Geophysical investigations of the origins and effects of density variations in the crust and upper mantle beneath the western and central United States

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A thesis submitted to the Faculty of the Graduate School of the University of Colorado in partial fulfillment of the requirements for the degree of Doctor of Philosophy in Geophysics Department of Geology 2014 This thesis entitled: Geophysical investigations of the origins and effects of density variations in the crust and upper mantle beneath the western and central United States

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The final copy of this thesis has been examined by the signatories, and we find that both the content and the form meet acceptable presentation standards of scholarly work in the above mentioned discipline.

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Thesis directed by Prof. Craig Jones

Abstract

Variations in density within the earth are the dominant cause of both surface topography--generating mountains, valleys, and plateaux--and convection, leading to plate tectonics. Density varies as a function of chemistry, mineralogy and temperature, containing information about physical state and history. I develop methods to estimate the density of the crust and upper mantle from seismic, gravity, heat flow, and topographic data. Decomposing density variations into thermal and compositional components provides insight into the origin of topography, tectonic history, and active processes. These techniques are applied to the inspiring landscapes of the western United States. The modern density structure of the Sierra Nevada, California, suggests that post-Miocene range uplift occurred in response to removal of dense mantle lithosphere. A numerical model of the flexural response of the surface to mantle loads shows that this material is likely now found in the upper mantle just west of the range, where it has created the Tulare basin. A broader density model of the entire western U.S. highlights a dichotomy in upper mantle buoyancy between the low-relief Great Plains and regions modified in the Cenozoic. Relief within the Cordillera is generated by varying degrees of crustal thermal and compositional buoyancy. A targeted thermal modeling study of the Colorado Plateau shows that ~2 km of Cenozoic uplift--in the absence of crustal shortening--can be ascribed to removal of tens of km of mantle lithosphere and related hydration of the lower crust. Overall, these four studies highlight the utility of density as a window into tectonic processes and a record of lithospheric history.

Dedication

To Cecie--for almost half my life's worth of laughter, inspiration, and adventure.

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CHAPTER 1: Introduction

Causes and effects of lithospheric density variations

Density variations within the earth give rise to surface topography, generate stress in the lithosphere, and can record the tectonic history of the crust and upper mantle. Isostatic balance requires that lithostatic pressure be equal at some depth, and thus lateral density variations are offset by topographic relief on the Earth's surface: mountains are like icebergs. Isostatic topography is modulated by flexure of the lithosphere, which acts to low-pass filter the relief supported by laterally variable buoyancy, and surface processes such as erosion and faulting superimpose short wavelength relief on the buoyancy-derived elevation field. Despite the fact that pressure may be roughly equal at some depth, surface topography and lateral density variations preclude its being equal at all depths. These pressure gradients induce stress--tensional in regions of high pressure--that tend to encourage collapse of topography and lateral homogenization of density: mountains are like honey. This stress combines with the other four tectonic stresses--edge shear, edge normal, basal shear, and basal normal--to produce the net stress field in the lithosphere.

Tectonic processes modify the distribution of mass within the lithosphere, while more enigmatic processes, such as heating, hydration, and phase transitions, change the density of lithospheric material. Thus, decomposing density variations into thermal and compositional components can give insight into the history and physical state of the lithosphere, the origin of topography, and the causes of lithospheric stress. Density variations in the crust and upper mantle can be estimated from several geophysical observables and with several methods. To first order, surface topography represents a vertical integral of density through the lithosphere. Small variations in the force of gravity across the Earth's surface arise from the threedimensional distribution of mass. Heat flow variations may provide a clue to the thermal state of the lithosphere. Geophysical imaging techniques, especially seismic tomography, are useful in that seismic wavespeed is sensitive to many of the same physical and chemical properties that affect density, with wavespeed and density generally positively covarying.

In this thesis, I present incremental improvements in techniques to estimate subsurface density from geophysical observations and/or models and present improvements in tracking and quantifying the uncertainties of these density models. These methods are applied on both regional (<1000 km) and continental (~3000 km) scales. Snapshots of modern density are then interpreted with an eye to understanding the topographic evolution of the beautiful landscapes of the American West, the chemical evolution of the lithosphere, and the mechanics of lithospheric deformation during tectonic--and more enigmatic--events.

Outline

Each of the following chapters is a stand-alone scientific paper, and may thus be considered individually. Nevertheless, Chapter 3 is a logical extension of chapter 2 and 5 of 4. Further, incremental progress in methods for estimating density with geophysical tools can be seen by examining chapters 2 and 4. Taken together, these four studies show the process, progress, and promise of developing and interrogating lithospheric density models.

In Chapter 2, "Seismological estimates of means of isostatic support of the Sierra Nevada" (Levandowski et al., 2013b), I use P-velocity models and receiver functions to derive a regional-scale density model with an eye to understanding the origin of high topography in the Sierra. Exploiting the differing sensitivities of velocity and density to thermal and compositional variations in the crust, I present a density model that satisfies both wavespeed variations and flexural isostasy. This model shows that at least 1 km of relief between the range crest and the foothills is generated by mantle temperature, suggesting that much of the post-Miocene surface uplift of the Sierra could be due to removal of cold, dense, lithospheric mantle.

The subsequent chapter of this thesis, "A flexural model linking Sierra Nevada uplift and Tulare Basin subsidence to lithospheric removal" (Levandowski and Jones, *in preparation*), explores whether this mantle material now comprises a high-wavespeed body previously imaged just SW of the Sierra or whether these high-wavespeeds record a neutrally buoyant stalled oceanic slab fragment. Estimating mantle density as in Chapter 2, I use a 2D flexural model to show that the pattern of post-Miocene subsidence above this body better accords with a continental (thermal) than oceanic (compositional) origin for this high velocity material. Further, the mass anomaly represented by this body with respect to the sub-Sierran upper mantle suffices to account for ~1.5 km of uplift. Thus, I suggest that redistribution of mass in the upper mantle dominates Pliocene topographic evolution of the region.

Chapter 4, "Origins of topography in the western U.S.: Mapping crustal and upper

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mantle density variations using a uniform seismic velocity model" (Levandowski et al., 2014b), follows a similar tack to the first chapter but on a much broader scale, seeking to understand what supports topography from the Great Plains west across the Cordillera to the Pacific. I expand on the technique outlined in Chapter 2 by incorporating heat flow data in order to estimate crustal temperatures. The ensuing density model is able to explain the surface elevation of \sim 85% of the region within uncertainty and reproduces gravity variations to within tens of mGal. Decomposing topographic support into crustal thermal, crustal chemical, mantle thermal, mantle compositional, and dynamic components shows that mantle temperature exerts a first-order control on relief between the Great Plains and the Cordillera, but the buoyancy derived from the mantle is broadly uniform within the latter. Instead, elevation differences among the Rockies, Colorado Plateau, Basin and Range, Sierra Nevada, and Cascades are crustal in origin and dominated by composition with a secondary component (+/- 350 meters) derived from thermal variations. Dynamic topography is not needed away from the subduction zone and is implausible beneath the southern Rockies and Colorado Plateau. This density model provides insight into myriad issues of current debate and will be explored further in the future; one example of such work is presented in Chapter 5.

Both the Proterozoic Colorado Plateau and Great Plains are blanketed with undeformed Cretaceous marine strata and host similar Paleozoic sedimentary sections, but the Colorado Plateau is now up to 1.5 km higher in elevation. The lithosphere is thinner under the Plateau, and the density model presented in Chapter 4 shows that ~1 km of topographic relief arises from differences in mantle temperature, suggesting that the lithosphere has thinned since the Cretaceous. Chapter 5, "Cenozoic uplift of the Colorado Plateau by lithospheric removal and crustal hydration: Insight from quantitative density models" (Levandowski et al., *in preparation*), presents a thermal model that shows that removal of some 80 km of lower lithosphere suffices to account for 1 km of uplift. An additional 500 meters of elevation with respect to the Plains is derived from crustal chemistry (rather than temperature or thickness), which accords with proposed hydration of the lower crust beneath the Colorado Plateau. This topic is revisited in Chapter 6, which explores the remaining 500 meters of Cenozoic Colorado Plateau uplift and a similar amount in the Great Plains.

CHAPTER 2: Seismological estimates of means of isostatic support of the Sierra Nevada

Abstract

Modern topography of the Sierra Nevada has been attributed to rapid uplift following foundering of negatively buoyant lithosphere into the asthenosphere since \sim 10 Ma. Uplift now manifests as \sim 2 km mean topographic relief between the crest of the southern Sierra and the western foothills and 1 to 2 km between the Sierran crest and adjacent Basin and Range. In this study we use seismic P-wave velocity structures derived from teleseismic tomography to estimate the lithospheric density structure in the region and thus infer the present sources of topographic support. We exploit the different derivatives of crustal density with temperature and wavespeed to attempt to identify a single solution for crustal density and temperature that satisfies flexural isostasy and the P-wave tomography. This solution yields both temperature variations compatible with observed heat flow and Bouguer gravity anomalies concordant with observations. We find that the topographic gradient between the crest and the eastern Great Valley is due to both crustal and mantle sources. Despite a greater thickness, the foothills crust is less buoyant than that beneath the range crest, accounting for ~ 1 km of the topographic difference. High densities are due principally to composition. High velocity upper mantle (~50-100 km depth) is also observed beneath the foothills but not the range crest, and this contrast explains an additional 1 km of topographic difference. Miocene or more recent removal of such upper mantle material from the Sierran crest, as inferred from xenoliths, would have triggered rapid uplift of ~1 km. Our

findings are consistent with the removal of negatively buoyant material from beneath the Sierra Nevada since the Miocene.

Introduction

The crust beneath the Sierra Nevada (Figure 2.1) is thinner (Fliedner et al., 2000; Frassetto et al., 2011; Jones et al., 1994; Ruppert et al., 1998) and more felsic than the worldwide average (e.g., Holbrook et al., 1992) and is substantially thinner under the high parts of the range than the low western foothills (Figure 2.2b). Despite this thin crust, the mean surface elevation within the range is ~2.5 km (Figure 2.2a). Gravity measurements show the range to be in isostatic equilibrium or even somewhat "overcompensated": having a greater gravity low than expected if the observed topography were compensated by an Airy root (Kennelly and Chase, 1989; Simpson et al., 1986). Such elevations in the absence of correlated variations in crustal thickness require lateral density variations within the crust and/or uppermost mantle (e.g., Jones et al., 1994).

The unusual crustal variations of the Sierra are matched by controversy over the origin of the topography. From the time of work by LeConte in the late 19th century (LeConte, 1886) to more recent times, geologists inferred that the Sierra rose during the late Tertiary (e.g., Christensen, 1966; Huber, 1981; Unruh, 1991; Wakabayashi and Sawyer, 2001). This assertion was challenged first by U/Th-He dating of erosion (House et al., 1998, 2001) and later by isotopic measurements inferred to reflect isotopic fractionation of precipitation associated with topography (Cassel et al., 2009; 2012; Mulch et al., 2006; 2008; Poage and Chamberlain, 2002). Although this interpretation remains controversial (Galewsky, 2009a, b; Jones et al.,

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2004; Molnar, 2010), inferences that the Sierran crest lost an anti-buoyant root between ~10 and 3 Ma (Ducea and Saleeby, 1996; 1998; Farmer et al., 2002; Manley et al., 2000) have suggested a "Sierran paradox" of an old range with a young means of support (Wernicke et al., 1996). Quantifying the contributions of compositional and thermal variations of crust and mantle to the modern support of the range can help to resolve this controversy and highlight areas where a dynamic component (here defined as one that arises from the convective regime rather than directly from density variations) is necessary.

The 2005-2007 Sierra Nevada EarthScope Project (SNEP) deployment of FlexArray broadband seismometers across much of the Sierra Nevada has led to a much fuller three-dimensional characterization of the seismic structure of the range (Frassetto et al., 2011; Gilbert et al., in review; Gilbert et al., 2012; Reeg, 2008). With such data in hand, we quantify the present day distribution of density consistent with seismic and topographic constraints of the region. The simple assumptions that seismic wavespeeds vary because of temperature variations in the mantle and because of a combination of compositional and temperature variations in the crust adequately recover topography. We thus do not find any need for dynamic topography in the region. Further, the derived density variations can then be interpreted to limit the range of plausible means of supporting the topography and thus the tectonic history of the region.

Method

It is useful to separate the contribution to topography (ϵ) from the crust (H_c) from that from the mantle (H_m). In this, we follow Lachenbruch and Morgan (1990) and define the following:

$$\varepsilon = H_c + H_m - H_0 \tag{2.1}$$

where the crustal and mantle topography are

$$H_{c} = \int_{-\varepsilon}^{z_{c}} \frac{\rho_{a} - \rho(z)}{\rho_{a}} dz$$

$$H_{m} = \int_{z_{c}}^{z_{a}} \frac{\rho_{a} - \rho(z)}{\rho_{a}} dz$$
(2.2)

 H_o is a correction term of 2.4 km to achieve isostatic equilibrium with an asthenospheric column (via mid-ocean ridges). The depth of the Moho is z_c as defined by receiver functions (Frassetto et al., 2011). In this study, we examine to a depth, z_a , of 195 km based on the loss of correlation between seismic velocity and topography below this depth. The density of the asthenosphere is ρ_a and is assumed to be 3250 kg/m³. As the motivation of this study is to explore the source of heterogeneity in the region, the exact choice of reference parameters (including asthenospheric density and crustal velocity) is of second-order importance. ε is the elevation above sea level with elastic flexural effects removed (Figure 2.2a) by smoothing by convolution with a zero-order Bessel function (Watts, 2001), discussed below.

In order to calculate the topography generated by the crust and mantle, we convert the *P* wavespeeds from the seismic tomography (Reeg, 2008; Jones et al., P-wave tomography of potential convective downwellings and their source regions, Sierra Nevada, California, *submitted to this Special Issue of Geosphere*. Hereafter referred to as Jones et al., *submitted*) into a density profile for each column, with a division at the Moho. Separating the support for smoothed topography into crustal and mantle components is necessary because we use different approaches in the crust and mantle to derive densities from seismic wavespeeds. Seismic data from SNEP are used to constrain crustal thickness (using receiver functions: Frassetto et al., 2011) and wavespeed variations in the crust and upper mantle (from P wave tomography, Jones et al., *submitted*).

Seismic Velocity Model

Jones et al. present seismic velocity models derived from inversion of travel time residuals subject to an array of starting models. Since teleseismic *P*tomography has notoriously poor vertical resolution (especially in the crust), we choose a tomography model that was derived from an inversion where the top 55 km of the starting model were defined by ambient noise and earthquake surface wave tomography (Moschetti et al., 2010a). The top of the starting model was held fixed for 14 of 21 iterations of the inversion, then allowed to vary as necessary to best fit teleseismic P delay times. Because of the greater depth sensitivity of surface waves (at the cost of lateral resolution), this inversion scheme provides a more robust view of crustal velocity variations than would be available from teleseismic P tomography alone. We have also completed our analysis with each of the other inversion schemes described by Jones et al. (from a variety of starting models). Our results are generally robust with respect to seismic models, excepting the statistically significant difference arising in west-central Nevada, somewhat outside of the Sierra Nevada region with which we are primarily concerned.

Effect Of Moho Variation On Velocity

Since we use different relationships between velocity and density (discussed below) in the crust and the mantle, we divide the tomography at the Moho depths determined by Frassetto et al. (2011). Nevertheless, the tomography of Jones et al. (*submitted*) was calculated assuming a flat Moho. Thus, while we consider crustal thickness variations in estimated densities and crust/mantle derived topography, we must also adjust the teleseismic tomography for biases introduced by variable crustal thickness. This correction and its second-order effect on predicted topography are described in the appendix.

Mantle Density Estimation

We calculate the mantle-derived topography by assuming that density and wavespeed variation is due to thermal heterogeneity. Isobaric heating will produce a decrease in both density and seismic velocity. Over a wide variety of lherzolite, harzburgite and peridotite mineralogies the temperature derivatives of density are nearly equal, though the absolute densities vary considerably (Hacker and Abers, 2004). Therefore, we make no assumption of mantle mineralogy other than that it is laterally constant across the study area at any depth. Using a compositionally independent conversion of velocity anomalies to density anomalies, we can then constrain the mantle contributions to isostasy for a purely thermally varying mantle, interpreting *P*-wavespeed variations reported by Jones et al. (*submitted*) as temperature variations and calculating the resulting density structure.

Recent laboratory data (Jackson and Faul, 2010) show a non-linear dependence of shear modulus on temperature, particularly within 150-250°C of the solidus. For this reason, we employ a decay in the density response to decreasing velocity for negative velocity perturbations, assuming that mean velocities in the upper mantle reflect temperatures of ~1050-1200 °C. This assumption is justified by estimates of minimum temperatures. At low temperatures velocity decreases less than 1% per 100 °C (Hacker and Abers, 2004). Thus a region with perturbation 5.5% (the maximum observed at 120 km) has a temperature at least 600 °C colder than background. This would then correspond to a potential temperature less than 600 °C, which we take as a reasonable lower limit at this depth.

Between 0% and -3% velocity anomaly, our estimate of $\delta \rho / \delta V p$ decreases from 10 to 6 kg/m³ per 1% velocity anomaly. This choice of parameters simulates the behavior of a variety of lherzolite, peridotite, and harzburgite mineralogies in the absence of anelasticity (i.e. at low T) (Hacker and Abers, 2004) and elastic response similar to that of millimeter-scale single crystal grains of olivine (Jackson and Faul, 2010). A schematic of the curvilinear relationship between velocity and density in the mantle is presented in Figure 2.3.

We also consider another endmember regression: a linear relationship between mantle velocity and density with no attempt to account for possible anelastic effects, but we find that our interpretations are relatively insensitive to the choice of regression, as discussed in the Results section, below.

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We note that P-wavespeed variations caused by melt (-3.6% per 1% in situ melt fraction) (e.g., Hammond and Humphreys, 2000) produce disproportionately small changes in bulk density (between 0 and 4 kg/m³ per 1% in situ melt fraction), which will cause the magnitude of variation in H_m to be too great. Because melt production and ponding depends on many parameters, we do not explicitly incorporate this effect in our wavespeed-density relationship but instead consider this effect later.

With this mantle topography in hand (Figure 2.4a), we examine the thermal and chemical variations in the crust.

Crustal Density Estimation

Seismic anomalies in the crust likely reflect variations in composition or temperature. We first convert P-wavespeeds to density within the crust using regressions of density onto P-wavespeed. We use velocities and densities reported by Christensen and Mooney (1995), correct the room temperature densities reported for the temperature at which velocities are reported (using the coefficients of thermal expansion from the same compilation and using their "average", ~15 °C per km geotherm), and regress density onto velocity for all non-volcanic and polymineralic rocks and for dunite and pyroxenite. From 0-10 km, ρ =827.5 + 316.6 x v_p . From 10-30 km, ρ =643 + 337 x v_p . From 30 km to Moho, ρ =552.0 + 350.5 x v_p . Because the velocity structure reported by Jones et al. is a perturbation from mean wavespeeds, we also choose a reference velocity structure of 5.8 km/s from the surface to 10 km, 6.5 km/s from 10-30 km depth, and 6.8 km/s from 30 km to the Moho based on a tomographic model determined using regional seismicity (Thurber et al., 2009b).

The initial assumption of an isothermal crust maximizes crustal density variations. For instance, a 3% increase in velocity from our reference at 20 km (v_p =6.5 km/s, ρ =2879 kg/m³) due to compositional variations predicts a ~65 kg/m³ higher density.

