Rayleigh Wave Constraints on Shear-Wave Structure 
and Azimuthal Anisotropy Beneath 
the Colorado Rocky Mountains

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We inverted Rayleigh wave data recorded in the Rocky Mountain Front Broadband Seismic Experiment for shear-wave velocity structure and azimuthal anisotropy. Distinctive structures are imaged beneath the southern Rocky Mountains, the western Great Plains, and the eastern Colorado Plateau. Beneath the southern Rockies, shear velocities are anomalously low from the Moho to depths of 150 km or more, suggesting replacement or delamination of the mantle lithosphere. The lowest velocities are beneath the extension of the Rio Grande rift into southern Colorado and are probably associated with partial melt. Beneath the Colorado Plateau, a thin, high-velocity lid is underlain by a low velocity layer to a depth of at least 160 km. Under the high plains, the velocities are above average down to ~150 km depth, but not as fast as beneath the cratonic core of the continent. A crustal, low-velocity anomaly is observed beneath the high elevations of central Colorado. Elsewhere, inferred crustal thickness correlates with elevation, with the thickest crust beneath the San Juan Mountains in southwestern Colorado. These crustal anomalies suggest that much of the isostatic compensation for the high topography takes place within the crust. We observe a simple pattern of azimuthal anisotropy in the Rocky Mountain region with fast directions rotated slightly counterclockwise from the absolute plate motion of the North America plate and strength increasing with period. The observed anisotropy can be explained by deep asthenospheric flow dominated by current plate motion and shallower and perhaps laterally variable anisotropy in the upper lithosphere.

INTRODUCTION

The Colorado Rocky Mountains are located between the tectonically stable Great Plains to the east and the elevated, but stable Colorado plateau to the west. The Rio Grande rift extends into the Rockies in southern Colorado and associ-
stand the formation of the southern Rockies and the stability and evolution of the Great Plains. For example, tomographic studies of North America have shown that the Rocky Mountain Front represents a transition from a seismically slow upper mantle in western United States to a fast, cratonic structure in central and eastern North America [Grand, 1994; van der Lee and Nolet, 1997]. This transition is imaged in more detail in regional body wave tomography in Colorado and western Kansas [Lee and Grand, 1996; Lerner-Lam et al., 1998]. These studies reveal a low velocity volume in the mantle beneath the southern Rockies, which the authors interpret as indicating isostatic support of the Rockies through thermal buoyancy and partial melt in the upper mantle. Seismic anisotropy, which may be sensitive to deformation in the lithosphere and flow in the asthenosphere, also provides important constraints on regional dynamic processes. There are rapid lateral variations in the fast direction of shear wave splitting in Colorado and many null measurements [Savage et al., 1996]. Savage and Sheehan [2000] suggested that this complicated pattern indicated a complex strain regime that might be consistent with asthenospheric upwelling sheared by plate motion.

In this paper, we use Rayleigh wave data recorded from the Rocky Mountain Front (RMF) Broadband Seismic Experiment (Figure 1) [Lerner-Lam et al., 1998] to obtain spatial and azimuthal variations in phase velocities, which we interpret in terms of crustal and upper mantle shear-wave structure beneath the Colorado Rocky Mountain region and surrounding terranes. Because there is a high density of stations in the RMF experiment, the lateral resolution in our study is greater than that in large-scale tomography studies [Grand, 1994; van der Lee and Nolet, 1997]. Because Rayleigh waves at different periods sample velocity structure in different depth ranges and the fundamental mode is insensitive to deep structure, measuring phase velocities from 20 to 100 s period provides better resolution for velocity structure in the upper 100 km than can be obtained in regional, body-wave tomographic studies with station spacing comparable to the RMF experiment.

Figure 1. Station locations on a relief map of the Colorado Rocky Mountains and surrounding terranes. The two areas with the highest elevations include the Sawatch Range and the San Juan Mountains, respectively. The black triangles represent seismic stations of the Rocky Mountain Front Seismic Experiment [Lerner-Lam et al., 1998]. Thin dashed lines mark tectonic boundaries and thin solid lines are state boundaries. Thick dashed lines are locations of profiles A-C in Plate 3.
addition, the frequency dependence of the azimuthal anisotropy of Rayleigh waves can provide constraints on variation of anisotropy with depth that complement the vertically averaged measure of anisotropy obtained from shear-wave splitting.

Our approach involves two steps. The first step is to invert Rayleigh wave amplitude and phase data for phase velocities of the waves. This gives a direct indication of the information contained in the data independent of any assumptions needed for a stable inversion for earth structure. The second step is to invert for shear-wave velocity structure from the phase velocities obtained in the first step. We construct models initially ignoring azimuthal anisotropy, seeking only to resolve lateral variations. Then we allow for the possibility of anisotropy and show the effects of this added model complexity on the estimates of lateral variations in phase velocity.

DATA SELECTION AND PROCESSING

Rayleigh wave data from 74 teleseismic earthquakes with body wave magnitudes larger than 5.0 and epicentral distances from 30° to 120° (Figure 2) were used in this study. The events were recorded by 35 RMF broadband seismic stations that operated during May to December in 1992 (Figure 1). Most of the stations were located in Colorado, with one in eastern Utah and two in western Kansas. The sites were not all occupied simultaneously, so the actual coverage is not as good as would be expected given the distribution of sources and receivers. Nevertheless, the coverage for surface waves is excellent (Figure 3), yielding many crossing paths that are needed for tomographic studies both inside and immediately outside the array.

