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# Multi-Frequency Inversion of Ps and Sp Receiver Functions: Methodology and Application to USArray Data

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1           **Multi-Frequency Inversion of Ps and Sp Receiver**  
2           **Functions: Methodology and Application to USArray**  
3           **Data**

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13           **Key Points:**

- 14           • We describe a method for inverting multi-frequency P-to-s and S-to-p receiver func-
- 15           tions for mantle thermo-chemical structure
- 16           • We distinguish tectonic regions based on thermo-chemical characteristics across
- 17           the USA from the active West to the cratonic-orogenic East
- 18           • We observe strong thermo-chemically induced and interrelated topographic vari-
- 19           ations of the lithospheric base and the 410-km discontinuity

## Abstract

We image the thermo-chemical structure of crust and mantle underneath the North American continent by inverting recordings of P-to-s (Ps) and S-to-p (Sp) converted seismic body waves (receiver functions; RF). Through careful data selection and processing, we construct a multi-frequency Ps (5, 8, 10 s) and Sp (10, 15 s) RF data set from USArray recordings. The inversion is interfaced with petrological phase equilibria computations to build self-consistent radial seismic velocity and density models for RF waveform simulations. Inverted models are combined through back-projection along converted ray-paths and interpolation to tomographic images of crust and mantle structure. Through clustering analysis we identify three major tectonic regions based on mantle thermo-chemical and seismic structure: the tectonically active West (TAW), the central transition regions (CTR), and the cratonic-orogenic East (COE). TAW is chemically more fertile with a  $Mg\# \sim 0.90$  (molar  $Mg\# = Mg/(Mg+Fe)$ ) and an elevated mantle potential temperature of  $1490 \pm 27$  °C relative to COE, which is chemically more depleted ( $Mg\# \sim 0.91$ ) and colder ( $1419 \pm 27$  °C). CTR is intermediate to TAW and COE. We find significant thermo-chemically-induced topography associated with the base of the lithosphere ( $\pm 90$  km), while the mantle transition zone is mostly influenced by thermally-induced topography on the 410-km discontinuity ( $\pm 15$  km). In contrast, the 660-km discontinuity, where variations are only  $\pm 5$  km, reflects a more complex thermo-chemical interplay. To place the results in a tectonic context, thermobarometric estimates from basaltic rocks across the western USA are integrated with the seismic inversions to produce a thermal model of the underlying mantle.

## 1 Introduction

Earth's thermal and compositional structure is the result of wide-ranging dynamical processes that have shaped and continue to shape the planet's surface and interior. Seismology provides tools for imaging traces of these processes, such as subducting slabs, thick continental roots, and mantle plumes, through their influence on material properties and thus seismic wave velocities and propagation.

These geologic features can be seismically imaged using converted, reflected, and refracted body waves (e.g. Farra & Vinnik, 2000; Flanagan & Shearer, 1998; Tan & Helmberger, 2007). For example, the construction of receiver functions (RFs) allows to extract timing and amplitude of P-to-s (Ps) and S-to-p (Sp) converted phases, which serve as proxy for imaging depth and sharpness of seismic discontinuities beneath a station, respectively (cf. Figure 1; Vinnik, 1977; Langston, 1979). Important seismic discontinuities include the base of the crust (Moho), the lithosphere-asthenosphere boundary (LAB), and the bounding discontinuities of the mantle transition zone (MTZ).

While the LAB primarily marks a change in thermal and rheological regime (e.g., Jaupart & Mareschal, 1999; Artemieva & Mooney, 2001; Eaton et al., 2009), two globally observable discontinuities bound the MTZ that are associated with the transformation of the mantle minerals olivine to wadsleyite and ringwoodite to bridgemanite+ferroperricline at depths of  $\sim 410$  and  $\sim 660$  km depth, respectively (Ringwood, 1975; Dziewonski & Anderson, 1981). The exact depth of these discontinuities (hereafter referred to as 410 and 660), is a non-linear function of local temperature, pressure, and composition (Bina & Helffrich, 1994; G. Helffrich, 2000; Xu et al., 2008). The RF technique serves as an excellent tool to study lateral thermo-chemically induced variations in discontinuity structure (e.g. Kumar et al., 2005; Lawrence & Shearer, 2006; Sodoudi et al., 2006; Andrews & Deuss, 2008; Tauzin et al., 2008; Geissler et al., 2010; Vinnik et al., 2010; Cottaar & Deuss, 2016; Knapmeyer-Endrun et al., 2017; Tauzin et al., 2017; Makushkina et al., 2019; Monna et al., 2019; Munch et al., 2020).

Here we consider Ps and Sp RFs to study crust, lithosphere, and mantle underneath the contiguous USA, which, from a geological point of view, offers a plethora of past and recent processes, encompassing areas from active tectonic regions in the west to precambrian and phanerozoic continental masses in the central and eastern parts (cf. Figure 2A–

73 B, Hoffman, 1989; Williams et al., 1991; Whitmeyer & Karlstrom, 2007; Mooney & Ka-  
 74 ban, 2010). Seismic studies of the North American lithosphere, its base, and intra-lithospheric  
 75 discontinuities have been considered extensively and provided insights into the processes  
 76 that formed these complex continental entities (e.g., Rychert et al., 2007; H. Yuan & Ro-  
 77 manovicz, 2010; Abt et al., 2010; H. Yuan et al., 2014; S. M. Hansen et al., 2015; Kind  
 78 et al., 2015; Afonso et al., 2016; Calò et al., 2016; Eilon et al., 2018; Boyce et al., 2019;  
 79 Kind et al., 2020). Seismic imaging of the mantle have tracked both recent (e.g., Juan  
 80 de Fuca slab, Yellowstone plume) and ancient tectonic features (e.g., Farallon slab) (e.g.,  
 81 van der Lee & Nolet, 1997; Burdick et al., 2008; Nettles & Dziewonski, 2008; Obrebski  
 82 et al., 2010; Sigloch, 2011; Porritt et al., 2014; C. Chen et al., 2015; Maguire et al., 2018;  
 83 Nelson & Grand, 2018; Krischer et al., 2018).

84 In this study, we aim to improve our current understanding of mantle structure and  
 85 its relation to past and present dynamical processes underneath the North American con-  
 86 tinent by providing a more detailed view of the combined thermo-chemical structure of  
 87 Moho, LAB, upper mantle, and MTZ. For this purpose, we will build upon and extend  
 88 the work of Munch et al. (2018), who combined mineral phase equilibrium computations  
 89 with geophysical inversion to self-consistently image the thermo-chemical structure of  
 90 the MTZ using P-to-s waves converted at the 410 (P410s) and 660 (P660s). There are  
 91 several advantages to this approach in that it anchors temperature, composition, seis-  
 92 mic properties, and discontinuities that are in laboratory-based forward models, while  
 93 simultaneously permitting, through inversion, to optimize profiles of physical properties  
 94 (e.g., S-wave velocity) to match seismic data.

95 We consider a number of improvements and extensions to our previous work by 1)  
 96 including Sp RFs; 2) considering multiple frequency bands for both Ps and Sp RFs; and  
 97 3) applying the methodology to USArray data to perform a tomographic study of the  
 98 North American continent. Points 1) and 2) allow for significant improvement in reso-  
 99 lution of lithosphere, upper mantle, and transition zone structure. With these improve-  
 100 ments in place, we are able to identify three major tectonic provinces based directly on  
 101 differences in mantle thermo-chemical and seismic structure that show significant thermo-  
 102 chemically induced variations across the LAB and MTZ. We further observe well known  
 103 tectonic features in our tomographic images, such as hot mantle upwelling underneath  
 104 the western USA as well as traces of active and ancient subduction.

105 This paper is divided into two parts: a methodological and an application part. In  
 106 part 1 (sections 2–3), we describe seismic data processing to construct a new set of multi-  
 107 frequency Ps and Sp RFs using USArray data with particular emphasis on Sp RFs (Ps  
 108 RFs are detailed in Munch et al. (2018)), including data selection, deconvolution, spa-  
 109 tial sensitivity analysis, stacking and uncertainty estimation, and RF modeling as part  
 110 of which we perform a benchmark of one-dimensional (1-D) synthetic waveforms against  
 111 three dimensional (3-D) full waveform synthetics to validate our 1-D modeling approach  
 112 to computing seismograms. In part 2 (sections 4–7), we apply the developed method-  
 113 ology to obtain self-consistent tomographic images of the crust, lithosphere, and man-  
 114 tle underneath the North American continent. Part 2 includes model parametrization,  
 115 inversion procedure, and discussion of results. Finally, to further test the results presented  
 116 here, we compare our thermal model with an independent model of the western USA con-  
 117 structed from thermobarometry data.

## 118 2 Data processing

119 The fundamental idea of the receiver function method is to separate signals gen-  
 120 erated by the direct and converted phases on three-component ground-motion record-  
 121 ings based on their distinct particle motion (e.g., Rondenay, 2009; Kind et al., 2012). In  
 122 the case of a P-to-s conversion, the P-motion is recorded on the vertical component and  
 123 the converted wave on the radial component, while the order is reversed for S-to-p con-  
 124 versions. The converted signal can be considered as the convolution of the direct wave-  
 125 field with the local structural response. Thus, deconvolving the “source”- (i.e., direct phase)



126 from the “response”-trace (i.e., converted phase) results in a time series representing the  
 127 structure below the receiver. This time series is called the receiver function (cf. Figure 1B–  
 128 C; Vinnik, 1977; Langston, 1979). Source and path effects between event and conver-  
 129 sion point are assumed to be removed through the deconvolution operation, as all this  
 130 information is contained in both, source- and response-traces (Kind et al., 2012).

131 In spite of certain advantages of Sp over Ps RFs, the use of the former for mak-  
 132 ing structural inferences is limited relative to the latter because of limitations that be-  
 133 set S-waves (e.g., Bock, 1994; Farra & Vinnik, 2000; X. Yuan et al., 2006; Wilson et al.,  
 134 2006). Firstly, the incidence angle of S-to-p converted phases is larger than that of the  
 135 original S-wave, causing post-critical incidence to restrict the usage of Sp RFs to nar-  
 136 rower epicentral distance ranges and depths shallower than 660 km (cf. Figure 1A). Also,  
 137 because of the large incidence angle, S-to-p conversion points are located farther away  
 138 from the station than in the case of P-to-s, which ultimately results in broader regional  
 139 sampling. Added to this, is a lower spatial resolution because of a lower frequency con-  
 140 tent of S-waves relative to P-waves. Secondly, the signal-to-noise ratio (SNR) of S-waves  
 141 is lower than for P-waves because of a range of interfering phases, which generally com-  
 142 plicates observation of S-to-p converted phases. In this context, Bock (1994) and Wilson  
 143 et al. (2006) have pointed out that amplitudes of P-wave multiples between the surface  
 144 and the MTZ can become significant to the extent that these can be mistaken for con-  
 145 versions. Thirdly, core-phases (ScS, SKS, and their respective conversions) in the epi-  
 146 central distance range  $\sim 85^\circ$ – $100^\circ$  interfere with S-to-p conversions, in particular S660p  
 147 phases (X. Yuan et al., 2006). These issues can, however, to a large extent be mitigated  
 148 through careful data selection and processing steps as described in e.g., Wilson et al. (2006);  
 149 Monna et al. (2019); X. Shen et al. (2019).

## 150 2.1 Data selection

151 We collected and processed data recorded at stations of the USArray (Figure 2C),  
 152 which includes about 100 reference stations in complementary use to the denser trans-  
 153 portable network. Seismic data were acquired from the IRIS datacenter (Incorporated  
 154 Research Institutions for Seismology, [ds.iris.edu/ds/nodes/dmc/](https://ds.iris.edu/ds/nodes/dmc/)) using ObspyDMT  
 155 (Hosseini & Sigloch, 2017). Event selection is based on information available from the  
 156 global centroid moment tensor catalogue (Dziewonski et al., 1981; Ekström et al., 2012,  
 157 [www.globalcmt.org](http://www.globalcmt.org)). Only events with a magnitude of at least 5.5 and a focal depth  
 158 shallower than 150 km were selected. The latter criterion is applied to reduce the inter-  
 159 ference of S-to-p conversions with surface-reflected multiples (Wilson et al., 2006). For  
 160 constructing Ps RFs, we select events in the epicentral distance ( $\Delta$ ) range  $40^\circ$ – $95^\circ$  (e.g.,  
 161 Andrews & Deuss, 2008). For Sp RFs, we are restricted to a  $\Delta$  range of  $55^\circ$ – $90^\circ$  because  
 162 of post-critical incidence. Also, because of the simultaneous arrival of SKS and ScS (in-  
 163 cluding their conversions) beyond  $85^\circ$ , the S660p conversion is very difficult to observe  
 164 (X. Yuan et al., 2006). We therefore focus on S-to-p conversions at discontinuities be-  
 165 tween the Moho and the 410. The selected distance ranges and the respective travel times  
 166 of direct and converted waves are shown in supplementary Figure S.1.

## 167 2.2 Receiver function processing

168 For each event, recordings are filtered between 0.02 Hz and 1 Hz using a Butter-  
 169 worth bandpass filter of 2<sup>nd</sup> order and then rotated to the ZRT (vertical-radial-transverse)  
 170 system under consideration of the back-azimuth computed from the source and receiver  
 171 locations. In the Sp case, we additionally rotate waveforms to the LQT-system, where  
 172 the orthogonal L and Q components are approximately aligned with the S and S-to-p  
 173 converted phases, respectively, improving their detectability (Rondenay, 2009; Kind et  
 174 al., 2012). The orientation of the L- and Q-axes is determined by maximizing the en-  
 175 ergy in the window from -30 s to +50 s on the Z-/L-component around the theoretical  
 176 S-wave arrival by means of principal component analysis (PCA; see supplementary Fig-

177 ure S.2). In order to determine the SNR and exact arrival of the direct phase, we im-  
 178 plemented the automated picking approach of Abt et al. (2010). An illustration hereof  
 179 is shown in supplementary Figure S.3. To ensure high-quality data, we only consider wave-  
 180 forms with a  $\text{SNR} > 5$ . Alternative SNR criteria for detecting lithospheric S-to-p conver-  
 181 sions have also been proposed by X. Shen et al. (2019) and Monna et al. (2019).