If, however, the velocity change, Δv_p , is due to a thermal perturbation, ΔT , then the density change is given by $\Delta \rho = \rho_0 \alpha \Delta T$, and since

$$\Delta T = \Delta v_p / \left(\frac{\partial v_p}{\partial T} \right), \tag{2.3}$$

$$\Delta \rho = \rho_0 \alpha \, \Delta v_p \, / \, (\partial v_p / \partial T) \tag{2.4}$$

with a known coefficient of thermal expansion, α . If the 3% velocity increase (0.195 km/s) discussed above was because this crust was colder than our reference, from eqn. (2.4) the density would only increase ~28 kg/m³, assuming a coefficient of thermal expansion of 2.5 x 10⁻⁵ °C⁻¹ and a $\delta v_p / \delta T$ of -0.5 m/s per °C (Christensen and Mooney, 1995). Thus, from eqn. (2.3) this region is 195 m/s / 0.5 m/s per °C, or 390 °C colder than reference.

Because of the linear independence of velocity-density scaling vectors, one for thermal effects and the other for compositional changes, one can employ them as a basis and thus match any arbitrary velocity-density data point. Given an estimate of mantle-derived topography, we thus calculate the crustal buoyant height necessary to match topography from eqn. (2.1):

$$H_c = \varepsilon + H_0 - H_m \tag{2.5}$$

From this value, we make use of knowledge of crustal thickness to calculate the mean crustal density necessary:

from eqn. (2.2),
$$\overline{\rho}_c = \rho_a \left(1 - \frac{H_c}{z_c} \right)$$
 (2.6)

We finally employ the thermal-compositional basis in velocity-density space to generate a thermal structure that reproduces both velocities and topography.

For example, a region with 3% velocity perturbation might need a maximum 47 kg/m³ average density anomaly in the crust relative to the background structure to be in isostatic equilibrium within uncertainty. As discussed above, a purely compositional source of heterogeneity would predict a 65 kg/m³ density anomaly. To decrease this density anomaly to 47 kg/m³, at least 50% of the velocity perturbation (1.5%, or 0.098 km/s for a 6.5 km/s background) is assigned to temperature. Following eqn. (2.3), this crustal section is then at least 195 degrees cooler on average than background.

We limit the thermal perturbation to a geologically plausible range ±250 °C. Miocene Moho paleotemperatures of ~350 °C (Molnar and Jones, 2004b) are probably a lower bound on plausible Moho temperatures and are about 1000°C below asthenospheric potential temperatures (1350 °C). Thus the Miocene crustal column would have an average temperature anomaly of 500 °C below the hottest plausible crust if geotherms are approximately linear. Given these limits, we calculate crust-derived topography (Figure 2.4b) that is a combination of thermal and chemical factors. This quantity represents an attempt to satisfy eqn. (2.5) subject to the limits on allowable thermal variations in the crust discusses above. The result of these calculations compares well with thermal constraints from surface heat flow (Figure 2.4c,d). The regions in which colder crustal geotherms are employed match almost exactly with areas of low reduced heat flow (Saltus and Lachenbruch, 1991) and low surface heat flow (Erkan and Blackwell, 2009), less than 25 mW/m² and 40 mW/m², respectively (Figure 2.4c).

Vertical Stress Coupling

We recognize that loads in the sublithospheric mantle are only partially coupled to the overlying crust and surface. Deeper loads (from velocities at 120 km and 170 km depth nodes) are reduced in amplitude to account for viscous response in the upper mantle (50% and 25% coupling, respectively) (Parsons and Daly, 1983). This reduction is an approximation, as the degree of reduction in topographic expression is a function of wavelength and viscosity structure. We have chosen these factors to be approximately correct for loads with wavelengths of ~100 km in an isoviscous mantle. Between endmember cases of full coupling and complete decoupling of the lower 100 km of the model from the surface, the mean change (absolute value) in predicted elevation at any given node is 140 meters. Our approximation cannot produce errors of greater mean absolute value than 100 meters, the difference from the fully coupled endmember.

Flexural Considerations

By smoothing the elevation field, we remove the effect of short-wavelength topographic variations generated by erosion/deposition, fault displacement and volcanism (i.e., removing the "surface loads" of Lowry et al., 2000). The elastic filter is based on an estimate of flexural rigidity (3.5 x 10²², elastic thickness 18 km) from basin geometry of the San Joaquin Valley, adjacent to the range (in accord with Saleeby and Foster, 2004). Although elastic thickness may differ across the region, varying the elastic thickness from 5 to 30 km changes our predicted topography by less than 250 m and not in a manner affecting the overall pattern of our results. Because flexural loads adjoining our study area will influence topography, we have examined these edge effects by extending our study outward 100 km into the region with tomographic coverage but no reliable receiver functions (and assuming a crustal thickness similar to the nearest RFs). These effects are less than 200 meters for models with 15 km elastic thickness and only affect the margins of our model.

The crustal and mantle topography are smoothed by the same flexural filter.

Error Analysis

Random Error

Uncertainties in seismologically based estimates of topography are calculated from the errors of the parameters used. The most important parameters in our error analysis are the velocity - density relations (Christensen and Mooney, 1995; Hacker and Abers, 2004; Jackson and Faul, 2010), wavespeed perturbations, crustal thickness variations, and background asthenospheric density. Two sigma uncertainties are estimated as 60 kg/m³ for the velocity to density conversion (Christensen and Mooney, 1995), 0.25% for velocity perturbation in the mantle, 3 kg/m³ per 1 % Vp perturbation, and 50 kg/m³ for asthenospheric density. Crustal thickness uncertainties are given by Frassetto and range between 1 and 5 km. These values yield typical 2σ uncertainties of the resulting synthetic topography of ~600 meters (Figure 2.5a).

Systematic Error

Systematic biases in our analysis may also be present to the degree that the seismic data (tomography and receiver functions) are themselves systematically biased. While travel time delay along a ray path is a robust observation in travel time tomography, the depth extent of velocity anomalies is less certain. We therefore consider the effects of the limited vertical resolution of teleseismic tomography, in particular the placement of anomalies from the mantle into the crust or vice versa. We have therefore conducted the same analysis under the physically unreasonable endmember cases that all travel time anomalies produced in the top 95 km are due to wavespeed variations solely in the crust or solely in the mantle. Since the velocity to density scaling relationships are different in the two domains but of the same polarity, the effect is one of magnification. Heterogeneities in predicted elevation (and thus residual topography) are maximized if the mantle is homogenous and the crust is responsible for all travel time delays. Because forcing all wavespeed variations into the crust generates crustal velocity heterogeneity of several tens of percent, values well beyond what is plausible, this calculation bounds our estimated topography's sensitivity to systematic biases in the tomography. The mean magnitude of change is less than 100 meters, and \sim 95% of the elevation predictions are affected by less than 400 meters.

A similar endmember investigation explores the robustness of our observations in the face of receiver function errors. In the analysis above we have treated the crust-mantle boundary as laterally variable. Nevertheless, identification of the Moho in many areas, especially those with a Moho significantly different from \sim 35-40 km depth, is challenging (Frassetto et al., 2011). Therefore, we explored similar topography calculations with a constant crustal thickness. To a great degree, the results of calculations with this architecture are consistent with those honoring RFs, as the practical effect of this change is to change the domain (crust or mantle) in which velocity anomalies are located, and consequently predicted elevations differ by up to \sim 100 meters.

Furthermore, anelastic effects are not considered in the crust. The signature of low crustal velocities (thus high H_c) and high estimated crustal temperatures might be indicative of super-solidus temperatures. If melt is present in the crust, density estimates will be too low (as described above), and positive gravity residuals will be produced.

Results

We have generated a lithospheric density model derived from seismic wavespeeds and a crustal thermal regime calculated to most closely match observed topography (summarized in Figure 2.4a,b). Where these combined crustal and mantle buoyancy variations reproduce observed topography, we suggest these variations are responsible for supporting the topography in the region. Elsewhere, other effects could be present, which could reflect the presence of (especially crustal) melt, dynamic topography (arising from convection-derived stress), compositional variations in the mantle, or compositional variations in the crust beyond those represented by the regressions of data from Christensen and Mooney (1995). To identify these areas more clearly, we calculate the residual topography, H_r , (Figure 2.5b) which represents the smoothed topography (Figure 2.2a) minus the calculated topography.

$$H_r \equiv \varepsilon + H_0 - H_c - H_m \tag{7}$$

As seen in Figure 2.5b, elevations in much of the study area are nearly matched by a combination of compositional and (geologically reasonable but not directly observed) thermal variations in the crust and thermal variations in the mantle. Furthermore, nearly all of the misfit is within the uncertainty of topography estimates (Figure 2.5c).

To explore possible errors from our scaling of mantle velocity into density, we also calculated results assuming a linear velocity-density relationship (i.e., one lacking correction for possible anelastic effects).. The results are nearly identical (Figure 2.5d), but with an additional -400 m of residual topography in the southern Cascades and southwesternmost Great Basin, near Death Valley. This residual is consistent with small amounts of partial melt or with anelastic effects, similar to the interpretations made above.

Gravity

A logical test of our 3D density model is a comparison with the observed Bouguer gravity anomaly, although we note that the limited ability to resolve the vertical position of density anomalies within the crust will lead to large uncertainties, even at the >100 km wavelengths of greatest interest here, and the limited width of the study area means that significant gravity anomalies could be generated from outside the model. Plausible ranges in density structures fitting the gravity field alone are illustrated by Kennelly and Chase (1989) and Jones et al. (1994). We have calculated the Bouguer gravity field derived from our seismically-derived model (Figure 2.6b), which largely recovers the observed Bouguer gravity variations (Figure 2.6a). The most striking feature of a Bouguer gravity map of the region is the ~300 mGal decrease from the foothills to the range crest. Our model recovers >200 mGal decrease into the range. Our model also recovers Bouguer gravity along range strike both NNW and SSE of the range crest. Remaining differences reflect a combination of masses outside the model bounds and the uncertainty in the vertical placement of density anomalies in the crust.

We note here that our gravity model seems to preclude the presence of variations in dynamic topography. To illustrate, consider a region at sea level with isopycnic mantle lithosphere (i.e., H_m =0) and 40 km thick crust of uniform density. If in isostatic equilibrium (i.e., H_c =2.4 km), the crust must be 3008 kg/m³. If a ~1 km of dynamic topography (basal normal force of ~30 MPa) is being generated by asthenospheric convection (i.e., H_c =1.4 km), then crustal density is 3088 kg/m³. The difference in the gravity signal from these two crustal columns is ~135 mGal in the infinite slab limit and 118 mGal if active over 300 km wavelength. Thus, given the absence of large magnitude, province-scale gravity residuals, we argue that the density structure that we estimate, and not dynamic topography, is responsible for the modern topography of the Sierra Nevada.

Discussion

In our model, the mean crustal density is estimated to be between 2900 and 3000 kg/m^3 beneath the foothills and $\sim 2750-2800 \text{ kg/m}^3$ beneath the range crest (Figure 2.7b). The $\sim 50-55 \text{ km}$ thick foothills crust provides $\sim 5 \text{ km}$ of topography while the thinner (40-45 km thick) but lighter crust under the range crest provides $\sim 6 \text{ km}$ of topography (Figure 2.4b).

The remaining ~1 km topographic difference between the two areas is generated within the upper mantle (Fig. 7d). A 10% velocity increase is reported by Jones et al. (*submitted*) over a lateral distance of ~150 km from beneath the range crest southwestward into the foothills (Figure 2.2c). With the anelastic scaling described above (following Hacker and Abers, 2004; Jackson and Faul, 2010), this velocity anomaly (if due as assumed to thermal variation alone) represents a density contrast of 84 kg/m³. Thus 1.0 km of relief is caused by this upper mantle anomaly, a magnitude comparable to the relief caused by crustal buoyancy variations.

As an alternative to decomposing the topography field spatially, into crustal and mantle components (Figure 2.4a,b) one can separate topography mechanistically into thermal, compositional, and dynamic components. We present such decomposition in Figure 2.7, where all values are demeaned. Since we succeed in explaining topography (within uncertainty) in nearly all of the study area and closely recovering gravity, with purely thermal variations in the mantle, we limit the need for chemical variations in the mantle or a spatially varying dynamic

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component to topography to less than our uncertainty of ~600m. Thus, compositional topography (Figure 2.7a) is simply crustal topography (Figure 2.4b) with the crustal thermal topography (Figure 2.7c) removed. Comparing the foothills to the range crest, we find little (~100 meters) difference in crustal thermal topography. Conversely, the mantle underlying the foothills provides ~1 km less thermal support than under the range crest (Figure 2.7d). As mentioned above, the remaining 1 km of discrepancy is due to crustal composition and thickness (Figure 2.7a).

These findings are consistent with the conceptual model proposed by previous and recent work (e.g., Ducea and Saleeby, 1996; Jones et al., 2004; Le Pourhiet et al., 2006; Molnar and Jones, 2004b; Saleeby et al., 2013a; Zandt et al., 2004) that negatively buoyant material has been removed from the mantle beneath the southern Sierra and is now foundering to the west of the range. If the mantle beneath the range crest had resembled that of the foothills until the Miocene, elevations would have been ~1 km lower. If either a thicker thermal boundary layer or compositionally antibuoyant mantle lithosphere had existed beneath the batholith, more uplift is admissible.

The \sim 35 km thick crust in the High Sierra has relatively low velocities that are due to lithology and not to high temperatures. The crust is thinner by 5-10 km but slower by \sim 0.25 km/s beneath the Basin and Range and Cascades than beneath the Sierra. We propose that the lower wavespeeds in the Death Valley region are due not to composition but to higher crustal temperatures and perhaps small average amounts of partial melt. The crust in the Death Valley region may be chemically denser (and thinner) than beneath the Sierra and thus supports 1 km less topography than crust under the Sierra.

Conclusions

Beginning from seismic velocities, we have created a density structure in the Sierra Nevada region that reproduces modern topography and is nearly concordant with the observed gravity field. The model assumes thermal and chemical heterogeneity in the crust and temperature variations in the mantle, but we find no need for topography derived from mantle chemistry or asthenospheric convection (i.e. dynamic topography) in order to recover observed elevations. Uppermost mantle densities account for ~1 km of relief between the Sierran crest and foothills. If such material was present beneath the range in the Miocene, we estimate ~1 km of surface uplift upon its removal and regional isostatic equilibration. Our work is consistent with the hypothesis that cold upper mantle has been removed from the southern Sierra but not the western foothills. Chemically buoyant crust of moderate thickness beneath the high southern Sierra supports >1 km of topography relative to the thinner, warmer, chemically denser crust of the Basin and Range and southern Cascades.

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Figure 2.1: The Sierra Nevada region with topography. Compare to Figure 2.2a (flexurally smoothed topography). Also shown are key geographic features: Lake Tahoe (LT), Mono Lake (ML), and Tulare Lake (TL). Seismic stations used in the tomographic inversion (Jones et al. *in preparation*) from the Sierra Nevada Earthscope Project (SNEP), the Sierran Paradox Experiment (SPE), and the Earthscope Transportable Array (TA) are shown. Miocene and younger xenolith localities in the southern Sierra separated into garnet-bearing (west) and garnet-free (east).


km layer of Jones et al. (*in preparation*). Note the high velocity material beneath the foothills, neutral velocity beneath the southern Sierra and low velocity in the southern Cascades and Basin and Range.



Figure 2.3: Mantle velocity-density relationship based on purely thermal effects. At low temperatures (positive velocity perturbations), the relationship is linear with a slope of 10 kg/m3 per 1% velocity change (~100 °C). Between 0% and -3% (~200 °C heating) velocity perturbation, anelastic effects begin to dominate, increasing the velocity

change for a unit temperature increase while density is still a linear function of temperature. At velocities lower than -3% (greater than 200 °C above background), we assume that material is above the solidus. Increased thermal input produces more melt, lowering velocity further, while melt has a very similar density to rock of the same temperature and thus bulk density remains constant (Hammond and Humphreys, 2000).





Figure 2.4: (a) Mantle topography calculated for the Sierra Nevada region for the purely thermal variation described in the text. Of particular note is the heterogeneity between the foothills (~-1.75 km) and the Sierra itself (~-0.75). (b) Crustal topography calculated for the Sierra Nevada region for the combined thermalcompostional variation described in the text. Of particular note is the

heterogeneity between the foothills (Hc \sim 5 km) and the Sierra itself (\sim 6 km). Also notable is the low buoyancy calculated for the southern Cascades and Basin and Range.

(c) Mean estimated thermal perturbation (half of Moho variation) estimated from the thermal-compositional basis approach described in the text-scale at right. Observed heat flow points in California (e.g. Erkan and Blackwell, 2009), cropped around study region are overlain for comparison (scale at left).



Figure 2.5: (a) Two sigma uncertainties in topographic prediction. Typical values are ~600 meters. (b) Residual topography (observed elevation minus predicted elevation). (c) Residual topography with 2 σ uncertainties removed. Nodes with residual magnitude within uncertainty are displayed as zero. Note change in scale from 5b. (d) Same as (c), but with densities linearly proportional to velocity (10 kg/m³ per 1% velocity perturbation)



Figure 2.6: (a) Bouguer gravity variations (i.e., regional mean removed) from the NGDC 1999 database (b) Map of demeaned Bouguer gravity field derived from density model. Areas outside of study are assumed to have neutral density.



Figure 2.7: Same color scale used for a, c, and d (a) Compositional topography: H_c (Figure 4b) minus thermal contribution from crust (Figure 7c) with the mean value removed. Note the contrast between southern Sierra range crest and regions NNW and SSE. (b) Mean crustal density

with estimated thermal effect removed. Note gradient from ENE to
WSW (Basin and Range to Sierra to Foothills).
Greater density of foothills crust is due principally to chemistry, not temperature.
(c) Crustal thermal topography. Note small magnitude of variation compared to mantle thermal (Figure 7d) and crustal compositional

(Figure 7a) contributions. Peak (Death Valley region) to trough (western foothills) amplitude is \sim 500 meters.

(d) Mantle thermal topography: Same as H_m (Figure 4a), but with the mean removed. Of particular note is the heterogeneity between the foothills (~-1 km) and the Sierra itself (~0 km).

CHAPTER 3: Linking Sierra Nevada uplift to Tulare Basin subsidence Abstract

A high seismic velocity body, the "Isabella Anomaly", has previously been imaged southwest of the Sierra Nevada beneath the Tulare basin, a region of $\sim 1 \text{ km}$ of anomalous Pliocene subsidence. In contrast, geomorphic, xenolith and seismic evidence suggest that the Sierra has risen more than 1 km since the Miocene in response to removal of dense lower lithosphere, and thus it has been suggested that the Isabella Anomaly comprises this lithospheric material. Nevertheless, while teleseismic P-tomography generally suggests a nearly vertical body from the Moho to \sim 250 km, surface wave images show an apparent connection to the Pacific plate, a gentle eastward dip and a vertical extent of only ~ 100 km, calling into question the possibility that this body is continental in origin. If the anomaly is an oceanic slab, high wavespeeds could be due to iron and water extraction during partial melting and the depleted body could be nearly neutrally buoyant. Conversely, cold continental lithospheric material would produce a large load on the overlying crust. We estimate upper mantle density variations from seismic velocities and present a 2D, distributed load flexural model of the surface response to mantle material, showing that mantle loads are capable of generating the anomalous subsidence and more. Beam-formed receiver functions suggest viscous thickening of the lower crust above this body, easily compensating for this mismatch. This pair of interpretations suggests that the body is not neutrally buoyant. Next, we calculate the excess mass within the anomaly with respect to the average mass in the region for a similar volume of mantle material. Removal of this anomalous 8.3 x 10¹⁶ kg from beneath

the central and southern Sierra would have triggered ~1.5 km of uplift. We thus argue that the Isabella Anomaly likely comprises continental rather than oceanic material and that its redistribution has played a significant role in the post-Miocene topographic evolution of the Sierra Nevada region.

