Deep structure beneath the Southern Rocky Mountains from the Rocky Mountain Front Broadband Seismic Experiment

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ABSTRACT

The deployment of a two-dimensional array of broadband seismometers during the PASSCAL Rocky Mountain Front experiment produced a teleseismic and regional event data set which provides constraints on both the laterally averaged and laterally varying structure of the crust and upper mantle beneath the Southern Rocky Mountains and Great Plains. Results from a spectrum of seismological imaging and inversion techniques indicate that the western edge of the North American tectosphere has been relocated from its position beneath the Paleozoic passive margin to a position several hundred kilometers east of the Colorado Front Range. The mantle and crustal structure presently beneath the Colorado Rockies suggest a laterally variable dynamic state, which, on average, must provide partial support for the high topography.

KEY WORDS: Seismology, structure of lithosphere and crust, Rocky Mountains.

INTRODUCTION

Large-scale seismic tomography images of the North American plate show that the west to east tectonic transition between the Colorado Rockies and western Great Plains is echoed by an equally dramatic transition in the velocity structure of the upper mantle between the western United States and the continental interior (e.g., Grand, 1994; van der Lee and Nolet, 1997). While the crustal deformation within the Rockies is ascribed to the integrated effect of a number of contractional events ending with the Laramide orogeny, the relationship of this deformation to the mantle transition, as well as the involvement of the mantle in present-day regional uplift and on-going extension are not well understood. Moreover, since the mantle transition marks the western margin of a thick, cold mantle lithosphere keel beneath the North American craton, the transition itself may provide insight into the evolution and stability of continental interiors.

The crustal deformation of the western United States has been dramatic, long-lasting, and complicated (Karlstrom and Humphreys, this issue). Throughout most of the Mesozoic, a volcanic arc was active along the present Sierra Nevada, and Antler, Sonoma, and Sevier fold-and-thrust belts developed to the east. In the late Cretaceous, volcanism ceased in the Sierra Nevada, and the locus of thrusting and magmatism shifted to the east of what is now the Colorado Plateau. In the Rocky Mountains, the style of contraction was thick-rather than thin-skinned and is termed the Laramide orogeny (~75-50 Ma). As the Laramide orogeny waned in the early Cenozoic, volcanic activity migrated west, north, and south from the Rocky Mountains. Tec-
tonic relief acquired during the Laramide orogeny in the Rocky Mountains was subsequently beveled down to a relatively gentle surface (Epis and Chapin, 1975; Dickinson et al., 1988; Pazzaglia and Kelley, this issue). Regional uplift of 1-2 km in late Cenozoic time has been suggested (Epis and Chapin, 1975; Eaton, 1986, 1987), but the paleontological corroboration for this inference (MacGinitie, 1953; Axelrod and Bailey, 1976) has been challenged and the uplift might be late Eocene (37-40 Ma; Gregory and Chase, 1992, 1994).

Models for the evolution of the lithospheric mantle that are compatible with this deformation are only beginning to be formulated. While the fold-and-thrust belts in and slightly east of the modern Basin and Range province seem compatible with "normal" contractile plate margins, the thick-skinned Laramide orogeny, far from a plate boundary, has been difficult to understand. One popular hypothesis is that the geometry of the subducting slab changed in late Mesozoic time, becoming subhorizontal and extending to the eastern edge of the modern Rockies (e.g., Lipman et al., 1971; Bird, 1984). This in turn has led to the inference that there has been wholesale transport of crust and upper mantle from west to east (Bird, 1984, 1988). In this particular model, the elevation of the High Plains and Rocky Mountains is due to an increase in the crustal thickness by lower-crustal flow during the Laramide orogeny. Others have compared the shape of the overall uplift to that of a mid-ocean ridge and inferred that the elevations are thermally supported from the mantle (e.g., Eaton, 1987), possibly reflecting the sinking of the subducted Farallon plate in the Cenozoic (Bird, 1984). An entirely different hypothesis is that there was not a horizontal slab beneath North America (Molnar and Lyon-Caen, 1988), or at least not one transmitting forces into the lithosphere (Livaccari, 1991), and that the Laramide orogeny reflects transmission of stress from the plate margin eastward through the lithosphere itself. Recent models offer a version of this hypothesis and suggest that Laramide deformation reflects collision at the west coast of north-translating arc terranes (Maxson and Tikoff, 1996).

A variety of geophysical data suggests that the easternmost extent of western United States crustal deformation may correlate with a lateral transition in upper mantle properties. The recent uplift of the North American mid-continent is centered on the Rocky Mountain Front, which separates the uplifted yet intact western Great Plains from the uplifted and broken crust of the western United States. This transition suggests a major strength contrast or contrast in effective flexural thickness somewhere beneath the western Great Plains (Bechtel et al. 1990). Evidence for a seismic velocity contrast across this transition was obtained by Cleary and Hales (1966), who found that teleseismic P-waves recorded on the craton arrived approximately 1 s earlier relative to the surrounding crust. This study and the subsequent station-residual studies of Wickens and Buchbinder (1980) (S-waves) and Dziewonski and Anderson (1983) (P-waves) gave additional support to the existence of the transition but lacked sufficient station density to resolve its sharpness. Grand (1987, 1994) used a tomographic method to invert several thousand hand-picked SH-wave residuals and found a geographic pattern of upper-mantle shear velocities that correlates with the patterns of station residuals and flexural thickness. This study, though groundbreaking in the quality and amount of data inverted, still suffers from the inherent sparseness of the existing global network of seismographs. Recent work by van der Lee and Nolet (1997) using surface-wave phase velocities, however, provides additional confirmation of the transition. Each of these studies places the largest lateral velocity gradients in the vicinity of the Rocky Mountain Front (RMF). If these gradients are taken as proxies for the mechanical properties, thermal structure, and chemical composition of the lithosphere and upper mantle, the implication is that a major regional boundary is present in the mantle.

