Shear velocity inversion guided by resistivity structure from the PI-LAB Experiment for integrated estimates of partial melt in the mantle

Nicholas Harmon[1], Shunguo Wang[2], Catherine A. Rychert[1], Steven Constable[3], J Michael Kendall[4]

[1] University of Southampton, Ocean and Earth Science, Southampton, UK
[2] Norwegian University of Science and Technology, Department of Electronic Systems, Trondheim, Norway
[3] University of California San Diego, Scripps Institution of Oceanography, San Diego, USA
[4] Oxford University, Department of Earth Science, Oxford, UK

Abstract

The lithosphere-asthenosphere system is fundamental to our understanding of mantle convection and plate tectonics. The different sensitivities of seismic and electromagnetic methods can be used together to better constrain the properties of the system. Here we re-examine the shear velocity model from Rayleigh waves in light of the magnetotelluric based resistivity models from the PI-LAB experiment near the equatorial Mid-Atlantic Ridge, with the goal of generating a structurally consistent velocity and resistivity model for the region. Cross-plots of the models suggest a linear or near-linear trend that is also in agreement with petrophysical predictions. We generate a new shear velocity model from the resistivity models based on petrophysical relationships. The new velocity model fits the phase velocity data, and the correlation coefficient between the shear velocity and resistivity models is increased. Much of the model can be predicted by expectations for a thermal half-space cooling model, although some regions require a combination of higher temperatures, volatiles, or partial melt. We use the petrophysical predictions to estimate the melt fraction, melt volatile content, and temperature structure of the asthenospheric anomalies. We find up to 4% melt, with the lowest resistivities and shear velocities explained by up to 20% water or 20% CO_2 in the melt or ~1% nearly pure sulfide melt, depending on the set of assumptions used. Melt is required in punctuated anomalies over broad depth ranges, and also in channels at the base of the lithosphere. Melt in the asthenosphere is dynamic, yet persistent on geologic time scales.

Introduction

Plate tectonic theory is predicated on the idea of a rigid lithosphere that overrides a weaker underlying asthenosphere (McKenzie & Parker, 1967), but the nature of the lithosphere-asthenosphere system remains the subject of vigorous debate. The oceanic lithosphere comprises the majority of the surface of the Earth and has the simplest evolution and history. It is classically thought to be thermally defined as a boundary layer in a simple thermal model (Parker & Oldenburg, 1973). In this model, increasing temperature with depth causes mantle rocks to weaken, creating the asthenosphere (e.g.,Goetze et al., 1978). However, a host of observations, including sharp seismic velocity discontinuities (Gaherty et al., 1996; Kawakatsu et al., 2009; Rychert et al., 2020; Rychert et al., 2018; Rychert & Shearer, 2011; Schmerr, 2012; Tan & Helmberger, 2007; Tharimena et al., 2017), low

velocity zones (Forsyth et al., 1998; Harmon et al., 2020), and low resistivity zones (Baba et al., 2006; Johansen et al., 2019; Key et al., 2013; Naif et al., 2013; Wang et al., 2020) in the asthenosphere, suggest that in addition to temperature other factors are likely required to explain the observations. Many potential explanations of these observations have been proposed including an increased effect of hydration (Karato, 2012), the presence of partial melt (Anderson & Sammis, 1970; Kawakatsu et al., 2009), and/or the enhanced effects at near sub-solidus conditions on seismic waves (Yamauchi & Takei, 2016). The debate centers around which of these explanations might be in operation and how widely they apply.

Partial melt is an attractive possibility given that it provides an explanation for a wide range of observations with different sensitivities (Rychert et al., 2020). Partial melt likely to exist in the asthenosphere, in particular near mid-ocean ridges and volcanic arcs where the volcanic systems must be fed by mantle melting (Anderson & Sammis, 1970). However, further away from volcanic plate boundaries its presence is debated (Kawakatsu et al., 2009; Priestley & McKenzie, 2006; Rychert et al., 2005). The amount of melt and its location is vital to our understanding of how the lithosphere-asthenosphere works, as the presence of partial melt is predicted to reduce the viscosity of the asthenosphere (Hirth & Kohlstedt, 1995; Jackson et al., 2006) and could also facilitate plate tectonics (Rychert et al., 2005; Rychert et al., 2007). However, different geophysical techniques with different sensitivities and resolutions have imaged anomalies that have been interpreted as melt in many forms (Rychert et al., 2020). For instance beneath mid-ocean ridges, seismic surface wave studies have interpreted a broad, hundreds of kilometers wide, melt triangle beneath the ultrafast spreading East Pacific Rise at 17 °S (Dunn & Forsyth, 2003; Forsyth et al., 1998) and the intermediate spreading Juan De Fuca Ridge (Bell et al., 2016; Gao, 2016), while other studies have imaged smaller scale and discrete melt zones beneath the slow spreading equatorial Mid-Atlantic Ridges on the order of 100-200 km wide (Harmon et al., 2020). The magnetotelluric (MT) method has typically imaged smaller and more discrete low resistivity zones interpreted as focused melt regions beneath the fast spreading East Pacific Rise at 9 °N and the ultra-slow spreading Mohns Ridge (Johansen et al., 2019; Key et al., 2013) that are typically < 100 km wide, although a broader region >200 km was inferred beneath the East Pacific Rise at 17 °S (Evans et al., 1999). Further off-axis, layered and/or pervasive melt in the asthenosphere has been inferred based on the imaging of discontinuities by scattered waves that require sharp drops in seismic velocity with depth (Kawakatsu et al., 2009; Rychert & Shearer, 2011; Rychert & Shearer, 2009; Tharimena et al., 2017). Active source seismic studies also find strong reflectors near the expected base of the tectonic plate, that have been interpreted as channelized melt (Mehouachi & Singh, 2018; Stern et al., 2015). Similar channelized structures have also been interpreted from thin low resistivity zones at 60-80 km depth (Naif et al., 2013; Wang et al., 2020). Whether or not differences among the inferred shape and location of melt are an artefact of resolution and sensitivities of the individual methodologies or representative of real Earth structure has remained unclear.

The complementary resolution and sensitivities of MT and seismic imaging techniques offer a promising means of probing Earth's physical properties to examine the thermal structure and the presence of partial melt. The Earth's mantle is primarily composed of olivine and pyroxene, and the conductivity of these minerals has a strong temperature dependence (Gardés et al., 2014; Naif et al., 2021), enhanced by the presence of conducting fluids such as partial melt (Naif et al., 2021; Ni et al., 2011) and the presence of water and other

91 crystallographic defects in the olivine mineral lattice (Gardés et al., 2014; Naif et al., 2021).
92 Water and other volatiles such as CO_2 are also thought to significantly increase the
93 conductivity of the fluid and therefore the overall conductivity of the mantle if present (Ni
94 et al., 2011; Sifre et al., 2014). On the other hand, seismic velocities are dependent on
95 temperature and pressure (e.g., Stixrude & Lithgow-Bertelloni, 2005), followed by the
96 presence of partial melt (Clark & Lesher, 2017; Hammond & Humphreys, 2000), particularly
97 for shear velocity, and are relatively insensitive to the presence of water as a
98 crystallographic defect (Abers et al., 2014) or as a component of the partial melt. These
99 differences mean that the two methods together have the potential to better constrain the
100 thermal properties of the mantle, the presence and amount of partial melt, and the amount
101 of hydration in the melt.
102
103 There have been two main approaches to cooperative or simultaneous joint inversion of
104 electromagnetic and seismic data: 1) inversion based on underlying petrophysical or
105 empirical relationships between velocity and conductivity (Abubakar et al., 2012; Jegen et
106 al., 2009; Sun & Li, 2016; Takougang et al., 2015) and 2) inversion based on model gradient
107 approaches, e.g,. forcing model changes in velocity and resistivity either in the same sense,
108 the opposite sense, or with no change in one of the models (Bennington et al., 2015;
109 Gallardo & Meju, 2004; Haber & Oldenburg, 1997; Moorkamp et al., 2011; Zhang et al.,
110 2020). The petrophysical or empirical approach requires either accurate models of the
111 physical properties of the rocks (Gardés et al., 2014), an (ideally) relatively simple system
112 that can be captured with simple linear or polynomial fits to data (Jegen et al., 2009), or a
113 guided fuzzy *c*-means clustering operator (Sun & Li, 2016), which is more likely the case in
114 locations with limited compositional and thermal variation. However, the success of this
115 approach depends strongly on the accuracy of the prior information and laboratory
116 measurements, which is specific to the area of interest since the relationship between
117 velocity and conductivity is not universal. The model gradient approach, such as cross
118 gradient (Gallardo & Meju, 2004) and normalized cross gradient (Zhang et al., 2020), on the
119 other hand, presumes that the gradient of resistivity structures is positively or negatively
120 correlated with the gradient of velocity structures. The minimization of model gradient or
121 cross-gradient can also be satisfied automatically where zero gradient is required by one or
122 both datasets. In contrast with the petrophysical approach, the gradient method is more
123 generic. The cross-gradient approach is probably more useful for detecting regions where
124 the physical and chemical properties of the Earth result in seismic and resistivity anomalies
125 that would not necessarily align. For instance, the presence of small amounts of certain
126 minerals such as magnetite in serpentine (Stesky & Brace, 1973) or graphite (Frost et al.,
127 1989) and other highly conductive minerals would generate a strong resistivity anomaly, but
128 may not be volumetrically significant enough to have a strong seismic signature. Choosing
129 between these two approaches or other approaches using Monte Carlo inversions
130 (Moorkamp et al., 2010) is dependent on the details of the particular datasets and the
131 structure involved. Combining the two approaches is possible, and exhibits enhanced
132 performance for structural similarity in the joint inversion (Colombo & Rovetta, 2018; Guo
133 et al., 2020).
134
135 The I-LAB (Imaging the Lithosphere-Asthenosphere Boundary) experiments including: 1)
136 Passive Imaging of the Lithosphere Asthenosphere Boundary (PI-LAB) experiment, 2)
137 Experiment to Unearth the Rheological Lithosphere-Asthenosphere Boundary (EURO-LAB),

138   and 3) the Central Atlantic Lithosphere-Asthenosphere Boundary (CA-LAB) experiment
139   presented a unique opportunity for interpretation of MT and seismic data in order to
140   understand the oceanic lithosphere-asthenosphere system at the equatorial Mid-Atlantic
141   Ridge. We deployed 39 ocean bottom seismometers (OBS) and 39 ocean bottom
142   magnetotelluric (OBMT) instruments on 0-80 Myr seafloor across the long offset Chain and
143   Romanche fracture zones (Agius et al., 2018; Harmon et al., 2018), allowing us to sample a
144   wide seafloor age range in one experiment. The OBS and OBMT were co-located (within 1-2
145   km), in three lines perpendicular to the ridge (Fig. 1). The experiment was designed to image
146   the uppermost mantle beneath the ridge system and examine the evolution of the oceanic
147   lithosphere-asthenosphere system and the nature of the lithosphere-asthenosphere
148   boundary.
149
150   Here we focus on two results for developing a structurally consistent model of the oceanic
151   lithosphere-asthenosphere system, the three-dimensional (3-D) shear-wave velocity model
152   from Rayleigh wave tomography and the two-dimensional (2-D) MT inversions from the two
153   southernmost lines (Fig. 1, 2). The shear velocity model images a high velocity lithosphere,
154   and several punctuated low velocity zones (<4.2 km/s) in the asthenosphere, that were
155   interpreted as melt (Harmon et al., 2020). Near the ridge axis, asthenospheric low velocity
156   zones are attributed to sub-ridge upwelling (Anomalies A and E in line I and line II,
157   respectively in Fig. 2), while further off-axis the low velocity anomalies are attributed to
158   melting due to upwelling caused by small scale convection (Anomalies B, C, and F in Fig. 2)
159   (Harmon et al., 2020; Wang et al., 2020). The MT result images similar structures to the
160   surface wave model, e.g., a high resistivity lithospheric lid ($\log_{10}(\rho)$ >2) and several low
161   resistivity anomalies ($\log_{10}(\rho)$ < 1) in the asthenosphere (anomalies A, B, C, D, E, and F in Fig.
162   2) (Wang et al., 2020). In Line I there is good agreement with the depth (50-80 km) and
163   lateral extent (~100-200 km) of the low resistivity anomaly and low seismic velocities
164   (Anomalies B and C) as well as evidence for a high resistivity, high velocity lithospheric drip
165   (anomaly D) that extended from 50 to 150 km depth. However, in line II (Fig. 2b and 2d) the
166   agreement in terms of the shapes of the anomalies is less remarkable, specifically anomaly
167   F, where the conductive anomalies suggest a channel structure < 20 km thick extending
168   from the ridge to 30 Myr seafloor, while the surface wave anomaly resembles a simple oval
169   ~200 km wide from 50-80 km depth. In addition, anomaly E is deeper in the resistivity
170   model, >100 km depth, than in the shear velocity model, where it extends from 50 to 100
171   km depth. While in line I, anomaly A is shallower at ~30 km depth and smaller, <50 km wide,
172   in the resistivity model than in the shear velocity model, where it is located at 50-80 km
173   depth and 150 km wide. In other words, while there is some similarity in the lateral
174   locations of the anomalies, the exact depth and morphologies are somewhat different.
175
176   Subsequent studies support the existence of these anomalies and suggest that apparent
177   discrepancies may be artefacts of resolution. For example, S-to-P receiver functions support
178   the existence of the anomalies (Rychert et al., 2021). The receiver functions image
179   discontinuities associated with sharp velocity decreases with depth above the locations of
180   the low shear velocity anomalies E, C, and F in the asthenosphere and also the locations
181   where the low resistivity anomalies gradually decrease with depth in the asthenosphere
182   (near anomaly E and directly beneath F) (Rychert et al., 2021). In addition, a short period
183   Rayleigh wave tomography study, which had better resolution in the upper 60 km than
184   Harmon et al. (2020), imaged a shallower anomaly for anomaly A beneath line I, more

185   consistent with the resistivity model (Saikia et al., 2021). The differences between the
186   surface wave models suggest that there are several possibilities for shear-wave velocity
187   models that will fit the Rayleigh wave data. Some of these shear velocity models could also
188   be consistent with the anomaly structure of the resistivity models as well, noting that the
189   MT method is preferentially sensitive to conductors, such as the anomalies C and F at the
190   LAB depths. Therefore, the primary motivation of this study is to find a satisfactory shear
191   velocity model that is also consistent with structural information from the resistivity models.
192
193   Here we jointly consider the Rayleigh wave phase velocities and the MT data to evaluate
194   differences and similarities between the seismic and MT anomaly structures, in particular to
195   determine an Earth structure that can satisfy both datasets within data errors. We compare
196   the models one-to-one to develop an empirical relationship between the two. Since the
197   observed relationship is very similar to laboratory-based predictions for shear velocity and
198   resistivity, we proceed using the laboratory-based relationship to translate the MT resistivity
199   to shear-wave velocity. We use the MT-derived shear velocity model as the new starting
200   model for the surface wave tomography inversion. This approach assumes that the
201   structure within the resistivity model is closer to the true earth structure, which may be the
202   case, for example, if a thin channel structure exists, which surface waves would not be able
203   to resolve without prior knowledge (e.g., Rychert et al., 2020). We discuss the validity of this
204   assumption in the discussion section. Finally, we compare the models to petrophysical
205   predictions for Earth properties in order to constrain temperature, the amount of partial
206   melt, and the amount of hydration, carbonization or sulfide weight percentage of the partial
207   melt in the asthenosphere.
208

