Regional Crustal Structure Derived from the CD-ROM 99 Seismic Refraction/Wide-Angle Reflection Profile: The Lower Crust and Upper Mantle

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The Continental Dynamics - Rocky Mountains project (CD-ROM) seismic refraction data were acquired along a ~950 km long profile extending from Fort Sumner, NM, to the Gas Hills in central Wyoming. This profile crossed many major structural features including the Jemez lineament, the Colorado mineral belt, and the Cheyenne belt. Velocity models derived using several techniques indicate that crustal thickness increases from ~45 under New Mexico to 55 km in central Colorado and thins to 40 to 45 km under southern Wyoming. A mid-crustal interface is very prominent within and can be thought of as the Conrad discontinuity. This interface lies at depths of ~25 to 30 km where velocities increase to about 6.8 km s\(^{-1}\). A high-velocity lowermost crustal layer (V\(_p\) = 7.0 to 7.4 km s\(^{-1}\)) with a thickness ranging from 5 to 10 km is evident in the Southern Rocky Mountains - Great Plains portion of the model. We interpret this layering of the middle and lower crust to be a result of the early conversion of the crust from an island arc composition to an intermediate composition that was enhanced during the extensive 1.4 Ga magmatic modification of the crust. A gravity model was also constructed for the profile. Combined with geologic data, the velocity and gravity models suggest that the main elements of the crustal structure today mainly emerged during the initial assembly of the continent (ca. 1.76 to 1.6 Ga) and during the extensive magmatism at 1.4 Ga. There are strong geophysical signatures of subsequent local modification of the crust, but the region was tectonically stable from 1.4 Ga to the Ancestral Rocky Mountains orogeny. The areas of thickest crust do correlate with Laramide uplifts suggesting that mechanical thickening played a role where the crustal thickness exceeds a value of ~45 km. Based on the gravity and seismic models, we conclude that compensation of the high topography in the southern Rocky Mountains results from a compound interplay of density variations within the lithosphere, most of which has probably developed since late Cretaceous time.
1. INTRODUCTION

The tremendous variety of tectonic activity that has occurred from the Precambrian to present increases the difficulty of understanding the processes at work during Precambrian assembly of the continent in the Rocky Mountain region. As several studies in this volume indicate, unraveling this tectonic history continues to be an ongoing challenge. Our major questions are how the North American continent was accreted during Precambrian time and how the structures formed during accretion influenced Phanerozoic tectonism and magmatism.

As discussed elsewhere in this volume, the geologic evolution of the Rocky Mountain region has included many diverse tectonic events. This evolution began with the Precambrian accretionary events that lead to the assembly of the North American craton (Laurentia). During the Proterozoic (1.4 Ga), extensive felsic magmatism spread across the southwestern portion of the North American craton (Laurentia) [Anderson, 1989]. Towards the end of the Proterozoic (1.0 - 1.1 Ga), the accretion of Laurentia was completed by the Grenville orogeny [Mosher, 1998] and was followed by widespread rifting [Adams and Keller, 1994]. Globally, the Grenville orogeny was part of the final assembly of the supercontinent Rodinia [e.g., Karlstrom et al., 1999]. Rodinia broke up near the end of the Proterozoic and passive continent margins formed along the rifted margins of Laurentia [e.g., Karlstrom et al., 1999]. Early Paleozoic tectonic stability was followed by late Paleozoic deformation (Ancestral Rockies), Late Cretaceous to early Tertiary shortening (Laramide orogeny), and Oligocene to Recent extension along the Rio Grande rift [e.g., Kluth and Cooney, 1981; Sloss, 1988; Bird, 1998; Keller and Baldridge, 1999].

As part of the Continental Dynamics – Rocky Mountain (CD-ROM) project, the University of Texas at El Paso and the University of Karlsruhe, Germany led the acquisition of seismic refraction data along a 950-km-long profile in 1999 (Figure 1). This profile extended from central Wyoming to east-central New Mexico (Figure 1). The emphasis in this project was to study the Precambrian architecture of the continent. Thus, this profile crosses the major northeast-striking Precambrian structural trends at a high angle (Figure 1).

The purpose of this paper is to present the results of the analysis of the data collected along this refraction profile from the standpoint of mid- to lower crustal and upper mantle structure. The crustal velocity model derived also contributes to analysis of the deep seismic reflection and
teleseismic data collected by the CD-ROM group [e.g. Magnani et al., this volume, a and b; Morozova et al., this volume; Dueker et al., 2001]. Our analysis of the refraction data was coordinated with the study of Rumpel et al. [this volume] that focused on the upper crust. In addition, it is a complement to the study of Levander et al. [this volume] who interpreted these data using a different approach.

2. GEOPHYSICAL BACKGROUND

The CD-ROM seismic profiles lie in an area with modest existing seismic constraints on crustal structure [Prodehl et al, this volume; Sheehan et al., this volume]. As discussed by Prodehl et al. [this volume], previous velocity models derived from refraction experiments usually lack detail, because of the widely spaced recording instruments and sources. The limited geographical extent of the previous seismic measurements and their low resolution has resulted in crustal models that show little correlation with exposed geologic features or topography [e.g., Sheehan et al., 1995]. However, these models provide a useful context in which to interpret the results presented here. The recent Deep Probe experiment (Figure 1) was of particular interest in this study, because the Wyoming shotpoint (SP 43) from Deep Probe was reused as the northernmost shotpoint in the CD-ROM seismic refraction experiment [Snelson et al., 1998; Henstock et al., 1998; Gorman et al., 2002]. Since the Colorado portion of the Deep Probe profile paralleled the CD-ROM profile, it was used as a starting model for our modeling effort. In addition, the two models may be compared to speculate on the crustal structure variations in an east-west direction.

