Imaging the Great Plains of the Central U.S. using Finite-Frequency Rayleigh Wave Tomography and Implications for Asthenosphere-Driven Uplift

by

Rachel Margolis

A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE

Master of Science

APPROVED, THESIS COMMITTEE:

Alan Levander, Carey Croneis Professor
Earth Science

Colin Zelt, Professor
Earth Science

Adrian Lenardic, Professor
Earth Science

HOUSTON, TEXAS
DECEMBER 2014
ABSTRACT

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Here we present a 3D shear velocity model for the lower crust and upper mantle beneath the Great Plains in the central United States using finite frequency Rayleigh wave tomography. We use USArray Transportable Array recordings of teleseismic Rayleigh waves and first invert for phase velocity using the two-plane wave method with finite frequency kernels, then invert the resulting dispersion curves for shear velocity structure. We characterize the lithospheric structure in this tectonically transitional regime to illuminate the differences between the actively deforming west and stable continental interior. The west is defined by slow velocities and thin lithosphere, whereas the east has fast velocities and thick lithosphere, with the thickest lithosphere under the Superior craton. From our tomography and heat flow data, we infer warm temperatures in the west and suggest that the asthenospheric mantle contributes to anomalously high elevation in the west with secondary contributions from crustal effects.
ACKNOWLEDGEMENTS

I am grateful to my wonderful support systems both professionally and personally that guided me through my graduate experience. Firstly, I would like to thank my advisor Dr. Alan Levander for giving me the opportunity to study and research at Rice University. Dr. Levander provided interesting discussions and expertise valuable to the completion of my thesis work. I would also like to thank the members of my thesis committee for giving their valuable time and feedback.

The S.T.A.R. (Seismology and Tectonics at Rice) group has been a constant source of insight, and it has been a pleasure learning from everyone. I want to specially acknowledge my officemate Sally Thurner who leant her great knowledge of the study region on numerous occasions as well as stimulating discussions on all topics and to post-doc Imma Palomeras who was very patient in introducing me to the code packages used in this thesis.

This work would not have been possible without the constant support of my family, especially my sisters Emily, Miriam and Sarah Margolis and my boyfriend Thomas K. Langin. Thank you all for being there through the thick and thin and always having confidence that I will succeed.
# TABLE OF CONTENTS

Title Page.........................................................................................................................i

Abstract...............................................................................................................................ii

Acknowledgements.........................................................................................................iii

Table of Contents..........................................................................................................iv

Chapter 1 Introduction.................................................................................................1

1.1 Motivation ..............................................................................................................1

1.2 Geologic setting of the Great Plains....................................................................2

1.3 Elevation.................................................................................................................3

Chapter 2 Data Processing.........................................................................................6

Chapter 3 Finite-frequency Rayleigh wave tomography.................................................8

3.1 Modified Two-Plane Wave Tomography.............................................................8

3.1.1 Two-Plane wave parameters.........................................................................8

3.2 Shear velocity inversion......................................................................................10

3.2.1 Resolution kernels and Damping Parameter..............................................11

Chapter 4 Shear Velocity Structure..........................................................................14

Chapter 5 Interpretation..............................................................................................17

5.1 Lithosphere asthenosphere boundary.................................................................17

5.2 LAB topography relationship to surface topography........................................19

5.3 Temperature of the Great Plains........................................................................22

5.3.1 Surface Heat Flow.......................................................................................22

5.3.2 Inferring temperature from seismic velocity..............................................22

5.4 Elevation..............................................................................................................25
5.4.1 Asthenospheric contribution to elevation…………………………………26
5.4.2 Crustal refinements to elevation………………………………………..28
5.5 Comparison of results………………………………………………………32

Chapter 6 Discussion and conclusions…………………………………………33

List of References………………………………………………………………35
CHAPTER 1
INTRODUCTION

1.1 Motivation

The goal of this study is to determine the lithospheric structure of the Great Plains and adjacent regions, including the recently deformed easternmost Southern Rocky Mountains (SRM) and Rio Grande Rift (RGR), the western High Plains and the stable continental craton of the east. Several regional tomographic studies of the North American mantle suggest a marked contrast in seismic shear wave speeds between the stable continental core and regions of active deformation (Bedle and van der Lee 2009, Burdick et. al. 2010, Marone et. al. 2007, Nettles and Dziewonski 2008, Porritt et. al. 2013, Sigloch 2011, van der Lee and Frederikson 2005, van der Lee and Nolet 1997). Before the installation of the USArray Transportable Array there was a very sparse station distribution in the midcontinent, and so only the most recent studies have the resolution to image the details of the transition between the western and eastern US, and despite the new models, a full study of the differences in lithospheric character between these two distinct regimes does not exist.