### 182 **2.2.1 Deconvolution**

183 Receiver functions are obtained from deconvolving source (direct phase) from re-  
 184 sponse waveform (converted phase). The source wavelet is extracted by trimming the  
 185 corresponding trace around the picked arrival of the direct phase within windows from  
 186 -50 s to +150 s and from -100 s to +30 s for Ps and Sp RFs, respectively. Analogously,  
 187 the response trace is cut within a window from -50 s to +150 s and from -150 s to +50 s  
 188 for Ps and Sp RFs, respectively. The tapered source and response waveforms are sub-  
 189 sequently deconvolved using iterative time-domain deconvolution (Ligorria & Ammon,  
 190 1999). For removing high-frequency noise from the RFs, a Butterworth lowpass filter of  
 191 2<sup>nd</sup> order is applied for different corner periods: 5 s, 8 s, and 10 s for Ps RFs and 10 s  
 192 and 15 s for Sp RFs. This choice is dictated by the increased noise-content at shorter  
 193 periods and lack of resolution at longer periods, respectively. Maximum wavelengths as-  
 194 sociated with these periods are 28–56 km for Ps and 90–135 km for Sp RFs, respectively.  
 195 Since the arrival times of converted phases not only depend on the structure below the  
 196 receiver but also on the slowness of the ray, we stretch and compress the time axis of each  
 197 RF to a reference slowness (moveout correction), that is  $6.5 \text{ s}/^\circ$  for Ps RFs and  $9.9 \text{ s}/^\circ$   
 198 for Sp RFs. For comparison with Ps RFs, the time axis and polarity of Sp RFs are flipped.

### 199 **2.2.2 Stacking and uncertainty estimate**

200 To further enhance SNR, we stack individual RF traces at a given station. This  
 201 ensures that information from many events, where focal depth, mechanism, and epicen-  
 202 tral distance differ, are combined coherently. Phases not aligned through the moveout  
 203 correction are cancelled and hence direct conversions enhanced. To determine uncertain-  
 204 ties on RFs, we employ a bootstrap resampling approach (Efron & Tibshirani, 1991) and  
 205 compute the standard deviation of 100 stacks, where each stack is formed from  $n$  ran-  
 206 domly recombined RFs, with  $n$  being the number of available RF traces. A crucial choice  
 207 has to be made on the selection of RF traces to be included in a stack, as this depends  
 208 on the spatial sensitivity of the data set, i.e., location of conversion points relative to sta-  
 209 tion and frequency content.

### 210 **2.2.3 Spatial sensitivity of Sp RFs**

Assuming the main sensitivity of a RF to be confined to the region directly beneath  
 the station allows us to sum all moveout-corrected RF traces available to form a stack.  
 While this assumption is reasonable for P-to-s conversions and therefore applied to the  
 Ps RF data set, it is not directly applicable to the S-to-p conversions, because of the high  
 incidence angles compared to the direct S-wave. Consequently, S-to-p conversion points  
 are located significantly further away from the station than those of P-to-s converted waves  
 from the same depth (cf. Figure 1A). Moreover, S-to-p converted waves propagate faster  
 and hence have longer wavelengths. These effects are illustrated for Sp RFs obtained at  
 station 39-COWI (‘39’ referring to the station-ID in Figure 2C and ‘COWI’ being the  
 station-abbreviation) and filtered at 10 s period by computing and plotting the Fresnel  
 volumes of converted waves at different depth slices following the approaches of Lekic  
 et al. (2011) and Cottaar and Deuss (2016). For each event present in the Sp RF data  
 set for station 39-COWI, we 1) determine S-to-p conversion points at depth  $Z$  using ray-  
 tracing through IASP91 (Kennett & Engdahl, 1991; Crotwell et al., 1999), and 2) com-  
 pute the Fresnel zone half-width,  $\delta^{\text{HW}}$ , at the same depth  $Z$ , which is a function of wave-

length,  $\lambda$ , via:

$$\delta^{\text{HW}} = \sqrt{\left(\frac{\lambda}{3} + Z\right)^2 - Z^2} \quad (1)$$

where  $\lambda$  is obtained by multiplying the P-wave velocity above  $Z$ , i.e., the velocity of the converted wave, with the dominant period (here 10 s). Knowing the location of conversion points and Fresnel zones, we evaluate the lateral sensitivity of the ray by comparing  $\delta^{\text{HW}}$  for each node on a discrete grid at distance  $\delta$  to the conversion point of interest via normalized cubic splines:

$$w = \begin{cases} \frac{3}{4}\bar{\delta}^3 - \frac{3}{2}\bar{\delta}^2 + 1, & \text{if } \bar{\delta} \leq 1 \\ \frac{1}{4}(2 - \bar{\delta})^3, & \text{if } 1 < \bar{\delta} \leq 2 \\ 0, & \text{if } \bar{\delta} \geq 2 \end{cases} \quad (2)$$

211 where  $\bar{\delta} = \delta/\delta^{\text{HW}}$ . We performed this analysis for conversions at 210 km and 410 km  
 212 depth (see Figure 3A–B). The weights of all rays included in a certain back-azimuth bin  
 213 are summed and re-scaled by the maximum value to facilitate comparison. Note that the  
 214 width of the Fresnel volume is not the resolution limit of a seismic wave (Spetzler & Snieder,  
 215 2004), but serves as a useful illustration of the latter.

216 For station 39-COWI, we observe that the group of events emanating from the west-  
 217 ern Pacific, South America, and Europe (Figure 3E) produce distinct sensitivity clus-  
 218 ters at 210 km and 410 km depth (Figure 3A–B). Their clear separation indicates that  
 219 summing RFs from the entire data set would lead to averaging of lithospheric and up-  
 220 per mantle structure from different regions. Instead, it requires forming stacks that are  
 221 sorted by back-azimuth (e.g., Vinnik et al., 2005; Vinnik et al., 2010; Monna et al., 2019).  
 222 However, back-azimuth binned stacks reduce the number of useful traces available in one  
 223 back-azimuth bin to between 10 and 80; yet, Sp RF stacks containing less than  $\sim 100$   
 224 traces tend to be noisy and, consequently, less robust (see supplementary section S.2).  
 225 A possible work-around is to increase the number of traces within a back-azimuth bin  
 226 by incorporating additional data as shown in the following.

#### 227 **2.2.4 Back-azimuth stacking using additional stations**

228 In addition to the reference network (Figure 2C), USArray also consists of a dense  
 229 network of transportable stations. Since stations are only spaced about 60 km apart, events  
 230 of similar back-azimuth recorded at neighbouring stations will produce S-to-p converted  
 231 waves that sample a similar region as illustrated in Figure 3C–D. Thus, the amount of  
 232 data in a back-azimuth bin for a given reference station can be increased by including  
 233 data recorded at near-by transportable stations.

234 This is illustrated in Figure 4 for the reference station 39-COWI, which shows how  
 235 the uncertainty and amplitudes of the Sp RF stack evolve as traces from the transportable  
 236 stations closest to 39-COWI are added for events in the western Pacific (back-azimuth  
 237 bin #1, panels A and B) and Europe (back-azimuth bin #3, panels C and D), respec-  
 238 tively. For both back-azimuth clusters, we see that the bootstrap-derived mean uncer-  
 239 tainty of the cumulative stack decreases and that RF amplitudes change when the num-  
 240 ber of traces is increased. The improvement in the quality of the Sp RF is particularly  
 241 evident in the case of the S410p phase that emerges in back-azimuth bin #3, which is  
 242 far less apparent in the Sp RF 39-COWI-only stack. Although the final number of traces  
 243 used ( $\sim 50$ ) for this particular event cluster remains below the threshold associated with  
 244 a robust stack, back-azimuth bin #1 includes more than 150 traces. In combination with  
 245 the clear signals emanating from Moho, lithosphere, and the 410-km discontinuity, we  
 246 therefore consider this to be a high-quality stack and retain it for the inversion.

247 An alternative approach would be common-conversion point (CCP) stacking, which  
 248 migrates each individual RF trace from time- to depth-domain according to its conver-  
 249 sion points (e.g., Dueker & Sheehan, 1997; Kosarev et al., 1999; Rondenay, 2009; Lekić  
 250 et al., 2011; Lekić & Fischer, 2014; Cottaar & Deuss, 2016; Lekić & Fischer, 2017; Kind

et al., 2020). This method, however, depends crucially on the background velocity model used for migration and, unless data are properly re-processed during inversion using the currently updated velocity model, is not self-consistent and therefore limited in accuracy.

### 2.3 Data workflow summary

We processed Ps and Sp RFs for all USArray reference stations (Figure 2C). This included the following steps:

- Construction of a multi-frequency Ps (5, 8, and 10 s low-pass period) and Sp (10 and 15 s low-pass period) RF data set.
- Formation of Sp RF stacks sorted by back-azimuth by combining data recorded at USArray reference and transportable stations that are located at a maximum distance of 140 km from the former.
- Selection of robust and only highest quality Ps and Sp RFs with clear signals emanating from the Moho, lithosphere, 410, and 660 for the inversion (cf. Table S.1 in the supplementary material).

## 3 Synthetic receiver function modeling

In order to generate a synthetic RF stack equivalent of the observed RF stack, we first compute waveforms using the reflectivity method (Fuchs & Müller, 1971; Müller, 1985). In order to ensure consistency, synthetic waveforms are processed the same way as the observed data (Munch et al., 2018). For optimal radiation of P and SV waves, we model the source as an explosion for Ps RFs and a normal-fault mechanism for Sp RFs. The source-time function is a Heaviside function and the modeled focal depth is imposed by the median of the depth distribution of the observed events. As the amplitudes of converted waves depend on slowness, their relative contribution to the final RF stack has to be mimicked in the synthetic RF modeling. The distribution of receivers and their relative weights in the stack therefore follows that of the distribution of the theoretical slownesses of the direct phase as imposed by the data set (Munch et al., 2018).

We exclude the uppermost layer of the model from the reflecting zone when computing waveforms for Sp RF (cf. supplementary Figure S.6). This suppresses multiple surface reflections which otherwise would constructively interfere and bias the RF. This approach is similar to the one by X. Yuan et al. (2006), who argue that propagation of multiples in a laterally heterogeneous Earth is ineffective. Ultimately, synthetic waveforms also have to conform to the following criteria: 1) the L- and Q-axis as imposed by the data set are employed to ensure consistency between observed and synthetic LQT-systems, since these affect RF amplitudes, and 2) the arrival time of the direct phase for the underlying model is determined via ray tracing (Crotwell et al., 1999), since the automated picking algorithm is unable to distinguish core phases from S-waves for  $\Delta > 85^\circ$ .

Finally, in order to validate our one dimensional (1-D) modeling approach to computing synthetic seismograms, we benchmarked the modeled receiver functions against three dimensional (3-D) full waveform synthetics (see supplementary section S.3.2). For this purpose, we employed the spectral element based software package Salvus (Afanasiev et al., 2019) to compute full waveforms in a 3-D velocity model. Comparison of 1-D and 3-D receiver functions is shown in supplementary Figure S.7 and the modeled waveforms are seen to be in good agreement. On the basis of this benchmark, the 1-D approach pursued here appears to be reliable for modeling both, Ps and Sp RFs, and, from a practical viewpoint, computationally much more efficient (1-D:  $\sim 15$ – $20$  sec on a normal desktop-computer, 3-D:  $\sim 25$  nodehours on a supercomputer).

## 4 Model parametrization

### 4.1 Petrological model

Building on previous experience (e.g., Khan et al., 2009; Munch et al., 2018), we model mantle composition by the weight fraction of basalt in a basalt-harzburgite mixture. This compositional model is based on the recognition that basaltic crust is generated at ridges through partial melting, leaving behind the depleted complement harzburgite (Xu et al., 2008). At subduction zones, this physically and chemically stratified oceanic lithosphere is continuously being recycled back into the mantle, where it remixes (e.g., G. R. Helffrich & Wood, 2001; Tackley et al., 2005). The resultant mineralogy depends on the degree of equilibration between basalt and harzburgite and two mixing models have been proposed (Xu et al., 2008): equilibrium assemblage (EA) and mechanical mixture (MM). The EA model assumes full equilibration, whereas in the MM model the mantle is a mixture of the end-members. Here, we focus on the EA model as Munch et al. (2018) have shown that RFs are less sensitive to differences between MM and EA.

For a given basalt fraction  $f$ , mantle composition  $\mathbf{X}$  is computed from

$$\mathbf{X} = f\mathbf{X}_B + (1 - f)\mathbf{X}_H, \quad (3)$$

where  $\mathbf{X}_B$  and  $\mathbf{X}_H$  are basaltic and harzburgitic end-member compositions, respectively. To predict the relative amount of stable minerals and their physical properties along a self-consistent mantle adiabat for harzburgite and mid-ocean ridge basalt (MORB) bulk chemical compositions in the  $\text{Na}_2\text{O}-\text{CaO}-\text{FeO}-\text{MgO}-\text{Al}_2\text{O}_3-\text{SiO}_2$  system (Table 1), we employ Gibbs free energy minimization and equation-of-state modeling (Perple\_X; Connolly, 2009). For this purpose, we adopt the thermodynamic formulation of Stixrude and Lithgow-Bertelloni (2005) and parameters and uncertainties as given by Stixrude and Lithgow-Bertelloni (2011). Bulk rock properties are subsequently obtained by means of Voigt-Reuss-Hill averaging (Hill, 1952; Watt et al., 1976). Density and elastic moduli are accurate to within  $\sim 0.5\%$  and  $\sim 1-2\%$ , respectively (Connolly & Khan, 2016). To obtain the mantle pressure profile, we integrate the vertical load from the surface pressure boundary condition. Within the crust and lithosphere, temperature is computed by assuming a conductive geothermal gradient, while the sublithospheric mantle adiabat is defined by the entropy of the lithology at the base of the lithosphere (see supplementary Figure S.8 for an illustration).

The present thermodynamic model of Stixrude and Lithgow-Bertelloni (2005, 2011) has limitations: 1) the equilibrium assumption is dubious at low temperatures as equilibrium is not likely to be achieved (e.g., Wood & Holloway, 1984). Consequently, we compute mineralogy for temperature below  $500^\circ\text{C}$  at  $500^\circ\text{C}$ , and the resultant mineralogy is then used to calculate physical properties at the temperature of interest; and 2) the thermodynamic model does not account for redox effects (e.g., Cline II et al., 2018), minor phases, and water and melt due to lack of thermodynamic data. Both, water and melt, can significantly affect elastic properties at lithospheric and MTZ depth. Negative gradients in S-wave velocity ( $> 5\%$ ) at mid-lithospheric depth in continental regions could arise from hydrous minerals (D. H. Green et al., 2010; Rader et al., 2015; Selway et al., 2015) and low wave velocities at the lithosphere-asthenosphere boundary in oceanic and active regions have been attributed to partial melt (e.g., Kawakatsu et al., 2009; Rychert & Shearer, 2009; Lekić & Fischer, 2014; Stern et al., 2015; Clark & Leshner, 2017; Karato & Park, 2019). The MTZ is considered as a plausible water reservoir within the mantle given the experimentally constrained storage capacities (1-3%) of wadsleyite and ringwoodite (Smyth, 1987; Kohlstedt et al., 1996; Férot & Bolfan-Casanova, 2012). However, mineral physics data suggest the effect of water on seismic velocities is negligible at MTZ conditions as a result of which seismic data are insensitive to water content (Thio et al., 2015; Schulze et al., 2018). Albeit, there is geophysical evidence for regional low-velocity layers atop the 410 (e.g., Song et al., 2004; Toffelmier & Tyburczy, 2007; Vinnik & Farra, 2007; Tauzin et al., 2010; Khan & Shankland, 2012; Kind et al., 2015; Xiao

**Table 1.** Model mantle compositions in wt% of mid-ocean ridge basalt (MORB) and its depleted residue, harzburgite (from Khan et al. (2009)). Mg# (molar) is given by  $\text{Mg}/(\text{Mg} + \text{Fe})$ .