Introduction

While the Great Valley of California has undergone late Miocene and more recent flexural subsidence due to asymmetric loading by the Sierra Nevada and Coast Ranges (Rentschler and Bloch, 1988), the Tulare subbasin in the southeast hosts an additional ~1.15 km of post-7 Ma sediment and ~700 meters of post-2.5 Ma sediment (Saleeby et al., 2013b). Further, the eastern extent of the Tulare basin is anomalous in that Quaternary sediments aggrade up drainages into the western foothills of the Sierra Nevada whereas drainages elsewhere along the range front do not exhibit this pattern. This area of anomalous subsidence (Figure 3.1) lies above a high seismic velocity body (Raikes, 1980), dubbed the Isabella Anomaly, that extends from near the base of the crust to ~245 km (Jones et al., *submitted*). Because of a dearth of stations, the Moho beneath the valley is virtually unsampled by conventional receiver functions (Frassetto et al., 2011; Zandt et al., 2004), and beam formed receiver functions from the eastern edge of the basin suggest a very thick (50-60 km) crust (Levandowski, 2007).

This seismic structure stands in contrast to the range crest of the central and southern Sierra Nevada some 100 km east. Here high elevations (up to 4 km) overlie a well-imaged, 35-40 deep Moho (Frassetto et al., 2011) and low velocity upper mantle (Crough and Thompson, 1977; Fliedner et al., 2000; Fliedner et al., 1996; Jones et al., 1994; Jones et al., *submitted;* Ruppert et al., 1998). These low velocities coincide with high electrical conductivity (Park, 2004), consistent with warm asthenosphere to near Moho depths as also inferred from young xenoliths (Ducea and Saleeby, 1996). Heat flow values (Saltus and Lachenbruch, 1991) are likely too low for these low velocities to reflect a steady state thermal structure, requiring that any change in the upper mantle temperature must be relatively recent.

Between the range crest and the Tulare basin, the western foothills of the Sierra display westward thickening, high velocity crust and high velocity upper mantle (Frassetto et al., 2011; Jones et al., *submitted*). Recent density estimates (Levandowski et al., 2013b) suggest that the thermal structure of the upper mantle accounts for ~1 km of relief between the range crest and the western foothills. Thus if similar high velocity, presumably cold, material had once existed under the range, its removal could account for ~1 km of uplift.

The dichotomy in seismic character between the foothills and the range crest nearly coincides with a change in xenolith populations. To the west, above or near fast material, a suite of ~10 Ma xenoliths records low temperature pyroxenitic assemblages from upper mantle depths, but to the east, ~3 Ma xenoliths were derived from warm peridotites at the same or shallower depths (Ducea and Saleeby, 1996; Ducea and Saleeby, 1998; Lee et al., 2001a; Lee et al., 2000). This ensemble of observations and P-T constraints from the xenoliths have been used to argue that ~100 km of cold, possibly eclogitic, lower crust and mantle lithosphere have been removed from beneath the southern Sierra since the Miocene. Because of its proximity to the Sierra it has been suggested that the Isabella Anomaly may comprise this removed, sinking root, with its emplacement under the Tulare basin leading to subsidence. Despite the possible importance of upper mantle mass redistribution in the topographic evolution of the Sierra Nevada region, a comparison of a seismically derived estimate of the load represented by the Isabella Anomaly and geological observations has not been performed. Previous models of the Tulare basin (Le Pourhiet et al., 2006; Saleeby and Foster, 2004; Saleeby et al., 2013a) have assumed a density anomaly rather than using one derived directly from seismic velocities.

Alternatively, the high wavespeed material beneath the Tulare basin could represent a stalled fragment of the Monterey microplate, a Farallon remnant (Benz and Zandt, 1993; Pikser et al., 2012; Wang et al., 2013). In this formulation, a young portion of the downgoing slab detaches from the older, denser plate (Lonsdale, 1991). Because it is relatively thin and melt-depleted, the slab fragment attains near-neutral buoyancy and ceases subducting. Its loss of water during melting increases viscosity, allowing this Farallon slab fragment (and possibly others) to couple to the Pacific plate after some time (Pikser et al., 2012). In this case, the colocation of high-velocity material and the Tulare basin is purely coincidental, with subsidence instead caused by a combination of thrust loading by the Coast Ranges and a flexural moat associated with Sierran uplift. Indeed, surface wave tomographic images (e.g., Moschetti et al., 2010a; Yang and Forsyth, 2006) generally show high velocity material extending westward beneath the Coast Ranges, possibly to the Pacific-North America plate boundary, having a shallow (~30 degree)

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eastward dip and a depth extent of only some 100 km. Conversely, teleseismic Pwave tomography (Jones et al., *submitted*) suggests a nearly vertical (>60 degree eastward) dip, ~100 km lateral extent, and ~200 km depth extent. This disagreement suggests that interpretations have been biased by the resolution of a given technique, with surface wave models smearing velocity variations laterally and teleseismic P models smearing them vertically.

Beyond the Sierra Nevada region, a growing number of seismologists (e.g., Schmandt and Humphreys, 2010; Sigloch and Mihalynuk, 2013; Wang et al., 2013) have suggested that many high velocity upper mantle anomalies beneath North America are remnant pieces of Farallon slab. This hypothesis suggests that slab fragmentation is an integral part of lithospheric growth and of stability of continental lithosphere over multiple Wilson cycles.

Estimates of the modern mantle density structure may provide a means of discriminating between a neutrally buoyant slab and a cold lithospheric root. If Tulare basin subsidence and a substantial portion of Sierra Nevada uplift are cogenetic, resulting from redistribution of the mass presently in the Isabella Anomaly, then two hypotheses remain to be tested: 1) that the modern <u>density</u> anomaly beneath the Tulare basin with respect to the rest of the Great Valley suffices to generate accommodation space for 0.7-1.15 km of sediment, and 2) the integrated <u>mass</u> anomaly beneath the Tulare basin with respect to the upper mantle beneath the Sierra Nevada could account for at least 1 km of range uplift. We estimate density variations in the upper mantle from teleseismic P-tomography, suggesting that if density estimates derived from wavespeeds satisfy the two tests

above, then the Isabella Anomaly is likely cold, foundering lithosphere. If not, it is more likely that high wavespeeds reflect melt-depleted, dry oceanic lithosphere with only a small thermal anomaly.

Mantle Density Calculation

We follow the approach of Levandowski et al. (2013b) in scaling upper mantle velocity perturbations to densities under the assumption that all variations are purely thermal in origin, with anelastic effects dominating the relationship near the solidus and with no change in density beyond the solidus. This methodology generated a regional density model that explains both modern topography and gravity in the Sierra Nevada region (to within ~150 meters and ~30 mGal, standard error).

We derive our density estimate from a P-velocity model from Jones et al. (*submitted*) starting from a 1-D model that includes a set of station travel-time corrections for the deep sediments of the Great Valley ("GV delay" model). We include their 70 km, 120 km, 170 km, and 220 km layers, effectively considering the depth range from ~50 to ~250 km. Since we are primarily interested in density and mass variations (the difference in mantle density under the Tulare basin and the rest of the region and the difference in mass beneath the Tulare basin and the Sierra Nevada, specifically), we remove the mean density from each layer (Figure 3.2).

We choose P-tomography rather than a surface wave model because, though velocity anomalies may be smeared vertically along the ray path, this smearing closely conserves the integrated travel time (Jones et al., *submitted*) and thus the

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vertically integrated density estimate. Surface wave inversions tend to smear velocity anomalies laterally and would therefore not produce a robust estimate of vertically integrated density variations. Since these integrals (convolved with the flexural filter of the lithosphere) create topographic relief, we choose P-tomograms over shear-wave velocity models.

Flexural Model Parameterization

Since our density estimates are derived from a tomographic model, they are discretized to the extent that the model is reported as velocity variations at a mesh of nodes. Therefore, assuming that the lithosphere responds as a thin plate, we can solve for the surface deflection due to the load beneath each surface node and superpose these solutions. As given by Watts (2001; eqs. 3.54 & 3.55), the deflection w(r) of a plate with flexural parameter α at radius r from the center of a cylindrical load of height h, radius R_d , and density ρ is:

 $w(r) = (h\rho / \rho_a - \rho_{infill}) (R_d / \alpha) [ker'(R_d / \alpha) ber(r / \alpha) - kei'(R_d / \alpha) bei(r / \alpha) + 1]; \quad (2.1a)$ within the load (i.e. for $r < R_d$), and

 $w(r) = (h\rho / \rho_a - \rho_{infill}) (R_d / \alpha) [ber'(R_d / \alpha) *ker(r / \alpha) - bei'(R_d / \alpha) *kei(r / \alpha)]; \quad (2.1b)$ outside of the load (i.e. $r > R_d$)

Here ber, bei, ker, kei, ber', bei', ker', and kei' are zero-order Kelvin-Bessel functions and their derivatives. We assume values of 3300 kg/m³ for asthenospheric density, ρ_a , and 2000 kg/m³ for infill density (the upper km or so of poorly indurated sediment). R_d , ~15 km, is chosen such that the cylinder surface area is equivalent to that of the tomography cells (~25 x ~28 km). The load density, ρ , is the average density anomaly beneath the surface node. In considering decoupling of deep or low-viscosity loads, we scale this density by a coefficient (0 for complete decoupling and 1 for mechanical coupling). Since we use mantle velocities from ~50 to 250 km depth, *h* is 200 km. Estimates of elastic thickness in the Sierra Nevada region range from ~5-20 km (Kirby and Swain, 2009; Lowry, 2012; Lowry et al., 2000; Watts, 2012). For simplicity, we discuss results for a uniform elastic thickness of ~15 km (*a*=40 km), noting that weaker lithosphere (lower *a*) would accentuate surface deflections and that stronger lithosphere would dampen variations.

Flexural Model Results

The mantle densities that we estimate from P-tomography recover the gross pattern of topography in the region. The Sierra is predicted to be highest, mantle loads depress the entire Great Valley, and the Tulare basin is a regional minimum (Figure 3.3). Variations in crustal thickness, chemical composition, and temperature--as well as in mantle composition--would be superimposed. Notably, mantle loads would allow ~1.8 km greater sediment thickness in the Tulare basin than in the rest of the Great Valley--some 0.7-1.1 km greater discrepancy than is observed (Saleeby et al., 2013b). We note that this discrepancy increases if the elastic thickness of the basin is lower (2 and 2.5 km for elastic thicknesses of 10 and 5 km) and decreases if it is higher.

The difference in predicted and observed deflection could arise from a variety of sources, and we will briefly mention several here. In a viscous medium, the surface expression of a load decreases with load depth (Parsons and Daly, 1983), but if seismic velocity is a proxy for temperature-dependent viscosity, the Isabella Anomaly is likely well-coupled to the overlying surface. Secondly, although low velocity (presumably warm, low viscosity) regions might not transmit vertical normal stress efficiently to the overlying surface, the Isabella Anomaly appears continuous on tomographic images, suggesting that loads are not transmitted through low-viscosity mantle. Finally, both conceptual (Zandt et al., 2004) and numerical models (Hoogenboom and Houseman, 2006; Stern et al., 2013) of lithospheric foundering have invoked viscous thickening of the lower crust, and we pursue such a possibility here.

Viscous Crustal Thickening

Loading by mantle anomalies may induce significant viscous thickening of the overlying crust (Hoogenboom and Houseman, 2006). Based on earlier seismic studies and their own difficulty with identifying Moho on receiver functions, Zandt et al. (2004) proposed that the lower crust of the southwesternmost Sierra in the vicinity of the drowned topography had thickened viscously in response to loading by the Isabella anomaly. In this conception, minimal impedance contrast across the Moho or a gradational crust-mantle boundary would produce low amplitude Moho P-s conversions. Near-surface scattering by the basin and complex structure in the ophiolitic foothills may further obscure this low-amplitude conversion.

To this end, Levandowski (2007) examined "beam-formed" receiver functions on three-station subarrays in the southern Sierra Nevada. In order to reduce stochastic and signal generated noise, the stations' radial and vertical traces for a given event are stacked, and the receiver function is the deconvolution of the stacked vertical from the stacked radial--effectively, three stations together function as single seismometer. This approach diminishes both random and signal-generated noise that is not common to all three stations, producing a pre-deconvolution increase in signal-to-noise ratio of $\sqrt{3}$. Consequently common arrivals (e.g., the Moho P-s conversion if the crustal thickness varies more slowly than near-surface structure) are essentially up-weighted in the receiver function.

The data used in this processing were from ~30 broadband seismometers deployed for ~6 months in the southern Sierra as part of the Sierra Paradox Experiment (SPE). Slant-stacked radial and vertical traces from the three stations encompassing 10 seconds before the direct-P arrival to 50 seconds after were filtered from 0.1-4 Hz. A cosine taper was applied to the first and last 3 seconds of each stacked trace. These filtered, tapered traces were used to calculate the receiver function by iterative, time-domain deconvolution (Ligorria and Ammon, 1999). 90 beam-event pairs passed quality control--variance reduction of 80%. 61 of these had an interpretable Moho arrival (assumed to be the greatest positive arrival between 3 and 7.5 seconds after the direct-P). Notably, beams that straddle rapid transitions (i.e., of shorter wavelength than station spacing) in crustal thickness do not recover the Moho P-s conversion, since these conversions arrive at different times and interfere destructively.

This receiver function analysis reveals similar patterns to previous estimates of crustal thickness in the Sierra Nevada region (e.g., Frassetto et al., 2011). Specifically, the lower elevation western foothills overlie thicker crust than the topographically highest southern Sierra. Nevertheless, it presents the first receiver function study to sample Moho depths beneath the Tulare basin and extends the trend of thickening crust beneath diminishing topography into the basin. Furthermore, it highlights intriguing strike-parallel variations in crustal thickness.

Arrival times from local earthquakes in the western foothills of the central Sierra (~100 km north of the region in question) suggest a crustal vp/vs of ~1.72 (Hurd et al., 2006). Crustal P-velocity is ~6.75 km/s (Thurber et al., 2009a). We bin and stack receiver functions by backazimuth quadrant, select the time to Moho from this stack (Table 1) and calculate depth using the values above. In order to estimate the conversion point for these receiver functions, we use the median backazimuth and ray parameter for each beam-quadrant bin and backproject from each vertex of the beam individually. As shown in Figure 3.4, these receiver functions image the Moho as far west as the center of the Tulare basin.

Here, we focus on the westernmost beams rather than the already wellstudied high Sierra. Receiver functions images and a brief discussion are presented in Appendix A. Along the western range front, there are two notable patterns in crustal thickness (Figure 3.4): 1) the west dip of the Moho continues under the Tulare basin, and 2) there is an additional northward dip in the Tulare basin, with crustal thickness reaching nearly 60 km at its north end. Systematically, the crust under the basin is some 10-15 km thicker than areas immediately ENE. 10 km of lower crustal material, with density 2950-3000 kg/m³ (Levandowski et al., 2013b), would provide 0.9-1.1 km of topography, since topography generated, ΔH ,

$$\Delta H = z_l (\rho_a - \rho_l) / \rho_a \tag{3.2}$$

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where z_l is layer thickness, ρ_a is density of compensating asthenosphere, and ρ_l is layer density. If this excess thickness is due to loading, then the surface

Beam name	Vertices			NW arrival time	# Events	SW arrival time	# Events
SW1	-119.0, 36.4	-118.6, 36.2	-119.0, 36.0	5.4	4	5.1?	3
SW3	-119.0, 36.4	-118.6, 36.2	-118.6, 36.5	??	0	5.0	2
WC1	-119.0, 36.6	-118.8, 36.6	-119.0, 36.4	6.4	2	6.1	5
WC4	-119.0, 36.6	-118.8, 36.6	-118.6, 36.8	6.2	2	5.9	5
NW2	-119.0, 36.6	-119.0, 36.9	-118.6, 36.8	5.8	3	5.8	4
NW1	-119.2, 36.4	-119.2, 36.7	-119.5, 36.5	??	4	5.9 or 7.7	4

manifestation of that loading would be reduced by ~ 1 km.

Linking the Isabella Anomaly to Sierra Nevada Uplift

It has been proposed (e.g., Biasi and Humphreys, 1992; Molnar and Jones, 2004a; Zandt, 2003) that the Isabella Anomaly is ~100 km of lowermost crust and mantle lithosphere removed from beneath the central and southern Sierra from ~36-38° N, east of the region where high velocity crust is imaged beneath the foothills and west of extended Owens Valley, a region of ~17,400 km² (Figure 3.6). South of here, mantle lithosphere was likely removed during the Laramide, west of here we presume that lower crust is still in place, and east of here one enters the Basin and Range. Some 1-2 km of uplift since the Miocene has been ascribed to this removal. Based on their tomographic images, Jones et al. (*submitted*) suggest that

the transport of this material is a complex, three dimensional process that leads to substantial internal deformation of the body and its mixing with surrounding asthenosphere. A similarly complex process is also favored by 2-D numerical models of convective instabilities (e.g., Le Pourhiet et al., 2006; Saleeby et al., 2013a; Saleeby et al., 2013b). Here, cold, dense material mixes with neutrally buoyant asthenosphere such that the volume of anomalous material increases, its average density anomaly decreases, and its mass anomaly is preserved. Even heat diffusion into surrounding asthenosphere preserves integrated density in the absence of vigorous convection, since asthenosphere cools even as lithospheric material heats. We are therefore more concerned with the integrated mass anomaly represented by the high seismic wavespeed body than its density contrast with the upper mantle beneath the southern Sierra.

The average density perturbation in the area between 120.5° and 118.5° W, 35.25° and 36.75° N, and 50-250 km depth with respect to the rest of the study region is 14.1 kg/m³. The Isabella Anomaly thus represents an excess mass of 8.3 x 10¹⁶ kg if it is a purely thermal anomaly. Applying such a mass to the 17,400 km² region of the southern Sierra is tantamount to suspending a 340 km thick load of the same density anomaly from the lithosphere. Following eqn. 3.2, this would depress the surface by 1.4 km.

The estimated mass also allows estimation of the thickness of material removed. We have assumed that all velocity and temperature variations are thermal in origin, and we now make the assumption, following receiver function interpretations of Frassetto et al. (2011) and Zandt et al. (2004), that the material was removed from as shallow as modern Moho depths (35-40 km). Based on surface heat flow and xenolith P-T constraints, Molnar and Jones (2004a) estimate a Moho temperature of 350 °C at ~10 Ma. If we assume that the geotherm was linear through the lower lithosphere, then its average temperature was ~850 °C (with 1350 °C asthenosphere below). This 500 °C thermal anomaly implies an average density contrast of 41 kg/m³ for a coefficient of thermal expansion of 2.5 x 10⁻⁵ per °C. A layer of such density contrast would have to be ~115 km thick to provide 1.4 km of antibuoyancy. Its thermal gradient would be 8.6 °C per km, giving a reduced heat flow of 25-35 mW/m², quite similar to values estimated in the central and western Sierra (Saltus and Lachenbruch, 1991).

Upper bound on lower crustal viscosity

By comparing the putative viscous thickening of the lower crust to the density anomaly and dimensions of the Isabella Anomaly, one can easily place an upper bound on the viscosity of the lower crust. Given the density anomaly of 14.1 kg/m³ mentioned above and a thickness (*H*) of 200 km, the Isabella Anomaly represents a load/unit area of :

$$\sigma_{zz} = \Delta \rho g H = 68.6 \, MPa \tag{3.3}$$

If we infer that there is no other mechanism of support for this load, 68.6 MPa is the maximum normal stress that the Isabella Anomaly could exert on the Moho. Further, we can speculate that the crust (elsewhere ~40 km, but ~50 km beneath the Tulare basin) has undergone 25% strain. If this deformation has occurred since 10 Ma (the age of pyroxenitic xenoliths), strain rate ($\dot{\varepsilon}$) is ~8.0x10⁻

¹⁶/s. Conversely, if this deformation has occurred since 3 Ma (the age of peridotitic xenoliths), strain rate is $\sim 2.7 \times 10^{-15}$ /s. The vertical normal stress (if we assume lithostatic pressure to be approximately laterally equal at Moho depths) generates strain following:

$$\sigma_{zz} = 2\eta \dot{\varepsilon} \rightarrow \eta = \sigma_{zz} / 2\dot{\varepsilon} \tag{3.4}$$

For the two timeframes listed above, the calculated lower crustal viscosity is of the order 10^{22} Pa s (1.3-4.3 x 10^{22} Pa s).

