Physical State and Structure of the Crust Beneath the Western-Central United States From Multiobservable Probabilistic Inversion

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Abstract  The Southern Rocky Mountains (SRM) have the highest regional elevations, but a comparable crustal thickness to that of the Great Plains. Isolating the contributions of the crust and upper mantle in supporting the topography of the SRM has been a topic of considerable controversy for the past two decades, leading to contrasting hypotheses. In order to constrain the relative contributions from the crust and mantle, reliable models of the physical state of the crust and upper mantle are necessary. Here we employ a recently developed multiobservable Bayesian inversion approach to study the present-day physical state (i.e., $V_p$, $V_s$, density, $V_p/V_s$, crustal thickness, and temperature) of the crust in a region encompassing the western-central United States. We image highly heterogeneous temperature and compositional structures, which combine in a nonintuitive way with total crustal thickness to create the observed topography and control other geophysical signatures. In the SRM, topography is supported by a combination of crustal and mantle contributions. South of 38°N, mantle sources (both static and dynamic) constitute the main support to the high elevations, whereas to the north of this latitude, crustal structure provides the main contribution. In the Colorado Plateau, Arizona transition zone, and Basin and Range provinces, most of the topography seems to be supported by mantle sources (within the lithospheric mantle and/or dynamic). At middle to lower crustal depths beneath the SRM, we image high temperatures, high $V_p/V_s$ values, anomalously low crustal velocities, and low densities. These observations correlate with high electrical conductivities and the location of recent volcanism, suggesting the occurrence of melts/liquids. We do not find low $V_p/V_s$ values at middle/low crustal depths beneath the SRM that may suggest felsic, weak lithologies.

1. Introduction

Despite being far away from any plate boundary, the western-central United States has experienced a series of complex tectonic events (e.g., extension, shortening, uplift, and volcanism) since the late Mesozoic (cf. Karlstrom et al., 2012). The causes of such a tectonism, especially in the Southern Rocky Mountains (SRM), have been the focus of numerous studies (e.g., Bailey et al., 2012; Drenth et al., 2012; Hansen et al., 2013; Karlstrom et al., 2012; Lerner-lam et al., 1998; Li et al., 2002; MacCarthy et al., 2014; Prodehl & Lipman, 1989; Sheehan et al., 1995). While it has been shown that there is no considerable crustal thickness difference between the SRM and the western Great Plains (GP; e.g., Gilbert, 2012; Gilbert & Sheehan, 2004; Prodehl & Lipman, 1989; Sheehan et al., 1995; Shen et al., 2013), the SRM has a significantly higher topography. Also, free-air anomalies over the SRM suggest that they are in near isostatic equilibrium state (e.g., Coblenz et al., 2011). Different models/hypotheses have been put forward in the past few decades to explain these observations, as well as the relative contributions from the crust and upper mantle, but some of their main conclusions remain contradictory (e.g., Bailey et al., 2012; Bird, 1984, 1988; Coblenz et al., 2011; Hansen et al., 2013; Gilbert & Sheehan, 2004; Jones et al., 2015; Lee & Grand, 1996; Levandowski et al., 2014; Li et al., 2002; Lowry & Pérez-Gussinyé, 2011; Moschetti et al., 2010; Sheehan et al., 1995).

Some studies ascribe the high topography of the SRM to variations in crustal properties. For example, crustal thickening resulting from the horizontal subduction of the Farallon plate beneath North America has been suggested by Bird (1988) as the main cause for this anomalous topography. Sheehan et al. (1995) noted that a pure Airy crustal root hypothesis cannot explain the higher topography of the SRM relative to the GP. Their
preferred model required half of the compensation in the crust and half in the mantle. Li et al. (2002) studied the crustal and upper mantle structures beneath the Colorado Rockies and found greater values for \( \frac{\partial \rho}{\partial V_s} \) in the crust. They suggested that the higher topography of this region is largely compensated by crustal density variations and crustal thickness. Bailey et al. (2012) modelled the subsurface density structure using seismic velocities estimated by joint inversion of receiver functions and dispersion curves and concluded that lateral density variations below \( \sim 50 \) km are unimportant. In a recent study, Levandowski et al. (2014) converted the shear-wave velocity model of Shen et al. (2013) into densities for the western United States. These authors separated the roles of the crust and upper mantle in the uplift of the Cordillera. According to their results, the crustal structure plays a more important role in supporting the elevations of the GP and the SRM. More recently, hydration of the lower crust has been suggested by Jones et al. (2015) as the main cause for the isostatic uplift of the Colorado Rockies.

Other studies have suggested that mantle buoyancy has significantly contributed to the surface elevation of the SRM (Boyd & Sheehan, 2013; Lee & Grand, 1996). The lack of correlation between the topography and crustal thickness (Gilbert & Sheehan, 2004; Hansen et al., 2013; Sheehan et al., 1995), together with the lack of thickened crust beneath the SRM relative to the adjacent GP, have been also interpreted as evidence for a substantial mantle support of the high topography. Coblenz et al. (2011) evaluated the geoid to topography ratios for the western United States to constrain compensation mechanisms. Their findings pointed to shallow compensation depths (<100 km) for the SRM and a more prominent role for subcrustal processes in supporting the high elevations.

It is clear therefore that the compensation mechanisms responsible for supporting the high topography of the SRM, and more generally of the western-central United States, remain contentious. Most of the controversy...
seems to be rooted in the use of different and/or incompatible crustal models (e.g., crustal thickness, density, \(V_p/V_s\), ...) derived from different data sets and methods. However, any valid crustal/mantle model should be compatible not with just one, but with all relevant data sets.

Here we adhere to this idea and characterize the 3-D thermal, physical and compositional crustal structure of the western and central United States through a simultaneous and internally consistent inversion of complementary geophysical data sets. For this purpose, we apply a recently developed multiobservable probabilistic inversion methodology (e.g., Afonso, Fullea, Griffin, et al., 2013; Afonso, Fullea, Yang, et al., 2013; Afonso, Rawlinson, et al., 2016; Qashqai et al., 2016) to a region encompassing the Colorado Plateau (CP), some parts of the Basin and Range (BAR), Arizona transition zone (ATZ), SRM, GP, Rio Grande Rift (RGR), and the Central Rocky Mountains (CRM; Figure 1). Data used in this study include Rayleigh wave dispersion curves, radial receiver functions (RFs), geoid anomalies, absolute elevation and surface heat flow.

The main aim of this work is to provide further constraints on the physical state and structure of the crust beneath western-central United States and to use this information to elucidate evolutionary models and test hypotheses regarding the present-day topography of the region. Although the focus is on the SRM and the GP, the crustal structure across the CP and surroundings, where some inconsistencies exist concerning their crustal structure (Bailey et al., 2012; Bashir et al., 2011; Frassetto et al., 2006; Gilbert & Sheehan, 2004; Gilbert, 2012; Hansen et al., 2015; Karlstrom et al., 2012; Liu et al., 2011; Lowry & Pérez-Gussinyé, 2011; Parsons et al., 1996; Sheehan et al., 1995; Shen et al., 2013; Zandt et al., 1995), is also addressed. We take advantage of complimentary data sets to provide robust models of the temperature, density, and seismic structure of the crust at a nominal resolution of \(\lessapprox 1 \times 1\) degree. Since our method relies on a probabilistic approach (section 3.4), the solution to the inverse problem is a multidimensional probability density function (PDF) rather than a single best model. This posterior PDF can then be interrogated to obtain relevant information on the average structure of the crust and associated uncertainties.

2. Tectonic Setting

Figure 1 shows the topography of the study region including physiographic provinces (Figure 1a) and major geological features (Figure 1b). After the removal foundering of the Farallon slab, widespread and voluminous magmatism occurred over a large region of the western United States during the middle Tertiary (Baldridge, 2004; Humphreys, 1995). During the late Cenozoic the Rio Grande Rift was created, separating the CP and the GP. However, the CP remained relatively undeformed during this time (Baldridge, 2004; Thompson et al., 2017). In the southern and western margins of the CP, the lithosphere was stretched and thinned, creating a new physiographic province with narrow mountain ranges separated by basins, referred to as the Basin and Range (Baldridge, 2004).

Archean and Proterozoic rocks underlie most of the west and southwest of the United States. They are now exposed at or near the surface in the cores of mountains in the west and southwest United States (Baldridge, 2004). Within the study region (Figure 1), Proterozoic rocks are mostly exposed along the SRM, including the Colorado Mineral Belt (CMB; Figure 1b). The SRM is a north-northwest trending high-elevation region that occupies the eastern part of the Cordillera orogen, extending from New Mexico through Colorado and into the Southern Wyoming (i.e., Central Rocky Mountains; Figure 1). Note that a major Precambrian structure known as the Cheyenne Belt (CB; Figure 1b) separates a Proterozoic-dominant basement in the south from a dominantly Archean basement in the north (Decker et al., 1988; Prodehl & Lipman, 1989). The CMB is characterized by Precambrian shear zones (Mccoy, Karlstrom, et al., 2013), Cenozoic magmatism (MacCarthy et al., 2014) and numerous igneous intrusions (Chapin, 2012). It is a \(\sim 500\)-km-long, 25- to 50-km-wide northeast-striking shear zone system, containing segments of Paleocene to Oligocene intrusions and related mineral deposits emplaced during and after Laramide orogeny (Chapin, 2012; Mccoy, Karlstrom, et al., 2013). The CMB is also located in a region of pronounced low Bouguer anomalies (Figure S1 in the supporting information) as well as low shear wave velocities, especially near the San Juan Mountains (SJ), the Sawatch Range (SR) and the Front Range (FR) (Gilbert, 2012; Levandowski et al., 2014; Mccoy, Karlstrom, et al., 2013; Shen et al., 2013). The general northeast trend of this mineral belt in the SRM is approximately paralleled to the Jemez Lineament in the south. The latter includes a series of middle Miocene to Quaternary volcanic fields, which intersects the RGR system and extends to the GP (Figure 1b).