Methods

210   MT data were inverted by Wang et al. (2020), which we briefly summarize here. The
211   determinant of the MT impedance tensor was used to invert logarithmic apparent resistivity
212   and linear phase along 2-D transects (line I and line II). The approach was chosen to
213   minimize 3-D coast effects from the nearby African coast (Wang et al., 2019). For the period
214   range chosen, 26-26,225 s, data quality and 3-D distortion were examined, and data points
215   with visible 3-D distortion were excluded in the inversions. The impedance polar diagrams of
216   the final selected data points were nearly parallel on either side of the Chain fracture zone
217   in lines I and II (Wang et al., 2020). Nevertheless, there may still be some 3-D effects due to
218   the fracture zones/3-D structure that influence the model, which we address in the
219   discussion section. Forward calculations and inversion were performed using the MAR2DEM
220   code (Key, 2016), modified to accept determinant data as an input (Wang et al., 2021).
221   Inversion of MT data with this approach is less dependent on the starting model than
222   surface wave inversion due to the diffusive nature of electromagnetic fields and the
223   smoothness and regularization of the inverse problem. Here we focus on varying the
224   starting model for the shear velocity inversion based on structural information from the
225   resistivity data, but not vice versa. We refer to this as resistivity structure guided shear
226   velocity inversion.
227
228   We first establish a relationship between shear velocity and resistivity in our study area. We
229   use two transects through the 3-D shear-wave velocity model of Harmon et al. (2020) in the
230   same locations of the two 2-D resistivity model transects of Wang et al. (2020). We make

231   cross-plots separately for the two lines. Cross-plots of the data suggest a linear relationship
232   between the two datasets, but with scatter (Fig. 3). The correlation coefficients of these
233   cross-plots for line I is 0.43±0.01 and line II 0.39±0.01. A linear regression of line I between
234   shear-wave velocity (km/s) and resistivity ($\log_{10}(\rho)$), yields a solution of Vs=4.19±0.02+
235   0.10±0.02*$\log_{10}(\rho)$, while for line II, Vs=4.22±0.02 + 0.08±0.02*$\log_{10}(\rho)$ and for line I and II
236   combined Vs=4.21±0.02+0.09±0.02*$\log_{10}(\rho)$.
237
238   We also consider predictions from laboratory petrophysical relationships between shear
239   velocity and resistivity for a half-space cooling model based on the error function solution to
240   the conductive heat transfer equation (Parker & Oldenburg, 1973; Turcotte & Schubert,
241   2002). To calculate the temperature structure, we assume a thermal diffusivity of 1×10^{-6}
242   m^2s^{-1}, and a mantle potential temperature of 1350 °C calculated for seafloor age from 0-40
243   Myr, the approximate range in of ages along lines I and II (Fig. 4). To model the predicted
244   shear velocity for a given temperature, pressure, and melt fraction we use the Very
245   Broadband Rheology calculator (Havlin et al., 2021), assuming a peridotite mantle
246   composition. We use the attenuation parameterization of (Jackson & Faul, 2010) that is
247   included in the calculator and use an average across the surface wave period range used
248   here, 18-143 s period. In this model, the addition of melt primarily affects shear velocity
249   with ~2-4% velocity reduction for 1% melt volume fraction depending on the dihedral angle
250   (Takei, 1998). The model of Takei (1998) assumes that melt is interconnected, without
251   necessarily proscribing a melt geometry. The associated predicted velocity reduction
252   depends on wetness, which is a measure of the amount of grain to grain contact relative to
253   the melt (Takei, 1998). Other models for the effect of melt on velocity exist based on
254   different assumptions of melt geometry (Clark & Lesher, 2017; Hammond & Humphreys,
255   2000; Schmeling, 1985) which we evaluate in the discussion section. For resistivity we use
256   the relationship for hydrated mantle peridotite (Gardés et al., 2014) and a model for the
257   conductivity of hydrous mantle melts (Ni et al., 2011). We then use the Hashin-Shtrikman
258   upper bound to calculate the total resistivity of a melt bearing peridotite mantle (Ni et al.,
259   2011), which, again, assumes interconnected melt. The predictions for an example case with
260   100 ppm water content in the background mantle and 1% melt in the melt triangle and
261   variable amounts of water in the melt from 4-20 weight % are shown in Figure 4. We
262   perform a linear regression on the melt-free mantle data points (black dots, Fig. 4), and find
263   a relationship of V_s = 4.14±0.02 + 0.11±0.01 *$\log_{10}(\rho)$. This relationship is very similar to the
264   one derived for the cross-plot in line I; the velocity intercept is 0.05-0.06 km/s lower than in
265   our cross-plot, and the slope is only 0.01 km/s/ $\log_{10}(\rho)$ higher than in the cross-plot. Given
266   the similarity between the two and that the petrophysical line visually fits the data from the
267   shear velocity and resistivity inversions, we opt to use the relationship from the
268   petrophysical modelling.
269
270   We use the aforementioned petrophysical relationship to translate the resistivity model
271   (Fig. 2a, b) to shear-wave velocity, creating a new starting model (Fig. 5c, d) for the shear
272   velocity inversion. We then invert the phase velocities from 18-143 s period from Harmon et
273   al. (2020) sampled along lines I and II, for shear velocity as a function of depth, sampling at
274   every point, 0.1 °. We calculate the partial derivatives relating Rayleigh wave phase velocity
275   to shear velocity using the Computer Programs in Seismology package (Herrmann, 2013),
276   and we assume a fixed Vp/Vs ratio of 1.8, which is consistent with the Preliminary Earth
277   Reference Model (PREM), a global one-dimensional (1-D) seismic velocity model

278 (Dziewonski & Anderson, 1981). We include a seawater layer along lines I and II in the
279 model based on the local bathymetry. We use a damped least-squares inversion and
280 assume an *a priori* model error of 0.2 km/s following choices from previous work (Forsyth &
281 Li, 2005; Harmon et al., 2020). We replace the upper 5 km of the model beneath the water
282 layer with average crustal values (3.5 km/s) from the 1-D model of Harmon et al. (2020). The
283 model is parameterized every 5 km in depth down to 400 km. This parameterization is finer
284 than that presented in Harmon et al. (2020) (Fig. 2). The finer parameterisation is necessary
285 to capture the smaller scale variations in the resistivity model. Therefore, we also present
286 an inversion using the 1-D starting model used in Harmon et al. (2020), but with the 5 km
287 thick layers down to 400 km depth used here for comparison purposes (Fig. 5).
288
289 We next determine the physical properties that explain the resulting anomalies including
290 temperature, melt fraction and volatile content of the melt. These quantities can trade off
291 with each other, and multiple combinations can fit the data, and therefore we use a grid
292 search approach. We presume that the thermal structure at any given point along each line
293 is similar to the predicted thermal structure from the half space cooling model calculated
294 above, but allow the effective seafloor age and corresponding thermal structure, resistivity
295 and shear velocity to vary. The effective age of the seafloor (e.g. lithospheric thickness)
296 could be greater due to "drips" (as in anomaly D) or younger if the lithosphere is thinned
297 due to upwelling. In other words, although seafloor age is known at each profile, we search
298 over the effective age of the seafloor given that our previously published models suggest
299 that the age progression of the lithosphere might not be monotonic everywhere. We
300 calculate the half space cooling thermal structure for seafloor from 0 to 40 Myr age in 1 Myr
301 intervals as described above. The thermal models have an adiabatic gradient added to them,
302 and as above, we assume a mantle potential temperature of 1350 °C. We did not vary the
303 mantle potential temperature to minimize the free parameters in the grid search. Then, for
304 each thermal structure from 0 to 40 Myr seafloor, we calculate the predicted shear velocity
305 and resistivity for melt fractions from 0.00 to 0.07 at 0.001 increments below 0.01 and 0.005
306 increment above 0.01 and melt water contents from 0 to 30 weight % in 1% increments for
307 all temperatures > 1100 °C at the corresponding depth/pressure values using the
308 relationships described above for the half-space cooling model presented in Fig 4. We then
309 examine the regions that cannot be explained by temperature alone, specifically, where the
310 shear velocity is <4.4 km/s and $\log_{10}$ resistivity is < 1.5 (< 30 Ωm), which are the nominal
311 limits of the melt free predictions of the half-space cooling model (black dots, Fig 4). We
312 perform a grid search over melt fraction, melt water content, and apparent seafloor
313 age/temperature for each point in lines I and II. We then determine the chi-squared residual
314 between the observed resistivity and shear velocity with the predicted resistivity and shear
315 velocity at the same depth in each thermal structure from 0 to 40 Myr. The chi-squared
316 residual is used to determine goodness of fit assuming an *a priori* standard deviation of 0.05
317 km/s for the shear velocity model and 0.10 $\log_{10}(\Omega m)$ for the resistivity model. A value of
318 melt, melt hydration and temperature is considered acceptable if the chi-squared value is <
319 1 for both the shear velocity and resistivity data. For most points, there are many
320 combinations of melt, melt hydration and temperature that satisfactorily fit the data. The
321 optimum value is the minimum summed value of the chi-squared values for resistivity and
322 shear velocity. We present the error as the maximum parameter value minus the minimum
323 acceptable parameter value divided by 2 for melt, melt water content and temperature,

324   which is the 95% confidence limit assuming symmetric error surfaces. We acknowledge that
325   this choice of reporting does not give a sense of the trade-offs in these parameters.
326

Results

328   The shear-wave velocity structure derived from translating the MT models to seismic
329   velocity according to the petrophysical predictions (Fig. 5c, d) closely resembles the MT
330   models (Fig. 2 a, b), which is to be expected. We impose a water layer of 0.0 km/s in the
331   model. The seismic velocities range from 4.5 km/s in the upper 20-50 km of the Earth, with a
332   minimum of 4.03 km/s associated with the lowest resistivity regions. Strong lateral
333   gradients are also visible in the starting model, with changes of 0.4 km/s over less than 50
334   km, particularly near anomaly C. The line II model has low velocity channels across the
335   transect at 20-70 km depth and several high velocity regions in the asthenospheric mantle.
336

337   When we use the shear-velocity model derived from MT (Fig. 5c, d) as the starting model for
338   the surface wave inversion we find a new shear-wave velocity model (Fig. 5 e, f) that more
339   closely resembles the MT models than the previously published model (Fig. 2). The highest
340   velocities are up to 4.81 km/s found in the fast lid, while the minimum velocity is 4.00 km/s,
341   found in anomaly B. The high velocity lid is more continuous than in the starting model but
342   follows a similar pattern of increasing thickness away from the ridges in both lines I and II. In
343   the asthenosphere, low velocity structures from the starting model are also retained.
344   Specifically, the channel structures in line II, near anomaly E and F, are retained throughout
345   much of the model, particularly in the east near anomaly F, with similar velocities (~4.0
346   km/s) to the starting model. The deep low velocity anomalies beneath anomaly E are also
347   retained in the final model. Anomaly E from the model of Harmon et al. (2020), has been
348   moved deeper into the mantle in the MT-derived starting model inversion, while anomaly F
349   has been compressed into the channel structure. In line I anomalies B and C are preserved
350   i.e., ~4.0 km/s from the starting model. Anomaly A is more pervasive beneath the ridge than
351   in the MT starting model. Anomaly D is also enhanced in the shear velocity model, with a
352   high velocity of 4.56 km/s relative to the starting model of 4.31 km/s at 100 km depth. The
353   chi-squared values indicating goodness of fit to the data are shown in Fig. 5a and 5b and are
354   ~1 or less for most of the profile indicating a fit that is within error. This goodness of fit is
355   similar to the values from Harmon et al. (2020).
356

357   When we use the 1-D starting model from Harmon et al. (2020) for the surface wave
358   inversion, and the parameterisation and damping used here we find similarities and
359   differences in comparison to Harmon et al. (2020) that illustrate the range of possible
360   models that fit the data (Fig. 5g, h). A high velocity lid is visible beneath the ridge and across
361   the region that ranges from 20-60 km in thickness. It shows low velocities beneath the ridge,
362   with a stronger and shallower low velocity region beneath the ridge than in the model of
363   Harmon et al. (2020), although in general the features are similar, and the magnitude of the
364   named velocity anomalies are similar in general, < 4.2 km/s but > 4.0 km/s. These
365   differences highlight the effect that even small changes in parameterization can have on the
366   final model. The normalized chi-squared fit to the data is shown in Fig. 5 a,b for Line I and II,
367   respectively. The chi-squared values are generally ~1 or less indicating that the model fits
368   the data within error and has a similar fit to the model with the MT starting model. It is
369   interesting to note, that prior to inversion (i.e., the 0th iteration), the fit of the 1-D model is

370 generally better than the MT starting model (chi-squared of ~2-3 vs 4-5, blue dashed v.s.
371 black dashed Fig. 5 a,b). This is likely because the 1-D model originated from the best fit 1-D
372 average seismic model for the region. Both converge to very similar final chi-squared values
373 after the iterative inversion. The new shear-velocity model with the 1-D starting model
374 presented in Figure 5g, h is primarily for comparison purposes. The goal of the paper is to
375 align the previously published shear-wave velocity and resistivity models, and so we do not
376 discuss the model of Figure 5g, h further except for the purposes of resolution discussions.
377
378 The correlation between resistivity and shear-velocity after inversion is higher when the MT
379 derived starting model is used in comparison to when the 1-D starting model is used. For
380 the 1-D starting model inversion result with the finer parameterization used here, there is a
381 slope visible in line I (Fig. 6a), but there is less of a visible relationship in line II (Fig 6b).
382 Visually, the cross-plots for the MT derived starting model inversion result are more linear,
383 with more of a slope visible in both lines I and II (Fig. 6 c, d). The correlation coefficients
384 between the resistivity model and the shear velocity model assuming 1-D starting model
385 presented here are 0.41 ± 0.01 and 0.29 ± 0.01 for lines I and II, in other words similar to that
386 between the resistivity and the original shear velocity model presented in Harmon 2020
387 above (0.43 ± 0.01 and 0.39 ± 0.01 respectively). The correlation coefficients are higher,
388 0.56 ± 0.01 and 0.62 ± 0.01 for lines I and II respectively, for the shear-wave model resulting
389 from the MT-derived starting model. With the two lines combined the correlation
390 coefficient is 0.60 ± 0.01 (Fig. 7).
391
392 We illustrate the behaviour of the effect of varying the amounts of melt and water in the
393 partial melt and compare it to the V_s and resistivity histogram for both lines I and II (Fig. 7).
394 We use the thermal structure from the half-space cooling model shown in Fig. 4 but now
395 allow partial melt at 0.1%, 1.0% and 3.0% where the mantle temperature exceeds 1100 °C.
396 We also vary the amount water in the partial melt between 4-20%. The smallest amount of
397 partial melt reduces the seismic velocity by << 1% in most cases, while the resistivity is
398 reduced by ~0.6 $\log_{10}(\Omega m)$ over the range of water contents presented here. At 1% melt the
399 shear velocity is reduced by ~2%, and the effect of increased water content is stronger,
400 reducing the resistivity up to ~1.5 $\log_{10}(\Omega m)$ at the highest water contents. Finally, at 3%
401 melt, the velocity is reduced by 4-5% and the resistivity reduction is up to ~2.1 $\log_{10}(\Omega m)$.
402 The span of partial melt and melt water contents considered here also generally spans the
403 range of most of the shear velocity and resistivity values from our inversions, i.e., the
404 petrophysical values overlie the peak in the histogram. There is a slight bias in the seismic
405 velocities with a longer tail towards higher values.
406
407 Given the good general agreement between the petrophysical modelling and the shear
408 velocity and resistivity model values, we map the amount of partial melt, water content of
409 the melt, and temperature relative to the half-space cooling model onto the transects of
410 lines I and II (Fig. 8). We only perform this mapping where shear velocity is < 4.4 km/s and
411 $\log_{10}(\rho) < 1.5$ $\log_{10}(\Omega m)$, which is the nominal lower limit of the melt free half-space cooling
412 model (Fig. 4 and Fig. 7). In line I we find partial melt contents up to 4-4.5% near anomalies
413 B and C and similar maximum values in line II for anomalies E and F. Lower values of partial
414 melt <2% are needed near anomaly A and for most of the other regions, typically requiring <
415 1%. The water content of the melts is typically < 10 weight % for most (~60 %) of the total
416 anomaly area (colored regions in Fig. 8), with the notable exception of anomaly C which

417 requires up to 24 weight % water content to account for the low resistivity found in this
418 region and anomaly B which requires up to 15 weight %. There are other smaller patches of
419 high-water content visible near the edges of some of the regions, and within the channel of
420 anomaly E. The temperature structure generally has cooler temperatures 1100-1200 °C at
421 depths < 100 km and temperatures > 1300 °C at greater depth. The grid search provides
422 formal error bounds corresponding to our presumed data errors (Fig. 9). The errors for the
423 melt percentages are typically <1%, while error for water content of the melt is on average 4
424 weight %, and the average errors for temperature are 26 °C.

