

1 **Supplementary information to “*Serpentinized peridotite versus thick mafic***
2 ***crust at the Romanche oceanic transform fault*” by Emma Gregory et al.**

3 This document contains the details of methods used in this study (p. 1-7), supplementary table 1 (p. 6),
4 a list of the references cited in this document (p. 7-10) and supplementary figures S1-S9 along with
5 their captions (p. 11-19).

6 **METHODS**

7 **OBS data**

8 Along the 400 km-long Romanche section of the seismic profile, 28 OBS recorded a total of 1334 shots
9 fired by an airgun array, with a spacing of 300 m. The airgun array consisted of 16 GGun airguns in
10 two strings, with a total volume of 4990 in³ (82 L), towed at a depth of 10 m and fired at a pressure of
11 140 bars. Each OBS was equipped with a hydrophone and 3-component geophone, and were spaced
12 ~14.2 km apart, except for where bathymetric depth did not allow this. The sampling rate was 4 ms and
13 the OBS data are generally of high quality (Supplementary Fig. S1). OBSs were relocated to their
14 seafloor positions using direct water wave picks and a robust least-squares inversion method. Phase
15 identification was aided by processing: predictive deconvolution, bandpass filtering between 4-18 Hz,
16 and a static correction to remove the topographic signature. Travel-times were then manually picked on
17 bandpass-filtered (4-18 Hz) hydrophone data, giving a total of 3127 Pg picks, 4259 Pn picks and 1987
18 PmP picks. Travel-time errors were assigned based on OBS relocation error, sampling interval and pick
19 uncertainty, and range from 20-82 ms (average 30 ms for Pg; 47 ms for Pn; 51 ms for PmP).

20 **Tomographic modelling**

21 The travel-time dataset was jointly inverted for velocity and a reflection boundary (the Moho) using the
22 method of Van Avendonk et al. (1998, 2004). Here, forward ray-tracing is solved using the shortest-
23 path method, and the inversion problem uses a least-squares approach to update the velocity model. The
24 starting velocity model was created as described as in Marjanović et al. (2020), using multibeam
25 bathymetry and coincident multichannel seismic (MCS) data to constrain the seabed and basement

26 interfaces, respectively. Sediment velocity is defined as 1.86 km/s, the average from semblance analysis
27 of MCS gathers. The base crustal starting velocity model follows a 1D profile based on White et al.
28 (1992) and Grevenmeyer et al. (2018a), with crustal thickness of 6 km. We define the model onto a 200 m
29 x 50 m grid, from the sea surface to 20 km depth.

30 We followed a top-down, layer-stripping inversion procedure: first inverting the Pg dataset to resolve
31 the upper crustal structure, before introducing PmP picks to resolve the lower crust and Moho depth,
32 and finally inverting the whole dataset together to resolve upper mantle structure. We apply both
33 smoothing and damping constraints during each iteration to keep the model realistic and to avoid
34 artefacts and local fit minima. We reach a final model after a total of nine iterations, finishing with a χ^2
35 fit of 2.4.

36 We then follow a Monte Carlo-type procedure to produce 100 possible final models (e.g. Korenaga et
37 al., 2000). An array of 100 starting models is produced by randomly perturbing the crustal thickness
38 (by up to ± 1 km), top mantle velocity (up to $-0.5/+0.2$ km/s), and one crustal velocity parameter (up to
39 ± 0.5 km/s) of the base starting velocity model. Subsequently, 100 travel-time datasets are produced by
40 adding random errors to the picked dataset: a common receiver (OBS) error up to 42 ms (the average
41 dataset travel-time error), followed by another error up to 42 ms, differing for each individual pick time.
42 Pairs of the random starting model and travel-time datasets are then inverted using the same procedure
43 and parameters. We take the mean average of the 100 final models to interpret in our study, which has
44 an average χ^2 fit of 2.4.