3. DATA CHARACTERISTICS

In order to span the Precambrian boundaries of interest and the southern Rocky Mountains, the CD-ROM refraction profile was 950-km in length. Thus to maintain the close station spacing desired, the experiment was carried out in two, ca. 475-km-long deployments (Figure 1). This geometry assumed that we would record waves arriving from the upper mantle, while maintaining high resolution with small receiver spacing (~ 800 m). An array of ca. 625 receivers consisting of 400 Texans (RefTek 125) and 225 RefTek (DAS) units were laid out for each deployment. Ten shotpoints were placed at ca. 100-km intervals along the profile. There were 15 successful shots during the experiment. SP 3 (Gardner, CO), SP 4 (Canon City, CO), SP 5 (Hartsel, CO), SP 7 (Kremmling, CO), and SP 10 (Day Loma, WY) (Figure 1) were shot twice and merged to increase data density. The final result was 10 merged record sections (shot gathers) of which only one (SP 6) produced just fair data quality due to shot point
loading problems. Gaps in the seismic record sections are a result of canyons or rivers and/or instrument failures.

In most cases, the data quality was high, providing clear first arrivals to offsets of > 200 km (Figures 2 – 4 and data displayed in Rumpel et al., this volume; Levander et al., this volume). The refracted phases identified in our analysis were Pg (upper and middle crust), Pl (lowermost crust), and Pn (upper mantle). Reflected phases identified were PcP (mid-crust), PIP (lowermost crust), and PmP (Moho). In addition to these phases, there are a number of upper and middle crustal reflections that are treated in detail by Rumpel et al. [this volume].

First arrival energy for Pg was strong on all shot records (Figures 2 – 5). A characteristic of this phase is that it mirrors the large changes in topography that occur along the profile. The first arrival from the lowermost crust (Pl) was identified on 7 record sections, and provides diving wave control on the velocity of the lowermost crust (e.g., Figure 2). One requirement for this phase to be a first arrival is that the thickness of the lowermost crustal layer must be at least 10 km and have a significant velocity contrast with adjacent layers. Pn was observed on only 6 record sections, because of lack of enough energy some of the shot points.

The mid-crustal reflection (PcP) was another important phase that is characterized by high amplitudes that overtake the Moho reflection (PmP) beyond the critical point making picking of post-critical PmP difficult (Figures 6 - 9). This pattern is common in most of the record sections and is the main cause for the different phase correlations used in this study and in Levander et al. [this volume]. One might pick the higher amplitude energy as PmP, because the highest amplitude event has a low apparent velocity, we identify this phase as PcP and not PmP (Figures 6 – 9). PmP is an earlier phase with an apparent velocity that approaches the average velocity of the crust. The moderate amplitude of post-critical PmP reflections indicates that the velocity contrast across the Moho is gradational (Figures 6 - 9). The lowermost crustal phase (PIP) often has high amplitudes as well, but its amplitude is exceeded by the mid-crustal reflector (PcP) on some shot records (Figures 6 - 9). The Moho reflection (PmP) is present on all record sections except for that from SP 6 (Figures 2 – 9), where the shot size was small.

4. VELOCITY MODELING

Because of the complexities caused by the overlapping nature of the main reflected phases (PcP, PIP, and PmP), we began with forward modeling to aid phase identification followed by inversion. The beginning velocity model [Snelson et al., 1998] was constructed from the adjacent Deep Probe model and modified to reflect the local geology
along the CD-ROM profile [e.g., MacLachlan et al., 1972; Woodward, 1988; Sloss, 1988; Blackstone, 1993]. Iterative forward modeling was undertaken using MacRay [Luetgert, 1992] to insure the pick quality. Over 2600 P-wave first arrivals times and over 850 reflected arrival times for PcP, PnP, and PnC were input to the inversion process. The picking error for first arrivals is estimated to be ~100 ms and for reflections ~200 ms at best.

The same set of arrival times were input to two different inversion algorithms that produced comparable velocity models. Detailed results for the upper crust using the approach of Zelt and Smith [1992] are described by Rumpel et al. [this volume], and a model for the entire crust was also obtained using this technique. Here we describe the results of first-arrival tomographic inversion followed by forward modeling of reflecting interfaces in the middle and lower crust. In addition, we describe reflectivity modeling of shot record amplitudes. This technique was used as another check on phase correlations and amplitude relationships primarily among the reflected phases, PcP, PnP, and PnC. Finally, we compare our results with two independently derived velocity models produced from the same data.

4.1 Seismic Travel Time Tomography

We derived a velocity model for the CD-ROM data set using the 3-D tomographic approach of Hole [1992]. Important parameters in our implementation of the tomographic inversion for the CD-ROM data set include (1) the choice to implement the code in 3-D, (2) choice of a starting model, and (3) choice of a smoothing schedule for updated models. The rugged terrain and limited accessibility in the Rocky Mountains necessitated an experiment lay out with a crooked-line character (Figure 1). Thus, to avoid any geometrical artifacts that would have affected a 2-D inversion, we implemented the code with a 3-D model space that was 1022 km in length (south-north) by 87 km wide (west-east) and 70 km deep using 1-km grid spacing.

The 3-D starting model was created by expanding a 1-D velocity model based on the forward modeling effort. Calculated travel times through this initial model produced an RMS misfit of 1.22 s. Tests showed that the success of the inversion scheme can be very sensitive to the starting model. For example, initial models that were slower than the one we finally chose would result in updated models for which ray paths calculations would fail. Faster models had larger starting RMS values, which simply would not reduce to reasonably low values after several iterations.

