaly is seen in P velocity as well as S velocity, it is unlikely that hydration alone is responsible for the anomaly and perhaps heat has been advected along with water and small amounts of melt.

Discussion

It is generally agreed that a flattening of the subducting Farallon plate and its subsequent removal from beneath the western United States has had a fundamental influence on the tectonic behavior of the region (see Humphreys et al., 2003 for a review). During flat slab subduction, the western U.S. lithosphere was under compression resulting in the Laramide orogeny and eastward migration of arc volcanism. Beginning about 40 Ma, it has been postulated that the flat slab beneath the western United States began to founder and roll back in some way towards the west (Coney and Reynolds, 1977; Humphreys, 1995). The foundering of the slab implies a replacement of relatively cold slab with hot asthenospheric mantle in the shallow upper mantle, resulting in widespread ignimbrite volcanism and ultimately in large scale extension through a large part of the western United States. Within this general framework, however, the details of how the flat slab interacted with the western North America lithosphere and how the roll back of the slab affected the lithosphere are still controversial. One general issue of importance is the role of pre-existing lithosphere structure in controlling the tectonics of the region and how that lithosphere was modified during the flat slab subduction and subsequent removal. A second issue is the role of active mantle processes, especially in recent times, in the tectonic behavior of the southwestern United States. Is the extension and volcanism of the past 20 Ma or so due to passive collapse of thickened crust or are externally applied forces responsible or are there forces acting on the lithosphere due to dynamic processes in the deeper mantle beneath the lithosphere?

The images of the mantle beneath the Ristra line (Figure 7) show lateral variations in P and S wavespeed at all depths. Above 200 km depth there is a clear relation between geologic province and mantle velocity. It is clear that regions that have exhibited magmatic and/or extensional activity during the past 40 Ma, the Rio Grande rift, the transition zone and the Navajo volcanic field, have slow mantle velocities beneath them bounded by mantle with relatively fast seismic velocity. The gradient in mantle seismic velocity also appears to be sharp between the regions. The slow anomalies beneath stations AZ50 (the Navajo field) and stations NM30-NM35 (the Jemez lineament) are both close to Proterozoic boundaries (Figure 1) (Karlstrom and Humphreys, 1998). This may imply a Proterozoic structural control on regions with mantle heating, melting, and deformation as proposed by Karlstrom and Humphreys (1998). The structural control may be related to variations in how fertile the mantle was and thus, how easily melt was generated, or to variations in fracture density that may have allowed easier melt and fluid migration through the lithosphere or perhaps to variations in the original thickness of the lithosphere (Humphreys et al., 2003; Smith et al., 2002). Removal of the flat Farallon slab and its replacement with hot asthenosphere about 30 Ma would then have modified the different lithospheres in different ways i.e. more fertile, fractured and thinner lithosphere would have generated more melt and been more mechanically weakened than surrounding lithosphere. At present, however, the seismic evidence discussed above indicates that most of the difference in mantle seismic velocity is due to temperature variations within the solid state. Also, there has been an increase in tectonic and magmatic activity during the past 10 Ma that is unlikely directly related to foundering of the Farallon plate 30 Ma. As discussed below, it may be that convection in the deeper mantle is responsible for recent magmatic and tectonic activity and that perhaps some of the older lithosphere beneath the Rio Grande rift and transition zone has been removed.

Below 200 km depth, the seismic variations are almost certainly related in some way to flow in the mantle, as this is below most estimates of lithosphere thickness. Buck (1986) and King and Ritsema (2000) showed that thinning of lithosphere due to extension should result in small scale convection due to the large lateral temperature differences created. Our seismically inferred lateral temperature variations are as large as Buck used in his calculations. Gao et al. (2003) claim to see evidence for such convection beneath the Baikal rift and the Rio Grande rift. In these studies upwelling occurs directly beneath the rift and downwellings occur symmetrically on either side of the rift. Our images show no indication of low seismic anomalies in the mantle beneath the rift below 150 km depth associated with upwelling mantle. Gao et al. (2003) claim that upwelling beneath the Baikal rift may align the a-axis of olivine in the vertical direction resulting in relatively fast P waves at near vertical incidence angle even if temperatures are hotter than average. This mechanism could explain the absence of deep slow P wave anomalies in our images but does not explain the absence of slow S wave anomalies. Thus we find no evidence for deep upwelling directly beneath the rift. However, we do see three large scale structures in the mantle below 200 km depth, labeled A, B, and C in Figure 7. Beneath the western edge of the model an eastward dipping high velocity anomaly (A) is seen in the transition zone. Resolution is very poor here as the anomaly is off the edge of the line, thus the detailed shape of the anomaly is unknown although it is certain there is anomalously fast mantle in the region. Van der Lee and Nolet (1997) also imaged high velocity in the same location in their study of shear velocity structure beneath North America. They interpret the transition zone anomaly near the Four Corners region to be the trailing edge of the Farallon plate. From the spatial progression of volcanism during the past 40 Ma, it appears that the region near the western end of the Ristra line is one of the last regions to have had the Farallon plate founder (Humphreys, 1995). Thus, we interpret the fast anomaly beneath the western part of the Ristra line to be remnant Farallon slab that may be stagnant in the transition zone or actively sinking. To the east of the slab anomaly is a region of slow seismic velocity (B) extending from near 500 km depth to the shallowest mantle (beneath seismic station 50). Resolution analysis and squeezing experiments indicate this feature is real. There appears to be some connection of the deeper slow anomaly with the strongest slow anomaly in the upper 200 km beneath Mount Taylor (station NM35) as well as with the Colorado mineral belt anomaly under station AZ50, both regions that have experienced recent volcanic activity. Thus it is likely that the deep slow anomaly represents some form of upwelling. The receiver function analysis of Wilson et al. (2003), however, shows no major deflection of the 410 km discontinuity in this region indicating it is unlikely that a large upwelling of hot mantle exists there. We suggest that the slow anomaly from 300 to 500 km depth beneath stations AZ45-AZ50 may be mantle that has been altered due to the release of water and other volatiles from the foundering Farallon slab just to the west. Recently, Bercovici and Karato (2003) have suggested that mantle that crosses the 410 km discontinuity from below would dehydrate due to the difference in water solubility between olivine and wadsleyite. They further suggest that partial melting could result just above 410 km depth due to this de-watering. The strongest part of anomaly B is indeed just above 410 km depth in our models and is stronger in S than P, perhaps indicating a small degree of partial melting. Previous studies have found indications of similar phenomena. Revenaugh and Sipkin (1994) interpreted a velocity reversal near 330 km depth beneath eastern China as a melt layer associated with subduction to the east. Zhao et al. (1997) found a slow anomaly to 400 km depth associated with the subducting Pacific plate in the Tonga subduction zone and associate the anomaly with dehydration of the slab and arc as well as back arc volcanism seen at the surface. Nolet and Zielhuis (1994) interpret a slow shear wave velocity anomaly at 300 to 500 km depth beneath the Tornquist zone as due to injection of water into the transition zone during ancient subduction. More recently, Abers (2000) has shown that hydrated oceanic crust exists to at least 250 km depth in subduction zones beneath Japan and Alaska, implying hydrous phases are carried beyond the volcanic front. The farallon slab beneath the Four Corners region is likely very young and it is not clear if hydrous phases in young slab can be carried to great depth but the dehydration behavior of slabs is still uncertain. If the slow deep anomaly beneath the Four Corners region is thermal in origin it may be that the Farallon slab upon reaching depths near 600 km has disturbed a thermal boundary layer creating an instability that results in rising hot mantle. Tan et al. (2002) show this phenomena in their simulations although the thermal boundary layer in their work is the core-mantle boundary.