Because several types of seismometers were used in the RMF stations, we corrected instrument responses to match a single type. We filtered vertical-component seismograms with a series of 10 mHz wide, zero-phase-shift, 4th-order Butterworth filters centered at frequencies of 50, 45, 40, 35, 30, 25, 20, 17, 15, and 10 mHz. These frequencies sample velocity structure to depths of 300–400 km and provide good vertical resolution to about 150 km depth. Fundamental mode Rayleigh wave trains were isolated by windowing each filtered seismogram. Frequency bands with signal-to-noise ratio less than 3:1 were rejected. A careful comparison of Rayleigh wave trains at all possible stations for each frequency was important for identifying problems such as timing errors and anomalous instrument responses. We then converted the filtered and windowed seismograms to the frequency domain to obtain phase and amplitude data.

PHASE VELOCITY

Phase velocity $c$ in a uniform slightly anisotropic medium varies as

$$c(\omega, \psi) = A_0(\omega) + A_1(\omega)\cos(2\psi) + A_2(\omega)\sin(2\psi) + A_3(\omega)\cos(4\psi) + A_4(\omega)\sin(4\psi)$$

(1)

where $\omega$ is frequency, $\psi$ is the azimuth of propagation of the wave, and $A_0$ to $A_4$ are velocity coefficients [Smith and Dahlen, 1973]. We neglected $A_3$ and $A_4$ terms here because they should be small for Rayleigh waves [Smith and Dahlen, 1973]. Phase velocities are represented as weighted averages of values of $A_0$, $A_1$, and $A_2$ at neighboring points on a grid of nodes (Figure 4). The spatial resolution is controlled by adjusting the characteristic scale length of the 2-D Gaussian weighting function, which is supposed to be a function of wavelength. However, to simplify the calculation, we use 80 km, an intermediate value, for all frequencies in this study. To account for wave propagation effects such as focusing and multipathing that occur between the sources and the array, each incoming wavefield is represented as the sum of two interfering plane waves with amplitude, initial phase, and propagation direction for each wave to be determined in the inversion [Forsyth et al., 1998; Li, 2001]. We solve simultaneously for these wavefield parameters and the velocity parameters in an iterative, least-squares inversion. Two stages are employed for each iteration: a simulated annealing method is used first to solve for the two-plane wave parameters; then a generalized linear inversion [Tarantola and Valette, 1982] with damping and
smoothing is applied to find both phase velocity coefficients at each node and the wave field parameters.

**Isotropic Phase Velocities**

To generate a reference model that will serve as a starting model in inversions for lateral variations in phase velocity, we first assume that phase velocities are uniform in the whole study area to obtain an average phase velocity at each frequency (Figure 5). Because the study area is characterized by three distinct tectonic provinces, the Colorado Plateau, the Rocky Mountains, and the Great Plains, we also solved for average phase velocities in each province by grouping the nodes by tectonic province (Figure 4). The nodes outside the array are treated as another group, but this region is heterogeneous and less well-constrained, so the results are not shown in Figure 5. This regionalized inversion is equivalent to a classic “pure-path” inversion, except that we use the average velocity for the whole study area as a starting model. The damping in the version is light and has virtually no effect except at the longest periods and in the least-well-constrained province, the Colorado Plateau.

The average phase velocities from 20 s to 100 s beneath the Rocky Mountains are overall much lower than those beneath the Colorado Plateau and the Great Plains (Figure 5), indicating the presence of high temperatures and/or partial melt in the crust and upper mantle. Phase velocities in the Colorado Rockies and the Great Plains increase gradually and smoothly with increasing period. The largest differences between these two provinces are at periods of 25 to 40s, which are primarily sensitive to structure in the crust and uppermost mantle down to depths of about 100 km (Figure 6). Diminishing differences at longer periods indicate a decrease in velocity contrast at depths greater than 100 km. The dispersion for the eastern Colorado Plateau lies between that for the Plains and the Rockies, but the velocity does not increase as smoothly with increasing period. Part of this oscillatory character may be due to larger uncertainties for this region, but the general feature of a decreased slope around 50 to 60 s is similar to that reported in an independent experiment in the Colorado Plateau [Lastowka et al., 2001] and is indicative of a pronounced low-velocity zone in the upper mantle underlying a high-velocity lithosphere.

