



Upper mantle structure beneath the eastern Colorado Plateau and Rio Grande rift revealed by Bouguer gravity, seismic velocities, and xenolith data

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[1] The lithospheric architecture of the Colorado Plateau–Rio Grande rift transition zone reflects a narrow corridor of active tectonism and magmatism between the tectonically stable Colorado Plateau and the stable interior of North America. In this paper we analyze Bouguer gravity data, crustal thickness variations, xenolith data, and seismic wave speed structure to argue that the Rio Grande rift and southeastern Colorado Plateau is underlain by a low-density upper mantle province that does not trend along upper crustal tectonic boundaries but, rather, is correlated with regions of middle to late Tertiary magmatism. We model uppermost mantle density contrasts using Bouguer gravity along five parallel profiles across the Colorado Plateau and validate our models by converting modeled density contrasts to seismic velocity variations for comparison with regional seismic data, as well as comparing modeled densities to calculated densities from xenoliths. We find that upper crustal features and crustal thickness variations do not explain long-wavelength (>200 km) variations in Bouguer gravity across the Colorado Plateau–Rio Grande rift. A large 200–440 km wide region of low uppermost mantle density is modeled beneath the eastern Colorado Plateau margin, Jemez Lineament, Rio Grande rift, and parts of the westernmost Great Plains. This region has a density contrast of -90 to -30 kg/m³ relative to the adjacent Colorado Plateau and Great Plains provinces and coincides with previously imaged regions of low upper mantle seismic velocities. In situ densities inferred from xenolith data in the Colorado Plateau support the modeled upper mantle density variations and support the idea that the eastern and western margins of the plateau have been significantly modified by metasomatism. We suggest that the region of low upper mantle density likely reflects some combination of temperature variations, partial melt contents, and compositional variations related to Tertiary magmatic activity and/or thinned lithosphere. The relationship of this zone of modified upper mantle to extension in the Rio Grande rift remains enigmatic, however, particularly given the lack of spatial correlation between the trends of upper mantle structures and surficial extensional structures.

Components: 10,905 words, 7 figures, 3 tables.

Keywords: Colorado Plateau; gravity; isostasy; Rio Grande rift; xenoliths.

Index Terms: 8120 Tectonophysics: Dynamics of lithosphere and mantle: general (1213); 8122 Tectonophysics: Dynamics: gravity and tectonics; 8178 Tectonophysics: Tectonics and magmatism.

Received 2 May 2005; **Revised** 27 July 2005; **Accepted** 29 August 2005; **Published** 21 October 2005.

Roy, M., J. K. MacCarthy, and J. Selverstone (2005), Upper mantle structure beneath the eastern Colorado Plateau and Rio Grande rift revealed by Bouguer gravity, seismic velocities, and xenolith data, *Geochem. Geophys. Geosyst.*, 6, Q10007, doi:10.1029/2005GC001008.

1. Introduction

[2] The region of the Rio Grande rift and adjacent provinces represents a lithospheric-scale corridor that has localized tectonic and magmatic activity during Tertiary time. The region underwent crustal shortening during the Laramide (70–40 Ma) orogeny followed by crustal extension during Neogene (35–20 Ma) Rio Grande rift extension. This transition in tectonic regimes was accompanied by a voluminous middle Tertiary bimodal magmatic episode (the “ignimbrite flare-up”) and late Tertiary to ongoing basaltic magmatism associated with the Rio Grande rift and the Jemez Lineament [Balch *et al.*, 1997; Baldrige *et al.*, 1991; McMillan *et al.*, 2000]. The region exhibits anomalously high elevation, active magmatism, high heat flow, and lateral variations in uppermost mantle seismic velocities of up to 10% [Gao *et al.*, 2004; West *et al.*, 2004; Wilson *et al.*, 2005] that may be explained by variations in upper mantle temperature of up to 300–470°C [Goes and van der Lee, 2002; West *et al.*, 2004]. The lithosphere of the Rio Grande rift region is therefore a zone of heterogeneity between the cold, thick, tectonically stable lithosphere of the western Great Plains and the tectonically stable Colorado Plateau block. The observed upper mantle heterogeneity beneath the Rio Grande rift province and surrounding regions is probably due to a combination of variations in lithospheric thickness, composition, temperature, and partial melt content, all of which are likely to be reflected in the regional upper mantle density structure. Our primary goal is to understand the state of the lithosphere in the Rio Grande rift corridor using an integration of gravity, seismic wave speed, and xenolith data to build a regional picture of upper mantle density structure.

[3] Previous studies have noted a long-wavelength, negative Bouguer gravity anomaly along the Rio Grande rift and the southeastern margin of the Colorado Plateau [Cordell *et al.*, 1982; Olsen *et al.*, 1987; Slack *et al.*, 1996; Spence and Gross, 1990] and we note here that this feature is oblique to the upper crustal expression of the rift axis (Figure 1). This has been variously attributed to an upwarping of low-density asthenosphere [Olsen

et al., 1987], partial melt beneath the Jemez Lineament [Spence and Gross, 1990], and a broad region of upper mantle extension in the rift region [Slack *et al.*, 1996; Wilson *et al.*, 2005]. To the north, this wide zone of negative Bouguer anomaly merges with the low-gravity province beneath the San Juan volcanic field and the Colorado Mineral Belt, which has also been interpreted in terms of upper mantle density variations [McCoy *et al.*, 2004; Roy *et al.*, 2004]. This study is the first attempt to quantitatively assess upper mantle density variations that can explain the broad zone of low gravity obliquely surrounding the Rio Grande rift and whether these variations are consistent with imaged upper mantle seismic wave speed structure and xenolith data. We present results of 2.5D forward gravity models of the crust and upper mantle to 100 km depth along five northwest-southeast trending profiles that cross the western Great Plains, Rio Grande rift, and Colorado Plateau perpendicular to the southeastern plateau margin (Table 1; Figure 1). The upper mantle density variations required in our gravity models are then compared to those expected from the seismic wave speed structure and from xenolith data to determine the robustness of our interpretations and to explore implications for the physical state of the upper mantle beneath the study area.

2. Upper Mantle Density and Relation to Topography

[4] The Rio Grande rift corridor, including the easternmost Colorado Plateau and westernmost Great Plains, is characterized by anomalously high topography that is unlikely to be supported by crustal thickness variations [Sheehan *et al.*, 1995; West *et al.*, 2004]. Crustal thickness in the eastern Colorado Plateau and western Great Plains is 40–45 km, and is 30–35 km beneath the basins of the Rio Grande rift [Gilbert and Sheehan, 2004; Keller *et al.*, 1998; West *et al.*, 2004; Wilson *et al.*, 2005]. East-trending topographic profiles clearly show that, although the basins of the Rio Grande rift occupy local topographic depressions due to crustal extension, the topographic expression of the rift is superposed on a long-wavelength topographic bulge (the 800–1000 km wide “Alvarado Ridge”