At the eastern edge of the La Ristra line a fast seismic anomaly (C) is seen below 200 km depth beneath stations TX01-TX05. We interpret this narrow fast anomaly as a downwelling associated with convection. There is also some indirect evidence that such convection must be occurring or has occurred in the recent past. Based on Nd and Sr isotope ratios Perry et al. (1988) and McMillan et al. (2000) have determined that the source region of Rio Grande rift volcanic rocks has changed from lithosphere to asthenosphere during the last 10 Ma and perhaps as recently as 4 Ma. The Perry et al. (1988) study included analyses of basalts from the Lucero volcanic field just south of the Ristra seismic line. Many of the younger volcanic rocks are alkali olivine basalts with isotopically depleted signatures. Alkali olivine basalt melt most likely forms at depths from 50 to 70 km (Perry et al., 1987) implying that the chemical lithosphere has thinned to less than those depths. Total extension across the rift at the location of the Ristra line is on the order of 25% (Cather et al., 1994; Chapin and Cather, 1994) with most of the extension occurring prior to 20 Ma and a recent episode of increased extension during the last 10 Ma. Thus if the present lithosphere thickness is 60 km and it has thinned solely by extension then its original thickness would be just 75 km. The lithosphere thickness beneath the northern Rocky Mountains has been estimated to be 200 km and 100 km beneath the northern Rio Grande rift (Dueker et al., 2001) and Riter and Smith (1996) find evidence for at least a 140 km thick lithosphere beneath the Colorado Plateau 25 Ma. It is thus likely that the lithosphere under the Rio Grande rift was well over 75 km thick prior to the extensional events of the past 40 Ma even if it was originally thinner than beneath the surrounding regions. If this is so it implies that some of the chemical lithosphere, mantle rock that has been in place since perhaps the Proterozoic, has been physically displaced from beneath the Rio Grande rift region. The downwelling imaged in Figure 7 beneath the Great Plains may consist partially of deeper lithosphere originally beneath the Rio Grande rift that has been entrained by the convective flow beneath the rift. Active convection with concurrent removal of some of the lithosphere beneath the rift and surrounding regions may also explain the recent rejuvenation of tectonic activity in the region. A period of extension and volcanism occurred in the Rio Grande rift region from 40 to about 20 Ma. This tectonic activity can be explained in terms of the foundering of the Farallon slab thought to have begun about 40 Ma in the rift region. Following 20 Ma there was a slowing of extension within the rift and a lull in magmatic activity. Beginning 10 Ma extension across the rift accelerated, magmatic activity picked up, and, though controversial, possible significant uplift of the region occurred (Morgan et al. 1986; Baldridge et al., 1991). Also during this time interval the magma source in the southern rift changed from lithospheric to asthenospheric (Perry et al., 1988). This recent activity may be associated with removal of lithosphere by the convection that we infer from our seismic image.

Another interesting aspect of our seismic model is that the strongest slow anomaly is beneath stations NM30-NM35. These stations are located to the west of the Rio Grande rift, within the transition zone. Station NM35 is located on Mount Taylor, part of the Jemez lineament. The very shallow mantle is slow directly beneath the rift but below 100 km depth, the mantle seems to have normal wavespeeds, at least relative to the very slow mantle to the west at those depths. West et al. (2003) have imaged the same structure in their analysis of surface wave dispersion along the Ristra line. If our interpretation of the seismic images in Figure 7 is correct, it appears as if deep upwelling (~400 km depth) is occurring beneath the San Juan basin (stations AZ45-AZ50) and is perhaps associated with volatiles from the Farallon slab to the west. This deep upwelling may be feeding the anomaly to 200 km depth beneath Mount Taylor. The mantle beneath the center of rift may be fed through largely lateral flow from the Mount Taylor region above 100 km depth. In this scenario there would be an overall eastward flow of mantle from the Jemez region, across the rift, with sinking beneath the Great Plains. This is consistent with the generally eastward dip of the deep structures in our image.

The shear wave splitting observations of Gok et al. (2003) show the fast polarization direction to be approximately perpendicular to the Ristra line and therefore appear to be in contradiction to our model. The Gok et al. study used a very limited number of back azimuths in their study and the amplitude of the shear wave splitting varied by a factor of two across the line so that a detailed flow model can not be determined from their data alone. Gao et al. (1997) note that shear wave splitting in continental rifts has generally been found to have fast polarization directions parallel to the rift. They interpret this to indicate the anisotropy is due to alignment of fluid filled cracks perpendicular to the maximum tension axis and not to preferred alignment of olivine in the flow direction. This is consistent with the Gok et al. (2003) results and does not preclude a generally eastward flow generated by downwelling beneath the thick great plains lithosphere. A further point to keep in mind is that our image is two dimensional and thus shows a projection of flow in just one direction, it clearly can not give the absolute flow direction in the region.

Figure 9 shows a cartoon model of our interpretation of the tomography images presented here. Upwelling from the deeper upper mantle, perhaps with higher than normal volatile content, is rising near the location of remnants of the sinking Farallon slab. The upwelling is causing melting in the shallow mantle under Mount Taylor and perhaps near the Navajo field. Both Mount Taylor and the Navajo field lie along Proterozoic suture zones that may have made the lithosphere more fertile there, thus more prone to melting, or more porous (Karlstrom and Humphreys, 1998; Smith et al., 2002). Downwelling of mantle is occurring beneath the thicker lithosphere of the Great Plains. Lithosphere under the Rio Grande rift and perhaps the transition zone, having already been weakened and thinned by previous extension, melting, and possibly hydration, has been mechanically removed to the east by the convective flow indicated by the Great Plains downwelling. Our interpretation is based on a single cross section so that the actual direction of flow is undetermined. Fully three dimensional images should further clarify the dynamics beneath the region.