Our implementation was comprised of multiple runs of the travel time calculation, followed by inversion. For the first run, a 200 by 60 by 40 km moving average filter was
used. For subsequent runs, each dimension of the filter was reduced by about half resulting in the final smoother of 30 by 30 by 10 km. This procedure produced a final model with an RMS error of ~0.16 s and the ray coverage shown in Figure 10a.

Overall, the calculated fit to the observed travel times for the final model is excellent (Plate 1a). Nevertheless, there are still places where the misfit is as much as 0.3 s (Plate 1a). These misfits appear to be a systematic, and probably result from the inversion forcing a fit of a smooth travel time curve to the observed travel times. Thus, there are always portions of the observed curve where the calculated times overshoot the observed times [Zelt et al., 1996].

The final velocity model shows only modest variations in the crustal structure from south to north as well as from the surface to the vicinity of the Moho (Plate 1). Most lateral variations occur in the upper crust and are consistent with the results of the detailed upper crustal analysis of Rumpel et al. [this volume]. Overall, the upper crust is about 15 to 20 km thick and has an average velocity of ~6.1 km s\(^{-1}\). The transition to the middle crust is defined by an increase in velocity at ~20 km, and the middle crust is about 10 km thick with an average velocity of ~6.7 km s\(^{-1}\). The lowermost crust is characterized by a high-velocity zone that ranges in thickness from 5 to 10 km and has an average velocity of 7.2 km s\(^{-1}\). The uppermost mantle velocity, where the resolution is good, is ~7.8 to 7.9 km s\(^{-1}\). The other portions of the model where this velocity is greater than 8.0 km s\(^{-1}\) are not well resolved. The model also shows that the crust thickens by about 5 km beneath central Colorado. This result is very similar to the tomographic model derived by Levander et al. [this volume] using a different algorithm [Zelt and Barton, 1998].

4.2 Forward Modeling of Wide-Angle Reflections

Because velocity models derived from tomographic methods are smooth and do not define discrete interfaces well and because the north end of our tomographic model (Plate 1) lacks ray coverage below the middle crust (Figure 11a), we chose to model the reflected phases, PcP, PnP, and PmP to improve the definition of major crustal features in our velocity model. To do this, we implemented the technique of Hole and Zelt [1995] that calculates travel times to reflecting interfaces suspended in a velocity grid. We output the final model grid and converted it to a ProMAX\(^\text{®}\) format and we then used the velocity field editing functions to interpolate the northern end of the model. In this section of the CD-ROM profile, there is relatively little first arrival data from long offsets. We defined the velocity field where data were lacking and used a smoothing function within the velocity editor to make a continuous model from south to north such that the
northernmost portion of the model was consistent with the Deep Probe model. The velocity model was then converted to a 3-D grid and used as the starting model for the reflected interface forward modeling.

Once a starting model was defined, a reflector interface was created within the model space. The location of the reflectors was chosen based on our initial forward modeling. The reflection travel times are calculated by propagating waves down to and up from a surface placed within the model that was derived from the tomography [Hole and Zelt, 1995]. The layer depth and geometry is then adjusted until the observed and calculated traveltimes fit well and the RMS is ~ 0.200 s. This procedure was used to determine the location of the top of the mid-crustal interface (PcP), the top of the lowermost crustal layer (PlP), and the top of the Moho (PmP).

The addition of reflecting interfaces in the model provided additional constraints on crustal structure along the profile. The final model (Plate 1b) shows that the mid-crustal interface is at a depth of about 25 km at the southern end of the model and increases in depth to about 30 km before rising to about 25 km at the northern end of the profile. The top of the lowermost crustal interface is at about 35 km depth at the southern end of the profile and deepens under central Colorado to about 45 km before thinning at the northern end of the profile to about 40 km. The Moho depth at the southern end of the profile is about 45 km. The Moho deepens to about 55 km under central Colorado and rises at the northern end of the profile to about 45 km.

4.3 Reflectivity modeling

The CD-ROM data are characterized by large amplitude arrivals associated primarily with post-critical reflections from deep interfaces on most of the record sections (e.g., Figures 2 - 9). At large offsets, the move-out of the arrival that was generally the most energetic phase was relatively low (~6 km s⁻¹). Because this apparent velocity is approximately the average velocity of the column of material above the reflector and because there were arrivals that appeared to originate from deeper interfaces, we suspected that this wide-angle reflection was not associated with the Moho, but originated in the mid-crust (PcP). To confirm that this key interpretation was valid, synthetic seismograms were created to better model amplitudes seen within the seismic record sections [Fuchs and Müller, 1971; Sandmeier, 1990]. This technique is limited to 1-D models of the earth, but calculates full wave field synthetic waveforms for a user-defined set of offsets and time interval.

The Ft. Sumner (SP 1) record section was chosen as the best candidate for amplitude modeling, because the data
quality was high and multiple phases from the deep crust were present (Figure 2). The 1-D reflectivity model is based on the final tomographic model (Plate 1b; Figure 10 inset). In order to create the large amplitude arrival that has a slow apparent velocity at post-critical offsets, the 1-D model includes a sharp boundary at ~22 km depth, which is correlative to the classic mid-crustal Conrad discontinuity (Figure 10a). In order to reproduce the PIP phase another relatively sharp boundary is necessary at ~35 km depth (Figure 10a). The high-velocity lowermost crust is represented by a zone of gradational velocity increase (7.0 to 7.4 km s\(^{-1}\)) down to a depth of 42 km, where a transition to mantle velocities of about 7.8 km s\(^{-1}\) occurs.