Component	MORB, $\mathbf{X}_B$	Harzburgite, $\mathbf{X}_H$
CaO	13.05	0.5
FeO	7.68	7.83
MgO	10.49	46.36
Al <sub>2</sub> O <sub>3</sub>	16.08	0.65
SiO <sub>2</sub>	50.39	43.64
Na <sub>2</sub> O	1.87	0.01
Mg#	0.709	0.913

348 et al., 2020), which might be due to partial melt (Bercovici & Karato, 2003; Hirschmann,  
 349 2006; Freitas et al., 2017). Future studies will consider this in more detail.

## 350 4.2 Crustal structure

351 The velocity structure of the crust is important for modeling receiver functions (see  
 352 supplementary sections S.5.1 and S.5.3.2). Here, we follow Munch et al. (2018) and model  
 353 the crust as a stack of three layers of variable thickness and S-wave velocity, which is as-  
 354 sumed to increase with depth. The ratios of density and P-wave velocity to S-wave ve-  
 355 locity are constant in the crust. We invert for all of these parameters by using conver-  
 356 sions at the Moho and P-to-s crustal reverberations.

## 357 4.3 Anelastic effects

358 In order to model the effect of anelasticity on seismic amplitudes (Liu, 2003), we  
 359 follow the scheme outlined in Munch et al. (2018), which considers a laboratory-based  
 360 viscoelastic dissipation model derived from grain-size-, temperature-, and pressure-sensitive  
 361 viscoelastic relaxation measurements. The dissipation model based on the extended Burg-  
 362 ers model is described in detail in Jackson and Faul (2010) and relies on laboratory ex-  
 363 periments in the temperature range 800–1200 °C of torsional forced oscillation data within  
 364 1–1000 s period on melt-free poly-crystalline olivine grains between 3–165  $\mu\text{m}$  size. Ra-  
 365 dial profiles of shear attenuation,  $Q_\mu$ , are determined for the viscoelastic extended Burg-  
 366 ers model using the implementation of Bagheri et al. (2019), for which rheological pa-  
 367 rameters are listed in supplementary Table S.2.

368 The advantage of this particular set-up is that it allows us to account for, through  
 369 the thermodynamic interface, the influence of temperature and pressure on the shear at-  
 370 tenuation model in a self-consistent manner (e.g., Bagheri et al., 2019). Alternative vis-  
 371 coelastic models have also been proposed (e.g., Jackson & Faul, 2010; Sundberg & Cooper,  
 372 2010; Karato et al., 2015), but the relatively high-frequency Ps and Sp RFs are not sen-  
 373 sitive enough to allow for proper discrimination (see discussion in supplementary sec-  
 374 tion S.6.3). In case  $Q_\mu$  exceeds 600,  $Q_\mu$  is set equal to 600 and bulk attenuation,  $Q_\kappa$ , is  
 375 held fixed at 57823 after PREM (Dziewonski & Anderson, 1981).

## 376 5 Inverse problem

### 377 5.1 Formulation of the inverse problem

Within the Bayesian framework, the solution to the inverse problem  $\mathbf{d} = \mathbf{g}(\mathbf{m})$ ,  
 where  $\mathbf{d}$  is a data vector containing observations and  $\mathbf{g}$  a typically non-linear operator



**Table 2.** Overview of model parameters, their quantity, and prior model parameter ranges and probability distributions

Description	Parameter	Quantity	Value/range	Distribution
Basalt fraction	$f$	1	0–0.5	uniform
Lithospheric temperature (°C)	$T_{\text{lit}}$	1	950–1600	uniform
Lithospheric thickness (km)	$Z_{\text{lit}}$	1	70–300	uniform
Crustal layer thicknesses (km)	$\delta Z$	3	1–20	uniform
S-wave velocity of uppermost crustal layer (km/s)	$V_S^1$	1	2.0–4.1	uniform
S-wave velocity change across crustal discontinuities (km/s)	$\delta V_S$	2	0–0.5	uniform
Density- and P-to-S-wave velocity crustal ratios	$\rho/V_S$ $V_P/V_S$	1 1	$0.768 \pm 0.0213$ $1.746 \pm 0.0195$	normal normal

that maps a model parameter vector  $\mathbf{m}$  into data, is given by (Mosegaard & Tarantola, 1995)

$$\sigma(\mathbf{d}, \mathbf{m}) = k \cdot h(\mathbf{m})\mathcal{L}(\mathbf{d}, \mathbf{m}), \quad (4)$$

where  $k$  is a normalization constant,  $h(\mathbf{m})$  is the prior probability distribution on model parameters (see Table 2) and contains information about model parameters obtained independently of the data under consideration,  $\mathcal{L}(\mathbf{d}, \mathbf{m})$  is the likelihood function, which can be interpreted as a measure of misfit between the observations and the predictions from model, and  $\sigma(\mathbf{d}, \mathbf{m})$  is the posterior model parameter distribution containing the solution to the inverse problem. The particular form of  $\mathcal{L}(\mathbf{d}, \mathbf{m})$  is determined by the observations, their uncertainties, and model data noise and will be discussed in the following section.

## 5.2 Definition of the likelihood function

We only consider the time windows that include Ps and Sp RF signal of interest, i.e., converted energy from the Moho (and reverberations thereof), the 410, and the 660 for Ps RFs, and the Moho, the lithosphere, and the 410 for Sp RFs. The windows are determined manually for each stack as illustrated in Figure 5 (gray areas). For each window  $i$ , observed RF, and low-pass period  $T$ , we consider a  $L_2$ -based likelihood function of the form

$$\mathcal{L}_{T,i}^{\text{RF}}(\mathbf{d}, \mathbf{m}) = \exp(-\Phi_{T,i}^{\text{RF}}), \quad (5)$$

where

$$\Phi_{T,i}^{\text{RF}} = \frac{1}{N} \sum_j \frac{\|\mathbf{d}_j^{\text{obs}} - \mathbf{d}_j^{\text{syn}}\|^2}{2\sigma_j^2} \quad (6)$$

with  $\mathbf{d}_j^{\text{obs}}$ ,  $\mathbf{d}_j^{\text{syn}}$ , and  $\sigma_j$  being synthetic and observed Ps or Sp RF amplitudes (for a given period and window) and data uncertainty of data point  $j$ , respectively, with  $N$  denoting the total number of points within each misfit window. Consequently, the total likelihood function for a given station, RF-type, and back-azimuth bin is the conjunction of the likelihood functions associated with each period and misfit window:

$$\mathcal{L}(\mathbf{d}, \mathbf{m}) = \prod_{T,i} \mathcal{L}_{T,i}^{\text{RF}}(\mathbf{d}, \mathbf{m}) \quad (7)$$

## 5.3 Sampling of the posterior

We sample the posterior distribution by means of the Metropolis-Hastings algorithm (Metropolis et al., 1953; Hastings, 1970). This algorithm is based on random sam-



pling of the model space, yet only models that result in a good data fit and are consistent with prior information are frequently sampled (importance sampling). At each step  $n$ , a model,  $\mathbf{m}^{n+1}$ , is proposed on the basis of the prior distribution and is either accepted or rejected based on the probability

$$P = \min \left[ 1, \frac{\mathcal{L}(\mathbf{m}^{n+1})}{\mathcal{L}(\mathbf{m}^n)} \right], \quad (8)$$

where  $\mathbf{m}^n$  represents the last accepted model. In a slight modification, we use the cascaded Metropolis algorithm (Mosegaard & Tarantola, 1995, 2002), which applies the acceptance criterion (Eq. 8) to each period  $T$  and misfit window  $i$  (crust, lithosphere, 410, and 660) separately for a given station and individual RF-type (and back-azimuth bin):

```

388 where  $\mathbf{m}^n$  represents the last accepted model. In a slight modification, we use the cas-
389 caded Metropolis algorithm (Mosegaard & Tarantola, 1995, 2002), which applies the ac-
390 ceptance criterion (Eq. 8) to each period  $T$  and misfit window  $i$  (crust, lithosphere, 410,
391 and 660) separately for a given station and individual RF-type (and back-azimuth bin):
392
393   (*) propose move:  $\mathbf{m}_{\text{RF}}^n \rightarrow \mathbf{m}_{\text{RF}}^{n+1}$  RF=Ps or RF=Sp
394   for  $T = 5, 8, 10$  s (if RF=Ps) /  $10, 15$  s (if RF=Sp)
395     for  $i = 1, 2, 3$ 
396       if  $\mathcal{L}_{T,i}^{\text{RF}}(\mathbf{m}_{\text{RF}}^{n+1}) \geq \mathcal{L}_{T,i}^{\text{RF}}(\mathbf{m}_{\text{RF}}^n)$  then
397         accept this step
398       else if  $\mathcal{L}_{T,i}^{\text{RF}}(\mathbf{m}_{\text{RF}}^{n+1}) < \mathcal{L}_{T,i}^{\text{RF}}(\mathbf{m}_{\text{RF}}^n)$  then decide randomly to move to
399         the next step or to reject the proposed move with probability
400          $P = \mathcal{L}_{T,i}^{\text{RF}}(\mathbf{m}_{\text{RF}}^{n+1}) / \mathcal{L}_{T,i}^{\text{RF}}(\mathbf{m}_{\text{RF}}^n)$ 
401       else reject  $\mathbf{m}_{\text{RF}}^{n+1}$  and return to (*)
402     end
403   end

```

Thus, the model is only retained if all of the individual steps are accepted. This acceptance criterion is more strict when compared to the “standard” acceptance criterion of the Metropolis-Hastings algorithm, but acts to prevent trade-offs between misfit windows as these are considered separately.

Finally, to improve MCMC performance, we reduced the burn-in of the sampling stage by employing a global optimisation method in order to obtain a good initial model for the subsequent model space search. For this purpose, we use the Covariance Matrix Adaption Evolution Strategy of Hansen and Ostermeier (2001). We first search for a crustal model by minimizing the misfit in the crustal time window, and subsequently refine both crustal and thermo-chemical model by minimizing misfit across all windows. In the MCMC stage, we sampled  $\sim 40000$  models in total by running 8 chains in parallel, of which  $\sim 4000$  models are retained for analysis here. Step lengths are adjusted so that the acceptance rate is between 30–50%. Synthetic test inversions and grid-search examples are described in the supplementary section S.5.

#### 5.4 A posteriori migration of inversion results

We created a series of maps and cross sections to visualize the spatial variation of model parameters based on mean models. The use of mean models is justified given that the posterior probability density distributions (Figure 6) are mostly Gaussian. Maps and cross sections are produced through linear interpolation of the Ps and Sp RF-derived mean models, which were migrated to their point of sensitivity, which, in the case of Ps RFs, is approximately the location of the station and for Sp RFs is the conversion point of the  $Sdp$  phase (cf. section 2.2.3), where  $d$  is the depth of sensitivity of a particular model parameter. For example, when migrating Sp RF-derived crustal and lithospheric thickness,  $d$  equals either  $Z_c$  or  $Z_{lit}$ , whereas in the case of temperature,  $d$  is equal to 410 km, since thermal information is primarily contained in the S410p phase (see supplementary Figure S.15).

For a given station, back-azimuth bin, and model parameter with sensitivity  $d$ , we compute the coordinates of the  $Sdp$  conversion points in IASP91 for all events included in that bin and average them to extract the final point of sensitivity. The so-computed Sp RF points are shown as diamonds in the maps that follow, while Ps RF points, i.e., station locations, are depicted as circles. All estimates related to discontinuity structure

434 (Moho, LAB, and MTZ) have been corrected for station elevation. Finally, while the present  
 435 inversion is not a tomographic inversion *sensu stricto*, it is an inversion for a set of (quasi-  
 436 local) radial profiles of thermochemical and physical structure, which, when pieced to-  
 437 gether, result in tomographic images that are based on the a posteriori migration pro-  
 438 cedure.

## 439 6 Results and Discussion

### 440 6.1 Datafit

441 The fits to observed Ps and Sp RFs at 5 and 10 s period, respectively, are shown  
 442 in Figure 5. Fits to Ps and Sp RFs at the other periods are summarised in supplemen-  
 443 tary Figure S.19. Conversions at the Moho, 410, and 660 discontinuities as well as litho-  
 444 spheric Sp RFs are seen to be captured well by the sampled models.

### 445 6.2 Crustal thickness

446 The inverted mean crustal thickness model is shown in Figure 7A. Results for  $\rho/V_S$   
 447 and  $V_P/V_S$  can be found in the supplementary material (Figure S.20). The abbrevia-  
 448 tions and extent of physiographic and tectonic provinces referred to hereafter are depicted  
 449 in Figure 2A–B. As expected, crustal thickness estimates are better constrained from Ps  
 450 than from Sp RFs (Figure 6A–B) given the higher frequency content of Ps RFs and the  
 451 additional information contained in crustal reverberations. Mean Ps and Sp RF derived  
 452 standard deviations are 1.6 km and 3.5 km, respectively. For comparison, we are also  
 453 showing the crustal thickness map based on Crust1.0 (Laske et al., 2013) in Figure 7B.  
 454 Our model is in good agreement with Crust1.0, including other models (not shown; e.g.,  
 455 Lowry & Pérez-Gussinyé, 2011; Levander & Miller, 2012; Schmandt et al., 2015; W. Shen  
 456 & Ritzwoller, 2016; Buehler & Shearer, 2017; Ma & Lowry, 2017) which indicate that  
 457 regions with relatively thin crust (25–35 km) are found in the western part of the USA,  
 458 i.e., beneath the Columbia plateau (physiographic province P2) and the Basin and Range  
 459 province (P3), as well as the passive margins of the continent in the southern and east-  
 460 ern parts (P10). Regions with thicker crust (45–55 km) are observed underneath the Wyoming  
 461 craton (tectonic province T2) and Trans-Hudson orogen (T3) as well as within the pre-  
 462 cambrian building blocks towards the east (P6–P8; T4–T8) and south-west that extend  
 463 into the Colorado plateau (P4).

### 464 6.3 Tectonic regionalization

465 In order to embed the results in a geological context, we first extracted radial pro-  
 466 files of temperature, S-wave velocity, and shear attenuation on a  $2.5^\circ \times 2.5^\circ$ -grid and  
 467 grouped them individually via k-means clustering according to their similarity, i.e., by  
 468 minimizing the  $L_2$ -distance to the mean profile of the respective cluster (MacQueen, 1967).  
 469 This approach allows for identification of tectonic regions without any a priori constraints  
 470 (e.g., Houser et al., 2008; Lekić & Romanowicz, 2011; Lekić et al., 2012; Calò et al., 2016).  
 471 Consistent results were found for three clusters that are shown in Figure 8. The radial  
 472 profiles for each parameter show specific characteristics that determine the clustering:  
 473 1) temperature and depth of the intersection point of conductive geotherm and mantle  
 474 adiabat; 2) amplitude and depth extent of upper mantle low-velocity zone; and 3) thick-  
 475 ness of high- $Q_\mu$  lid.