Given the low heat flow (Saltus and Lachenbruch, 1991), high P_n velocities (Buehler and Shearer, 2010), and deep seismicity (Hurd et al., 2006) in the western foothills of the Sierra, Moho temperatures are likely lower than in most non-cratonic regions. Using seismic models and topography, Levandowski et al. (2013b) estimated a Moho temperature near 350°C in the western foothills, or a geotherm of a mere ~7°C/km. Therefore, if viscosity in the lower crust were governed purely by temperature variations, one might suspect that the western foothills of the Sierra are a decent proxy for the maximum viscosity of lower crust. A fuller treatment of the vertical normal stress applied to the Moho by a sinking Isabella Anomaly awaits future work.

Conclusions

We have shown that Pliocene subsidence of the Tulare basin and \sim 1.5 km of uplift of the Sierra Nevada may be cogenetic, resulting from redistribution of mass in the upper mantle. A seismically-derived estimate of mantle density and a 2D flexural model show that mantle loads are capable of generating the observed 0.71.15 km of anomalous sediment accumulation in the Tulare. In fact, the excess sediment thickness with respect to the rest of the Great Valley may be slightly (< 1 km) less than is predicted from mantle density. Viscous thickening of the lower crust by ~10 km in response to mantle loading is our preferred explanation for this mismatch. We therefore suggest that it is unlikely that high wavespeed material beneath the Tulare basin is neutrally buoyant, likely precluding that it is dry, meltdepleted oceanic lithosphere. Instead, we favor the explanation that the Isabella Anomaly comprises cold, dense lithospheric material stripped from beneath the Sierra Nevada. Removal of this mass from beneath the central and southern Sierra could account for 1.4 km of uplift of the range.





Figure 3.2: Density perturbations (relative to layer mean) estimated from seismic velocities at the depths shown.





Figure 3.4: Crustal thickness map of the Sierra Nevada (after Frassetto et al., 2011). Additional constraints in the southwestern Sierra (circles) from beam-formed receiver functions (Levandowski, 2007) have been added. Two example receiver functions are shown. Note positive arrivals (Moho) near 6 seconds.



Figure 3.5: Velocity at 40 km depth from Jones et al. (*submitted*). The inferred source region of the Isabella Anomaly is the hachured polygon west of Owens Valley, east of the high-velocity lower crust, and between 36°N and 38°N. This area comprises 17,400 km².

CHAPTER 4: Origins of topography in the western U.S.: Mapping crustal and upper mantle density variations using a uniform seismic velocity model

Abstract

To investigate the physical basis for support of topography in the western U.S., we construct a sub-continent scale, 3D density model using ~1000 estimated crustal thicknesses and S-velocity profiles to 150 km depth at each of 947 seismic stations. Seismic signatures of temperature and composition are considered in the crust, but we assume that mantle velocity variations are thermal in origin. From these densities, we calculate crustal and mantle topographic contributions. Typical 2σ uncertainty of topography is ~500 meters, and elevations in 84% of the region are reproduced within error. Remaining deviations from observed elevations are attributed to melt, variations in crustal quartz content, and dynamic topography; compositional variations in the mantle, while plausible, are not necessary to reproduce topography. Support for western U.S. topography is heterogeneous, with each province having a unique combination of mechanisms. Topography due to mantle buoyancy is nearly constant (within ~ 250 m) across the Cordillera; relief there (>2 km) results from variations in crustal chemistry and thickness. Cold mantle provides ~ 1.5 km of ballast to the thick crust of the Great Plains and Wyoming craton. Crustal temperature variations and dynamic pressures have smaller magnitude and/or more localized impacts. Positive gravitational potential energy (GPE) anomalies ($\sim 2x10^{12}$ N/m) calculated from our model promote extension in the northern Basin and Range and near the Sierra Nevada. Negative GPE anomalies $(-3x10^{12}N/m)$ along the western North American margin and Yakima fold and thrust belt add

compressive stresses. We thus argue that stresses derived from lithospheric density variations may strongly modulate edge and basal force-derived stresses in many regions in the western U.S. continental interior.

Introduction

The Cordilleran orogen of the western United States is one of the broadest on Earth. Elevations above 2 km extend 1500 km from the plate boundary (Figure 4.1a) and active deformation extends 1000 km from the plate boundary. Unlike other relatively broad boundaries, this orogen lacks a continental collision or even subduction over much of its length. The processes producing such widespread uplift and deformation remain poorly understood largely because of the heterogeneous history of different parts of the orogen and the absence of uniformly collected and analyzed orogen-scale information on crustal and upper mantle structure. We address this deficiency through analysis of newly created seismic wavespeed models developed from ambient noise and earthquake surface wave observations at EarthScope Transportable Array (TA) stations spaced roughly every 80 km throughout the region.

Variations in continental elevation stem from some combination of variations in crustal density, crustal thickness, mantle density and basal normal stress at the model bottom, to the last of which we apply the unevenly defined term "dynamic topography." The mantle component of topography arises from variations in the density and thickness of the mantle lithosphere. Variations in the thickness of crust and mantle lithosphere are generally products of tectonism, whereas variations in densities are often the results of magmatism and thermal adjustments that can occur during more tectonically quiescent

times. Thus isolating the modern day mechanisms of support for provinces within the western U.S. contributes towards our understanding of the tectonic evolution of the continental lithosphere.

At the broadest scale, the elevation of the orogen is often attributed to a warm and buoyant mantle (e.g. Grand and Helmberger, 1984) emplaced after removal of the lower lithosphere because of "flat slab" subduction during the 75-45 Ma Laramide orogeny (e.g., Bird, 1988; Humphreys, 2009; Spencer, 1996). Challenges to this model range from disagreements over the geometry of the Laramide-age slab (e.g., Saleeby, 2003; Sigloch and Mihalynuk, 2013) through the post-Laramide presence of pre-Laramide mantle lithosphere in the western U.S. (e.g., Ducea and Saleeby, 1996; Livaccari and Perry, 1993) to unexplained >1 km modern elevations of the untectonized High Plains (Eaton, 1986). As a result, many workers have chosen to focus on pieces of the orogen, introducing a broad range of mechanisms for surface uplift of portions of the region. In the Colorado Plateau, for example, Roy et al. (2009) argue that ~ 2 km of Cenozoic surface uplift is due to conductive warming of the lithosphere, Levander et al. (2011) attribute elevation change to delamination of the lower crust and mantle lithosphere, and Moucha et al. (2008) and (Liu and Gurnis, 2010) favor dynamic support from the mantle convective regime. Such province-scale studies often lack a regional framework to contextualize and substantiate hypotheses. In this study, we provide such a framework and illustrate its application with specific examples.

The presence of Pleistocene to Recent deformation ~1000 km from the Pacific plate potentially shares a common origin with topography. The variations in stress manifest in observed strain are typically attributed to lateral variations in gravitational

potential energy that arise from lateral variations in the thickness, elevation, and density of the lithosphere (e.g., Flesch et al., 2000; Flesch et al., 2007; Humphreys and Coblentz, 2007; Sonder and Jones, 1999), although the significance of the stresses generated by GPE variations has been disputed (Parsons and Thatcher, 2011). Previous estimates of GPE (and thus the stresses that arise from lateral GPE variations) relied either on filtering geoid anomalies or compiling and interpolating seismic models produced by different techniques and then converting such structures into density. Geoid anomalies are equivalent to GPE if all the density anomalies contributing to the geoid are within the depth range appropriate for GPE calculations (Haxby and Turcotte, 1978). In the western U.S., however, a long wavelength contribution is probably sublithospheric, so most workers filter the geoid (e.g., Flesch et al., 2000; Jones et al., 1996). Although filtering removes the deeper contributions, it also can remove longer wavelength (e.g., provincescale) shallow contributions. Compiling seismic models in the literature and converting these to GPE estimates (e.g., Jones et al., 1996) carries the risk that biases between different workers and techniques will create geographic biases in GPE estimates. For many geodynamic applications, these discrete seismic models must be interpolated in some manner (e.g., CRUST 2.0, as used, for instance, by Flesch and Kreemer (2010)) that can further amplify biases and errors. A uniformly calculated estimate of GPE derived from an evenly distributed set of seismic observations would, at minimum, reduce any intra-orogenic uncertainty due to these biases.

The motivation for this work is to leverage the passage of Transportable Array (TA) seismometers across the western U.S. (Figure 4.1a) and the development of new seismic techniques (Shen et al., 2013b) to produce a spatially pseudo-uniform 3D density

model across the entire western U.S.. The distribution of accepted velocity models (detailed below) provided by Shen et al. (2013a) removes inter-investigator biases while providing a robust measure of seismological uncertainty. In turn, the envelope of densities estimated from those velocities allows us to quantify the mechanisms of modern topographic support and decompose this field into crustal and mantle or thermal and compositional components. Finally, the density estimates consistent with topography and seismic velocities determine variations in the body forces that contribute to the modern stress field. This workflow overcomes many of the challenges faced in previous studies, which had to rely upon spatially variable data coverage, non-uniform data processing techniques, and models that may be highly dependent on the chosen inversion parameters.

Such an improved model set allows us to pursue answers to technically and geodynamically important questions. Can seismic velocities, in concert with heat flow measurements, be used to reliably estimate densities? We check our density estimates quantitatively against predicted topography and gravity. Where do these predictions fail? We examine regions where dynamic topography, crustal melt, and anomalously felsic crust are likely. To what extent are thermal, compositional and dynamic topography each responsible for surface elevations, and is one dominant? We decompose the elevation field into these components. What are the magnitudes of GPE variations in the western U.S., and how do these variations compare with modern strain? We quantify the GPE with respect to the asthenosphere throughout the region. For use in future work, we include an electronic supplement that contains our estimated parameters at each station.

Seismic Models

The passage of the Transportable Array across the western U.S. has allowed pseudo-uniform seismic coverage of the region and new processing techniques have allowed for more robust interpretation of the data in terms of seismic wavespeed and lithospheric structure. The models thus derived represent a tool that was unavailable to previous workers, who had to rely upon velocity structures available at the time. Ideally, models would sample the lithosphere uniformly and could distinguish wavespeeds in the crust from those in the mantle, as the relationship of wavespeed to density differs in the two layers.

Shen et al. (2013b) present a technique that creates velocity models of unprecedented utility in calculating the buoyancy of crustal and mantle components in that it 1) generates uniformly sampled and processed velocity models to 150 km depth and with ~100 km lateral resolution, 2) includes crustal thickness constraints, and 3) allows easy tracking of uncertainty. At each of 947 TA stations in the western U.S., Shen et al. (2013a) began with a loosely constrained prior distribution of seismic V_{SV} velocities with depth and derived posterior distributions of ~1000 shear-wave velocity profiles (0-150 km; Figure 4.2d) and crustal thicknesses (mean shown in Figure 4.1b) that jointly satisfy surface wave dispersion curves and receiver functions. The entire posterior distribution of (acceptable) velocity profiles can be used in later calculations, allowing robust tracking of uncertainty (Figure 4.2). The inclusion of receiver function constraints greatly improves depth resolution of velocities when compared to surface wave dispersion simulations alone (Shen et al., 2013b).

The models derived from this technique offer advantages over those previously available for investigations of lithospheric density and buoyancy, which were primarily derived from active-source profile, surface tomography, local earthquake tomography, or teleseismic body wave tomography. Active-source profiles are necessarily scattered and interpretations, particularly of secondary arrivals, frequently differ between different workers (e.g., contrast Holbrook (1990) with Catchings and Mooney (1991), or Prodehl (1979) with Wolf and Cipar (1993)). Further, while quite powerful in resolving crustal velocity, these models rarely extend into the mantle lithosphere. Surface wave models have more uniformly sampled the lithosphere in this region with the deployment of the TA, but tradeoffs between wavespeeds of the crust and mantle are typically large. Local earthquake tomography is possible only where events occur and typically has poor resolution at depths in the upper mantle and lower crust below the deepest events. Teleseismic body-wave tomography and receiver functions recover only lateral gradients or contrasts and not absolute values and typically contain little information within the crust.

Because Shen et al. (2013a) produce a distribution of posterior models that satisfy the original observations, we can properly account for the effect of uncertainty in the seismological models on the derived density profile (Figure 4.2e,g). Previous work often relied on forward modeling of seismic travel-time observations lacking formal estimates of uncertainty. Additionally, because wavespeed structures intrinsically carry trade-offs between different depths (that is, uncertainties at one depth will covary with those at other depths), by estimating derived parameters (such as mean density) for each individual structure and then calculating the uncertainty in the derived parameter, we avoid overestimating the uncertainties arising from the seismological uncertainties. As explained in greater detail in section 4, we find that this seismological uncertainty dominates the uncertainty in our predicted topography, exceeding the uncertainty from the scatter in velocity to density regressions (Figure 4.2).

Density Estimation and Decomposition of Topography

We investigate the source of topographic relief in the western U.S. by exploiting the relationship between wavespeed and density. It is useful to separate the contribution to topography (*e*) from the crust (H_c) from that from the mantle (H_m). In this, we follow Lachenbruch and Morgan (1990) and define the following:

$$\varepsilon = H_c + H_m - H_0 \tag{4.1}$$

where the crustal and mantle contributions to buoyant height are

$$H_{c} = \int_{-\varepsilon}^{z_{c}} \frac{\rho_{a} - \rho(z)}{\rho_{a}} dz$$

$$H_{m} = \int_{z_{c}}^{z_{a}} \frac{\rho_{a} - \rho(z)}{\rho_{a}} dz$$
(4.2)

 H_0 is a correction term of 2.4 km to achieve isostatic equilibrium with an asthenospheric column (via mid-ocean ridges). *z* is positive downward, such that the depth of the Moho below sea level is z_c . We assume the asthenosphere to be laterally uniform below the base of the seismic models (z_a) at 150 km, and we discuss the biases introduced by this assumption below. The density of the asthenosphere, ρ_a , is assumed to be 3200 kg/m³. Because the motivation of this study is to explore the source of

topographic variation in the region, the exact choice of reference asthenospheric density is of second-order importance. ε is the isostatically supported elevation above sea level, or the convolution of surface elevation with the flexural filter of the lithosphere (e.g., following Jones et al., 1996). This convolution, in essence a low-pass filtering of surface elevation, removes flexurally supported topography from ε . The filter is a zero-order Bessel function (Watts, 2001) appropriate for elastic thickness estimates at each station (Figure 4.1c) (Lowry, 2012; Lowry et al., 2000). A model of isostatic and dynamic topography is viable if, when convolved with this same filter, it reproduces this smoothed elevation within uncertainty.

In order to calculate H_m and H_c , at each point we convert each of the ~1000 members of the distribution of v_s models into a density profile. Separating the support for smoothed topography into crustal and mantle components is necessary because we use different approaches in crust and mantle to derive densities from seismic wavespeeds. The crustal and mantle topographic contributions are smoothed by the same flexural filter described above to produce an estimate of the surface expression of these loads.

Mantle-supported Topography

We initially solve for H_m (which we will term the mantle topography for clarity) by assuming that density and wavespeed variations are a product of thermal heterogeneity. Isobaric heating will produce a decrease in both density and seismic velocity. Over a wide variety of lherzolite, harzburgite and peridotite mineralogies the temperature derivative of density is nearly the same, though the absolute densities vary considerably (Hacker and Abers, 2004). Therefore, we make no initial assumption of mantle mineralogy other than that it is laterally constant across the study area at any depth. This simplification is clearly not a robust characterization of the lithosphere and upper asthenosphere, so we will later examine regions where compositional variations may manifest themselves as residuals in topography and gravity calculated assuming uniform mineralogy. Converting velocity anomalies to density anomalies, we can then constrain the mantle contributions to isostasy for a purely thermally varying mantle, interpreting *S*-wavespeed variations reported by Shen et al. (2013a) as temperature variations and calculating the resulting density structure.

Laboratory data (Jackson and Faul, 2010) show a non-linear dependence of shear modulus on temperature, particularly within 150-200 °C of the solidus, which we assume to be ~1350 °C. To account for increasing anelastic effects with increasing temperature, we must relax the linear relationship between density and velocity at low velocities (Figure 4.3). Using the empirical relationship between shear modulus and temperature for 1 mm diameter olivine crystals (Jackson and Faul, 2010) and a coefficient of thermal expansion of $3x10^{-5}$ per °C, we calculate that the last ~150-200 °C before the solidus manifests as decrease in seismic velocity of ~3%.

To estimate the velocity that corresponds to the 1200 °C isotherm, or the approximate temperature below which an elastic velocity-density relationship is valid, we note that the maximum velocity found by Shen et al. (2013a) at 120 km depth is 4.75 km/s and is observed in the Wyoming craton. Here, the thermal boundary layer is ~200 km thick (e.g., Schutt et al., 2011). If the geotherm is approximately linear, the expected temperature at 120 km depth is ~820 °C (surface temperature 20 °C). Again using the shear modulus data for 1 mm olivine (Jackson and Faul, 2010) and a coefficient of
thermal expansion of 3.2×10^{-5} per °C, we find that this 380 °C temperature contrast corresponds to a 5.5% velocity difference. The 1200 °C isotherm therefore corresponds roughly to a velocity of 4.75 / 1.055 = 4.50 km/s. For wavespeeds below that of the solidus, we expect most seismic variation to be due to the increased presence of melt. Because melt produces small changes in bulk density (between 0 and 4 kg/m³ per 1% in situ melt fraction) (e.g., Hammond and Humphreys, 2000), we assume that density is constant for wavespeeds more than 3% lower than 4.50 km/s (i.e., <4.37 km/s).

We assume that there is no variation with depth of our velocity to density relationship largely because of uncertainty in the depth variation of anelastic effects. Certainly the solidus occurs at increasingly lower velocities at greater depth, but the volume of material affected is small and has little effect on our calculations. If the solidus occurs at 2% lower velocity at 150 km than at the Moho, then we will underestimate density an average of up to 7 kg/m³ in the mantle. Thus, by eqn. 4.2, errors introduced by the constant-solidus approximation might yield errors in our estimate of topography of up to 200m.

We assume that mantle loads are fully coupled to the overlying crust and surface (although resultant topography is modulated by the flexural filter). The degree to which loads present in a deforming viscous medium affect surface topography depends on the viscosity structure of the medium and load wavelength (e.g., Parsons and Daly, 1983). Nevertheless, the lateral resolution (~100 km) of dispersion curve inversions and long wavelength (200-300 km) of velocity anomalies reported by Shen et al. (2013a) are broad enough that we treat mantle loads as fully coupled to the surface. We then smooth these

values by the estimated flexural response of the lithosphere. Following these assumptions, we calculate the mantle topography (Figure 4.4a).

Crust-supported Topography

We assume that seismic wavespeeds in the crust depend on some combination of composition and temperature. We convert *S*-wavespeeds to density within the crust using *Brocher's* (2005) regression of density onto *S*-wavespeed and a correction for thermal variations based on estimates of temperature variations in the crust, discussed below.