Using the average phase velocity for the region as a whole as a starting value, we solved for 2-D phase velocity variations across the array without a priori regionalization. Phase velocities at the grid points (Figure 4) are used to generate maps on a finer grid of 0.1° by 0.1° using a smoothing length of 80
km. Plate 1 shows the maps of phase velocity anomalies at periods of 25, 33, 40, 50, 59, and 67 s. A striking and consistent feature in these maps is a band of low velocities confined to the Colorado Rockies region. Confirming the results of the a priori regionalization, average velocities are highest in the western Great Plains, lowest in the Rockies, and intermediate in the eastern Plateau. At 25 s, there is an almost circular pattern of low velocities in central Colorado. This pattern suggests that the crust must be thicker and/or slower in this region than elsewhere in the southern Rockies, because phase velocities at 25 s are primarily sensitive to the upper 50 km (Figure 6). At 40 s, there are low phase velocities in the northern extension of the Rio Grande rift into Colorado. Because this anomaly is absent at 25 s, it suggests that there must be a pronounced, low shear-velocity anomaly in the uppermost mantle beneath the rift, probably indicating the presence of partial melt. Compared to the shorter periods, the lateral contrast in phase velocities at 50 to 67 s is reduced, but the contrast at 67 s is still about 3% from the slowest area beneath the Rockies to the fastest beneath the Great Plains. Because these periods are relatively insensitive to the crust, these maps provide a clear image of a transition from a tectonic to a cratonic upper mantle in the vicinity of the RMF.

Resolution

To understand the significance of lateral variations in the phase velocity maps presented in Plate 1, a description of the resolution is needed. In our approach, the value at each point in a map represents a Gaussian weighted average over adjacent node points, with the value of the weight decreasing to 1/e of the maximum at the characteristic distance of 80 km. Lateral variations in resolution are indicated by lateral variations in the standard error or uncertainty in these averages, taking into account the covariance between values of adjacent node points. In the generalized linear inversion for phase velocities, the a posteriori covariance matrix, C_{MM}, for model parameters (phase velocity at each node) can be directly calculated from

\[
C_{MM} = (G^T C_{nn}^{-1} G + C_{nn}^{-1})^{-1}
\]

where G is the partial derivative or sensitivity matrix relating predicted changes in phase and amplitude to perturbations in phase velocity, and C_{nn} and C_{mm} are the a priori data and model covariance matrices, respectively [Tarantola and Valette, 1982]. The a priori model covariance acts to damp the least squares solution by assigning an uncertainty to the starting model and retaining it as a constraint in the inversion. In an undamped least squares inversion, very large velocity anomalies tend to be assigned to poorly constrained regions of the model. In a highly damped solution, lateral velocity variations are usually underestimated and velocity variations in the poorly constrained regions are very small. We choose rel-
Plate 1. Variations of phase velocities and phase velocity uncertainties. Plate 1 A to E show phase velocity anomalies at 6 periods (25 s, 33 s, 40 s, 50 s, 59 s, 67 s) relative to the average phase velocities (solid line) in Figure 5. Although the inversion is performed in the whole study region, the shown area is outlined by the error contour of 2.2% in Plate 1G which shows twice the standard errors of phase velocity anomalies at 33 s. Plate 1H is a map of azimuthally anisotropic phase velocities at 50 s. Both the average velocity and azimuthal anisotropy were smoothed with a Gaussian characteristic length of 80 km. Azimuthal anisotropy is represented by the black bars. The orientation of the bars indicates the fast direction of azimuthal anisotropy and the length of the bars is proportional to the strength of anisotropy. The absolute plate motion of the North America Plate according to Gripp and Gordon [1990] is marked as the big white arrow. Black dashed lines indicate tectonic boundaries.
atively light damping that tends to leave the amplitude of the velocity variations roughly constant throughout the study area, but assigns larger a posteriori errors to the parts of the model that are poorly constrained.

We plot in Plate 1G twice the standard error of the weighted average velocities for a period of 33 s. Maps of the standard error at other periods are similar in form, but the errors increase for longer periods in the same manner as in Figure 5. The reason that variance increases with period is that the same magnitude error in Rayleigh wave phase produces larger uncertainty in travel time at longer periods and the signal-to-noise ratio and the number of acceptable signals also decrease. Not surprisingly, the errors are smallest where the density of crossing paths is greatest (compare to Figure 3). In the maps of velocity in Plate 1A-F, we use the 2.2% contour from 33 s as a mask, eliminating the illustration of velocity variations in regions outside this contour for being relatively poorly constrained. Twice the standard deviation yields a rough guide to the 95% confidence level for velocity anomalies; if the change in velocity from one point to another in the maps is greater than two standard errors, then there is only about a 5% likelihood that this difference would have arisen by chance. Thus, at 40 s, for example, the difference in phase velocity between the Rio Grande rift region and the central Rockies at 40° N is significant at the 95% confidence level.

**SHEAR-WAVE STRUCTURE IN THE CRUST AND UPPER MANTLE**

**Methodology**

Although the variations in Rayleigh wave phase velocity yield measures of the lateral and vertical variations in structure that are easily interpreted qualitatively, for geological