The general features of the SRM are in sharp contrast with those of the adjacent CP and GP provinces. The CP is characterized by \(\sim 2\) km of Cenozoic uplift (e.g., CC: the Circle Cliffs, MM: the Miners Mountain, KU: the Kaibab
Uplift, MU: the Monument Uplift, SRS: the San Rafael Swell, and UU: the Uncompahgre Uplift in Figure 1b), significant magmatism around its western and southern boundaries (cf. Crow et al., 2011), a series of faults zones (e.g., Holbrook Lineament—see Figure 1b) and sharp mantle velocity and lithospheric thermal gradients near its boundaries (e.g., Afonso, Rawlinson, et al., 2016; Levander et al., 2011). Although dynamic topography is commonly proposed to be the main cause for the uplift of the CP and surroundings (e.g., Bird, 1984; Liu & Gurnis, 2010; van Wijk et al., 2010), a number of recent studies based on geophysical data have argued that the current surface topography of the Colorado Plateau can be explained by the internal buoyancy of the lithosphere, with dynamic sublithospheric processes playing a secondary, localized role (e.g., Afonso, Rawlinson, et al., 2016; Coblenz et al., 2011; Hansen et al., 2013; Levandowski et al., 2014).

3. Data and Methodology
3.1. Receiver Functions and Processing
Three-component broadband teleseismic data, recorded by 245 USArray TA stations (Figure 1a), were used to compute radial receiver functions. For each station, earthquakes with magnitudes >5.5 and epicentral distances between 30° and 90° were selected. A time window of 60 s before and 240 s after direct P wave arrival time was used to truncate each selected seismogram. The AK135 velocity model (Kennett et al., 1995) was used to compute the slowness of these seismograms.

All seismograms were resampled to 10 Hz. To calculate P-to-S receiver functions for each seismic station, the horizontal components of the three-components seismogram were rotated from ZNE to the ZRT coordinate system using the computed back azimuth. To isolate the effect of local structure from the source time function, the vertical component was deconvolved from the horizontal components by using the time-domain iterative deconvolution approach of Ligorria and Ammon (1999). To remove higher frequencies (e.g., >0.5 Hz), the estimated receiver functions were multiplied by a Gaussian function of the following form in the frequency domain (\(\omega\)):

\[
G(\omega) = \exp \left( -\frac{\omega^2}{4a^2} \right)
\]

where \(a\) is a width factor controlling the cutoff frequency. In this paper, we used \(a = 1.0\) s which corresponds to a cutoff frequency of \(-0.5\) Hz. Since it is expected that some details of the internal crustal structure will be obscured with such a low corner frequency, we tested higher values of \(a\) (2.5 – 5 s). However, this required additional model parameters (i.e., more crustal layers) and computational time. We found that the choice of \(a = 1.0\) and four crustal layers offered a good balance between parsimony of model specification, computational time, details of crustal structure and fitting statistics (see, e.g., Tkalčič et al., 2006; Tkalčič et al., 2011).

Radial receiver functions from different events were stacked to increase the level of signal-to-noise ratio at each station. It should be noted that in the presence of dipping layers, arrival times, and amplitudes of the P-to-S converted phases and reverberations associated with dipping layers vary significantly with back azimuths and ray parameters (e.g., Cassidy, 1992). Therefore, two constraints were considered in the stacking procedure: (i) similar to Bodin et al. (2013), the range of ray parameters was narrowed (0.058 – 0.072 s/km) and (ii) only coherent receiver functions were chosen by using a cross-correlation matrix approach described in Tkalčič et al. (2011). This approach thus removes any noisy/erroneous receiver functions and also suppresses the effect of azimuthal variations. The mean (stack) and square root of the variance of all radial RFs were taken as the input data and its associated uncertainty, respectively.

3.2. Rayleigh Wave Dispersion Curves
Dispersion curves and their associated uncertainties at periods of 8–80 s were extracted from the phase velocity maps of Shen et al. (2013) at all locations where seismic stations were present. These maps were constructed from cross-correlations of ambient noise between USArray TA stations using an Eikonal tomography method (e.g., Lin et al., 2009) for periods 8–40 s. Rayleigh wave phase velocity maps at 25 to 80 s were generated from earthquake surface waves using a Helmholtz tomography method (Lin & Ritzwoller, 2011). More details about how these phase velocity maps are constructed are described in Shen et al. (2013). For periods between 80 and 150 s, we use the data set of Afonso, Rawlinson, et al. (2016), who inverted surface wave data from around 450 teleseismic events with magnitudes >5.5 using two-plane-wave tomography (Yang & Forsyth, 2006). In this method, the wavefield of each incoming teleseismic event to our study region is modelled using the interference of two plane waves. The sensitivities of each plane wavefield to velocity heterogeneities are described by 2-D finite frequency sensitivity kernels calculated following Ying et al. (2004).
For the tomography, phase velocities and two-plane wavefields are jointly inverted using phase and amplitude data of teleseismic events recorded at each station. Uncertainties of phase velocities are evaluated from the posterior model covariance matrix during the inversion. More details about the procedures of constructing phase velocity maps using two-plane-wave tomography are discussed in Yang et al. (2008) and Afonso, Rawlinson, et al. (2016).

3.3. Geoid, Surface Heat Flow, and Elevation

Geoid height was taken from the global Earth model EGM2008, which contains spherical harmonic coefficients up to degree 2190 and order 2159 (Pavlis et al., 2012). The total geoid was filtered to remove the long wavelengths associated with anomaly mass sources deeper than ~400 km (e.g., degrees 2–9 removed). Surface heat flow values were obtained by interpolating existing data points in the geothermal map of North America (http://www.smu.edu/dedman/academics/programs/geothermalab). Some values >200 and <25 mW/m² were treated as outliers and removed, as they are not considered representative of the deeper thermal structure (mostly due to local shallow conditions, e.g., underground water flow). ETOPO2v2 Global Database (http://www.ngdc.noaa.gov/mgg/fliers/06mgg01.html) was used to extract elevation data (see below).

Input values of geoid, surface heat flow, and elevation as well as their associated uncertainties were obtained as follows. The study area was subdivided into 140 columns or cells (14 by 10), each $1^\circ$ by $1^\circ$ (Figure 1a). Within each cell, all available values in the respective data sets were used to compute a mean value and associated 1σ standard deviation. These were the values used in the joint inversion. We note that the obtained 1σ standard deviations are not true observational uncertainties, but rather a measure of the natural variability of the fields within each cell’s surface (observational uncertainties are much smaller and thus ignored here). A similar procedure was used for surface heat flow where enough data points were available. In regions where the original coverage was not dense enough, we assigned a minimum uncertainty of 15%.

When there is more than one seismic station in a cell (Figure 1a), the mean values of the nonseismic data (geoid, elevation, and heat flow) and their related 1σ uncertainties in that cell were used as input values. By doing this, we force the crustal models associated with all stations in each cell to be compatible with the average nonseismic data characterizing that cell.

3.4. Inversion Approach

In this study, we employ a recently developed multiobservable probabilistic inversion approach (Afonso, Fullea, Griffin, et al., 2013; Afonso, Fullea, Yang, et al., 2013; Guo, Afonso, et al., 2016; Jones et al., 2017; Afonso, Rawlinson, et al., 2016; Qashqai et al., 2016; Shan et al., 2014) based on a Bayesian inference framework. In this framework, prior information on both data and model parameters is encoded in a joint probability density function (PDF) denoted by $\rho(d,m)$. Similarly, prior knowledge about the physical theory relating the unknown model to data is given by a joint PDF represented by $\Theta(d,m)$. The general solution to the inverse problem is then obtained by conjunction of these states of information and given by a joint PDF, known as the posterior PDF (Tarantola, 2005). In the case of linear data and parameter spaces, the posterior PDF can be written as

$$\sigma(d,m) \propto \rho(d,m) \Theta(d,m)$$

Under most practical circumstances, this joint PDF can be integrated over the data space to obtain the marginal PDF in the parameter space as (cf. Afonso, Fullea, Griffin, et al., 2013; Tarantola, 2005)

$$\sigma(m) \propto \rho(m)L(m)$$

where $\rho(m)$ is the prior distribution over $m$ describing all that is known about the model parameters independently of the data $d$ used in the inversion; $L(m)$ is the likelihood function, which expresses how probable are the observed data given a specific combination of model parameters (i.e., is a measure of how well a specific model $m$ can explain $d$).