Conclusion

Tomographic inversion of travel time delays recorded by the long and densely spaced Ristra seismic line show variations in seismic wavespeed at all scales and depths in the upper mantle. The strongest variations are within the upper 200 km of the mantle, unsurprisingly, and correlate well with surface tectonic behavior. In particular, it is clear that the upper 100-200 km of mantle beneath the magmatically and tectonically active Rio Grande rift and transition zone to the Colorado Plateau is seismically distinct from the mantle beneath the stable Colorado Plateau and Great Plains. Furthermore, the boundaries between the active and stable mantle are very sharp and are located near ancient suture and shear zone boundaries indicating that old lithospheric structure plays a key role in tectonic behavior as suggested by Karlstrom and Humphreys (1998) and Dueker et al. (2002). The evolution of the mantle from a state of relative uniformity, in density at least, to the extremely heterogeneous state observed today is not clear. It is possible that some heterogeneity in fracture density, fertility, and lithosphere thickness existed since Precambrian times and that this affected the behavior of the lithosphere during flat slab subduction and subsequent rollback. In this scenario melting and lowering of viscosity would have preferentially occurred in lithosphere that was more fertile, fractured, and thinner. At present, it appears that most of the variation we see is due to temperature variations however. Convective removal of at least some of the lithosphere beneath the tectonically activated regions is likely to have occurred. Our seismic images of the deeper mantle, from 200 to 600 km depth, appear to show evidence for such convection with upwelling associated with the trailing edges of the Farallon plate and downwelling occurring under the thicker lithosphere of the Great Plains.

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Figure Captions

Figure 1. Map of the southwestern United States showing the location of the La Ristra array seismic stations. The location of the main physiographic and tectonic provinces discussed in the text, are also shown. Dashed lines show boundaries of two Proterozoic provinces from Karlstrom and Humphreys (1998). The Colorado mineral belt lies along the northwestern Proterozoic boundary shown in the figure passing through Aspen.

Figure 2. Map showing the earthquakes used in the tomographic inversion. Figure 2a shows the events used that produced clear P waves and Figure 2b shows the events used that produced clear S waves. The star in the center of the figures shows the LA RISTRA array. Not shown are a few earthquakes at distances beyond 100^0 that were used to measure SKS wave travel times.

Figure 3. Example of P and S data recorded by La Ristra stations from a single earthquake. Predicted origin times are marked on the data. Note the similarity in waveform at the beginning of the P and S waves and the variations in arrival time with respect to the predicted times.

Figure 4. Average P and S residuals at each station in the La Ristra array (open symbols). The closed symbols show the average residuals corrected for topography, crustal velocity variations, and Moho depth.

Figure 5. Average corrected P and S residuals as a function of backazimuth for La Ristra stations. The open circles are average residuals from earthquakes to the southeast of the array and stars are residuals from earthquakes to the northwest of the line.

Figure 6. Synthetic resolution test. The top panel shows an input structure used for P and S velocity. The model consists of blocks of 50 km dimension used to compute a synthetic data set. The lower two panels show the result of inverting the synthetic P and S data respectively using the same smoothing and algorithm applied to the real data.

Figure 7. The P and S models obtained through inversion of travel time residuals.

Figure 8. A plot of travel time residual of P versus S for common earthquakes and stations. The least squares fit line through the data has a slope of 2.9.

Figure 9. Schematic illustration of our interpretation of present day mantle structure beneath the Rio Grande rift and surrounding regions.The full three dimensional flow of mantle in the region is unknown but we indicate certain features projected onto the two-dimensional plane beneath the Ristra line that are indicated by the tomography results. The stippled pattern in the shallow mantle indicates lithosphere. Arrows indicate mantle flow direction and wavy lines indicate mantle enriched in volatiles. Volatiles released from the Farallon plate are entrained in mantle upwelling eventually rising under Mount Taylor. Downwelling is occurring under the Great Plains. In our model a significant portion of the lithosphere from Mount Taylor to Carrizozo has been removed and is entrained in the eastern downwelling. FC is the Four Corners region, MT is Mount Taylor, CZ is Carrizozo, CB is Carlsbad,NM, and PS is Pecos,TX.