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**Figure 5.** Average phase velocities in tectonic regions. Circles and the solid line are for the whole area (Ave.), squares and the thick dashed line for the Colorado Plateau (CP), triangles and the thin dashed line for the Rockies (RM), and diamonds and the intermediate dashed line for the Great Plains (GP). The bars represent one standard error on each measurement. Note that the errors are larger in the Colorado Plateau.
interpretation, we need to invert these observations for the velocity structure of the earth. Because the problem is non-linear and the structure is imperfectly resolved, the particular models we present will depend somewhat on the starting model. To obtain a reference model for this area, we started with the phase velocities near station CCR and an initial, average earth model (AK 135 from Kennett et al. [1995]) that was modified to have a 48 km crust as constrained from receiver functions at CCR [Sheehan et al., 1995]. We chose CCR because it lies in the middle of the study area (Figure 1) where velocities are best constrained and the velocities are intermediate between that of the central Rockies and the Great Plains. Model parameters are shear velocities in approximately 20-km-thick layers extending to a depth of 200 km, with thicker but ultimately unresolvable layers extending to 600 km. P-wave velocities were coupled to S-wave variations with the same Poisson’s ratio in each layer as in the starting model, so the sensitivity kernels shown in Figure 6 actually represents the combined response to coupled velocity changes, not just the shear velocity. In isotropic models, the dominant effect is S wave velocity, but P velocity is important near the surface. For example, much of the sensitivity to crustal velocity at 67 s (Figure 6) is due to P wave sensitivity, but the mantle response is almost entirely due to S wave sensitivity. There is not sufficient resolution to uncouple the P and S velocities in the inversion. We add crustal thickness as a parameter, preserving a velocity discontinuity at the Moho, by increasing the thickness of the lower crustal layer at the expense of the thickness of the uppermost mantle layer. We adjusted the relative damping of crustal thickness and velocity variations to maximize agreement between Rayleigh wave and receiver function data, although the latter were not explicitly employed as constraints in the inversion.

Synthetic phase velocities were computed using Thomson’s algorithm [Thomson, 1997; Martin et al., 1997] which allows for general anisotropy. Partial derivatives of phase velocity with respect to changes in model parameters were obtained by finite differences from a series of forward models in which individual parameters were perturbed. When inverting the maps of phase velocity for 3-D structure, we assumed for computational efficiency that as long as the crust is constant thickness, the partial derivatives are constant, because the primary non-linearity in the problem stems from variations in thickness of the layers. We tabulated partial derivatives for 17 models with crustal thickness ranging from 34 to 66 km, then after each iteration in an inversion, selected two sets of partial derivatives whose corresponding crustal thickness are most close to the crustal thickness in the current model, and obtained the most appropriate partial derivatives by interpolating the selected two sets. The shear velocity models are damped by assigning a priori standard deviations of 0.1 km/s to the starting model velocities and smoothed by introducing off-diagonal terms in the model covariance matrix that enforce a 0.4 correlation in the changes to adjacent layers.

In this reference model at CCR with crustal thickness fixed at 48 km, the S-wave velocity in the upper mantle is indistinguishable from the AK135 model, never differing by more than 0.01 km/s (Figure 7). The S-velocity in the crust is 3.32 km/s in the top layer (0–20 km) and 3.73 km/s in the bottom layer (20 km-Moho), about 0.13 km/s slower than in model AK135. The similarity of mantle structure is coincidental; there are substantial deviations from these reference models in other parts of the study area.

Average Structure in Tectonic Provinces

One-dimensional shear-wave models for the Colorado Plateau, southern Rockies, and the Great Plains are distinctly different (Figure 7). Shear velocity under the Rockies is remarkably low from the surface to ~140 km depth. Velocities in the model are also slightly slow compared to AK135 at greater depths where our data do not have good resolution; although the difference is probably real, the depth distribution...
of the anomaly is not well constrained. If such differences do exist, as is suggested by body-wave tomography [Lee and Grand, 1996; Lerner-Lam et al., 1998], the lack of resolution from surface waves coupled with the damping of the model inversions would lead to an underestimate of the magnitude of the anomaly. Our model agrees with the preferred pure-path model for the southern Rockies shown in the paper of Lerner-Lam et al. [1998]. There is no discernible, fast, lithospheric lid.

Vertical resolution depends on the period range sampled and the precision with which the phase velocities are measured. One measure of the resolution is the rank of the inverse matrix, which indicates the number of linearly independent combinations of model parameters that are resolved, or, equivalently, the number of pieces of information about the velocity structure that the data provides. For the southern Rockies and western Great Plains regions, the rank is 3.8; for the Colorado plateau, with larger errors on the velocities (Figure 5) the rank is 2.8; and for typical points in the maps of Plate 1, the rank is 2.9. The vertical distribution of the information is described fully by the resolution matrix.Crudely, three well-resolved, independent pieces of information about the vertical velocity structure are: the average velocity in the crust; the Moho depth and/or average velocity from 40 to 70 km; and average velocity from 70 to ~150 km. For the better-resolved, average structure of the Rockies and Great Plains, there is some additional information extending to depths greater than 150 km. A posteriori standard errors for individual layers are smaller where the information density is higher, like in the crust (Figure 7). Average velocities over resolvable depth ranges are better constrained than the standard errors indicated for individual layers. The typical standard error for crustal thickness is a few kilometers, but it should be recognized that standard errors in any inversion of this type do not provide extreme limits.

The model of the Colorado Plateau shows a fast lid from Moho to ~100 km and a strong low velocity layer underneath it. Velocities in the low-velocity zone approach those beneath the Rockies in the same depth range. This relatively thin lid beneath the Colorado Plateau supports the model that nearly horizontal subduction of the Farallon slab mechanically thinned the lithosphere [Humphreys and Dueker, 1994; Spencer, 1996], in agreement with the surface wave interpretation of Las-towka et al. [2001]. Our model velocity in the lid is unusually high; this may be caused partially by chance variations in the observed phase velocities within the bounds expected given the larger standard errors (Figure 5), but there is also some tradeoff possible between lid velocity and crustal thickness. Our best fitting crustal thickness for this average dispersion curve is about 47 km. If we constrained the thickness to be in the 40–44 km range typical of seismic refraction and receiver function estimates for the Colorado Plateau [Sheehan et al., 1997; Keller et al., 1998], the lid velocity could be reduced to 4.6 to 4.7 km/s.