The interpretation of the features of the shear velocity model of the central US produced in this study suggests a relationship between the lithospheric structure and the high surface elevations observed in the westernmost Great Plains. There have been numerous studies discussing the topography of the western US and SRM with investigation of such mechanisms as climate and erosional feedback, tectonics, crustal structure, mantle structure and mantle flow (Coblentz et. al. 2011a; Coblentz et. al.
2011b; Eaton 2008; Hansen et. al. 2013; Humphreys and Dueker 1994; Nereson et. al. 2013, Levandowski et. al. 2014, Li et. al. 2002). To our knowledge, there has been no such analysis east of the Rocky Mountain Front (RMF). This study examines the role of thermally expanded asthenosphere under a thin lithosphere and includes the secondary effects of crustal isostasy and crustal thermal expansion as compensation mechanisms in the SRM and High Plains.

1.2 Geologic setting of the Great Plains

Most of the history of the Great Plains is a story of stability after the initial formation of North America, with only very recent modification of the west producing relief that distinguish this region from the low-lying mid-continent. The Great Plains contains provinces ranging in age from Archean to Mesoproterozoic. Several Archean cratons form the core of North America: the Rae-Hearne, Wyoming, and Superior. These distinct Archean units accreted during the Trans Hudson Orogeny between 1.85-1.78 Ga, forming a deformational zone several hundred kilometers wide (Whitmeyer and Karlstrom 2007). The subsequent tectonic history of North America involved a series of juvenile oceanic terranes accreting onto the southwestern edge of the cratonic core, growing the landmass southward. The major accretionary periods were: (1) the Yavapai Orogeny (1.71-1.68 Ga), (2) the Mazatzal Orogeny (1.65-1.60 Ga), (3) the Granite Rhyolite province accretion (1.55-1.30 Ga), (4) and the Grenville Orogeny (1.3-0.9 Ga). The Cheyenne belt preserves the accretionary boundary of the 1.78-1.75 Ga collision between the Wyoming province and Yavapai province, and the Jemez lineament preserves the accretionary boundary between the Yavapai and Mazatzal provinces. A
major break up occurred from 1.2-1.1 Ga, resulting in the failed Midcontinent Rift that extends from Kansas to the Great Lakes (Green 1983).

Phanerozoic events have also shaped the geology of the interior of the continent, most notably, the formation of the Ancestral Rocky Mountains at 300 Ma. The subduction of the Farallon slab initiated the formation of the Sevier fold-and-thrust belt along the western margin of the US at 150 Ma (Eaton 1987). At ~80 Ma, the Laramide Orogeny commenced in the western US (Gao 2004) and at 40 Ma, slab rollback toward the west initiated, with asthenospheric mantle filling the void left by the slab.

1.3 Elevation

The western Great Plains abut the Southern Rocky Mountains where elevation exceeds 3 km, and gently slopes to the low elevations (approximately sea level) in the interior of the US (Figure 1.1). The Great Plains have an unusual elevation profile that possesses a much broader region of uplifted elevation and lower relief than the piedmonts of other orogenic systems (Figure 1.2). Early work suggested that this topography resembles the bathymetry at a mid-ocean spreading ridge (Eaton 1987, Eaton 2008). Some of the topography of this region was contributed by the Laramide Orogeny and associated crustal thickening (Bird 1984), but the somewhat thickened crustal root underneath the Rocky Mountains is insufficient to explain the highest elevation of the SRM and the broad uplifted region away from the mountain front (Karlstrom et. al. 2012). This residual topography suggests the existence of other contributions to elevation besides orogenic uplift. Recent geochemistry work supports the hypothesis of post-
Laramide uplift: analysis of volcanic flows around the Jemez lineament indicates uplift of the surface ongoing for at least the past 5 Ma (Nereson et. al. 2013).

Fig. 1.1 Topographic map of the central United States with the study region outlined by the black rectangle. The white solid lines indicate the major age boundaries: the Superior craton, Trans-Hudson Orogen (THO), Wyoming Craton, Yavapai Province, Mazatzal Province, Granite Rhyolite Province. The white dashed lines indicate the Southern Rocky Mountains (SRM), Rio Grande Rift (RGR), Grenville deformation front and MidContinent Rift (MCR). The USArray Transportable Array stations used in this study are shown as black triangles.
Fig. 1.2 Elevation along the flanks of major orogens taken from the base of each mountain. Each elevation profile is fixed to 0 km at 1000 km for distance comparison from the deformation front (adapted from Eaton 2008)
CHAPTER 2
DATA PROCESSING

Vertical-component seismograms from 451 USArray TA stations were downloaded from the IRIS Data Management Center. A total of 208 shallow-focus ($\leq$70 km depth) teleseismic events with magnitude $M_w \geq 5.5$ that occurred between 01-01-2008 and 06-01-2012 were used in the analysis. The epicenters of the events are located between 30° and 120° from the center of the study region to ensure that the Rayleigh wave phases are sufficiently separated from S waves so that the fundamental mode Rayleigh waves can be identified on the seismograms (Figure 2a).
**Fig. 2.** Azimuthal Distribution of the 208 earthquakes (red dots) used in this study. Ray paths (green lines) are plotted from each event to the center of study region at 40 N, -100 W (a). Number of raypaths for each period (frequency) band from 20 to 167 s (b).