The synthetic record section produced from this 1-D model (Figure 10) shows waveforms that are similar to the original data. At pre-critical offsets (~100 km), the relative amplitudes of the PcP, PIP, and PmP phases are comparable to the observed data, as is the moveout of the reflections (Figure 10). Both the Pl and PIP phases from the lowermost crust match those in the original data. At post-critical offsets (>220 km), PcP has an apparent velocity of about 6.0 km s\(^{-1}\), as in the observed data. PmP and PIP merge into a complex waveform whose relative amplitude is similar to the observed data near the critical point for PmP, but at greater offsets, the amplitude of this phase is larger in the synthetic seismograms than in the observed data. We could not simultaneously match the observed amplitudes of the pre-critical and post critical amplitudes of the PIP/PmP reflections, which suggest complexity that is beyond the 1-D approximation of the reflectivity method. In general, the observed data (Figures 2 - 10) include a considerable amount of reflectivity that follows the PmP and PcP phases at pre-critical offsets and the PcP phase at post-critical offsets. We consistently picked the beginning of this reflectivity as the arrival time of the phase, but this reflectivity is an indication of complex layering that is not 1-D in nature.

The same pattern of deep reflectors (PcP, PIP, and PmP) is evident along the profile as far north as southernmost Wyoming. However, the confidence with which we could identify all three deep reflections decreased at the northern end of the profile. Our approach of identifying these phases on each record section represents a bias towards making consistent phase correlations along the profile. Because of the integrated nature of our modeling approach, this bias does not greatly alter the overall crustal structure. However, in the following analysis, the reflecting interfaces appear more continuous than they probably are in the earth. This approach to phase correlations is different than that of Levander et al. [this volume] and explains some of the differences in the velocity models derived. However, comparison of the various velocity models shows that the
tectonic implications of these differences is not significant [Keller et al., this volume].

5. MEASURES OF RESOLUTION

A sense for the model resolution can be gained by evaluating the traveltime fits, ray coverage, and RMS error. Unfortunately with the tomographic inversion, a resolution matrix is not created because the technique is non-linear [Hole, 1992]. We have also used the uncertainties obtained from RAYINVR by Rumpel et al. [this volume] as a guide for discussing the relative uncertainties in our model.

5.1. Ray Coverage

The ray coverage or hit count represents the number of rays that pass through a particular cell. The velocity estimate for a given cell may be considered more reliable for a cell with higher hits counts. Considering the modest number of shots in the experiment, the ray coverage from first arrivals is adequate except for the deep portion of the northern third of the profile (Figure 11a). The inclusion of the reflected phases greatly increases the ray coverage in that part of the model in particular. Using the velocity model produced by Rumpel et al. [this volume], the ray coverage for the reflected arrivals as output from RAYINVR is shown in Figure 11b. Figure 11c shows the bounce points for the reflections as well as some of the refracted arrivals, but it is apparent that the main control for the lower portion of the model is the reflected arrivals.

5.2. Estimated Resolution

Based on forward and inversion modeling in collaboration with Rumpel et al. [this volume] and the tomographic inversion of the first arrivals and reflections, we feel that the resolution of the depth for the deep interfaces is +/- 2 km if the velocity structure and phase identification is assumed to be completely accurate. However, if one considers the uncertainty of the velocity, then the uncertainty for deeper interfaces could be as much as +/- 3 km. The estimated uncertainty related to the upper crustal velocity field is +/- 0.1 km s\(^{-1}\) a depth of about 25 km and then increases up to about +/- 0.2 km s\(^{-1}\) in the lower crust and upper mantle. Figure 11c shows the locations where rays crossed the various interfaces within the model space. These reflection points show where each interface is being sampled in the model space. As expected, the resolution is best for the central portion of the model.

6. INTEGRATION WITH GRAVITY DATA
The Bouguer gravity map of the southern Rocky Mountains (Plate 2) is characterized by a long wavelength, ca. 150 mGal gravity low centered on the highest topography in Colorado. Definition of the source of this anomaly is controversial and is critical in understanding the mechanism of isostatic compensation of the topography. Elements of this anomaly have been variously interpreted as resulting from a shallow crustal source [Isaacs and Smithson, 1976; McCoy et al., this volume], a combination of lateral density variations in the crust and Moho relief [Li et al., 2002] and significant density variations in the mantle [Sheehan et al., 1995]. The new controls on crustal velocity and thickness as well as upper mantle velocity presented here, together with the Deep Probe results [Gorman et al., 2002] allow us to investigate the source of this anomaly and address the implications for isostatic compensation.

6.1. Density Modeling

In our gravity modeling, we used an implementation of the 2.5D algorithm of Cady [1980]. We employed an iterative forward modeling approach that incorporated all available geological and geophysical constraints in order to match the observed Bouguer anomaly values. The initial gravity model was derived from the final CD-ROM tomographic velocity model and our Deep Probe velocity model [Snelson et al., 1998]. Initial density values were calculated from experimentally determined velocity/density relationships [e.g., Christensen and Mooney, 1995]. To guide modeling of the upper crust, a geologic cross-section along the profile was constructed using geologic maps and compiled drill hole data in the region [Suleiman and Keller, 1985; Blackstone, 1993; Snelson et al., 1998; Treviño and Keller, this volume]. Additional constraints were provided by various seismic reflection and refraction profiles [Behrendt et al., 1969; Wellborn, 1977; Applegate and Rose, 1985; Beggs, 1985; Gries and Dyer, 1985; Kaplan and Skeen, 1985; Lange and Wellborn, 1985; Prodehl et al., this volume]. Features discernible in the detailed upper crustal velocity model of Rumpel et al. [this volume] correlate very well with surface geology and were used as a constraint in the density modeling and taken as an additional guide to the location of major upper crustal features important to the gravity model (Figure 12).