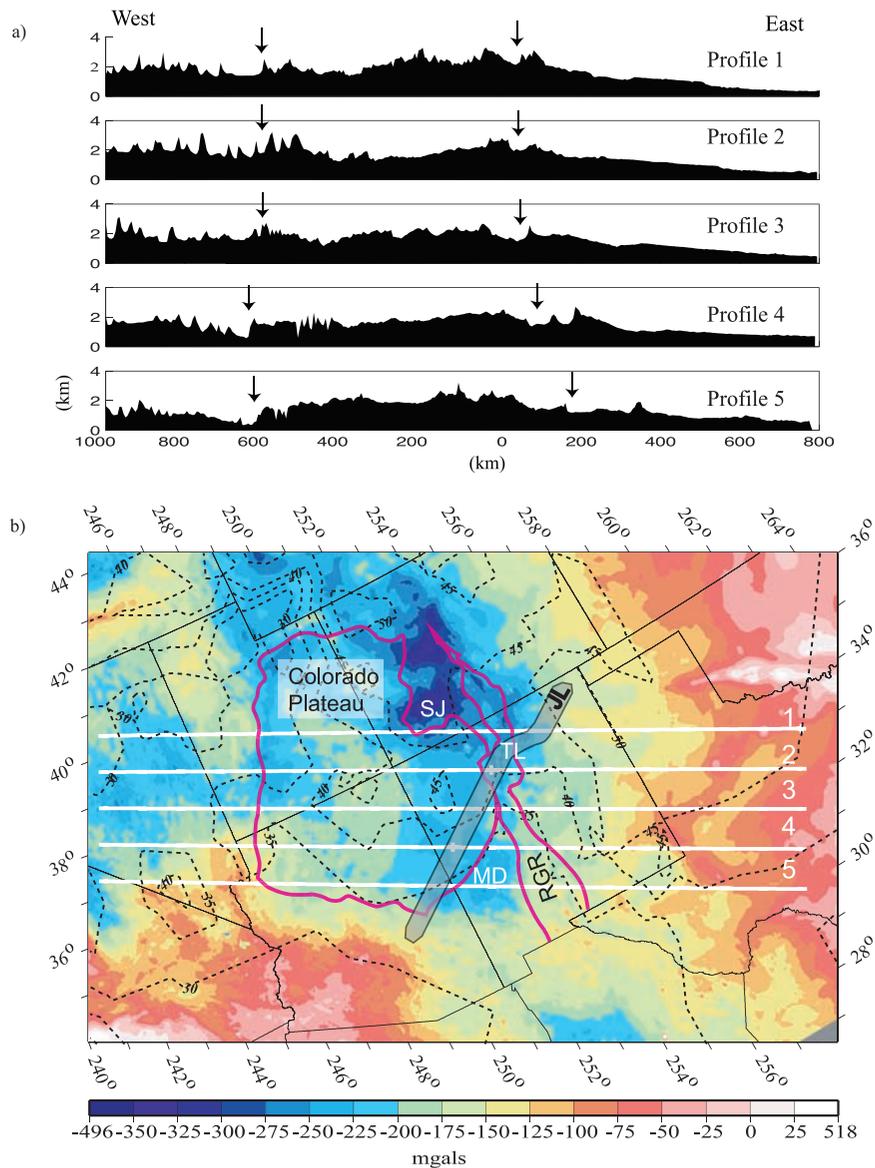


Figure 1. (a) Elevation along five study profiles across the Colorado Plateau region. The somewhat symmetric, long-wavelength topographic high surrounding the eastern margin of the Colorado Plateau is the “Alvarado Ridge” [Eaton, 1986]; arrows indicate the approximate locations of the western margin of the plateau and the eastern margin/Rio Grande rift transition. (b) Gridded Bouguer gravity (colors) with crustal thickness variations (dashed contours; km) gridded from a combined data set of LA RISTRA data [Wilson *et al.*, 2005], data from Gilbert and Sheehan [2004], and the Crust 2.0 model (G. Laske, REM Web page, <http://mahi.ucsd.edu/Gabi/rem.dir/rem.home.html>, 2005). Locations of profiles 1–5 (white lines) were chosen to cross perpendicular to the negative Bouguer anomaly at the eastern margin of the Colorado Plateau and Rio Grande rift. Tectonic province boundaries are as indicated (magenta); RGR, Rio Grande rift. Major middle Tertiary volcanic fields include the Mogollon-Datil (MD), the Taos-Latir (TL), and the San Juan (SJ) volcanic fields. Late Tertiary volcanism associated with the Jemez lineament (JL) coincides with the gray area.

[Eaton, 1986]), with highest average elevations at the eastern Colorado Plateau margin (Figure 1).

[5] The Rio Grande rift is oriented roughly north-south, from Leadville, Colorado, in the north, widening and merging with the southern Basin and Range province over 1000 km to the south

[Chapin and Cather, 1994; Keller and Baldrige, 1999]. Extension in the rift varies from 6% in the north to 20–30% at the Albuquerque basin and increases to 50% in southern parts of the rift [Chapin and Cather, 1994]. Flexural uplift models [Brown and Phillips, 1997; Peterson and Roy, 2005; Roy *et al.*, 1999] and regional low-

Table 1. Forward Gravity Model Profile Locations^a

Profile	Western Endpoint (Lon, Lat)	Eastern Endpoint (Lon, Lat)
p1	-116.476, 40.468	-98.440, 33.071
p2	-116.882, 39.688	-99.034, 32.285
p3	-117.282, 38.906	-99.615, 31.496
p4	-117.676, 38.121	-100.183, 30.706
p5	-118.065, 37.334	-100.740, 29.913

^aEach profile is 1800 km long.

temperature thermochronology data on rock cooling [House *et al.*, 2003; Kelley *et al.*, 1992; Roy *et al.*, 2004] show that flexural responses to extension in the Rio Grande rift can account for locally elevated rift-flanks adjacent to the basins of the central and southern Rio Grande rift, however, extension-related uplift cannot explain the regionally high elevations (Alvarado Ridge) surrounding the Rio Grande rift corridor [Roy *et al.*, 1999, 2004].

[6] At the western margin of the Colorado Plateau, the transition from the highly extended Basin and Range province to the Colorado Plateau is accompanied by thickening of the crust by 5–10 km, suggesting that the crust provides at least some buoyancy required for isostatic support of the plateau [Gilbert and Sheehan, 2004] (Figure 1). At the eastern margin of the Colorado Plateau adjacent to the Rio Grande rift, plateau elevations reach a maximum, without additional thickening of the crust [Wilson *et al.*, 2005] (Figure 1). Observed geoid to topography ratios suggest that the plateau is compensated near a depth of 50 km, implying that the upper mantle plays a minor role in supporting high elevation [Chase *et al.*, 2002]. However, seismic and xenolith studies have observed that the Colorado Plateau crust is more mafic in composition than adjacent provinces [Condie and Selverstone, 1999; Nelson and Harris, 2001; Parsons *et al.*, 1996; Zandt *et al.*, 1995], relatively cold [Goes and van der Lee, 2002; Thompson and Zoback, 1979], and not thick enough to support all of the high topography [Gilbert and Sheehan, 2004; Lastowka *et al.*, 2001]. Several workers also note the poor correlation between crustal thickness, topography, and smoothed free air anomaly values, suggesting that the crust is not the major source of buoyancy [Gilbert and Sheehan, 2004; Keller *et al.*, 1979; Sheehan *et al.*, 1995].

[7] The observations above have prompted our study of regional variations in average upper man-

tle density across the Colorado Plateau, Rio Grande rift, and westernmost Great Plains. Our interdisciplinary approach, combining gravity, xenolith, and seismic wave speed data provides a robust interpretation of upper mantle density provinces, although we recognize that a fundamental uncertainty is the depth-distribution of density. We confine our interpretations, therefore, to variations in average upper mantle density, with the primary goal of determining if the upper mantle can be considered a significant source of buoyancy in the Rio Grande rift region.

3. Gravity Modeling Methods

[8] Over 350,000 irregularly spaced regional Bouguer gravity points were obtained from the Pan-American Center for Earth and Environmental Studies (PACES) data repository, at the University of Texas at El Paso. The data are a compilation of disparate Bouguer gravity data sets (legacy data using older surveying techniques and newer collections using survey-grade GPS), with average errors for the compilation ranging from 0.05 to 2 mgal (R. Aldouri, personal communication, 2005). The data are gridded to a 0.01 degree spacing (~ 1 km) by Delauney triangulation and extracted every five kilometers along five parallel profiles, trending 118° (Figure 1) using the Generic Mapping Tools (GMT) program [Wessel and Smith, 1991]. Each profile is 1800 km in length and samples gravity from the Basin and Range, Colorado Plateau, Rio Grande rift, and Great Plains provinces (Table 1). Profile orientations are chosen to be perpendicular to the trend of the gravity low at the eastern Colorado Plateau margin (Figure 1). Ungridded data from 5 km-wide swaths were also projected orthogonally onto the profile lines and compared to the gridded data; these are found to match within 1.3% for points ± 2 km perpendicular from the profile lines. The following analysis therefore is restricted to the gridded Bouguer gravity data.