45 We use a variety of methods to quantify the coverage, resolution and uncertainty of the final velocity
46 model. The derivative weight sum (Supplementary Fig. S2A - DWS – Toomey & Foulger, 1989) shows
47 good ray coverage of the crustal and upper mantle. Checkerboard tests, performed with anomaly sizes
48 of 25 km x 4 km and 15 km x 3 km at an amplitude of 5%, show good recovery of these anomalies with
49 a little distortion from ray smearing (Supplementary Fig. S2B-E). To quantify the uncertainty in velocity
50 and Moho depth, we take the mean average standard deviation of the 100 final models from the mean
51 model (Supplementary Fig. S2F). Uncertainties in velocity are generally low: varying from 0.1-0.2 km/s
52 in the upper crust to <0.1 km/s in the lower crust, and 0.1-0.2 km/s in the upper mantle. Moho depth

53 standard deviations vary from ~0.1-0.5 km along the model. To examine whether the model beneath
54 the transform valley, including a velocity contrast at the Moho, is reasonable, we perform 2D finite
55 difference modeling to compute synthetic seismograms. The synthetic data show that a model
56 containing a velocity contrast at the Moho provides a significantly better match to the observed arrivals
57 than a model with smoothly increasing velocity with depth (Supplementary Fig. S4).

58 **Gravity analysis**

59 Gravity data were acquired along the profile during the ILAB-SPARC cruise using a Bondenseewerk
60 KSS31 gravimeter. Marjanović et al. (2020) describe the data corrections and the calculation of the
61 Free-Air Anomaly (FAA – Supplementary Fig. S5A).

62 In order to perform forward modeling to produce a crustal density model, we then correct the FAA for
63 the gravity anomaly produced by differences in mantle temperature across the transform, due to plate
64 cooling (Kuo & Forsyth, 1988). We use the age of the lithosphere (Müller et al., 2019), bathymetry
65 (GEBCO, 2019), half-spreading rate of 1.75 m/yr, half-space cooling model (Turcotte & Schubert,
66 2002), and mantle density of 3300 kg/m³, to calculate the mantle thermal anomaly in the region of the
67 Equatorial Atlantic Ocean. To find the most appropriate value of the mantle thermal expansion
68 coefficient in our study area to use in the calculation, we perform analysis along two flowlines, one
69 from the northern ridge segment and one from the southern. Firstly, we calculate mantle bouguer
70 anomaly (MBA) along the flowlines (Supplementary Fig. S6) and compare the fit of the MBA to the
71 mantle thermal correction calculated using coefficient values ranging from 2.8-4.8 x 10⁻⁵ K⁻¹
72 (Supplementary Fig. S7a). We find that for the northern flowline, the best-fit value of 4.4 x 10⁻⁵ K⁻¹ is
73 outside the range of possible values for oceanic lithosphere of 3.0-3.8 x 10⁻⁵ K⁻¹ (Afonso et al., 2005),
74 and for the southern profile, although the best-fit value of 3.4 x 10⁻⁵ K⁻¹ is realistic, an insignificant
75 difference in fit (<1 mGal) is found for values ranging from 2.9-3.9 x 10⁻⁵ K⁻¹. We therefore then
76 estimate the parameter by comparing observed bathymetry with predicted seafloor subsidence,
77 calculated using a half-space cooling model (e.g. Parsons & Sclater, 1977) with different values of the
78 thermal expansion coefficient (Supplementary Fig. S8, S7b). For the half-space cooling model, we use
79 mantle and water densities of 3330 kg/m³ and 1000 kg/m³, respectively, a base mantle temperature of

80 1350°C, thermal conductivity of 3.5 W/m/K, and specific heat capacity of 1170 J/kg/K. Ridge depths
81 are calculated as the average of the west and east ridge wall depths on each flowline and are 2650 m
82 and 2550 m for the northern and southern flowlines, respectively. Using a least-squares fitting
83 algorithm, we find best-fit values of $3.5 \times 10^{-5} \text{ K}^{-1}$ for the northern flowline and $3.6 \times 10^{-5} \text{ K}^{-1}$ for the
84 southern (Supplementary Fig. S8). As both are realistic values and are similar to one another, we take
85 the average of $3.55 \times 10^{-5} \text{ K}^{-1}$ to use in our final calculation of the mantle thermal anomaly. To quantify
86 uncertainty in the thermal anomaly correction caused by potential errors in plate age, we use the age
87 error grid provided by Müller et al. (2008), as there is not one provided with the 2019 version of the
88 model. The greatest errors in age along our profile occur within the transform valley, with an average
89 of ± 5.7 Myr, and the resulting errors in the mantle thermal correction are shown in Supplementary Fig.
90 S5A.

91 Aside from the mantle thermal correction, we also correct the observed FAA for the effect of 3D
92 bathymetry by calculating the difference in the Bouguer correction when using the 2D bathymetry
93 extracted along the model profile and the 2.5D talwani2d GMT program, and the 3D bathymetry from
94 the ship swath grid supplemented by GEBCO satellite data, using GMT's gravfft program. Applying
95 these corrections to the FAA produces the observed, corrected gravity profile, with error bounds
96 resulting from the uncertainty in the mantle thermal correction (Supplementary Fig. S5B).