We applied the following workflow to the raw data: (1) traces were resampled to 20 Hz, (2) instrument responses were removed, (3) and a series of narrow band 4th-order Butterworth filters were applied to the data, producing seismograms centered in 18 different frequency bands from 0.006 Hz (167 s) to 0.05 Hz (20 s) with a 0.01 Hz (10 s) bandwidth. The fundamental mode Rayleigh wave was isolated by selecting the waveform by hand and windowing the time series records using a cosine taper. Traces that were noisy or windowed inappropriately were discarded, and the most high quality traces were those at the middle frequencies (Figure 2b).
CHAPTER 3
FINITE-FREQUENCY RAYLEIGH WAVE TOMOGRAPHY

3.1 Modified Two-Plane Wave Tomography

Surface waves can be scattered by heterogeneity along the path from the source to the receiver, resulting in multipathing and a nonplanar wavefield, as evidenced by the complicated recordings at the receivers. Here we employ the two-plane wave method to model the observed wavefield as the result of two interfering plane wavefields (Forsyth and Li 2005). For each earthquake, first simulated annealing is performed to calculate the six parameters that best describe the incoming two-plane wavefield (amplitude, phase, and deviation from great circle path for both the primary and secondary waves). In a second stage, the wave parameters and the phase velocity model are updated using an iterative linearized inversion procedure to yield final phase velocities.

Each wave samples structure beyond an infinitesimal ray path and these finite frequency effects are accounted for by calculating 2D phase and amplitude sensitivity kernels (Yang and Forsyth 2006). Accounting for the off-ray sampling increases the lateral resolution of features with a similar scale as the wavelength. The calculated sensitivity of both the phase and amplitude is greatest at the receiver and decreases away from it, representing Fresnel zone structure.

3.1.1 Two-plane wave parameters

The two-plane wave method is appropriate only for small regions where the curvature of the earth does not invalidate the assumption of a planar wavefront.
Therefore, in our large study region of 20°x12°, we have divided the region into eight partially overlapping subregions of dimension either 7.5°x6.5° or 7.5°x8.0° and calculated the phase velocities within each separately. The phase velocities in the zones of overlap between the subregions were averaged and smoothed to produce a final phase velocity map for the entire study region.

In the two-plane wave method the incoming wavefield is divided into a primary wave, with a path roughly coincident with the great circle arc path between the source and the receiver, and a secondary wave. Additionally, the secondary wave amplitude should be much smaller than the primary wave, leaving the ratio of secondary to primary <1. In Figure 3.1 we show histograms for these parameters for a 45 s wave in one subregion. The mean amplitude ratio is 0.3, the mean primary wave angle deviation is -0.8°, and the mean secondary wave angle deviation is -4.4°. Additionally we show the observed versus predicted amplitude, and observed versus predicted phase. The predicted amplitudes are slight overestimates when compared to the observed amplitudes at the 121 recording stations, but still trend along a line with a slope of 1. The predicted phases and observed phases also are very similar, and approximate a line of slope 1 with less spread than the amplitude trend. This analysis demonstrates that the two-plane wave assumption is a good approximation to the observed data.
Fig. 3.1 The top row shows histograms of parameters for the two-plane wave description for a 45 s wave. The amplitude ratio is less than 1, the primary wave deviation angle from the great circle arc peaks at 0, and the secondary wave deviation from the great circle arc has more variation and is approximately symmetric about 0. The bottom row shows the predicted versus observed amplitudes and phases calculated at each of the 121 stations that recorded an event on March 7, 2010. These are typical of the dataset.

3.2 Shear velocity inversion

The 1D phase velocity curves calculated by the two-plane wave method serve as the observed data input into an inversion for shear velocity structure using the DISPER80 code package that calculates normal mode solutions (Saito 1988). The inversion for shear
velocity proceeds by perturbing phase velocity and calculating changes in earth model parameters from the initial 1D layered earth model (with density, P-wave velocity, and S-wave velocity). Using a grid parameterization of 0.25°x0.25°, 3969 unique initial 1D models were used for the inversion. Each initial model averaged P and S wave velocity values from Crust2.0 in the upper 50 km (Bassin et. al. 2000), and P and S wave velocities from regional body wave tomography from 50-400 km (Schmandt unpublished), with densities from the global model AK135 (Kennett et. al. 1995). An estimate of the Moho at each node was obtained from PdS Receiver Functions (RF) and also served as input into the inversion (Thurner et. al. in prep). The final 3D model is the synthesis of all 1D shear velocity profiles.