Figures

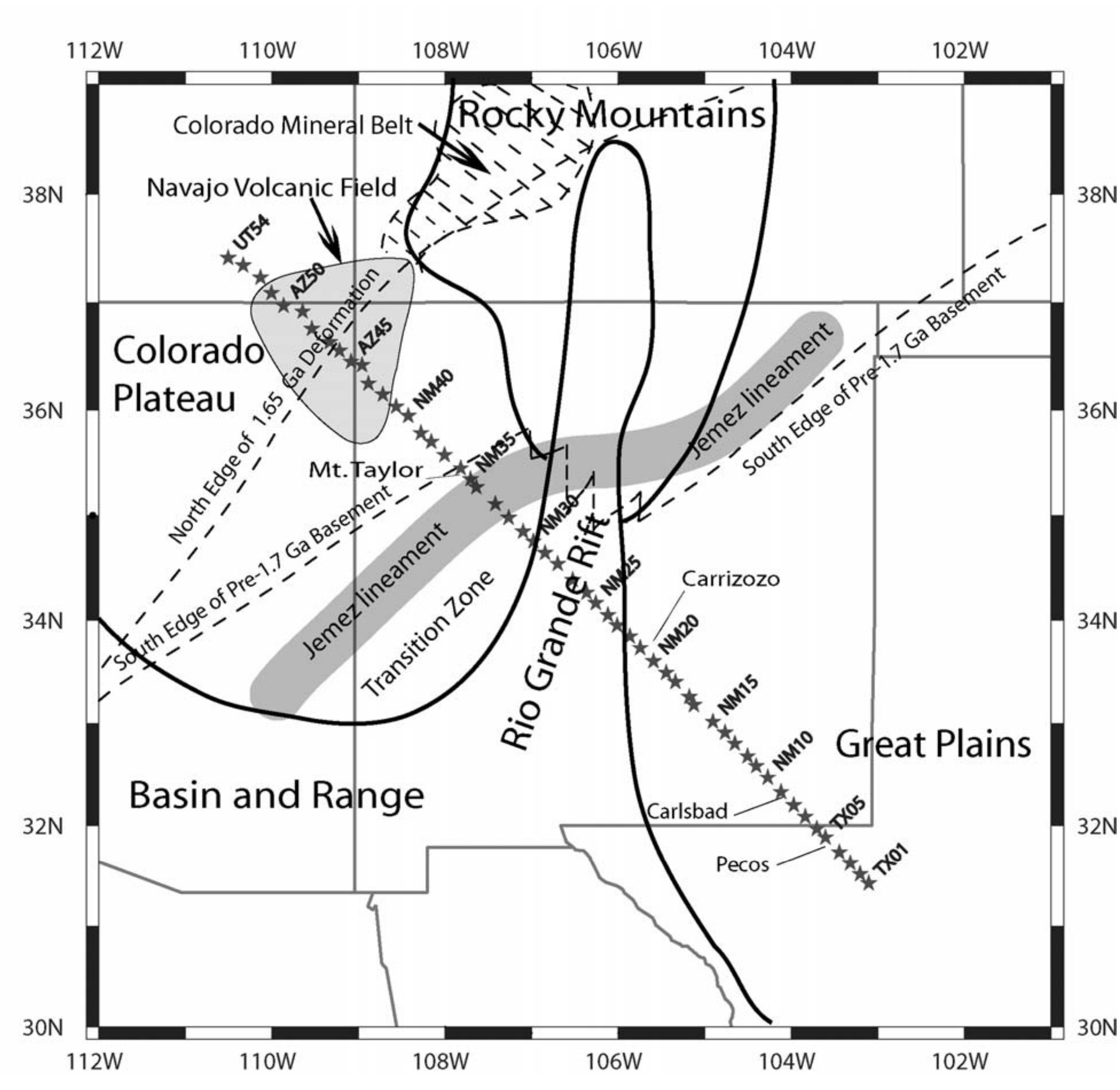


Figure 1

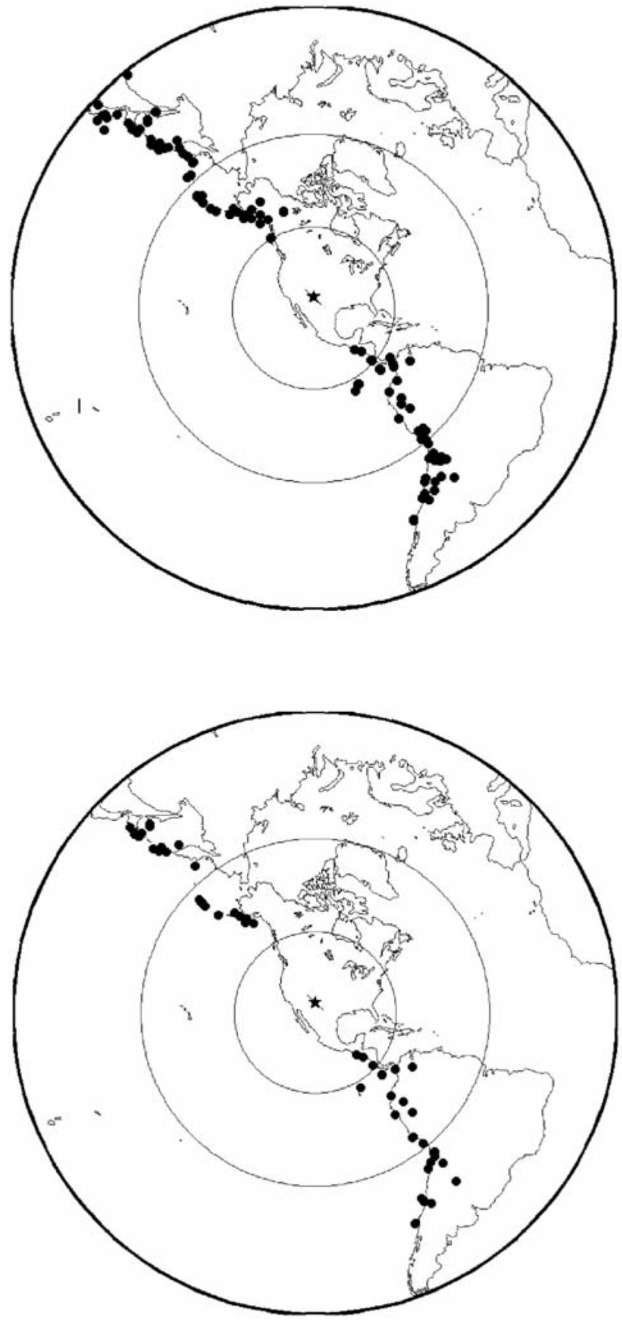


Figure 2

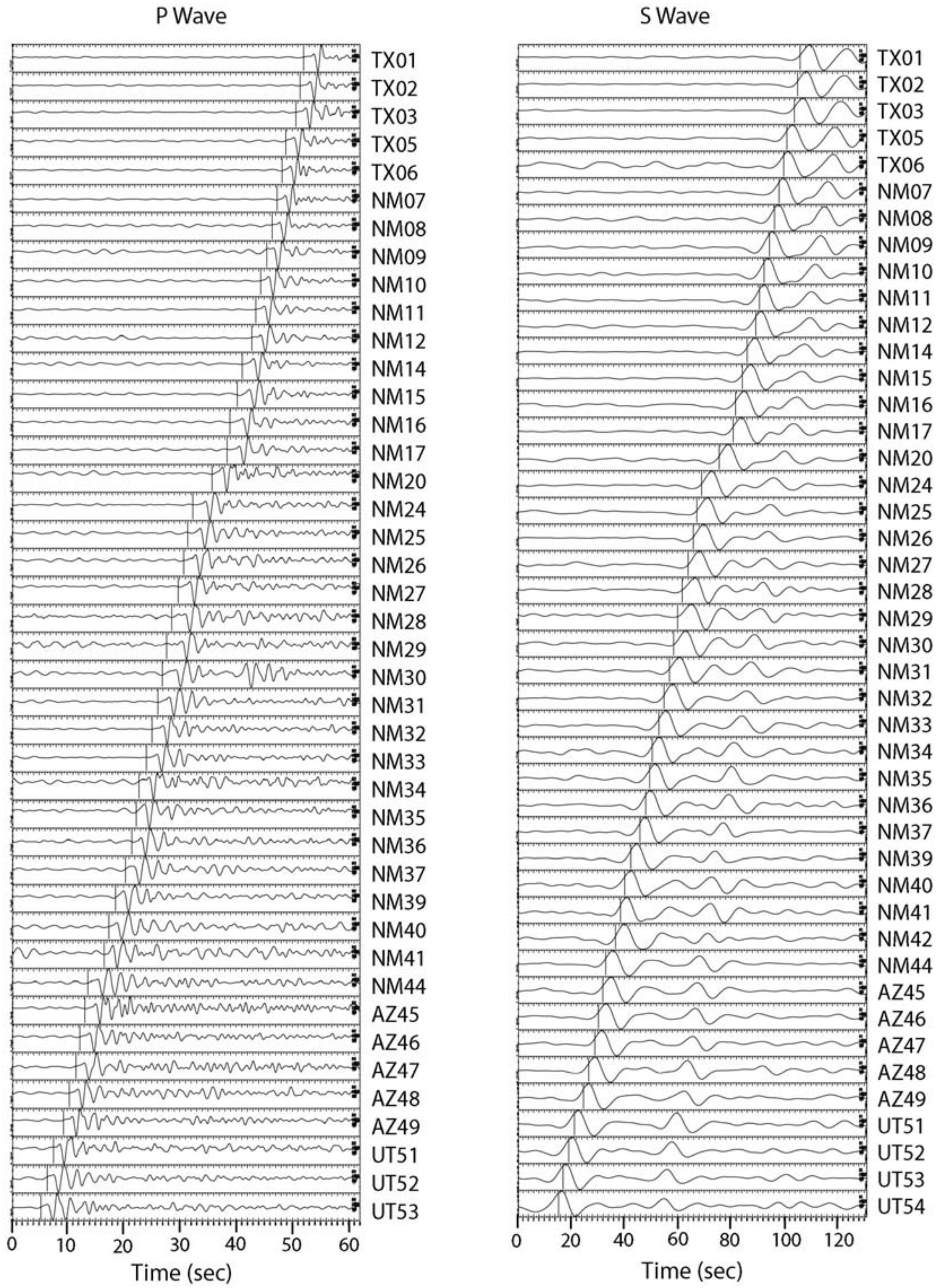
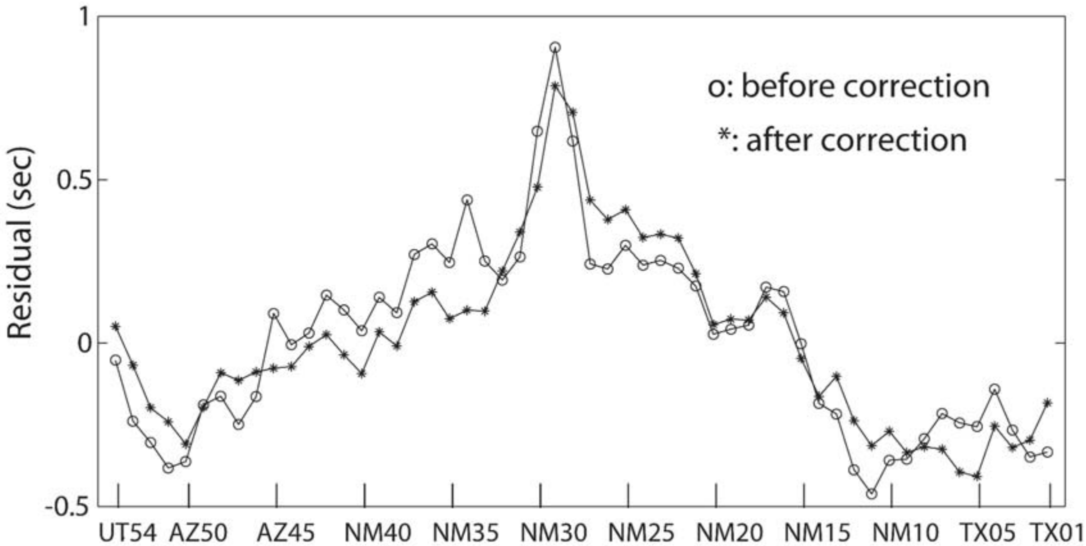

P Wave
S Wave
TX01
TX02
TX03
TX05
TX06
NM07
NM08
NM09
NM10
NM11
NM12
NM14
NM15
NM16
NM17
NM20
NM24
NM25
NM26
NM27
NM28
NM29
NM30
NM31
NM32
NM33
NM34
NM35
NM36
NM37
NM39
NM40
NM41
NM42
NM44
AZ45
AZ46
AZ47
AZ48
AZ49
UT51
UT52
UT53
UT54
0 10 20 30 40 50 60
0 20 40 60 80 100 120
Time (sec)
Time (sec)

Figure 3


o: before correction
*: after correction
Residual (sec)
1
0.5
0
-0.5
UT54
AZ50
AZ45
NM40
NM35
NM30
NM25
NM20
NM15
NM10
TX05
TX01