476 When the mean profile for each location is projected onto a map (see insets in Fig-  
 477 ure 8), a well-known trend that follows the tectonic division of the North American con-  
 478 tinent appears. In particular, we are able to distinguish the tectonically active west (en-  
 479 compassing physiographic provinces P1–P5), which is characterised by being relatively  
 480 hot, of slow shear-wave velocity, and highly dissipative (red dots), from the colder, faster,  
 481 and less dissipative older eastern parts (blue dots) that include the cratonic core and the

**Table 3.** Summary of thermo-chemical characteristics and discontinuity structure of the main tectonic regions identified through cluster analysis in Figure 8A. Parameters are: crustal thickness ( $Z_c$ ), potential temperature ( $T_{\text{pot}}$ ), lithosphere thickness ( $Z_{\text{lit}}$ ), basalt fraction ( $f$ ), Mg# (molar Mg/(Mg+Fe)), and depth to 410 ( $Z_{410}$ ) and 660 ( $Z_{660}$ ). Values represent the mean and standard deviation based on the distributions shown in Figures 9C,E and 12B,D.

Parameter	Tectonic region		
	TAW (tectonically active West)	CTR (central transition regions)	COE (cratonic-orogenic East)
$Z_c$	$40 \pm 5$ km	$44 \pm 5$ km	$44 \pm 4$ km
$T_{\text{pot}}$	$1490 \pm 27$ °C	$1429 \pm 37$ °C	$1413 \pm 27$ °C
$Z_{\text{lit}}$	$100 \pm 23$ km	$146 \pm 31$ km	$208 \pm 33$ km
$f$	$13 \pm 4.5\%$	$11 \pm 3.0\%$	$10 \pm 3.6\%$
Mg#	$0.904 \pm 0.003$	$0.906 \pm 0.002$	$0.907 \pm 0.003$
$Z_{410}$	$424 \pm 7$ km	$417 \pm 6$ km	$416 \pm 6$ km
$Z_{660}$	$660 \pm 2$ km	$659 \pm 2$ km	$658 \pm 2$ km

482 Appalachian mountains, i.e., parts of provinces P6–P9 and T2–T8. The third cluster (green  
 483 dots) represents an intermediate group of profiles distributed over the central and east-  
 484 ern parts that are sandwiched between red and blue regions. While these features sug-  
 485 gest that tectonic regions are mainly controlled by the thermal structure of the conti-  
 486 nental lithosphere, a compositional signal is also present (section 6.6). For the remain-  
 487 der of the discussion, we will refer to the tectonic regions identified in Figure 8A as the  
 488 tectonically active West (TAW; red), the central transition regions (CTR; green), and  
 489 the cratonic-orogenic East (COE; blue). Smoothed outlines of these regions are indicated  
 490 in the maps and an overview of the values observed for the thermo-chemical parameters  
 491 is given in Table 3.

#### 492 6.4 Mantle temperature

493 For ease of comparison with literature estimates, we rely on the concept of poten-  
 494 tial temperature,  $T_{\text{pot}}$ , which is the temperature of the adiabatic mantle when extended  
 495 to the Earth’s surface (McKenzie & Bickle, 1988). We compute  $T_{\text{pot}}$  by linear extrap-  
 496 olation of the adiabatic mantle geotherm:  $T_{\text{pot}} = T_{\text{lit}} - \nabla T \cdot Z_{\text{lit}}$ , where we have em-  
 497 ployed an adiabatic gradient  $\nabla T = (T_{410} - T_{\text{lit}})/(Z_{410} - Z_{\text{lit}})$  with  $T_{410}$  and  $Z_{410}$  rep-  
 498 resenting temperature at and depth of the 410, respectively. Uncertainties on  $T_{\text{pot}}$  av-  
 499 erage 35 °C and 43 °C for Ps and Sp RFs, respectively (Figure 6C–D). Figure 9A dis-  
 500 plays the interpolated values of the mean value of  $T_{\text{pot}}$ . As in Figure 8A, we observe a  
 501 clear east-west trend from low to high potential temperatures, reaching values of 1430–  
 502 1520 °C in region TAW, 1360–1470 °C in CTR, and 1350–1450 °C in COE, respectively.  
 503 We find large positive thermal anomalies underneath the Cascade Range (P1), the Basin  
 504 and Range province (P3), and Yellowstone Hotspot. The same general east-west pattern  
 505 was also observed in the thermal anomaly maps obtained by Khan et al. (2011) from in-  
 506 version of surface wave phase velocities, although of lower resolution, by Maguire et al.  
 507 (2018) from stacking of Ps RFs, and S. M. Hansen et al. (2015) from analysis of surface  
 508 wave tomographic models. We should note that thermal anomalies constrained by Sp  
 509 RFs could be biased by the basalt-rich nature of the inverted compositions (see Figure 6H).

510 We estimate that in these regions “normal” basalt fractions would cool the mantle lo-  
 511 cally by up to  $\sim 30\text{--}50$  °C (cf. supplementary Figure S.16).

## 512 6.5 Lithospheric thickness and structure

513 A map of the thickness of the thermal lithosphere is shown in Figure 9B. We em-  
 514 phasize that lithospheric thickness refers to the depth (defined by  $Z_{\text{lit}}$ ) at which the con-  
 515 ductive geotherm intercepts the mantle adiabat. There is no apparent discontinuity as-  
 516 sociated herewith and hence no converted phase; yet, the long-period part of the Sp RFs  
 517 (cf. supplementary Figure S.15) is affected. Note that uncertainties on thermal lithospheric  
 518 thickness are generally much smaller for Sp (17 km) than for Ps RFs (32 km) (Figure 6E–  
 519 F). This arises because lithospheric Sp RFs are not obscured by crustal reverberations,  
 520 as is the case for Ps RFs. Consequently, Ps RFs only provide indirect insight into litho-  
 521 spheric structure via the sensitivity of the P410s and P660s to mantle temperature, which  
 522 is controlled by both  $Z_{\text{lit}}$  and  $T_{\text{lit}}$ . However, as S-to-p measurements are sparse, we in-  
 523 clude the results from Ps RFs for interpolating  $Z_{\text{lit}}$ .

524 The thinnest lithosphere (79–141 km) is found in region TAW, as also observed in  
 525 previous studies that considered Ps and/or Sp RFs (e.g., Li et al., 2007; Abt et al., 2010;  
 526 Lekić et al., 2011; Levander & Miller, 2012; Hopper et al., 2014; S. M. Hansen et al., 2015).  
 527 The exact shape of the thin-lithosphere branches that extend across the Rocky Moun-  
 528 tains (P5) towards the Central plains (P6) are only constrained by few data points and  
 529 hence are not interpreted further here. Region CTR shows an intermediate range in litho-  
 530 spheric thickness of 85–187 km and coincides mostly with the western Great plains (P6),  
 531 the New Madrid Seismic Zone (e.g., Zhang et al., 2009; C. Chen et al., 2014), and the  
 532 mid-continental rift system (e.g., Stein et al., 2015). Finally, region COE is, as expected  
 533 on account of its cratonic and orogenic nature, composed of relatively thick lithosphere  
 534 (125–247 km). This negative correlation between potential temperature and lithospheric  
 535 thickness observed in the maps is clearly visible in Figure 9C and is further illustrated  
 536 in the tomographic cross sections shown in Figure 11, where lateral variations in thick-  
 537 ness go hand-in-hand with variations in S-wave anomalies (see section 6.7) such that thick  
 538 lithosphere is predominant in “cold” regions, whereas thin lithosphere is found in “hot-  
 539 ter” regions.

540 We should note that some Sp RFs (e.g., 17-JCT, 32-SUSD, and 39-COWI) show  
 541 complexities in the lithospheric part that are difficult to fit, which could potentially lead  
 542 to a bias in thermal thickness estimates. These complexities have been attributed to mid-  
 543 lithospheric discontinuities (MLDs) that were observed in Sp RFs from North America  
 544 (e.g., Abt et al., 2010; Miller & Eaton, 2010; H. Yuan & Romanovicz, 2010; Lekić & Fis-  
 545 cher, 2014; Kind et al., 2015; Calò et al., 2016) and elsewhere (e.g., Heit et al., 2007; Sav-  
 546 age & Silver, 2008; L. Chen, 2009; Ford et al., 2010; Wölbern et al., 2012; Sodoudi et al.,  
 547 2013). Processes capable of creating MLDs have been discussed extensively by Selway  
 548 et al. (2015). In a recent study, Kind et al. (2020) has pointed out that MLDs could re-  
 549 sult from the use of “inexact” filtering and deconvolution. Instead Kind et al. (2020) di-  
 550 rectly summed waveforms to obtain RF stacks as a result of which MLDs, for example,  
 551 were less apparent and point to the importance of careful data processing. While unob-  
 552 jectionable, this is of less importance in this study, inasmuch as the problems that be-  
 553 set imaging and interpretation are avoided in an inversion, where the same processing  
 554 steps are applied to synthetic and observed data.

## 555 6.6 Mantle composition

556 Mantle composition (basalt fraction) primarily affects the sharpness of the 410 and  
 557 660 seismic discontinuities through its influence on RF amplitudes rather than the tim-  
 558 ing of converted waves (see supplementary Figure S.15). This is opposite to the effect  
 559 of temperature, which primarily affects the depth of the 410 and 660 and, through that,  
 560 travel time of the converted phases. This separation of sensitivity allows us to decou-

561 ple the parameters. Basalt fractions estimated from Ps RFs vary on average from 0 to  
 562 25%, i.e., within the Harzburgite to Pyrolite range with uncertainties around 5% (cf. Fig-  
 563 ure 6G). In comparison, Sp RF-estimated basalt fractions show larger uncertainties (9.5%;  
 564 cf. Figure 6H) and appear to cluster toward the upper end of the prior ( $f = 50\%$ ), in  
 565 disagreement with the Ps RF-derived results and other studies (e.g., Khan et al., 2011;  
 566 Maguire et al., 2018). The inferred basalt-enrichment is likely related to the reduced sen-  
 567 sitivity of the S410p to composition, because of the generally small amplitudes of S410p  
 568 relative to P410s. However, as noted earlier, this only has a small effect on thermal es-  
 569 timates due to the first-order decoupling of temperature and composition. Consequently,  
 570 we exclude Sp RF-derived information from the compositional map (Figure 9D).

571 For present purposes, we express composition by means of the Mg#, which is com-  
 572 puted as  $\text{Mg}/(\text{Mg}+\text{Fe})$  (molar) (cf. Table 1), where high/low Mg# is indicative of de-  
 573pleted/fertile mantle, i.e., low/high basalt fraction, respectively. While most of the man-  
 574 tle across the continent appears to be made up of relatively depleted compositions (Fig-  
 575 ure 9D), region TAW is found to be slightly more fertile ( $f = 13 \pm 4.5\%$ ,  $\text{Mg\#} \sim 0.90$ )  
 576 in comparison to regions CTR and COE (COE:  $f = 10 \pm 3.6\%$ ,  $\text{Mg\#} \sim 0.91$ ), similar  
 577 to the results of Khan et al. (2011). Local enrichment of basalt underneath the conti-  
 578 nent could correspond to traces of active (Cascadia) and ancient subduction zones (Far-  
 579 allon). The location of these anomalies are in agreement with the results by Maguire et  
 580 al. (2018).

## 581 6.7 Shear-wave velocity structure

582 In the following, we focus on models of isotropic elastic S-wave velocity. Figure 10  
 583 shows S-wave velocity images of the upper mantle and MTZ. For comparison, analogous  
 584 plots for two tomographic models, US-SL-2014 (Schmandt & Lin, 2014) and SL2013NA  
 585 (Schaeffer & Lebedev, 2014), are shown in the supplementary material (Figures S.22–  
 586 S.25). To illustrate lateral variations in structure further, we also present cross sections  
 587 (Figure 11) along three east-west and north-south profiles (magenta lines in Figure 2C).  
 588 From these tomographic images, the following observations can be made:

589 1) S-wave velocity anomalies in the upper mantle (100–300 km) closely follow the  
 590 tectonic regionalization and appear to be dominated by the lithospheric signature in that  
 591 velocities are high/low in regions with a deep/shallow lithosphere, respectively. These  
 592 observations are particularly apparent in the east-west trending cross sections, where the  
 593 lithosphere slowly thickens and S-wave velocities decrease when passing from region TAW  
 594 to CTR underneath the eastern Rocky Mountains flank (P5 to P6). The large variations  
 595 in lithosphere thickness (70–250 km) affect the extent and amplitude of the upper man-  
 596 tle low-velocity zone (Figure 8B), including large lateral velocity variations of  $\pm 5.1\%$  at  
 597 100 km depth. Amplitudes aside, similar observations were made in models SL2013NA  
 598 and US-SL-2014 (supplementary Figures S.22–S.25) and other regional tomographic mod-  
 599 els (e.g., Grand & Helmberger, 1984; van der Lee & Frederiksen, 2005; Nettles & Dziewon-  
 600 ski, 2008; Bensen et al., 2009; H. Yuan & Romanovicz, 2010; Khan et al., 2011; H. Yuan  
 601 et al., 2011, 2014; W. Shen & Ritzwoller, 2016; Krischer et al., 2018).

602 2) We are able to image mantle features underneath the western margin of the con-  
 603 tinent (region TAW), where volcanism, accretion, and extensive tectonics are the likely  
 604 surface expression of ancient and active subduction (e.g., Sonder & Jones, 1999; Dick-  
 605 inson, 2004; Humphreys, 2009). We observe localized high-velocity anomalies in the north  
 606 that move further inland and southward at greater depths (Figure 11B) and probably  
 607 represent fragmented pieces of the Juan de Fuca, Gorda, and Farallon slabs (e.g., Sigloch,  
 608 2011; Portner & Hayes, 2018). We further see a dominant low-velocity band parallel to  
 609 the Pacific coast, which, at shallow depths, forms a circular structure around the Col-  
 610 orado plateau (P4) that extends towards Yellowstone. Similar features were identified  
 611 by Becker (2012) in a comparison of tomographic models of the region (Roth et al., 2008;  
 612 Burdick et al., 2008, 2010; Obrebski et al., 2010, 2011; Schmandt & Humphreys, 2010,  
 613 2011; James et al., 2011; Sigloch, 2011). However, it has to be pointed out that: 1) res-



614 olution of our model is low because of reduced data coverage at shallow depth at the flank  
 615 of the eastern Rocky Mountains (P5); and 2) high basalt fractions, as inferred from Sp  
 616 RFs, tend to decrease velocity within the MTZ, which exaggerates low-velocity anom-  
 617 alies in regions where Ps RF data points are limited.