The final gravity model (Figure 12) reveals a number of new aspects of the lithospheric structure of the Rocky Mountains. As expected, known geologic features of the uppermost crust contribute significantly to short-wavelength gravity anomalies. Four upper crustal density anomalies play a significant role in fitting the intermediate wavelength (~300 km) features in New Mexico and Colorado. The occurrence of these features is evident in the detailed velocity modeling of Rumpel et al. [this volume]
and is consistent with other geologic evidence for both low and high-density intrusions within the crystalline crust. In particular, to match the steep gravity gradients on either side of the gravity low in the central portion of the profile, an upper crustal body with density 2600 kg m\(^{-3}\) must be surrounded by two high-density bodies of 2900 kg m\(^{-3}\) with similar depth extent to 10 km. When these bodies are removed from the model (Figure 12a), a much broader, lower-amplitude (ca. 80 mGal) gravity low remains suggesting a deeper feature is controlling the gravity signature.

The large gravity low at a distance of ~650 km in the model is part of a prominent northeast-trending anomaly that correlates with the Colorado mineral belt (Figure 12) [McCoy et al., this volume]. A number of Laramide felsic intrusions crop out along this belt, which also has the same approximate trend as Precambrian terrane boundaries. No younger features are present that could explain this long-wavelength anomaly, and thus, it is attributed to a series of large felsic bodies, probably intrusions, in the upper crust.

The Sierra Grande uplift region (northeastern New Mexico) is along the northwest extension of the Wichita-Amarillo uplift [Suleiman and Keller, 1985], which is part of the Southern Oklahoma aulacogen. Particularly in Oklahoma, mafic intrusions of Cambrian age are well documented under the Wichita uplift [e.g., Keller and Baldridge, 1995]. In addition, the 1.1 Ga Pecos mafic igneous complex extends northward from Texas and southeastern New Mexico, where it is well documented in drill holes and seismic reflection data [Adams and Miller, 1995]. The gravity modeling of Suleiman and Keller [1985] showed that the basement relief in northeastern New Mexico is insufficient to produce the observed gravity high that crosses the Sierra Grande uplift region, and thus they include mafic material in the upper crust of their crustal models. Thus, the southern mafic body in Figure 12d could either be due to Cambrian or late Proterozoic magmatism.

The Wet Mountains area is associated with a gravity high and the dense upper crustal body placed in the model is consistent with the deep origin of the Precambrian rocks exposed in the mountains [C. Andronicos, personal communication]. If the highly metamorphosed rocks at the surface are from depths of 25 km or more, then rocks that were once lower crustal are present in the upper crust below them. The origin of the gravity high associated with the Sierra Madre/Park Range is less clear. The simple body used in the model does not fit some short wavelength features in the observed data, and it is centered just south of the Cheyenne belt. The exposed geology in this area is multifaceted and includes large mafic and felsic plutons [e.g., Foster et al., 1999]. In addition, the presence of the Cheyenne belt and several major shear zones probably
produce local anomalies that are beyond the scale of this study.

With the upper crustal density structure well constrained by known geology and velocity modeling, a southward decrease in mantle density is required to match the observed data. This density decrease is completely consistent with the southward decrease in mantle velocity observed in data from the Deep Probe experiment [Henstock et al., 1998; Snelson et al., 1998]. When the modeled gravity is recalculated using a single density for the mantle (Figure 12b) the overall shape of the calculated curve remains similar to the observed curve, but the calculated values are less than the observed values by >50 mGal on the north end of the model. Thus, the long wavelength gravity low that extends from about 200 to 1000 km in the model clearly correlates with the both the shape of the Moho and the topography, but lateral density variations in the upper mantle (Figure 12c) are needed to produce a fit on the north end.

7. DISCUSSION

7.1. Mid-Crustal Velocity Boundary

Below the laterally heterogeneous velocity field of the upper crust [Rumpel et al., this volume], the middle crust is characterized by a well-defined boundary at 20 to 25 km depth, which is marked by a change in velocity from 6.4 to 6.6 km s\(^{-1}\). This interface location is defined by prominent PcP reflected energy found on all the shot records. The energy from the PcP is characterized by a long coda that may result from multiple reflections within the upper crust (Figures 2 and 10). Our modeling suggests that the main reflecting boundary dips gently northward beneath the Great Plains to reach a maximum depth of ca. 26 km, as the profile crosses the Rocky Mountain Front near model coordinate 500 km (Plate 1). It then rises gently northward to ca. 23 km depth at the north end of the model. Whereas reflections from this boundary are prominent in the wide-angle data presented here, the near-vertical incidence data collected as part of the CD-ROM effort exhibit no corresponding sub-horizontal reflectivity at this level [Magnani et al., this volume a, b; Morozova et al., this volume]. This observation suggests that the boundary is characterized by a velocity gradient that it is transparent to the shorter wavelength content of the near-vertical incidence data.

More recent results from inferred exposures of the Conrad discontinuity [Salisbury and Fountain, 1994; Lana et al., 2003] and from xenoliths [Sachs and Hansteen, 2000] indicate that the Conrad is both a metamorphic and compositional boundary. Comparison of representative 1-D velocity-depth functions from the CD-ROM velocity model
to laboratory measurements of rock velocities (Figure 13) shows that the mid-crustal velocity discontinuity along the CD-ROM profile also marks the transition from model velocities appropriate for lower-grade rocks with more felsic compositions to velocities appropriate for higher grade rocks with more mafic compositions. Refraction results from the central United States [Braile, 1989] and earlier results from the Rocky Mountains [Prodehl and Lipman, 1989] map a similar velocity discontinuity near 20 km depth, although the Deep Probe results from western Colorado and New Mexico do not require its presence [Snelson et al., 1998] (Figure 13). Thus, most of the evidence suggests that this boundary is a widespread feature of the crust in the Great Plains and Rocky Mountains.