In the typical case of independent and identically (normally) distributed observations, we can adopt the following form for the likelihood function (e.g., Hauser et al., 2011; Pasyanos et al., 2006; Tarantola, 2005):

$$L(m) = \mathcal{N}(\mathbf{d} | \mathbf{m}, \mathbf{C})$$
Figure 2. Discretization scales used to model the 1-D Earth structure beneath each station (Triangles in Figure 1a), modified from Afonso, Fullea, Yang, et al. (2013); Shan et al. (2014). The finest mesh with vertical grid step of 2 km (red dots) is used to solve the forward problems associated with elevation, geoid, and surface heat flow data (see text). The second mesh (intermediate discretization) is constructed to use in solving the Gibbs free energy minimization problem (blue squares). Here 10 and 5 nodes are used in the lithospheric (Compositional layer 1) and the sublithospheric mantle (Compositional layer 2), respectively.

\[ L(m) \propto \exp \left[ -\frac{1}{2} \sum_j \Phi(m)_j \right] \]  

where

\[ \Phi(m)_j = w_j \sum_i \left[ \frac{d_{i}^{obs} - g_i(m)}{\sigma_i} \right]^2 \]  

In equation (5), \( \Phi(m)_j \) is the so-called sum-of-squares function (also known as misfit function) associated with each data, \( j \) and \( i \) refer to data type (i.e., surface waves, geoid, elevation, and SHF) and data points, respectively, \( \sigma_i \) is the uncertainty associated with each data types at point \( i \), \( w_j \) are arbitrary weights assigned to each data type and \( g_i(m) \) denotes synthetic data computed using model \( m \). The weights control the influence of each data set to the total misfit function (sum of \( \Phi(m)_j \)) and thus to the final solution. For example, up-weighting the importance of RFs would result in crustal models with more vertical structure than those obtained by up-weighting, for example, surface waves. As in our previous study (Qashqai et al., 2016), our main interest is to obtain models that jointly fit the seismic data as well as possible while being simultaneously consistent with, but not strictly controlled by, noneismic data. We therefore utilize the same scaling (weighting) factors as in Qashqai et al. (2016; see their Appendix A). Noneismic data sets are “naturally” weighted only by their respective uncertainties.
In general, when the model space is multidimensional and there is no simple form for the posterior PDF, Markov Chain Monte Carlo (MCMC) sampling methods can be employed to obtain estimates of \(\sigma(m)\). We sample the posterior PDF with the Delay Rejection Adaptive Metropolis (DRAM) algorithm of Haario et al. (2006).

In the DR algorithm (Mira, 2001), upon the rejection of a model by the MH rule, a new sample is drawn from a different proposal distribution (typically, a scaled version of the original proposal distribution), which may or may not depend on the rejected value. If the new model is rejected again, one can consider a third attempt and so on. Therefore, the algorithm improves the sampling by reducing the number of rejected samples.

In the AM algorithm (Haario et al., 2001), the basic idea is to update/adapt the proposal distribution (needed for sampling the model parameters) based on all previously accepted models. Adaptation approximates the proposal distribution to the posterior distribution, hence avoiding sampling regions of low probability. The optimum number of samples, needed for the adaptation, as well as the number of samples between adaptations (adaptation intervals), strongly depends on the actual problem (Guo, Chen, et al., 2016; Haario et al., 2006). Numerous tests and previous studies (Afonso, Rawlinson, et al., 2016; Guo, Chen, et al., 2016; Qashqai et al., 2016; Shan et al., 2014) have indicated that five adaptations are sufficient to estimate the posterior PDF in the current problem. We use 150,000 simulations in the initial nonadaptation stage and the proposal distribution of the model is adapted five times, every 30,000 samples (total number of simulations per station: 300,000).

### 3.5. Model Parameterization and Priors

Since the parameterization of the crust and the upper mantle have been fully described elsewhere (e.g., Afonso, Fullea, Yang, et al., 2013; Afonso, Rawlinson, et al., 2016; Qashqai et al., 2016; Shan et al., 2014), here we only summarize the most relevant aspects for this study. Our 3-D domain is made up of a collection of adjacent and nonoverlapping 1-D columns coincident with the location of the seismic stations. Each 1-D column is defined by four crustal layers (a sedimentary layer and three crystalline layers, Figure 2) over two compositional upper mantle layers (lithosphere and sublithospheric mantle). As mentioned in section 3.1, the choice of four crustal layers is a compromise choice given the filtering of RFs, parsimony of the model and fitting statistics. As the focus of this paper is crustal structure, a smooth structure is assumed in the upper mantle (i.e., no seismic discontinuities).

Similar to our previous works (e.g., Afonso, Rawlinson, et al., 2016; Qashqai et al., 2016; Shan et al., 2014), each crustal layer is described by six parameters including the coefficient of thermal expansion (\(\alpha\)), compressibility (\(\beta\)), thermal conductivity (\(k\)), bulk density at surface P-T conditions \(\rho_0(P_0, T_0)\), thickness variation (\(\Delta h\)), and \(V_p/V_s\). In addition to these layer-specific parameters, the average crustal radiogenic heat production (RHP) is also considered an unknown and inverted for. Due to the fact that predicted data are dominantly affected by changes in \(\rho_0, \Delta h, V_p/V_s,\) and RHP (cf. Afonso, Rawlinson, et al., 2016), they are treated as unknowns in the inversion while the first three (\(\alpha, \beta, k\)) are assumed constant (Table 1). This reduces the total number of crustal unknowns to 13 per column.

The discretization of the upper mantle is similar to that used in Afonso, Fullea, Yang, et al. (2013); Shan et al. (2014); Guo, Chen, et al. (2016); and Afonso, Rawlinson, et al. (2016). The entire 1-D model (crust + upper mantle) is discretized with a finite-difference mesh with a resolution of 2 km (red dots in Figure 2), to solve the forward problems associated with geoid height, elevation and the computation of lithospheric geotherms (see next section). In addition, 15 “thermodynamic” nodes are used in the upper mantle (10 within the lithosphere and five in the sublithospheric mantle; blue diamonds in Figure 2) for solving the Gibbs free energy minimization problem which informs all the forward problems (see below).

Given the strong and direct relationship between rock strength and temperature (cf. Afonso, Moorkamp, et al., 2016), the bottom of the thermal lithosphere (hereafter referred to as the LAB) is defined as the 1250 °C isotherm \((T_{1250})\), which is consistent with results from numerical modeling of upper mantle convection with realistic viscosities (Afonso et al., 2008; Afonso & Zlotnik, 2011; Ballmer et al., 2011; Gerya, 2010; van Wijk et al.,

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### Table 1

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Values</th>
</tr>
</thead>
<tbody>
<tr>
<td>CTE in the first crustal layer (°C⁻¹)</td>
<td>2.7 × 10⁻⁵</td>
</tr>
<tr>
<td>CTE in the second crustal layer (°C⁻¹)</td>
<td>2.7 × 10⁻⁵</td>
</tr>
<tr>
<td>CTE in the third crustal layer (°C⁻¹)</td>
<td>2.6 × 10⁻⁵</td>
</tr>
<tr>
<td>CTE in the fourth crustal layer (°C⁻¹)</td>
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<tr>
<td>(k) in the second crustal layer (W m⁻¹ °C⁻¹)</td>
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<td>(k) in the fourth crustal layer (W m⁻¹ °C⁻¹)</td>
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Table 2
The List of All Unknown Parameters and Priors

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<th>Maximum value</th>
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<td>from Afonso, Rawlinson, et al. (2016)</td>
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<td>5.0</td>
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<td>$\text{FeO}$ in the lithosphere (wt%)</td>
<td>6.0</td>
<td>9.2</td>
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<tr>
<td>$\text{MgO}$ in the lithosphere (wt%)</td>
<td>34</td>
<td>55</td>
</tr>
<tr>
<td>$\text{CaO}$ in the lithosphere (wt%)</td>
<td>0.1</td>
<td>5.5</td>
</tr>
<tr>
<td>$\text{Al}_2\text{O}_3$ in the sublithosphere (wt%)</td>
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<td>5.0</td>
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<tr>
<td>$\text{FeO}$ in the sublithosphere (wt%)</td>
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<td>9.2</td>
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<tr>
<td>$\text{MgO}$ in the sublithosphere (wt%)</td>
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<td>55</td>
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<tr>
<td>$\text{CaO}$ in the sublithosphere (wt%)</td>
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<tr>
<td>STP density (fourth crustal layer) (kg/m$^3$)</td>
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<td>3150</td>
</tr>
<tr>
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<td>2.2</td>
</tr>
<tr>
<td>$V_p/V_s$ (second crustal layer)</td>
<td>1.6</td>
<td>1.9</td>
</tr>
<tr>
<td>$V_p/V_s$ (third crustal layer)</td>
<td>1.6</td>
<td>1.9</td>
</tr>
<tr>
<td>$V_p/V_s$ (fourth crustal layer)</td>
<td>1.6</td>
<td>1.9</td>
</tr>
<tr>
<td>$\Delta h$ (first crustal layer) (km)</td>
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<td>$\Delta h$ (second crustal layer) (km)</td>
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<td>$\Delta h$ (fourth crustal layer) (km)</td>
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<td>7.0</td>
</tr>
<tr>
<td>RHP (μWm$^{-3}$)</td>
<td>0.4</td>
<td>1.8</td>
</tr>
</tbody>
</table>

2010; Zlotnik et al., 2008). The sublithospheric temperature structure is entirely controlled by three unknown parameters (or temperatures) at different node-depths (Figure 2 - $T_{\text{buffer}}$, $T_{\text{int}}$, and $T_{\text{bottom}}$). Once these temperatures are randomly sampled by the MCMC sampler (see their ranges in Table 2), the temperature profile for the sublithospheric mantle is constructed by linearly interpolation between $T_{1250}$, $T_{\text{buffer}}$, $T_{\text{int}}$, and $T_{\text{bottom}}$ (e.g., see Afonso, Rawlinson, et al., 2016; Shan et al., 2014). Since the focus of the present work is on the crust, we use a prior for LAB depth based on the results in (Afonso, Rawlinson, et al., 2016), who focused on the mantle thermochemical structure in the same region using a full 3-D multiobservable thermochemical tomography approach.