618 3) Considering the central and eastern part (regions CTR and COE), we find mul-  
 619 tiple high-velocity anomalies that extend to the MTZ, where a roughly north-south aligned  
 620 band underneath the Great plains (P6) is the most pronounced feature (see east-west  
 621 cross section). These anomalies could be interpreted as remnant pieces of the Farallon  
 622 or Laramide slabs, although differences in shape relative to other studies are apparent  
 623 (e.g., Sigloch, 2011; Schmandt & Lin, 2014; Porritt et al., 2014).

624 4) Toward the bottom of the upper mantle, amplitudes of shear-wave anomalies  
 625 generally decrease, but again increase significantly at and around 400 km depth as olivine  
 626 transforms to wadsleyite. Lateral velocity variations are larger than those above and be-  
 627 low the transition and emphasizes the importance of considering phase-induced varia-  
 628 tions in addition to variations arising purely from volumetric velocity perturbations. Struc-  
 629 ture in and around the MTZ is complex and shows significant topography, particularly  
 630 on the 410 (Figure 12A). As shown in Figure 12B, we find that the 410 occurs deeper  
 631 (409–432 km) in “hot” regions (TAW) than in “colder” parts (404–423 km in CTR; 401–  
 632 422 km in COE). To compositionally achieve an equivalent 410-depth variation (>30 km),  
 633 basalt fraction would have to change by >50% (see supplementary Figure S.16), which  
 634 corresponds to more than the total compositional variation observed across the whole  
 635 study area (Figure 9D). This suggests that composition is of secondary nature in deter-  
 636 mining topography on the 410.

637 5) In contrast to the large peak-to-peak amplitudes of >30 km for the 410, lateral  
 638 variations of the 660 are only  $\pm 5$  km (Figure 12C) and there appears to be little corre-  
 639 lation with temperature (Figures 12D). Instead, the dependence of the 660 on temper-  
 640 ature and composition is highly non-linear (Figure S.14) and also affected by the trans-  
 641 formation of garnet, whose relative importance increases as temperature and basalt frac-  
 642 tion increases (e.g., Weidner & Wang, 1998; Hirose, 2002; Xu et al., 2008; Khan et al.,  
 643 2009) (see also supplementary Figures S.3 and S.4 of Munch et al. (2018)). Since the Clapey-  
 644 ron slopes for the transformations of garnet and ringwoodite are opposite in sign (Hirose,  
 645 2002; Liu et al., 2018), the garnet phase transition would follow the behaviour of the 410  
 646 and move deeper in regions of elevated temperature.

647 6) As a result, MTZ-thickness (Figure 12E) is largely controlled by the topogra-  
 648 phy on the 410 and is thinner in region TAW ( $\sim 220$ – $250$  km) than in CTR and COE  
 649 ( $\sim 230$ – $260$  km). The 410 beneath the western USA (TAW), where two strong depres-  
 650 sions separated by a high-elevation band across northern Nevada (see Figure 11B) are  
 651 observable, agrees well with other Ps and Sp RF studies of the region (Cao & Lavender,  
 652 2010; Tauzin et al., 2013; Gao & Liu, 2014) and possibly relates to subduction. In con-  
 653 trast, the 660 and hence MTZ thickness varies more across the aforementioned studies.  
 654 This first and foremost reflects the use of different imaging methods and different ref-  
 655 erence models in order to convert receiver functions from time to depth. Furthermore,  
 656 to the east of the Rocky Mountain front (CTR and COE), our inferred 410 and hence  
 657 MTZ thickness tends to “normal” values, in overall agreement with the continental-scale  
 658 MTZ model of Gao and Liu (2014). Further to this, Gao and Liu (2014) found, in agree-  
 659 ment with Maguire et al. (2018), the MTZ to be locally thicker underneath the Great  
 660 Plains (P6), which they attributed to remnants of the Farallon or Laramide slabs. Al-  
 661 though less prominent, similar MTZ anomalies are also observed here.

## 662 7 Comparison of seismic and thermobarometric constraints

663 Assessment of the chemical and petrological characteristics of erupted lavas can pro-  
 664 vide complementary information to seismic inversions to understand the thermo-chemical  
 665 structure of the underlying mantle. In order to facilitate this comparison, we compiled  
 666 whole rock chemical analyses of  $\sim 500$  primitive basaltic rocks from the western USA,

667 filtered for basalts with eruption ages younger than 5 Ma and with MgO contents  $\geq 8$  wt%  
 668 (see section S.7 for details). The composition of each sample was inverted in order to solve  
 669 for the mean pressure and temperature of melt segregation according to the thermobarom-  
 670 eter of Lee et al. (2009) to create an internally-consistent database. The resultant es-  
 671 timates on (minimum) lithosphere thickness and mantle potential temperature for each  
 672 sample have uncertainties of about 17 km and 50 °C, respectively. Our results are con-  
 673 sistent with those of Plank and Forsyth (2016) to within ( $\pm 20$  °C) and ( $\pm 15$  km) for over-  
 674 lapping localities.

675 Comparison between petrological and seismological T-P estimates is subject to caveats  
 676 associated with i) the differing spatial coverage, and ii) the degree to which phase- and/or  
 677 compositional changes influence the seismic inversion (section 4.1). To account for the  
 678 locally-confined spatial sensitivity of thermobarometry relative to seismic wavelengths  
 679 (cf. section 2.2.3), we masked cells beyond a radius of  $1.5^\circ$  to a sample location. Litho-  
 680 spheric thickness is more difficult to compare given the reduced data coverage in the western-  
 681 most USA and the imposed seismic lower bound of 70 km.

682 With these considerations in mind, the interpolated results from thermobarome-  
 683 try are shown in Figure 13A–B, while the difference to seismically derived potential tem-  
 684 perature estimates are presented in Figure 13C–D. The major magmatic provinces are  
 685 demarcated on the basis of both seismic and petrological inversions (Figure 9 and 13).  
 686 Thermally anomalous mantle is observed underneath the Cascade range (northern P1)  
 687 and the Snake River Plain (eastern P2) towards Yellowstone Hotspot, and particularly  
 688 under the Basin and Range province (P3) in southern Nevada, extending towards the  
 689 southeast in the Rio Grande rift. Both methods agree on the relative amplitude of ther-  
 690 mal anomalies, but differ in absolute values.

691 The largest temperature discrepancies, in which seismological estimates are 100–  
 692 250 °C higher with respect to thermobarometric inversions (cf. Figure 13C–D), are pri-  
 693 marily observed along the Cascade arc parallel to the Pacific coast (P1). The petrologically-  
 694 derived potential temperatures, 1300–1350 °C, are in agreement with other thermobarom-  
 695 etric estimates (Grove et al., 2002; Leeman et al., 2005), whereas temperatures derived  
 696 from the seismic model may reflect the cumulative effects of increased  $H_2O$ ,  $fO_2$ , and  
 697 melt fraction underneath this region, none of which are considered. Influx of water and  
 698 oxidising melts from the downgoing slab into the mantle wedge are characteristics of sub-  
 699 duction zones, with the net effect being that increasing  $Fe^{3+}$  in olivine results in increased  
 700 dissipation of the shear modulus (Cline II et al., 2018), such that more oxidised/water-  
 701 rich mantle appears seismically slower (and therefore hotter). Other regions show smaller  
 702 differences (0–100 °C), and pertain to largely anhydrous tholeiitic provinces (Snake River  
 703 Plain, Southern Nevada, and New Mexico) formed by decompression melting of upwelling  
 704 asthenosphere (Fitton et al., 1991; Feuerbach et al., 1993; Leeman et al., 2009). This agree-  
 705 ment therefore empirically verifies the accuracy of the seismic inversion in quantifying  
 706 the thermo-chemical characteristics of more typical upper mantle.

707 The strength of this comparison lies in that surficial expressions of thermal anoma-  
 708 lies may be linked to deep mantle structures. It also highlights the sensitivity of seismologically-  
 709 derived temperature-depth estimates to the specific petrological characteristics of the man-  
 710 tle, which changes with tectonic setting. Correctly accounting for these variations could  
 711 help solve ambiguity in the chemical and isotopic signatures of eruptive provinces by imag-  
 712 ing the thermal state of the underlying mantle in more detailed seismic and petrolog-  
 713 ical surveys in future.

## 714 8 Summary and conclusions

715 In this paper, we proposed a methodology for processing, modeling, and inversion  
 716 of multi-frequency P-to-s and S-to-p converted body waves (receiver functions) and ap-  
 717 plied it to data recorded by USArray to perform a tomographic study of crust and man-  
 718 tle beneath the North American continent.



Careful data selection and processing has been found to be a central step in the construction of Sp receiver functions, since these are typically prone to be noisy and sense distinct mantle regions depending on source-receiver configuration. To properly account for this, we binned data into different back-azimuth bins and considered these separately. To further increase the robustness of our Sp RF stacks, we augmented the latter by combining data from USArray reference and transportable stations. Ultimately, only high-quality Ps and Sp receiver functions with clear conversions at the Moho, lithosphere, 410, and 660 were retained for inversion. Finally, to ensure consistency between observed and synthetic RF waveform amplitudes, processing of modeled waveforms mimicked that applied to real data in detail. As part of waveform verification, we also conducted a validation of our approach to computing synthetic waveforms in 1-D models by comparing these to full waveform synthetics computed in a 3-D velocity model.

Relying on a thermodynamic formulation, the forward model consisted of first computing seismic velocity and density models via Gibbs' free energy-minimization, followed by seismic waveform modeling to construct maps of the crust and thermo-chemical mantle structure underneath the contiguous USA. We were able to identify three major tectonic regions based on similarities in radial temperature and S-wave velocity structure:

1) The tectonically active West (TAW), comprising regions west of the Rocky Mountains, is characterised compositionally as more fertile ( $\text{Mg}\# \sim 0.90$ ), physically by a thinner lithosphere ( $100 \pm 23$  km), and thermally by higher mantle potential temperatures ( $1490 \pm 27$  °C). Consequently, shallow upper mantle S-wave velocity is strongly diminished (by up to  $-5\%$ ), shear-attenuation is increased, and the 410-km seismic discontinuity is shifted to greater depths ( $424 \pm 7$  km).

2) The cratonic-orogenic East (COE), including most of the continent to the east of the Great Plains, is compositionally more depleted ( $\text{Mg}\# \sim 0.91$ ) and colder ( $1413 \pm 27$  °C), underlain by an overall thicker lithosphere ( $208 \pm 33$  km). Accordingly, S-wave velocity is higher (by up to  $+5\%$ ), shear-attenuation is decreased, and the 410 appears at shallower depths ( $416 \pm 6$  km).

3) The Central transition regions (CTR) represent tectonic regions with intermediate properties in-between those of TAW and COE.

Despite being able to image well-known mantle features such as traces of active and ancient subduction or the Yellowstone plume, we note that resolving the exact shape and extent of such 3-D volumetric features by our method is limited by two factors: 1) we do not solve the inverse problem in a fully 3-D framework, but rather use a 1-D formulation to build a 3-D model a posteriori, and 2) converted waves are not directly sensitive to volumetric velocity changes, but rather to velocity contrasts across discontinuities. Further limitations of our method arise from the use of isentropes in modeling mantle thermal structure, end-member compositional model, and the assumption of an equilibrium assemblage configuration. Further, the absence of thermodynamic data, for now, involving melt-, and water-bearing systems might bias our results in certain regions, such as underneath the Cascades arc, as evidenced in the apparent mismatch with petrological estimates. Yet in spite of these limitations, a fully equilibrated, compositionally uniform, adiabatic, and dry mantle nevertheless provides an excellent fit to the observed seismic data. Our combination of geophysically-derived potential temperature- and lithospheric thickness estimates to those obtained from thermobarometric inversions of volcanic rock chemistry in the western USA show first-order agreement, permitting the assessment of broad-scale mantle melting conditions and their magmatic expression. This integrated approach represents a promising avenue for providing multidimensional insight into crustal tectonic structures and regimes and their link with deep mantle processes.