We interpret the observation that the mid-crustal boundary extends from the Great Plains through the Rocky Mountains as evidence that it is a Paleoproterozoic/Mesoproterozoic-age boundary that has remained essentially undisturbed by Phanerozoic tectono-magmatic events such as the late Paleozoic Ancestral Rockies event and the Laramide orogeny. Reflectivity interpreted as the signature of the Proterozoic orogens in the near-vertical incidence data [Magnani et al., this volume b; Morozova et al., this volume] crosscuts the discontinuity suggesting that the formation of this boundary must have been contemporaneous with the last Proterozoic orogenic pulse. Otherwise, this reflectivity should have been more profoundly disrupted by subsequent differentiation processes.

It is likely that subsequent magmatic events may have served to enhance this boundary. Models for stabilization and evolution of continental crust [e.g., Bohlen and Mezger, 1989; Nelson, 1991; Keller et al., this volume] commonly invoke episodic injection of the crust by mafic magmas as a mechanism for thickening the crust and pushing the lower crust towards a more mafic composition. The 1.4 Ga “anorogenic” magmatic event that affected both the Great Plains and the Rocky Mountains, as well as voluminous Tertiary magmatism in the southern Rockies, are at least two candidates for times when this velocity boundary may have been enhanced.

7.2. High Velocity Lower Crust

The lowermost crust of our velocity model is characterized by a ca. 10 km thick layer with a velocity of ca. 7.2 km s\(^{-1}\). This feature is defined from wide-angle reflections (PlP) (Figures 2 - 9) and a number of shot records (Figures 2 - 3, 5 - 6, 8 - 9) that we interpret to contain a refracted arrival from the lower crust. A single continuous layer is the simplest way to model this feature; however, the data allow for the possibility that the
thickness of this layer may vary significantly beneath New Mexico and southern Colorado, as is the case in the velocity model of Levander et al. [this volume]. The model presented in this paper is considered a thick end member for the high-velocity lower crust compared to the smaller thicknesses derived by Levander et al. [this volume]. Furthermore, the data do not require its presence at the north end of the profile.

Laboratory measurements (Figure 13) [e.g., Christensen and Mooney, 1995] and theoretical calculations [Furlong and Fountain, 1986] show that velocities ca. 7.2 km s\(^{-1}\) are appropriate for mafic lithologies including gabbroic rocks and mafic garnet granulite. A high-velocity lower-crustal layer is common to many regions of the central and western U. S. including the mid-continent [Braile, 1989], the Colorado Plateau [Wolf and Cipar, 1993], and the Deep Probe profile [Snelson et al., 1998]. Such a layer is also observed in other regions of the world, with the Baltic shield [e.g., EUROBRIDGE Seismic Working Group, 1999] being a notable example. These features are most commonly interpreted as mafic magmatic underplates emplaced during a major melting event, and are consistent with observed high velocities and magmatic history of the region.

We also interpret the high-velocity layer on the CD-ROM profile as a magmatic underplate. It probably is in part a relic of the initial formation of the crust, but based on its continuity across the Great Plains-Rocky Mountain tectonic boundary, we propose that a significant portion of it was likely emplaced during the voluminous ca. 1.4 Ga magmatic event [Anderson, 1989]. This event profoundly modified the crust of the Great Plains to form the southern Granite-Rhyolite province and also led to the intrusion of many plutons along Proterozoic shear zones in the Yavapai and Mazatzal provinces [Karlstrom and Humphreys, 1998]. The high-velocity layer beneath the Baltic Shield is also associated with magmatism that is chemically similar to, but, at ca. 1.5 Ga, somewhat older [e.g., Gaál and Gorbatschev, 1987]. Subsequent magmatic episodes, including events at ca. 1.1 Ga, 500 Ma, and in Tertiary time may also have contributed material to this layer in the southern Rocky Mountains [Keller et al., this volume].

The unusually thick (~20 km) high-velocity (~7.2 km s\(^{-1}\)) lower-crustal layer observed along the Deep Probe transect (Figure 13) [Henstock et al., 1998; Snelson et al., 1998; Gorman et al., 2002] likely represents a mafic underplate that is Archean or Paleoproterozoic in age rather than Mesoproterozoic and younger. This layer extends for 700 km from central Wyoming into southern Alberta, where it terminates at the north side of the Medicine Hat block [Gorman et al., 2002; Clowes et al., 2002]. These regions lack evidence for ca. 1.4 Ga magmatism. However, in Wyoming and Montana, Chamberlain et al. [2003] have
mapped at least five thermal events of Archean age that could represent the signature of underplating events. By contrast, xenoliths interpreted as originating from the high-velocity layer beneath the Medicine Hat Block in southern Alberta have Paleoproterozoic ages of 1814-1745 Ma [Davis et al., 1995]. In addition, the layer beneath these Archean crustal blocks is not physically continuous with the high-velocity lower-crustal layer present in the CD-ROM model (Figure 13).

7.3. Crustal Thickness and Upper Mantle Velocity

One observation is that the depth to these interfaces constrained by reflections is typically shallower than the depth one would have picked based on the tomographic velocity model alone (Plate 1b). This suggests that the velocity in the crust may be somewhat slower at shallower depths than is required by the tomographic results.

Our final velocity model (Plate 1c) shows that the crust is 40 to 45 km thick crust beneath the Great Plains of New Mexico and that the crust thickens to about 50 km beneath the high topography of southern and central Colorado. Previous models [Johnson et al., 1984; Prodehl and Lipman, 1989; Snelson et al., 1998; Prodehl et al., this volume] suggest that the crust thins again to about 40 to 45 km beneath southern Wyoming, although our data do not require such thinning.