Prior distributions used for crustal and upper mantle parameters are given in Table 2. Priors for the rest of parameters are compatible with laboratory measurements of physical properties of crustal rocks (e.g., Brown & Rushmer, 2006; Jaupart & Mareschal, 2011).

3.6. Forward Problem

Upon proposing a model in each simulation of the MCMC algorithm, the forward problem can be summarized as follows: (1) compute the lithospheric geotherm by solving the steady-state heat transfer equation subject to Dirichlet boundary conditions at the surface of the Earth (10 °C) and at the base of the lithosphere ($T_{1250}$); (2) obtain all physical properties of interest (e.g., $V_s$, $V_p$, and density) for the crust and mantle (see below); (3) solve the isostatic balance equations; (4) compute the geoid height; (5) calculate radial receiver functions; and (6) compute fundamental-mode Rayleigh wave dispersion curves considering anelasticity effects.

In the crust, the thicknesses of crustal layers are obtained by randomly perturbing an initial crustal model. Here CRUST1.0 (Laske et al., 2013) is used as the initial model, and the minimum and maximum bounds for perturbations ($\Delta h$) are given in Table 2. The P-T-dependent density for each crustal layer is computed in a finite-difference mesh (fine nodes—red dots in Figure 2) using the following equation:
\[
\rho(P, T) = \rho_0 [1 - \alpha(T - T_0) + \beta(P - P_0)] \tag{6}
\]

where \(\rho_0\) (i.e., bulk density at surface conditions) in each crustal layer is a free parameter sampled from its prior distribution during the inversion (see prior ranges in Table 2). At each finite difference node within the crust, lithostatic pressure is calculated at any depth by integrating the weight of the overlying material. This pressure is then used in equation (6) to obtain the density at any node. Note that the computation of pressure and density at each node is done iteratively, as one depends on the other (e.g., Afonso, Fullea, Yang, et al., 2013; Fullea et al., 2009). This density then is used to update the pressure at the next node. When the in-situ (P-T dependent) density is determined within each crustal layer, the \(V_p\) is related to density according to a relationship proposed by Brocher (2005), assuming the validity of Birch’s law of correspondent state as an approximation (cf. Karato, 2008). \(V_s\) is then obtained for each layer from the computed \(V_p\) and the unknown \(V_p/V_s\) parameter sampled by the MCMC algorithm (\(V_s = V_p/V_p/V_s\)).

The thermodynamic database of Xu et al. (2008) and components of the Perple_X software (Connolly, 2009) are used in the Gibbs free-energy minimization stage for obtaining the stable mineral assemblages in the mantle and their relevant bulk properties (velocities, densities, anharmonic elastic moduli, etc) within the system of CFMAS (CaO-FeO-MgO-Al_2O_3-SiO_2). The reader is referred to Afonso, Fullea, Griffin, et al. (2013); Afonso, Fullea, Yang, et al. (2013); Afonso, Rawlinson, et al. (2016) for more information on the compositional priors, their correlations and strategies used during the sampling as well as the forward problems associated with geoid and elevation. Once anharmonic velocities are obtained from the energy minimization problem, anelastic corrections to these velocities are implemented following a modified Burgers model (Jackson & Faul, 2010) calibrated for the region of study (cf. Afonso, Rawlinson, et al., 2016).

Synthetic seismograms for each station are computed with the algorithm of Randall (1989), which uses the matrix formalism of Kennett (1983). For the calculation of radial receiver functions, synthetic seismograms were then subjected to the same processing parameters and steps (e.g., same time domain iterative deconvolution and low-pass filter) as for observations, in order to have the same frequency contents for both observed and calculated receiver functions. Rayleigh wave dispersion curves are computed using a modified version of the program disp96 (Herrmann, 2013), based on the Thompson-Haskell eigenvalue problem (Haskell, 1953).

### 4. Data Fits

#### 4.1. Seismic Data

Figure 3 shows misfit maps for RFs and dispersion curves (square root of equation (5). Considering data and model errors, misfit values > 10 (orange to red colors) for RFs and >4 (yellow to red color) for surface waves are indicative of relatively poor fits to seismic data. Individual examples of fits to seismic data are given in Figure S2 for some illustrative locations (large blue triangles in Figure 1; e.g., station N27A) and also in Figure S3, including stations located on sedimentary basins (e.g., Denver, Black Mesa, and Green River basins). The results in Figure S2 demonstrate that our inversion method and employed parameterization (e.g., a four-layered crust) can successfully fit converted phases associated with crustal discontinuities (e.g., sedimentary layer or Moho) and their related reverberations. For example, the observed receiver function for station N27A in the GP is characterized by the lack of a direct \(P\) wave pulse at zero-lag time, which is a usual feature of sedimentary layers (Chen & Niu, 2016; Shen et al., 2013). This phase and its related multiples are well fitted by the inversion. In Figures S4 and S5 we also show some representative examples that demonstrate that the posterior distributions for the main model parameters are not significantly affected by the chosen prior bounds (given in Table 2).

The relative high misfits observed in some of the RFs can be attributed to three main factors. The first one is due to the strong azimuthal variability in observed RFs. Such high misfits can be seen mostly in the Green River (stations N18A, N20A, and N21A; Figure 3). Unita and the Piceance basins. Stacked RFs calculated from ray paths coming to station N18A from different directions are shown in the Figure S6, which clearly shows a strong azimuthal variation. Therefore, using a simple stack of all RFs from all azimuths produces an unrepresentative trace that is difficult to fit during the inversion.

The second cause for high misfits can be attributed to the chosen (ad-hoc) scaling factors for the different seismic data types and/or odd fluctuations in the dispersion curves in some stations. In such cases, jointly fitting RFs and dispersion curves is difficult and usually leads to the need for relaxing the fit to one type of data in favour of the other. Stations P15A and M19A can be included in this category. For example, the observed
Figure 3. Nonnormalized misfit maps for RFs and dispersion curves. The locations of all seismic stations are denoted by red triangles. Seismic stations that are mentioned in the text are marked. Locations of major sedimentary basins in the study region are also determined: GRB = Green River Basin; UB = Unita Basin; PB = Piceance Basin; DB = Denver Basin; SJB = San Juan Basin; and BMB = Black Mesa Basin.

dispersion curve for station M19A has unusually low velocity at periods 8–10 s (Figure S3) as well as odd irregularities at periods between 50 and 100 s, indicative of processing/quality issues. Since we tried to use the same scaling factors for all stations, those with data irregularities resulted in larger relative misfits.

Finally, although upper mantle discontinuities are usually masked by crustal multiples, especially when a thick sedimentary layer exists, the presence of such seismic discontinuities cannot be ruled out. Since we neglected any possible discontinuity within the upper mantle in our parameterization, the existence of it would affect the fit to RFs. Nevertheless, the large majority of the observed RFs are well fitted within the used time window without requiring mantle discontinuities in the parameterization.

4.2. Nonseismic Data
The fits to nonseismic data (elevation, geoid and heat flow) are shown in Figure 4. Maps in the middle column depict the mean of the posterior distributions of predicted data obtained from the inversion. The maps in the left and right columns represent bounds (as 1σ) in the variability of the observed data sets. Geoid and elevation residuals are within their 1σ uncertainties. Predicted surface heat flow values are well within 1σ of the observed values, except for two isolated locations where the mean predicted heat flow values are slightly smaller than the lower bound of the heat flow data (∼107.5W, 38.5N and ∼112.5W, 42.7N).

5. Main Results and Discussion
All maps and profiles shown in this section depict the mean of the posterior distributions associated with the respective parameter. Information about the uncertainties in these fields are provided in Figures S7–S10.