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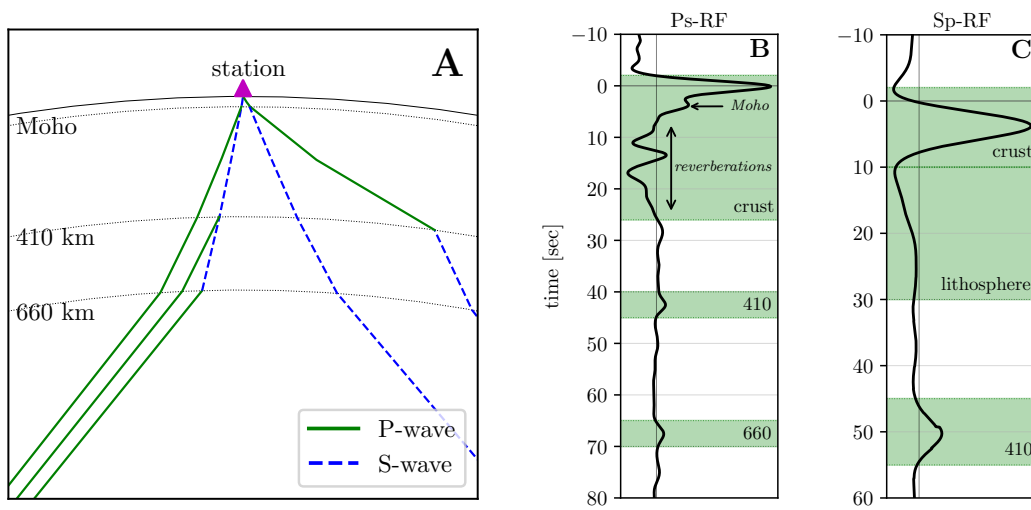
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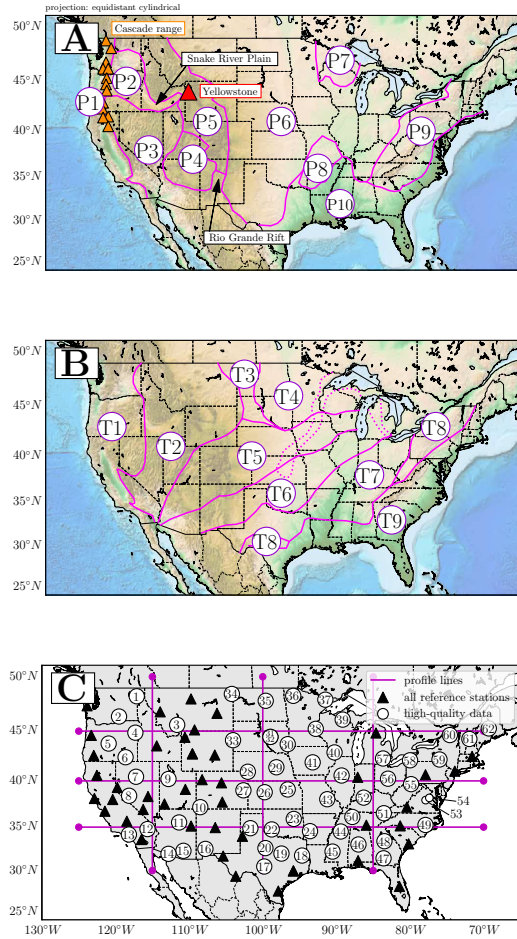
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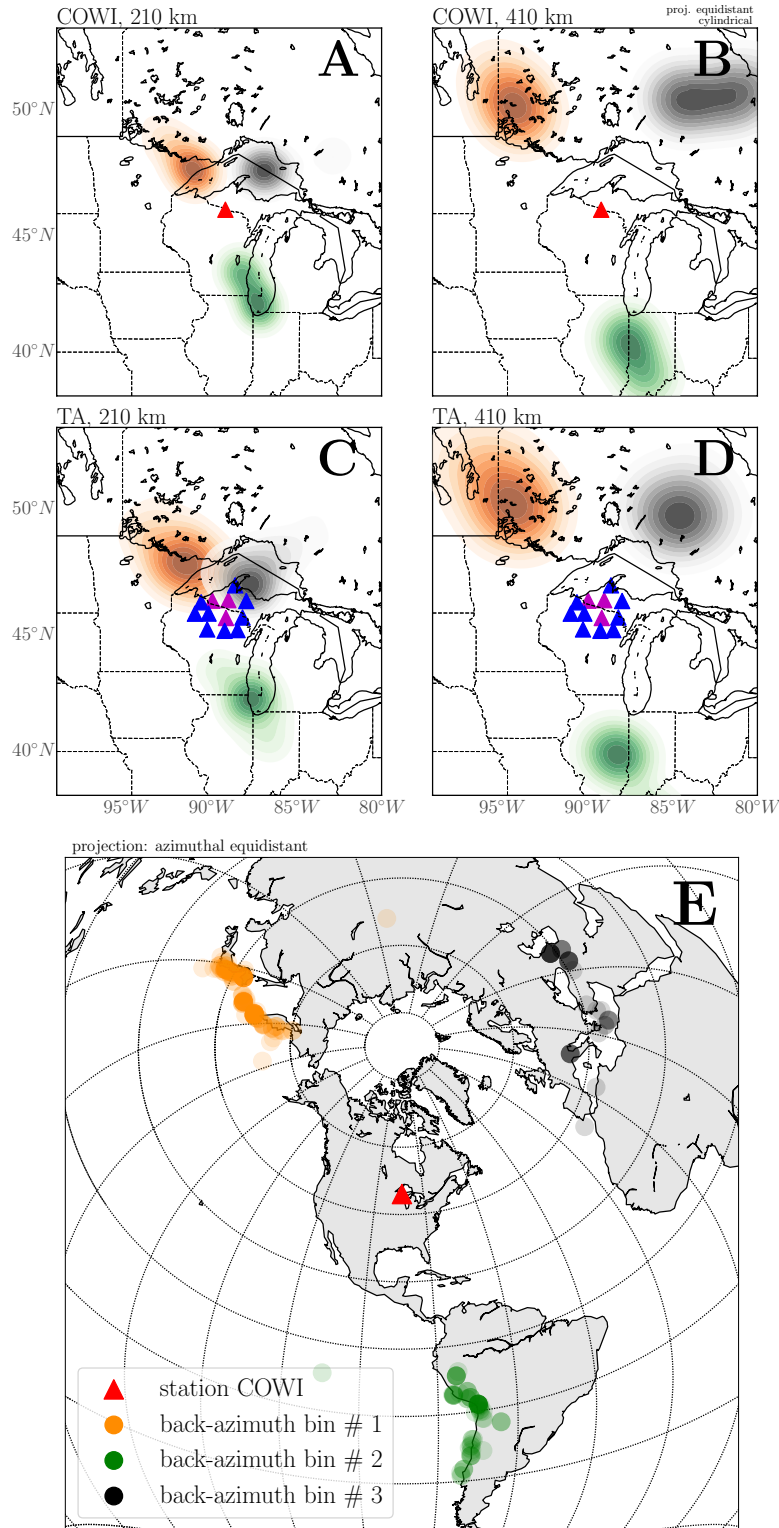




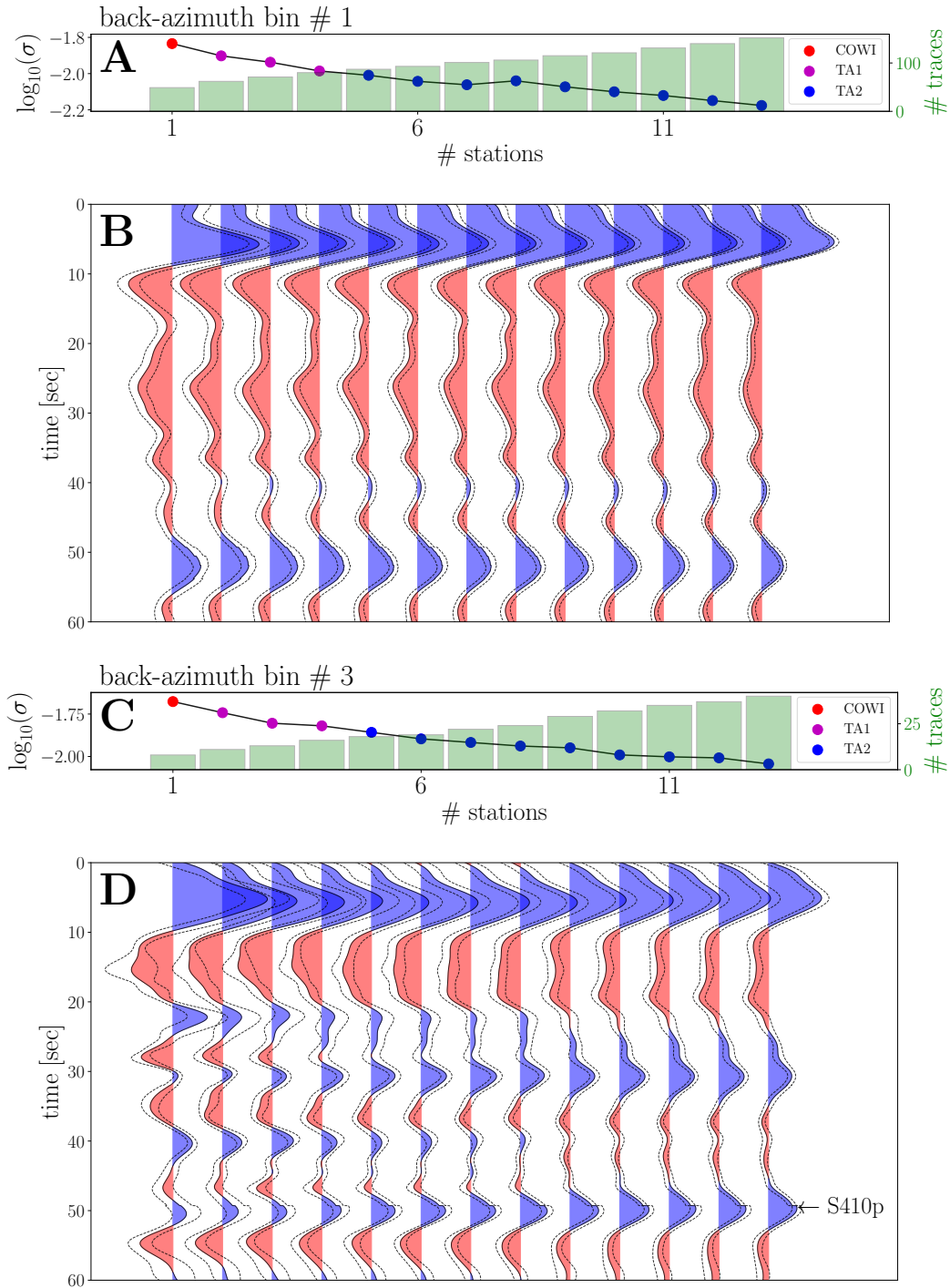
**Figure 1.** Illustration of conversions of body waves at the Moho and the 410- and 660-km seismic discontinuities in the Earth (A) and corresponding synthetic P-to-s (B) and S-to-p receiver functions (C) based on the seismic reference model IASP91 (Kennett & Engdahl, 1991). P- and S-waves are shown as green (solid) and blue (dashed) lines, respectively. Time windows including arrivals of converted waves are shaded in green. Note that the polarity and time axis of Sp RFs are flipped as a result of which arrivals with positive travel times arrive prior to the S-wave and are not masked by crustal reverberations as is the case for P-to-s conversions



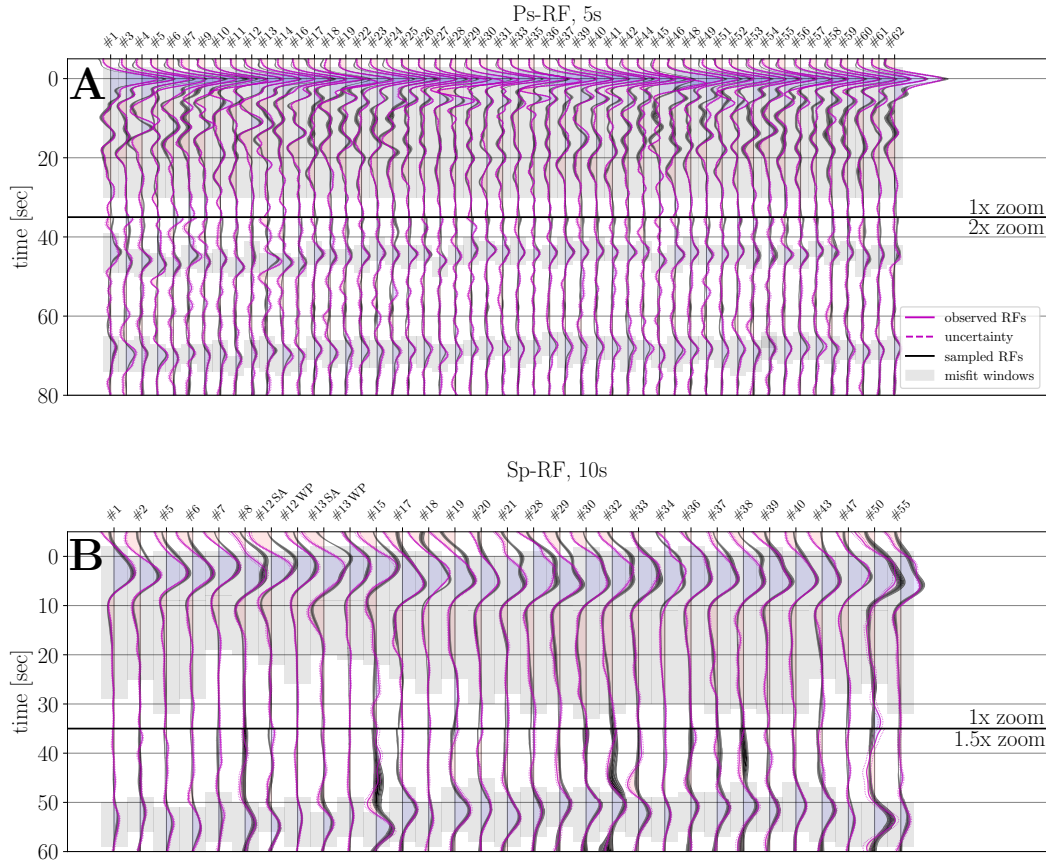
**Figure 2.** Main physiographic features (Fenneman, 1928) of the contiguous USA (**A**): P1: Pacific mountain system (incl. Cascade Range); P2: Columbia plateau (incl. Snake River Plain and Yellowstone Hotspot); P3: Basin and Range (incl. Rio Grande Rift); P4: Colorado plateau; P5: Rocky mountain system; P6: Interior Highlands; P7: Laurentian Highlands; P8: Interior Highlands; P9: Appalachian Highlands; P10: Atlantic Plain. The tectonic provinces (Hoffman, 1988; Whitmeyer & Karlstrom, 2007) represent the long-term geological history of the continent **B**: the oldest parts of the continent, i.e., of precambrian age, include the Greater Wyoming province (T2) and Superior craton (T4), which are connected along the Trans-Hudson orogen (T3). Proterozoic orogenies accreted the Yavapai (T5), Mazatzal (T6), Granite-Rhyolite (T7), and Grenville provinces (T8) to the cratonic core. During episodes of extension in proterozoic times, the mid-continental rift was formed but abandoned, and later the Pacific and paleo-Atlantic oceans were opened. The closure of the latter during the paleozoic formed the Appalachian mountain belt (T9, P9), which, nowadays, is bordered by the Atlantic plains (P10), i.e., the relatively young passive continental margin, to the east and south. In contrast to the east, the western margin (T1) is dominated by subduction processes acting since the mesozoic (e.g., Sonder & Jones, 1999; Dickinson, 2004; Humphreys, 2009), which have resulted in accretion of terrane to the American continent, volcanism (Cascade range, P1), and extensional tectonics (Basin and Range, P3). The topography model underlying the maps is that of ETOPO1 (Amante & Eakins, 2009). **C**: location of USArray reference stations ([www.usarray.org](http://www.usarray.org)). White circles indicate stations, where either P-to-s or S-to-p receiver functions or both were inverted, while black triangles represent stations for which data were processed but because of quality constraints were not considered in the inversion. Station numbering and details are given in Table S.1. Magenta lines indicate profiles along which cross sections of mantle seismic properties are shown



**Figure 3.** A–D: normalized Fresnel-zone weights of S-to-p piercing points (color coded by back-azimuth; see panel E) at a depth of 210 km (A and C) and 410 km (B and D) for the events recorded at 39-COWI (red triangle) and 12 near-by transportable USArray stations within a distance of  $\sim 70$  km (TA1; magenta triangles) and  $\sim 140$  km (TA2; blue triangles), respectively. E: location of station 39-COWI and the seismic events used for computing S-to-p receiver functions coloured by back-azimuth (dots)