Discussion

426 The inversion result from the MT derived starting model fits the phase velocity data within
427 error. Initially, the misfit of the MT-derived starting model before inversion is only a factor
428 of 2 greater than the misfit of the 1-D starting model before inversion in most places. This
429 suggests general agreement between the MT-derived starting model and the phase velocity
430 data. The inversion result using the MT-derived starting model fits the phase velocity data
431 just as well as the shear velocity model of Harmon et al. (2020) which used a 1-D starting
432 model and also the shear velocity model using the 1-D starting model and the finer
433 parameterisation presented here for comparison purposes. All of these have a normalized
434 chi-squared value 1 or less for most of the transects. The MT-derived shear-velocity model
435 improved the visual agreement and correlation coefficient between the resistivity and shear
436 velocity model.
437
438 Overall, many of the common features of the original studies are retained and several of the
439 anomalies come into better agreement. For example, the MT-derived shear velocity model
440 retains the thickening of the lithosphere and the drip feature at anomaly D observed in the
441 Harmon et al. (2020) model. The lithospheric thickening with distance from the ridge is
442 more pronounced in the MT-derived shear velocity model in comparison to that of Harmon
443 et al. (2020), presumably a result of removing the 1-D influence on the model. Anomalies B
444 and C are also retained in the MT-derived model, although anomaly B is more prominent
445 than in the Harmon et al. (2020) study. In the asthenosphere, better agreement between
446 the resistivity model and the MT-derived shear velocity model is achieved for the channel
447 features in line II associated with anomaly F. Anomaly C in the MT-derived shear velocity
448 model has a morphology more similar to the MT model than in the Harmon et al. (2020)
449 model. Other anomalies such anomaly A shifts shallower than the Harmon et al. (2020)
450 model and align better with a weak shallow anomaly directly beneath the ridge in the
451 resistivity model. Anomaly E is deeper than that in the Harmon et al. (2020) model, again in
452 better agreement with the resistivity model.
453
454 The differences in the shear velocity models here highlight some of the limitations of the
455 approach. Specifically, inversion of Rayleigh wave phase velocities for shear velocity
456 structure is non-unique, and this is well-known (Rychert et al., 2020) as many previous
457 works have demonstrated that a variety of models can fit a given dispersion curve. The
458 differences between Harmon et al. (2020) (Fig. 2), the 1-D starting model with smoothing,
459 damping, and parameterization of this study (Fig. 5e, f) and the MT-derived starting model
460 (Fig. 5g, h) illustrate this fact again and highlight that the strength of an anomaly can vary
461 from model to model depending on the starting model, even if similar damping is used and
462 the same fit is achieved as was the case here. For instance, the MT-derived shear-wave

463 velocity model includes velocities in Anomalies B, C that are up to 1% slower in comparison
464 with Harmon et al. (2020), which impacts interpretation in terms of the presence of partial
465 melt. Suitable additional constraints are needed to determine which structure is the most
466 likely, such as information from receiver functions or resistivity.
467
468 The cross-plots indicate that the shear-wave velocity model and resistivity are in good
469 agreement with the petrophysics predictions for the half-space cooling model and variable
470 partial melt concentrations and melt water contents. About 80% of the shear velocity data
471 lie within 0.1 km/s of the petrophysical predictions for reasonable temperature structure,
472 melt and melt water contents (Fig. 7). The resistivity model is completely spanned by the
473 petrophysical predictions. Shear velocity appears to be biased towards higher values, which
474 may be a result of either the inversion process or a physical process. Shear-wave velocity
475 inversions can trade off velocities at shallow depths with deeper asthenospheric anomalies,
476 by compensating low asthenospheric values with higher lithospheric values. On the other
477 hand, other physical effects such as depletion (Schutt & Lesher, 2006) of peridotite through
478 ridge melting toward more harzburgitic compositions (Hacker & Abers, 2004) could cause
479 higher velocities by ~1-2%. In addition, anisotropy could also enhance the apparent velocity
480 by up to 1-3% (Rychert & Harmon, 2017; Saikia et al., 2021). In reality, it is likely some
481 combination of these physical effects and model artefacts which are not accounted for in
482 the calculations used for predicting shear velocities.
483
484 In this work we chose to force the shear velocity structure towards a closer match to the
485 resistivity model, because the MT method has better resolution for certain features such as
486 electrically conductive thin channels, which is an assumption that is worth examination. We
487 presumed the resistivity model has better structural resolution, but this assumption has
488 limitations, since the 2-D assumption for the resistivity model may break down. For
489 instance, anomaly E is part of a larger 3-D velocity anomaly that extends to the south along
490 the Mid-Atlantic Ridge in Harmon et al. (2020), and the depth of the anomaly is much
491 greater in the resistivity anomaly, perhaps owing to issues of dimensionality. The deep
492 conductive anomalies beneath anomaly E are also poorly resolved in the MT data (Wang et
493 al., 2020), suggesting there is still some uncertainty about this particular location. Other
494 observations, such as S-to-P receiver functions, suggest there may be a shallower shear
495 velocity anomaly associated with anomaly E, which is necessary in order to produce a sharp
496 velocity contrast in these regions (Rychert et al., 2021). In addition, the lateral extent of the
497 S-to-P discontinuity agrees well with the lateral extent of the Harmon et al. (2020)
498 anomalies, suggesting this is a 3-D feature that is not well resolved in the 2-D MT inversion.
499 However, given that we prefer the MT-derived shear-wave velocity structure for some of
500 the major anomalies (A, B, C, D, and F), we proceed by interpreting our estimates for mantle
501 melting and melt water content, bearing the limitations of the inversions in mind.
502
503 The thermal structure predicted from our grid search (Fig. 8e, f) suggests relatively warm
504 temperatures beneath Anomalies B and C as well as the deeper parts of E (>1300 °C), while
505 Anomalies A and F have relatively low temperatures (1100-1200°C). This variability is likely a
506 result of the pressure dependence of the seismic waves. The low temperatures are generally
507 consistent with the interpretation that the shallow anomalies, particularly the channel
508 structures in F, are interacting with the base of the lithosphere (Harmon et al., 2020; Wang

509   et al., 2020). The deeper, hotter anomalies (anomaly B and C) are also generally consistent
510   with the interpretation of upwelling from depth associated with small scale convection.
511
512   The predicted melt fractions are in general agreement with our previous work from the
513   region, considering the various assumptions. Our melt fraction of up to 0.04 agrees with the
514   0.01 – 0.07 values previously reported based on the resistivity model alone (Wang et al.,
515   2020). It is higher than the 0.005 to 0.015 reported by the previous shear-wave velocity
516   model (Harmon et al., 2020). However, this can be explained by two main differences: 1)
517   The anomalies in the new shear velocity model presented here are up to 1 % slower than
518   those of the previous study (Harmon et al., 2020) and 2) We used the Takei (1998)
519   relationship between melt and velocity here, which corresponds to about a 2 % velocity
520   reduction for 0.01 melt fraction in comparison to the 7.9% reduction for 0.01 melt fraction
521   from the work of (Hammond & Humphreys, 2000) used by Harmon et al. (2020). Our melt
522   fraction result of up to 0.04 is also consistent with the 6 – 11 % velocity drop with depth
523   required by receiver functions after correcting for the maximum effect of temperature
524   (Rychert et al., 2021), which would require melt fractions of 0.03 – 0.06 assuming the same
525   melt-velocity relationship from Takei (1998) that we used here.
526
527   A different parameterization choice for the effects of melt on velocity due to different
528   assumptions on the melt geometry could yield lower melt fraction requirements by the
529   seismic constraints and still satisfy the resistivity model. The Takei (1998) model used here
530   presumes interconnected melt, which permits current flow and affects resistivity, but does
531   not prescribe a specific melt geometry rather using "wetness" and dihedral angle.
532   Unconnected melt geometries such as isolated pockets (Schmeling, 1985) do not affect
533   resistivity and so we can rule those out (Naif et al., 2021). Assuming interconnected films
534   and organized cuspate tubules (Hammond & Humphreys, 2000), as used in Harmon et al.
535   (2020), reduces the maximum amount of partial melt fraction to < 0.02. Melt in the form of
536   interconnected tubules and cuspate geometries (Hammond & Humphreys, 2000), which
537   have a velocity reduction of 14.5% per 0.01 melt fraction would suggest even lower melt
538   fractions (< 0.01). Resistivity does not depend strongly on the geometry of connected melt.
539   This is mostly due to the fact that the greatest resistivity reduction occurs at melt fractions <
540   0.03, with a more gradual reduction in resistivity at higher melt fractions (Fig. 10). However,
541   since resistivity also has a strong dependence on the volatile content in the melt, the lower
542   melt fractions predicted for the interconnected tubules and cuspate geometries could also
543   satisfy the resistivity anomalies with additional volatiles. More work would be required to
544   determine the most likely partial melt geometry and relationship for shear velocity
545   reduction to place better constraints on the 3-fold variation predicted from differing
546   assumptions.
547
548   Predicted water contents are typically < 10 weight % for the melt but are surprisingly high,
549   up to weight 24%, in the centers of anomalies C and F. Simple fractional or batch melting
550   calculations suggest that for a typical MORB mantle source with 100 ppm and an average
551   6% melting of the mantle water contents of the melt should be ~0.2 weight % (Workman &
552   Hart, 2005). Higher water melt contents are possible for low degrees of partial melting, for
553   example <0.005 melt fraction yields > 1% weight water for 100 ppm in the mantle source,
554   and >7% weight water for 800 ppm in the mantle source. One possible explanation is that
555   these off-axis anomalies represent coalesced low-degree melts of a moderately wet mantle

556   with high water content. There is some geochemical evidence for a moderately wet mantle
557   from basalts collected from the ridge segments in the study area, with estimated water
558   contents that range from 110-770 ppm (~ 0.01-0.08 weight %) for the mantle source (Le
559   Voyer et al., 2015). The advantage of this model in which volatile rich melts coalesce is that
560   wet melts are stable and can persist in the mantle for long periods of time as has been
561   suggested as an explanation for inferred melt channels beneath older oceanic lithosphere
562   imaged by active source imaging (Mehouachi & Singh, 2018).
563
564   High CO_2 in the mantle melts instead of high-water content is another possible explanation
565   for the low resistivities observed in region (Sifre et al., 2014). Carbonated peridotite is
566   thought to exist in the mantle, although the abundance of carbon is relatively low, likely <
567   100 ppm, as it is present in ancillary phases, rather than being hosted in olivine or pyroxene
568   (Dasgupta & Hirschmann, 2010). Carbonated melts are generated and stable at greater
569   depths, and only small degrees of partial melt are likely to be generated (<0.001 melt
570   fraction) (Dasgupta & Hirschmann, 2010; Hirschmann, 2010). However, the melts could
571   percolate upwards and coalesce, generating higher CO_2 contents in the melt (Hirschmann,
572   2010). Fig. 10 shows the trade off in effective resistivity for 1 weight % water in the melt,
573   and 10% and 30% CO_2 by weight in the melt as a function of disequilibrium melt fraction
574   assuming 100 ppm in the un-melted mantle background for a depth of 80 km and a
575   temperature of 1350 °C. The figure is for demonstrative purposes since; (i) melt fraction is
576   imposed rather than generated using batch melting or fractional melting, (ii) we did not vary
577   temperature as we did in the silicate case, and (iii) the melt may not necessarily be stable.
578   At 30% CO_2 weight percent the resistivity is similar to the high-water content (20 weight %)
579   case. However, geochemical estimates of CO_2 in the primary ridge basalts range from 104
580   ppm to 1.9 weight % (Le Voyer et al., 2019), which is much lower than the >30 CO_2 weight %
581   needed to explain our results. To reach our high values, again aggregation of extremely low
582   degree partial melts would be required, and this also cannot be the melt that directly erupts
583   at the ridge.
584
585   Another possible explanation for the observed anomalies besides high water contents
586   (>10%) is sulfide melts, which are extremely conductive, >10^4 S/m (Ducea & Park, 2000).
587   Small amounts of sulfide melts can rapidly reduce the effective resistivity of the aggregate.
588   To illustrate this we follow the parameterization of Ducea and Park (Ducea & Park, 2000),
589   using the (Gardés et al., 2014) parameterizations for the solid olivine and the Ni et al. (2011)
590   parameterization for the silicate melt. We assume a conductivity of 10^4 S/m for sulfide
591   melts. Fig. 10 shows a comparison between the effective resistivity for an olivine matrix
592   with wet disequilibrium melts and also for sulfide/wet disequilibrium melt mixtures with
593   predominately sulfide melt. Like the CO_2 case, this is for demonstrative purposes, without
594   varying a full suite of parameters. A nearly pure sulfide melt has a similar resistivity as a
595   silicate melt with 20% water, reaching values below 1 Ωm at < 0.01 melt fraction. So, in this
596   case, regions of high melt water contents in Fig. 8, e.g., anomaly C, could also be regions of
597   high sulfide melt content. Given the bulk abundance of sulphur measured in basaltic glasses
598   in the region typically < 0.1 weight % (Le Voyer et al., 2015) and in <0.3 weight % in
599   xenoliths from continents (Ducea & Park, 2000), it is unlikely that 0.04-0.05 sulfide melt
600   fraction exists in the mantle. However, a more conservative sulfide melt fraction of ~0.01
601   could at least partially explain anomaly C (Hammond & Humphreys, 2000). There is also
602   some evidence that melts from the nearby ridge segments are sulphur saturated (Le Voyer

et al., 2015), and this may therefore suggest that sulfide melts may exist in higher abundance away from the ridge melt triangle where silicate melts are in high abundance. Sulfide melts have also been proposed to explain low seismic wave speeds in the asthenosphere (Helffrich et al., 2011). Further work is needed to test whether sulfide melts would be compatible with small scale convection and explain our off-axis anomalies, as they have a higher density than silicate melts.