[9] We use the observed Bouguer gravity anomaly curves extracted from the gridded regional data for forward models based on assumed subsurface density variations. Calculations follow the Talwani method, with subsequent revisions by Cady [Cady, 1980; Talwani *et al.*, 1959]. Input parameters for the calculation include (1) vertices of uniform-density polygonal bodies in the vertical plane below the profile, (2) density contrasts of modeled bodies with respect to some chosen background

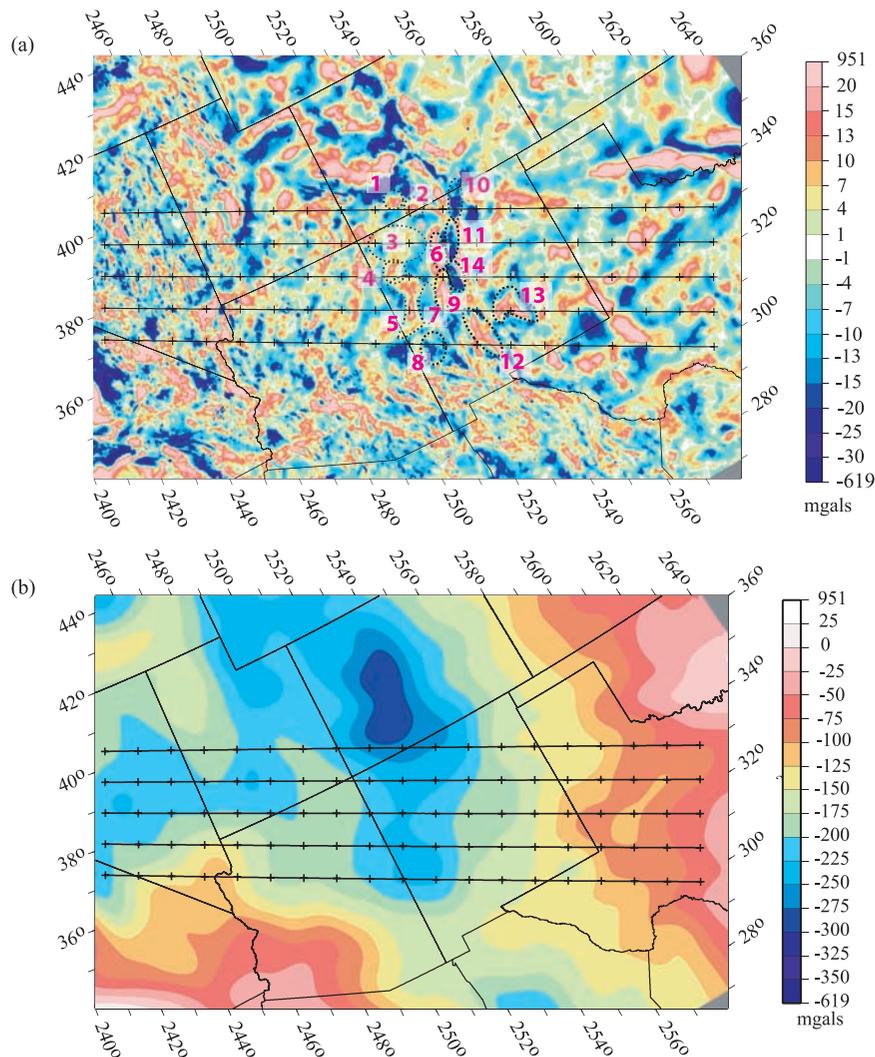


Figure 2. (a) High-pass filtered Bouguer gravity data highlighting features less than 100 km in wavelength. Significant features modeled in this study are in the dashed circles and numbered corresponding to descriptions in Table 2. (b) Low-pass filtered Bouguer gravity, retaining wavelengths >100 km, showing the prominent low at the eastern margin of the Colorado Plateau of amplitude ~ -200 mgal over ~ 400 km.

density, and (3) the in/out strike distance of the modeled bodies. The background density model we use varies as a function of depth: upper crustal background (0–15 km) is 2700 kg/m^3 ; middle crustal background (15–30 km) is 2850 kg/m^3 ; lower crustal background (30 km – Moho) is 3000 kg/m^3 ; and the uppermost mantle background (below Moho) is 3300 kg/m^3 . We note that the gravity models can only resolve density contrasts and our results are independent of the background density model.

[10] The shape of each observed gravity curve is treated as the sum of four contributions (in order of decreasing wavelength): (1) a decreasing linear

regional trend from higher gravity in stable, interior North America to lower values in the tectonically active western US; (2) contributions from upper mantle density variations, (3) contributions from crustal thickness variations; and (4) shortest wavelength contributions from mid to upper crustal features. Lower crustal features are not included here, however, in the Appendix we explore whether upper and lower crustal density variations alone can explain the long-wavelength gravity data. This exercise shows that unreasonable lower crustal density contrasts of $200\text{--}850 \text{ kg/m}^3$ across $200\text{--}400$ km wide provinces are required to explain the data, supporting our focus on the upper-mantle origin of the long-wavelength Bouguer anomalies.

Table 2. Upper Crustal Features Modeled in Gravity Profiles^a

	Description/Name	Data Type and Source	Rock Types/Densities Used
1	High: adjacent to Piedra River, CO	unconstrained	These features were modeled as high-density (2800 kg/m ³) intrusive bodies of limited extent (<50 km) and ~3 km depth; they contribute <25 mgal to the gravity.
2	High: near Chromo, CO; modeled as mafic intrusive	unconstrained	
3	Low: San Juan Basin	geology [Woodward and Callender, 1977]	Modeled as Pz to Mz sediments at ~2600 kg/m ³ , in the upper 0–3 km.
4	High: near Zuni Mts., NM	unconstrained	Modeled as a mafic body of limited lateral dimension (10 km) at shallow depth (0.5 km) at 3000 kg/m ³ with low-density felsic volcanics/seds. 2550 kg/m ³ in the surrounding region.
5	High: Zuni Mts. and Low: El Malpais, NM	gravity model [Ander and Huestis, 1982]	Modeled as a mafic intrusive body of limited lateral dimension (5–10 km) at shallow depth (2–4 km) at 2900 kg/m ³ .
6	High: Nacimiento Mts.	geology [Woodward and Callender, 1977]	Low-density felsic volcanics and sediments modeled at 2350–2400 kg/m ³ in the upper 2–4 km.
	Low: Jemez Mts.	gravity model [Nowell, 1996]	
7	Low: adjacent to El Malpais, NM	gravity model [Kelley and Reynolds, 1989]	Modeled as Mz-Cz sediments in a 5 km-wide region with density 2400 kg/m ³ in the upper 1 km.
8	Low: Mogollon-Datil Volcanic Field	gravity model [Schneider and Keller, 1994]	Modeled as a 2650 kg/m ³ granitic intrusion in the upper 10 km.
9	High: Lucero Uplift and Sierra Ladrões	geology and seismic reflection [Russell and Snelson, 1994]	This feature falls between profiles 3 and 4 and does not contribute significantly to either.
10	Low: southern San Luis Basin	geology [Lipman and Mehnert, 1979]	Modeled as Mz to Cz sediments, ~2550 kg/m ³ in the upper 5 km
11	Low: Espanola Basin and Sangre de Cristo Mountains	geology [Kelley, 1978]	Modeled as Mz to Cz sediments, ~2350–2400 kg/m ³ in the upper 0–2 km.
12	High: San Andres Mts.	gravity and seismic refraction [Adams and Keller, 1994]	Layered mafic intrusion modeled at 2800 kg/m ³ in upper 6 km.
13	High: Sierra Blanca	geology [Moore et al., 1991; Elston and Snider, 1964]	Modeled as a layered mafic intrusive, from 2 to 6 km depth with density 2800 kg/m ³ .
14	High: Mafic Intrusive Low: Albuquerque Basin	gravity [Bowsher, 1991] geology and seismic reflection [Russell and Snelson, 1994]; gravity model of Peterson and Roy [2005] geology and seismic refraction [Roberts et al., 1994]	Modeled as a suite of Pz, Mz, and Cz sediments, ~2600–2650 kg/m ³ in the upper 0–9 km.

^aDensities are relative to an upper crustal background of 2700 kg/m³. Numbers correspond to numbers in Figure 2a.

We describe below our analysis of gravity contributions from upper crustal, Moho, and upper mantle features.

[11] To model gravity contributions from the upper crust, we start by identifying major short wavelength gravity features along our profiles using a high-pass 2D cosine arch to filter the gridded

Bouguer data, suppressing wavelengths greater than 100 km (Figure 2a). The resulting high-pass map is overlain on bedrock geologic maps to identify gravity features that are likely geologically controlled by upper crustal density variations. Within the region of the long-wavelength low-gravity anomaly of interest (Figures 1 and 2b), we identify fourteen major features for modeling

(Table 2; Figure 2a). Upper crustal models are used to fit to the gravity features identified in our high-pass map (Figure 2a) after isolating the feature by detrending with a linear best fit line. For the short-wavelength gravity features of interest that correlate with mapped geologic structures, models are constructed using published geometries and densities where available (Table 2). Short wavelength gravity features that do not correlate well with bedrock geology or are apparently not previously investigated are modeled as due to unconstrained upper crustal density variations. Upper crustal contributions are ignored outside of the region of broad gravity low at the eastern margin of the Colorado Plateau region (Figure 2b). This is not due to a lack of geologic constraints, but rather because our focused upper crustal gravity models include some rather prominent upper crustal features (the Albuquerque and San Juan basins, the Mogollon-Datil volcanic field, the Jemez caldera; Figure 2a) and show that they contribute only minor amplitude gravity signals ($<40\text{--}50$ mgal) over short (<50 km) wavelengths. The major, long-wavelength gravity feature we are most interested in has an amplitude of -100 to -150 mgal over ~ 400 to 600 km, and hence detailed, exhaustive upper crustal modeling is unwarranted when focusing on this signal. For completeness and to display the minor contribution of the upper crust, the results discussed below do include the effects of the key upper crustal features identified in Figure 2a.