A second, though no less important, issue is the correlation between localized or sub-regional variations in crustal deformation and the mantle structure below. Though regionally broad scale, the uplift and extension centered on the Rockies is locally heterogeneous, with variations associated with the northern Rio Grande rift and other features. From a crustal dynamics standpoint, it is important to understand whether these variations are a consequence of heterogeneous reactivation of variable preexisting crustal weaknesses, or are related to a complex pattern of subcrustal mantle flow during Laramide and/or Tertiary tectonism with relic structures observable at present.

Seismological imaging provides an effective means for addressing these issues, but previous imaging efforts have been limited by datasets with insufficient ray-path density. Only a handful of stations from the old analog WWSSN (World Wide Standard Seismograph Network) were located close to the RMF, and only one, station GOL near Golden, Colorado, was actually on it. There were no stations in eastern Colorado, western Kansas, or Nebraska. The number of stations in the replacement digital
networks is even smaller. There were several active-source refraction experiments done across the Front Range (e.g., as summarized by Prodehl and Lipman, 1989 and references therein), and these are useful as constraints on crustal velocities and thicknesses.

It was in this context that the Rocky Mountain Front Experiment was designed. The Program for Array Seismological Studies of the Continental Lithosphere (PASSCAL) of the Incorporated Research Institutions of Seismology (IRIS), funded by the National Science Foundation, provides individual investigators use of a new class of portable, autonomous, digital seismographs with sensitive broadband seismometers. This facility is the ideal tool for regionally dense imaging of the crust and upper mantle, and is capable of recording both teleseisms and regional earthquakes with very high fidelity. Relatively large numbers of instruments can be deployed in an array concentrated geographically over the structures of interest, and the array dimensions, or aperture, and the interstation spacing can be designed according to a priori imaging density and resolution criteria.

THE PASSCAL ROCKY MOUNTAIN FRONT EXPERIMENT

The Rocky Mountain Front Experiment was conducted in two phases. Ten seismographs were deployed in a preliminary reconnaissance phase for 2.5 months during the summer of 1991. These were oriented on an east-west line extending from eastern Utah to eastern Colorado at approximately 39° 20' N latitude with a nominal inter-station spacing of approximately 100 km. Subsequently, thirty-six seismographs were deployed for nearly seven months in 1992 for the main data acquisition phase (Fig. 1). The instruments were arranged in a two-dimensional pattern extending east-west from eastern Utah into western Kansas and north-south between the state borders of Colorado, and included reoccupation of most of the reconnaissance sites.

Three-component sensors capable of recording teleseisms across a broad frequency band were installed at all sites, and the PASSCAL data loggers were programmed to record both continuous broadband and triggered high-sample-rate data streams. The latter programming is useful for high-fidelity recording of regional earthquakes, which provides average constraints on the structure of the uppermost mantle and crust. During the second phase of the experiment, the array was deployed in a two-dimensional rectangular pattern and 446 earthquakes with adequate body-wave phases were recorded. Approximately 100 of these are classified as regional events (located within 1000 km of the array), and the rest are teleseisms.

There are several ways to process these data for information on crust and upper-mantle structure. Teleseismic body waves travel through the upper mantle and crust beneath the receiver at steep angles of incidence, and their traveltime anomalies are ascribable to the integrated velocity structure along the ray path. Tomographic inversion of these observations for velocity structure yields good lateral resolution of vertically integrated structure, but vertical resolution is critically dependent on the azimuthal distribution of events and their ray-path crossing pattern beneath the array. Even if the ray paths are well distributed in azimuth, the depth above which the velocity structure can be resolved is limited by simple geometry and the nominal interstation spacing within the array.

Waveform analysis of the body waves is one way of overcoming this limitation, at least in part. The body of techniques known as “receiver-function” or “boundary-interaction-phase” analysis relies on using portions of the three-component record as a reference signal against which the wave interactions with near-receiver horizontal or slightly dipping structure can be measured. Forward or inverse modeling of these body wave interactions can be used to estimate the depth to major velocity discontinuities and the interval velocities. These techniques are particularly useful for estimating crustal thickness and vertically averaged seismic velocities in the crust and upper mantle. The available lateral resolution depends once again on the nominal station spacing within the array, while the vertical resolution largely depends on the bandwidth over which the signal is recorded and modeled successfully.

Neither of these techniques is particularly accurate in retrieving the velocity of the uppermost mantle just beneath the Moho, a critical region for discussing the role of the mantle in crustal deformation. The appropriate class of techniques for examining this region involves the analysis of seismic waves propagating horizontally through it, such as crustal phases from regional events and intermediate-period surface waves from teleseisms. The analysis of these waves provides constraints on both the absolute value of the seismic velocities as well as the magnitude of velocity gradients, but the inferences are limited by the effect of lateral heterogeneities along the propagation path that cause unmodeled lateral refraction. Thus these methods are useful principally for constraining the regionally averaged velocity structure.
In this paper, we summarize the results of previously published analyses of the Rocky Mountain data set. We will show how the combined analysis provides constraints on the average regional structure and the smaller-scale variability superimposed on it. Taken together, these studies provide the most complete picture of crust and mantle velocity structure beneath the Southern Rocky Mountains.

**Laterally averaged structure**

One of the goals of this work is to establish where and over what scale average mantle in the stable continental keel gives way to average mantle beneath the tectonized western United States. To do this, we first must show that the laterally averaged structure beneath the east and west portions of the RMF array can be characterized as "cratonic" or "tectonic" in comparison with other regional models. Second, from the point of view of inversion theory, we should find the location of the transition that maximizes the difference between these averaged structures. The station location density of the RMF array is not well suited to a formal inversion for this location, and so we proceed with a forward-modelling analysis that seeks to demonstrate that dispersion and traveltime measurements fall naturally into "cratonic" and "tectonic" populations. An additional benefit of these regional models is that they provide the background against which lateral variation can be assessed. In simple terms, it is helpful to know whether observed lateral variation is relative to an average tectonic mantle or an average cratonic mantle in addressing dynamic issues.