Upper mantle velocities range from 7.8 to 7.9 km s\(^{-1}\) along the entire profile, and the analysis of Levander et al. [this volume] shows a distinct southward decrease of velocity southward to the vicinity of the Jemez lineament. These values are similar to those obtained in the Deep Probe velocity model south of the Cheyenne belt [Henstock et al., 1998] and the earlier results of Prodehl and Lipman [1989]. These values are low compared to averages for stable continents and suggest that the present-day mantle is warm and buoyant. Low upper mantle velocities are broadly consistent with large-scale mantle velocity models [e.g., Grand et al., 1997; van der Lee and Nolet, 1997] that map slow, hot upper mantle beneath the uplifted orogenic plateau of western North America, and with more detailed teleseismic tomography results [Zurek and Dueker, this volume] that map generally low P-wave velocities in the upper 200 km of the mantle in this region. Asthenospheric upwelling beneath the modern Rio Grande rift likely plays an important role in reducing upper mantle velocities along the CD-ROM profile.

7.4. Implications for Isostatic Compensation of Topography

By constraining crustal thickness in Colorado through analysis of receiver functions, Sheehan et al. [1995] were
able demonstrate that a significant portion of the support for the excess topography must come from the mantle. However, they were unable to document the role of crustal density variations in detail because the receiver function method does not produce information on lateral velocity variations in the crust, nor can it estimate Moho depth as accurately as refraction data because of the long wavelengths involved. Li et al. [2002] used shear wave velocity structure based on measurements of Rayleigh wave dispersion concluded that the density anomaly mostly responsible for the high topography must reside in the crust rather than the mantle. Surface wave data are characterized by even longer wavelengths than those used in receiver functions, thus leading to lower resolution results. Furthermore, the inversion for a match to the Bouguer anomaly employed by Li et al. [2002] may have underestimated density variation in the mantle due to the sensitivity of shear wave velocity to the presence of melt.

The density model presented here (Figure 12) has the advantage of having been constructed from a shorter wavelength seismic source sensitive to lateral variations in crustal velocity. The modeling serves to separate the three main elements that contribute to low Bouguer gravity values in the southern Rocky Mountains: (1) thickened crust beneath the high topography; (2) a decrease in mantle density beneath the high topography; and (3) low density intrusive bodies in the upper crust. Thickened crust contributes to a long wavelength gravity low centered on the Colorado Mineral belt (Figure 12a), whereas the difference in upper mantle density contributes to a mass deficit in the southern portion of the profile (Figure 12b). Upper crustal low-density bodies increase the magnitude of the low (Figure 12c).

We conclude that compensation of the high topography in the southern Rocky Mountains results from a compound interplay of density variations within the lithosphere, most of which has probably developed since late Cretaceous time. Crustal thickening may have occurred during Laramide shortening as a result of lower crustal flow [e.g., Bird, 1998] or large-scale detachments [e.g., Erslev, this volume]. Buoyancy in the upper crust would have been added during Tertiary-age magmatism in the Colorado mineral belt and the San Juan volcanic field [e.g., Stein and Crock, 1990; Decker, 1995]. Thermal support from the mantle may have been added in Cenozoic time, possibly as a result of removal of the mantle lithosphere [e.g., Decker, 1995; Humphreys et al., 2001].

8. CONCLUSIONS

The CD-ROM refraction profile crosses a major modern physiographic province and a number of major Precambrian tectonic boundaries, yet displays remarkably
simple velocity structure. We interpret this as evidence that the crustal architecture developed primarily during Paleo- and Mesoproterozoic tectonism and that subsequent Phanerozoic events caused only minor modification or enhancement of the existing structure. We propose that a two-layer crust developed during accretion of Yavapai and Mazatzal crust from 1.76 to 1.6 Ga. Differentiation of the crust into an upper felsic layer and a lower more mafic layer probably occurred primarily through melting of the lower crust to produce widespread felsic plutons at ca. 1.6 Ga that now crop out at the surface. Later magmatic episodes, particularly at 1.4 Ga and in Tertiary time may have enhanced this basic structure through melting of lower crustal rocks, emplacement of felsic rocks in the upper crust, and segregation of mafic components to the lower crust. Underplating of basaltic melts to the base of the crust at 1.4 Ga regionally, and perhaps at 1.1 Ga and 500 Ma locally caused formation of a third, high-velocity layer at the base of the crust, as well as crustal thickening. At a crustal scale, the only evidence for Laramide shortening is a modest deepening of the mid-crustal discontinuity beneath the high topography and a corresponding thickening of the crust.

The control on lithospheric architecture provided by the refraction velocity model permit a new assessment of the factors that contribute to the isostatic compensation of high topography in the southern Rocky Mountains. A density model derived from the velocity model shows that a multifaceted interplay of density variations within the lithosphere is required to explain the Bouguer gravity low associated with the range in central and southern Colorado. These density variations include crustal thickening, a transition from high density to low-density mantle, and low-density material in the upper crust. All of these features likely developed since late Cretaceous time as a result of Laramide-age shortening and magmatism. This is yet another demonstration that, in general, a number of mechanisms are commonly at work to accomplish compensation of high topography on the continents.