3.2.1 Resolution kernels and Damping Parameter

The predicted phase velocities generated in the shear velocity inversion procedure respond when each of the Vs, Vp, and densities of an initial model or an iteration of the model are perturbed. This response varies at different periods, and we plot the sensitivities of the phase velocity at a representative inversion node in Figure 3.2. The sensitivity of phase velocity to shear velocity is the greatest, followed by compressional velocity, and then is the least sensitive to density by a large degree. The peak of the curve represents the depth at which the wave of a given period has the most sensitivity. For all parameters, the resolution kernels peak deeper than the surface and decay rapidly with depth, leading to the most well resolved depths from 25 km to 250 km in our study. In the right-most panel of Figure 3.2 we show the rows of the resolution matrix versus depth for a representative inversion node. This plot illustrates the resolution of the shear wave
model and indicates that no layer is independent. The broadness of the peak indicates the relative number of layers adjacent to the depth of interest necessary to calculate an independent value of the model. Note that the layer thickness in both the initial and final models are nonuniform: the crustal layers are thinnest (5 km, 7.5 km, 7.5 km, 10 km and 10 km) and the mantle layers increase in thickness to 50 km.
Fig. 3.2 Sensitivity of phase velocity with respect to model parameters density (left), compressional velocity (second to left), and shear velocity (second to right) at center periods 20 s, 40 s, 67 s, 100 s, and 167 s. Rows of the resolution matrix for a representative inversion node (right).

Regularization parameters are often used in inversion methods to minimize the model norm and produce a smooth model. Here we used a damping parameter of 0.20 determined by an L-Curve test in which the optimum tradeoff between the norm of the model (Vs) and data residuals (phase velocity) was calculated (Aster et. al. 2013).
CHAPTER 4
SHEAR VELOCITY STRUCTURE

Figure 4 displays the average 1D structure and the shear velocity anomaly maps at 40 km, 60 km, 80 km, 100 km, 125 km, 150 km, 200 km, and 250 km depth. The anomaly at each depth is with respect to the average velocity of that depth (3.95 km/s, 4.54 km/s, 4.58 km/s, 4.55 km/s, 4.52 km/s, 4.50 km/s, 4.48 km/s, 4.50 km/s respectively). At all depths there is a striking contrast between the convex fast velocity anomaly (the Superior craton and eastern portions of the Proterozoic terranes) and the concave slow velocity anomaly in the west. The lowest velocities are concentrated under the SRM while the fastest velocity is in the northeast in the Superior craton.

At 40 km, a predominantly lower crustal depth, the region is characterized by a large negative velocity anomaly that extend as far southward as the Grenville Deformation Front, and throughout the Yavapai and Mazatzal Provinces. This negative anomaly is centered about the SRM and is discontinuous with the negative anomaly in the northern part of the Wyoming craton and THO. The Superior craton, Granite Rhyolite province, and the Gulf Coast are markedly different because they are characterized by a positive velocity anomaly.

In the uppermost mantle (from 60 to 100 km), the negative velocity anomaly is stronger and reflects longer wavelength features, suggesting increased heterogeneity in the velocity structure. The lowest velocities are concentrated under and slightly east of the Southern Rocky Mountains with the shape of the low velocity anomaly distinctly concave to the west. There are three extensions of the low velocity body that trend
roughly northeast: one along the Cheyenne belt, a second through the Mazatzal Province, and a third through the Yavapai Province. The presence of low velocity in localized zones roughly corresponding to accretionary boundaries suggests that these boundaries are features that are preserved over long time scales and that they remain zones of weakness where asthenosphere may preferentially flow (Karlstrom and Humphreys 1998; Magnani et. al. 2005). Similar low velocity features have been seen in previous studies beneath the Snake River Plain, Saint George volcanic trend and the Jemez lineament, all of which also parallel Proterozoic boundaries (Karlstrom and Humphreys 1998). The northern and eastern regions of the Great Plains have an overall fast velocity structure, with the fastest velocities concentrated in the northeast, coincident with the Superior craton. In the Gulf Coast region there are low velocities that are unlikely to be related to those under the SRM.

At 200-250 km, the high velocity anomaly in the northeast is evidence of the lithospheric root beneath the Superior craton. Even at this depth, the study area is not homogeneous, suggesting differences in structure between the Superior craton and the west that persist through the upper mantle. The Wyoming craton exhibits no clear root, which suggests that it may have a history of modification possibly very early lithospheric removal associated with a ~2.0 Ga rifting event (Karlstrom and Humphreys 1998) or Phanerozoic tectonism (Henstock et. al. 1998).
Fig. 4. Average shear velocity structure over the study region (left panel). Shear velocity anomaly maps for 40 km, 60 km, 80 km, 100 km, 125 km, 150 km, 200 km, and 250 km depth. The anomaly is taken with respect to the average shear velocity for that depth. The color bar for the top row is from -10 to 10% velocity perturbation, and the color bar for the bottom row is from -5 to 5% velocity perturbation. The tectonic and age boundaries are labeled: Superior craton (Sup), Wyoming craton (WC), Trans-Hudson Orogen (THO), Yavapai province (Yav), Mazatzal province (Maz), Southern Rocky Mountains (SRM), Rio Grande Rift (RGR), Mid-Continent rift (MCR), and Gulf Coast (GC), Grenville deformation front (GF). The extensions of the low velocity body are shown at 100 km with the double-headed arrows.
CHAPTER 5
INTERPRETATION