The assumption of an isothermal crust would maximize estimates of crustal density variations, as the partial derivatives $\partial \rho / \partial v_{s(temperature)}$ and $\partial \rho / \partial v_{s(composition)}$ are different. Regressions of density onto velocity (Brocher, 2005; Christensen, 1996) show that near v_s =3.9 km/s (v_p =6.7 km/s and ρ =2900 kg/m³), $\partial \rho / \partial v_{s(composition)} \approx 523$ kg/m³ per km/s, while $\partial \rho / \partial v_{s(temperature)} = 249.2$ kg/m³ per km/s. Here, we assume a coefficient of thermal expansion of 2.5 x 10⁻⁵ °C⁻¹, a v_{ρ}/v_s of 1.78 that is insensitive to temperature, and a $\partial v_{\rho}/\partial T$ of -0.5 m/s per °C (Christensen and Mooney, 1995); this calculation is discussed below. Because of this difference, and because we aim to quantify the tectonic significance of crustal temperature variation, we seek to separate the minor (-0.281 m/s °C⁻¹) velocity (and thus inferred density) variations due to temperature from those due to composition, and to do so we must estimate the mean temperature of the crust.

We use surface heat flow observations (from SMU Geothermal Database; http://smu.edu/geothermal/georesou/DataRequest.asp, accessed on 11/15/2012) smoothed over a 100 km radius as a proxy for crustal temperature (Figure 4.4c). Obviously such a dataset places only some constraints on the overall thermal structure of the crust as

hydrological effects, varying thermal conductivity, variable radioactive heat generation and disequilibrium geotherms all will disrupt the relationship between surface heat flow and subsurface thermal structure. We follow Hasterok and Chapman (2007a) and avoid any attempt to correct for these issues, as observational constraints on all of these parameters are weak and spatially irregular.

We instead assume a simple linear geotherm through the crust and that this geotherm is accurately reflected in the heatflow. For example, if we assume a conductivity of 3 W/m°C, then a heatflow of 75 mW/m², such as in the southern and central Basin and Range, corresponds to a crustal geotherm of ~25 °C/km and a Moho (30-35 km) temperature of ~850 °C. Since the maximum reasonable Moho temperature is that of convective asthenosphere, we limit the temperature to 1350 °C. Also, for a region with a 200 km thick thermal boundary layer, a 50 km deep Moho would be no colder than 350 °C, so we place this as the minimum temperature. This approximation neglects heat production in the upper crust, however. We explore the consequences of this simplification briefly below.

Our estimate of crustal density is thus derived from our inferred temperature variation and the observed shear wavespeed:

$$\rho = \rho_{\text{Brocher}} \left(v_s - \frac{\partial v_s}{\partial T} \Delta T \right) + \frac{\partial \rho}{\partial T} \Delta T$$

$$= \rho_{\text{Brocher}} \left(v_s - \frac{v_s}{v_p} \frac{\partial v_p}{\partial T} \Delta T \right) (1 - \alpha \Delta T)$$

$$= \rho_{\text{Brocher}} \left(v_s + 0.28 \frac{\text{m/s}}{^{\circ}\text{C}} \Delta T \right) (1 - 2.5 \cdot 10^{-5} (^{\circ}\text{C})^{-1} \Delta T)$$
(4.3)

where $\rho_{\text{Brocher}}(v_s)$ is the combined regression of v_s on v_p and v_p on density of Brocher.

For example, the regression of density onto velocity suggests that a 0.1 km/s increase in velocity due to compositional variations corresponds to a ~52.3 kg/m³ higher density. Alternatively, a temperature difference of -356 °C would also increase velocity by ~0.1 km/s. This temperature difference corresponds to a density increase of only ~25 kg/m³. The compositional regression thus overestimates the density of "cold" material by a factor of ~0.08 kg/m³ °C⁻¹. Modest velocity variations due to temperature can lead to tectonically significant errors in predicted density (in a 40 km crust, the error above would produce ~400 meters of topography), and ascribing all velocity heterogeneity to composition or to temperature will lead us to calculate inaccurate densities.

Assuming a linear geotherm in the crust will overestimate mean crustal temperatures. For example, a region of 75 mW/m² surface heat flow and 40 km thick crust -- both typical of the western U.S. -- would be predicted to have a Moho temperature of ~1020 °C limit if radioactive heat production is ignored. If, instead, heatflow largely reflects radiogenic heat in the upper 10 km and reduced heatflow is ~40 mW/m² -- a low value for the western U.S. (Blackwell, 1983; Saltus and Lachenbruch, 1991) -- the Moho temperature would be a more modest 670 °C. The average temperatures of these crustal columns are 520 °C and 390 °C, respectively. The error introduced by this 130 °C mismatch is, on average, ~10 kg/m³. Following eqn. 4.2, we would underestimate H_c by 130 meters. Thus, we will tend to underestimate H_c in areas with high radiogenic heat production. Nevertheless, since the variation in heat production within provinces is similar (Morgan and Gosnold, 1989) we expect that this effect mostly

produces a relatively uniform underestimate in H_c near or under 100m. The approximation of the thermal regime is thus likely adequate for our purpose owing both to the unknowns in the thermal structure and the relatively small contribution to topography from thermal variations within the crust, which, as we discuss below, has a total range of about 500 m (Figure 4.4e).

Taking these inferred temperature perturbations (Figure 4.4d) into account, we calculate crustal topography throughout the western U.S. (Figure 4.4b). We note that *S*-wavespeed variations caused by melt (7.9% decrease per 1% in situ melt fraction) produce far smaller changes in bulk density than composition or temperature (between 0 and 4 kg/m³ per 1% in situ melt fraction)(Hammond and Humphreys, 2000). This bias will cause H_c as calculated here to be too great in areas with crustal melts, a bias we consider in addressing areas of topographic misfit below.

Topography Uncertainties

The posterior distribution of wavespeed structures from the Monte-Carlo investigation of seismic models of Shen et al. (2013b) allows for a direct analysis of the uncertainty of our topographic calculations due to seismic velocity and crustal thickness uncertainties. Each individual velocity profile and crustal thickness is converted into a density profile, and the attendant crustal and mantle topographies are calculated. The resulting ~1000 estimates of H_c and H_m at each point define the seismic uncertainty in our results (Figure 4.4f).

We have quantified the uncertainty in our topography estimates and investigated its origins. We find that the variation in elements of the posterior distribution of v_s models

overwhelms the uncertainties in converting velocity to density. As illustrated in Figure 4.2, basing predictions of elevation on a single velocity profile not only may produce systematically biased results but also underestimates the uncertainty of the prediction, even if the uncertainty in density derived from velocity is considered. In fact, once the full posterior distribution of velocity models is analyzed, incorporation of such uncertainty yields no further variation in the predicted topography. Even if deviations from the presumed velocity-density relationship correlate over layers 10 km thick, uncertainties in H_c increase by less than 50 meters. Vertical correlations would have to be crustal in scale, significantly greater than the \sim 5 km length suggested by investigations of the Ivrea Zone (e.g., Goff et al., 1994; Holliger and Levander, 1992; Levander and Holliger, 1992), to have an impact comparable to the uncertainty in velocity profiles. Thus, we do not include uncertainties in the velocity to density conversion in our uncertainty of H_c . Substantial and systematic deviations of a region from the assumed velocity-density relationship will produce equally systematic deviations of the calculated topography from that observed. We discuss such occurrences below.

A potential difficulty arises if large magnitude, long wavelength variations in radial anisotropy are present. We have assumed that the v_{SV} profiles used are sufficiently close to the mean shear wave speed of the crust and mantle for calculation of densities. However, the presence of variations in radial anisotropy would produce biases; we would be underestimating the Voigt average shear velocity by ~1.5% in areas where radial anisotropy was ~5%, in which case we would overpredict topography by about 800m relative to isotropic sections if the anisotropy extended through the entire crust and mantle to ~150 km depth. Variations inferred by Moschetti et al. (2010b) for the area west of 110°W suggest we might be underpredicting elevation in the Colorado Plateau and further underpredicting elevation in the Sierra Nevada by some few hundred meters. At a longer wavelength, the model of Marone et al. (2007) suggests that we could be overpredicting elevations in the southern Rockies by several hundred meters. Owing to the present low-resolution of and variations among existing of models of radial anisotropy, we do not explicitly correct for this effect.

Other limitations that could affect our results arise from the parameterization of the seismological model. Crustal low-velocity zones are prohibited, and sharp increases in wavespeed are only permitted at the base of sediments and at the Moho. Since surface wave dispersion at a given frequency is sensitive to a wide depth range, low velocity zones would, to first order, result in a broad region of lower wavespeeds in the surface wave models. The topography calculations, however, are integrals through the crust and upper mantle. Thus, the fact that low velocity zones are not allowed in the crust does not substantially affect our conclusions. Strong discontinuities at depth other than the Moho could result in material being assigned the wrong velocity to density function; this is presumably most likely in areas where "double Mohos" are present (e.g., southern Wyoming (Karlstrom et al., 2005)). The error here depends on whether the seismic inversion has selected the top or bottom Moho and the velocity of the material between the two Mohos. Take, for example, a 10 km thick underplate with a velocity of 4.3 km/s $(v_p \sim 7.5 \text{ km/s})$. We would calculate a density of 3137 kg/m³ and H_c of 197 meters from this body (following eqn. 4.2). If the receiver functions show "Moho" above this body, however, we would calculate a density of 3185kg/m^3 and an H_m of 47 meters. The total error in topography is thus ~150 meters.

Comparison to Topography and Adjustments to Densities

Where our combined crustal (Figure 4.4b) and mantle (Figure 4.4a) variations reproduce observed topography acceptably (to the limits shown in Figure 4.4f), lithospheric thermal and crustal compositional variations are sufficient to support the topography. Elsewhere, other factors presumably affect the velocity-density relationship or the surface elevation. The factors include the presence of crustal melt, compositional variations in the mantle, lithospheric mantle extending below the model, or normal stress derived from the convective regime of the asthenosphere (dynamic topography). To identify these areas more clearly, we calculate the residual topography, H_r , (Figure 4.4g) which represents the smoothed topography (Figure 4.1a) minus the topography calculated from our initial assumptions, e_c .

$$H_r \equiv \varepsilon - e_c = \varepsilon + H_0 - H_c - H_m \tag{4.4}$$

Thus, positive residual topography denotes a higher observed elevation than predicted by a given model. In light of the appreciable uncertainties (mean 2σ =590 meters), we pay particular attention to regions where H_r exceeds our calculated uncertainty (Figure 4.5a).

As seen in Figure 4.5a, elevations in ~84% of the study area are matched within uncertainty by a combination of compositional and thermal variations in the crust and thermal variations in the mantle. In the Yellowstone region, Cascadian forearc, and Southern Rocky Mountains, elevations are coherently predicted to be 0.5-1 km higher than observed. This discrepancy can be eliminated in one of two ways: imposing a <u>downward</u> normal stress of 15-30 MPa on the lithosphere (i.e., dynamic subsidence), or systematically increasing lithospheric density. Since we use Brocher's regressions of v_s onto v_p and v_p onto density, pervasive quartz-poor crust and its high v_p/v_s ratio would lead us to underestimate density for an observed seismic velocity. Alternatively, the presence of melt drastically reduces velocities but has a lesser effect on density, and would thus cause us to underestimate density.

Near Yellowstone and in the southern Rockies, negative residual topography coincides with heat flow in excess of 100 mW/m² (Figure 4.4c), high seismic attenuation in the crust (e.g., Phillips and Stead, 2008), somewhat low crustal v_p/v_s ratio (Lowry and Perez-Gussinve, 2011), and inferred near- or supra-solidus mantle temperatures (see Figure 4.4a). For these reasons, we favor the explanation that partial melt is present in the crust in these areas (Figure 4.5b), though we recognize that presence of melts in the mantle at sub-solidus temperatures would also produce an error in calculated density. In a crustal column with original S-velocity of 3.15 km/s that contains an average 1% melt, wavespeeds decrease by 0.25 km/s (following Hammond and Humphreys, 2000). We would misinterpret such a decrease as a 135 kg/m³ (~5%) density decrease, and, when integrating through a 40 km crustal column, would overestimate crustal topography by 1.7 km. Thus, ~0.3% in situ partial melt throughout the ~40 km crust near Yellowstone would account for ~ 0.6 . km residual topography, as would 3% in a 4 km zone; we cannot discriminate between distributed and concentrated crustal melt especially as the shear wave structures we use prohibit crustal low velocity zones. An similar average amount of melt would resolve the discrepancy in the southern Rockies, though this could be lessened if radial anisotropy is indeed stronger in this area.

Conversely, -0.5 to -1 km residual topography in the Cascadian forearc coincides with low heat flow. Geologically, the presence of substantial amounts of serpentine, with its unusual wavespeed to density relationship, might be expected to contribute to this error. Although serpentinization lowers v_s substantially, using the non-linear regressions of Brocher (2005) of v_s to v_p and v_p to density produces a misfit of only 111 kg/m³ (~4%). A 10 km layer that is 50% serpentine increases estimated topography by only 175 meters. Thus, the ~30 km crust would have to be nearly entirely serpentinite to account for the topographic residual. Instead, we propose that the forearc is depressed by downward basal normal stresses of ~15-30 MPa exerted on the lithosphere by subduction zone processes (Figure 4.5d).

Elevations in the southern Sierra Nevada and northern Basin and Range (Figure 4.5a) and to a lesser extent Wyoming and the Idaho batholith (Figure 4.4g) are higher than expected by as much as 500 meters. Overestimating density by 40 kg/m³ (~1.4%) throughout a 40 km crust would account for this discrepancy. All of the regions in question (Figure 4.5c) coincide with low v_p/v_s estimated from receiver functions that Lowry and Perez-Gussinye (2011) interpret as reflecting a high quartz content. Indeed, in the Sierra Nevada, the crust is thought to comprise only the upper, felsic portion of a Mesozoic-Cretaceous batholith (Ducea and Saleeby, 1998; Fliedner et al., 2000; Levandowski et al., 2013b).

Particularly felsic crust, and its attendant low v_p/v_s would lead us to calculate systematically high densities, since we use Brocher's regressions of v_s onto v_p and v_p onto density. To estimate the mean amount of quartz increase necessary to reconcile seismic velocities and topography, we compare the observed and predicted densities of pure

quartzite (Christensen, 1996), assuming that the polynomial regression (Brocher, 2005) is appropriate for average continental crust of $\sim 60\%$ SiO₂ containing $\sim 10\%$ quartz. The density estimated from our application of Brocher's regressions for a v_s of 4.035 km/s (200 MPa quartzite) is 2975 kg/m³, but the density of quartzite is only 2652 kg/m³. Thus an increase in the modal abundance of quartz of $\sim 90\%$ corresponds to a 325 kg/m³ (~11.6%) bias in density. Thus, a 500 meter elevation error can be explained by an increase in the modal abundance of quartz of ~11% throughout the crust.

Overestimated topography could also be attributed to variations in mantle chemistry. Increasing Mg# (Mg#=MgO/[MgO+FeO]) of olivine in mantle lithosphere both increases velocity (~0.3% per 0.01 increase in Mg#) and decreases density (~8.5 kg/m^3 or ~0.02% per unit increase in olivine Mg#) (Schutt and Lesher, 2010). Thus, an increase of ~ 0.01 in Mg# can resolve the apparent discrepancy between seismic velocity and elevation. Because we might suspect significant iron depletion in the Wyoming craton and we observe large magnitude, long wavelength, positive residual topography there (Figure 4.5a), we discuss this possibility below.

We desire to account for the effects of melt and varying quartz content in order to more clearly examine the contributions of crust and mantle to topography; thus we modify the density structures calculated assuming thermal variations throughout the lithosphere and compositional variations in the crust (topography of which is shown in Figure 4.4). These modifications affect $\sim 12\%$ of the study area. To recover topography, a mean crustal density adjustment of $\Delta \rho$:

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$$\Delta \rho = -H_r \left(\frac{\rho_a}{z_c}\right) \tag{4.5}$$

is necessary, with residual topography, H_r , as defined in eqn. 4.4, asthenosphere density ρ_a =3200 kg/m³, and crustal thickness z_c . Adding this term to the previously derived structures yields an adjusted density structure and final estimate of crustal compositional topography (Figure 4.6d).

These adjustments are relatively small, especially when compared to the ~60 kg/m³ (~2.1%) standard errors associated with a linear velocity-density scaling (Christensen and Mooney, 1995). Where applied, the mean increase in crustal density due to melt is 23.5 kg/m³ (~0.8%), and the mean decrease in crustal density from quartz content is -17.7 kg/m³ (-0.6%).

Results

With the adjusted density estimates as described above, we examine three characteristics: the decomposed topography, predicted gravity, and gravitational potential energy. The values calculated for these and other parameters are presented in the electronic supplement.

Topography

We have determined a set of mantle and crustal densities that accord with both seismic velocity and topography. Nearly all of the variation in topography (Figure 4.1a, 6a) across the western U.S. arises from compositional (Figure 4.6c,f) and thermal (Figure 4.6b,e,h) variations expressed in wavespeed variations. Elsewhere (Figure 4.5a), in areas where crustal melt or highly felsic crust (Lowry and Perez-Gussinye, 2011) are likely, the relationship between velocity and density must be slightly adjusted. Taking these small adjustments (eqn. 4.5) into account, we can further separate modern topography into thermal and compositional components (Figures 4.6-4.7).

In addition to dynamic topography (Figure 4.4.4.5d, 4.6i), there are exactly four isostatic components of topography: mantle thermal (Figure 4.6h), mantle compositional, crustal thermal (Figure 4.6e), and crustal compositional (Figure 4.6f; where both thickness (Figure 4.1b) and chemistry (Figure 4.8) are considered).

We have assumed that all density and velocity variations in the mantle are thermal in origin and have found no locations violating this assumption beyond uncertainty. We acknowledge two shortcomings of this assumption, however. First, the seismic velocity models only extend to 150 km. The high velocity, presumably lithospheric material below this depth that is seen in other velocity models would provide negative buoyancy if included in our analysis. For example, in the Wyoming craton, where our current analysis suggests a need for a small amount ($\sim 200 \text{ m}$) of additional buoyant support (Figure 4.5a), a $\sim 2\%$ velocity anomaly extends to ~ 250 km depth (e.g., Schmandt and Humphreys, 2010). This 14 kg/m³ thermal density anomaly could plausibly be countered by Mg# increase of $\sim 1.6\%$. This depletion would supply ~ 400 m of buoyant height if present over 100 km thickness. The second shortcoming of our analysis with regard the mantle (and crustal) chemistry is the combination of >500 m analytical uncertainty (Figure 4.4f) with systematic patterns of residuals (Figures 4.4g-h, 4.5a, and 4.7). Our technique is insensitive to province scale deviations from mantle isochemistry (or similar scale, systematic variations in crustal chemistry) that produce up to \sim 500 m of topography. To illustrate the magnitude of mantle depletion that our current analysis may be failing to recognize, we return to the Wyoming craton. The initial misfit to topography (Figure

4.4g) is ~700 m. If due to the upper 100 km of the lithospheric mantle (~50-150 km), this misfit would correspond to an increase in Mg# of ~2.6. Combining this plausible depletion of the upper 100 km of the lithosphere with the plausible depletion from 150-250 km, mantle composition could account for ~1.1 km of buoyant height in the Wyoming craton.

In the crust, we have estimated a mean temperature, and thus (following eqn. 4.2) the effect of thermal expansion and contraction, Hc_{thermal} , is (Figure 4.6e):

$$Hc_{\text{thermal}} = z_c \,\rho_0 \,\alpha \,\Delta T / \rho_a \tag{4.6}$$

where z_c is crustal thickness, ρ_0 is the crustal density, α is the coefficient of thermal expansion (2.5 x 10⁻⁵ °C⁻¹), and ΔT is derived from heat flow. Then, the crustal compositional topography (Figure 4.6f) is given as:

$$Hc_{\rm comp} = H_c - Hc_{\rm thermal}$$
 (4.7)

Examining the different components of topographic support (Figures 4.6-4.7), it is clear that differences in elevation among the southern Basin and Range, the Great Basin, the Colorado Plateau and the southern Rockies are mostly to be found in differing crustal characteristics (Figure 4.6d-f) rather than heterogeneity in the mantle.