5.1. Crustal Thickness
Crustal thickness and its associated 2σ uncertainty map, derived from the inversion are given in Figure 5. Estimated uncertainties are relatively low (<2.5 km) in most of the region, with some isolated larger values (>3 km). The resulting crustal thickness map (Figure 5a) shows a thin crust (<32 km) beneath the eastern and southern regions of the BAR province. Within the ATZ province, crustal thickness reaches ~35-40 km, increasing to ≥45 km beneath the CP. Although the CP generally has a relatively thick crust, crustal thickness within this province varies significantly, ranging from ~38 km in its southern and western margins to ~54 km in northeastern margins.

Figure 5a shows that relatively thick crust underlies parts of the SRM and CRM (and the Rocky Mountain Front), particularly beneath the Cheyenne Belt/Park Range area. Our results also indicate that the crust beneath the Front Range is ~4 to 6 km thicker than that beneath the San Juan Mountains and the Sawatch Range, which
is consistent with previous seismic studies (e.g., Prodehl & Pakiser, 1980). In contrast, a thinner crust is found beneath the San Luis Basin (SLB) and the extension of the RGR, ranging from ~42 to 46 km. Thick crust in the Cheyenne Belt seems to extend into the central Rockies. However, the crust north of the Cheyenne belt (central Rockies) thins toward the west, reaching ~38 km near the Wasatch Range. Similar to the CP, the GP are also characterized by relatively thick crust with some local variabilities.

Sharp gradients in the crustal thickness are observed mostly within the transition domains from one province to another. For instance, in the transition from the RGR to the SRM and the GP, the crust thickens abruptly from

Figure 4. Fit to nonseismic data (geoid, elevation, and heat flow). Middle column: mean of the posterior distributions for predicted data. The left and right columns contain the mean of the observations minus and plus 1σ uncertainty, respectively.
Figure 5. (a) Average crustal thickness map with topography superimposed as transparency. Some of the seismic stations whose crustal thickness values agree closely with those calculated by Bashir et al. (2011) are also shown. (b) Uncertainties in crustal thickness as $2\sigma$ from the posterior probability density function.

Despite some localized differences (e.g., near the Kaibab uplift, west of the Grand Canyon), the overall pattern of crustal thickness derived here is similar to that reported in numerous previous studies (e.g., Bashir et al., 2011; Frassetto et al., 2006; Gilbert & Sheehan, 2004; Gilbert et al., 2007; Gilbert, 2012; Hansen et al., 2013; Lowry & Pérez-Gussinyé, 2011; Prodehl & Pakiser, 1980; Purevsuren, 2014; Sheehan et al., 1995; Shen et al., 2013; Velasco, 2009; Zandt et al., 1995), pointing towards a general agreement for the region. For example, thick crust ($\sim 46$ km) found beneath the SRM (e.g., Bailey et al., 2012; Gilbert & Sheehan, 2004; Gilbert, 2012; Prodehl & Pakiser, 1980; Shen et al., 2013; Shen & Ritzwoller, 2016) is broadly consistent with our results. Our crustal thickness map in the GP is also compatible with those given by Shen et al. (2013); Shen and Ritzwoller (2016) and Gilbert (2012). As another example, our estimates of crustal thickness are in the range of $\sim 38$ to $\sim 44$ km. Other examples with such sharp variations are the boundaries between the CP and the ATZ and NBR provinces as well as near the Wasatch Range (WF).
excellent agreement with those obtained by Frassetto et al. (2006) in the CP. Likewise, our estimates beneath stations shown in Figure 5a (except for R17A) agree closely with the crustal thicknesses calculated by Bashir et al. (2014), particularly in the western margins of the CP where a thick crust (~50 km) was reported by this author.

5.2. Temperature Structure

Figure 6 shows the thermal structure as a series of depth slices from 10 to 60 km depth; all values are the means of the posterior PDF and associated uncertainties are shown in Figure S10. The SRM and margins of the CP show larger lateral and vertical geothermal gradients relative to the GP and the central Rockies. Lateral thermal variations up to ~200–300 °C are estimated in the crust between the SRM and the interior of the CP or GP at depths ≥15 km. This lateral temperature difference is larger towards the north and east of the SRM, reaching maximum values of ~400–600 °C at 40 km depth; they are also larger than the corresponding uncertainties obtained from the posterior PDF (Figure S10). Interestingly, the thermal structure of the CP does not follow well the physiography of the province as indicated by surface features (black lines in Figure 6). While the core of the CP is relatively cold at all depths, its margins are significantly warmer. This is mainly related to the thin lithosphere that characterizes these regions (e.g., Afonso, Rawlinson, et al., 2016; Hansen et al., 2015; Levander et al., 2011). The locations of recent basaltic volcanism within the CP province and its immediate surroundings (dots in Figure 6) also correlate well with regions of relatively high temperature. The estimated temperature gradients in the western and southern edges of the CP are of the order of 15–20 °C/km at middle and lower crustal depths. The BAR, the ATZ, and the RGR are also characterized by high temperatures at depths ≥20 km comparable with the temperatures obtained for the SRM.

Beneath the SRM, temperature values at ~40 km depth (Figure 6) are above the wet solidus of most crustal rocks, suggesting that the existence of melts at these depths is likely (Afonso, Rawlinson, et al., 2016; Levandowski et al., 2014). If this is true, our temperatures are expected to be overestimated, as the parameterization used in this study does not consider the effects of partial melting on crustal seismic velocities. Within this context, Afonso, Rawlinson, et al. (2016), which included partial melt effects in their 3-D study (albeit in a simplified way), obtained temperatures of the order of 100–200 °C lower than those reported here beneath the SRM. Therefore, our highest temperatures are likely to represent upper bounds. A more sophisticated modeling of partial melting effects within the crust is left for a future study.

A peculiar observation is that the crust of the San Francisco Volcanic Field (SFVF) seems to be slightly colder than other volcanic regions in the study area. This was also pointed out in Qashqai et al. (2016). However, this does not mean that the SFVF is a “cold” continental area. In fact, the estimated temperatures are high enough to consider the SFVF anomalously hot in comparison to average continental crust. Moreover, while at first glance it may seem that the relatively lower surface heat flow values measured in this region (likely associated with groundwater flow, Paul Morgan et al., 2004; Shearer & Reiter, 1981; Sass et al., 1982) are controlling our results, this is not the case, as this observable puts only mild constraints on the final thermal structure in comparison to seismic and geoid data. In this context, it has been shown that the effect of compositional heterogeneities on temperature structure within the crust may result in variations of ≥100 °C at middle crustal depths over short distances (Furlong & Chapman, 1987). Considering therefore (1) that this region is bounded by important structural lineaments (e.g., Holbrook Lineament—HL) and (2) its velocity, density and Vp/Vs structures (sections 5.3 and 5.6), it seems plausible that its relative colder structure is the direct reflection of a distinct crustal bulk composition. This interpretation is also supported by the large magnetic and density contrasts across the Holbrook lineament (Bouligand et al., 2009; Hendricks & Plescia, 1991).

At uppermost mantle depths (e.g., depths 50 and 60 km), places where temperatures reach values close to 1200 °C include the SRM, the RGR, the ATZ, the BAR, the margins of the CP and the eastern end of the Jemez lineament in the GP. Such high temperatures in these regions imply that the lithosphere-asthenosphere boundary (LAB) is not much deeper than 60–70 km, which is in agreement with the estimates of Levander et al. (2011); Liu et al. (2011); and Afonso, Rawlinson, et al. (2016). In the southeast border of the CP, Klöcking et al. (2017) has recently reported similar values for the LAB based on trace-element modeling of volcanic rocks. This is also in agreement with the high crustal temperatures and high heat flow values (Decker, 1995; Minier & Reiter, 1991) in these provinces. In contrast, lower temperature values beneath the center of the CP and the GP at 50-60 km depths clearly point toward a thicker lithosphere for these regions, consistent with numerous previous estimates (e.g., Lee et al., 2001; Liu et al., 2011; Roy et al., 2009) and the lack of significant internal deformation (Bashir et al., 2011; Morgan & Swanberg, 1985). Interestingly, the temperature profile for
Figure 6. Depth slices from 3-D temperature model. Locations of major volcanic fields are also shown (red circles: <5 Ma, green circles: 5-10 Ma; SFVF: San Francisco Volcanic Field).
station O24A in the western edge of the GP (Figure S2) shows significantly higher temperatures than other stations in the GP at 30- to 40-km depths (e.g., N27A, S29A; Figures 1 and S2). Yet there are no significant changes in \( v_p \) among these stations, which suggests the presence of compositional differences (section 5.6).

### 5.3. Seismic Velocity and Density Structures

Figures 7–10 show our results for \( v_p \) and density structures. As compressional wavespeeds correlates with density (section 3.6), we do not show \( v_p \) results here, but we include them in Figures S11–S12. To better illustrate the complex horizontal and vertical patterns in seismic velocities and density, we show both depth slices (Figures 7 and 9) and vertical cross sections along eight illustrative profiles (Figures 8 and 10). We discuss these below.