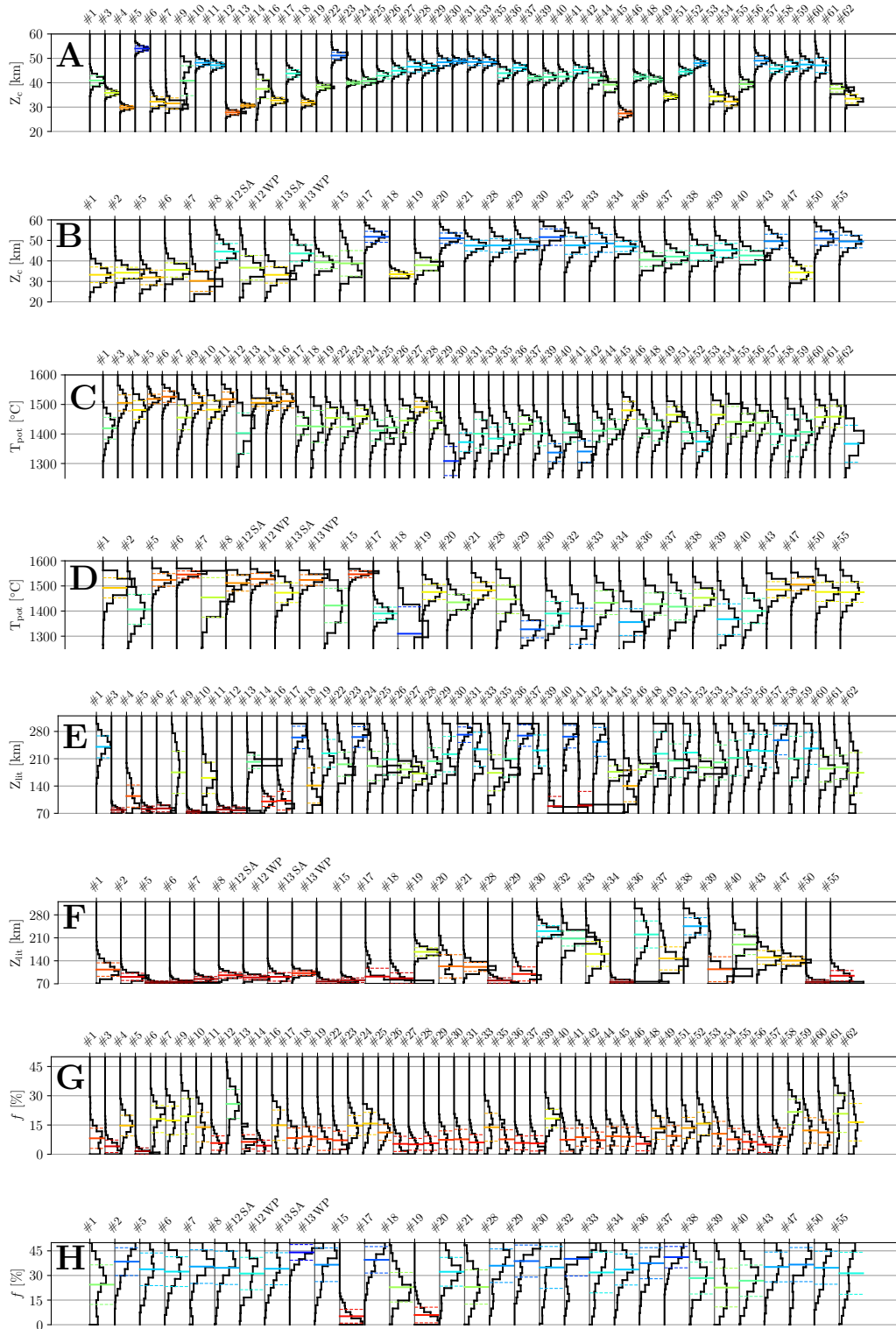


**Figure 4.** S-to-p receiver function stacks, mean uncertainty ( $\sigma$ ), and number of summed traces (green bars) for two distinct back-azimuth bins related to events in the west Pacific (#1; **A** and **B**) and Europe (#3; **C** and **D**), respectively. Including data recorded at transportable stations (blue and magenta dots) (TA1 and TA2, respectively) of the reference station (39-COWI, red dot) reduces uncertainty and enhances the S410p phase in back-azimuth bin # 3 (**D**)

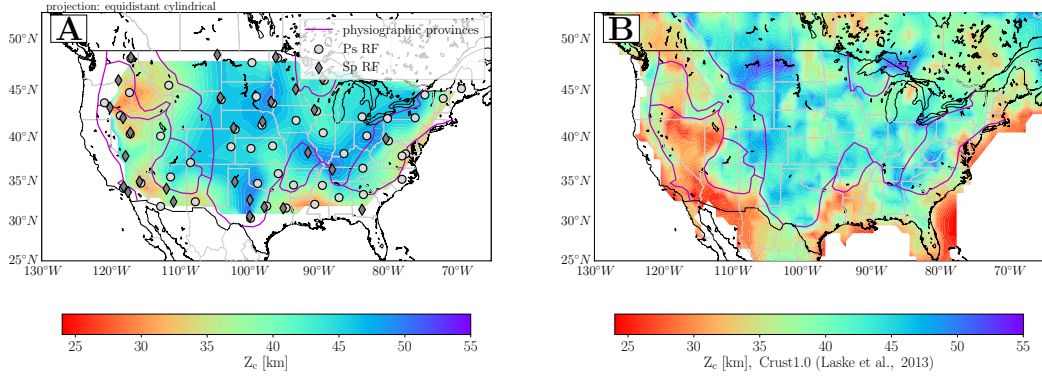


**Figure 5.** Data misfit between inverted and observed high-quality P-to-s (5 s low-pass period, **A**) and S-to-p receiver functions (10 s low-pass period, **B**). Observations and uncertainties are shown as magenta lines, sampled receiver functions as contours (dark bands), and misfit windows in light gray. Station numbering is shown in Figure 2C, whereas ‘SA’ and ‘WP’ denote events located in South America and the West Pacific, respectively. For improved visibility, amplitudes of conversions from the 410- and 660-km seismic discontinuities have been scaled up. Equivalent data fits for P-to-s and S-to-p receiver functions at low-pass periods of 8, 10, and 15 s are shown in supplementary Figure S.19

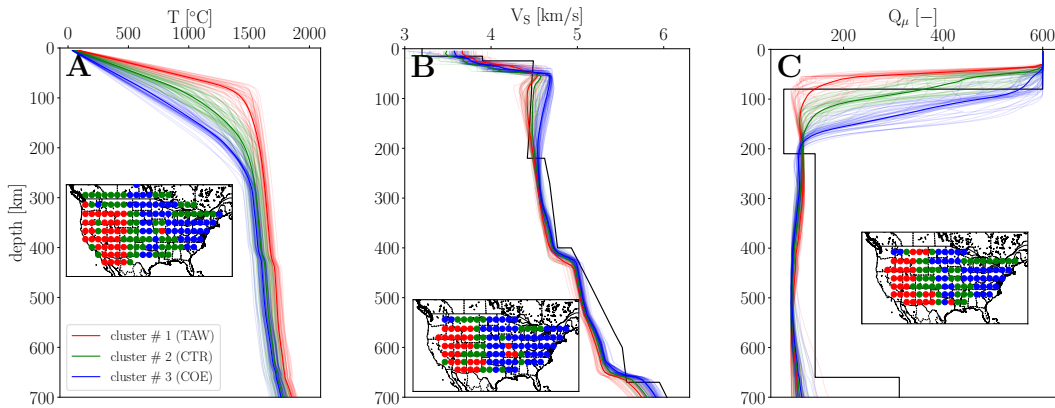




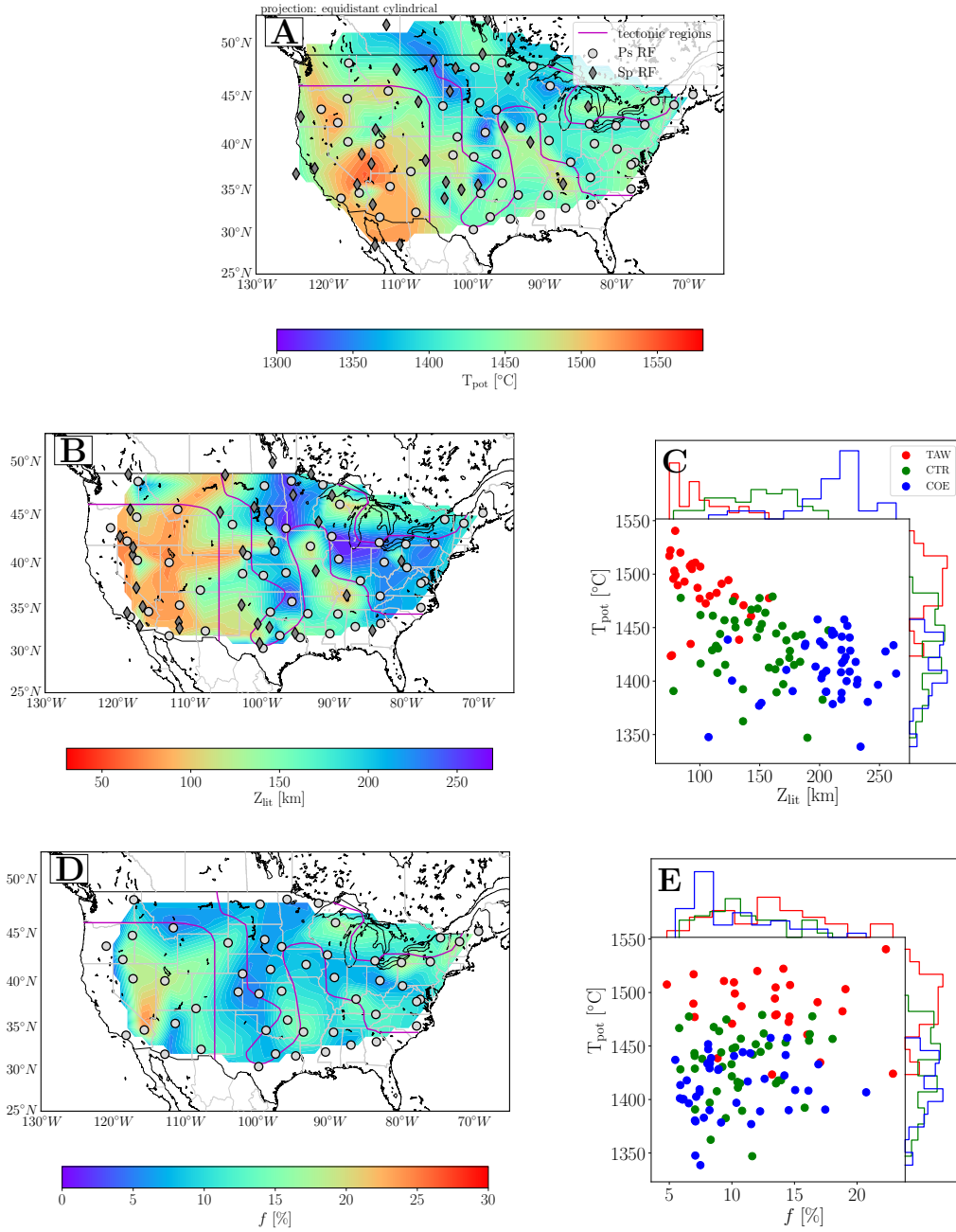
**Figure 6.** Sampled marginal posterior probability distributions (black lines) of crustal thickness,  $Z_c$  (A, B), potential temperature,  $T_{pot}$  (C, D), lithosphere thickness,  $Z_{lit}$  (E, F), and basalt fraction,  $f$  (G, H) from inversion of P-to-s and S-to-p receiver functions, respectively. Mean values and standard deviations of the distributions are shown as thick and thin lines, respectively, where choice of color is based on the colormaps used in Figures 7 and 9. Station numbering refers to Figure 2C, whereas ‘SA’ and ‘WP’ refer to events from South America and the West Pacific, respectively



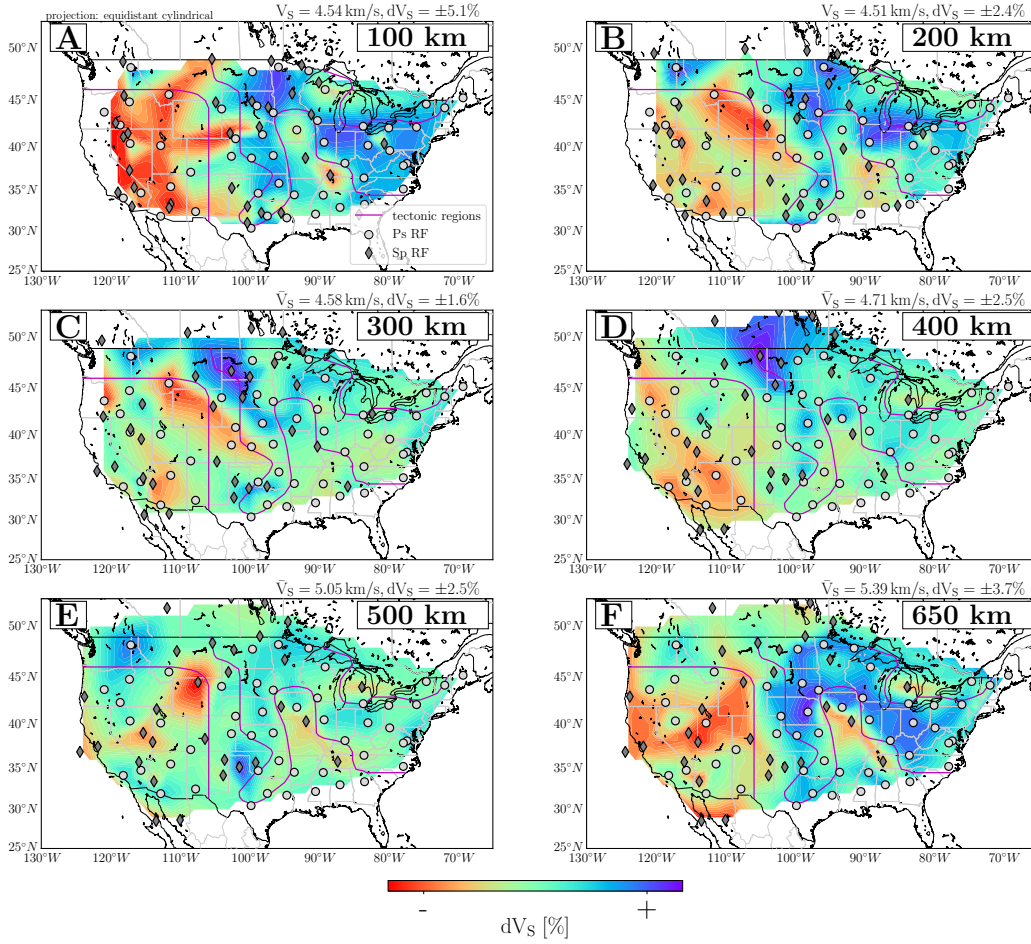
**Figure 7.** **A:** Mean crustal thickness,  $Z_c$ , across the contiguous USA interpolated from migrated P-to-s (circles) and S-to-p (diamonds) receiver functions. **B:** Crustal thickness map from model Crust1.0 (Laske et al., 2013). Physiographic provinces are outlined in magenta (Fenneman, 1928, cf. Figure 2A)