**Localization of the transition**

The dispersion of fundamental-mode Rayleigh waves is a sensitive indicator of changes in average upper mantle and crustal properties. Chen and
Lerner-Lam (1993) used a cross-correlation technique to measure the change of phase of Rayleigh fundamental modes from station to station within the RMF array, and obtained a map of interstation phase velocities for periods greater than 10 s. Although there were significant lateral variations and there is evidence for lateral refraction, Chen and Lerner-Lam (1993) found a significant increase in fundamental-mode phase velocities well to the east of the Rocky Mountain Front. The dispersion measurements were grouped in two populations east and west of the observed phase-velocity increase and were averaged above 25 s periods to obtain regionalized dispersion curves. These curves were then inverted for regionalized one-dimensional shear-velocity structures. The resulting models are shown in Figure 2.

It is possible to test the resolution of small differences in velocity structure below a fixed depth by inverting jointly the differences between regionalized dispersion for differences in velocity structure. The velocities are allowed to vary independently above a fixed depth, but are required to have the same variation below. As a series of inversion experiments are performed while varying the fixed depth, and the change in variance reduction is monitored. The result is a bound on the minimum depth to which differences in regionalized structure must persist. The bound is sensitive, however, to assumptions about the homogeneity of the paths contributing to the regionalization. This "squeezing" technique was developed by Lerner-Lam and Jordan (1987) to test for differences between continental and oceanic upper mantle.

Chen and Lerner-Lam (1993) applied this technique and found that the differences between the east and west regionalizations could be confined to depths above 180 km. The differences approach 9% in the seismological lid (rheological lithosphere) and decrease nearly monotonically with depth. A quantitative examination of the interstation phase velocities placed this transition 200-300 km east of the Rocky Mountain Front.

The question of whether this transition is a formal lateral discontinuity, in the sense of a "step" in the velocity model, or a gradual transition cannot be addressed by this intermediate-period surface-wave analysis. However, the grouping of interstation phase velocities into two populations suggests that the transition must occur over only one or two interstation lengths, that is, 100-200 km. Further examination of refracted surface waves, or a search for scattered phases, could resolve this issue.

Figure 2. Chen and Lerner-Lam's (1993) preferred "pure-path" models for the Southern Rocky Mountains (dotted line) and eastern Colorado-western Kansas (solid line).

Laterally averaged structure of crust and uppermost mantle

Higher-frequency surface waves and crustal phases can be analyzed to constrain the laterally averaged structure of the crust and uppermost mantle. Guo (1995) analyzed the Pnl wavetrain and short-period Rayleigh waves produced by the St. George, Utah earthquake (September 2, 1992; mb = 5.7; Pechmann et al., 1994; Pearthree and Wallace, 1992) and recorded on 15 RMF array stations stretching from western Colorado to western Kansas. Pnl, described by Helmberger and Engen (1980), is a combination of the mantle head wave Pn and the crustal reverberation PL and is sensitive to laterally-averaged crust and upper mantle structure (e.g., Wallace, 1983; Shaw and Orcutt, 1984; Clouser and Langston, 1990; Holt and Wallace, 1990). Figure 3 shows a typical Pnl waveform from the St. George event recorded by the RMF array (station BTO).

The Pn portion of the waveform arrives first and represents high-frequency seismic energy propagating in the uppermost mantle or the mantle-lid wave guide. It has been modeled successfully both as a "whispering gallery" phase (Menke and Richards,
1980) and a mantle-lid refraction (Sereno and Orcutt, 1985). PL is the long-period wavetrain following Pn (Fig. 3), characterized by normal dispersion, rapid amplitude decay with range, and prograde elliptical particle motion in the radial plane. Numerous observational and synthetic modeling studies (e.g., Helmberger and Engen, 1980; Wallace, 1983; Shaw and Orcutt, 1984; Holt and Wallace, 1990; Clouser and Langston, 1990) have demonstrated that PL can be successfully modeled as the summation of multiply reflected and converted P and SV energy trapped within the crust with associated evanescent shear energy loss to the mantle. Thus, the Pn portion of Pnl provides constraints on uppermost mantle structure similar to the constraints provided by post-critical body waves in refraction profiles, whereas the PL portion provides constraints on crustal structure similar to a multimode surface wave. For example, several studies (Hill, 1971; Clouser and Langston, 1990; Holt and Wallace, 1990) show that the gradient in the upper mantle can profoundly affect the Pn waveform, but has little or no effect on the PL portion. A positive mantle gradient enhances Pn amplitudes, and vice versa, with respect to a constant-velocity lid. This single parameterization of Pnl has been modeled, for example by Langston (1982), but the ratio of Pn to PL amplitudes has been shown to be a more robust measure of the mantle gradient (Holt and Wallace 1990; Clouser and Langston, 1990, Guo, 1995). The complexity of the waveform arises from the constructive and destructive interference of multiple reflections, refractions, and mode conversions, and its constraints on structure are best inferred through forward modeling with synthetic seismograms.

Synthetics were calculated for one-dimensional models using the vectorized wavenumber integration (reflectivity) scheme of Fuchs and Muller (1971) and Muller (1985), which accounts automatically for all P-SV interactions. The synthetic seismograms were calculated in a slowness window of 0.04-0.35 sec/km and a frequency window of 0.01-2.0 Hz. Guo (1995) assigned a representative attenuation of Qp = 1000 and 300, and Qs = 500 and 200.