[12] Crustal thickness is determined by combining three data sources and gridding them to approximately 10 km spacing (Figure 1). The LA RISTRA passive teleseismic array provides high-resolution crustal thickness from receiver functions for a swath from southeastern New Mexico to southeastern Utah [Wilson *et al.*, 2005]. Crustal thickness for much of Colorado, Utah, Nevada, and Wyoming are obtained from the compilation by Gilbert and Sheehan [Gilbert and Sheehan, 2004], which is at a lower resolution than the LA RISTRA data. Crustal thickness for the remaining regions in the study area is obtained from the Crust 2.0 2×2 degree crustal model (G. Laske, REM Web page, <http://mahi.ucsd.edu/Gabi/rem.dir/rem.home.html>, 2005). These Moho depths are subject to disparate errors, depending on the resolution of the seismic experiment and analysis. The error estimates range from ± 2 km for the LA RISTRA data set (D. Wilson, personal communication, 2005) to ± 5 km for the Crust 2.0 compilation.

[13] After the contribution of the Moho is removed, a linear trend representing a continent-scale transition from the stable, cratonic interior of North America to the tectonically active western US is removed from the data. This trend, equivalent to a linearly changing background model density in the upper mantle, is removed because it represents a contribution from continent-scale density variations that are not the focus of this study.

[14] Following removal of the contributions above, the remaining Bouguer anomaly observed in each profile is attributed to variations in upper mantle densities. Although seismically imaged upper mantle heterogeneity extends to depths of ~ 200 km beneath the Southwestern US [Dueker *et al.*, 2001; Gao *et al.*, 2004] and some published estimates of lithospheric thickness beneath the Colorado Plateau and Great Plains exceed 100 km [West *et al.*, 2004], we find that when the bodies extend to 200 km depth density contrasts required in the upper mantle are reduced to $\leq \pm 10$ kg/m³ (or $<0.5\%$, below the uncertainties in density estimates from supporting data, e.g., xenoliths and seismic wave speed scaling relations). In our first-order analysis, therefore, we keep density contrasts larger to highlight the heterogeneity and constrain our bodies to ≤ 100 km depth. We recognize the fundamental tradeoff between density contrast and depth-extent in gravity models and note this as a source of uncertainty in our results. Upper mantle density contrasts are modeled with simple vertical sides and large in/out strike distances, typically 3500 km, to minimize edge effects. Crustal features for which it would be inappropriate to assign large in/out strike distances, such as volcanic edifices or batholiths, are assigned distances corresponding to the size of the feature.

4. Gravity Modeling Results

[15] Here we present the gravity models along profiles 1–5, including features in the upper crust that were included in the model for each profile, Moho variations, and the upper mantle density bodies required.

[16] First, we note that on all of the profiles, crustal thickness variations produce a poor fit to the observed long-wavelength Bouguer gravity (Figure 3). A persistent asymmetry is observed in the predicted gravity from crustal thickness variations, where predicted gravity is consistently

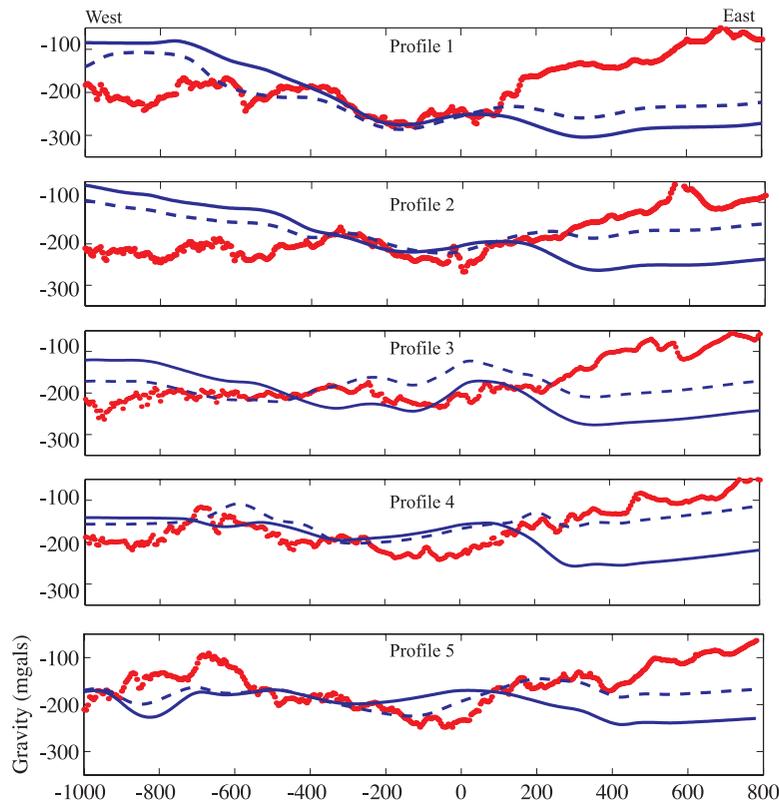


Figure 3. Observed gravity (red dots) and predicted gravity (solid blue lines) from crustal thickness variations along profiles 1–5. The dashed lines show the predicted gravity for the case where the Moho is allowed to vary within the error bounds discussed in the text so as to minimize the misfit to the observed gravity.

higher than observed in the west, and lower than observed in the east (Figure 3). (One exception is the southernmost profile (5) where the predicted gravity in the west is lower than observed.) To investigate the effects of uncertainties in the crustal thickness from the various data sets (the LA RISTRA data have the lowest uncertainties of ± 2 km and the Crust 2.0 compilation has uncertainties of ± 5 km), we considered a range of crustal thickness models that all lie within the error bounds. None of these models can match the observed long-wavelength gravity variations and the asymmetry in the residual persists, with consistently higher-than-observed gravity to the west and lower-than-observed to the east. As an example, we consider the case where the crustal thickness in each profile is allowed to be a minimum in the east and maximum in the west (the Moho is linearly adjusted within error bounds as a function of distance from the center of the profile) to minimize the regional misfit asymmetry (dashed lines in Figure 3). This exercise shows that even when Moho depths are adjusted within error bounds to best fit the data, crustal thickness

is a poor predictor of the Bouguer anomaly. The mismatch between gravity predicted by crustal thickness variations and the observed Bouguer anomaly becomes more positive from west to east, with undulations roughly 100–400 km in wavelength (Figure 4). We interpret this long wavelength positive-to-the-east trend to be a continental-scale signature arising from the high gravity of cratonic North America on the east and the low gravity of the western United States on the west. As discussed above, this regional trend is approximated by a linear best fit line (Figure 4) and removed. In our de-trending, we allow the slope of the linear fit to vary among the profiles, and we find a systematic shallowing of this slope from north to south. We suggest that this is because the wavelength of the continental-scale transition in upper mantle physical properties likely increases from north to south, with varying tectonic setting. This longest-wavelength character in the Bouguer gravity is also reflected in continental-scale V_s variations in the upper mantle and was recognized by Kane and Godson [1989] as likely representing effects due to long-