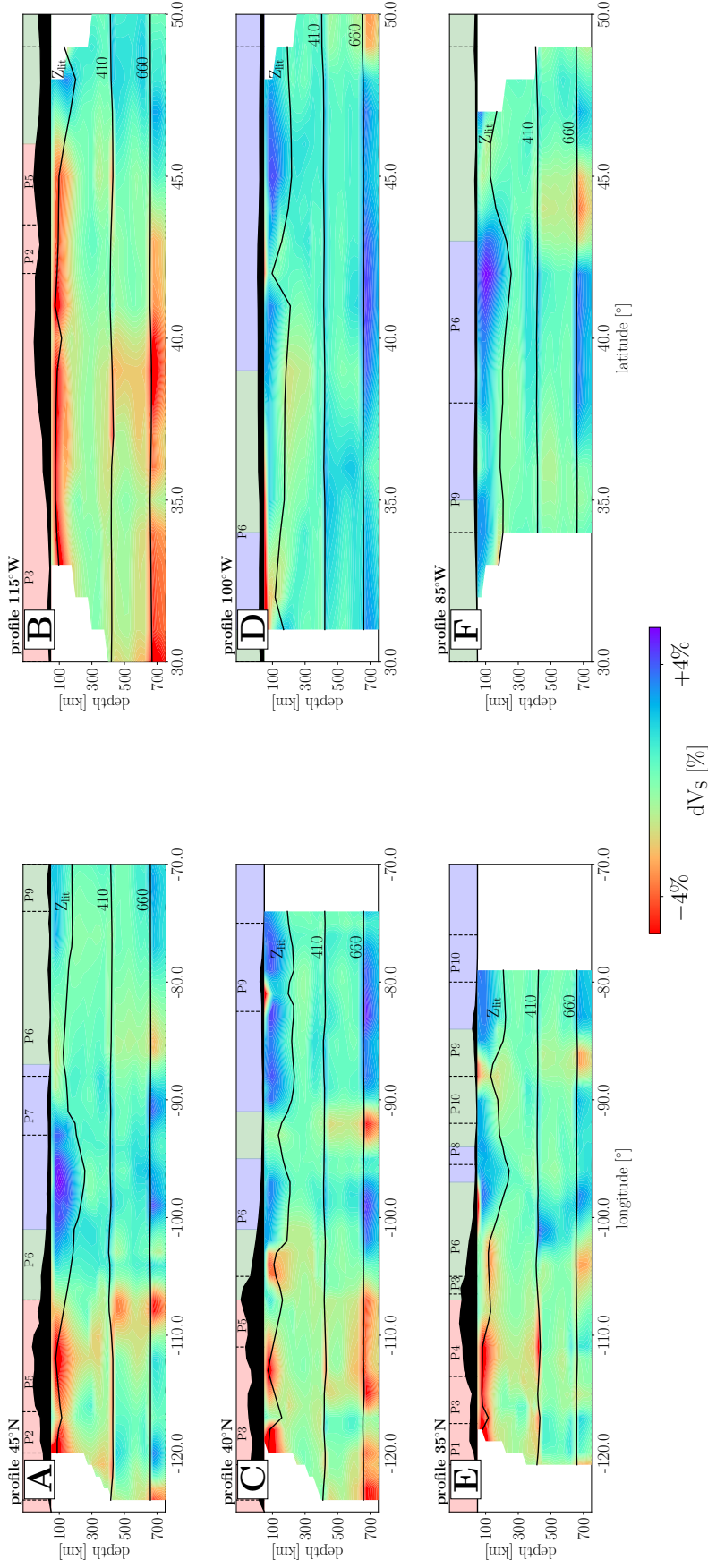
**Figure 8.** Profiles of temperature ( $T$ , **A**), S-wave velocity ( $V_s$ , **B**), and shear attenuation profiles ( $Q_{\mu}$ , **C**) obtained from inversion of P-to-s and S-to-p receiver functions and clustering analysis. Profiles are clustered into three clusters according to similarity (color-coded) and displayed in light color, whereas the mean profile of each cluster is shown as a thick line. The tectonic regions identified in panel **A** will be referred to as tectonically active West (red; TAW), central transition regions (green; CTR), and cratonic-orogenic East (blue; COE) in the following. The maps shown in the insets are interpolated from the inversion results on a  $2.5^{\circ} \times 2.5^{\circ}$ -grid. PREM profiles are shown in black in panels (**B**) and (**C**)



**Figure 9.** Models of the North American thermo-chemical mantle structure: potential temperature,  $T_{\text{pot}}$  (A), lithosphere thickness,  $Z_{\text{lit}}$  (B), and basalt fraction,  $f$  (D). Tectonic regions identified earlier (cf. Figure 8A) are outlined in magenta. The distribution and correlation of  $T_{\text{pot}}$  with  $Z_{\text{lit}}$  and  $f$  for each of these tectonic regions (color coded) are shown in panels C and E, respectively. Maps derive from P-to-s (circles) and S-to-p (diamonds) receiver functions, respectively, except the compositional map showing basalt fraction (D), where S-to-p receiver function estimates are excluded because of the large uncertainties (see section 6.6). Differences in lateral extent of these maps arise from the migration of S-to-p receiver function results along the converted raypath to different depths (see section 5.4)

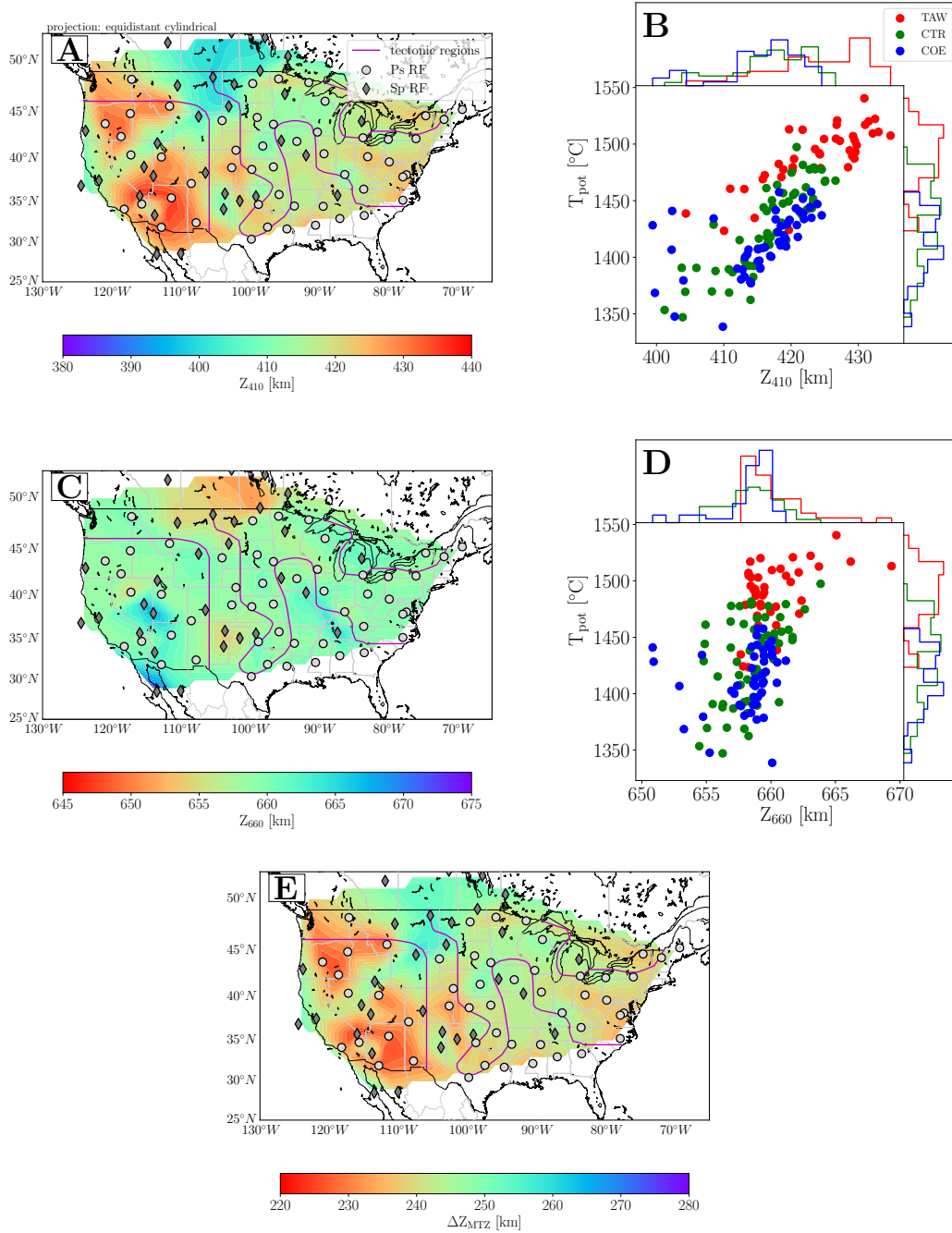


**Figure 10.** Interpolated S-wave velocity perturbations,  $dV_S$ , along horizontal slices between 100–650 km depth. All maps share the same colormap but differ in absolute mean velocity and standard deviation (indicated on the top right of each panel). Earlier identified tectonic regions (cf. Figure 8A) are outlined in magenta. Differences in lateral extent of these maps arise from the migration of S-to-p receiver function results along the converted raypath to different depths (see section 5.4)

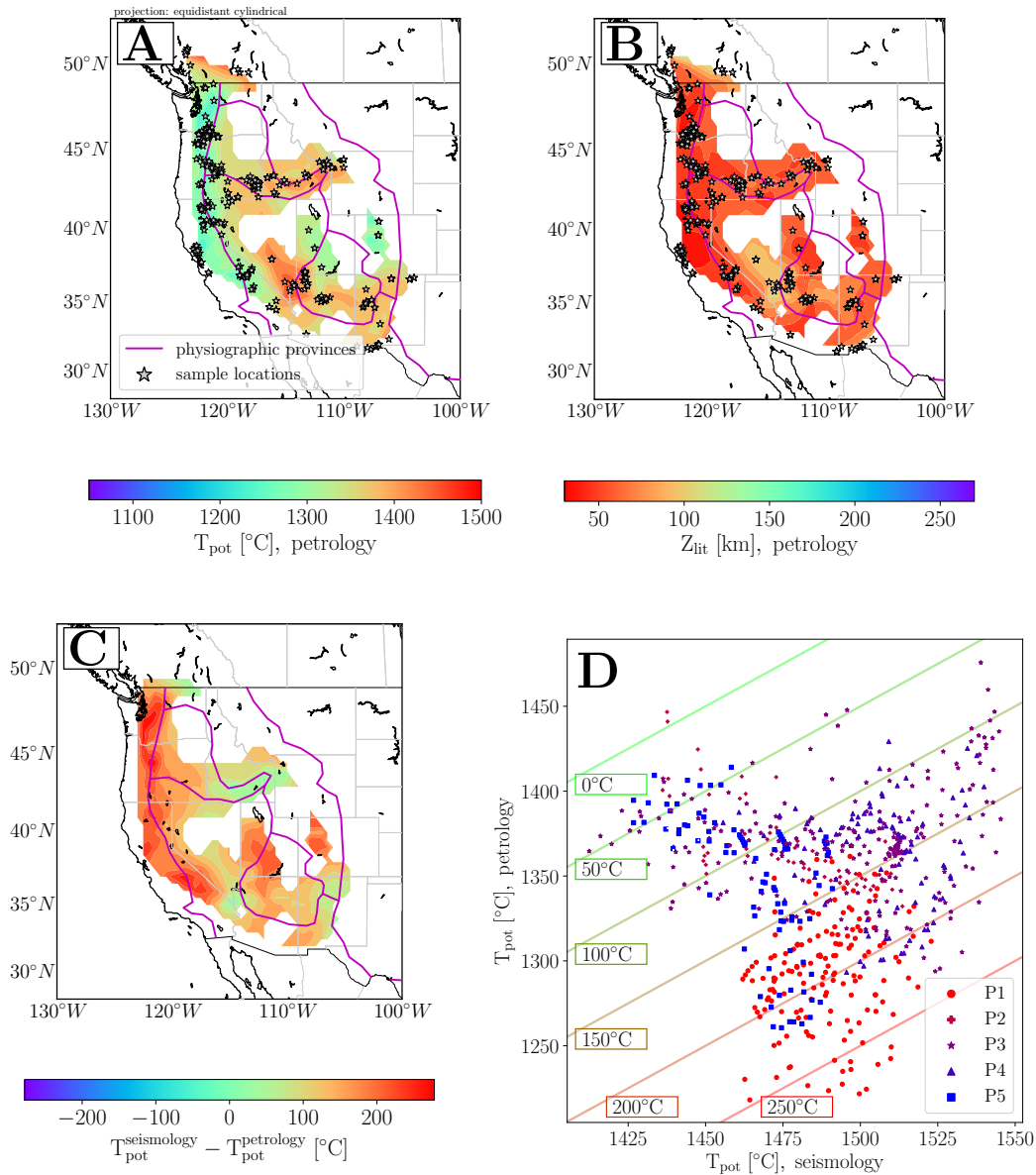


**Figure 11.** Longitudinal cross sections of S-wave velocity perturbations (**A**, **C**, and **E**) at 45°N, 40°N, and 35°N latitude, respectively, and latitudinal cross sections (**B**, **D**, and **F**) at 115°W, 100°W, and 85°W longitude, respectively. Lines indicating location of cross sections are shown in Figure 2C. The perturbation is computed relative to the mean velocity with depth (not shown). Surface topography (Laske et al., 2013) is exaggerated 50× and black lines represent topography on lithosphere and 410- and 660-km seismic discontinuities. Differences in lateral extent with depth arise from the migration of S-to-p receiver functions along the converted raypath to different depths (see section 5.4). Above each panel, the extent of the tectonic regions identified in Figure 8A are indicated by different colors: TAW (red), CTR (green), and COE (blue). Physiographic provinces (Fenneman, 1928, cf. Figure 2A) are outlined with dashed vertical lines





**Figure 12.** Interpolated maps of mean depth to the major mantle seismic discontinuities at 410 and 660 km depth,  $Z_{410}$  (A),  $Z_{660}$  (C), and mantle transition zone thickness,  $\Delta Z_{MTZ}$  (E). Tectonic regions identified in Figure 8A are outlined in magenta. The distribution and correlation of mantle potential temperature  $T_{pot}$  with  $Z_{410}$  and  $Z_{660}$  for each of these regions (color coded) is shown in panels B and D, respectively



**Figure 13.** Thermo-chemical structure of the lithosphere underneath the western USA as estimated from thermobarometry: interpolated potential temperature ( $T_{\text{pot}}$ , **A**) and lithosphere thickness ( $Z_{\text{lit}}$ , **B**). The deviation to seismically derived temperature estimates (cf. Figure 9A) is shown in panel **C**. Cells of the interpolation grid farther away than  $1.5^\circ$  from the nearest sample location (stars) are masked. Physiographic provinces are outlined in magenta (Fenneman, 1928, cf. Figure 2A). Panel **D** compares the petrologically and seismically derived temperature estimates (markers colored by physiographic province). Isotherms represent absolute temperature differences, i.e.,  $T_{\text{pot}}(\text{seismology}) - T_{\text{pot}}(\text{petrology})$  (color coded and labeled)