**Figure 3.** Comparison between observed Pnl and Sn at RMF site BTO (bottom trace) and synthetic seismograms computed for different models (see Guo [1995] for complete explanation). Arrows point to features of the waveform for which sensitivity tests were performed, and which are critical to the interpretation. Model m5 gives the best fit, and is the model from which RMF740 is derived.
in the crust and mantle respectively. Data and synthetics were band-passed between 0.01 and 0.25 Hz.

Summary of Preferred Model results

Guo's (1995) preferred model, RMF740, comprises a two-layer crust overlying a thin-lid/ thin-LVZ upper mantle with a lid gradient of 0.0015 sec¹. This model represents an average structure integrated along the path between the St. George earthquake and the RMF array. While the lateral variability of these structures cannot be determined with this analysis, model RMF740 represents an adequate background against which smaller scale variations can be assessed.

Figure 4 shows RMF740 plotted against other significant one-dimensional structures obtained for the western US and some cratonic regions. In Figure 4A, RMF740 is compared with two other P-wave models obtained for the western United States: model GCA (Walck, 1984) derived for the Gulf of California spreading center and the more general Basin and Range model T7 (Burdick and Helberger, 1978) derived from long-period-body-wave forward modeling. Figure 4B shows model RMF740 plotted against the continental models S25 of LeFevre and Helberger (1989), representing the Canadian Shield, K8 of Given and Helberger (1980), representing northwest Eurasia, particularly the Baltic Shield, and WCH of Beckers et al. (1994), representing the mixed structures of western China. Except for the case of GCA, which is derived from mostly vertical ray paths, these models were derived using seismological imaging techniques that, like Guo (1995), average the lateral heterogeneity along pseudo-horizontal propagation paths. WCH, K8 and S25 typify the range of crust and upper mantle structures observed for stable continental interiors, while T7 and GCA are representative of the upper mantle structure seen beneath spreading crust.

A comparison of these models shows that the upper mantle beneath the Southern Rocky Mountains is substantially different than the mantle beneath continental shields. Continental shield models are characterized by very high Pn velocities (compressional velocity just beneath the Moho), and, if present at all, relatively deep low-velocity zones with small velocity decreases. RMF740, on the other hand, is much closer to the models of "tectonized" upper mantle (T7 and GCA), with similarly low values of the Pn velocity, thin lid, and shallow low velocity zone.

There are distinct differences among the models, however, which suggest that the average crust and mantle beneath the Southern Rockies is distinct from the average structure beneath the Basin and Range. RMF740 has a faster Pn velocity (8.00 km/s vs. 7.95 km/s) and a much smaller lid velocity gradient (0.15 vs. 0.5 km/s per 100 km) than T7. Moreover, the velocity difference between the lid and the LVZ is smaller in RMF740 than T7, although the difference in LVZ thickness is not well resolved by the Pn forward modeling procedure with these data. Finally, the crust of RMF740 is thicker and has higher velocity than the crust of the Basin and Range.

Laterally varying structure

Summary of Receiver Function Analysis of Moho converted phases

The coda of first-arriving, teleseismic P body waves propagating upward through the upper mantle and crust beneath the receiver contains mode conversions between compressional and shear energy caused by interactions with velocity discontinuities. If these discontinuities are subhorizontal, the mode conversions are large on the radial component of the seismogram while the vertical component remains relatively undisturbed. Thus the vertical component provides a reference for source mechanism, deep-mantle propagation effects, and instrument filtering, against which the perturbations to the radial component caused by boundary interactions can be measured. "Receiver function analysis" treats the radial component as the linear convolution of the vertical reference signal with a time series, or "receiver function", representing the near-receiver mode conversions. Typically, the most significant mode conversions are the conversion of P to S at the Moho, and the subsequent first reverberation. Thus receiver function analysis is used widely to estimate crust thickness and structure (e.g., Phinney, 1964; Langston, 1977, 1989; and Owens et al. 1984, 1987).

With data from a two-dimensional deployment such as the RMF experiment, receiver function analysis provides a plan view of crustal thickness variations, which can then be interpreted in terms of the major structural trends exhibited in the geology. Sheehan et al. (1995) computed receiver functions at 20 Rocky Mountain Front deployment sites for 12 teleseismic events with mb >5.5. The receiver functions were interpreted in terms of depth to Moho by computing synthetic receiver functions (corrected for sediment cover) for a suite of crustal thicknesses and determining the RMS misfit for each event-receiver pair. The crustal thickness val-
Figure 4. (a) Comparison of RMF740 with models derived for active tectonic provinces: T7: Basin and Range (Burdick and Helmberger, 1978); GCA: Gulf of California spreading center (Walck, 1984). (b) Comparison with models derived for more stable provinces: K8: northwest Eurasia (Given and Helmberger, 1980); S25: Canadian Shield (Lefevre and Helmberger, 1989); WCH: western China (Beckers et al., 1994). Global average IASP91 (Kennett and Engdahl, 1991) shown for comparison.

ues found at the RMS minimum for each event-receiver pair were then averaged over the event ensemble to produce a value for that receiver site. Sheehan et al. (1995) performed a sensitivity analysis to show that for reasonable values of the compressional and shear velocities in the crust, the estimates of crust thickness using this technique are robust and have a standard deviation of about 2 km. The crustal thickness estimates thus obtained are used to map the Moho topography beneath the RMF array.

Figure 5 (Sheehan et al., 1995) shows an example of the observed and synthesized receiver functions computed for a single event, displayed as a pseudo-record section across the RMF array. It is apparent immediately that the mode conversion at the Moho does not vary significantly across the Rocky Mountain Front. To facilitate a geological interpretation, Sheehan et al. (1995) mapped the receiver function results onto four physiographic provinces with the following results for regionally-averaged mean crustal thickness: the Kansas Great Plains, 43.8 ± 0.4 km; the Colorado Great Plains, 49.9 ± 1.2 km; the Colorado Rocky Mountains, 50.1 ± 1.3 km; and the northeast Colorado Plateau, 43.1 ± 0.9 km.