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FIGURE CAPTIONS

Figure 1. Basemap for the CD-ROM geophysical experiments. The seismic refraction line is the thin solid black line with stars representing the shot point locations. Thick short black lines are the seismic reflection profiles acquired. The long thick black line is the location of the Deep Probe profile. The gray diamonds are the teleseismic arrays. Precambrian outcrops are in a solid dark gray. The Colorado Plateau is outlined in short dashed lines. Precambrian suture zones are represented by large dashed lines. Geologic features associated with the density model are as follows: SG (Sierra Grande Arch), RB (Raton Basin), WM (Wet Mountains), SoP (South Park), CMB (Colorado Mineral belt), NP (North Park), SM (Sierra Madre), HB (Hanna Basin), WRB (Wind River Basin), and BHB (Big Horn Basin). This map was partially produced using Generic Mapping Tool [Wessel and Smith, 1998].

Figure 2. Seismic record section for Fort Sumner, NM (SP 1). Shot point location is always at 0 km. All record sections are reduced at 6 km s−1. Observed picks are overlain onto the record section as gray lines and phases are labeled on all record sections. Pg is an upper crustal refraction, Pl is a lower-crustal refraction, Pn is Moho refraction, PcP is a mid-crustal reflection, PIP is a lower-crustal reflection, and PnP is the Moho reflection. These phase definitions will carry forward to Figures 3-10. Bar at base of figure indicates the lateral extent of Figure 6 relative to this figure. The seismic record sections were produced using Generic Mapping Tool [Wessel and Smith, 1998].

Figure 3. Seismic record section for Wagon Mound, NM (SP 2).

Figure 4. Seismic record section for Rawlins, WY (SP 9).

Figure 5. Seismic record section for Day Loma, WY (SP 10). Bar at base of figure indicates the lateral extent of Figure 9 relative to this figure.

Figure 6. Close-up of the seismic record section from Figure 2 for the shot point at Fort Sumner, NM (SP 1). Notice the clear first arrivals of the Pl phase.

Figure 7. Close-up of the seismic record section for Hartsel, CO (SP 5) to the south. Notice the high amplitude energy of the PcP phase.

Figure 8. Close-up of the seismic record section for Kremmling, CO (SP 7) to the south. Notice the high amplitude energy of the PcP phase. Also there are clear Pl first arrivals.

Figure 9. Close-up of the seismic record section from Figure 5 for Day Loma, WY (SP 10). Notice the high amplitude energy of the PcP phase.

Figure 10. (a) The gradational 1-D model used to produce the synthetic record section. (b) The seismic record section for Fort Sumner, NM (SP 1) reduced at 8 km s−1. (c) The synthetic seismic record for Fort Sumner, NM (SP 1). Notice the high amplitude energy of the reflected phases, which is similar to (b).

Figure 11. (a) Ray coverage for the 3-D velocity model. Hit count is the number of rays that encountered each cell. High hits are dark gray to black and low hits are lighter shades of gray. The 3-D model has been compressed to two dimensions for display. Thus, the hit count in a given cell represents a sum of hits from cells in the y-direction. Shot points are noted at the top of the plot. (b) Ray coverage from travelt ime modeling by Rumpel et al. [this volume]. The northern portion of model shows only reflections and no refractions as indicated by the lack of diving waves. Light gray rays are upper crustal, medium gray is middle crustal, and dark gray to black are lower crustal and mantle reflections. (c) Reflection points from travelt ime modeling by Rumpel et al. [this volume]. The model is well resolved from the wide-angle reflections. Dark black lines are the bounce points for the model.

Figure 12. (a) 2.5-D gravity values along the CD-ROM transect showing the effects of removing the intrusive bodies in the upper crust; (b) 2.5-D gravity values along the CD-ROM transect showing the effects of including a homogeneous mantle; (c) 2.5-D gravity values along the CD-ROM transect showing the preferred model fit to the observations; and (d) 2.5-D density model along the CD-ROM transect. The density values are in kg m−3.

Figure 13. Comparison of laboratory measurements of selected rocks types [Christensen and Mooney, 1995] to 1-D velocity-depth functions extracted from the (a) Deep Probe and (b) CD-ROM velocity models. Horizontal dashed line at 20 km and 30 km depth indicate possible range for the Conrad discontinuity. Velocity is in km s−1.
Plate 1. (a) Traveltime fits from the 3-D tomographic velocity model. The plot is reduced at 6 km s\(^{-1}\). Red plus signs are the observed picks, blue triangles are the calculated picks from the inversion, and the green circles are the residuals or difference between the observed and calculated. The residuals are shifted by 7 s to show them more clearly. (b) First arrival velocity model from the tomography. Velocity is in km s\(^{-1}\). Contour interval is 0.5 km s\(^{-1}\). The final 3-D model has been compressed to two dimensions for display. To do this, the velocity in a given cell was calculated from a weighted average of cells in the y-direction, where the weight is the hit count. Thus, velocity values in cells with higher hit counts receive a larger weight in the average than those with only a few hits. Shot points are noted at the top of the plot. (c) The final velocity model with the reflected interfaces overlain on the velocity field. Velocity is in km s\(^{-1}\). Shot points are noted at the top of the plot. Data were gridded and smoothed for display with Generic Mapping Tools [Wessel and Smith, 1998].

Plate 2. Bouguer anomaly map of the southern Rocky Mountains. Contour interval is 50 mGal. Red line is the seismic refraction profile with shot points represented by black stars. Map is based on data compiled by the National Geodetic Survey (http://www.ngdc.noaa.gov) that were gridded and smoothed for display with Generic Mapping Tools [Wessel and Smith, 1998]. Black dots are locations of gravity values within a distance no greater than 5 km from the seismic profile that were used in modeling (Figure 12).

DEEP CRUSTAL STRUCTURE OF THE ROCKY MOUNTAINS

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Figure 2. Snelson et al.
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Figure 11. Snelson et al.
Figure 12. Snelson et al.
Plate 1. Snelson et al.
Plate 2. Snelson et al.