In the model for the average western Great Plains, shear wave velocity is everywhere faster in the crust and upper mantle than in AK135 and the reference model, with a subtle low velocity zone, indicating a cratonic lithosphere. Defining a thickness of the lid or lithosphere is difficult because there is a relatively small velocity contrast between lid and low-velocity zone. The structure is quite similar to that found for the western Australian craton from Rayleigh wave tomography [Simons et al., 1999], but the velocities are clearly lower than beneath the core of stable North America [Brune and Dorman, 1963]. The lower average velocities compared to the continental interior are not surprising considering that the Rayleigh wave

Figure 7. One-dimensional shear-wave velocity structures. The model of AK135 is plotted as a dotted line. The thick solid line represents our reference model at station CCR. The thin solid line is for the southern Rocky Mountains (RM). The thin and thick dashed lines correspond to the western Great Plains (GP) and the eastern Colorado Plateau (CP), respectively. These models are based on the phase velocities for average regions shown in Figure 5. One standard errors of shear velocity under the Rocky Mountains are plotted at the center of each layer, based on the a posteriori covariance of the damped inversion.
phase velocities systematically decrease and the topography systematically increases from western Kansas to the Rocky Mountain Front (Plate 1 and Figure 1). The form of this variation is shown by our inversion for three-dimensional structure.

3-D Structure

Variations of crust thickness and velocity anomalies are shown in Plate 2 and 3. The crust is generally thick (48–52 km) beneath the Colorado Rockies and thins gradually to 44 km in the nearby Colorado Plateau and to 40 km in the Great Plains. There is a good, general correlation of crustal thickness with regional elevation. The thickest crust is beneath one of the two most elevated areas of the Rockies, the San Juan Mountains (Plate 2A and 3A). An exception to this correlation is local crustal thinning beneath the other highest area of the southern Rockies near the Sawatch Range (Plate 2A and 3B), but in this area there is a strong, low velocity (-3%) anomaly in the crust (Plate 2B, 2C, and 3B) that may indicate low crustal densities which could provide local isostatic compensation for the high topography [Li et al., 2002]. Some trade-off between the low velocity anomaly and crustal thickness is possible, but the crustal anomaly is resolvable different than the effects of a change in crustal thickness, and the local thinning is confirmed by receiver function analysis beneath stations of the RMF experiment [Sheehan et al., 1995; Keller et al., 1998].

The variations of Moho depth in Plate 2A are significantly different than those reported in compilations of results from seismic reflection and refraction profiles and receiver function studies [Prodehl and Lipman, 1989; Sheehan et al., 1995; Keller et al., 1998]. Those compilations show the thickest band of crust to the east of the Rocky Mountain Front and a generally poor correlation of crustal thickness with elevation in Colorado. The combination of this lack of correlation with the discovery that P and S velocities beneath the southern Rockies are lower than beneath either the Great Plains or the Colorado Plateau led to the hypothesis that isostatic compensation of the southern Rockies takes place largely in the mantle [Eaton, 1987; Sheehan et al., 1995; Lee and Grand, 1996; Lerner-Lam et al., 1998; Karlstrom and Humphreys, 1998]. In contrast, Li et al [2002] showed that the Bouger gravity anomaly in Colorado could be matched by a combination of crustal thickness variations as mapped in Plate 2A and intracrustal density variations that are proportional to the velocity anomalies mapped in Plate 9B and 9C, indicating that no significant contribution of buoyancy from the mantle is required.

It is possible to reconcile the estimates of crustal thickness based on surface waves with the estimates from seismic reflection, refraction and receiver function analysis. First, there is some positive correlation between S-wave vertical travel times through the crust estimated from receiver function analysis and the times predicted from our crustal models. Second, the compilations include data from various types of studies with differing quality and there may be some ambiguous possible interpretations that are necessarily hardened into a definite value in compiling a map. The Rayleigh wave data set is more uniform. Third, and most importantly, Rayleigh waves do not directly detect seismic discontinuities. They are sensitive to average velocities over depth ranges, so a 5 km change in Moho depth with a 0.7 km/s velocity contrast at 50 km is roughly equivalent to a 0.1 km/s velocity change over a depth range of 35 to 70 km. The important point, however, is that the Rayleigh wave data require either substantial variations in crustal thickness that correlate with topography, as shown in Plate 2A, or substantial variations in velocity of the lowermost crust and uppermost mantle that correlate with topography. Either way, the combination of upper crustal velocity variations and variations near Moho depths suggest that much of the isostatic compensation of the southern Rocky Mountains takes place in the shallow lithosphere.