The melt anomalies inferred here extend to the base of our well-resolved region, ~150 km depth, which is greater than the 60 – 80 km predictions of a dry melting curve (Katz et al., 2003). This suggests that water or CO_2 induced melting is occurring at depth or the presence of sulfide melts or some combinations are active to produce melts so deep. In addition, the largest melt fractions are associated with anomalies B, C, E and F, which are far from the ridge axis. This suggests melt generation occurs away from the ridge either owing to off-axis small scale upwellings, the presence of volatiles, or the combination of the two. Persistent melt near the base of the lithosphere and apparent channelization near anomaly F also suggests a role for water or other volatiles in the melts in order to stabilize them at relatively cool temperatures near the base of the lithosphere (Mehouachi & Singh, 2018).

Our joint seismic-MT constraints require melt fractions (> 0.01) over large swaths of the asthenosphere mantle, several hundred kilometers, and hundreds of kilometers off the ridge axis. Such high percentages are not expected to persist over time and length scales that would enable seismic imaging (Spiegelman & Elliott, 1993). For instance, melt fractions > 0.01 could be explained by a lack of a drainage route for the melt. Melt may coalesce at a permeability boundary at the lithosphere-asthenosphere boundary, as suggested by recent numerical models that include 2-phase flow (Sim et al., 2020). Asthenospheric porosity in these models at a given snapshot in time can reach up to 10-20%, which could explain our melt fraction observations in the channels (Sim et al., 2020). The melt may also reduce the asthenospheric viscosity (Hirth & Kohlstedt, 1995; Jackson et al., 2006) potentially further promoting small scale convection.

Partial melt is inferred in different geometries beneath the two different ridge segments in our study region, including punctuated anomalous regions impinging on the base of the plate, channels of melt beneath the plate, and punctuated regions at deeper depth. Interestingly, the deeper melt exists at depths greater than predicted from simple adiabatic upwelling models and are completely disconnected from the shallower melts. The variable geometries may be related to the 3-D nature of the study area, and may also suggest that we are imaging different stages in melt generation and migration, rather than steady-state equilibrium melt. Our observations in light of these geodynamic models suggests that melt is dynamic but may be persistent on geological timescales.

Conclusions

We developed a simple relationship for shear velocity and resistivity of the oceanic lithosphere and asthenosphere that can be used to initialize these quantities for joint inversions based on data from the I-LAB experiments and petrophysical modelling. We used the relationship to create a shear-wave starting model that we used to re-invert the phase velocities. The new shear-wave velocity model more closely resembles the resistivity models, in particular by including a low velocity channel and also in terms of the location

and shape of slow velocity anomalies. The apparent lithospheric drip was also enhanced. Overall, the correlation between the surface wave and MT datasets increased. This suggests that apparent discrepancies between the original models are more likely an artefact of resolution and inversion schemes. Surface waves cannot resolve thin channel structures unless significant prior knowledge is used in the starting model in the inversion. We also demonstrate the utility for shear velocity inversion guided by resistivity structure for mantle melting and thermal structure based on petrophysical modelling. We show that shear velocity can place good constraints on melt volume, while resistivity can place good constraints on melt water content, CO_2 content or presence of sulfide melt given a simple thermal structure such as the half-space cooling model.

Our estimates of melt, melt water content and temperature are in general reasonable and within the expectations given geochemical outputs from the nearby ridge segments. The one exception is very high water or CO_2 contents (>15%) estimated in the slowest and least resistive anomalies. These high melt water or CO_2 contents could be real but would require coalescing low degree partial melts of moderately wet or carbon-rich mantle sources. Alternatively, nearly pure sulfide melts at small fractions could potentially partially explain these anomalies. Overall, joint interpretation and/or inversion of resistivity and shear velocity models holds promise for resolving debates about the lithosphere-asthenosphere system and the presence and character of partial melt in the mantle.

Acknowledgements

Figure 1. Map of the PI-LAB study region (Harmon et al., 2020). Circles indicate stations for seismic and MT locations with names indicated. MT stations are within 1-2 km of the seismic stations. Bold black lines I and II indicate transects used in this study. Background colors indicate bathymetry (Smith & Sandwell, 1997), white contours indicate seafloor age from Seton et al., (2020), and thick, dark grey line indicates the location of the Mid-Atlantic Ridge. Red box in inset map indicates study area.
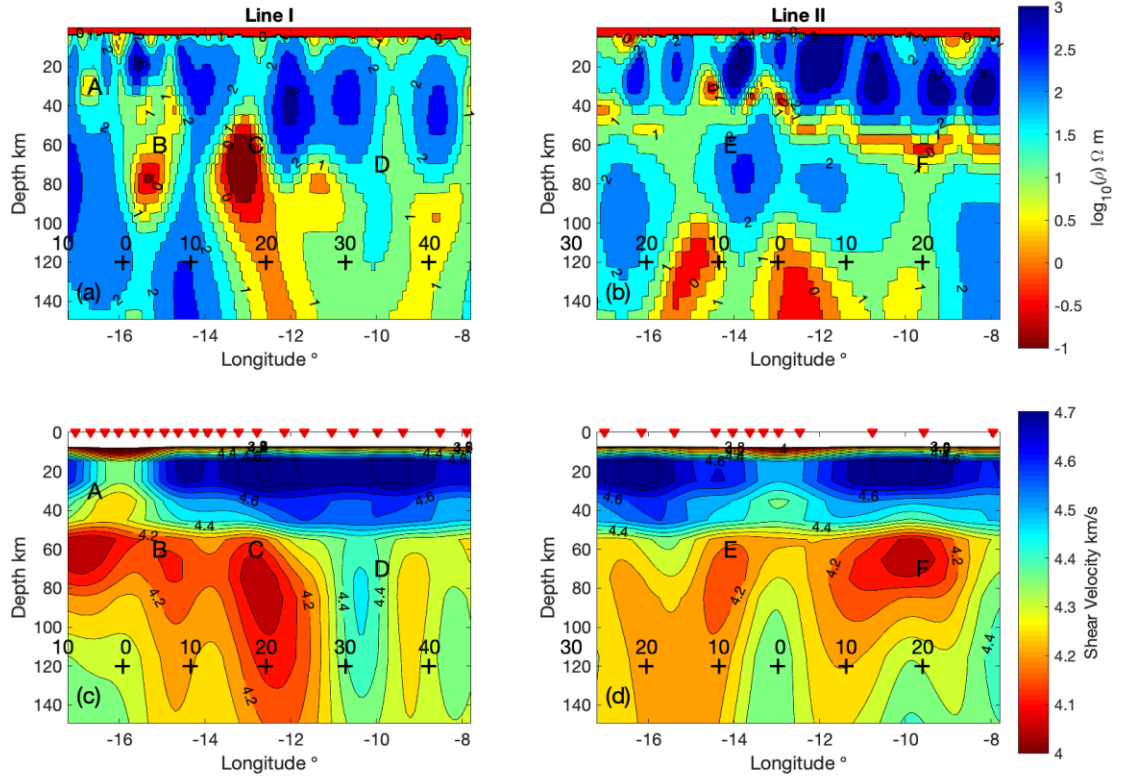
693
694 **Figure 2. Resistivity model and shear-wave velocity model from previous work**. Panels (a)
695 and (b) show contoured resistivity transects from line I and line II, respectively, from Wang
696 et al. (2020). Contour interval is 0.5 log units. Panels (c) and (d) show contoured shear
697 velocity transects for line I and II, respectively, from Harmon et al. (2020). Contour interval
698 is 0.05 km/s. Anomalies A, B, C, D, E and F from Harmon et al. (2020) are indicated. Red
699 triangles show seismic/MT station locations along the lines. Crosses at 120 km depth in plots
700 indicate the seafloor age, in Myr, with 0 indicating the ridge location.
701
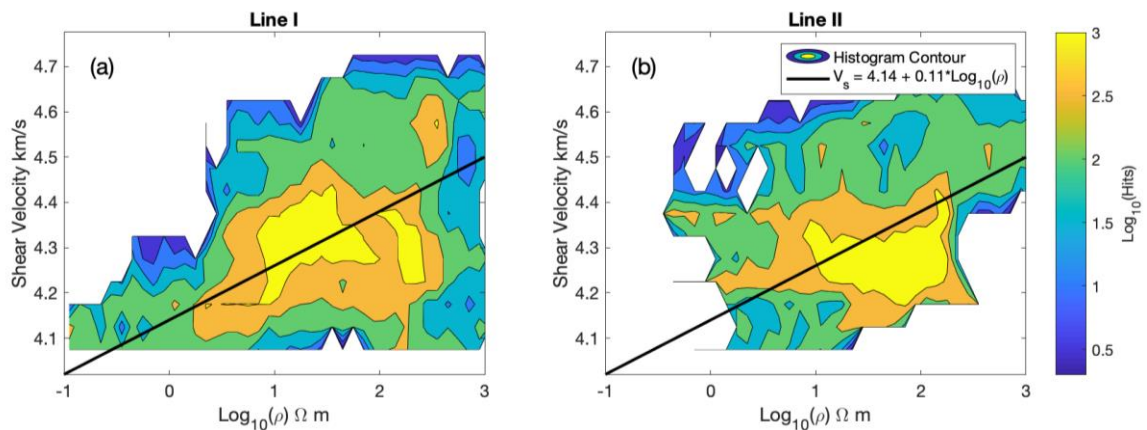
702
703 **Figure 3. Cross-plot histograms of resistivity and shear-wave velocity from previous work.**
704 Panels a and b shows the histograms for line I and line II, respectively. Black line indicates
705 preferred linear relationship from petrophysical modelling shown in Figure 4.
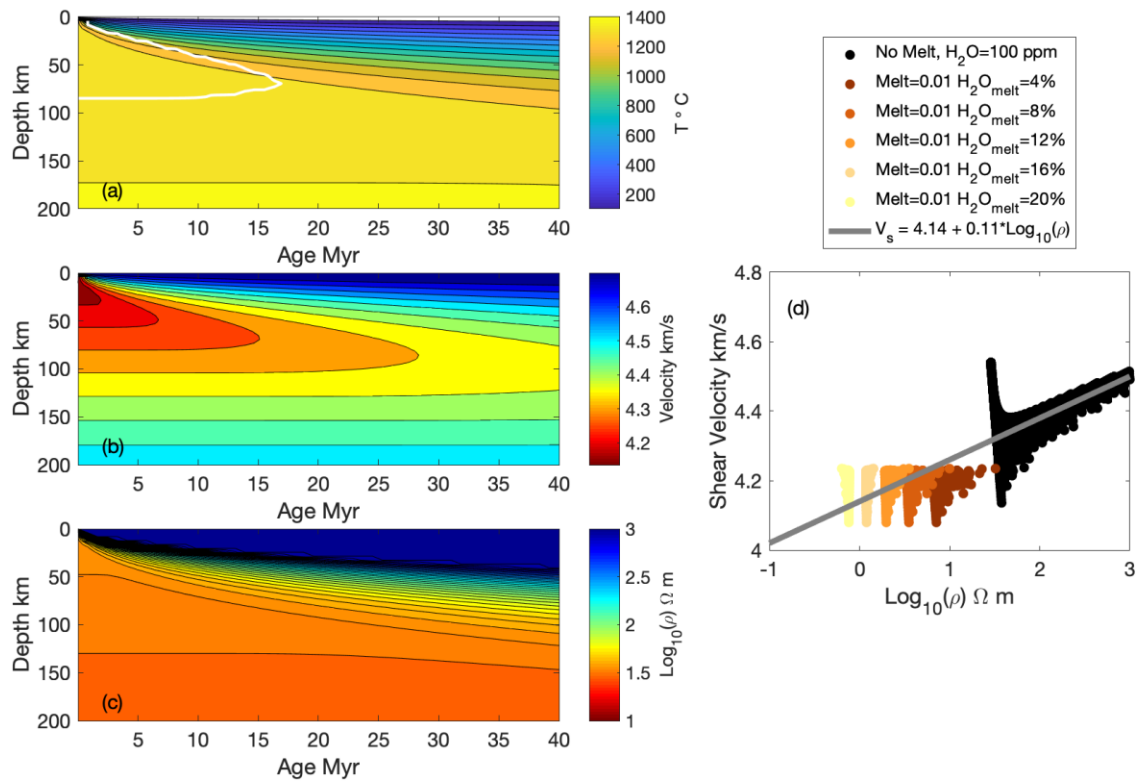
706
707 **Figure 4. Petrophysical predictions for resistivity and shear-wave velocity for half-space**
708 **cooling model.** Panel a shows the thermal structure for the half-space cooling model, b
709 shows the predicted shear-wave velocity structure, and c shows the predicted resistivity
710 structure predicted from petrophysics calculated as described in the text. White line in
711 panel a indicates the predicted melt triangle for 100 ppm water in a background mantle
712 (Katz et al., 2003). Panel d shows the cross-plot of predicted resistivity and shear velocity
713 without melt from panel b and c (black circles) and with a presumed melt fraction (0.01)
714 containing different amounts of water (4-20%), within the predicted melt triangle (yellow
715 and brown circles). Grey line in Panel d shows preferred linear relationship between
716 resistivity and shear velocity based petrophysical modelling presented here and consistent
717 with the cross-plot histograms presented in Figure 3.
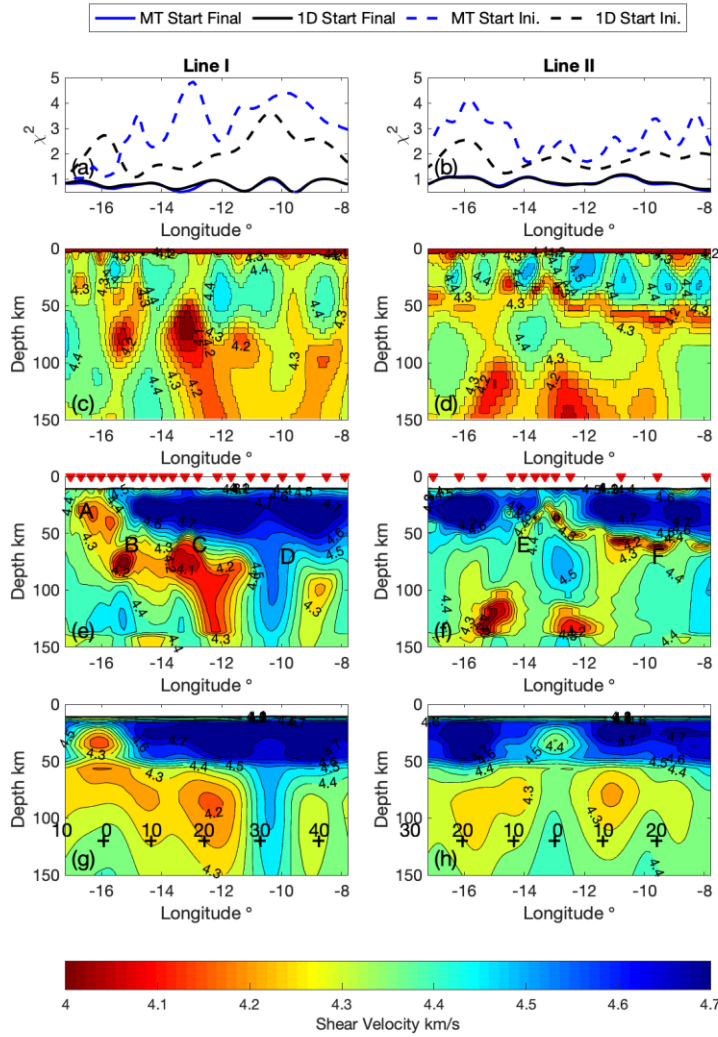
Figure 5. Shear-wave velocity inversions based on resistivity predictions. Panels a and b show misfit along lines I and II using normalized chi-squared. Panels c and d show the shear-wave velocity models that result from translating the resistivity model shown in Fig. 2 to velocity using the linear relationships based on petrophysical modelling. Panels e and f show the shear-wave velocity inversion results using panels c and d, respectively, as starting models. Red triangles show seismic/MT station locations along the lines. Panels g and h the show shear velocity inversion results using the 1-D starting model from Harmon et al. (2020) and the smoothing, damping, and model parameterisation used here. The model shown in panels g and h is for comparison purposes. It is different than that shown in Figure 2 panels c and d from Harmon et al. (2020) because we used a finer depth parameterisation here in order to capture the fine scale structure of the resistivity model. Contour interval is 0.05 km/s. Asthenospheric anomalies A, B, C, D, E and F from Harmon et al. (2020) and Wang et al. (2020) are shown for reference. Crosses at 120 km depth in plots indicate the seafloor age, in Myr, with 0 indicating the ridge location.