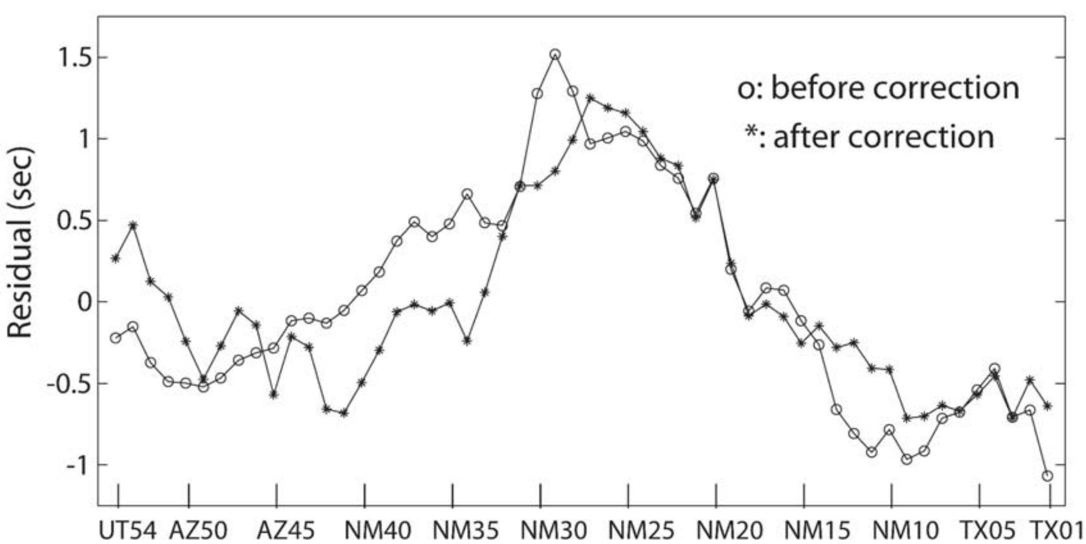
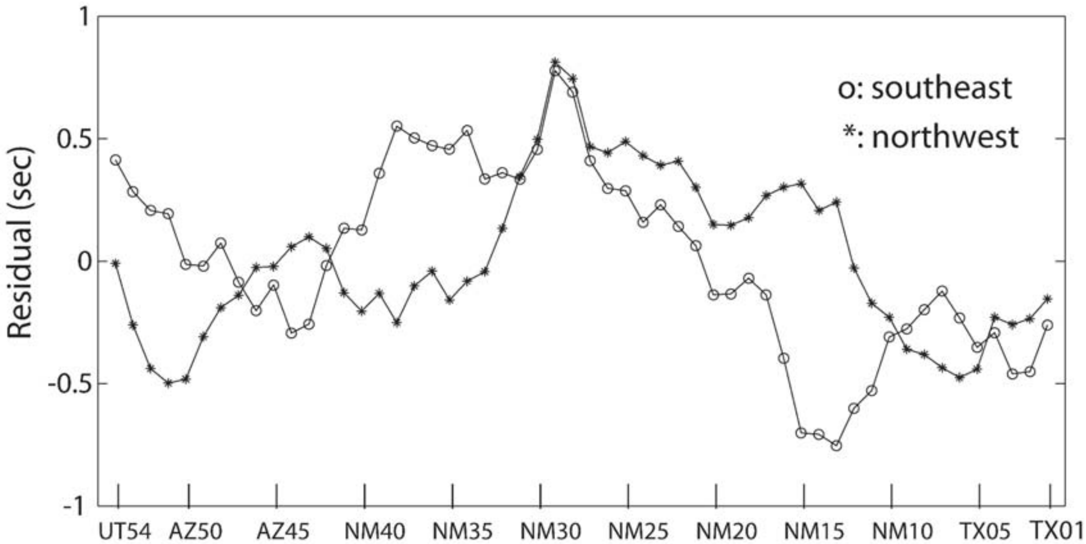

o: before correction
*: after correction
Residual (sec)
1.5
1
0.5
0
-0.5
-1
UT54
AZ50
AZ45
NM40
NM35
NM30
NM25
NM20
NM15
NM10
TX05
TX01

Figure 4


o: southeast
*: northwest
Residual (sec)
1
0.5
0
-0.5
-1
UT54
AZ50
AZ45
NM40
NM35
NM30
NM25
NM20
NM15
NM10
TX05
TX01

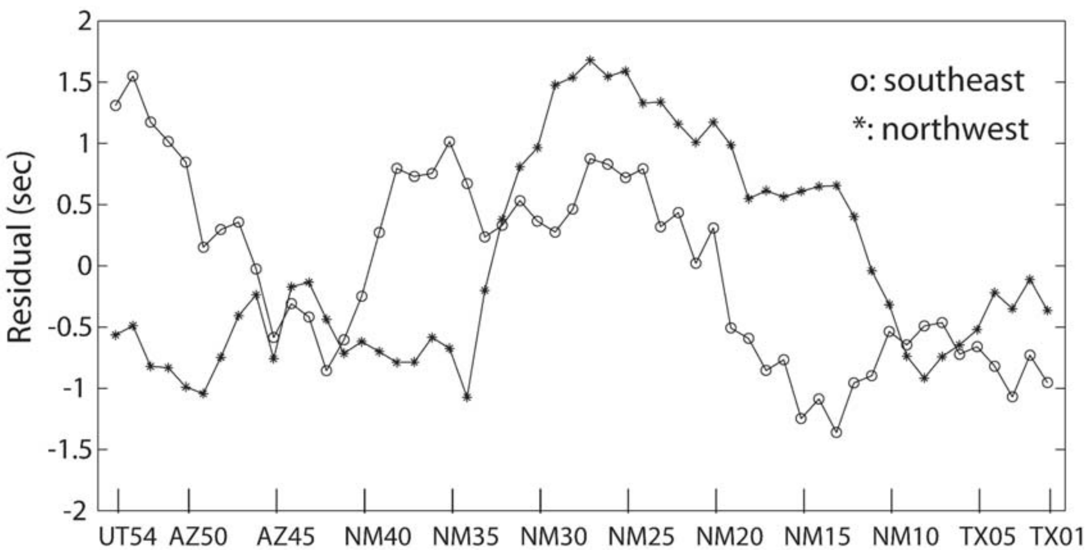
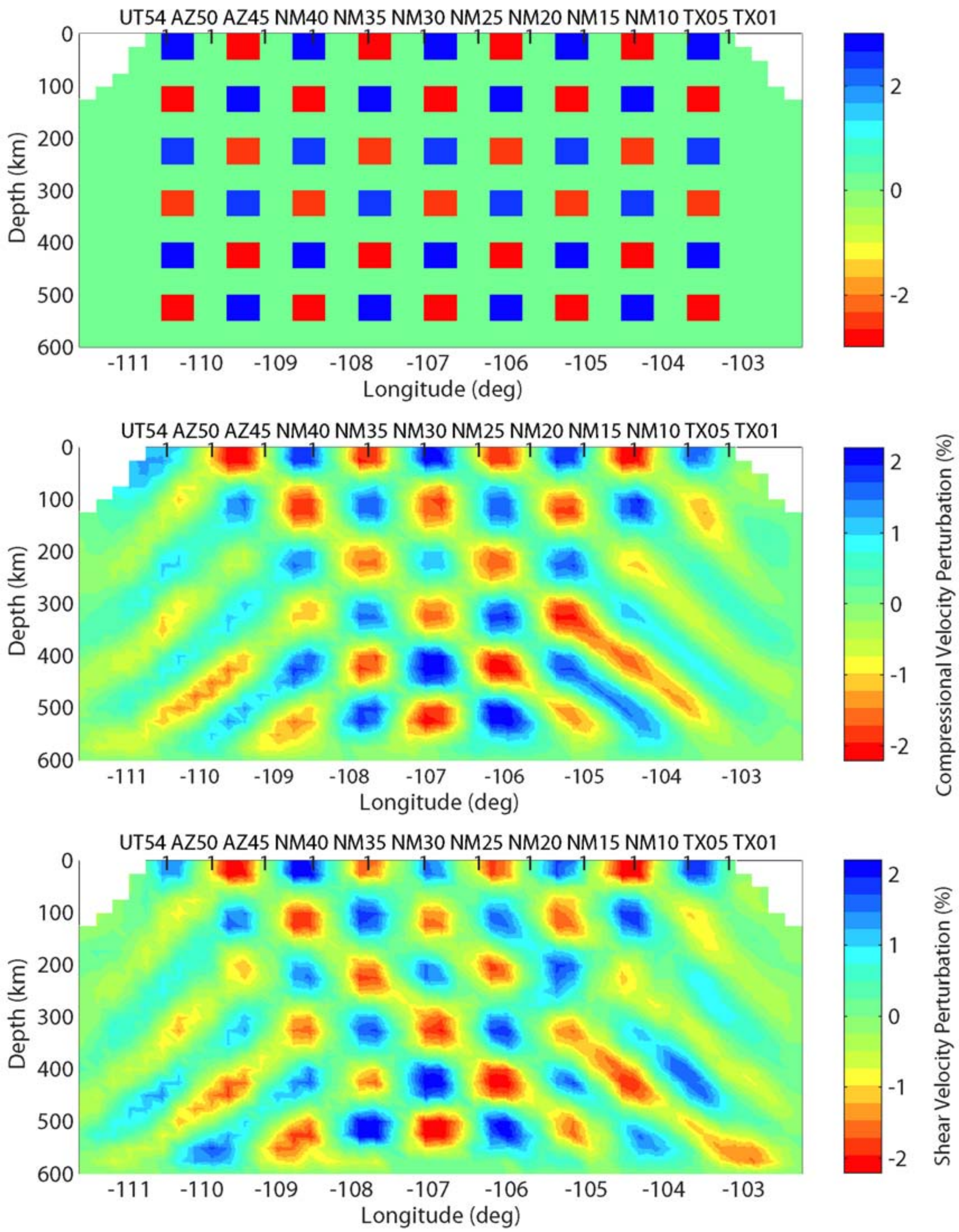