In the upper mantle from the Moho to 140 km depth, our velocity images reveal two primary features: a band of low velocities under the Colorado Rockies with a gradual transition to high velocities beneath eastern Colorado and western Kansas (Plate 2D to H, and 3C); and a local low velocity anomaly centered beneath the northern extension of the Rio Grande rift into southern Colorado that is most pronounced just below the Moho (Plate 2D to F, and 3A). Previous studies have found similar slow anomalies beneath the Rio Grande rift in New Mexico [Parker et al., 1984; Davis et al., 1993], although these anomalies have also been attributed to the northeast trending Jemez lineament in New Mexico [Humphreys and Dueker, 1994; Dueker et al., 2001]. Our observations agree with the S-wave tomography model of Lee and Grand [1996] in terms of the overall pattern of lowest velocity beneath the Rockies, but differ from it in detail. For example, their lowest velocity anomaly in the upper 100 km is essentially uniform in amplitude extending from central Colorado southward into the Rio Grande rift. In our models, there are three components to this anomaly: the crustal velocity anomaly centered on the Sawatch range, the Moho depth anomaly near the San Juan mountains, and the sub-Moho anomaly in the Rio Grande rift.

Lee and Grand also show a pronounced low-velocity anomaly beneath the central Colorado Rockies extending from 100 to 300 km. Because the horizontal outline of this anomaly is remarkably similar to our crustal anomaly and they did not allow for any crustal velocity anomalies, we are concerned that some of the shallow structure may have “leaked” into the deeper parts of their model. In an independent, P-wave tomog-
Plate 2. Variations of crustal thickness and shear-wave velocity anomalies in 7 layers from the surface to 140 km depth. The velocity anomalies are calculated relative to the 1-D reference model (CCR) in Figure 7.
Plate 3. Profiles of shear-wave velocity anomalies (similar to Figure 2D-F in [Li et al., 2002]). (A) Profile along 38°N. (B) Profile along 39.2°N. (C) Profile along 106.5°W. Thick black line denotes crustal thickness from Plate 2A.
raphy study that did allow for the possibility of shallow structure, however, Dueker et al. [2001] report the existence of a strong (~1.5%), slow, P-wave velocity anomaly extending from the Moho to about 250 km. They call this feature, which coincides in location roughly with the shallow structure (Plate 2B and C), the Aspen anomaly. We see hints of the Aspen anomaly in our deepest slices (Plate 2H), but our images suggest that it does not connect to the crustal anomaly. Body wave tomography with the existing station distribution provides little vertical resolution within the upper 100 km. A tomographic study with denser station distribution in central Colorado is needed to resolve this question.

The transition from a slow tectonic mantle in the southern Rockies to a fast cratonic lithosphere in the Great Plains is imaged near the Rocky Mountain Front, consistent with large-scale tomography studies [Grand, 1994; van der Lee and Nolet, 1997] and regional body-wave tomography [Lee and Grand, 1996; Lerner-Lam et al., 1998]. This transition represents the western edge of the North American craton. Although this study cannot image a sharp lateral boundary, it is clear that in the mantle the transition has significant breadth and extends well into the High Plains east of the Rocky Mountain Front. At depths of 80 km and more in southeastern Colorado, east of the Rio Grande rift, the cratonic lithosphere has been eroded or is absent, suggesting that the old, cold, and presumably more rigid tectosphere can be modified by small-scale convection or rifting events beyond the surface expression of extension. Similar modification of cratonic lithosphere has also been observed in Brazil, and the northeastern U.S./Canadian shield [VanDecar et al., 1995; van der Lee and Nolet, 1997; Rondeauy et al., 2000]. Dueker et al. [2001] suggest that the initial compositional lithosphere may be intact and that localized low velocity anomalies may be created by partial melting of hydrated, olivine-poor lithologies embedded in the lithosphere during earlier, Proterozoic suturing events.

In the upper mantle at depths of 50 to 80 km, the average shear-wave velocity contrasts is 4.5 to 5% from the southern Rockies to the Great Plains (Figure 7 and Plate 2). The range decreases with increasing depth, but may be somewhat underestimated because smoothing and damping tend to minimize the variation. Body wave tomography shows a ~9% S-wave velocity contrast across the region [Lee and Grand, 1996], but this contrast includes a ~3% variation within Kansas, an area which is not well-resolved by surface waves, and, as discussed above, the contrast may be exaggerated by leaking of crustal anomalies into the mantle part of the model. According to Nataf and Ricard [1996], a 100 °C increase in temperature decreases the shear velocity in the shallow mantle by about 1.1%. Our result thus suggests a minimum of 400°C variation in temperature from the Colorado Rocky Mountains to the Great Plains, which is compatible to the temperature contrast in this area obtained by Goes and van der Lee [2002], although Karato [1993] suggests that thermal coefficients of velocity may be larger if Q is low. The required temperature contrast is reduced if partial melt is present. Faul et al. [1994], for example, estimated that 1% melt distributed in inclusions with a range of aspect ratios would decrease shear-wave velocity by about 3.3%. The large, lateral velocity contrasts led Lee and Grand [1996] and Lerner-Lam et al. [1998] to hypothesize that upwelling and mantle melting is occurring beneath the Rockies, perhaps associated with the northern continuation of rifting and extension from the Rio Grande rift into the Colorado Rockies. Although the absence of a high-velocity lid is consistent with such a scenario, the minimum, absolute, shear velocity for the average southern Rockies is about 4.33 km/s (Figure 7), much faster than the ~4.0 km/s found beneath the East Pacific Rise at comparable depths [Nishimura and Forsyth, 1989]. Rather than melt being widely distributed beneath the Rockies, it is more likely that it exists in only a few localities, like beneath the Rio Grande rift or possibly in the Aspen anomaly.