We will discuss the importance of these results in the discussion section, but we point out that there is essentially no variation in estimated crustal thickness between the Colorado Great Plains and Rockies, a result that is consistent with earlier refraction studies (Pakiser, 1963; Jackson et al., 1963; Jackson and Pakiser, 1965; Roller and Jackson, 1966; Prodehl, 1979; Prodehl and Pakiser, 1980).

Shear-wave tomography

Shear-wave tomography results from the Rocky Mountain Front experiment are summarized in Lee and Grand (1996). Traveltimes were obtained for 624 S-wave arrivals from teleseismic events from 30° to 120° epicentral distance. Traveltimes were measured by cross-correlating the S-wave data with synthetic seismograms computed using the WKBJ technique (Chapman, 1978) and using the Harvard CMT solutions for the source mechanisms. The shear-wave residuals are corrected for known crustal, sediment, and topographic variations (Sheehan et al., 1995) and are thus sensitive to mantle shear-velocity variations beneath the Colorado array. Up to 5 s differences in shear-wave traveltimes are observed between the central Rockies in Colorado and station PKS in western Kansas. These traveltime variations are nearly as large as those between the Canadian Shield and the East Pacific Rise (Grand and Helmberger, 1984). The shear-wave residuals are inverted for shear-wave velocity variations from...
50- to 425-km depth using a back-projection technique (Humphreys et al., 1984). The block dimensions used in the inversion are 0.6 by 0.6 degrees in lateral dimension and 50 km thick in the vertical direction. The tomographic models show slow shear-wave velocities beneath the Rocky Mountains relative to the Great Plains, extending from the surface to 200 km depth (Fig. 6). The amplitude of the variation ranges from 9% at 50–100 km depth to 7% at 150–200 km depth. At greater depths, lateral variations are smaller and not well correlated with surface tectonics. The tomographic images reveal three mantle structures beneath the study area. Beneath the Rockies, the mantle is extremely slow. Intermediate velocities are found to the west and east of the Rockies. At the Colorado-Kansas border there is an abrupt increase in mantle shear velocity, with faster velocities to the east.

Compressional-wave tomography

The data for the compressional-wave tomography consist of approximately 2700 hand-picked P wave arrival times from the Rocky Mountain Front experiment. The estimated error on the picks is 0.1 s, and the range in traveltime residuals observed is 5 s. Relative residuals are created by subtracting the average absolute residual for each event. Corrections for sedimentary basins, topography, and crustal thickness variations are made to the traveltime residuals before inversion. The inversion consists of breaking the study area into 50 x 50 x 30 km blocks, tracing rays through this structure, and inverting using the SIRT algorithm (Humphreys and Clayton, 1988).

The results of the compressional-wave inversion reveal similar patterns to those found with the shear-wave inversion (Fig. 7). The most prominent low-velocity volume is found beneath the Rocky Mountains in Central Colorado, with a transition to higher velocities found beneath western Kansas. The peak-to-peak velocity variation in the model at 150 km depth is 4%. The largest amount of heterogeneity is found in the upper 250 km of the upper mantle. The tomographic model only explains about 60% of the signal variance, which may indicate that unmodeled small-scale heterogeneity is present.

Overview of tomographic results

If due solely to thermal effects, a 9% shear-wave velocity variation implies an 800° variation in temperature (Nataf and Ricard, 1996). This is thought to be unreasonably large, as this is nearly twice the thermal variation predicted between oceans and continents, and would also elevate the temperature of the mantle rocks above their solidus and cause massive melting. Given thermal expansion, a temperature variation of 800° would predict a density anomaly more than double the estimate of Sheehan et al. (1995) for the mantle density beneath the Rockies needed to account for the topography and gravity signature of the Rockies. The presence of melt can have a profound effect on seismic velocity, and almost certainly is a contributor beneath the Rockies. Assuming Faul et al.'s (1994) relation between partial melt and shear velocity and density and Nataf and Ricard's (1996) relation between...
shear velocity and temperature, a model with 350 °C difference in temperature between the mantle beneath central Kansas and the Rockies as well as a 1.5% partial melting of the shallow mantle beneath the central Rockies predicts a shear velocity contrast of 8.8% and a density contrast of 1.1% between the two regions, in close agreement with our observations. Chemical heterogeneity may also play a role and could lessen the amount of thermal or melt heterogeneity required.

Using the scaling relations given in Humphreys and Dueker (1994) a 4% P-wave velocity variation implies a 640 °C temperature anomaly. This is significantly less than the 800 °C suggested by the shear-wave tomography model. Some differences in model amplitudes may be due to the parameterization of the models, with larger amplitude velocity variations needed if heterogeneity is confined to a smaller depth interval, or smaller amplitudes if a larger depth interval is used. Another possibility is an anelastic contribution to the velocity variation, which, for an attenuation parameter, Q, of 50, would halve the thermal variation required (Karato, 1993). For a given thermal variation, Q for shear waves (Qs) is smaller than Q for compressional waves (Qp). Therefore, following Karato (1993), the anelastic contribution to the variation in velocity for a given thermal anomaly is expected to be larger for shear waves than for compressional waves. This may help explain the large shear-velocity anomalies relative to compressional-velocity anomalies beneath the Rocky Mountains.
Figure 7. Four horizontal slices through a compressional-velocity tomogram, at depths of 120, 150, 180, and 210 km. Eastward increasing velocities below the Colorado-Kansas border are displaced somewhat east from the shear-velocity results. The correlation with the shear-tomogram suggests laterally varying P/S ratios is the mantle.