o: southeast
*: northwest
Residual (sec)
2
1.5
1
0.5
0
-0.5
-1
-1.5
-2
UT54
AZ50
AZ45
NM40
NM35
NM30
NM25
NM20
NM15
NM10
TX05
TX01

Figure 5



UT54 AZ50 AZ45 NM40 NM35 NM30 NM25 NM20 NM15 NM10 TX05 TX01
Depth (km)
Longitude (deg)
Compressional Velocity Perturbation (%)
Shear Velocity Perturbation (%)

Figure 6

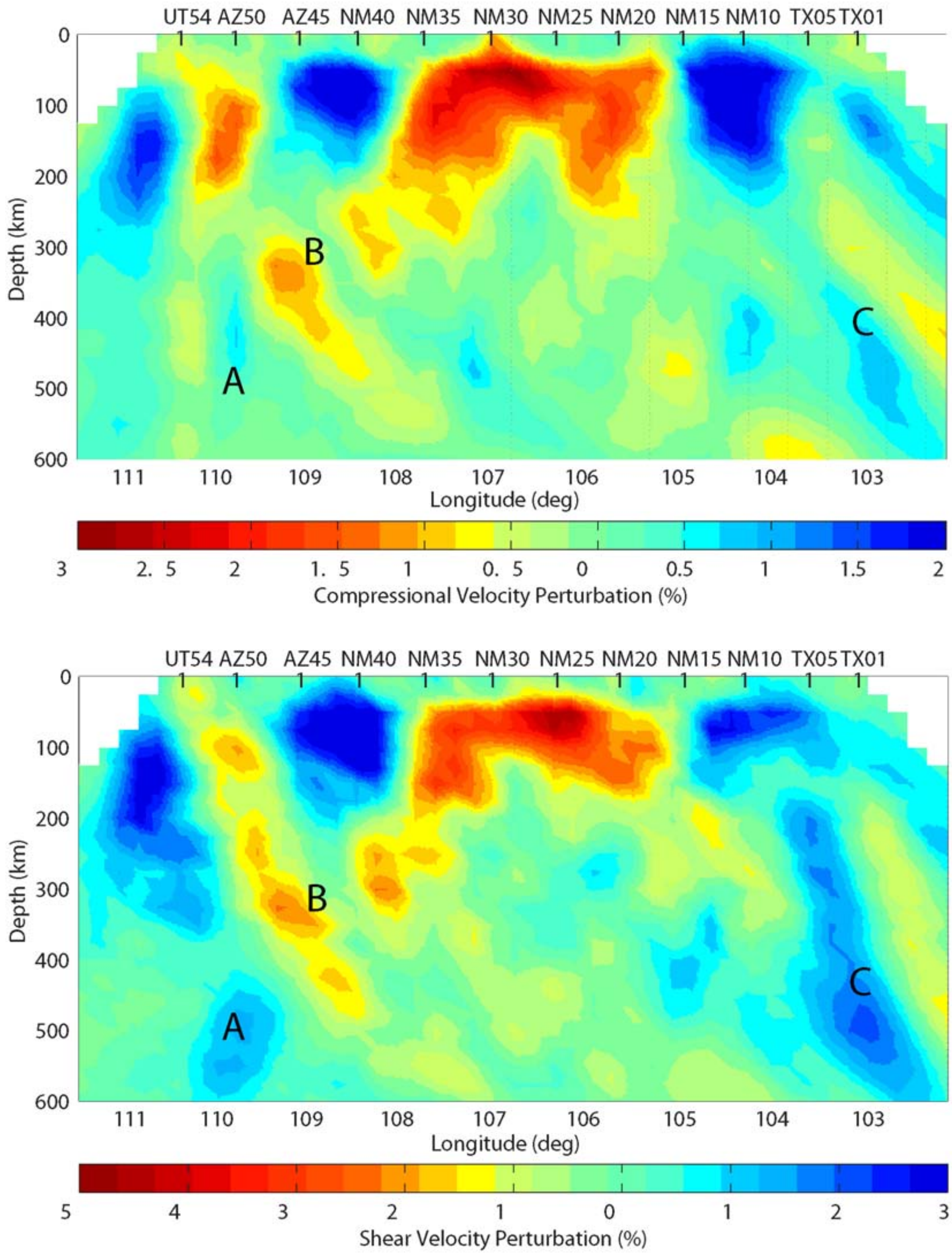

UT54 AZ50 AZ45 NM40 NM35 NM30 NM25 NM20 NM15 NM10 TX05 TX01
Depth (km)
0
100
200
300
400
500
600
A
B
C
111 110 109 108 107 106 105 104 103
Longitude (deg)
3 2. 5 2 1. 5 1 0. 5 0 0.5 1 1.5 2
Compressional Velocity Perturbation (%)
UT54 AZ50 AZ45 NM40 NM35 NM30 NM25 NM20 NM15 NM10 TX05 TX01
Depth (km)
0
100
200
300
400
500
600
A
B
C
111 110 109 108 107 106 105 104 103
Longitude (deg)
5 4 3 2 1 0 1 2 3
Shear Velocity Perturbation (%)