19
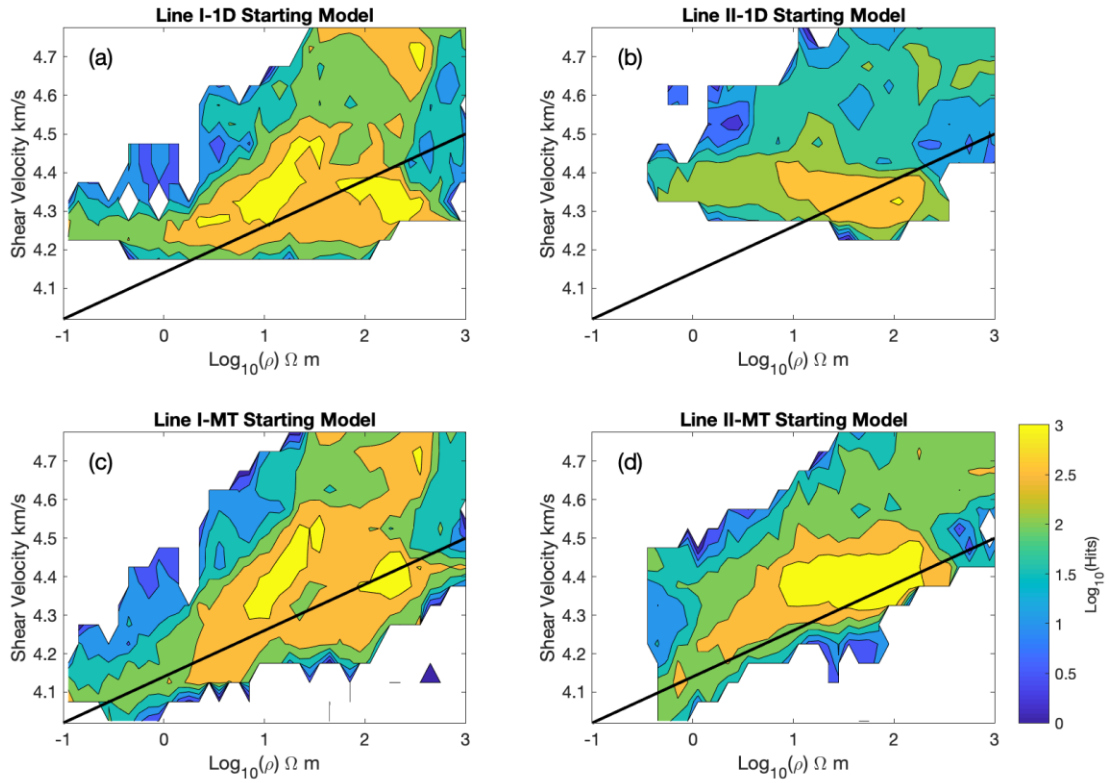
733
734 **Figure 6. Cross-plot histograms of resistivity and shear-wave velocity models.** Panels a and
735 b show the cross-plot histograms for line I and line II, respectively, for the shear-wave
736 velocity model derived from using the 1-D velocity starting model from Harmon et al. (2020)
737 and the damping, smoothing, and parameterisation used here (Figure 5g, h). Panels c and d
738 show the cross-plot histograms of the MT-derived shear-wave velocity model (Fig. 5 e, f).
739 Black line indicates preferred linear relationship from petrophysical modelling shown in
740 Figure 4.
741
742

743
744

Legend:
- Data Distribution
- No Melt, H_2O=100 ppm
- Melt=0.001 H_2O_{melt}=4%
- Melt=0.001 H_2O_{melt}=8%
- Melt=0.001 H_2O_{melt}=12%
- Melt=0.001 H_2O_{melt}=16%
- Melt=0.001 H_2O_{melt}=20%
- Melt=0.01 H_2O_{melt}=4%
- Melt=0.01 H_2O_{melt}=8%
- Melt=0.01 H_2O_{melt}=12%
- Melt=0.01 H_2O_{melt}=16%
- Melt=0.01 H_2O_{melt}=20%
- Melt=0.03 H_2O_{melt}=4%
- Melt=0.03 H_2O_{melt}=8%
- Melt=0.03 H_2O_{melt}=12%
- Melt=0.03 H_2O_{melt}=16%
- Melt=0.03 H_2O_{melt}=20%
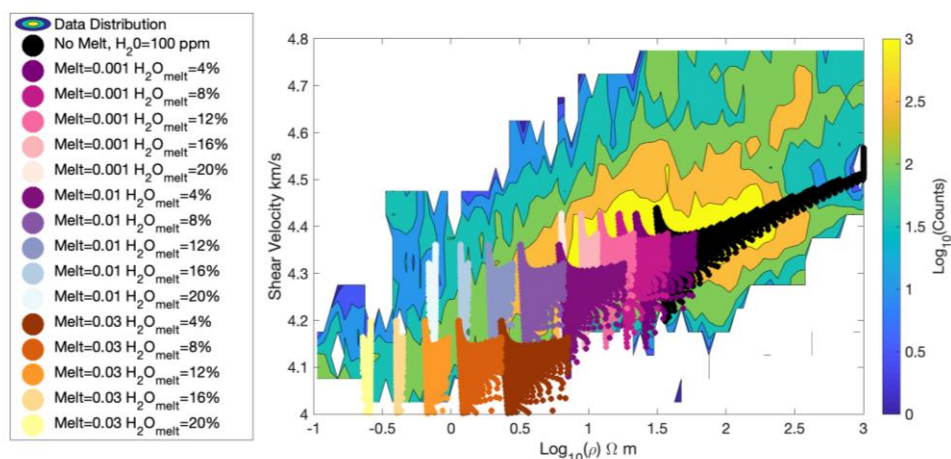
745
746 **Figure 7. Cross-plot histogram of resistivity and shear-wave velocity from the MT-derived**
747 **shear-wave velocity model for both lines I II and petrophysical predictions.** Purely thermal
748 predictions are shown as black dots. Colored dots show predictions for various melt
749 fractions and melt water contents. Legend indicates the amount of imposed disequilibrium
750 melt fraction (0.001, 0.01 and 0.03) and water content of the melt in weight % (4-20%).
751
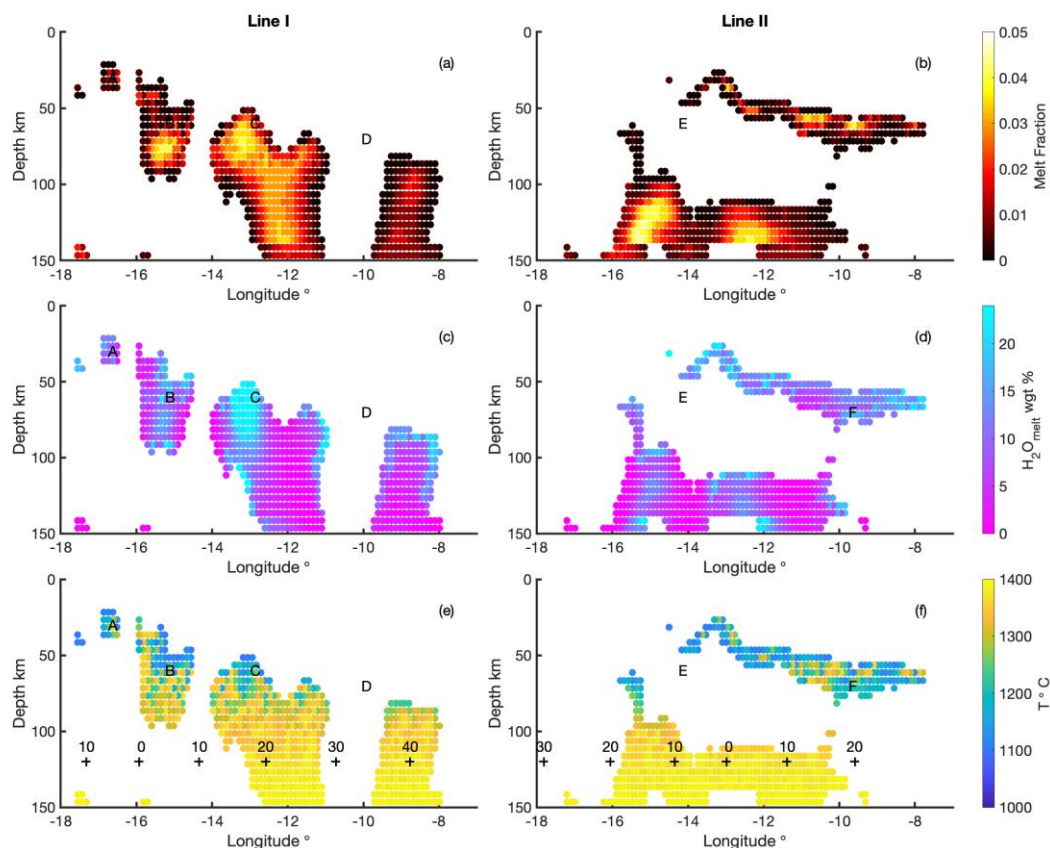
Figure 8. Results of grid search for partial melt, melt water content and mantle temperature. Panels a and b show results for partial melt fraction, panels c and d show water content of the partial melt, and panels e and f show the result for temperature for lines I and II, respectively. Anomalies A, B, C, D, E and F are plotted at the same locations as in Figure 2 for reference. Crosses at 120 km depth in plots indicate the seafloor age, in Myr, with 0 indicating the ridge location.
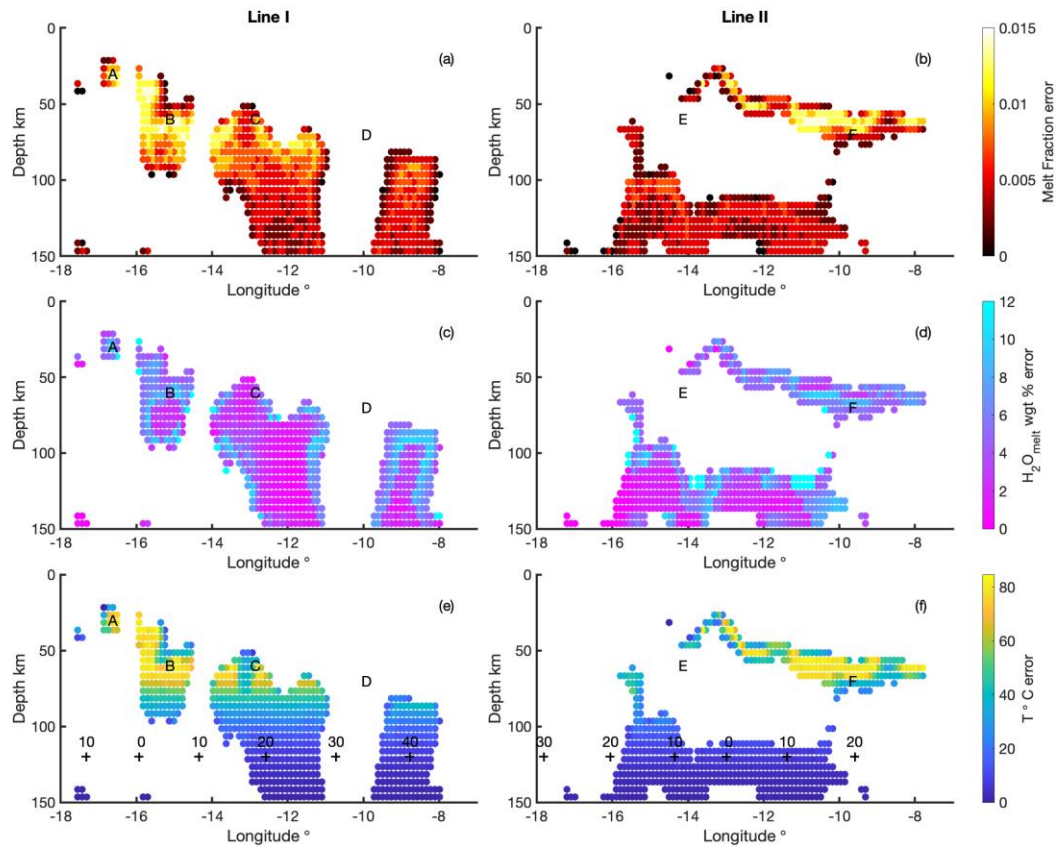
Figure 9. Error estimates of grid search for partial melt, water content of the melt and mantle temperature. Panels a and b show partial melt fraction error, panels c and d show water content of the partial melt error, and panels e and f show temperature error for line I and line II, respectively. Anomalies A, B, C, D, E and F are plotted at the same locations as in Figure 2 for reference. Crosses at 120 km depth in plots indicate the seafloor age, in Myr, with 0 indicating the ridge location.