Wyoming, the one Cordilleran province lacking warm mantle, is higher than the plains because of higher crustal compositional topography (Figure 4.6f). We note also that since our density models recover topography and gravity (presented below) in the Wyoming craton reasonably well, the high velocities observed below 150 km (e.g., Burdick et al., 2008) either represent cold but iron-poor isopycnic material or require somewhat lower densities in the mantle above 150 km.

Mantle topography accounts for the eastward descent from 2.5 km elevations in the Rockies to less than 1 km in the Great Plains.

Comparison to Gravity

Our adjusted density structure can be tested by calculating gravity anomalies from it and comparing these to the observed Bouguer anomaly (an alternative approach, as followed by Mooney and Kaban (2010), uses gravity as a primary observable and deduces density variations from gravity). We note first, however, that the predicted gravity field at a given station is strongly dependent on the shallow structure beneath that station. The top few km is poorly constrained seismically because of limited sampling at higher frequencies. Furthermore, the use of receiver functions in determining acceptable seismic models can impart a local bias; the structure beneath a station may not be representative of the surrounding ~70 km.

We do not have a direct way to track the uncertainty of gravity, because we make use the entire posterior distribution of 1D velocity profiles and gravity is sensitive to density in 3D. Instead, we may estimate the uncertainty from the mean topographic uncertainty, ~600 m. This uncertainty corresponds with an uncertainty of 48 kg/m³ through 40 km crust. Such an error would produce 47 mGal of gravity misfit. Alternatively, we can calculate the uncertainty at each station that results from the seismological uncertainty in its 1D posterior distribution. These uncertainties range from +/-10 to +/-40 mGal (2σ). Uncertainties for our 3-D gravity will be greater to the degree that errors in seismic wavespeeds are laterally correlated. Thus, in considering gravity residuals, we bear in mind these crude estimates of uncertainty. We estimate the Bouguer anomaly from our preferred 3D density, including adjustments for inferred crustal melt and quartz enrichment; details of this calculation are presented in the appendix. The 3D gravity prediction (Figure 4.9a) recovers the overall Bouguer anomaly variations of the western U.S. (Figure 4.9b) to within a few tens of mGal (Figure 4.9c). The misfit has mean magnitude of 25 mGal, well below our crude estimate of uncertainty. The misfit is less than 60 mGal in 95% of the study area and below 30 mGal in 75%. Nevertheless, there are systematic provincial residuals that suggest coherent errors in estimated density. A regional misfit of 30 mGal corresponds to a consistent density error of 20 kg/m³ throughout a 40 km crust, though this density error would only produce 250 meters of topography. We do not investigate the origins of low magnitude but regionally coherent misfit in gravity (Figure 4.9c) and/or topography (Figures 4.4g,h, 4.5a, 4.7), but these could be due (among many other things) to regional differences in mantle chemistry (as discussed above), crustal geotherm, anisotropy, or crustal mineralogy.

Gravitational Potential Energy

Lateral variations in pressure that arise from topography and density differences generate stresses within the lithosphere, with areas of high integrated pressure (or GPE) exerting compressive stress on adjacent regions of lower GPE. From distributions of density, we can calculate the GPE available to modulate the stress field imposed by basal and edge forces:

$$GPE = \int_{0}^{150+E} \rho(z)gz\,dz$$
 (4.8)

where *z* is positive upward from the model base (in this case 150 km depth, such that mean sea level is at *z*=150 km), and *E* is the surface elevation. We compare GPE to that of an asthenospheric column (Jones et al., 1996) of density 3200 kg/m³ that extends from 150 km to 2.4 km depth (order 10^{14} N/m). This column is calculated to be in isostatic equilibrium with a mid-ocean ridge (Lachenbruch and Morgan, 1990). Such a column of asthenosphere should be free of deviatoric stresses, making it a useful reference state. High potential energy (positive anomaly, Δ GPE, relative to an asthenospheric column) increases horizontal deviatoric extensional stresses while negative Δ GPE favors contractional deformation. Lateral variations in Δ GPE are of the order 10^{12} N/m (Figure 4.10), and uncertainties are of the order 10^{11} N/m, up to 10^{12} N/m.

The mean deviatoric stress acting between two idealized columns is the difference in GPE divided by the column, or lithospheric, thickness (e.g., Sonder and Jones, 1999). To illustrate, for two adjacent regions of 200 km thick lithosphere with a GPE contrast of 2×10^{12} N/m, the mean deviatoric stress exerted is 10 MPa. The magnitudes of these stresses are used by geodynamicists to calculate the magnitude of plate boundary stresses in the continental interior and to estimate the bulk viscosity of the lithosphere in thin viscous sheet models (e.g., Flesch et al., 2007). For example, a 10 MPa stress in lithosphere with a bulk viscosity of 10^{23} Pa s would generate a strain rate of 10^{-16} /s, similar to values in the Idaho Batholith (e.g., Payne et al., 2013).

Positive Δ GPE is most prominent in the Sierra Nevada and the northern Basin and Range. The eastern front of the Sierra is, in fact, a locus of modern extension (e.g., Unruh and Hauksson, 2009) and the northern Basin and Range has been previously suspected to be a region of highly positive GPE (Humphreys and Coblentz, 2007; Jones et al., 1996). The large-scale negative Δ GPE along the western margin of North America may be the result of surface depression due to subduction related dynamic pressures (especially north of the Mendocino Triple Junction). A limb of negative Δ GPE projects eastward from the Cascade margin at ~46 °N. This anomaly coincides with the Yakima Fold and Thrust Belt, a zone of Quaternary deformation that may be connected to compressional strain along the Cascade margin (Blakely et al., 2011). We propose that body forces modulate edge and basal stresses to create this pattern of contractional deformation.

Discussion

Topography and earlier studies

The explanation of western US topography presented here differs from that inferred in earlier work; we consider here the origins of those differences and the implications for the validity of our results. Jones et al. (1996) did not use any seismological information for the mantle and instead inferred variations in H_m by assuming isostatic compensation in the asthenosphere. Values of H_c were mainly derived from *P*-wave refraction profiles using the Christensen and Mooney (1995) wavespeeddensity regressions with no correction for lateral thermal variations. Most of our values of H_c are quite similar where seismic models were available to *Jones et al.*; the most notable differences are significantly lower H_c values (Figure 4.6d) in the California Central Valley and Colorado High Plains (which are due at least in part to the thermal effects on crustal wavespeed-density relations that *Jones et al.* ignored) and somewhat lower values in the northern Basin and Range. We only find ~350 m variation in support from the mantle within the Cordillera outside Wyoming (Figure 4.6f), about one quarter that of Jones et al. (1996). The differences mainly reflect the explicit inclusion of mantle wavespeed anomalies here and suggests that most of the topography Jones et al. attributed to mantle density variations is caused by other effects.

Hasterok and Chapman (2007b) focused on a more complex thermal analysis of North America but overall used nearly identical assumptions as *Jones et al.* in correcting for varying compositional H_c in trying to reproduce topography across the region. We limit the use of surface heat flow to estimate crustal temperatures but Hasterok and Chapman [2007b] extended its use into the mantle. Although we share an assumption of a thermal origin for mantle density anomalies, we rely on seismic wavespeeds to estimate mantle temperatures and thus density. Furthermore, we adjust observed wavespeeds to account for thermal variations before interpreting chemical variations in the crust. They estimated, as we do (compare their Figure 4.4c with our Figure 4.6b), that thermal variations account for ~3 km of relief. The differences between surface heat flow and seismic wavespeeds at depth suggests that much of the scatter *Hasterok and Chapman* [2007b] found can be attributed to non-steady-state thermal structure within the lithosphere. Unlike an extrapolation of surface observations into the mantle, our approach permits different thermal structures in the crust and mantle, implicitly allowing nonsteady state geotherms, which are reflected by seemingly inconsistent crustal (Figure 4.6d) and mantle (Figure 4.6g) thermal topography as in the Sierra Nevada and Colorado Plateau.

Our estimates of crustal compositional topography (Figure 4.6c) variations are generally of the same polarity but of different magnitude from Hasterok and Chapman (2007a). Specifically, we tend to calculate much greater variations of crustal buoyancy

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within the Cordillera. For example, comparing the northern and southern Basin and Range, we propose that nearly all of the ~1 km of relief is compositional in origin, as are differences between these provinces and the southern Rockies (Figure 4.6c). In each case, Hasterok and Chapman (2007b) ascribe this relief to thermal variations.

Lowry et al. (2000) inferred from an analysis considering gravity and some seismic refraction models that about 2 km of topographic variation was caused by dynamic stresses applied to the lithosphere. Although our estimates of crust-derived topography are fairly close to theirs (compare our Figure 4.6d and their Plate 3b), we have a very different appraisal of the topography due to thermal effects in the mantle largely because we are interpreting seismic models in the mantle, but they projected surface heat flow measurements into the mantle. This disparity suggests that the lithosphere in the region is either not in a conductive steady-state or has large deviations in conductivity or heat production from values presumed by *Lowry et al.*, and this difference is why we do not infer the significant dynamical component to topography that they reported, at least away from the subduction zone.

Examples of applications to province-scale tectonics

One can interrogate this subcontinental-scale model of the sources of topography to examine province-scale tectonics in a regional context. The comparison of two provinces, for example, allows for an explanation of modern relief; as an example, we explore the topographic disparity of the southern and northern Basin and Range (Figure 4.1a, 4.6a). Alternatively, with a constraint on paleoelevation and a knowledge the

modern topographic components, one can examine the changes that may be responsible for surface uplift or subsidence; we do so for the Colorado Plateau.

The ~800 meters of relief between the southern and northern Basin and Range has been variously attributed to plume-derived dynamic topography (Saltus and Thompson, 1995), variations in mantle lithospheric thickness (Jones et al., 1996) and/or chemistry (Schulte-Pelkum et al., 2011) and variations in crustal density (Eaton et al., 1978). Examining Figures 4.6-8, we conclude that relief is generated by crustal compositional variation, not by mantle variations. Furthermore, this elevation difference is due not to crustal chemistry; mean thermally-corrected densities (Figure 4.8) are 2726 kg/m³ in the southern and 2716 kg/m³ in the northern Basin and Range, which contributes ~ 100 meters of relief. Instead topography arises from a crustal thickness difference of 4.5 km (Figure 4.1b), which accounts for 700 meters of relief. Note that this interpretation is at odds with earlier estimates based on refraction studies (e.g., Catchings and Mooney, 1991) that showed a \sim 30 km crustal thickness throughout the Basin and Range. The receiver functions used here and other continent-scale receiver function studies (e.g., Gilbert, 2012) allow for a more uniform sampling of crustal thickness whereas the sparse available seismic refraction lines may preferentially sample anomalously thin or thick crust in a given region.

The Colorado Plateau has risen ~2 km since the Cretaceous, and this uplift has been attributed to a variety of processes, including 1) warming of the uppermost mantle either conductively (Roy et al., 2009) or by removal of the lower lithosphere recently (Levander et al., 2011) or during the Laramide (Spencer, 1996), 2) dynamic support from the mantle convective regime (Moucha et al., 2008), or 3) crustal thickening due to lower

crustal flow (McQuarrie and Chase, 2000) or a lower crustal phase change (Jones et al., 2011; Morgan, 2003). We find that the mantle thermal topography (compare Colorado Plateau and Great Plains in Figures 4.6g,h and 7) and the crustal chemistry (compare Colorado Plateau to Plains and southern Rockies in Figure 4.8) are responsible for the modern elevation, and modern topography does not require dynamic support (Figure 4.5d). The 40 kg/m³ difference in crustal chemical density between the Plains and Colorado Plateau that we estimate lends ~500 meters of relative support to the latter. Hydration of lower crust, as recorded in xenoliths (e.g., Butcher, 2013) is one possible means of changing crustal density since the Cretaceous. The remaining 1.5 km of uplift is suspiciously similar to the difference in mantle thermal topography between the Colorado Plateau and the lower part of the Great Plains (Figure 4.6g,h). If the continental interior serves as an estimate for the pre-Cretaceous Colorado Plateau (Spencer, 1996), then a change in the mantle thermal structure largely explains the change in topography. The magnitude of this inferred change suggests that mechanical replacement of the lower thermal boundary layer is more likely than conductive heating. For a ~90 km thick lithosphere (e.g., Levander and Miller, 2012), the mean lithospheric temperature would have to change by 520 °C (and thus the base of the lithosphere by 1040 °C, even in the endmember case of a linear geotherm) to produce 1.5 km of uplift. Alternatively, removal of ~75 km of thermally equilibrated mantle lithosphere (i.e., with a linear geotherm) could produce 1.5 km of uplift (Levandowski et al., 2013a) over 70 Ma. A more detailed investigation of this process is deferred to another manuscript (Levandowski et al., in preparation).

Implications for dynamic topography

Previous workers have invoked dynamic topography, or basal normal stresses exerted by the convective regime of the asthenosphere, to explain elevations of the Colorado Plateau (Liu and Gurnis, 2010; Moucha et al., 2008), the southern Rockies (Karlstrom et al., 2012) or Yellowstone (e.g., Pysklywec and Mitrovica, 1997) although some of these studies include erosion of mantle lithosphere with the effects of basal normal stresses. Nevertheless, we present densities that recover modern elevations reasonably well, and since the gravity misfit (Figure 4.9c) is within expectations of our topographic uncertainties, we largely reject the role of dynamic pressures in supporting topographic variations in the Cordillera, except east of the vicinity of the Cascadia subduction zone (Figure 4.5d).

To illustrate, consider a region at sea level with isopycnic mantle lithosphere (i.e., H_m =0) and 40 km thick crust of uniform density. If in isostatic equilibrium (i.e., H_c =2.4 km), the crust must be 3008 kg/m³. If a ~1 km of dynamic topography (basal normal force of ~30 MPa) is being generated by asthenospheric convection (i.e., H_c =1.4 km), then crustal density is 3088 kg/m³. The difference in the gravity signal from these two crustal columns is ~135 mGal in the infinite slab limit and 118 mGal if active over 300 km wavelength. Thus, given the absence of large magnitude, province-scale gravity residuals, we argue that the density structure that we estimate, and not dynamic topography, is responsible for the modern elevation of the western U.S..

GPE and earlier studies

Previous attempts to estimate GPE have necessarily relied upon the filtered geoid or interpolations of seismic models. With the availability of more uniform seismic coverage, however, we have been able to improve upon the limitations of former by including long-wavelength variations due to shallow (<150 km) structure and upon the latter by utilizing a near-uniform model coverage and uniform seismic data processing methods.

The locations of relative GPE anomalies vary substantially in earlier studies. In a study of similar spatial dimensions to ours, Flesch et al. (2007) estimate a GPE high in the southern Rocky Mountains and a general gradient downward toward the Pacific margin. Using the geoid somewhat differently, Humphreys and Coblentz (2007) suggested that the northern Basin and Range broadly, and northeastern Nevada specifically, was a region of high GPE and that the Rockies were nearly without GPE-derived deviatoric stresses. Jones et al. (1996) also found high GPE in northeastern Nevada, when using seismic velocities instead of the geoid. But unlike later work, they also found high GPE in the Sierra Nevada, low GPE on the western margin, and variable GPE in the southern Rockies. Our work, perhaps not surprisingly, more closely resembles this previous effort that uses seismic velocity than those using geoid. We find GPE highs in NE Nevada and the Sierra Nevada and a coherent, consistent GPE low along the western margin of the continent (Figure 4.10).

The magnitudes of GPE anomalies that we calculate are comparable to previous estimates (Flesch et al., 2007; Humphreys and Coblentz, 2007; Jones et al., 1996). Ranges have been estimated at 4.5 TN/m, 9 TN/m, and 4.5 TN/m, respective to the citations above. Our estimated range is \sim 7 TN/m (with the exception of the unreliable edges of our model).

Although the implications of our new GPE estimates require a more complete analysis, certain effects can be illustrated by simple comparison with published work. In general, approaches to modeling lithospheric deformation tend to combine the stress field from a particular GPE distribution with that derived from boundary and basal stresses with the goal of matching some observable, typically the magnitude and/or orientation of the stress, strain rate or velocity field (e.g., Flesch et al., 2000, 2007; Humphreys and Coblentz, 2007); the total effect of a difference in GPE distribution will depend on how this influences the derived boundary stresses, which in turn relates to the relative importance of boundary and internal stresses in driving deformation. If the differences do not much alter the estimated boundary stresses, then differences will be accommodated by changes in effective viscosity and the relative role of body forces. For instance, Flesch et al. [2000] used the ratio of stress to strain rate to estimate effective viscosity. Higher values of \triangle GPE than used in that paper, such as in the eastern Sierra and parts of the Basin and Range would produce higher effective viscosities than published estimates and further reinforce the importance of body forces in these areas while lower GPE values, such as in the southern Rockies relative to Flesch et al.'s [2007] estimate, would tend to yield lower effective viscosities but would reduce the significance of GPE in encouraging deformation in the region.

Conclusions

We have generated a density model of the western U.S. lithosphere from surface heat flow and seismic models at the well-distributed Transportable Array stations and quantitatively checked it against predicted topography and gravity. Large overestimates of elevation (>600m) near Yellowstone and in the southern Rocky Mountains are attributed to the presence of lithospheric melt, while we attribute some underestimates of topography to anomalously quartz-rich crust. Overestimated elevations near the Cascadia subduction zone probably are caused by dynamic effects up to ~1 km. Correcting for these effects yields our final density structure.

The origin of topographic variations within the western U.S. can be examined by decomposing the elevation field into its five components: crustal thermal, mantle thermal, crustal compositional, mantle compositional and dynamic topography (Figure 4.6). Crustal composition (Figure 4.6f) and mantle temperatures (Figure 4.6h) dominate both in magnitude and heterogeneity. Dynamic topography (Figure 4.6i) is only locally important along the plate boundary, whereas crustal thermal topography (Figure 4.6e) is of low magnitude across the region. We find no statistically significant need for relief generated by variations in mantle composition, though variations of several hundred meters are possible.

The Cordillera overlies nearly constant-density mantle (Figures 4.6g, 4.6h), and topographic relief generally reflects variations in crustal thickness and chemistry.

The Wyoming craton overlies cold, dense mantle (Figure 4.6h), but thick crust (Figure 4.1b) allows modest elevations. High velocities observed below 150 km (e.g., Burdick et al., 2008) presumably record cold mantle that is either itself isopycnic with surrounding asthenosphere or requires the mantle lithosphere above 150 km to be depleted and less dense than we infer here. Elevation decreases eastward into the Great Plains are due to chemically denser crust (Figure 4.8).

Away from the Cascadia subduction zone, our results limit topographic effects of dynamic stresses to under a few hundred meters. Our seismologically based density structure reproduces elevations within 600m at the 2σ level. Significant dynamic effects should produce large errors in our predicted gravity field, but the differences between observed and predicted gravity are as expected from seismologically derived uncertainties.

Finally, we have uniformly quantified the variations in gravitational potential energy throughout the western U.S. (Figure 4.10). Positive GPE anomalies favor horizontal extension in the Northern Basin and Range and along the eastern front of the Sierra Nevada. Compression in the Yakima fold and thrust belt, conversely, coincides with negative anomalies.