5.1 Lithosphere asthenosphere boundary

The boundary between the lithosphere and asthenosphere is the region where heat transfer changes from convective to conductive, the former associated with the asthenosphere and the latter with the rigid plate (Eaton 2009; Fischer et. al. 2010; Levander and Miller 2012; Rychert and Shearer 2009). This thermal regime discontinuity is not well understood (see for example Lee et al., 2005) and has been identified using a multitude of geophysical proxies (Eaton 2009). The seismic definitions of the LAB include a change in velocity from a fast “lid” to a deeper slow asthenosphere, a negative impedance contrast on PdS and SdP RFs, or a change in seismic anisotropy. Here we identified the seismic LAB using two criteria and then smoothing the final result: (1) the midpoint between the depth of the maximum shear velocity (the lid) and the depth of the minimum shear velocity in the uppermost mantle which we take as the center of the asthenospheric channel, and (2) absolute velocity contours (Eaton 2009). Surface waves are sensitive to average bulk velocity which enables our tomography to detect only gradual transitions (see the broad peaks of the resolution kernels in Figure 3.2), and the LAB is most likely a diffuse boundary.

In Figure 5.1 we plot the final LAB depth picks on several latitudinal profiles. In the lower latitudes the LAB depth estimates of the averaging method are very plausible, but at higher latitudes the estimate appears too shallow. This is a limitation of the averaging method because if the LAB is deep, the depth of the minimum Vs is beyond
the depth resolution of our dataset and therefore the method calculates the midpoint as too shallow. Numerous studies show that LAB estimates are difficult to achieve in cratonic regimes (Eaton 2009), we find a reasonable value by interpreting the LAB under the craton by following the 4.6 km/s velocity contour. At the highest latitudes in the east (see 48 N in Figure 5.1) this contour extends to the limit of resolution of the surface wave tomography (~250 km), which suggests that the LAB may be even deeper than interpreted. A greater lithospheric depth estimate would be consistent with the ~350 km estimate of Van der Lee and Nolet 1997, which is similar to thicknesses of cratons worldwide (Artemieva 2009; Yuan and Romanowicz 2010).
Fig. 5.1 Latitudinal profiles (32, 34, 36, 38, 40, 42, 46 and 48 North) with the LAB shown as white dots overtop the shear velocity structure.

The LAB depth map is given in Figure 5.2a. A first order trend is the thickening of the lithosphere radially away from the RMF. An exception to this pattern is the thin lithosphere around the Gulf Coast (coincident with the arc of low velocities in the southern part of the study area in Figure 4); the LAB depth here is influenced by the lithospheric thickness of the oceanic basin that has an upper limit controlled by convective instability (Nagihara et al. 1996). The thin lithosphere underneath the Gulf Coast has also been inferred from a 2D seismic transect (Ainsworth et al. 2014) and from a study of seismic anisotropy (Gao et al. 2008). The lithosphere is deepest in the northeastern region (reaching a maximum of ~250 km), which corresponds to the keel of the Superior craton. The lithosphere under the Wyoming craton is thinner than the Superior, with no deep keel. Our results are in agreement with a previous detailed survey of the thermal and seismic LAB of cratons (Artimieva 2009; Yuan and Romanowicz 2010).

5.2. LAB topography relationship to surface topography

The shape of the LAB mimics the arc in the surface topography contours, as both decrease radially away from the RMF. For most of the region away from the coast, where there is thin lithosphere there is high elevation and vice versa. If we consider depth and elevation both as positive quantities, the 2D correlation coefficient between the LAB and the surface elevation is 0.61, and is higher, 0.76 when only including latitudes greater
than 34 N. This selection excludes the Gulf Coast where we do not believe there is a correlation between the LAB and elevation because the main control on LAB thickness is the transition from oceanic to continental structures.

We take several diagonal transects that originate at (39 N, -106 W) and cut roughly perpendicular to the contours of constant elevation, and therefore are examining profiles that are almost monotonically decreasing functions of elevation (Figure 5.2b). The 1D correlation coefficients are 0.71, 0.86, 0.93, 0.90, 0.94 for profiles A, B, C, D, and E respectively which suggests a clear correlation between the two observables. Profile A and B have slightly lower coefficients because their southeastern ends truncate around the Gulf Coast. Due to this correlation analysis, we hypothesize that the thickness of the lithosphere may have an effect on features at the surface, and we discuss this in the following section.
Fig. 5.2 LAB map with the average depth and standard deviation displayed (a). Elevation (b top) and lithospheric depth (b bottom) along 5 diagonals profiles each originating at (39 N, -106 W). Locations of the profiles are shown by the white dashed lines in (a).
5.3 Temperature of the Great Plains

5.3.1 Surface heat flow

A characterization of temperature variation among tectonic provinces is important in understanding the difference between the western and eastern US. Regions of active tectonics are normally associated with elevated temperatures and high surface heat flow (HF), like in the SRM. Whereas stable continents are associated with lower temperatures and low surface HF, as in the Superior craton (Sclater et al. 1980). While the province-wide trends are general, specific HF measurements should be used with caution as the HF measured at the surface may be influenced by shallow crustal sources such as radiogenic compounds in groundwater (Gosnold 1990). HF measurements within the Great Plains average between 40 and 60 mW/m\(^2\) (Gosnold 1990); the lowest HF values are characteristic of cratons (Chapman 1986), and the measurements available in the Superior craton are within that range (http://smu.edu/geothermal/georesou/alldata.htm). The Rocky Mountains have an elevated surface HF of 80-100 mW/m\(^2\) and the RGR branch in the western Plains has an average HF of 72 mW/m\(^2\) (but with a very high standard deviation of 35 mW/m\(^2\), suggesting a complicated thermal state in the RGR). In general, greater HF measurements are coincident with the tectonically active SRM and RGR, and the HF measurements are considerably less in the stable interior.