23
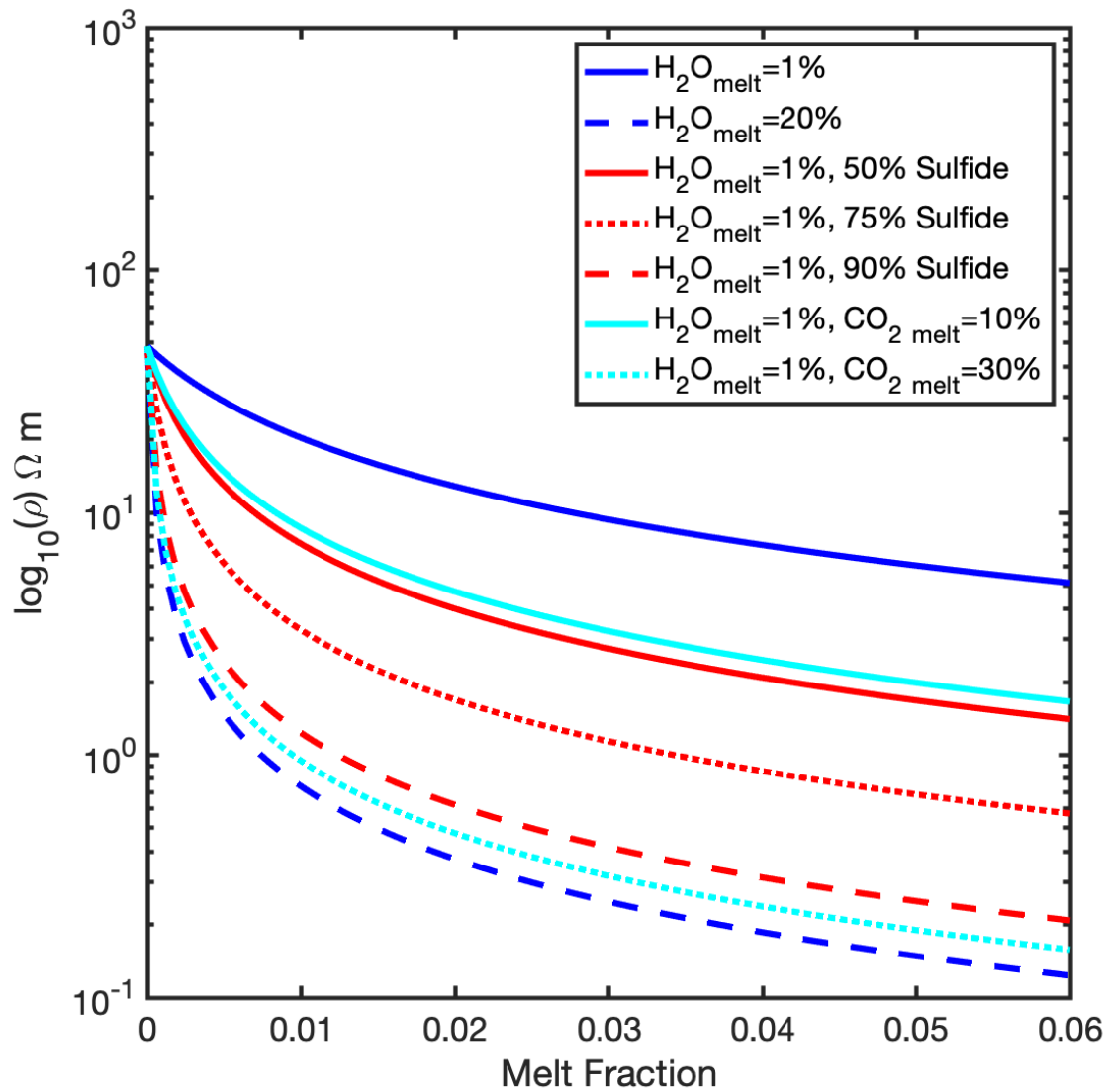
771
772 **Figure 10. Effective resistivity predictions for water, CO_2 and sulfide in silicate melts as a**
773 **function of melt fraction.** We assume a solid mantle with 100 ppm water and disequilibrium
774 melt at 1300°C. Legend indicates the respective water, CO_2 and sulfide concentrations. Melt
775 with water only is shown as blue lines. Melt that includes water and CO_2 is shown as cyan
776 lines and melt that includes sulfide is shown as red lines.
777

References

Abers, G. A., Fischer, K. M., Hirth, G., Wiens, D. A., Plank, T., Holtzman, B. K., . . . Gazel, E. (2014). Reconciling mantle attenuation-temperature relationships from seismology, petrology, and laboratory measurements. *Geochemistry Geophysics Geosystems, 15*(9), 3521-3542. doi:10.1002/2014gc005444

Abubakar, A., Gao, G., Habashy, T. M., & Liu, J. (2012). Joint inversion approaches for geophysical electromagnetic and elastic full-waveform data. *Inverse Problems, 28*(5). doi:Artn 055016

10.1088/0266-5611/28/5/055016

Agius, M., Harmon, N., Rychert, C. A., Tharimena, S., & Kendall, J. M. (2018). Sediment Characterization at the Equatorial Mid‐Atlantic Ridge From P‐to‐S Teleseismic Phase Conversions Recorded on the PI‐LAB Experiment. *Geophysical Research Letters, 45*. doi:https://doi.org/10.1029/2018GL080565

Anderson, D., & Sammis, C. (1970). Partial Melting in the Upper Mantle. *Phys. Earth Plan. Interiors, 3*, 41-50.

Baba, K., Chave, A. D., Evans, R. L., Hirth, G., & Mackie, R. L. (2006). Mantle dynamics beneath the East Pacific Rise at 17 degrees S: Insights from the Mantle Electromagnetic and Tomography (MELT) experiment. *J. Geophys. Res., 111*(B02101). doi:doi:10.1029/2004JB003598

Bell, S., Ruan, Y. Y., & Forsyth, D. W. (2016). Ridge asymmetry and deep aqueous alteration at the trench observed from Rayleigh wave tomography of the Juan de Fuca plate. *Journal of Geophysical Research-Solid Earth, 121*(10), 7298-7321. doi:10.1002/2016jb012990

Bennington, N. L., Zhang, H. J., Thurber, C. H., & Bedrosian, P. A. (2015). Joint Inversion of Seismic and Magnetotelluric Data in the Parkfield Region of California Using the Normalized Cross-Gradient Constraint. *Pure and Applied Geophysics, 172*(5), 1033-1052. doi:10.1007/s00024-014-1002-9

Clark, A. N., & Lesher, C. E. (2017). Elastic properties of silicate melts: Implications for low velocity zones at the lithosphere-asthenosphere boundary. *Science Advances, 3*(12).

Colombo, D., & Rovetta, D. (2018). Coupling strategies in multiparameter geophysical joint inversion. *Geophys. J. Int., 215*(2), 1171-1184. doi:10.1093/gji/ggy341

Dasgupta, R., & Hirschmann, M. M. (2010). The deep carbon cycle and melting in Earth's interior. *Earth and Planetary Science Letters, 298*(1-2), 1-13. doi:10.1016/j.epsl.2010.06.039

Ducea, M. N., & Park, S. K. (2000). Enhanced Mantle Conductivity from Sulfide Minerals, Southern Sierra Nevada, California. *Geophysical Research Letters, 27*(16), 2405-2408. doi:https://doi.org/10.1029/2000GL011565

Dunn, R. A., & Forsyth, D. W. (2003). Imaging the transition between the region of mantle melt generation and the crustal magma chamber beneath the southern East Pacific Rise with short-period Love waves. *J. Geophys. Res., 108*(B7), 2352, doi:2310.1029/2002JB002217.

Dziewonski, A. M., & Anderson, D. L. (1981). Preliminary Reference Earth Model. *Phys. Earth Plan. Inter., 25*(4), 297-356.

Evans, R. L., Tarits, P., Chave, A. D., White, A., Heinson, G., Filloux, J. H., . . . Unsworth, M. J. (1999). Asymmetric electrical structure in the mantle beneath the East Pacific Rise at 17 degrees S. *Science, 286*(5440), 752-756.

824   Forsyth, D. W., & Li, A. (2005). Array-analysis of two-dimensional variations in surface wave
825   phase velocity and azimuthal anisotropy in the presence of multipathing
826   interference. In A. Levander & G. Nolet (Eds.), *Seismic Data Analysis and Imaging*
827   *with Global and Local Arrays*: AGU Monograph.
828   Forsyth, D. W., Scheirer, D. S., Webb, S. C., Dorman, L. M., Orcutt, J. A., Harding, A. J., . . .
829   Team, M. S. (1998). Imaging the deep seismic structure beneath a mid-ocean ridge:
830   The MELT experiment. *Science, 280*(5367), 1215-1218.
831   Frost, B. R., Fyfe, W. S., Tazaki, K., & Chan, T. (1989). Grain-Boundary Graphite in Rocks and
832   Implications for High Electrical-Conductivity in the Lower Crust. *Nature, 340*(6229),
833   134-136. doi:DOI 10.1038/340134a0
834   Gaherty, J. B., Jordan, T. H., & Gee, L. S. (1996). Seismic structure of the upper mantle in a
835   central Pacific corridor. *J. Geophys. Res., 101*(B10), 22,291-222,309.
836   Gallardo, L. A., & Meju, M. A. (2004). Joint two-dimensional DC resistivity and seismic travel
837   time inversion with cross-gradients constraints. *Journal of Geophysical Research-*
838   *Solid Earth, 109*(B3). doi:Artn B03311
839   10.1029/2003jb002716
840   Gao, H. Y. (2016). Seismic velocity structure of the Juan de Fuca and Gorda plates revealed
841   by a joint inversion of ambient noise and regional earthquakes. *Geophysical*
842   *Research Letters, 43*(10), 5194-5201. doi:10.1002/2016gl069381
843   Gardés, E., Gaillard, F., & Tarits, P. (2014). Toward a unified hydrous olivine electrical
844   conductivity law. *Geochemistry, Geophysics, Geosystems, 15*(12), 4984-5000.
845   doi:https://doi.org/10.1002/2014GC005496
846   Goetze, C., Poirier, J. P., Kelly, A., Cook, A. H., & Greenwood, G. W. (1978). The mechanisms
847   of creep in olivine. *Philosophical Transactions of the Royal Society of London. Series*
848   *A, Mathematical and Physical Sciences, 288*(1350), 99-119.
849   doi:doi:10.1098/rsta.1978.0008
850   Guo, R., Yao, H. M., Li, M., Ng, M. K. P., Jiang, L., & Abubakar, A. (2020). Joint Inversion of
851   Audio-Magnetotelluric and Seismic Travel Time Data With Deep Learning Constraint.
852   *IEEE Transactions on Geoscience and Remote Sensing*, 1-14.
853   doi:10.1109/TGRS.2020.3032743
854   Haber, E., & Oldenburg, D. (1997). Joint inversion: a structural approach. *Inverse Problems,*
855   *13*(1), 63-77. doi:10.1088/0266-5611/13/1/006
856   Hacker, B. R., & Abers, G. A. (2004). Subduction Factory 3: An Excel worksheet and macro for
857   calculating the densities, seismic wave speeds, and H2O contents of minerals and
858   rocks at pressure and temperature. *Geochemistry Geophysics Geosystems, 5*.
859   doi:Artn Q01005,10.1029/2003gc000614
860   Hammond, W. C., & Humphreys, E. D. (2000). Upper mantle seismic wave velocity: Effects of
861   realistic partial melt geometries. *J. Geophys. Res., 105*(5), 10975-10986.
862   Harmon, N., Rychert, C., Kendall, J., Tharimena, S., Bogiatzis, P., & Agius, M. (2020).
863   Evolution of the oceanic Lithosphere in the equatorial Atlantic from Rayleigh Wave
864   tomography, evidence for small-scale convection from the PI-LAB experiment.
865   *Geochem. Geophys. Geosys.*(10.1029/2020GC009174). doi:10.1029/2020GC009174
866   Harmon, N., Rychert, C. A., Agius, M. R., Tharimena, S., Le Bas, T. P., Kendall, J. M., &
867   Constable, S. (2018). Marine geophysical investigation of the Chain Fracture Zone in
868   the equatorial Atlantic from the PI-LAB Experiment. *J. Geophys. Res., 123*, 11,016–
869   011,030. doi:10.1029/2018JB015982

870 Havlin, C., Holtzman, B., & Hopper, E. (2021). Inference of thermodynamic state in the
871   asthenosphere from anelastic properties, with applications to north American upper
872   mantle. *Phys. Earth Plan. Inter.*, 106639.
873   doi:https://doi.org/10.1016/j.pepi.2020.106639
874 Helffrich, G., Kendall, J.-M., Hammond, J. O. S., & Carroll, M. R. (2011). Sulfide melts and
875   long-term low seismic wavespeeds in lithospheric and asthenospheric mantle.
876   *Geophysical Research Letters, 38*(11). doi:https://doi.org/10.1029/2011GL047126
877 Herrmann, R. B. (2013). Computer Programs in Seismology: An Evolving Tool for Instruction
878   and Research. *Seismological Research Letters, 84*(6), 1081-1088.
879   doi:10.1785/0220110096
880 Hirschmann, M. M. (2010). Partial melt in the oceanic low velocity zone. *Phys. Earth Planet.*
881   *Inter., 179*(1-2), 60-71. doi:10.1016/J.Pepi.2009.12.003
882 Hirth, G., & Kohlstedt, D. L. (1995). Experimental Constraints on the Dynamics of the
883   Partially Molten Upper-Mantle .2. Deformation in the Dislocation Creep Regime. *J.*
884   *Geophys. Res., 100*(8), 15441-15449.
885 Jackson, I., & Faul, U. H. (2010). Grainsize-sensitive viscoelastic relaxation in olivine: Towards
886   a robust laboratory-based model for seismological application. *Phys. Earth Plan.*
887   *Inter., 183*(1-2), 151-163. doi:Doi 10.1016/J.Pepi.2010.09.005
888 Jackson, I., Faul, U. H., Fitz Gerald, J. D., & Morris, S. J. S. (2006). Contrasting viscoelastic
889   behavior of melt-free and melt-bearing olivine: Implications for the nature of grain-
890   boundary sliding. *Materials Science and Engineering a-Structural Materials*
891   *Properties Microstructure and Processing, 442*(1-2), 170-174. doi:Doi
892   10.1016/J.Msea.2006.01.136
893 Jegen, M. D., Hobbs, R. W., Tarits, P., & Chave, A. (2009). Joint inversion of marine
894   magnetotelluric and gravity data incorporating seismic constraints: Preliminary
895   results of sub-basalt imaging off the Faroe Shelf. *Earth and Planetary Science Letters,*
896   *282*(1), 47-55. doi:https://doi.org/10.1016/j.epsl.2009.02.018
897 Johansen, S. E., Panzner, M., Mittet, R., Amundsen, H. E. F., Lim, A., Vik, E., . . . Arntsen, B.
898   (2019). Deep electrical imaging of the ultraslow-spreading Mohns Ridge. *Nature,*
899   *567*(7748), 379-+. doi:10.1038/s41586-019-1010-0
900 Karato, S.-I. (2012). On the origin of the asthenosphere. *Earth Planet. Sci. Lett., 321-322*, 95-
901   103.
902 Katz, R. F., Spiegelman, M., & Langmuir, C. H. (2003). A new parameterization of hydrous
903   mantle melting. *Geochem. Geophys. Geosys., 4*(9), 1073,
904   doi:1010.1029/2002GC000433.
905 Kawakatsu, H., Kumar, P., Takei, Y., Shinohara, M., Kanazawa, T., Araki, E., & Suyehiro, K.
906   (2009). Seismic Evidence for Sharp Lithosphere-Asthenosphere Boundaries of
907   Oceanic Plates. *Science, 324*(5926), 499-502. doi:10.1126/science.1169499
908 Key, K. (2016). MARE2DEM: a 2-D inversion code for controlled-source electromagnetic and
909   magnetotelluric data. *Geophys. J. Int., 207*(1), 571-588. doi:10.1093/gji/ggw290
910 Key, K., Constable, S., Liu, L., & Pommier, A. (2013). Electical image of passive mantle
911   upwelling beneath the northern East Pacific Rise. *Nature, 495*, 499-502.
912   doi:doi:10.1038/nature11932
913 Le Voyer, M., Cottrell, E., Kelley, K. A., Brounce, M., & Hauri, E. H. (2015). The effect of
914   primary versus secondary processes on the volatile content of MORB glasses: An
915   example from the equatorial Mid-Atlantic Ridge (5 degrees N-3 degrees S). *Journal of*
916   *Geophysical Research-Solid Earth, 120*(1), 125-144. doi:10.1002/2014jb011160