#### 5.3.1. Depth Slices and Vertical Profiles

One of the most notable features in these models are the strong low velocity (~3.4–3.6 km/s for \( v_p \) and ~5.7–6.7 km/s for \( v_s \)) and low density (~2,750–2,850 kg/m\(^3\)) anomalies from upper to lower crustal depths (~2–40 km) beneath the SRM, around the CMB and the SJM (Figures 7–10). This low velocity and density anomaly seems to have an “inverted cone” shape, more visible in the \( v_s \) structure (Figure 8). In this region, one of the largest and most negative Bouguer anomalies within North America (Figure S1) is believed to be caused by low density granitic batholith bodies (e.g., McCoy, Roy, et al., 2013; Snelson et al., 2013). Our crustal density structure along profiles B, C, E, F, G, and H illustrates and supports this feature. Also, the density model along profile G is compatible with that estimated by Snelson et al. (2013) in this region. We will further discuss this in the next section.

Low velocity (and density) anomalies can also be seen at the boundaries of the CP with the neighboring provinces (excluding the northern margins), as well as beneath the ATZ, RGR, and the BAR provinces (see the depth slices at 30 km), coincident with the locations of Cenozoic volcanism and high crustal temperatures (shown in Figure 6). Compared to the SRM, the center of the CP, the Central Rockies and the GP are characterized by seismically faster crustal velocities and significantly higher crustal densities at depths ~20 km. For instance, in these regions (e.g., northwest of the Four Corners area: latitude: ~37°N, Longitude: ~109°W), relatively high crustal \( v_p \) (~3.7–3.8), \( v_s \) (~6.8–7.0 km/s), and densities (~2,900–3,000 kg/m\(^3\)) are visible, reaching values of ~4.0–4.2 km/s, 7.1–7.2 km/s, and ~3,100 kg/m\(^3\) for \( v_p \), \( v_s \), and density, respectively, at ~40 km depth (see profiles B, E, F, H, and J in Figures 8, 10, and S10).

Another prominent feature of the results is that the velocity structure along profile G (latitude >39°N) shows sharp lateral variations from the south of the Cheyenne Belt to its northern boundaries. For instance, \( v_p \) varies from ~6.8 km/s in the south of the Cheyenne Belt to ~7.3 km/s in its northern boundaries, corroborating results from previous studies (e.g., Gorman et al., 2002; Snelson et al., 2013). Since \( v_p \) and density are correlated in our inversion (see section 3.6), the density structure in the lower crust also varies sharply along profile G from ~2,900 (at ~41N) to ~3,100 kg/m\(^3\) near the latitude of 43°N. This agrees well with the density of ~3.170 kg/m\(^3\) calculated by Jones et al. (2015), who analyzed the crustal xenoliths from several Tertiary magmatic centers along a profile from Montana to Colorado.

#### 5.3.2. Comparison of the \( v_p \) Model With Previous Models

Although our \( v_p \) model is largely compatible with a number of recent models (e.g., Bailey et al., 2012; Li et al., 2002; Shen et al., 2013; Schmandt et al., 2015; Shen & Ritzwoller, 2016), some slight discrepancies exist. For example, the absolute variability in \( v_p \) reported in Bailey et al. (2012) beneath the center of the CP is larger than that obtained here. Similarly, Shen et al. (2013) reports a crustal velocity of ~3.9 km/s at 40 km depth beneath the south-central of the CP (Figure 11, profile DD’ in Shen et al., 2013) whereas we obtain ~4.0–4.2 km/s, comparable to the \( v_s \) beneath the GP (e.g., see Figure 8, profile A). This difference is interesting because we used the same Rayleigh wave dispersion curves for periods <40 s. The above inconsistencies can therefore be attributed to the different constraints and parameterizations used in their inversions, such as constant \( v_p/V_s \), (e.g., Bailey et al., 2012; Shen et al., 2013), smooth splines and forced positive gradient beneath the Moho (e.g., Shen et al., 2013) and/or the lack of short period phase velocities (e.g., Bailey et al., 2012), for example, periods <20s, which may lead to spurious results in the upper and middle crust.

### 5.4. Mafic Lithologies and Discrepancies in Crustal Thickness in the CP

Discrepancies among previous crustal thickness models for the western, central and southern margins of the CP (e.g., Bailey et al., 2012; Bashir et al., 2011; Gilbert, 2012; Lowry & Pérez-Gussinyé, 2011; Shen et al., 2013) can be attributed to at least two related factors. The first one is the presence of mafic lithologies in the lower crust.
Figure 7. Depth slices of shear-wave velocity model at 10-, 20-, 30-, 40-, 50-, and 60-km depths. Profiles A–J, needed for the next figure, are shown as well by dashed lines. Symbols for volcanism are as in Figure 6.
**Figure 8.** Vertical cross sections of $V_s$ along the profiles shown in Figure 7. Tectonic provinces and major lineaments are also shown (e.g., HL = Holbrook Lineaments—see Figure 1). The Moho along each profile is shown by a dashed line.
Figure 9. Depth slices of density model at 10-, 20-, 30-, 40-, 50-, and 60-km depths. Profiles A–J, needed for the next figure, are shown as well by dashed lines.
Figure 10. Vertical cross-sections of density along the profiles shown in Figure 9. The Moho along each profile is shown by a dashed line. Texts and symbols are as Figure 8.
The presence of high velocity mafic lower crust would not only reduce the velocity contrast across the Moho, but it would also produce a marked contrast at the top of the mafic layer, making the Moho signature in RFs less apparent compared to that from the top of the mafic layer (when the latter is overlaid by more felsic material). Gilbert (2012) has discussed this difficulty and showed multiple low-amplitude receiver function arrivals between depths of 30 and 50 km within the CP. Therefore, depending on the thickness of the high-velocity layer and differences in data processing, a range of Moho depths could result from RFs analysis alone.

The second factor can be ascribed to the low velocity in the sub-Moho layer(s), which can be caused, for example, by high temperatures or alteration. The effect of such a sub-Moho layer in RFs, if it is overlaid by high-velocity lower crustal materials, would be an unclear (weak and low amplitude) Moho converted phase. In such circumstances (i.e., high-velocity lower crust and/or reduced velocity contrast across the Moho), joint inversions such as those performed in this study, can help narrow the range of acceptable models and thus provide a more representative estimate of the true structure. In this context, our results suggest that at least in some parts of the study region, the Moho depth selected in Gilbert (2012) may actually be the top of a mafic lower crust layer.

Two typical examples of such cases can be seen beneath stations W17A and R17A (see Figure 5 for locations), where both conditions apply: (i) crustal $V_s$, $V_p$ and density maps show that mafic lithology exist beneath the CP at depths $\geq$ 20 km, consistent with the previous geophysical (e.g., Bailey et al., 2012; Bashir et al., 2011; Wilson et al., 2010; Wolf & Cipar, 1993; Zandt et al., 1995), geochemical and crustal xenoliths studies (e.g., Condie & Selverstone, 1999; McGechin & Silver, 1972; Nelson et al., 1992; Nelson & Davidson, 1993) (see Figures 7–10 and S11–S12); (ii) the upper mantle temperature maps (e.g., at 50 and 60 km) beneath the most of western CP show higher temperatures compared to the east and north of the plateau (~1200 °C at 60 km), suggesting a thin and relatively hot lithosphere beneath the southern and western CP. This leads to low sub-Moho velocities beneath the two stations. The fits to the observed seismic data as well as resulting 1-D temperature and S-wave velocity structures for these stations (R17A and W17A) are provided in Figure S13.

In relation to the presence of the high density lower crust beneath the CP, the observed Bouguer gravity anomaly map in Figure S1 shows a sudden increase from the southeast of the plateau to the west and southwest of the Four Corners region (~ by 50 mGal). This rapid increase in a short distance supports that the source of this variation should be in the crust. The existence of such a silica-poor and anomalously high density thick lower crust beneath the center of the plateau, compared to the estimates of crustal composition beneath neighboring provinces, could be one of the reasons for its enigmatic tectonic stability (Bashir et al., 2011; Reid et al., 2012; Roy et al., 2009).

5.5. Gravity Low Beneath the SRM

Figure S1 shows the most negative Bouguer anomaly (~ $-225$ mGal) in the SRM centered on the CMB, especially beneath the SJM, the SR and the FR. Numerous studies have attempted to map the source of these anomalies using gravity modeling (Drenth et al., 2012; McCoy, Roy, et al., 2013; Rumpfhuber et al., 2009; Snelson et al., 2013), resulting in conflicting density structures.

Our estimates of crustal densities, show pronounced low density values underneath the SRM, correlating well with the remarkable gravity low observed in this region. We note here that while we have inverted for long-wavelength geoid anomalies, no input data on gravity anomalies have been considered. Therefore, in order to assess whether our density model is also consistent with observed Bouguer gravity anomalies, we model 2-D gravity anomalies along profile G (Figure 11) using our density model. Observed gravity anomalies were taken from the WGM2012 model (Bonvalot et al., 2012) and wavelengths <50 km were filtered out. Importantly, we allowed densities to vary only within 1σ of their posterior PDF (supporting information). Despite the relatively coarse resolution of our models, it shows a relatively good fit ($\pm 15$ mGal) to the observed Bouguer anomalies, particularly the lowest gravity beneath the FR (near ~39.5N).