5.3.2 Inferring temperature from seismic velocity

Tomographic models suggest a strong relationship between seismic velocity and temperature since fast velocities are generally associated with the cold, stable continent and slow velocities are associated with zones of active deformation, volcanic activity and
hot spots. Quantifying the effect temperature has on seismic velocity presents many challenges because there are possible second-order effects due to composition, phase transformation, melt, attenuation, seismic frequency, anisotropy, and the presence of fluids (Karato 1993; Sobolev et. al. 1996). However, it has been shown that temperature has the primary influence on seismic velocity (Sobolev et. al. 1996), and therefore estimating temperatures from velocity without including other parameters is a good first-order approximation.

Laboratory experiments show that shear moduli sensitivities to temperature have very little dependence on composition as summarized by Lee 2003. For our calculations, we will assume a mantle composition of pure forsterite since the composition of the earth’s upper mantle is predominantly Mg-rich olivine (McDonough and Sun 1995). An experiment on synthetic forsterite determined the shear modulus sensitivity to temperature of this end-member species of olivine at near-mantle temperatures and pressures (Faul and Jackson 2005; Jackson and Faul 2010). The shear modulus behaves approximately linearly at temperatures below the mantle adiabat but at higher temperatures, the behavior becomes somewhat nonlinear. This change is due to anelastic effects or partial melt which can cause a decrease up to 8% in shear velocity per 1% melt fraction. It is difficult, however, to constrain the effect of melt on seismic velocity as estimates of the decrease in Vs vary widely in the literature (see Hansen et. al. 2013 and references therein).

We calculate the temperature perturbations relative to a reference temperature at constant pressures (i.e depth). We begin with an end-member calculation considering temperature effects only. We later provide a brief discussion of the corrections required if
(1) a small percentage of melt is assumed and (2) a mantle with multiple compositions (reflected in depletion or enrichment in Fe) was considered. At any depth the temperature perturbation relative to a reference is

$$\Delta T = \left( V_{s_{obs}} - V_{s_{ref}} \right) \frac{\partial V_s}{\partial T}$$

Where $V_{s_{obs}}$ is the observed velocity and $V_{s_{ref}}$ is the reference velocity that is an average of a subsection of the craton (since the craton is stable we expect no positive thermal anomalies here). Here we use a $\partial V_s/\partial T$ value of -0.00077 km/s °C that is calculated for 1 mm diameter olivine at 100 km (Hansen et. al. 2013; Jackson and Faul 2010).

An overestimation of the temperature anomaly could occur if the SRM and surrounding areas of very low velocity had some percentage of partial melt. For example at 100 km, the reference velocity is 4.67 km/s and parts of the SRM have velocities of 4.2 km/s. The calculated temperature anomaly is 610°, which is consistent with Goes and van der Lee, 2003, in which the authors inverted the NA00 shear velocity model for absolute temperature: they found temperature contrasts around 600° at 110 km depth. However, if 8% of the variation in velocity could be attributed to partial melt, then the temperature difference we calculate would only be 561°, rendering our original estimation ~50° too high. In conclusion, if even a small amount of partial melt is present then the uncertainty in the temperature calculations can be 10% or greater. An estimation of partial melt percentage is challenging since it is highly dependent on the geometry of the melt (Hammond and Humphreys 2000; Shankland et. al. 1981), so here we assume that there is no partial melt. This enables us to place an upper limit on the temperature anomalies from our calculations.
The sensitivity of shear velocity to composition is small with a $\frac{\partial V_s}{\partial Mg\#}$ value of $\sim 0.0143$ km/s (Lee 2003). If we consider the reference craton to be highly depleted ($Mg\#=93$) and the SRM to be of a primitive composition ($Mg\#=88$), this contributes less than 2% of the velocity variation (if compared to a reference velocity of 4.5 km/s). A similar analysis by Jackson and Faul (2005) in Australia, concluded that the $\sim 10\%$ observed velocity variation between the craton and the southeast was not due to differences in Mg# alone. Assuming the full range in composition, the correction for composition attributes about 10-15% of our calculated temperature anomalies. Therefore, even though it is likely that there is compositional variation across the Great Plains, this effect is small even when the maximal Mg# variation is assumed, which is not true over most of the region and so we will not consider it further.