917 Le Voyer, M., Hauri, E. H., Cottrell, E., Kelley, K. A., Salters, V. J. M., Langmuir, C. H., . . . Füri,
918   E. (2019). Carbon Fluxes and Primary Magma CO2 Contents Along the Global Mid-
919   Ocean Ridge System. *Geochemistry, Geophysics, Geosystems, 20*(3), 1387-1424.
920   doi:https://doi.org/10.1029/2018GC007630
921 McKenzie, D., & Parker, R. (1967). The North Pacific: an Example of Tectonics on a Sphere.
922   *Nature, 216*, 1276–1280 doi:https://doi.org/10.1038/2161276a0
923 Mehouachi, F., & Singh, S. (2018). Water-rich sublithospheric melt channel in the equatorial
924   Atlantic Ocean. *Nat. Geosci., 11*, 65-69. doi:doi:10.1038/s41561-017-0034-z
925 Moorkamp, M., Heincke, B., Jegen, M., Roberts, A. W., & Hobbs, R. W. (2011). A framework
926   for 3-D joint inversion of MT, gravity and seismic refraction data. *Geophys. J. Int.,*
927   *184*(1), 477-493. doi:10.1111/j.1365-246X.2010.04856.x
928 Moorkamp, M., Jones, A. G., & Fishwick, S. (2010). Joint inversion of receiver functions,
929   surface wave dispersion, and magnetotelluric data. *Journal of Geophysical Research-*
930   *Solid Earth, 115*. doi:Artn B04318
931 10.1029/2009jb006369
932 Naif, S., Key, K., Constable, S., & Evans, R. L. (2013). Melt-rich channel observed at the
933   lithosphere-asthenosphere boundary. *Nature, 495*(7441), 356-359.
934 Naif, S., Selway, K., Murphy, B. S., Egbert, G., & Pommier, A. (2021). Electrical conductivity of
935   the lithosphere-asthenosphere system. *Phys. Earth Plan. Inter., 313*, 106661.
936   doi:https://doi.org/10.1016/j.pepi.2021.106661
937 Ni, H., Keppler, H., & Behrens, H. (2011). Electrical conductivity of hydrous basaltic melts:
938   implications for partial melting in the upper mantle. *Contributions to Mineralogy and*
939   *Petrology, 162*(3), 637-650. doi:10.1007/s00410-011-0617-4
940 Parker, R. L., & Oldenburg, D. W. (1973). Thermal Model of Ocean Ridges. *Nature, Phys. Sci.,*
941   *242*(122), 137-139.
942 Priestley, K., & McKenzie, D. (2006). The thermal structure of the lithosphere from shear
943   wave velocities. *Earth Plan. Sci. Lett., 244*(1-2), 285-301.
944 Rychert, C. A., Fischer, K. M., & Rondenay, S. (2005). A sharp lithosphere-asthenosphere
945   boundary imaged beneath eastern North America. *Nature, 436*(7050), 542-545.
946 Rychert, C. A., & Harmon, N. (2017). Constraints on the anisotropic contributions to velocity
947   discontinuities at ~60 km depth beneath the Pacific. *Geochemistry, Geophysics,*
948   *Geosystems, 18*. doi:10.1002/2017GC006850
949 Rychert, C. A., Harmon, N., Constable, S., & Wang, S. G. (2020). The Nature of the
950   Lithosphere-Asthenosphere Boundary. *Journal of Geophysical Research-Solid Earth,*
951   *125*(10). doi:ARTN e2018JB016463
952 10.1029/2018JB016463
953 Rychert, C. A., Harmon, N., & Tharimena, S. (2018). Seismic Imaging of the Base of the
954   Ocean Plates. In H. Yuan (Ed.), *Lithospheric Discontinuities* (pp. 71-87). Washington
955   DC: AGU Monographs.
956 Rychert, C. A., Kendall, J. M., & Harmon, N. (2016). *Passive Imaging of the Lithosphere-*
957   *Asthenosphere Boundary*.
958 Rychert, C. A., Rondenay, S., & Fischer, K. M. (2007). P-to-S and S-to-P imaging of a sharp
959   lithosphere-asthenosphere boundary beneath eastern North America. *Journal of*
960   *Geophysical Research-Solid Earth, 112*(B8). doi:Artn B08314
961 10.1029/2006jb004619

962 Rychert, C. A., & Shearer, P. (2011). Imaging the lithosphere-asthenosphere boundary
963   beneath the Pacific using SS waveform modeling. *J. Geophys. Res., 116*(7), B07307,
964   doi:07310.01029/02010JB008070.
965 Rychert, C. A., & Shearer, P. M. (2009). A Global View of the Lithosphere-Asthenosphere
966   Boundary. *Science, 324*(5926), 495-498. doi:Doi 10.1126/Science.1169754
967 Rychert, C. A., Tharimena, S., Harmon, N., Wang, S., Constable, S., Kendall, J. M., . . .
968   Schlaphorst, D. (2021). A dynamic lithosphere-asthenosphere boundary near the
969   equatorial Mid-Atlantic Ridge. *Earth and Planetary Science Letters, 566*, 116949.
970   doi:https://doi.org/10.1016/j.epsl.2021.116949
971 Saikia, U., Rychert, C., Harmon, N., & Kendall, J. M. (2021). Upper Mantle Anisotropic Shear
972   Velocity Structure at the Equatorial Mid-Atlantic Ridge Constrained by Rayleigh
973   Wave Group Velocity Analysis From the PI-LAB Experiment. *Geochemistry,*
974   *Geophysics, Geosystems, 22*(3), e2020GC009495.
975   doi:https://doi.org/10.1029/2020GC009495
976 Schmeling, H. (1985). Numerical-Models on the Influence of Partial Melt on Elastic, Anelastic
977   and Electric Properties of Rocks .1. Elasticity and Anelasticity. *Phys. Earth Plan. Inter.,*
978   *41*(1), 34-57. doi:Doi 10.1016/0031-9201(85)90100-1
979 Schmerr, N. (2012). The Gutenberg Discontinuity: Melt at the Lithosphere-Asthenosphere
980   Boundary. *Science, 335*(6075), 1480-1483. doi:10.1126/science.1215433
981 Schutt, D. L., & Lesher, C. E. (2006). Effects of melt depletion on the density and seismic
982   velocity of garnet and spinel lherzolite. *Journal of Geophysical Research-Solid Earth,*
983   *111*(B5), doi:10.1029/2003JB002950. doi:Artn B05401
984 Doi 10.1029/2003jb002950
985 Seton, M., Müller, R. D., Zahirovic, S., Williams, S., Wright, N. M., Cannon, J., . . . McGirr, R.
986   (2020). A Global Data Set of Present-Day Oceanic Crustal Age and Seafloor Spreading
987   Parameters. *Geochemistry, Geophysics, Geosystems, 21*(10), e2020GC009214.
988   doi:https://doi.org/10.1029/2020GC009214
989 Sifre, D., Gardes, E., Massuyeau, M., Hashim, L., Hier-Majumder, S., & Gaillard, F. (2014).
990   Electrical conductivity during incipient melting in the oceanic low-velocity zone.
991   *Nature, 509*(7498), 81-+. doi:10.1038/nature13245
992 Sim, S. J., Spiegelman, M., Stegman, D. R., & Wilson, C. (2020). The influence of spreading
993   rate and permeability on melt focusing beneath mid-ocean ridges. *Phys. Earth Plan.*
994   *Inter., 304*. doi:ARTN 106486
995 10.1016/j.pepi.2020.106486
996 Smith, W. H. F., & Sandwell, D. T. (1997). Global seafloor topography from satellite altimetry
997   and ship soundings. *Science, 277*, 1957-1962.
998 Spiegelman, M., & Elliott, T. (1993). Consequences of melt transport for uranium series
999   disequilibrium in young lavas. *Earth and Planetary Science Letters, 118*(1), 1-20.
1000   doi:https://doi.org/10.1016/0012-821X(93)90155-3
1001 Stern, T. A., Henrys, S. A., Okaya, D., Louie, J. N., Savage, M. K., Lamb, S., . . . Iwasaki, T.
1002   (2015). A seismic reflection image for the base of a tectonic plate. *Nature,*
1003   *518*(7537), 85-+. doi:10.1038/nature14146
1004 Stesky, R. M., & Brace, W. F. (1973). Electrical Conductivity of Serpentinized Rocks to 6
1005   Kilobars. *Journal of Geophysical Research, 78*(32), 7614-7621. doi:DOI
1006   10.1029/JB078i032p07614

1007 Stixrude, L., & Lithgow-Bertelloni, C. (2005). Mineralogy and elasticity of the oceanic upper
1008   mantle: Origin of the low-velocity zone. *J. Geophys. Res., 110*, B03204,
1009   03210.01029/02004JB002965.
1010 Sun, J., & Li, Y. (2016). Joint inversion of multiple geophysical and petrophysical data using
1011   generalized fuzzy clustering algorithms. *Geophys. J. Int., 208*(2), 1201-1216.
1012   doi:10.1093/gji/ggw442
1013 Takei, Y. (1998). Constitutive mechanical relations of solid-liquid composites in terms of
1014   grain-boundary contiguity. *Journal of Geophysical Research: Solid Earth, 103*(B8),
1015   18183-18203. doi:https://doi.org/10.1029/98JB01489
1016 Takougang, E. M. T., Harris, B., Kepic, A., & Le, C. V. A. (2015). Cooperative joint inversion of
1017   3D seismic and magnetotelluric data: With application in a mineral province.
1018   *Geophysics, 80*(4), R175-R187. doi:10.1190/geo2014-0252.1
1019 Tan, Y., & Helmberger, D. V. (2007). Trans-Pacific upper mantle shear velocity structure. *J.*
1020   *Geophys. Res., 112*(8), B08301, doi:08310.01029/02006JB004853.
1021 Tharimena, S., Rychert, C., Harmon, N., & White, P. (2017). Imaging Pacific lithosphere
1022   seismic discontinuitiesInsights from SS precursor modeling. *Journal of Geophysical*
1023   *Research-Solid Earth, 122*(3), 2131-2152. doi:10.1002/2016jb013526
1024 Turcotte, D., & Schubert, G. (2002). *Geodynamics* (2nd ed.). Cambridge: Cambridge
1025   University Press.
1026 Wang, S., Constable, S., Reyes-Ortega, V., Jahandari, H., Farquharson, C., & Esquivel, T. A.
1027   (2021). Two-dimensional determinant inversion of marine magnetotelluric data and
1028   a field example from the Gulf of California, Mexico. *Geophysics, 86*(1), E37-E57.
1029   doi:10.1190/geo2019-0735.1
1030 Wang, S. G., Constable, S., Reyes-Ortega, V., & Rychert, C. A. (2019). A newly distinguished
1031   marine magnetotelluric coast effect sensitive to the lithosphere-asthenosphere
1032   boundary. *Geophys. J. Int., 218*(2), 978-987. doi:10.1093/gji/ggz202
1033 Wang, S. G., Constable, S., Rychert, C. A., & Harmon, N. (2020). A Lithosphere-
1034   Asthenosphere Boundary and Partial Melt Estimated Using Marine Magnetotelluric
1035   Data at the Central Middle Atlantic Ridge. *Geochemistry Geophysics Geosystems,*
1036   *21*(9). doi:ARTN e2020GC009177
1037 10.1029/2020GC009177
1038 Workman, R. K., & Hart, S. R. (2005). Major and trace element composition of the depleted
1039   MORB mantle (DMM). *Earth and Planetary Science Letters, 231*(1-2), 53-72.
1040   doi:10.1016/j.epsl.2004.12.005
1041 Yamauchi, H., & Takei, Y. (2016). Polycrystal anelasticity at near-solidus temperatures.
1042   *Journal of Geophysical Research, 121*(11), 7790-7820. doi:10.1002/2016jb013316
1043 Zhang, R., Li, T., Deng, X., Huang, X., & Pak, Y. (2020). Two-dimensional data-space joint
1044   inversion of magnetotelluric, gravity, magnetic and seismic data with cross-gradient
1045   constraints. *Geophysical Prospecting, 68*(2), 721-731.
1046   doi:https://doi.org/10.1111/1365-2478.12858
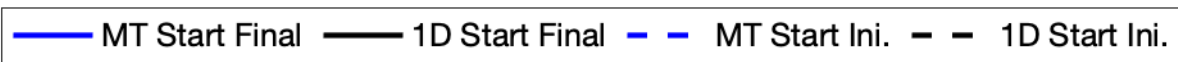1047

Figure 1.

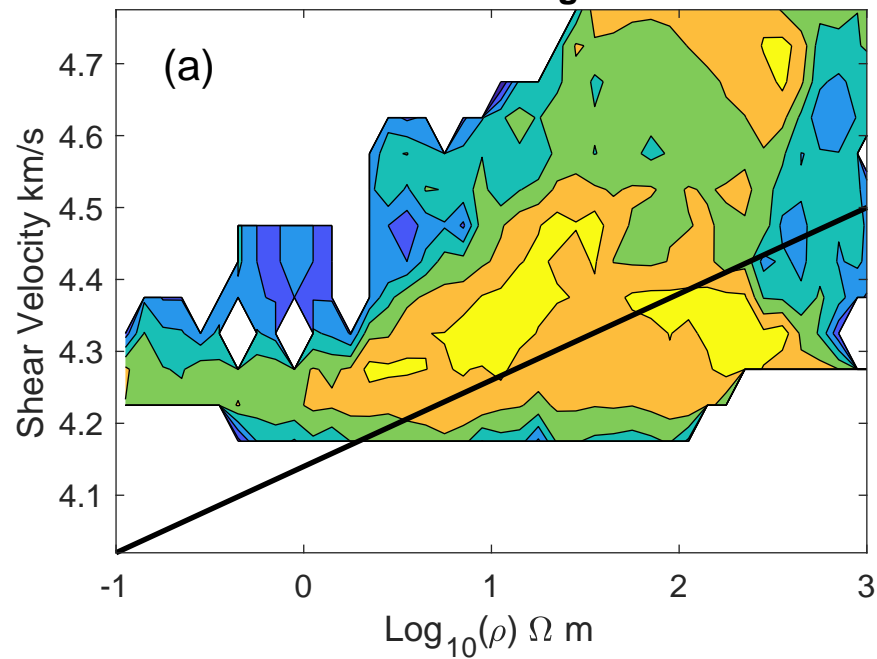**Figure 2.**

Figure 3.