The large-scale shallow low density body (~2,700–2,750 kg/m$^3$) and the high density layers (~2,850–2,900 kg/m$^3$) at depths ~20–40 km, revealed by the inversion, are needed to fit the observed short wavelengths (low and high) gravity anomalies along the profile (e.g., underneath south of 35N, the Wet Mountains-WM and north of 39N in Figure 11). High density materials in north of 39N are consistent with exposed metavolcanic and large mafic plutonic rocks seen in the Park Range (see Figure 1b for location) of Colorado (Foster et al., 1999; Snelson et al., 2013). We emphasize here that, contrary to what is typically
assumed in gravity modelling, the lateral density changes within the shallow lithospheric mantle play a 
non-negligible role in fitting the long-wavelength anomalies.

The first-order density model along profile G is compatible with the density structure modeled from seismic 
refraction/wide angle reflection study, “the Continental Dynamics of the Rocky Mountains (CD-ROM) 
experiment” (Snelson et al., 2013). For example, the 2,900–3,000 kg/m³ or 2,950–3,100 kg/m³ bodies (or layered 
lower crust) discovered here in the south of the Wet Mountains (~38N) and north of the Cheyenne Belt (~41N), 
respectively, are very similar to the high density lowermost crust proposed by Snelson et al. (2013). In addi-
tion, despite its comparatively lower resolution, the density structure in our model correlates well with the 
inter-wedging structure (shown by dashed blue lines in Figure 10) imaged along profile G by seismic reflection 
data (Tyson et al., 2002).

5.6. \( V_p/V_s \) Structure

The \( V_p/V_s \) ratio for common rock types varies from ~1.63 for felsic rocks to ~2.1 for mafic and ultra mafic 
rocks or unconsolidated sediments (Christensen, 1996; Fountain et al., 1992; Wang et al., 2009; Zandt & Ammon, 
1995) and is commonly considered a useful proxy for bulk crustal composition (Chevrot & Van Der Hilst, 2000; 
Christensen, 1996; Gercek, 2007; Liu & Gao, 2010; Nair et al., 2006; Shillington et al., 2013; Tarkov & Vavakin, 
1982; Zandt & Ammon, 1995). However, it also known that the presence of melts/fluids can significantly affect 
the \( V_p/V_s \) ratio and mask the bulk compositional signature (Caricchi et al., 2008; Chatterjee et al., 1985; Hacker 
et al., 2014; Johnson & Poland, 2013; Nakajima et al., 2001; Owens & Zandt, 1997; Takei, 2002; Tripoli et al., 
2016; Ueki & Iwamori, 2016; Zhang et al., 2014). With this caveat in mind, we will make use of our Vp/Vs model 
(Figures 12-13) to provide some first-order constraints into possible compositional structures for the crust in 
the study region. We note that the crustal \( V_p/V_s \) ratio is a direct output of our inversion.

Average \( V_p/V_s \) values (Figure 12) vary across the region from ~1.68 up to 1.85, well within the common range 
found in continents (Zandt & Ammon, 1995). Higher values (~ > 1.78) are mostly found beneath the eastern 
and western parts of the CP, the SRM, the RGR, and in some parts of the GP. Average \( V_p/V_s \) values agree well 
with those obtained in previous receiver functions studies, especially the high \( V_p/V_s \) values (~1.75–1.85) in 
the SRM (e.g., Bashir et al., 2011; Frassetto et al., 2006; Gilbert & Sheehan, 2004; Purevsuren, 2014). However, 
our results in the CP, the GP, the Southern and Central Rockies differ from the \( V_p/V_s \) ratios ≤1.72 derived by 
Buehler and Shearer (2014) from a joint analysis of Pn and Sn phases. Also, we find no evidence for low \( V_p/V_s \) 
values beneath the SRM, as reported in Lowry and Pérez-Gussinyé (2011).
While the average crustal $V_p/V_s$ is useful for comparison with previous works, it smears out important information on the vertical structure of the crust. We therefore show cross-sections of $V_p/V_s$ along profiles A–J (shown in Figure 12) in Figure 13. In addition, maps of $V_p$, $V_s$ and $V_p/V_s$ for each of the crustal layers are given in Figures S14–S16. At upper crustal depths ($\leq 10$–15 km), $V_p/V_s$ is restricted to values $\leq 1.73$ in most of the region, indicative of a dominant felsic upper crust. For example, $V_p/V_s$ values at these depths beneath the SJM ($V_p/V_s \sim 1.65$–1.67; $V_s \sim 5.8$–5.9) are compatible with values associated with (low-density) silicic batholith bodies that were modelled and interpreted by Drenth et al. (2012) in this area. There are, however, a few exceptions, perhaps the most notable being the Cheyenne Belt area (e.g., along profiles D and G; 106W, 42N), where significantly higher values are observed. This seems consistent with the geology and structure of this shear zone (cf. Houston et al., 1989).

In addition to silica content, cracks with small volumes of aqueous fluids can also play an important role in reducing $V_p/V_s$ values, as discussed by Takei (2002), Nakajima et al. (2001) and Hauksson and Unruh (2007). These factors tend to reduce $V_p$ more than $V_s$, resulting in bulk $V_p/V_s$ values. The modeling of Ma and Lowry (2017) and Guerri et al. (2015) showed that hydration processes can also decrease $V_p/V_s$ values in the crust, except when melt is present. While a recent magnetotelluric study in the SRM (Feucht et al., 2017) does not report a significant reduction in bulk resistivity within the upper crust in this region, felsic materials containing unconnected cracks filled with $H_2O$-rich fluids can be an alternative interpretation for the low $V_p/V_s$ low $V_p$ and low $V_p/V_s$ values observed at shallow depths ($\leq 10$–15 km) beneath the SRM and southern margins of the CP. Similar conditions have been reported by Zhang and Lin (2014) beneath the Coso geothermal area in California.

Interestingly, our results at depths $\geq 15$ km (i.e., middle and lower crustal depths) beneath the SRM and southern boundaries of the CP do not confirm the values reported in Lowry and Pérez-Gussinyé (2011) and Ma and Lowry (2017). At these depths, $V_p/V_s$ varies between 1.75–1.9, suggesting the presence of either intermediate to mafic lithologies, high temperatures and/or partial melts. This is a robust feature of our $V_p/V_s$ model (e.g., cross sections in Figure 13) in the SRM, the margins of the CP and the RGR system. These regions are characterized by relatively low $V_p$ and low $V_s$ and elevated $V_p/V_s$ values ($\geq 1.85$–1.9; see profiles A, B, E, F, G, H, and J, as well as the depth slices at 30 and 40 km in Figures 7, 8, and S11 and S12). Since the wavespeeds are relatively low, mafic compositions at these depths seem unlikely as an explanation. The presence of small amounts of melts or fluids can better explain this feature. As argued by Chatterjee et al. (1985), Ferrari et al. (2016), and others (e.g., Christensen, 1996; Dawson et al., 1999; Hacker et al., 2014; Lin et al., 2014; Mueller & Massonne, 2001; Nakajima et al., 2001; Owens & Zandt, 1997; Xu et al., 2007), when melting commences both $V_p$ and $V_s$ are decreased, but $V_s$ is decreased more rapidly than $V_p$, thus increasing the total $V_p/V_s$ ratio. These observations and interpretation are consistent with the recent MT results by Feucht et al. (2017), who imaged an anisotropic high conductivity region in the middle and lower crust of the SRM, which they interpreted as strike-parallel fractures/dikes along a north-south direction containing melts and/or aqueous fluids. Although the resolution of our parameterization does not allow us to capture such small-scale fracture zones, we note that the regions of significantly high $V_p/V_s$ values beneath the SRM and the RGR system (latitude $\sim 39$N and longitude $\sim 106$W; e.g., profiles B, C, D, E, F, G, and H in Figure 13) coincide well with the regions of high conductivity reported by Feucht et al. (2017). The spatial correlation between low wavespeeds, high $V_p/V_s$, elevated temperatures (see Figure 6) and the locations of Cenozoic volcanism in the SRM and margins of the CP, offers compelling support for the occurrence of partial melting within the middle to lower crustal depths in these regions (e.g., beneath the Jemez zone and the SRM). Thermochronologic and magnetotelluric studies in the Jemez zone (e.g., Ander et al., 1984; Feucht et al., 2017; Shaw et al., 2013) are also consistent with this interpretation.

### 5.7. Crustal $V_p/V_s$ VS Crustal Thickness and Elevation

Figure 14 shows scatter plots of the averaged crustal thickness versus crustal $V_p/V_s$ and observed topography for the study region. If we neglect the two outliers (113.1W,35.2N and 112.3W,33.95N) with thin crust ($\sim 33$ km).
Figure 13. Vertical cross sections extracted from 3-D $V_p/V_s$ model along profiles A–J (shown in Figure 12). Texts in the vertical cross sections are as in Figure 8.

and 30 km) and high $V_p/V_s$ ratios (∼1.81 and 1.825) in the ATZ, Figure 14a shows that a number of provinces (BR, CRM and SRM are exceptions) show only a weak positive trend.