AZIMUHTAL ANISOTROPY

Shear wave splitting measurements [Savage et al., 1996] demonstrate that anisotropy exists beneath at least some parts of our study area in a form that should cause azimuthal variations in Rayleigh wave phase velocities. Although our data have good azimuthal coverage, they are not sufficient for resolving continuous 2-D variations in azimuthal anisotropy. Consequently, we required anisotropy to be constant in each of three tectonic regions, the eastern Colorado Plateau, Colorado Rocky Mountains, and western Great Plains (Figure 4). We inverted simultaneously for the A1 and A2 terms needed to describe azimuthal anisotropy along with the A0 terms at each node and the two-plane-wave parameters describing the incoming wavefield from each earthquake. Our first inversions showed that the Colorado Plateau terms were very poorly constrained, so in the inversions presented here, those terms are fixed at zero. Including anisotropy in the inversions makes only minor improvements in the least-squares fit to the data, but adding the four anisotropic parameters (A1 and A2 for two regions) increases the rank by 3 to 4, depending on period, indicating that the anisotropic terms are well resolved.

Observations

Anisotropy varies laterally between the tectonic regions and is also frequency dependent (Plate 1H and Figure 8). Anisotropy under the Rockies shows a simple pattern. It increases with period from near zero at 40 s to over 7% at 100 s with a nearly constant NE-SW fast direction. At periods
less than 40 s, the strength is less than 2% and the fast direction progressively rotates to nearly N-S as the period decreases and the waves become primarily sensitive to crustal structure. The strikes of young normal faults in the Rockies [Eaton, 1987] are also nearly N-S, suggesting that this apparent crustal anisotropy may be caused by the alignment of cracks. In the Great Plains, anisotropy is generally weaker at most periods than in the Rockies, the 95% confidence limits at several periods include zero anisotropy, and the directions are inconsistent from one period to the next. We do not consider the existence of azimuthal anisotropy in this region to be resolved convincingly; the average effect is likely to be less than 1%. In both regions, the uncertainty increases with increasing period as the velocities decrease in accuracy, the noise increases, and the azimuthal coverage worsens.

**Trade-offs**

There is always a possible trade-off between velocity and anisotropy when they are jointly inverted. Travel-time variations caused by true azimuthal anisotropy can always be perfectly mimicked by allowing in the model sufficiently strong lateral variations in isotropic velocity on sufficiently short distance scales. Generally, however, the resulting models are highly heterogeneous with large velocity variations over distances comparable to the station separation. With the length scale of velocity variations allowed by the smoothing in this study, there is no significant tradeoff between lateral velocity variations and regional anisotropy. This is illustrated in Plate 1D and 1H at a period of 50 s. Adding the anisotropic terms causes only a slight diminution of the amplitude of the low velocity anomalies in the Rockies. Even though the ampli-

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**Figure 8.** Variations of azimuthal anisotropy with period beneath the Rocky Mountain region (A and C) and the Great Plains (B and D). Black circles indicate the strength of anisotropy in percent. Vertical bars represent one standard error in strength. Fast directions of azimuthal anisotropy are indicated by the orientations of black bars as if in a map view with north up on the diagram. Gray bars show one standard error of the fast directions. For a given frequency, anisotropy in A and B is solved simultaneously in the inversion. For C or D, the anisotropy in that region is allowed to vary, but in the rest of the study area, it is fixed to zero.
the anisotropy beneath the Rocky Mountains is dominated by a source at depths of 100 km or greater, probably a highly anisotropic asthenosphere strongly sheared by the North America plate. Beneath the Plains, asthenospheric shearing may occur at greater depths, beyond the range of detection of Rayleigh waves in this study.

Our results of strong and consistent anisotropy at longer periods appear to contradict the observations of shear-wave splitting at stations in the northern Colorado Rockies [Savage et al., 1996], which are characterized by variable direction and many null measurements. The strength of apparent azimuthal anisotropy we observe should produce strong and consistent shear wave splitting. Although the SKS and SKKS phases employed in the splitting studies sample velocity structure from the core-mantle boundary to the earth’s surface, analytic studies [Saltzer et al., 2000] suggest that shear-wave splitting measurements tend to be dominated by the structure at shallow depth. If this is true, then the null and variable measurements of shear-wave splitting in the northern Colorado Rockies could be consistent with rapid lateral and vertical changes in anisotropy in the upper lithosphere, although no consistent evidence of a two-layer structure has been reported for the splitting.