5.4 Elevation

The westernmost Great Plains is the eastern flank of the Rocky Mountains and rises to high elevations over a broad region; this elevation profile is in contrast to the flanks of other mountain piedmont slopes where there is high relief as shown in Figure 1.2 (Eaton 2008). The reason for this anomalous elevation has driven debate with analyses of sources of uplift ranging from climate and erosional feedback, tectonics, crustal structure, mantle structure and mantle flow. Previous studies usually focus on part of the SRM and the western US, but here we consider the elevation east of the RMF. Contributions to topography can be divided into an isostatic component, i.e. the buoyancy of the crust floating atop lithospheric mantle (LM), and an asthenospheric component.
While the western High Plains has the elevation profile more similar to a mid-ocean ridge than an orogen, it is objectively different as at the RMF no new lithosphere is being generated (likely this is occurring southward in the RGR). Here, a source at depth is vertically forcing the surface from below causing broad patterns of uplift, and is refined on a smaller scale due to variations in crustal thickness. Our study provides a unique treatment of this deep source, considering thermally expanded asthenosphere to influence the surface elevation. The analysis presented here is different than “dynamic topography” calculations where topography is due to vertical stress produced by instantaneous mantle flow from a much deeper mantle source (Forte et al. 2010; Moucha et al. 2008; Becker et al. 2013). Here we consider that the temperature contrasts between the craton and active west are stable and long-lived, and do not require the dynamics of mantle flow.

5.4.1 Asthenospheric contribution to elevation

We model the topography of the Great Plains by first calculating the sublithospheric contribution to surface elevation constrained from the seismic tomography and lithospheric depth estimates of this study. The hypothesis that structure under the lithosphere may be influencing the elevation in this region is based on two observations: (1) the topography of the LAB has a high correlation with the surface elevation, and (2) the large velocity (temperature) contrast between the east and west.

Temperatures inferred from our seismic shear wave velocity model suggest warm mantle under the SRM and western High Plains. At constant pressure (i.e. constant depth) warmer material is more thermally expanded. We calculate the one-dimensional
expansion in each vertical column, with differential expansion of each layer segment given by

\[ \Delta L = \alpha \Delta T L \]

Where \( \alpha \) is the 1D thermal expansion coefficient of forsterite, \( \Delta T \) is the temperature differential and \( L \) is the total length of the segment, here 5 km. The total variation in topography is the summation of the differential expansions between the depth of the lithosphere averaged over a subsection of the craton (205 km depth) and the depth of the lithosphere at any other location. We assume that the uplift caused by the differential expansion is fully coupled to the overlying lithosphere and surface, and thus the amount of variation in the calculated expansion will be observed at the surface.

In Figure 5.3 we plot the calculated topography and the residual topography (calculated topography minus observed topography smoothed with a Gaussian filter with a standard deviation of 200 km). The Colorado Rockies and RGR have the largest contribution to topography due to a combination of warm temperatures and shallow asthenosphere. The Wyoming craton and the eastern halves of the Yavapai and Mazatzal have only a moderate contribution from the asthenosphere; this is due to a combination of colder temperatures and only slightly thickened lithosphere relative to the west. The Superior craton has almost no contribution (and even a slightly negative value), implying this cold region lacks thermal expansion and has a deep lithosphere. The Gulf coast elevation is overestimated by up to ~1km. The thin lithosphere and velocities in this region lead to misfits in our thermal modeling, as expected. The peak elevations under the SRM are underestimated by between 1 and 1.5 km, however this thermal expansion of the asthenosphere still accounts for about ~50-67% of the topography here. Even
though the majority of the topography can be attributed to asthenospheric expansion, and
the long wavelength features ($\lambda > 250$ km) are recovered, we can refine our study by
exploring the mechanisms contributing to the residual elevation.

Fig. 5.3 (a) Calculated topography from thermally induced expansion in the
asthenosphere. (b) The residual topography (calculated – smoothed elevation).

5.4.2 Crustal refinements to elevation

While we have concluded that the majority of elevation and long-wavelength
behavior is likely to be due to thermally expanded asthenosphere, in this section we
assess the possible contributions from a crustal source. We first consider the effects of
warm crust and then consider the effect of a crustal root under the SRM.

We perform a calculation of thermal expansion in the crust by constraining
temperature anomalies at each depth using geotherms calculated for different surface HF
regimes (Chapman 1986). We define the craton (-94 W) temperature using the 40 mW/m² geotherm (Chapman 1986), the temperature of the SRM at (-106 W) as the 100 mW/m² geotherm, and the temperature of the eastern edge of the SRM (-104 W) as the 80 mW/m² geotherm (Chapman 1986). We estimate the temperature at any other longitude point by a piecewise linear interpolation between the three geotherms. For each point we calculate differential expansion in the same way as we did for the asthenosphere, but this time integrate the crust from 45 km depth to the surface. We find that a warm crust could contribute as much as ~0.63 km of elevation under the SRM.