Figure 4.1:

(a) Elevation of the western U.S., smoothed as discussed in the text. Physiographic boundaries are shown in black outline. SN: Sierra Nevada; SRP: Snake River Plain; NBR: Northern Basin and Range; SBR: Southern Basin and Range; CP: Colorado Plateau; SRM: Southern Rocky Mountains; WC: Wyoming craton; GP: Great Plains

(b) Crustal thicknesses from Shen et al. (2013a). Each of the 947 seismic stations used is marked with a small circle.

(c) Elastic thickness estimated from Lowry (2012).



Figure 4.2:

(a) A single member of the posterior distribution of S-velocity profiles for station S22A, Creede, CO.

(b) The envelope of 671 density profiles derived from (a), with random error in velocity-density conversion at each node as given by Christensen and Mooney (1995) in the crust and with 30% uncertainty in the mantle. Uncertainty is not vertically correlated.

(c) Histogram of the elevations predicted from (b). Uncertainty is 2σ .

(d) The 671 S-velocity profiles in the posterior distribution at station S22A.

(e) The envelope of densities derived from (d), with no uncertainty in velocity-density conversion.

(f) Histogram of the elevations predicted from (e). Note different mean and much larger uncertainty than (c).

(g) The envelope of densities derived from (d), but with uncertainty as in (b).

(h) Histogram of elevations predicted from (g). Note similar mean and uncertainty to (f).



Figure 4.3: Mantle velocity-density relationship based on purely thermal effects. At lc temperatures (positive velocity perturbations relative to 4.5 km/s), the relationship linear with a slope of 7 kg/m³ per 1% velocity difference (\sim 70 °C). Between 0% and -3 (\sim 150 °C heating) velocity perturbation, anelastic effects begin to dominate, augmenti the velocity decrease for a unit temperature increase while density is still a linear functi of temperature. At velocities lower than -3% (greater than 150 °C above background), v assume that material is above the solidus. Increased thermal input produces more me lowering velocity further, while melt has a very similar density to rock of the sat temperature and thus bulk density remains constant (Hammond and Humphreys, 2000).



Figure 4.4:

(a) Initial estimate of mantle topography. Note large, negative values in the Wyoming Craton and Great Plains, especially when compared to the relatively constant value in the southern Rockies, Colorado Plateau, and Basin and Range.

(b) Initial estimate of crustal topography. Note large magnitude of support from the crust of the southern Rockies and Wyoming craton.

(c) Observed surface heat flow from SMU Geothermal Database. Colorscale is chosen to reflect conversion into mean crustal temperature (Figure 4d), which is described in the text.

(d) Estimated mean crustal temperature variations, based on heat flow.

(e) Topography variations arising from estimated crustal thermal structure. Note \sim 500 meter peak-to-trough amplitude.

(f) 2σ uncertainty in predicted elevation, derived from the envelope of acceptable velocity profiles. Mean 2σ uncertainty is 590 meters.

(g) Residual topography, H_r , as defined in eqn. 4.4. Negative values indicate an underestimate of density or existence of a positive downward basal normal force being exerted on the lithosphere that is not reflected in the seismic velocity. Positive values indicate upward basal normal force or density overestimate.

(h) Accepted misfit between predicted and observed topography. All values are within uncertainties shown in (f). Color scale as in (g).



Figure 4.5:

(a) Statistically significant residual, or H_r (Figure 4g) +/- uncertainty (Figure 4f).

(b) Minimum amount of in-situ melt, averaged through the crust that we propose contributes to H_r .

(c) Minimum amount of quartz increase, averaged through the mantle lithosphere that we propose contributes to H_r .

(d) Minimum amount of dynamic (downward) topography that we propose contributes to H_r . ~30 MPa downward normal force would produce 1 km of surface depression.



Figure 4.6: Components of topography. Left column is a combination of columns to the left. Top row is a combination of rows below. Same scale is used in (b), (c), (f), and (h).

(a) Flexurally smoothed topography of the western U.S.. Same as Figure 1a.

(b) Topographic variations due to thermal variations (i.e. Hc_{thermal}+Hm_{thermal} with the mean removed to facilitate comparison). Note consistent values in Basin and Range, Snake River Plain, and Southern Rockies. Note also low values in Wyoming craton and Great Plains.

(c) Topographic variations due to compositional variations (i.e. $Hc_{comp}+Hm_{comp}$ with the mean removed to facilitate comparison). Note very high values in the Wyoming craton, high values in the southern Rockies and Colorado Plateau, and low values in the Basin and Range.

(d) Final estimate of crustal topography, representing initial estimate (Figure 4b), corrected for the effect of proposed melt and quartz content (Figure 5b-c).

(e) Same as Figure 4e. Topography variations arising from estimated crustal thermal structure.

(f) Crustal compositional topography, representing total crustal topography (Figure 8c) corrected for estimated thermal topography of the crust (Figure 4e, 7d). Values are presented with the mean removed for more ready comparison. Note high values in the Great Plains, Rockies, and Colorado Plateau (0.5 to 2 km) as compared to the Snake River Plain and Basin and Range (<0 km).

(g) Mantle topography, same as Figure 4b.

(h) Mantle thermal topography, same as Figure 7g, but with the mean removed and then plotted on the same scale as Figures 7b, 7c, and 7f. Note large contrast between the Wyoming craton and Great Plains (values -0.5 to -1.4 km) and the Southern Rockies, Snake River Plain, Colorado Plateau and Basin and Range (nearly constant values of (0.6 to 0.85 km).

(i) Dynamic topography as in Figure 5d.


Figure 4.7: Bar graph of the average components of topography by province (CA=Cascades, SN=Sierra Nevada, SRP=Snake River Plain, SBR=Southern Basin and Range, GB=Great Basin/Northern Basin and Range, SRM=Southern Rocky Mountains, WC=Wyoming, HP=High Plains—smoothed elevations above 1 km, LP=Low Plains—smoothed elevations below 1 km). The minimum of each component is set to zero to better examine variations. Note similar mantle thermal topography from the Cascades through the Rockies and the strong difference between these regions and Wyoming/Plains. Other topography is mostly crustal in origin and dominated by compositional variation. Average smoothed elevation is shown in black for comparison. Misfits between predicted and observed are within uncertainty (see Figure 4h).



Mean Crustal Density, thermal variation removed

Figure 4.8: Crustal density from seismic velocities after the estimated thermal variations (Figure 4d) are removed. Note contrast between Wyoming and Southern Rockies.



Figure 4.9:

(a) Predicted Bouguer gravity field from our proposed density model. A correction (described in the text) is applied to mimic the effect of the Juan de Fuca slab.

(b) Observed Bouguer gravity field.

(c) Observed-predicted gravity field. 90% of the study area is matched within 40 mGal.



Figure 4.10: Gravitational potential energy (GPE) variations predicted from our preferred density model. Note positive GPE anomalies in the extending Northern Basin and Range (NBR) and eastern front of the Sierra Nevada (SN). Also note negative GPE along the western margin of North America, especially in the Cascades Forearc, with an arm of negative GPE extending eastward at the latitude of the Yakima Fold and Thrust Belt (YFTB).

CHAPTER 5: Cenozoic uplift of the Colorado Plateau by lithospheric removal and crustal hydration: Insight from quantitative density models

Abstract

The Colorado Plateau has risen from near sea level during the Cretaceous to its modern elevation of ~2 km without significant internal deformation and while other Proterozoic North American terranes, notably the Great Plains, have not attained such elevations. The origin of this topographic change, however, is enigmatic. We interrogate a seismically-derived continental-scale density model of the western United States with an eye to understanding this uplift. Of the suite of possible differences between the modern state and that of the Cretaceous, we argue that transient thinning of the thermal boundary layer by 100-125 km since 70 Ma is the dominant source (>1 km) of uplift. Additionally, our calculations suggest a contribution (<1 km) from changes in lower crustal mineralogy, potentially due to hydration by a dewatering Farallon slab.

Introduction

Shallow marine sediments constrain the late Cretaceous elevation of the Colorado Plateau to near sea level. The mechanism(s) by which the region has attained its modern, ~2 km, elevation (Figure 5.1a) continues to provide fodder for much research and debate. The age of Colorado Plateau uplift is similarly contentious, and each suggested cause of uplift carries with it an implicit range of possible uplift histories and rates. Hypotheses fall into several basic divisions, and readers are directed to references herein for fuller discussions of each: Laramide-related, early-mid Cenozoic crustal shortening/thickening (e.g., Bird, 1979; McQuarrie and Chase, 2000); thermal or chemical, mid-late Cenozoic mantle lithosphere modification (e.g., Morgan and Swanberg, 1985); Tertiary or more recent stream incision and/or isostatic adjustment (e.g., Pedersen et al., 2002); and recent-ongoing dynamic uplift (Liu and Gurnis, 2010).

Just as a change in the elevation of a region requires a change in isostatic buoyancy or dynamic topography, modern relief between two areas requires a difference in one or both of these quantities. For example, based on their 3D density model of the western U.S., Levandowski et al. (2014b) speculate that the ~1 km elevation difference between the Colorado Plateau and Great Plains (referring to the region west of 100°W and east of the Rocky Mountain front; here the average elevation is ~1 km) is due dominantly to differences in the mantle thermal structure. This spatial comparison is presented because the modern stable interior of North America may be a good proxy for the pre-Laramide Colorado Plateau (Spencer, 1996), because thicknesses of Paleozoic platform strata are similar in the two regions, indicating that they once had the same lithospheric buoyancy, and each was at sea level at ~70 Ma.

In the present study, we leverage the seismically-derived 3D density structure for the western U.S. of Levandowski et al. (2014b) to constrain the possible magnitudes of dynamic and isostatic contributions to Cenozoic uplift of the Colorado Plateau, and our inferences are informed by comparison with the modern Great Plains. We note that this model of density from the surface to 150 km depth reproduces both topography and gravity in the Colorado Plateau and thus presumably offers a decent estimate of the mean 1D density structure in the province.

Contributors to surface elevation

The surface elevation, before being smoothed by lithospheric flexure (e.g., Watts, 2001) is the combined expression of isostatic (i.e. density related) and dynamic (i.e. related to basal normal forces exerted by convection) components. The isostatic component, ε , can be subdivided into crustal and mantle components, H_c and H_m (following Lachenbruch and Morgan, 1990):

$$\varepsilon = H_c + H_m \tag{5.1}$$

where the crust- and mantle-supported elevations are:

$$H_{c} = \int_{-\varepsilon}^{z_{c}} \frac{\rho_{a} - \rho(z)}{\rho_{a}} dz$$

$$H_{m} = \int_{z_{c}}^{z_{a}} \frac{\rho_{a} - \rho(z)}{\rho_{a}} dz$$
(5.2)

The depth of the Moho below sea level is z_c . Models used here extend to a depth, z_a , of 150 km, and we assume the asthenosphere to be laterally uniform below. The density of the asthenosphere, ρ_a , is 3200 kg/m³. Elastic flexural effects are removed by convolving H_c and H_m with a zero-order Bessel function (Watts, 2001) defined by estimates of the flexural rigidity of the lithosphere (Lowry, 2012; Lowry et al., 2000).

As a corollary to eqns. (5.1-5.2), the change in isostatic topography, $\Delta \varepsilon$, due to a uniform density change, $\Delta \rho$, in a layer of thickness *z*, is given as:

$$\Delta \varepsilon = z \Delta \rho / \rho_a \tag{5.3}$$

Density variations in the crust and mantle are functions of both composition (chemistry and mineralogy) and temperature. Thus, there are exactly four isostatic components: crustal thermal, mantle thermal, crustal compositional, and mantle compositional. For ease of discussion, we include crustal thickness variations (Figure 5.1b) in the crustal compositional term. The derivations of each term are given by Levandowski et al. (2014b) and in Appendix A.

In comparing a region to itself at two disparate points in time, we seek variations in these four isostatic terms and/or in the dynamic topography. Following McGetchin and Merrill (1980) and Morgan and Swanberg (1985), uplift requires 1) crustal heating, 2) mantle heating, 3) crustal thickening, 4) chemical (non-thermal) crustal density loss, 5) chemical (non-thermal) mantle density loss, 6) downward dynamic stress in the past, or 7) upward dynamic pressure currently. We will now semi-quantitatively investigate these seven possibilities.

Crustal Heating

The modern heat flow of 50-60 mW/m² in the Colorado Plateau (e.g., Blackwell and Richards, 2004; SMU, 2012) is low compared to the broader western U.S. Furthermore, lower crustal and upper mantle seismicity suggests a low geothermal gradient (Wong and Humphrey, 1989), and a joint analysis of heat flow and P_n velocities (Schutt et al., 2011) suggests a Moho temperature of ~700 °C. Under the approximation of a linear geotherm, even a 300 °C heating of the Moho since the Cretaceous (i.e., Moho originally at 400 °C) would only produce an average density change of 150°C x 2800 kg/m³ x 3.0 x 10⁻⁵/°C=12.5 kg/m³. Following equation (3) and using a crustal thickness of 40 km, this thermal expansion would trigger only 160 meters of uplift. Crustal heating is thus only capable of small amounts of uplift.

Mantle Heating

Estimates of modern mantle thermal buoyancy are derived by pairing an empirical scaling of velocity variations to temperature variations that accounts for anelasticity (Jackson and Faul, 2010) with an assumption of a linear dependence of density on temperature. The topography generated by this buoyancy (Figure 5.2) is similar in the Colorado Plateau to other regions modified in the Cenozoic (the southern Rockies, Basin and Range, Sierra Nevada, and Cascades). When compared to the Great Plains, however, the difference in mantle temperature (from Moho to 150 km) creates ~1 km of topographic relief. If cold mantle lithosphere extends below 150 km beneath the Plains, this difference is greater still. Thus, a change in the thermal structure in the mantle since the Cretaceous is a possible source of substantial uplift of the Colorado Plateau and will be addressed in more detail below.

Crustal thickening

Modern crustal thickness (Figure 5.1b) of the Colorado Plateau is ~40 km (Gilbert, 2012; Shen et al., 2013a). Following eqn. (5.2), with modern crustal density ~2770 kg/m³ (Levandowski et al., 2014b), a thickening of 5 km would have produced ~670 meters of uplift. Thus, thickening of the crust is a viable contributor to increased surface elevation. Note, however, that if the proto-Colorado Plateau to Great Plains analogy holds, then any increase in crustal thickness must have been minor, as the latter region has modern Moho depths of ~50 km (Figure 5.1b).

Crustal chemical density decrease

Based on seismic velocities (Shen et al., 2013a) and observed surface heat flow, the estimated mean crustal density in the Colorado Plateau is ~2770 kg/m³ (Levandowski et al., 2014b). This value is lower than in the Great Plains (~2810 kg/m³), and this discrepancy increases with depth. The Colorado Plateau crust is ~20 (+/- 10) kg/m³ denser than the Plains from 0-10 km, then 30 (+/-7), 50 (+/-6), and 95 (+/-9) kg/m³ less dense in the 10-20, 20-30 and 30-40 km layers (Figure 5.3). Because of the aforementioned comparison of the pre-Laramide Colorado Plateau to the modern Great Plains, crustal density decrease merits consideration as a mechanism for post-Cretaceous uplift; this process is discussed below. Following eqn. (5.3), and with a crustal thickness of 40 km, a density change of 40 kg/m³ would produce 500 meters of uplift.

Mantle chemical density decrease

Xenoliths record anomalously magnesian (i.e., melt-depleted) mantle lithosphere beneath the Colorado Plateau (e.g., Alibert, 1990; Lee et al., 2001b; Smith, 2000), but uplift since the Cretaceous would require depletion (i.e., replacement of iron with magnesium) in the Cenozoic. Depletion is generally associated with high temperature partial melting such that Archean and Proterozoic melt extraction generally produces more magnesian mantle than Cenozoic modification (e.g., Jordan, 1975). Moreover, we (Levandowski et al., 2014b) did not find a need for magnesium enrichment in the mantle lithosphere of the Colorado Plateau in order to explain modern topography, gravity, and seismic velocity. We thus do not favor the possibility that the Colorado Plateau lithosphere has been further depleted since the Cretaceous.

Dynamic topography: Down then

An alternative to modern uplift generated by dynamic effects is the possibility that modern topography is the result of relaxation of previous long-lived dynamic subsidence (Mitrovica et al., 1989). The Colorado Plateau was not far above sea level from at least the Mississippian until the Cretaceous, as evidenced by a series of units with submarine members (e.g., Mississppian Redwall Limestone, Permian Kaibab Limestone (Condon, 1997), Triassic Moenkopi (Blakely, 1973), and Jurassic Carmel (Wilson and Palmer, 2008)). Because eustatic sea level rose more than 200 meters between the Triassic and Cretaceous [REF?], only minimal amounts of surface depression beyond that generated by thrust loading are required to create adequate accommodation space for sediments shed from the Sevier Highlands (e.g., White et al., 2002). Furthermore, given the absence of sedimentological evidence for large magnitudes of subsidence at a discrete time, any dynamic topography would have been nearly constant in magnitude for ~200 m.y., with this period spanning the onset of subduction along western margin of North America. Because the initiation of subduction would modify the existing pattern of asthenospheric flow and thus dynamic topography, we reject the possibility of long-lived downward dynamic topography that was relieved since the Cretaceous.

Dynamic topography: Up now

Geodynamic models (e.g., Liu and Gurnis, 2010) have allowed that small-scale convection, particularly related to post-Farallon subduction return flow, could elevate the surface of the Colorado Plateau by some 700 meters. Nevertheless, we have generated a density model (Levandowski et al., 2014b) that accords, within uncertainty, with seismic

velocity, heat flow, gravity and topography. Consequently, we suggest flexurally modulated isostatic forces and not dynamic pressures compensate topography.

Quantification of influences on topography

We argue that changes in two lithospheric characteristics have contributed to the ~2 km change in surface elevation of the Colorado Plateau since the Cretaceous: mantle thermal structure and crustal composition. We now explore the mechanisms and magnitudes by which the topography associated with each factor may have been modified.

Mantle heating

Heating of the mantle may either be conductive or advective. The fact that Miocene xenoliths sample Proterozoic mantle lithosphere beneath the Colorado Plateau (e.g., Alibert, 1990; Lee et al., 2001b; Smith, 2000; Smith and Riter, 1997) indicates that passage of the Farallon slab did not entirely remove mantle lithosphere and suggests conductive heating of the stable lithosphere.

For example, Roy et al. (2009) argue that uplift has occurred after lithosphere refrigerated by the Farallon slab was re-exposed to the convective asthenosphere in the middle Tertiary. With a thicker lithosphere than surrounding regions, the Colorado Plateau would heat not just from the lithosphere-asthenosphere boundary upward but also laterally in from its edges, hastening isostatic response. Their calculations require a temperature of ~600 °C at the base (150-200 km depth) of the lithosphere during refrigeration in order to subsequently produce 1 km of surface uplift. In the absence of lithospheric thinning, however, the base of the lithosphere would have cooled from being in contact with ~1300 °C asthenosphere. If we assume that the lithosphere reached thermal equilibration in each case, average temperature would have decreased from ~650 °C to ~300 °C. Density would have increased some 35 kg/m³, leading to more than 1 km (for a 100 km thick lithosphere) of air-filled subsidence during shallow subduction, creating accommodation space for several (~3) km of post-Cretaceous sediments that are not observed. The negative buoyancy of the slab itself would further this subsidence. A 75 km thick slab of oceanic lithosphere with an average temperature contrast of 300 °C with the underlying asthenosphere, as suggested by numerical modeling of Farallon subduction (Currie and Beaumont, 2011), would have an average density anomaly of at least ~30 kg/m³ and by eqn. 5.3 would generate 700 meters of additional subsidence (and space for ~2 km of sediment for a total of ~5 km).