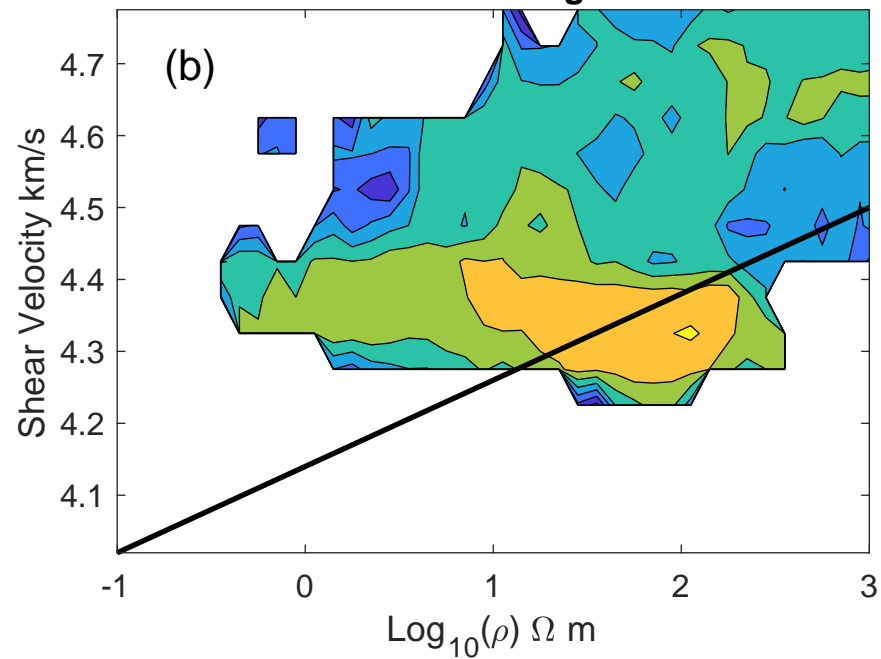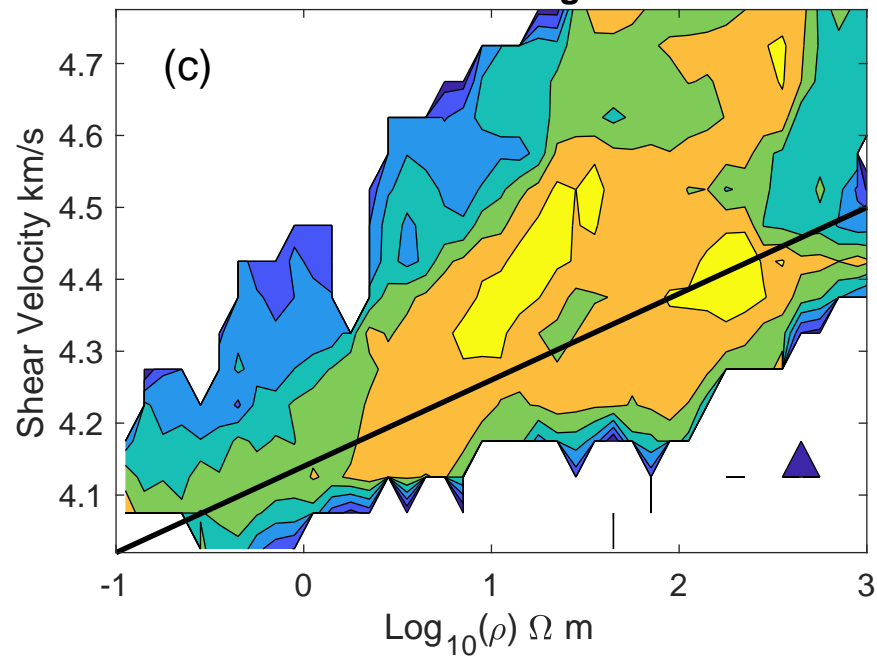**Figure 4.**
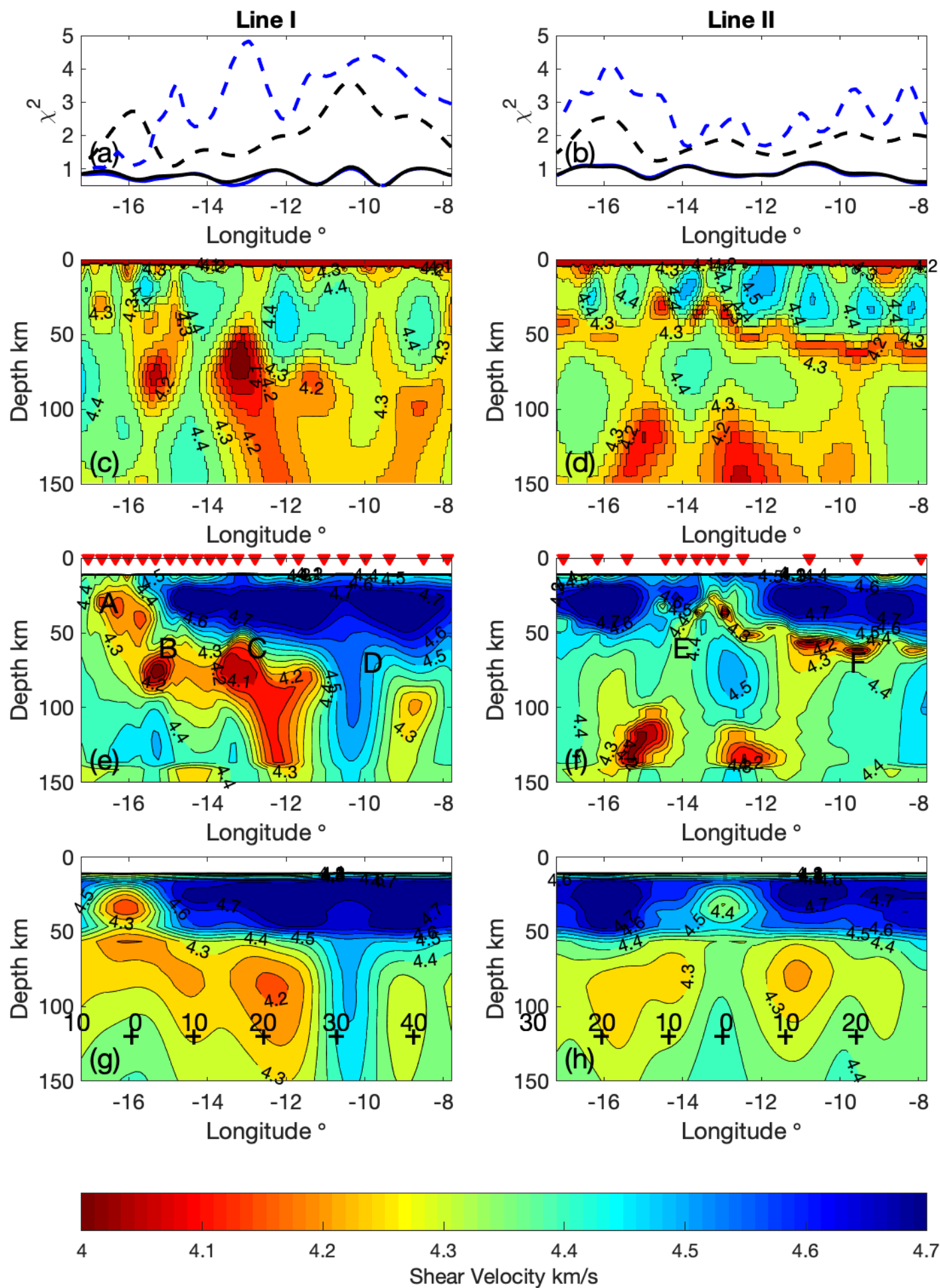
Figure 5.

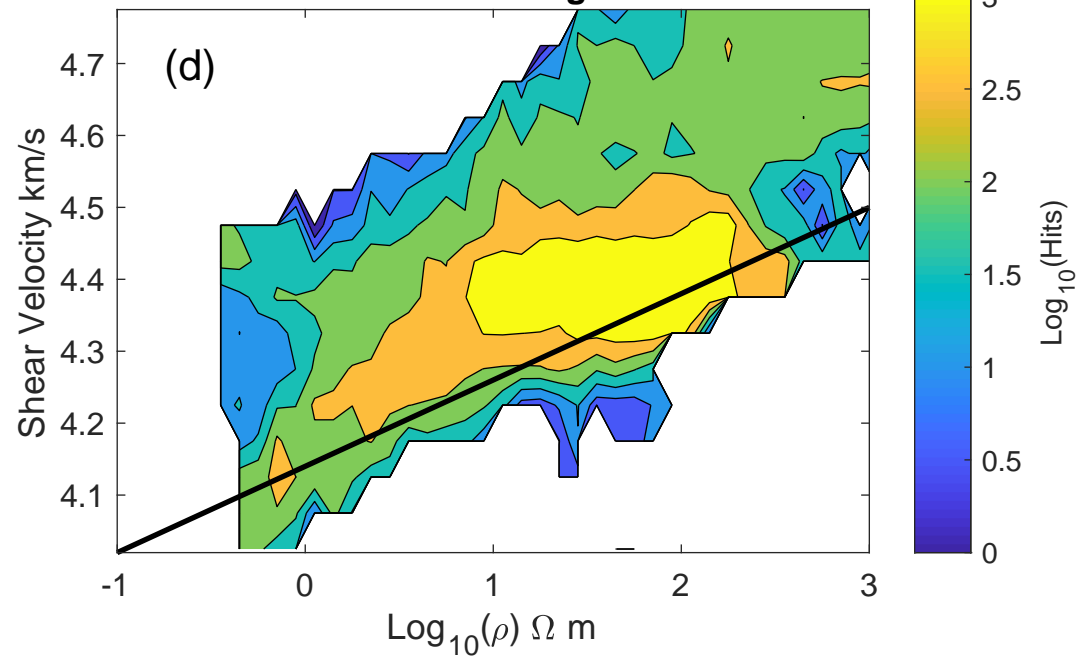**Figure 6.**

Line I-1D Starting Model (a)

Line II-1D Starting Model (b)

Line I-MT Starting Model (c)
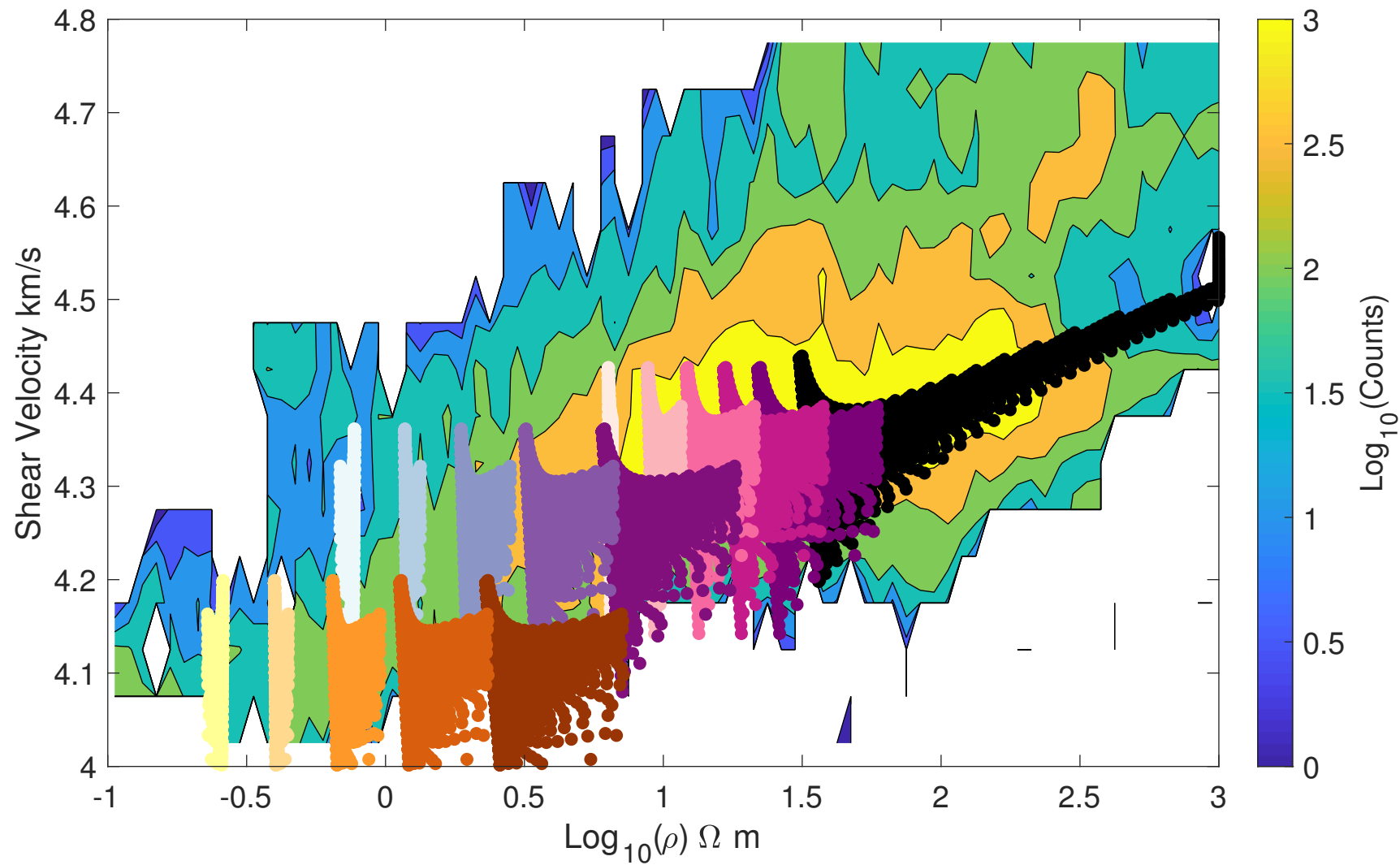
Line II-MT Starting Model (d)

Figure 7.

Figure 8.

Figure 9.

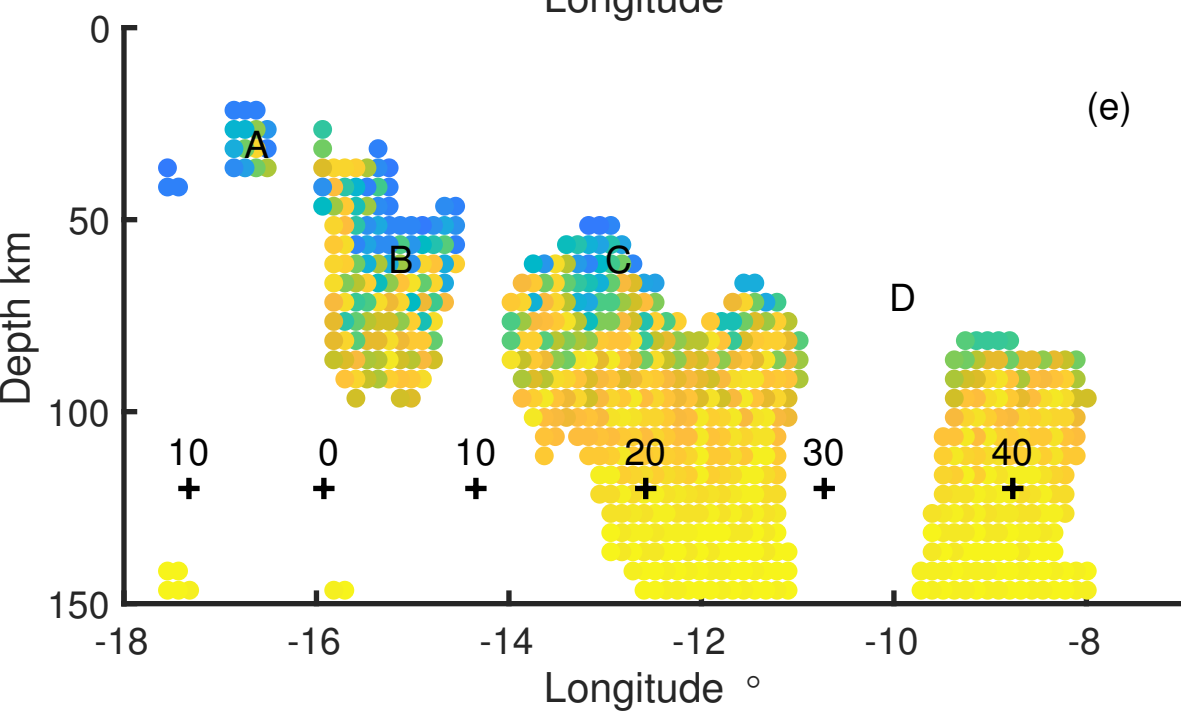Line I   Line II

Figure 10.