The large apparent anisotropy at longer periods is not yet well understood. Maximum anisotropy is ~3% at 83 s and 100 s assuming that the upper mantle is 70% olivine and 30% orthopyroxene and their fast axes are perfectly aligned horizontally, and only ~1% anisotropy distributed throughout the upper mantle (0–400 km) is needed to obtain about 1 s observed delay times of shear-wave splitting. There are several possible solutions to this problem. First, with large uncertainties at 83 and 100 s, it could just be a fortuitous accident that the estimated values continue the trend established at intermediate periods. The maximum could be no more than 3% without violating any of the Rayleigh wave observations at the 95% confidence level. However, even at 3%, predicted shear-wave splitting would still exceed observed delay times, unless mitigated by variable anisotropy at shallower depths not resolved by the Rayleigh waves. Second, problems with resolving lateral heterogeneity at finite wavelengths may exist. Except for our weighting function, we implicitly assume ray theory in our interpretation, which is strictly valid only when the wavelength is much smaller than the heterogeneity. At long periods, the wavelength, ~400 km at 100 s, exceeds the width of the Rocky Mountain region and the actual sensitivity of the splitting will be spread over a broad region around the ray path [Marquering et al., 1998, 1999] much like a Fresnel zone [Gudmundsson, 1996], so that banded lateral heterogeneity with a scale less than a wavelength may be more effectively represented as azimuthal anisotropy. But if the large anisotropy comes from a low velocity anomaly beneath the Rocky Mountain region, the average fast direction is expected to be per-

Discussion

The small and variable anisotropy in the Great Plains could indicate that the crust and uppermost mantle are nearly azimuthally isotropic, but our estimates might be biased by the assumption of uniform anisotropy in each tectonic province that artificially averages out lateral variations of anisotropy within the province. Significantly different anisotropy within the high plains is indeed revealed in shear-wave splitting analyses [Savage et al., 1996; Savage and Sheehan, 2000], changing from a NE fast direction in northeastern Colorado to a SSE direction in southeastern Colorado. Therefore, the apparent abrupt change in direction with frequency that we observe could be due to a changing population of paths sampling the two areas differently at different frequencies. Unfortunately, there is simply not enough resolution to productively break down the study area into smaller subregions.

The most interesting observation is the simple pattern of azimuthal anisotropy in the Rocky Mountain region (Figure 8A) at periods longer than 40 s. A remarkably consistent fast direction is oriented NE-SW. The observed fast directions on average are rotated 10 to 20 degrees counterclockwise from the direction of the absolute plate motion of the North American plate [Gripp and Gordon, 1990] and agree well with the fast direction of shear wave splitting found in the northern Rio Grande rift area of northern New Mexico and southern Colorado [Sandvol et al., 1992; Savage and Sheehan, 2000]. Although the absolute values are less believable at the longest periods due to relatively large uncertainties, the trend of anisotropy increasing with period is significant. Therefore, the pattern suggests that anisotropy beneath the Rocky Mountains is not due to the trade-off with true velocity anomalies, as can be seen by the fact that the anisotropy is in general stronger at longer periods even as the velocity anomalies become weaker.

A trade-off of anisotropy between different tectonic provinces is also possible. We were concerned because the anisotropy is unexpectedly large in both the Rockies and Great Plains at long periods (Figure 8A and B), but their fast directions are nearly orthogonal to each other. Could this be due to the trade-off of anisotropy between the two regions? To answer this question, we solved for anisotropy in the Rockies and the Great Plains independently. In these inversions, we allowed uniform anisotropy in just one tectonic province and kept other areas isotropic. The complete results from this experiment are displayed in Figures 8C and D. The trade-off does reduce the anisotropy slightly, by less than 1%, but the patterns of anisotropy in both the Rockies and the Plains are unchanged.
pendicular to the band in the E-W direction, not NE as observed. Third, the anisotropy could be due to dipping fast axes that yield greater variation in shear velocity if the propagation is not in symmetry planes [Babuska and Cara, 1991] and affect Rayleigh waves and shear waves differently. For instance, if the upper mantle is pure olivine, the strongest azimuthal anisotropy at 83 s is 5.6% when the fast axis is 35° dipping from horizontal. Dipping fast axes should be observable from the variation of splitting with azimuth if SKS phases propagate slightly off a vertical ray path. Further synthetic calculations are needed in order to better understand the large amount of anisotropy from the Rayleigh waves and to compare this result with shear-wave splitting observations.

CONCLUSIONS

Using Rayleigh waves propagating across the Rocky Mountain Front seismic array, we observed distinct structures beneath the western Great Plains, southern Rocky Mountains and the eastern Colorado Plateau. There is no high velocity lid beneath the Rockies and anomalously low velocities continue to depths of 150 km and greater. Beneath the Plateau, there is a high velocity lid and pronounced low velocity zone and beneath the western Plains, velocities are high and the low-velocity zone is almost absent. The largest velocity contrast across the region is in the crust, with the lowest velocities beneath the elevated region around the Sawatch Range, suggesting that, in addition to crustal thickening, density variations within the crust play an important role in compensating the high topography of the Rocky Mountains. In the shallow upper mantle, the strongest low-velocity anomaly is imaged in southern Colorado, near the Rio Grande rift, indicating the presence of partial melt. The transition from cratonic mantle in the eastern part of the study area to tectonic mantle beneath the Rockies is gradual and not confined to the vicinity of the Rocky Mountain Front.

We constrained azimuthal anisotropy in a joint inversion that included spatial variations in the azimuthally averaged isotropic phase velocities. With the scale length of spatial variations allowed in this study, there is little trade-off between azimuthal and spatial variations in velocity. Azimuthal anisotropy beneath the Rocky Mountains increases with period and its fast direction at depth is close to the absolute plate motion of the North American plate. This pattern is consistent with deep asthenospheric flow dominated by the current plate motion.

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