97 Subsequently, we convert the final velocity model into a density model, maintaining a constant density
98 in the water (1035 kg/m^3), sediments (1900 kg/m^3), crust (2700 kg/m^3) and mantle (3300 kg/m^3). We
99 use the talwani2d code in GMT, employing the Talwani method (Talwani et al., 1959) for calculating
100 gravity over density anomalies in 2.5D, to calculate FAA over the density model. This uses gravity
101 contrasts as opposed to absolute values, so we then apply a bulk shift to the modelled gravity profile so
102 the mean fit to the observed profile is ~ 0 mGal (Supplementary Fig. S5B). The constant density model
103 fits the observed profile relatively well (RMS of ~ 14 mGal), but a clear low gravity anomaly of
104 ~ 20 mGal is present within the transform valley (considering uncertainties this could potentially vary
105 from -32 to -6 mGal), along with a high gravity anomaly of ~ 20 mGal south of Romanche (11 to 37
106 mGal).

107 We begin forward modelling by adding in a lower density (2500 kg/m^3) upper crust, following the
108 6.6 km/s (White et al., 1992) contour extracted from the velocity model, which improves the fit along
109 the whole 0-400 km profile, and subsequently re-adjust the bulk shift. This addition effectively
110 introduces a large low-density anomaly into the transform valley. The fit within the valley is further
111 improved through introducing a low-density anomaly (3100 kg/m^3) to the upper mantle, with the extent
112 of the anomalies guided by the velocity anomaly in the final inversion model. The fit to the south of
113 Romanche can be improved through the addition of a high-density anomaly (2700 kg/m^3) to the upper
114 crust, although due to the ambiguity of gravity data this could be equally fit with a lower crust density
115 anomaly. The final RMS fit between the observed and modelled data is 8.2 mGal . As mentioned above,
116 this method is sensitive to the density contrasts as opposed to absolute density, so within the transform
117 valley crust and mantle the interpretable density anomalies are both -200 kg/m^3 from the background
118 crust and mantle densities and in the crust to the south of Romanche the anomaly is $+200 \text{ kg/m}^3$ from
119 the background crust density.

120 **Effective medium analysis**

121 In investigate the cause of the V_p anomaly, we undertake an effective medium analysis, following the
122 differential effective medium (DEM) theory (Taylor & Singh, 2002). To simulate fracturing within a
123 transform fault, we add elongate, aligned, fluid-filled fractures, with aspect ratios of 5-100, to a host
124 medium and assess the effect on the physical properties with increasing porosity.

125 We take the intrinsic properties of the different host media from the literature, with values used
126 summarised in Supplementary Table 1. For mafic material, we use values from laboratory
127 measurements of gabbro samples from the MAR and Oman ophiolite (Hyndman & Drury, 1976; Saito
128 et al., 2015), as we can obtain more reliable, intrinsic ‘zero-porosity’ parameters for gabbro due to its
129 lack of natural porosity compared to recovered basalts and dolerites. Due to its ubiquity in dredge
130 samples from the eastern Romanche (Seyler & Bonatti, 1997; Tartarotti et al., 2002), we assume
131 melt-impregnated peridotite (MIP) is the dominant peridotite composition within the transform valley,
132 and include unaltered peridotite for the composition beneath the surrounding crust. For these lithologies,
133 we use modal and mineral compositions from dredged sample analyses from Romanche TF to predict

134 the physical properties using the toolbox from Abers & Hacker (2016). We average the composition of
 135 samples from Tartarotti et al. (2002) for a melt-impregnated peridotite composition of 59.2% olivine
 136 (90% Fo, 10% Fa), 16.5% orthopyroxene (90% En, 10% Fs), 11.8% clinopyroxene (Di), 1.5% spinel
 137 and 11% plagioclase (90% An, 10% Ab). We use the composition of the undepleted lherzolite from
 138 eastern Romanche from Seyler & Bonatti (1997), similar to the western Romanche undepleted
 139 lherzolites, for an unaltered peridotite composition of 65.2% olivine (90% Fo, 10% Fa), 22.3%
 140 orthopyroxene (90% En, 10% Fs), 10.8% clinopyroxene (Di) and 1.7% spinel. For the properties of
 141 serpentinized melt-impregnated peridotite, we take standard Vp and density changes from
 142 serpentinization for ‘standard’ peridotite (~90% olivine harzburgite) of -3 km/s in Vp and -700 kg/m³
 143 in density for 100% serpentinization (e.g. Roland et al., 2012), and scale these for the olivine
 144 composition of the melt-impregnated peridotite. This assumes that the alteration of olivine to serpentine
 145 is the dominant reaction with the greatest effect on the physical properties. We then use a Vp/Vs value
 146 of 2.0 to calculate the Vs (after Grevemeyer et al., 2018b; Christensen, 1966).