The isostatic mechanisms under the SRM and western High Plains are complicated, but where there is a thickened crustal root there will be some positive contribution to elevation. The Free Air anomaly is shown in Figure 5.4c (Gravity Database of the US: http://www.research.utep.edu) and the large short-wavelength anomaly in the mountains may be due to edge effects directly adjacent to a balanced region. In an isostatic model, the elevation above sea level due to a crustal root is dependent on the thickness of the crustal root and on the densities of the crust and LM: here we assume the density of the LM is 3250 kg/m³ (Lachenbruch and Morgan 1990), an average crustal density of 2850 kg/m³, and a root consisting of mafic lower crustal material of density 3000 kg/m³. We use crustal thicknesses determined from smoothed PdS RFs (Figure 5.4b) and define the crustal root as any crustal thickness in excess of the reference crustal thickness, 45 km. The crustal root in the Rocky Mountains is thickest north of 38° latitude and is much thinner in the south. This variation in crustal structure is due to a variety of factors including initial variations inherited from
continental assembly, thickening during Rocky Mountain formation, and rifting by the RGR.

Here the thickened crustal root beneath the northern SRM contributes elevation up to 0.5 km greater than the elevation calculated from thermal expansion alone (Figure 5.4 d-f solid blue line). In the south, the absence of a root, or slightly negative root means that the inclusion of crustal isostasy produces smaller estimates of elevation than thermal expansion only. While somewhat qualitative due to assumed density structure, the purpose of this discussion is to indicate that: (1) there is some topographic contribution from a thickened root in the north, and (2) our results corroborate previous suggestions that isostasy alone does not fully explain the observed elevation (Sheehan et. al. 1995).
Fig. 5.4 Smoothed elevation map contoured every 0.25 km (a). Smoothed crustal root map contoured every 1 km (b). The free air anomaly (c). The latitude profiles with the smoothed elevation (dashed black line), topography calculated from asthenospheric thermal expansion only (solid red line), topography from asthenospheric and crustal thermal expansion (solid green line), and topography from the sum of crustal and asthenospheric thermal expansion and crustal isostasy (solid blue line) (d-h). The
numbers in the bottom left hand corner of each profile indicate the RMS error between the estimated curves and the observed elevation (from left to right: asthenospheric expansion only, asthenospheric and crustal expansion, and sum of thermal expansion and crustal isostasy).

5.5 Comparison of results

The results shown in Figure 5.4 d-h reproduce the smoothed elevation within reason. The asthenospheric expansion tends to slightly underestimate the elevation everywhere (red solid line) and the asthenospheric and crustal thermal expansion tends to underestimate elevation just under the SRM and slightly overestimate elevation east of the SRM (green solid line). The curves that includes the crustal isostasy (solid blue line) are good estimates for 37 N and 38 N, but contributes large misfits east of the SRM. For each profile, adding the first crustal refinement (crustal expansion) to the original asthenospheric thermal expansion calculation decreases the RMS misfit between observed and predicted elevation. Adding the second refinement of crustal isostasy does not always decrease the overall misfit.

These results suggest that a large contribution to the topography of the High Plains could be attributed to thermal expansion and a relatively gently deepening asthenosphere, i.e., thickening lithosphere, over a long distance eastward of the RMF. However, sometimes the thermal model only does not achieve the highest elevations under the SRM and the inclusion of crustal isostasy improves our estimate. The presence of some large residuals east of the RMF suggests that there may exist a complicated density or Moho structure that our simple isostatic model does not take into account.
CHAPTER 6
DISCUSSION AND CONCLUSIONS

This study has provided a 3D shear velocity model of the Great Plains of the central US. We image the fast Superior cratonic keel to depths of \(~250\) km, the slow velocity anomalies associated with the SRM and RGR, and possible asthenospheric flow along Proterozoic suture zones. Our results show that the velocity structure of the Wyoming craton is distinctly different from the Superior craton, suggesting lithospheric thinning.

An important result derived from the dataset is a characterization of seismic LAB: the LAB is thinnest under the SRM and deepens east and northeastward reaching a maximum thickness under the Superior craton. The thinned lithosphere that follows the arc of the Gulf Coast is a prominent feature that does not follow the overall pattern of lithospheric thickening outward and away from the RMF. The correlation between lithospheric thickness and surface topography is clear, which suggests that the depth to which the asthenosphere rises may have control on the surface geology.

We calculate the topography of the study region as a combination of expansion of the asthenosphere from thermal anomalies in the asthenosphere and consider the secondary effects of expansion in the crust and crustal isostasy. The long wavelength behavior of the elevation is accounted for by the thermal modeling of the asthenosphere as the broad region of high topography adjacent to the SRM is due to the very large contrast in both temperature and depth to which the asthenosphere rises between the stable craton and the actively deforming west. Our estimates are always improved when
including a crustal thermal expansion term, but only the estimates in the northern
latitudes under the SRM are improved by including crustal isostasy. Our approach has led
to a quantification of the elevation in the Great Plains by considering a physical
mechanism contributing to uplift that exhibits first-order control on topography, thermal
expansion of the asthenosphere, but so far has been absent from investigations of
elevation.


doi: 10.1016/j.epsl.2010.03.017

doi:10.1029/2003JB002743


Gravity Database of the US. N.p.: n.p., Web.


Thurner et. al. (in prep)