Figure 7

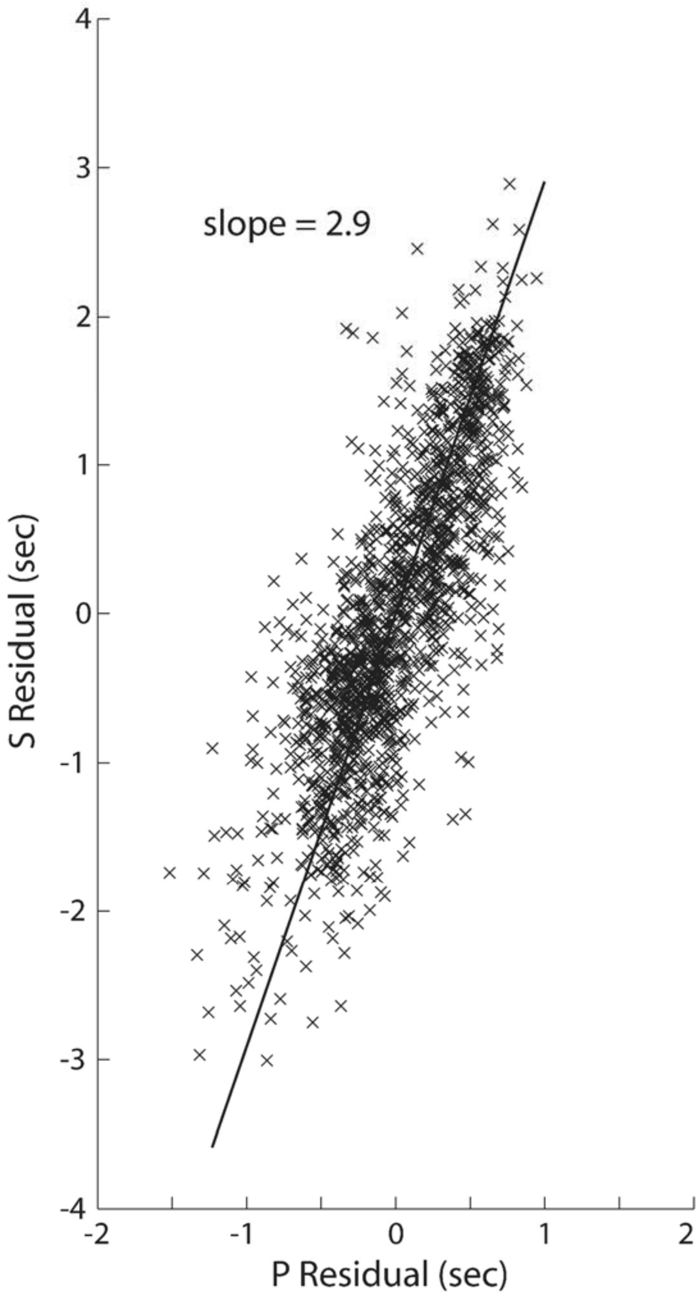


Figure 8

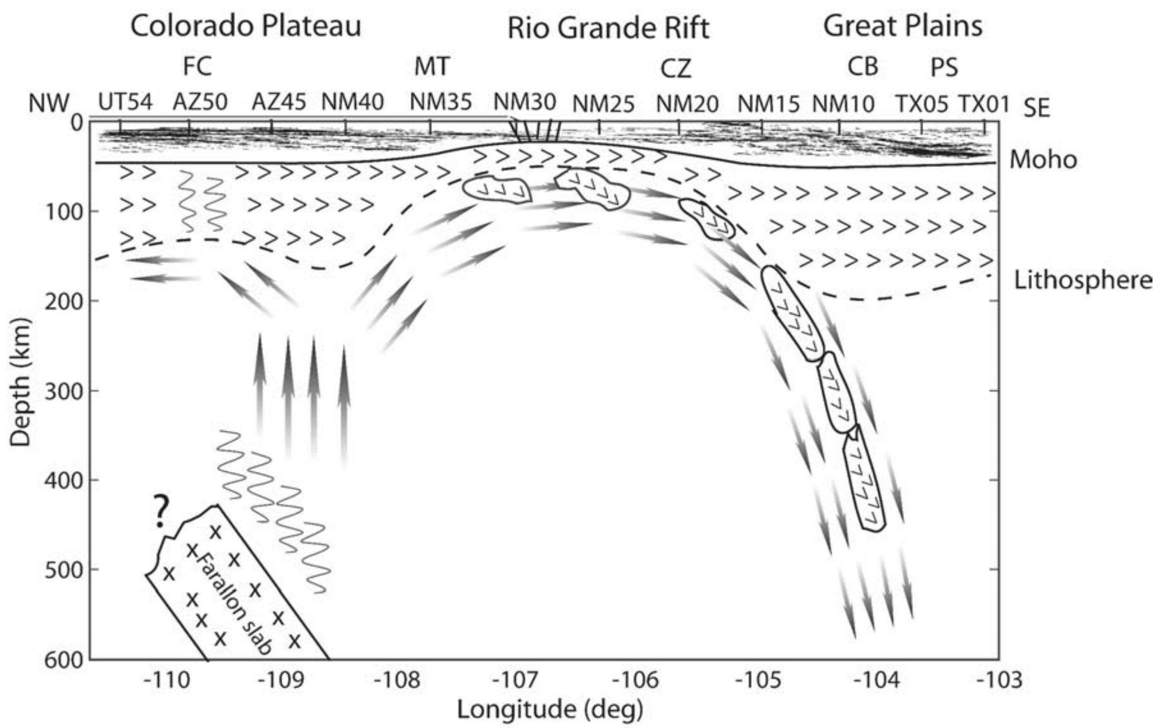

Colorado Plateau
Rio Grande Rift
Great Plains
FC
MT
CZ
CB
PS
NW
UT54
AZ50
AZ45
NM40
NM35
NM30
NM25
NM20
NM15
NM10
TX05
TX01
SE
Moho
Lithosphere
Depth (km)
0
100
200
300
400
500
600
?
Farallon slab
-110
-109
-108
-107
-106
-105
-104
-103
Longitude (deg)

Figure 9