Therefore we argue that heating is at least partially advective. Advection by large magnitude igneous intrusion is not supported by volcanic evidence or crustal thickness estimates, and we thus favor advection by lithospheric thinning. Cenozoic lithospheric thinning beneath the Colorado Plateau has previously been proposed: suggested mechanisms include sinking Rayleigh-Taylor instability (Levander et al., 2011), delamination (Bird, 1979), and ablation by the Farallon slab (Bird, 1988). If this thinning takes the form of removal of only the lower mantle lithosphere, then remaining depleted material could later be sampled by xenoliths (Spencer, 1996). Further, it is plausible that lithosphere was removed during emplacement of a flat slab, and the slab was removed at some later time. Indeed, if the topographic effect of removing lithosphere were similar to

that of emplacing the slab, then little change in surface elevation would be recorded until the slab is removed.

We now consider the case in which a thermal boundary (comprising both the crust and mantle lithosphere) of thickness z_{tbl} was thinned at some time, t=0, to a thickness of L by effectively instantaneous removal of its lower portion. (Note that this time t=0 could represent the time at which a Rayleigh-Taylor blob becomes unstable or the time at which the flat slab is removed.) The remaining lithosphere certainly begins to warm, but there is some uncertainty in the thermal consequences in the asthenosphere. The most conservative estimates of the topographic evolution of the Colorado Plateau will be derived from assuming that the asthenosphere begins to cool as a semi-infinite conductive body. The greatest magnitude of surface uplift would be derived if the asthenosphere were allowed to convectively remove heat from the interface with the lithosphere, maintaining a fixed temperature. The latter would require a change in steady-state basal heat flow, and the former ignores convection entirely, so we recognize that neither is a perfect formulation of the problem. Therefore, we pursue the more conservative possibility, noting that the amount of uplift for a given amount of lithospheric thinning is a minimum.

We assume that the lithosphere $(y=z_{ibl})$ begins in equilibrium with asthenosphere of temperature $T_a=1350$ °C and that the surface temperature is held constant at $T_s=20$ °C. If we also assume a linear geotherm, the temperature distribution prior to the moment of convective removal as a function of depth, y, is

$$T(y,0) = T_s + (T_a - T_s)y/z_{tbl}$$
(5.4)

An immediate pulse of uplift occurs in response to the removal of the thermally anti-buoyant material of thickness z_{tbl} -L. The magnitude of this uplift is controlled by this thickness. Since the mean temperature of the removed section is:

$$(T(L,0)+T_a)/2 = (T_s + (T_a - T_s)L/z_{tbl} + T_a)/2,$$
(5.5)

the mean temperature contrast with the asthenosphere is

$$T_{a} - (T(L,0) + T_{a})/2 = T_{a} - (T_{s} + (T_{a} - T_{s})L/z_{tbl} + T_{a})/2,$$
(5.6)

and the density anomaly with respect to asthenosphere is

$$[T_{a} - (T_{s} + (T_{a} - T_{s})L/z_{tbl} + T_{a})/2] \alpha \rho_{l}$$
(5.7)

where α is the coefficient of thermal expansion, and ρ_l is the reference lithospheric density. Since we assume that the mantle is compositionally homogeneous, $\rho_l = \rho_a$. Finally, following eqn. (5.3) the instantaneous uplift is:

$$U_{immediate} = [T_a - (T_s + (T_a - T_s)L/z_{tbl} + T_a)/2] \alpha \rho_l (z_{tbl} - L)/\rho_a$$

= $\alpha (T_a - T_s)/2 (z_{tbl} - L)^2/z_{tbl}$ (5.8)

As the asthenosphere begins to cool, the surface slowly subsides. The thermal structure in the remaining lithosphere returns to a linear geotherm. We assume that the temperature at depth z_{tbl} is held constant at T_a and the surface at T_s . The thermal history of the remaining lithosphere is calculated by separation of variables, where the temperature profile is the sum of the steady-state temperature, v(y), and a transient (decaying) perturbation, w(y,t), to that temperature:

$$T(y,t) = v(y) + w(y,t)$$

with v(y) as a linear geotherm from the surface to z_{tbl} :

$$v(y) = (T_a - T_s)y/z_{tbl} + T_s$$

The Fourier expansion of this system of equations is:

$$T(y,t) = v(y) + \sum_{n=1}^{\infty} c_n e^{-\frac{n^2 \pi^2 \kappa t}{L^2}} \sin \frac{n \pi y}{L}; \ c_n = \frac{2}{L} \int_0^L w(y,0) \sin \frac{n \pi y}{L} dy$$
(5.9)

Here, *L* is the modern lithospheric thickness, and w(y,0)=T(y,0)-v(y), or

$$w(y,0) = 0 \text{ for } y \le L$$

$$w(y,0) = (T_a - T_s)(z/z_{tbl}) \text{ for } L \le y \le z_{tbl}$$
(5.10)

The amount of uplift as a function of time is then the sum of the immediate uplift and the uplift or subsidence due to thermal equilibration:

$$U_{total}(t, L, z_{tbl}) = U_{immediate}(L, z_{tbl}) + \alpha L \int_0^L T(y, t) dy$$
(5.11)

 $U_{immediate}$ is given by eqn. (5.7), and T(y,t) is given by eqns. (5.9-5.10). This set of equations holds if a slab of subducting lithosphere replaced mantle lithosphere; $U_{immediate}$ is then controlled by the lithospheric properties while t=0 refers to the time at which the lithosphere or slab is removed (still considered instantaneous). These solutions are nearly identical to those of Bird (1979), who considered the process and consequences of delamination of continental mantle lithosphere.

The final result of this model is the interplay among previous lithospheric thickness, amount of material removed, and surface elevation as a function of time. Here, we again leverage the comparison between the Mesozoic Colorado Plateau and the modern Great Plains and investigate a range of paleothickness of 200-250 km--the lithospheric thickness in the modern midcontinent (Van der Lee and Nolet, 1997; Yuan et al., 2014). We explore the range of 50-150 km of material removed (i.e., lithosphere thinned to 50-200 km).

In order to discriminate among various combinations of previous lithospheric thickness, amount of lithosphere removed, and time at which the remaining lithosphere

was exposed to the asthenosphere, we sequentially leverage three separate calculations: uplift, reduced heat flow, and temperature at 150 km depth as functions of time. First, Figure 5.4a shows the uplift as a function of time for several combinations of initial lithospheric thickness and amount of thinning. In order to produce ~1 km of uplift, some ~100 km of material must have been stripped away. We now turn our attention to the predicted temperature at a depth of 120 km, where Hansen et al. (2013) suggest a temperature (at least under the northeastern part of the Colorado Plateau) of >1200°C. Only models with a previous lithospheric thickness of 200 km attain this temperature (Figure 5b). The effect of transient lithospheric thinning on surface heat flow is minor (Figure 5c) and is not a good discriminant among models; the change in heat flow is less than 10 mW/m² for all plausible models. We therefore favor a scenario in which 100-125 km of lithospheric material has been removed from a 200-km thick Colorado Plateau lithosphere during the Cenozoic.

Crustal density decrease

Xenoliths from ~20 km depth beneath the Four Corners region (Butcher, 2013; *in prep.*) record hydration, with major anhydrous phases--plagioclase and garnet--breaking down to form albite and hydrous phases like phengite and actinolite. Secondary monazite associated with the hydrated assemblage forms a dominant population (~60%) with latest Cretaceous-earliest Cenozoic (~90-60 Ma) Th-Pb crystallization dates. The remaining monazites are mainly Paleoproterozoic. The percent of garnet loss is uncertain, but estimates based on volumetric abundance of pseudomorphic amphibole are between 15% and 60%. Attendant density decrease is ~20-80 kg/m³ (0.75-3%), quite similar to the

modern mid-crustal density difference between the Colorado Plateau and Great Plains. The increasing density difference below 20 km (Figure 5.3) is consistent with hydration by fluids derived from below such as from the Farallon slab (e.g., Li et al., 2008; Smith et al., 1999).

Jones et al. (2011) have suggested that crustal hydration and especially garnet breakdown could be responsible for density loss and thus buoyancy increase since the Laramide, particularly in the Colorado Plateau and Great Plains. To illustrate, for a volume increase of factor V (V=1 implies no volume change), crustal thickness, z_c , would increase by a factor V, and density would change as 1/V. Topography, then, (using eqns. (5.1-5.3)) changes as:

$$\Delta \varepsilon = z_c \left(V - l \right) \tag{5.12}$$

For $\Delta \varepsilon$ =500 meters, and knowing that modern crustal thickness is 40 km, we find V=1.013, or 1.3% expansion, and we find that z_c =39.5 km. Thus only 500 meters of crustal thickening and an average density loss of 1.3% are required for 500 meters of elevation increase. Paleo-crustal densities would have been ~2806 kg/m³, similar to those in the modern Great Plains.

Following eqn. (5.3), the density differences in the 10-20, 20-30, and 30-40 km layers produce 91 (+/-22), 153 (+/-17), and 300 (+/-28) meters of topographic relief, totaling ~550 (+/-40) meters. If only half of this difference has developed since the late Cretaceous, then a minimum estimate of Cenozoic uplift by hydration is ~300 meters. Since the Plains themselves may be hydrated (Jones et al., 2011), these estimates of relief between the Colorado Plateau and the Great Plains are likely minimum estimates of the overall magnitude of surface uplift.

Synthesis

We have argued that Cenozoic uplift of the Colorado Plateau has two components: transient thinning of the lithosphere and crustal hydration. The principal timing constraint available to us is the 90-60 Ma period of monazite growth (Butcher, 2013), and this period is contemporaneous with passage/emplacement of the Farallon slab, suggesting that fluids were derived from dewatering oceanic lithosphere. Since several (~5) km of early Cenozoic sedimentation is not observed, we suggest that the slab replaced lower lithosphere of similar buoyancy and that the surface experienced little elevation change due to this replacement. Surface uplift would have occurred only in response to hydration, and volcanism would have been limited since the lithosphere remained insulated. Upon removal of the slab the surface would rise, and hydrated lithosphere would melt readily, potentially leading to the Oligocene-Miocene volcanism that samples this iron-depleted material (e.g., Alibert, 1990; Lee et al., 2001b; Smith, 2000; Smith and Riter, 1997). Thermal equilibration of the thinned lithosphere would then ensue, modestly raising the surface.

These two uplift mechanisms must sum to generate 2 km of surface uplift and 1-1.5 km relief with respect to the Great Plains. Crustal hydration plays a minor role--only some 300 meters--in relief generation, but this relief may be superimposed on broader scale crustal hydration that elevated Plains by some 450 meters (Levandowski et al., 2014a), leading to 750 meters of total Cenozoic Plateau uplift. The remaining ~1 km of relief and uplift is due to lithospheric thinning and heating. Early Cenozoic uplift by crustal hydration that is followed by mass redistribution in the upper mantle is consistent with uplift and/or drainage reorganization in the early Cenozoic (e.g., Flowers and Farley, 2012) and Neogene volcanism that samples continental lithosphere. Remaining lithosphere would warm (and potentially hydrate), leading to ongoing lateral erosion of the mantle lithosphere (e.g., Roy et al., 2009).

Conclusions

We have shown that lithospheric removal and subsequent heating of the remaining lithosphere combined with compositional density changes in the crust in the Colorado Plateau can produce both ~2 km of surface uplift since the Cretaceous and the modern difference in mantle velocity structure and surface elevation between the Colorado Plateau and the Great Plains. Thinning of the lithosphere by 100-125 km would account for 1 km of uplift, and a minor contribution from crustal density decrease provides ~300 meters of additional relief with respect to the Plains. More hydration-induced uplift is plausible if both regions were hydrated in the Cenozoic. Thus we argue for a purely isostatic mechanism for uplift of the Colorado Plateau, suggesting that Laramide-aged hydration is potentially cogenetic with lithospheric thinning, with both being due to emplacement of the Farallon slab. Subsequent removal of the slab triggered isostatic adjustment of the overlying surface and then a protracted period of thermal equilibration with the asthenosphere.



Figure 5.1:

(a) Flexurally smoothed elevation of the western US. Thin black lines are state boundaries. Bold black lines in this and other figures are boundaries of physiographic provinces, which are labeled. SN: Sierra Nevada, NBR: Northern Basin and Range, SBR: Southern Basin and Range, SRP: Snake River Plain, CP: Colorado Plateau, WC: Wyoming Craton, SRM: Southern Rocky Mountains, GP: Great Plains.

(b) Crustal thickness (Shen et al., 2013).



Figure 5.2: Variations in surface elevation due to mantle thermal structure (the mean has been removed to facilitate comparison). There is little difference in mantle buoyancy among the forearc, Cascades, Sierra Nevada, Basin and Range, Snake River Plain, Colorado Plateau, and Southern Rockies.

- ^{-0.7} Wyoming craton and Great Plains
- is substantially denser.



Figure 5.3: Mean densities of crustal layers. Areas with crustal thickness less than the median depth for each layer are blacked out. Note that the difference in density between the Colorado Plateau and Great Plains increases with depth. Average values in kg/m³ for these two provinces are shown.





Figure 5.4:

(a) Uplift vs. time for varying lithospheric thickness and amounts of thinning (following eq. 5.11). Initial lithospheric thickness of 250 km is shown in blue, 200 km in black. Note that only thinning of 100-125 km produces 1 km of uplift at any point in the Cenozoic.

(b) Temperature at 120 km depth for the four plausible curves from (a)--100 and 125 km thinning of either 200 or 250 km lithosphere. Note that only 200 km thick lithosphere ever reaches a temperature near the estimated temperature.

(c) Reduced heat flow evolution. Note that maximum change is $\sim 10 \text{ mW/m}^2$.

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Appendix A: Receiver function images from the southern Sierra Nevada

Overview maps are provided below, and beneath each receiver function image, the seismic station names, locations and elevations are given.


Figure A.1 Crustal thickness in the Sierra Nevada region, showing location of Figure A.2-A.4



Figure A.2: Geometry of beams, overlain on basemap from Zandt et al. (2004). Beams are labeled, and comprise the seismic stations located beneath the vertices.



Figure A.3: Rough geometry of transects shown below



Figure A.4: Beams displaying a mid-crustal negative arrival are outlined in magenta. Note NNE-SSW strike.

Beam-formed receiver functions from the western foothills

Of particular note is the step from arrivals at ~6s on beams NW1, NW2, WC1, and WC4 to arrivals at <5.5s on SW1 and SW3. The Moho P-s conversion is visible on most beams from the northwest and southwest, but is of low amplitude (compare to eastern Sierra beams, below). Southeastern backazimuth events show no, or lower amplitude still, Moho P-s, suggesting profound variations in crustal thickness beneath the western foothills at wavelengths comparable to station spacing. Note the profound mid-crustal negative SE of SW3.



NW1 BGR: 36.63°N, 119.02°W, 954 METER BRR: 36.91°N, 119.04°W, 1259 METERS HVY: 36.7°N, 119.32°W, 193 METERS



NW2 BGR: 36.63°N, 119.02°W, 954 METER BRR: 36.91°N, 119.04°W, 1259 METERS CPR: 36.8°N, 118.58°W, 1603 METERS



WC4 CCC: 35.52°N, 117.36°W, 670 METERS CPR: 36.8°N, 118.58°W, 1603 METERS BGR: 36.63°N, 119.02°W, 954 METER



WC1 CCC: 35.52°N, 117.36°W, 670 METERS BGR: 36.63°N, 119.02°W, 954 METER LMC: 36.36°N, 119.03°W, 211 METERS



SW1 LMC: 36.36°N, 119.03°W, 211 METERS WMD: 36.2°N, 118.58°W, 2592
 METERS PDC: 36.03°N, 118.98°W, 167 METERS



SW3: LMC: 36.36°N, 119.03°W, 211 METERS WMD: 36.2°N, 118.58°W, 2592
 METERS MKW3: 36.45°N, 118.61°W, 2358 meters

Central Sierra

Two notable patterns emerge on a transect from north to south through the central Sierra. First, the step in Moho P-s arrivals (here from ~5.5s to <5s) is once again observed near 36.5°N. Second, large, negative mid-crustal signals (between 1.5 and 4s) are observed on NC1, SW2, and SC1. Possible negatives also appear on HS3 and HS4, but--qualitatively--these receiver functions are poor.



NC1 FLL: 37.28°N, 118.97°W, 2237 METERS SRF: 36.97°N, 118.63°W, 1807 METERS BRR: 36.91°N, 119.04°W, 1259 METERS



NC3 CPR: 36.8°N, 118.58°W, 1603 METERS SRF: 36.97°N, 118.63°W, 1807 METERS BPC: 37.13°N, 118.43°W, 2370 METERS



HS3 CPR: 36.8°N, 118.58°W, 1603 METERS SRF: 36.97°N, 118.63°W, 1807 METERS OVY: 36.78°N, 118.33°W, 2704 METERS



HS4 CPR: 36.8°N, 118.58°W, 1603 METERS CCC: 35.52°N, 117.36°W, 670 METERS JUN: 36.58°N, 118.41°W, 2471 METERS



SW2 WMD: 36.2°N, 118.58°W, 2592 METERS TWR2: 36.35°N, 118.41°W, 1946

METERS MKW3: 36.45°N, 118.61°W, 2358 METERS



SC1 SFT: 36.23°N, 118.06°W, 1753 METERS ARC2 WMD: 36.2°N, 118.58°W,
2592 METERS TWR2: 36.35°N, 118.41°W, 1946 METERS

Eastern Sierra

A transect along the eastern front of the Sierra displays two important patterns. First, the apparent Moho step near 36.5°N is absent; P-s conversions arrive consistently at 4-5s along the entire eastern front. Secondly, the negative arrivals between 1 and 4 seconds that were observed mainly on stations in the south are more pervasive. Negative arrivals can be seen (from north to south) or NC2 from the northwest (3.6s), speculatively on NE1 from the southwest (0.8s and 2.1s, but only two events populate this backazimuth and stack is unreliable), NE2 from all backazimuths (3.6s), HS1 from northwestern and southwestern events (3.6s), and ES1 and ES2 from the northwest and southeast (~3.5s from NW and 2.5s from SE).



NC2 FLL: 37.28°N, 118.97°W, 2237 METERS SRF: 36.97°N, 118.63°W, 1807 METERS BPC: 37.13°N, 118.43°W, 2370 METERS



NE1 SRF: 36.94°N, 118.11°W, 2148 METERS SRF: 36.97°N, 118.63°W, 1807 METERS BPC: 37.13°N, 118.43°W, 2370 METERS



NE2 SRF: 36.94°N, 118.11°W, 2148 METERS OVY: 36.78°N, 118.33°W, 2704 METERS BPC: 37.13°N, 118.43°W, 2370 METERS



HS1 WHP: 36.59°N, 118.22°W, 2388 METERS JUN: 36.58°N, 118.41°W, 2471 METERS OVY: 36.78°N, 118.33°W, 2704 METERS



ES2 WHP: 36.59°N, 118.22°W, 2388 METERS JUN: 36.58°N, 118.41°W, 2471 METERS TWR2: 36.35°N, 118.41°W, 1946 METERS



ES1 WHP: 36.59°N, 118.22°W, 2388 METERS TWR2: 36.35°N, 118.41°W, 1946 METERS SFT: 36.23°N, 118.06°W, 1753 METERS

Mid-crustal negative

As seen in Figure A.4, there is a negative mid-crustal arrival that strikes subparallel to the Sierra Nevada, SSW-NNE. Generally, this negative arrival deepens to the north (see eastern transect), though some eastward dip also appears (compare SW3, SW2, and ES1/ES2). A mid-crustal negative has previously been observed in the eastern Sierra (Jones and Phinney, 1998; Zandt et al., 2004) and was interpreted as shear zone, where Basin and Range extension is encroaching on the Sierra Nevada. Indeed, the NNE strike of this negative arrival closely parallels the Kern Canyon Fault through the southernmost Sierra.