Regions of the study area with high heat flow (Morgan and Gosnold, 1989; Blackwell and Steele, 1992) correlate strongly with regions where we observe low velocities. While crustal heat production likely contributes to part of the heat-flow anomaly observed, the heat-flow anomaly is of such significant magnitude (60 mWm^-2) that a mantle contribution ascribable to conduction, convection, or both, is very likely.

Summary of measurements of mantle anisotropy

It is well known that shear waves propagating through a transversely anisotropic but otherwise homogeneous medium perpendicular to its axis of symmetry split into orthogonally polarized components which travel at different speeds. Once these components have passed through the region of anisotropy, the time delay between them is proportional to the magnitude of the anisotropy and thickness of the anisotropic layer, while the azimuth of one of these components is parallel to the symmetry axis of the anisotropy. In the mantle, transverse anisotropy is thought to arise from the lattice-preferred orientation of olivine resulting from finite strains (McKenzie, 1979) or stress-mediated recrystallization. In this circumstance, the axis of the fast polarization lies along the symmetry axis of the anisotropy. As a consequence, measurements of splitting of shear body waves are used as proxies for the orientation of olivine fabrics and hence as an indication of the orientation and strength of mantle deformation and mantle flow. Obviously, only the present-day fabric,
which is a result of the total strain, is observed. Circumstantial arguments and/or finite-strain flow models must be developed to constrain the time history. See Silver (1996) for a recent review.

Shear-wave splitting parameters were determined for RMF array sites by Savage et al. (1996), who examined core (SKS and SKKS) and body phases (S) from 28 events recorded during both the first and second phases of the experiment. The two horizontal components of each shear-arrival complex were digitally combined and subsequently rotated and time-shifted in a two-dimensional parameter grid until the waveform distortion caused by splitting was minimized. The azimuth of the fast polarization, \( \phi \), and the time delay between the fast and slow polarizations, \( \delta t \), were estimated along with formal uncertainties. Savage et al. (1996) also reported null measurements, that is, a deterministic observation of an unsplit arrival.

The interpretation of \( \phi \) and \( \delta t \) in terms of mantle anisotropy requires a choice of parameterization for the medium. An effective parameterization, assumed by Savage et al. (1996), localizes transverse (hexagonal) anisotropy with a horizontal axis of symmetry within a single homogeneous layer. (Other parameterizations are possible; see Silver and Savage, 1994; Levin et al., 1996.) The utility of this parameterization lies in its role as a natural proxy for the horizontal component of mantle flow, as it is thought that the finite strain associated with mantle flow results in the olivine constituent having a preferred orientation (McKenzie, 1979). Since the incoming S, SKS, and SKKS phases are propagating through the upper mantle with near-vertical incidence, their splitting will be optimally sensitive to the horizontal component of the olivine fabric. In this model, \( \phi \) will be parallel to the horizontal symmetry axis, and \( \delta t \) will be proportional to the thickness of the layer and the volume fraction of oriented olivine.

Savage et al. (1996) found splitting parameters varying among the array sites, as well as several null measurements (Fig. 8). The interpretation of null measurements of shear-wave splitting is particularly interesting although somewhat subjective. A null measurement could indicate isotropy, but it is also consistent with transverse isotropy having a vertical symmetry axis that might arise from vertical flow. In recent work, Saltzer et al. (1997) showed that null measurements may also be consistent with a stack of thin layers having randomized fabric, orthorhombic fabric, or scattering heterogeneities.

While any of these scenarios would constitute a useful null hypothesis for interpreting null splitting observations, the results of Savage et al. (1996) suggest the presence of rapidly varying anisotropy beneath the RMF array. If measurable finite strain has thus occurred in the mantle beneath the RMF array, it may be more reasonable to assume that anisotropy is present rather than absent. For this reason, Savage et al. (1996) interpret null measurements to indicate vertical fabric in this region.

The average \( \delta t \) found by Savage et al. (1996) is 0.7 s, which is consistent with an anisotropic path length of 50 to 150 km using observed bounds on the aligned volume fraction of olivine obtained from kimberlite nodules (Mainprice and Silver, 1993). The lower bound is consistent with confinement of the anisotropy to the lithosphere or crust, or consistent with a longer path length through a more heterogeneous anisotropic structure.

**DISCUSSION**

The strength of an experiment configuration such as the RMF array lies in the ability to make hierarchical inferences based on the different spatial scales resolved by various observations and modeling techniques. The RMF array provides data constraining both the regional geophysical context and the intra-regional variation needed for large-scale geological interpretation. It is useful to examine these constraints beginning with the largest spatial scales.

There are two persistent problems in understanding the large-scale tectonics of the western United States: First, what is the role of the mantle in the series of orogenies culminating in the Laramide? Second, what is the source of the regional-scale uplift across most of the region (Eaton, 1987; Gregory and Chase, 1992, 1994). The results of the analyses of the RMF data summarized in this paper support the view that the mantle has participated in past deformation, and continues to contribute to the regional uplift. There are several lines of argument.

The laterally-averaged shear velocities found beneath most of the RMF array are much lower than the shear velocities found beneath cratons. There is a large body of geophysical and geochemical evidence indicating that stable continental cratons are underlain by “keels” of depleted mantle (the “tectosphere”; see Jordan et al. [1989] for a summary discussion) which is stabilized against advective disruption by compositional differences with the surrounding, “normal” mantle. This cratonic mantle is characterized by high average velocities which persist to depths greater than 200 km and often
deeper beneath the Archean cores (e.g., Grand, 1987, Lerner-Lam and Jordan, 1987). The low velocities found on average beneath Colorado are not consistent with cratonic mantle.