A scatter plot of crustal thickness versus elevation (Figure 14b) shows no clear correlation between these two parameters in the GP and SRM. In the CRM, while the topography remains nearly constant, the crustal thickness increases by ∼15 km from ∼40 km to approximately 55 km. Topography in the ATZ, however, have a positive linear relationship with crustal thickness (correlation coefficient or $R ≃ 0.51$; see also Qashqai et al., 2016). This plot for stations with elevations >1,200 m (in the ATZ) demonstrates that an addition of 15 km (from ∼30 to ∼45 km) crustal material corresponds to 1-km increase in elevations (from ∼1,250 to ∼2,250 m). This
Figure 14. (a) Scatter plot of crustal thickness versus $V_p/V_s$ ratio for all tectonic provinces. If the two points with high $V_p/V_s$ and low crustal thickness values are removed, the plot for the ATZ shows a strong positive trend with correlation coefficient (or R) of $\sim 0.7$. Correlation coefficients for others are lower than 0.5 ($R_{CP} = 0.13$, $R_{GP} = 0.32$, $R_{RRG} = 0.48$). Fitting a positive trend/line to all points in this plot gives $R = 0.26$. (b) Scatter plot of crustal thickness versus surface elevation for all provinces. Tectonic provinces are denoted by different symbols.

plot also shows that a broad positive correlation between elevation and crustal thickness can be argued for the region as a whole when the stations with low relative elevations (<1,500 m) in the GP are excluded. A possible explanation for the departure of these low-elevation stations from a positive trend, as well as for the lack of a clear overall correlation, is that the mantle plays a significant role in supporting the observed elevations (either via compositional, thermal, or dynamic components). This is consistent with the highly heterogeneous LAB imaged in this region (e.g., Afonso, Rawlinson, et al., 2016). In addition, if significant lateral compositional (and density) anomalies exist within the crust (as suggested by our results), it is expected that each compositional domain will have a distinct trend (i.e., different slope). Therefore, combining a small number of stations with distinct compositions into a single plot may result in a lack of overall correlation.

5.8. Isostatic Support

The lack of correlation between observed elevations and crustal thickness in the SRM and GP (Figure 14b) points towards internal crustal or mantle density anomalies supporting the high elevations. However, separating the role of the crust is subject to simultaneously determining crustal and upper mantle structures. Several studies have attempted to achieve this using different methods, which have led to contrasting results (e.g., Bailey et al., 2012; Boyd & Sheehan, 2013; Coblentz et al., 2011; Gilbert & Sheehan, 2004; Hansen et al., 2013; Lee & Grand, 1996; Levandowski et al., 2014; Li et al., 2002; Sheehan et al., 1995).

In this section, we use our inverted density and crustal thickness models in a regular grid of 0.5° by 0.5° to isolate the first-order isostatic contribution of the crust in supporting the high elevations in the SRM relative to the adjacent GP. We employ the principle of crustal isostasy (Figure 15a), considering the effect of lateral density and crustal thickness variations simultaneously, to derive an equation that predicts the elevation at each point relative to a reference column, $(\Delta E_i)_p$:

$$(\Delta E_i)_p = \left( \frac{\rho_m - \rho_i}{\rho_i} \right) h + \left( \frac{\rho_i - \rho_r}{\rho_i} \right) t$$

where $\rho_i$ represents the density of the $i$th column/point, $\rho_r$ and $t$ are the reference density and crustal thickness values, respectively; $h$ is the amount of crustal thickening for column $i$ and $\rho_m$ is lithospheric density (assumed constant, $\rho_m = 3300$ kg/m$^3$). Densities, observed elevations and crustal thicknesses in the GP are averaged and
considered as reference values \((\rho_r = 2,900 \text{ kg/m}^3, t = 46.77 \text{ km}, \text{average elevation} = 1,226.2 \text{ m})\). We can see that equation (7) is a combination of the Airy and Pratt-Hayford hypotheses. If there is no lateral density changes \((\rho_i = \rho_r)\) then the second term would be zero and the topography will be predicted based on pure Airy crustal model. On the contrary, if no crustal thickening occurs \((i.e., h = 0)\), equation (7) reduces to the Pratt-Hayford crustal root model.

A weighted average (based on the thickness of each crustal layer) is used for calculating average crustal densities (Figure S17). The amount of topographic relief (in %), predicted only by crustal structure \((\Delta E_i)^p\), as well as the residual (misfit) between the observed and predicted topographic differences for each point \(i\), are given as maps in Figure 15b. Here \((\Delta E_i)^o\) is the observed topography difference between the reference area and each point \(i\). The misfit in the right panel of Figure 15b \((\delta \epsilon)\), refers to the quantity

\[
(\delta \epsilon) = (\Delta E_i)^o - (\Delta E_i)^p
\]

We only show values between 0 and 100 in Figure 15b (left). According to equations (7) and (8), negative \((\Delta E_i)^p\) can be caused by crustal thinning \((h < 0)\) or density variation \((\rho_i < \rho_r)\) and \((\delta \epsilon) < 0\) results when predicted values of \((\Delta E_i)^p\) are greater than the actual \((\Delta E_i)^o\).

Figure 15b (left) implies that both crustal thickness and lateral density variations play significant roles in controlling the isostatic compensation of the SRM. This is particularly evident between 38° N and 40° N, where thick crust and low crustal densities exist (e.g., the SF, FR, and the Park Range). Overall, crustal structure in the SRM and the CRM is responsible for \(\approx\)30–80% of the isostatic support. The lower values for the RGR \((\sim 30–50%)\) are largely consequence of a thinner and less dense crust relative to the reference values.

The mantle contribution to the topography of the SRM has been discussed by a number of authors before. While some have concluded that the mantle plays a dominant role (e.g., Coblenz et al., 2011; Sheehan et al., 1995), other have favoured crustal structure as the main source of buoyancy (e.g., Bailey et al., 2012; Becker et al., 2013; Hansen et al., 2013; Levandowski et al., 2014; Li et al., 2002). As stated above, our results suggest...
that the present-day topography in the SRM is supported predominantly by the crustal structure at latitudes >38°N, whereas those regions to the south of 38°N are largely supported by the mantle. Although not entirely surprising given the similarity of the methods used, this result is consistent with the lithospheric structure and sublithospheric flow pattern obtained by Afonso, Rawlinson, et al. (2016). We also note here that the plot in Figure 15c supports the results and suggestions made by Hansen et al. (2013), who also found a strong negative correlation between elevation and mean crustal density.

The contribution of crustal structure to topographic relief in other provinces are found to be notably smaller (~<30% or even negative), particularly in places where the density is greater or the crust is thinner than the reference values. In particular, it appears that most of the isostatic support in the plateau come from the upper mantle rather than the crust. This result complements those reported by Afonso, Rawlinson, et al. (2016), who found that most of the topography in the CP was due to lithospheric contributions (lumping crust and lithospheric mantle together), with some localized dynamic (i.e., sublithospheric convection) contributions, mostly around the edges of the plateau.

6. Concluding Remarks

We have simultaneously inverted Rayleigh wave phase velocity dispersion curves, P-to-S receiver functions, surface heat flow, geoid height and elevation data to characterize the thermophysical state (temperature, seismic velocities, density, etc.) of the crust and uppermost mantle beneath the SRM and surrounding provinces. Our main findings can be summarized as follows:

- Despite some localized discrepancies, our crustal thickness and \( V_s \) maps corroborate previous estimates in most of the Colorado Plateau. Thick crust (≥50 km) is found beneath most of the major uplifted areas in the western side of the plateau.
- Thick crust (~48–56) underlies the SRM, comparable to that of the adjacent GP, especially around the Cheyenne Belt, the Frontal Range and the Rocky Mountain Front regions.
- The temperature structure of the region is highly heterogeneous at most crustal levels, showing differences of up to ~400–600 °C at 40 km depth. Major thermal anomalies are found at lower crust levels beneath the SRM and margins of the CP.
- Low seismic wave speeds together with low \( V_p/V_s \) values at shallow crustal depths beneath the SRM suggest a felsic upper crust containing fractures/cracks filled with aqueous fluids. At middle to lower crustal depths, however, our models show a large-scale thermal anomaly beneath parts of the SRM and margins of the CP, consistent with the high \( V_p/V_s \) values, anomalously low crustal velocities and low densities retrieved by the inversion. In the SRM, these features correlate with high electrical conductivities found at depths ≥ 25 km by Feucht et al. (2017) and the location of recent volcanism; taken together, these observations suggest the presence of melt/liquids in the mid/lower crust beneath parts of the SRM.
- The high topography of the SRM is supported by a combination of crustal and mantle contributions. North of 38°N, crustal structure is the main contributor to the high elevations, whereas to the south of this latitude, mantle buoyancy sources (both static and dynamic) provide the largest contribution. The latter is also true for other provinces (i.e., CP, ATZ, and BAR).
- The horizontal and vertical patterns of \( V_p/V_s \) in the study region are complex and reflect a highly heterogeneous crust in bulk composition. In some regions, the Vp/Vs ratio seems to be affected by more than one factor (i.e., temperature, fluids, melt, and lithology), complicating its interpretation as a compositional proxy. We do not find low \( V_p/V_s \) values at mid/lower crustal depths beneath the SRM that may suggest felsic lithologies.

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