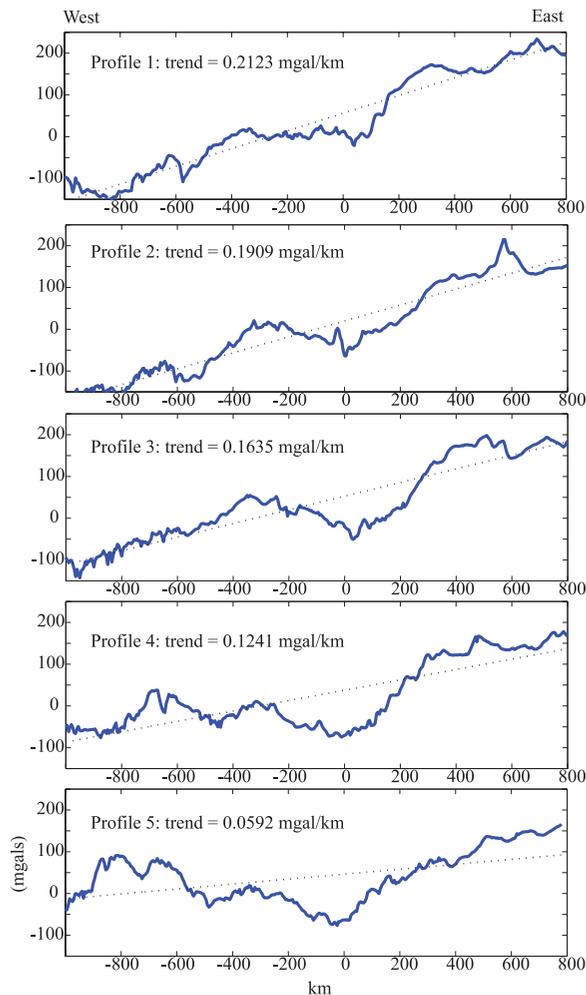


Figure 4. Observed minus predicted gravity residual from Figure 3. The trend in the residual is approximated by a best fit line (dashed) and removed before modeling upper mantle contributions. The 200–400 km undulations in the residual determine the edges of upper mantle provinces in Figure 5.

wavelength density and structural changes in the upper mantle beneath North America. In our modeling, removal of this trend is equivalent to removing the effects of a slowly changing background density and normalizing to a uniform background in the upper mantle. As noted above, the gravity models are sensitive to density contrasts and not absolute density, so our results are independent of the assumed background density.

[17] The 100–400 km undulations in the residual gravity (Figure 4) roughly correspond to tectonic provinces in the region of interest, and are the basis for defining the edges of our modeled upper mantle bodies in Figure 5. The models

in Figure 5 contain the (minor) effects of the upper crustal features identified in Figure 2a (sources in Table 2) and upper mantle density variations (see Appendix for discussion of lower crustal density variations). Three regions of distinct upper mantle densities are observed: high-density upper mantle beneath the western Great Plains, low density beneath the Rio Grande rift and eastern Colorado Plateau margin and, in the three northern profiles, high density in the central Colorado Plateau (Figure 5). The high-density region at the western Great Plains has a model density contrast of $\Delta\rho = +20\text{--}40 \text{ kg/m}^3$ and in the three northern profiles (1–3) our modeling constrains the eastern extent of this region (approximately 100–300 km wide). In the southern two profiles (4 and 5) the high-density bodies extend further east into the Great Plains province and we cannot constrain their eastern edges. Considering the variable resolution and sources of crustal thickness data which may affect the placement of upper mantle bodies, we note that on profiles 4 and 5 the eastern high-density bodies are well-constrained on the west by the intersection of these profiles with the high-resolution LA RISTRA seismic line. However, the persistence of the region of high density in profiles 1–3, where models are not as strongly controlled by the LA RISTRA data, supports the existence of the feature regionally beneath the western Great Plains province.

[18] The region of low density along the Rio Grande rift and eastern margin of the Colorado Plateau ranges in density contrast from $\Delta\rho = -10$ to -50 kg/m^3 , generally narrowing from a broad region of subdued density contrast in the north into a more focused zone of increasing density contrast in the south. After the removal of Moho contributions, upper crustal features, and the long-wavelength regional trend, the gravity anomaly attributed to the upper mantle is as much as -75 mgal and 440 km wide in the north, and -120 to -150 mgal and 200 km wide in the south (Figure 5). The entire low-density upper mantle province underlies the Rio Grande rift, the eastern margin of the Colorado Plateau, and parts of southern Rocky Mountains and western Great Plains.

[19] The western high-density region modeled in profiles 1–3 is within the central Colorado Plateau. Model density contrasts are $\Delta\rho = 20$ to 30 kg/m^3 , and the region is 225 km to 260 km wide along the profiles. We constrain the north-south extent of this area of high density to at least 200 km (as it does

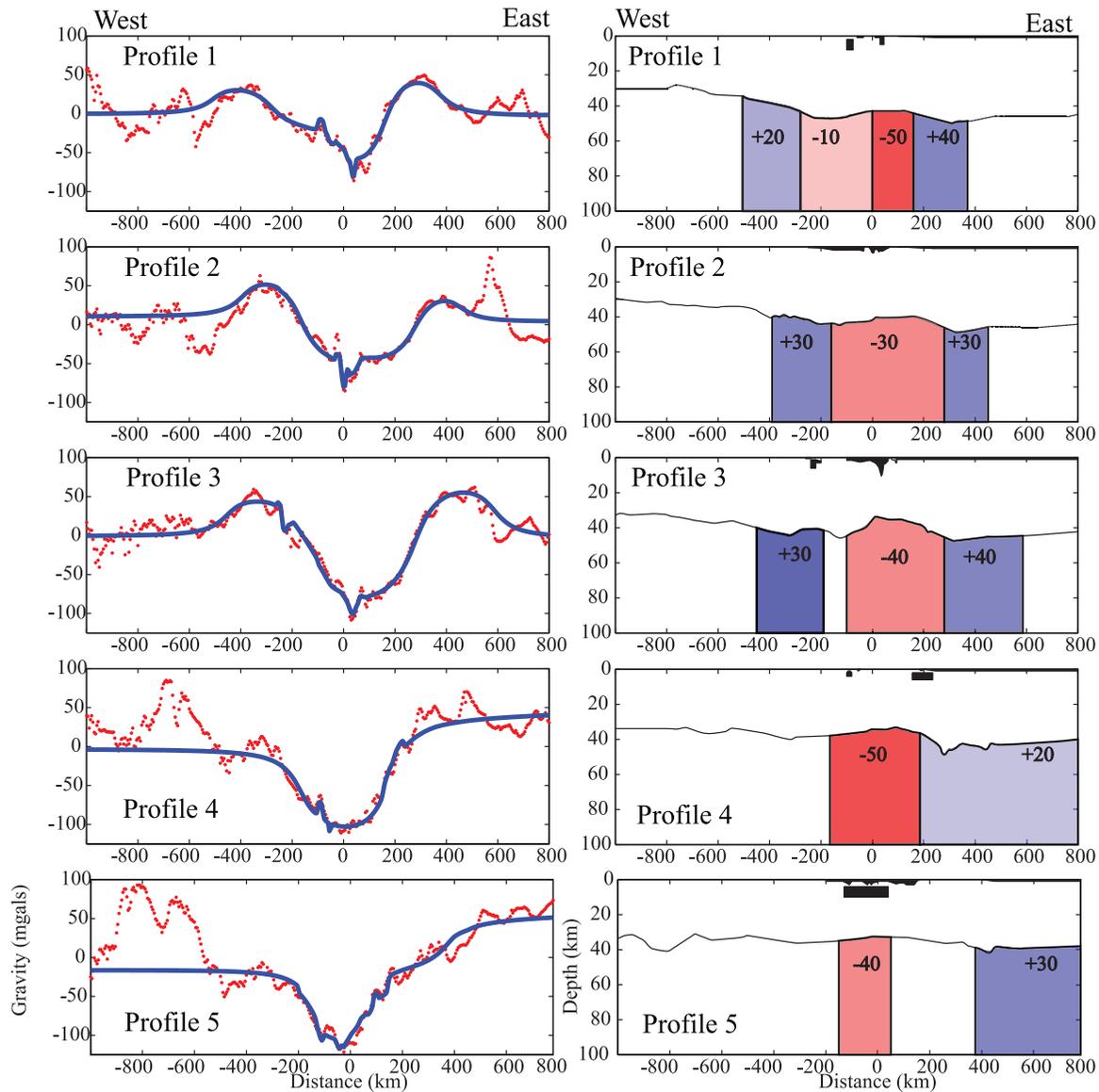


Figure 5. Left panels: Predicted gravity (solid blue lines) compared to “corrected gravity” (red dots) representing observed gravity minus the contributions from crustal thickness variations (blue line in Figure 3) and regional trend (dashed line in Figure 4) along profiles 1–5. Right panels: Density contrasts in kg/m^3 (numbers) and locations of bodies used to derive the predicted gravity in the left panels including upper crustal and upper mantle bodies. Upper crustal models are as summarized in Table 2, but generally contribute <-40 to -50 mgal amplitude over short (<50 km) wavelengths.

not appear in the two southern profiles) and its east-west dimension is 200–300 km.