The surface-wave and tomographic studies summarized above place the transition between Rocky Mountain mantle and cratonic mantle approximately 200 km east of the Rocky Mountain Front, and suggest that this transition persists to at least a depth of 180 km. A similar transition in mantle structure beneath the eastern United States has been detected some 400-450 km west of the North Atlantic passive margin (Fischer, personal communication, 1997; also Grand, 1994; and van der Lee and Nolet, 1997). Together, these may be interpreted as the western and eastern edges of the North American tectosphere. If a similar scaling operated in the Paleozoic beneath the western United States passive margin, which extended from the Sierra Nevada batholith on the west to the Colorado Plateau on the east, the western edge of the North American tectosphere would have extended to the vicinity of the western edge of the Colorado Plateau. This is supported by the observation that the sedimentary hingeline of the Cordilleran miogeocline was at the western edge of the Colorado Plateau, suggesting that the plate was thinned at least this far east. The RMF experiment results for present-day mantle structure show that to first order, this presumed correlation between the pre-orogenic Paleozoic passive western margin and craton-like upper mantle structure has been lost, and that the tectosphere boundary has migrated eastwards since the Paleozoic. We suggest that the lateral mantle transition observed beneath western Kansas and eastern Colorado represents the “new” western boundary of the North American tectosphere, and that this new boundary has been relocated on the order of several hundred kilometers eastward by the successive actions of the cordilleran orogenies and the subsequent uplift. This boundary thus is a marker for the easternmost extent of mantle modification during Mesozoic-Cenozoic deformation in the western United States.

An outstanding question thus concerns the mechanisms that relocated the western mantle margin of the tectosphere. A complete discussion is be-

Figure 8. Shear wave splitting measurements from the Rocky Mountain Front experiment [Savage et al., 1996]. A positive measurement is a single thick dark line oriented parallel to fast direction of polarization $\phi$, with length corresponding to delay time $\delta t$ in seconds (scale on lower right). Null measurements are plotted as two small crossed lines with directions equal to the two allowed orientations of anisotropy. All measurements are plotted at the 220 km depth projection of the split ray. Small open circles denote seismograph station locations. Thin solid lines define state boundaries, black dashed lines define physiographic provinces.

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beyond the scope of this work, but there are three somewhat related conceptual candidates: (1) advective removal of previously resident continental tectosphere through the action of subhorizontal subduction (Bird, 1984, 1988), (2) advective removal of tectosphere through large-scale vertical transport, lithosphere delamination, or the development of Rayleigh–Taylor instabilities (Houseman and Molnar, 1997), and (3) convective disruption of the tectosphere through the action of accelerated thermal or chemical diffusive mechanisms (Shapiro et al., 1991). In the first and second cases, the location of the mantle boundary 200 km east of the easternmost extent of Laramide crustal deformation suggests a scale length for the distance between the decoupling of the lithosphere from crustal deformation and its interaction with the remaining cratonic lithosphere. In the third case, the thickness of the lateral transition between Rocky Mountain mantle and cratonic mantle should give a sense of the scale of the diffusive processes. Whatever the mechanism, the resulting forces were strong enough to overcome whatever processes stabilized the tectosphere. It will be important to examine this boundary, and other tectosphere margins, in detail. Nevertheless, the implication of these findings is that the regional upper mantle beneath the Rockies has undergone substantial modification since the time of the Paleozoic passive margin, presumably through mechanisms that produced the successive contractional orogenies through the Laramide and lithospheric thinning in the Cenozoic. Whether this location represents a translation of the orogenic boundary east, or a marker for the easternmost extent of tectosphere removal is not directly constrained by RMF observations. However, the anisotropy observed at station CES, in western Kansas, is not consistent with finite compressional strain in the mantle along an east–west azimuth. Thus we do not favor the translational explanation.

Regardless of the details of the mechanism that removed the Paleozoic tectosphere, the average residual mantle and crust structure observed at present suggests continued mantle involvement in western United States dynamics. It is useful to examine whether the average mantle structure argues for any particular model. The surface-wave analysis yields models that are consistent with models of mantle beneath other tectonically active regions, but the Pn1 regional phase analysis summarized earlier provides more detail. The analysis of these phases makes a case for an average structure with a mantle lid with positive-downward velocity gradient, which is not consistent with models of mid-ocean ridges or the Gulf of California (model GCA, Walck, 1984), which predict little or no lid, and nil or negative gradients. On the other hand, at a thickness of 30 km, the RMF mantle lid is less than half the thickness of the mantle lid under stable continent, where it is found (e.g., Lefèvre and Helmberger, 1989) or beneath Phanerozoic platforms. Equivalent lid thicknesses can be found beneath oceanic plates that have cooled conductively for approximately 10–20 m.y. (Leeds, 1975; Schlué and Knopoff, 1977; Nishimura and Forsyth, 1988; Jordan et al., 1989), but neither the conductive time constant nor the implied subsidence is consistent with late Eocene uplift. Seismological lids are also found beneath currently active continental collisional orogenies, such as the southern Himalaya (Molnar and Lyon-Caen, 1988) or in complicated regions such as western China (Beckers et al., 1994). Although in some cases these lids may be orogenically thickened, these structures are in almost all cases much thicker than 30 km. It is difficult to draw robust conclusions from any single path, but we suggest that the Pn1 model, with its 30-km lid and similarities to other regional tectonic models, rules out coherent, shallow, unidirectional spreading and subsidence. At the same time, it is incompatible with excessively thickened lithosphere resulting from collision or contraction. Thus this model argues indirectly for a mantle in a transitional dynamic state rather than steady-state in the long-term. As we discuss below, however, some elements have been quasi-static since the late Eocene.