147 *Supplementary Table 1: Intrinsic physical properties used for different lithologies in DEM analyses*

	Vp (km/s)	Vs (km/s)	Density (kg/m³)
<i>Gabbro</i>	7.00	3.83	3000
<i>Unaltered peridotite</i>	8.23	4.80	3316
<i>Melt-impregnated peridotite (MIP)</i>	8.05	4.65	3252
<i>100% serpentinized MIP</i>	6.13	3.07	2804

148

149 We use the intrinsic physical properties of each lithology as the properties of the host medium in the
 150 DEM calculations. We add increasing proportions of fractures with aspect ratios of 1, 5, 10 and 100 and
 151 plot the porosity against the resultant P-wave velocity and density (Supplementary Fig. S9). To assess
 152 the cause of the crustal low velocity anomaly, we compare the DEM results for gabbro and 100%
 153 serpentinized MIP with the range of Vp observed in the transform valley crust. We also show the effect
 154 of progressive serpentinization on the velocity of MIP. To further quantify the causes of the transform
 155 valley anomalies, we show the change in density associated with the increasing porosity for different

156 lithologies (Supplementary Fig. S9b), compared with the density anomalies within the crust and mantle
157 as modelled using the gravity data.

158 REFERENCES CITED

- 159 Abers, G. A., & Hacker, B. R. (2016). A MATLAB toolbox and Excel workbook for calculating the
160 densities, seismic wave speeds, and major element composition of minerals and rocks at
161 pressure and temperature. *Geochemistry, Geophysics, Geosystems*, *17*(2), 616–624.
162 <https://doi.org/10.1002/2015GC006171>
- 163 Afonso, J. C., Ranalli, G., & Fernández, M. (2005). Thermal expansivity and elastic properties of the
164 lithospheric mantle: Results from mineral physics of composites. *Physics of the Earth and
165 Planetary Interiors*, *149*(3–4), 279–306. <https://doi.org/10.1016/j.pepi.2004.10.003>
- 166 Christensen, N. I. (1966). Elasticity of ultrabasic rocks. *Journal of Geophysical Research*, *71*(24),
167 5921–5931. <https://doi.org/10.1029/JZ071i024p05921>
- 168 Grevemeyer, I., Ranero, C. R., & Ivandic, M. (2018a). Structure of oceanic crust and serpentization
169 at subduction trenches. *Geosphere*, *14*(2), 1–24. <https://doi.org/10.1130/GES01537.1>
- 170 Grevemeyer, I., Hayman, N. W., Peirce, C., Schwardt, M., Van Avendonk, H. J. A., Dannowski, A.,
171 & Papenberg, C. (2018b). Episodic magmatism and serpentized mantle exhumation at an
172 ultraslow-spreading centre. *Nature Geoscience*, *11*(6), 444–448. [https://doi.org/10.1038/s41561-
173 018-0124-6](https://doi.org/10.1038/s41561-018-0124-6)
- 174 Hyndman, R. D., & Drury, M. J. (1976). The physical properties of oceanic basement rocks from deep
175 drilling on the Mid-Atlantic Ridge. *Journal of Geophysical Research*, *81*(23), 4042–4052.
176 <https://doi.org/10.1029/JB081i023p04042>
- 177 Korenaga, J., Holbrook, W. S., Kent, G. M., Kelemen, P. B., Detrick, R. S., Larsen, H. C., Hopper, J.
178 R., & Dahl-Jensen, T. (2000). Crustal structure of the southeast Greenland margin from joint

179 refraction and reflection seismic tomography. *Journal of Geophysical Research: Solid Earth*,
180 105(B9), 21591–21614. <https://doi.org/10.1029/2000jb900188>

181 Kuo, B.-Y., & Forsyth, D. W. (1988). Gravity Anomalies of the Ridge-Transform System in the South
182 Atlantic Between 31 and 34.5° S: Upwelling Centers and Variations in Crustal Thickness.
183 *Marine Geophysical Researches*, 10, 205–232.