5. Discussion

[20] The density contrasts chosen for the upper mantle bodies beneath the Colorado Plateau, Rio Grande Rift, and western Great Plains in this study are allowed to vary in such a way as to best fit the Bouguer gravity data, and are thus unconstrained by other data sets. In order to interpret the physical

character and dimensions of the upper mantle provinces we have identified, we explore the consistency of our upper mantle density variations with inferences based on mantle xenoliths and on numerous seismic surveys in the eastern Colorado Plateau–Rio Grande rift region.

5.1. Consistency With Upper Mantle Properties Inferred From Xenoliths

[21] We choose upper mantle xenoliths from three regions of Oligocene or younger volcanism: the

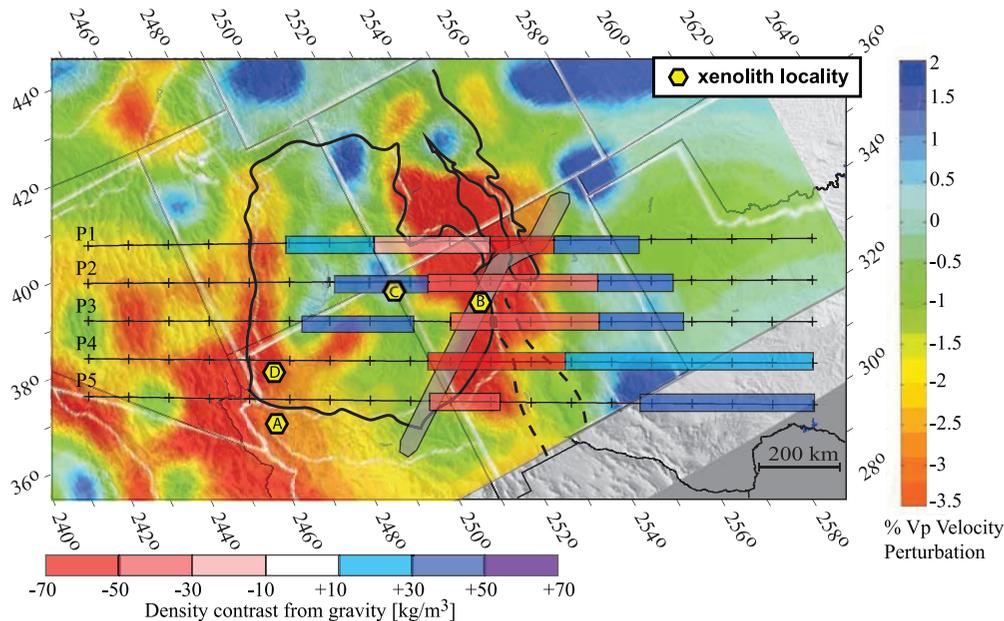


Figure 6. Modeled density heterogeneity at 100 km depth based on gravity (colored bars represent upper mantle density bodies as in Figure 5) overlain on topography and seismic wave speeds at 100 km depth [Dueker *et al.*, 2001]. Areas along profiles with no bars are at background density of 3300 kg/m³. Tectonic province boundaries (solid lines) are indicated for reference, and letters correspond to xenolith localities in Table 3.

western Colorado Plateau [Alibert, 1994; Smith, 2000], the central Colorado Plateau [Alibert, 1994; Lee *et al.*, 2001; Smith, 2000], and the eastern Colorado Plateau–Rio Grande rift transition zone (Puerco Necks) [Porreca and Selverstone, 2004] (Figure 6; Table 3). The relative paucity of xenoliths from the Great Plains province limits our interpretations in this region. Host rock ages of the selected xenoliths are ≤ 1 Ma in the western Colorado Plateau [Alibert, 1994; Riter and Smith, 1996], 20–30 Ma in the central Colorado Plateau [Alibert, 1994], and ≤ 5 Ma in the eastern Colorado Plateau [Hallett *et al.*, 1997], and available xenolith data include some combination of rock name, modal mineralogy, mineral compositions, and overall descriptions of entire xenolith populations. We recognize that the relatively older host rock ages in the central Colorado Plateau will limit our interpretations of present-day variations in upper mantle physical properties in this region.

[22] The distribution and composition of pyroxenite versus lherzolite xenoliths varies among the localities in our study area and also within a given locality. In general, we note the relative lack of pyroxenite xenoliths from the central Colorado Plateau, whereas pyroxenite xenoliths are abundant at the margins of the plateau at the transition zones

to extensional provinces, the Basin and Range and Rio Grande rift [Wilshire *et al.*, 1988]. Conversely, we note that garnet lherzolite xenoliths are only found in the center of the plateau and are lacking at the plateau margins. For example, in the eastern Colorado Plateau some of volcanic necks within the Rio Puerco volcanic field contain dominantly pyroxenite xenoliths (including garnet websterites), as opposed to the lherzolite-dominated populations that are common elsewhere, suggesting that overall, pyroxenites may compose as much as 15–20% of the upper mantle beneath the Colorado Plateau–Rio Grande rift transition zone [Selverstone *et al.*, 2004]. Using the distribution of xenolith populations as a guide, we estimate the average in situ density of upper mantle columns at the center of the Colorado Plateau and at its western and eastern margins.

[23] Our calculations follow the Hacker and Abers [2004] method of estimating in situ physical properties based on mineral compositions of rocks, where inputs for the calculation are (1) modal mineralogy with selected compositions and (2) equilibrium temperature and pressure. Where only rock type data are available, reasonable modal mineralogy is estimated on the basis of the International Union of Geological Sciences classification system for ultramafic rocks, as

follows: olivine is Fo_{90} , garnet is pyrope, orthopyroxene is enstatite $_{90}$ and ferrosilite $_{10}$, clinopyroxene is diopside $_{90}$ and hedenbergite $_{10}$, and spinel is MgAl spinel. Mineral modifiers are assigned 1–9% modal concentration. Where only modal mineralogy is available, reasonable compositions are similarly estimated. We start by calculating the properties for individual rock types in a particular locality (e.g., a population of lherzolites and pyroxenites using published equilibration temperatures and pressures where available), followed by average properties of mantle columns for various pyroxenite-lherzolite proportions. The density calculations are sensitive to the range of temperature and pressure at depth as well as the chemical composition of the estimated mineralogy, so these parameters are varied within published ranges or within those ranges presented above to constrain possible densities for each sample (example calculations are shown in Table 3, but this is not a complete listing of available data). The densities depend strongly on the modal concentration of fayalite, diopside and spinel, and are most sensitive to fayalite because of its iron content.

[24] To calculate densities for average upper mantle columns, we choose varying proportions of lherzolites (70–90%) and pyroxenites (10–30%) to represent the average upper mantle at the margins of the plateau. At the center of the plateau, the average upper mantle column is taken to be 100% lherzolite, with roughly equal proportions of garnet and spinel lherzolite (assuming a garnet stability depth of 70 km, Moho depth of 40 km, and a 100 km lithosphere column). Average upper mantle densities calculated in this fashion range from 3300 to 3325 kg/m^3 for the central plateau and 3270 to 3300 kg/m^3 at the margins of the plateau; densities at the eastern margin tend to be somewhat lower than those at the western margin, driven mainly by the low-density pyroxenites (e.g., 3250 to 3280 kg/m^3 in Table 3). The sign and amplitude of the variation in upper mantle density from xenoliths across the Colorado Plateau ($\Delta\rho = -50 \text{ kg/m}^3$) is consistent with our findings from gravity models, although we interpret the densities with caution as these are less sensitive to varying composition than other physical properties such as seismic wave speeds (for example, in the Puerco necks the calculated density difference between spinel lherzolite and a garnet websterite at 1.5 GPa and 1000°C is $\sim 1.4\%$, whereas seismic wave speeds are predicted to vary by as much as 3–4%; Table 3). Seismic

wave speeds calculated on the basis of xenolith compositions are generally lower for regions where pyroxenites account for >10% of the upper mantle composition, e.g., the margins of the plateau, and higher for the central, lherzolite dominated region (Table 3). This pattern of high seismic wave speeds for the central Colorado Plateau surrounded by regions of lower wave speeds is consistent with observations from seismic tomography [Dueker *et al.*, 2001; Humphreys *et al.*, 2003].