The evidence that this transitional structure plays an active role in the current dynamics of the Rockies and western Great Plains is less circumspect. The crust thicknesses derived by Sheehan et al. (1995) fall into two populations, with a thicker crust of approximately 50 km underlying the Rockies and Colorado Great Plains, and thinner crust of 43–44 km beneath the northeast Colorado Plateau and the Kansas Great Plains. The main variation in crustal thickness is thus between western Kansas and eastern Colorado, and not between the Colorado Great Plains and the Rocky Mountains. Noting the observation that much of the western United States, including the Rockies, is in isostatic equilibrium, Sheehan et al. (1995) argued that to first order, the crustal thickness results imply that Rocky Mountain topography cannot be compensated by Airy-type crustal roots. An additional, quantitative analysis by Sheehan et al. (1995) incorporating the gravity signature demonstrated this more precisely, concluding that compensation of Rocky Mountain topography can not be accom-
plished solely in the crust, and requires a mantle component. Sheehan et al. (1995) calculated that about half of the excess topography in the Rocky Mountains is supported by the mantle.

There are several possible mechanisms for this excess support, including thermally- and compositionally-driven buoyancy, and vertically oriented dynamic forces. The results of the shear- and compressional-wave tomography show a large amount of small-scale variability with the implication that there may be several mechanisms operating at once. However, both the shear and compressional tomography show a large region of relatively slow velocities centered on the Southern Rockies in the vicinity of the Rio Grande rift near Leadville. The magnitude of the anomaly is too large to be consistent with a compositional origin and is probably of thermal origin. Moreover, Savage et al.'s (1996) measurements of anisotropy are null in this region, but are arrayed systematically around the locus of low velocities and are similar to measurements made elsewhere in the northern Rio Grande rift (Sandvol et al., 1992). This region also correlates with high heat flow, as noted above. Taken together, these observations are consistent with a spatially restricted zone of vertical dynamic support beneath the southern Rocky Mountains near the Rio Grande rift. It is possible that this localized vertical support could be manifested through vertical flow, but this is not demonstrated by our data, and does not preclude other mechanisms operating elsewhere. As noted by Savage et al. (1996), the near isostatic compensation of the Rockies argues for mantle flow driven by shallow deformation rather than the reverse, at least on a regional scale. This issue could be resolved by dense lateral sampling and estimation of the depth variation of anisotropy.

This excess buoyancy most likely has been a quasi-steady state feature of the dynamics since the late Eocene, if the conjecture by Gregory and Chase (1992, 1994) on the timing of the topographic uplift is correct. This timing would also argue against a single thermal pulse of regional scale because, as suggested earlier, the subsequent topographic subsidence through conductive lithospheric cooling is not observed. It is entirely possible, however, that smaller scale features could be the result of successive thermal pulses and conductive cooling, although these may be below the resolving limit of this experiment. Under these circumstances, it is not clear what drives the passive flow, for large amounts of post late-Eocene extension are not observed. The flow may be a residual effect of the disruption of the tectosphere, or a "back-flow" associated with continued Rayleigh-Taylor instabilities of the sort proposed by Houseman and Molnar (1997).

In circumstances such as those operating beneath the Rockies, it is legitimate to ask whether the standard regionalized notions of lithosphere and asthenosphere make sense. The removal of large amounts of continental tectosphere by an unspecified mechanism may leave behind a mantle which does not exhibit steady-state dynamics at the local level. The motion is never coherent enough laterally to cause further deformation (extension or contraction) in the crust. However, the motion in the aggregate provides enough vertical support so that the uplift is regionalized and appears to be the consequence of steady-state flow. This implies that there must be a balance between the net upward intrusion of asthenosphere and the net downward advection of lithosphere. Perhaps better spatial resolution of the uplift history of the southern Rockies will provide constraints on this balance.

CONCLUSIONS

The strength of an array deployment such as the Rocky Mountain Front Experiment is its ability to constrain features with scale lengths comparable to the dimensions of the array as well as the nominal station spacing. When analyzed by a variety of techniques, constraints on both the structure and the dynamics can be developed.

Analysis of body-wave-traveltime and surface-wave-dispersion measurements show that a fast-velocity mantle tectosphere presumably present beneath the western United States during the Paleozoic has been transformed in the Rocky Mountains into a velocity structure having on average a 30-km lid and slow velocities. The transition between this structure and the cratonic upper mantle underlying stable North America is constrained to occur beneath the Colorado-Kansas border and appears to be sharp. The transition represents the western edge of the North American tectosphere and is a marker for the eastern limit of upper mantle deformation associated with the western United States contractional orogenies and subsequent uplift. This eastern limit of mantle deformation appears to be displaced from the eastern limit of Laramide crustal deformation by approximately 200 km.

The most significant feature appears to be a restricted zone of passive upwelling located to the north of the northern Rio Grande rift, in south-central Colorado. This upwelling supplies a mantle partially supporting excess topography.
Rockies, which cannot be solely supported by mechanisms confined to the crust. The magnitude of the velocity anomalies in this upwelling zone require the presence of partial melt. A significant problem is bracketing the melt zone both laterally and vertically.

Smaller scale features are observed in the topographic models and the variation of anisotropy, but it is not apparent whether these features are correlated with smaller scale features in the crust. An unresolved problem is whether these small-scale variations represent dynamics operating in concert to provide the source of the post-late-Eocene uplift.

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DEEP STRUCTURE BENEATH THE SOUTHERN ROCKY MOUNTAINS


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Seismicity of COLORADO and Surrounding Areas
1800 - 1993

Figure 1. Seismicity of Colorado and surrounding areas, 1800-1993 [USGS, 1993]. Cities indicated by stars, earthquakes by circles. Size of symbols corresponds to earthquake magnitude. Concentration of seismicity north of Denver is associated with fluid injection at Rocky Mountain Arsenal [Healy et al., 1968], and clustering of seismicity in northwest Colorado (on Utah border) is largely associated with secondary oil recovery near Rangely, Colorado [Gibbs et al., 1973].