5.2. Comparison to Seismic Wave Speed Structure

[25] We now compare the low-density upper mantle province identified beneath the Rio Grande rift and easternmost Colorado Plateau from gravity and xenolith data above with seismic wave speed structure derived from a regional inversion [Dueker *et al.*, 2001]. Our upper mantle density models at 100 km depth are overlain on a tomographic slice at the same depth to highlight the coincident spatial distribution of density and wave speed anomalies (Figure 6). The low seismic wave speed region, $\Delta V_p = -2.0$ to -3.5% , imaged beneath parts of the southern Rocky Mountains, Rio Grande rift, and eastern Colorado Plateau agrees with the size and shape of the low-density upper mantle region in our gravity models. A low-wave speed feature of similar shape and magnitude is also imaged in P and S-wave tomography by [Humphreys *et al.*, 2003]. Both the low wave speed and the low-density regions are broader in the north and narrow in the south (unlike the upper crustal expression of the Rio Grande rift), and trend roughly parallel to the southeastern margin of the Colorado Plateau and not the upper crustal trend of the Rio Grande rift.

[26] The Rio Grande rift region is a zone of high heat flow and middle to late Tertiary volcanic activity [Keller and Baldrige, 1999; Olsen *et al.*, 1987; Thompson and Zoback, 1979], so we first investigate the implications of the low-density upper mantle province for the temperature structure of the mantle. Using a linear relationship between density and temperature (and a bulk coefficient of expansion for upper mantle $\alpha = -2.5 \times 10^{-5} \text{ K}^{-1}$), the temperature change required to reduce upper mantle density from a reference value of 3300 kg/m^3 by the density contrast $\Delta\rho = -10$ to -50 kg/m^3 obtained from gravity models and xenolith data is $\Delta T \sim +121$ to 606 K. We note that if the upper mantle density

reduction in the Rio Grande rift corridor is attributed to thermal differences alone, the predicted temperature variation based on linear scaling from both xenolith data and from gravity models shows a greater range than temperature variations derived from seismic tomography at 100–200 km in the upper mantle, $\Delta T = 300\text{--}470$ K [Gao *et al.*, 2004; Goes and van der Lee, 2002]. We note that part of this discrepancy may be due to the breakdown in linear temperature-density scaling over such large changes in temperature, although in this area of high heat flow and Tertiary volcanism, there may also be effects from variations in composition and partial melt content.

[27] A simple calculation shows that if partial melt is present in situ beneath the eastern margin of the Colorado Plateau–Rio Grande rift and we attribute all of the observed upper mantle density variations to the presence of partial melt, we expect a 2–3% reduction in V_p for melt fractions ranging from 0.4 to 3% using the scaling relations of Hammond and Humphreys [2000]. These predicted melt fractions are reasonable for regions of intraplate volcanism, suggesting that partial melt may play a role in reduced upper mantle densities beneath regions of active magmatism such as the Jemez lineament and parts of the Rio Grande rift (e.g., the Socorro magma body). If we assume that both temperature variation and, say, a 1% in situ partial melt fraction is responsible for the observed density reduction beneath the Jemez lineament, then the required upper mantle temperature variations need only be 160–320 K (M. B. Magnani *et al.*, Seismic and geomorphic evidence for rejuvenation of topography in the Southern Rocky Mountains, manuscript in preparation, 2005), closer to the values inferred from seismic tomography [Gao *et al.*, 2004; Goes and van der Lee, 2002].

[28] If we scale our density contrast model to seismic wave speed variation using the empirical upper mantle wave speed-density scaling relation of Humphreys and Dueker [1994], the predicted V_p perturbation in this region is a reduction by 2.2–6.7%, encompassing the range of variations imaged by regional tomography [Dueker *et al.*, 2001; Humphreys *et al.*, 2003]. We note that this scaling is based on empirical observations for velocity-temperature derivatives [e.g., Anderson and Bass, 1984] and linear expansion, assuming a reference upper mantle velocity of 8 km/s. This is a first-order estimate that ignores the effects of basalt removal or in situ partial melt [Humphreys and Dueker, 1994] and the expected range in V_p vari-

ation is therefore limited by the assumption of purely thermal control on physical properties.

5.3. Dynamic Significance of Low-Density Upper Mantle and Relation to Prior Tectonism

[29] As noted above, the low-density upper mantle province beneath the Rio Grande rift and easternmost Colorado Plateau is likely controlled by upper mantle temperature variations with contributions from partial melt content and inherent compositional variations. Significant Tertiary magmatic events in the region that may have modified both upper mantle composition and partial melt content include the middle Tertiary ignimbrite flare-up and the late Tertiary magmatism in the Jemez Lineament and Rio Grande rift. Of these, the middle Tertiary ignimbrite flare-up likely involved the eruption of a regional-scale zone of caldera-complexes extending from the Mogollon-Datil to the Taos-Latir and San Juan volcanic fields [Lipman *et al.*, 1972; Steven, 1975], and generated significant volumes of bimodal volcanism sustained by protracted basalt extraction from the mantle [Johnson and Thompson, 1991; McMillan *et al.*, 2000]. Removal of significant fractions of basaltic melt from mantle peridotite leaves a lower-density residuum with a higher olivine Mg# [Dawson, 1987; Jordan, 1979; Lee, 2003], which likely contributes to the regional upper mantle density reduction [Johnson, 1991; McCoy *et al.*, 2004; Roy *et al.*, 2004]. For example, 1–5% basaltic melt extraction over roughly a 6×10^6 km³ volume (roughly a cubical region 150 km on a side) beneath the middle Tertiary San Juan volcanic field can reduce average upper mantle density by 1%, consistent with the long-wavelength low Bouguer gravity across the region [Roy *et al.*, 2004]. A similar modification of upper mantle density might be expected regionally along the eastern margin of the Colorado Plateau, along the zone of middle Tertiary ignimbrite centers including the Mogollon-Datil and Taos-Latir volcanic fields (Figure 1). Compositional modification due to voluminous basalt extraction is a process that is often overlooked when considering the upper mantle structure of the margins of the Colorado Plateau, and might explain the strong spatial correlation between the broad zone of low upper mantle density and the middle Tertiary magmatic centers at the eastern margin of the plateau (Figures 1 and 6). Interestingly, lherzolites from the Rio Puerco volcanic field at the transition between the Rio Grande rift and the eastern Colorado Plateau show evidence for

significant melt depletion in their elevated Mg# of 90.4–92 compared to a common value for upper mantle of Mg# = 88–90 [McDonough and Rudnick, 1998; Porreca and Selverstone, 2004; Smith, 2000].

[30] Late Tertiary and ongoing volcanism in the Rio Grande rift and Jemez Lineament, and associated elevated temperature and partial melt distributions, are almost certainly influencing present-day upper mantle temperatures and partial melt content. Regional reduction of residuum density following basalt removal is not a likely consequence of the late Tertiary magmatism in the Jemez Lineament and Rio Grande rift because of lower volumes of inferred upper mantle melt extraction during dominantly basaltic magmatism [Johnson and Thompson, 1991]. The pattern of late Tertiary magmatism at the eastern margin of the Colorado Plateau shows a distinct inward progression from the Rio Grande rift toward the present-day locus of active volcanism, the Jemez lineament, sweeping across the region of low Bouguer gravity anomaly over the last 5 Ma and perhaps contributing to modification of thermal and partial melt fields in the upper mantle as evidenced in the metasomatically altered upper mantle xenoliths entrained in the 0–5 Ma Rio Puerco volcanic field.

[31] The arguments above suggest that low upper mantle density and the coincident reduction in seismic wave speed at the eastern margin of the Colorado Plateau–Rio Grande rift are consistent with reasonable modifications of the physical state of North America lithosphere over a 200–300 km wide region. The relationship of this region of low upper mantle density to the orientation of and deformation in the Rio Grande rift is enigmatic because the zone of low density is oblique to and displaced from the rift (Figures 1, 2b, and 6). Magmatism within the Rio Grande rift shows a distinct progression toward asthenospheric sources following a ~10 m.y. relatively amagmatic interval between the middle Tertiary ignimbrite episode and the late Tertiary volcanism [Baldridge *et al.*, 1991; McMillan *et al.*, 2000; Perry *et al.*, 1998]. These observations are consistent with lithospheric thinning beneath the rift and indeed, the regional low-gravity feature has been interpreted as evidence for broad upwarping of the asthenosphere beneath the thinned lithosphere of the region [Cordell *et al.*, 1982; Slack *et al.*, 1996]. The LA RISTRA experiment imaged a symmetric zone of reduced V_s in the upper mantle along a profile that is oblique to

the Rio Grande rift [West *et al.*, 2004; Wilson *et al.*, 2005]. This feature has been interpreted to reflect an asthenospheric upwarp over a broader region of deep lithospheric thinning than is evident from the upper crustal extension in the Rio Grande rift [Wilson *et al.*, 2005]. If the zone of low upper mantle density (low Bouguer gravity) is interpreted as a zone of modified and/or thinned lithosphere, which is certainly consistent with petrologic, geochemical, and geophysical data, we would argue that this region is oblique to the upper crustal expression of the Rio Grande rift, and that kinematic linkages between upper mantle structure and upper crustal extension remain enigmatic.

6. Conclusions

[32] We forward model Bouguer gravity along five parallel profiles crossing the Colorado Plateau, Rio Grande rift, and western Great Plains. Crustal thickness variations and upper crustal features in the models are constrained by existing geological and geophysical data. We find that the long-wavelength Bouguer gravity variations across the Colorado Plateau require a region of low-density upper mantle straddling the Rio Grande rift and extending beneath the eastern margin of the Colorado Plateau. At the eastern Colorado Plateau, the low-density upper mantle region aligns with the trend of prominent middle Tertiary magmatic centers, including the Mogollon-Datil, Taos-Latir, and San Juan volcanic fields and encompasses the Jemez Lineament. This region is broad in the north and narrows to the south, consistent with regional seismic tomographic models, and inconsistent with along-strike variations in width of the Rio Grande rift extensional system. The upper mantle density variations beneath the Rio Grande rift corridor from our gravity models are consistent with densities inferred from xenolith data and from seismic wave speed structure. We explore temperature variations, partial melt content, and compositional variations that are consistent with the low-density zone in the upper mantle and interpret it to be due to a combination of modified and/or thinned lithosphere. Upper mantle xenolith populations from the eastern margin of the Colorado Plateau reflect this in metasomatic alteration and in the greater abundance of pyroxenites and lack of garnet lherzolites compared to the central Colorado Plateau. Upper mantle densities are higher beneath the central Colorado Plateau where upper mantle xenolith populations show a distinct lack of pyroxenites and the presence of garnet lherzolite,

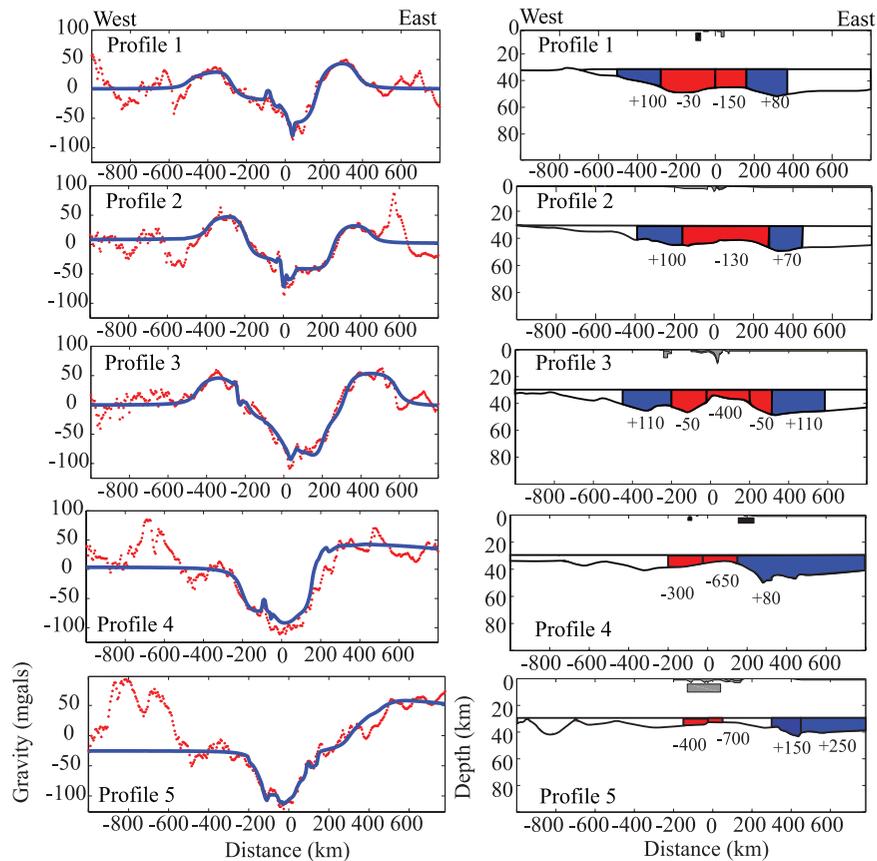


Figure A1. Left panels: Predicted gravity (solid blue lines) compared to “corrected gravity” (red dots) representing observed gravity minus the contributions from crustal thickness variations (blue line in Figure 3) and regional trend (dashed line in Figure 4) along profiles 1–5. Right panels: Density models used to derive the predicted gravity in left panels including crustal bodies (contrasts are relative to a background lower crustal density of 3000 kg/m^3). Upper crustal models are as summarized in Table 2.

supporting the idea that the plateau is internally heterogeneous.

Appendix A: Exploring Possible Lower Crustal Contributions

[33] To test the whether or not observed gravity requires density heterogeneity in the uppermost mantle, we consider an alternative model that forces density heterogeneity to be only in the crust. To explain the long-wavelength gravity variations, we model lower crustal bodies that are constrained to be between 30 km depth and the Moho with very large in/out strike lengths, as in the upper mantle case. The lower crustal bodies have less volume than the upper mantle bodies, and therefore need larger density perturbations in order to produce a gravity signature of the amplitude required by the observed gravity (Figure A1). Density contrasts modeled in the lower crust beneath the Rio Grande rift vary from $\Delta\rho = -700 \text{ kg/m}^3$ in profile 5 to $\Delta\rho =$

-130 kg/m^3 in profile 2, relative to an assumed background of 3000 kg/m^3 between 30 km and the Moho. Density contrasts modeled beneath the central Colorado Plateau are $+100$ to $+110 \text{ kg/m}^3$, and those beneath the western Great Plains are $+70$ to $+250 \text{ kg/m}^3$. We note that the zone of reduced lower crustal density is broader, with more subdued contrast in the north and localizes into a narrower region of sharper density contrast in the south.

[34] Forcing density heterogeneity into the crust produces a model that is inconsistent with seismic data. Very large lateral variations in lower crustal density, such as $\Delta\rho = 130$ to 850 kg/m^3 modeled between the Rio Grande rift and the western Great Plains in profiles 3 and 4, would produce correspondingly large lower crustal seismic wave speed gradients, which are not observed [Olsen *et al.*, 1987; Snelson *et al.*, 1998]. For example, the density contrast between the enigmatic high wave speed lower crust (the “7.xx” layer) imaged in the

CD-ROM experiment and normal lower crust was modeled as $\sim 150 \text{ kg/m}^3$ using both seismic and gravity constraints [Snelson et al., 2005]. (We note that the LA RISTRA experiment observed a 200 km wide region in the crust beneath the Rio Grande rift that has slightly reduce shear wave speed (by 0.2 km/s) relative to the surroundings, which may reflect the presence of partial melt in the crust [Wilson et al., 2005]). The large density contrasts required within the crust from gravity models would generally require very large changes in rock lithology. For example, at 25 km depth, the density contrast between andesite and mafic eclogite is roughly 850 kg/m^3 [Christensen and Mooney, 1995]. (Note that near surface density contrasts in the upper crust rarely exceed $600\text{--}700 \text{ kg/m}^3$, the density contrast between unconsolidated sediments and crystalline or intrusive igneous rocks [Christensen and Mooney, 1995].) Given the subdued nature of the required lower crustal density contrasts in the northern profile, we would argue that although lower crustal buoyancy variations must contribute to the Bouguer gravity field, for most of our central and southern profiles we can attribute much of the long-wavelength variations in gravity to upper mantle density contrasts.

Acknowledgments

[35] The authors are grateful to P. vanKeken, S. Goes, and an anonymous reviewer for comments and suggestions that significantly improved this paper. We also thank D. Wilson and R. Aster for sharing the LA RISTRA crustal thickness data, H. Gilbert for sharing the crustal thickness compilation for the Colorado Plateau, S. Kelley and J. Pederson for discussions on the evolution of the Colorado Plateau, C. Porreca for discussions on the Rio Puerco xenoliths, C. Peterson for help with GIS files, G. R. Keller for access to the Talwani2.5 program and discussions on gravity modeling, and V. E. Langenheim for comments on a poster of an early version of this work. This work was partially funded by NSF grant EAR0408513 to M.R.

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