184 Marjanović M., Singh, S. C., Gregory, E. P. M., Grevemeyer, I., Growe, K., Vaddineni, V., Laurencin,
185 M., Carton, H., Gómez de la Peña, L., & Filbrandt, C. Seismic Crustal Structure and Morpho-
186 tectonic Features Associated with the Chain Fracture Zone and their Role in the Evolution of the
187 Equatorial Atlantic Region. *Journal of Geophysical Research: Solid Earth*.

188 Müller, R. D., Sdrolias, M., Gaina, C., & Roest, W. R. (2008). Age, spreading rates, and spreading
189 asymmetry of the world's ocean crust. *Geochemistry, Geophysics, Geosystems*, 9(4), 1–19.
190 <https://doi.org/10.1029/2007GC001743>

191 Müller, R. D., Zahirovic, S., Williams, S. E., Cannon, J., Seton, M., Bower, D. J., Tetley, M. G.,
192 Heine, C., Le Breton, E., Liu, S., Russell, S. H. J., Yang, T., Leonard, J., & Gurnis, M. (2019). A
193 Global Plate Model Including Lithospheric Deformation Along Major Rifts and Orogens Since
194 the Triassic. *Tectonics*, 38(6), 1884–1907. <https://doi.org/10.1029/2018TC005462>

195 Parsons, B., & Sclater, J. G. (1977). An analysis of the variation of ocean floor bathymetry and heat
196 flow with age. *Journal of Geophysical Research*, 82(5), 803–827.
197 <https://doi.org/10.1029/jb082i005p00803>

198 Saito, S., Ishikawa, M., Arima, M., & Tatsumi, Y. (2015). Laboratory measurements of ‘porosity-
199 free’ intrinsic V_p and V_s in an olivine gabbro of the Oman ophiolite: Implication for
200 interpretation of the seismic structure of lower oceanic crust. *Island Arc*, 24(2), 131–144.
201 <https://doi.org/10.1111/iar.12092>

- 202 Seyler, M., & Bonatti, E. (1997). Regional-scale melt-rock interaction in lherzolitic mantle in the
203 Romanche Fracture Zone (Atlantic Ocean). *Earth and Planetary Science Letters*, 146, 273–287.
- 204 Talwani, M., Worzel, J. L., & Landisman, M. (1959). Rapid Gravity Computations for Two-
205 Dimensional Bodies with Application to the Mendocino Submarine Fracture Zone. *Journal of*
206 *Geophysical Research*, 64(1).
- 207 Tartarotti, P., Susini, S., Nimis, P., & Ottolini, L. (2002). Melt migration in the upper mantle along the
208 Romanche Fracture Zone (Equatorial Atlantic). In *Lithos* (Vol. 63).
209 www.elsevier.com/locate/lithos
- 210 Taylor, M. A. J., & Singh, S. C. (2002). Composition and microstructure of magma bodies from
211 effective medium theory. *Geophysical Journal International*, 149(1), 15–21.
212 <https://doi.org/10.1046/j.1365-246X.2002.01577.x>
- 213 Toomey, D. R., & Foulger, G. R. (1989). Tomographic inversion of local earthquake data from the
214 Hengill- Grendalur central volcano complex, Iceland. *Journal of Geophysical Research*,
215 94(B12). <https://doi.org/10.1029/jb094ib12p17497>
- 216 Turcotte, D. L., & Schubert, G. (2002). *Geodynamics*. Cambridge University Press.
- 217 Van Avendonk, H. J. A., Harding, A. J., Orcutt, J. A., & McClain, J. S. (1998). A two-dimensional
218 tomographic study of the Clipperton transform fault. *Journal of Geophysical Research: Solid*
219 *Earth*, 103(B8), 17885–17899. <https://doi.org/10.1029/98JB00904>
- 220 Van Avendonk, H. J. A., Shillington, D. J., Holbrook, W. S., & Hornbach, M. J. (2004). Inferring
221 crustal structure in the Aleutian island arc from a sparse wide-angle seismic data set.
222 *Geochemistry, Geophysics, Geosystems*, 5(8). <https://doi.org/10.1029/2003GC000664>

223 White, R. S., McKenzie, D., & O’Nions, R. K. (1992). Oceanic crustal thickness from seismic
224 measurements and rare earth element inversions. *Journal of Geophysical Research*, 97(B13),
225 19683–19715. <https://doi.org/10.1029/92JB01749>

226

227

228

229

230

231

232

233

234

235

236

237

238

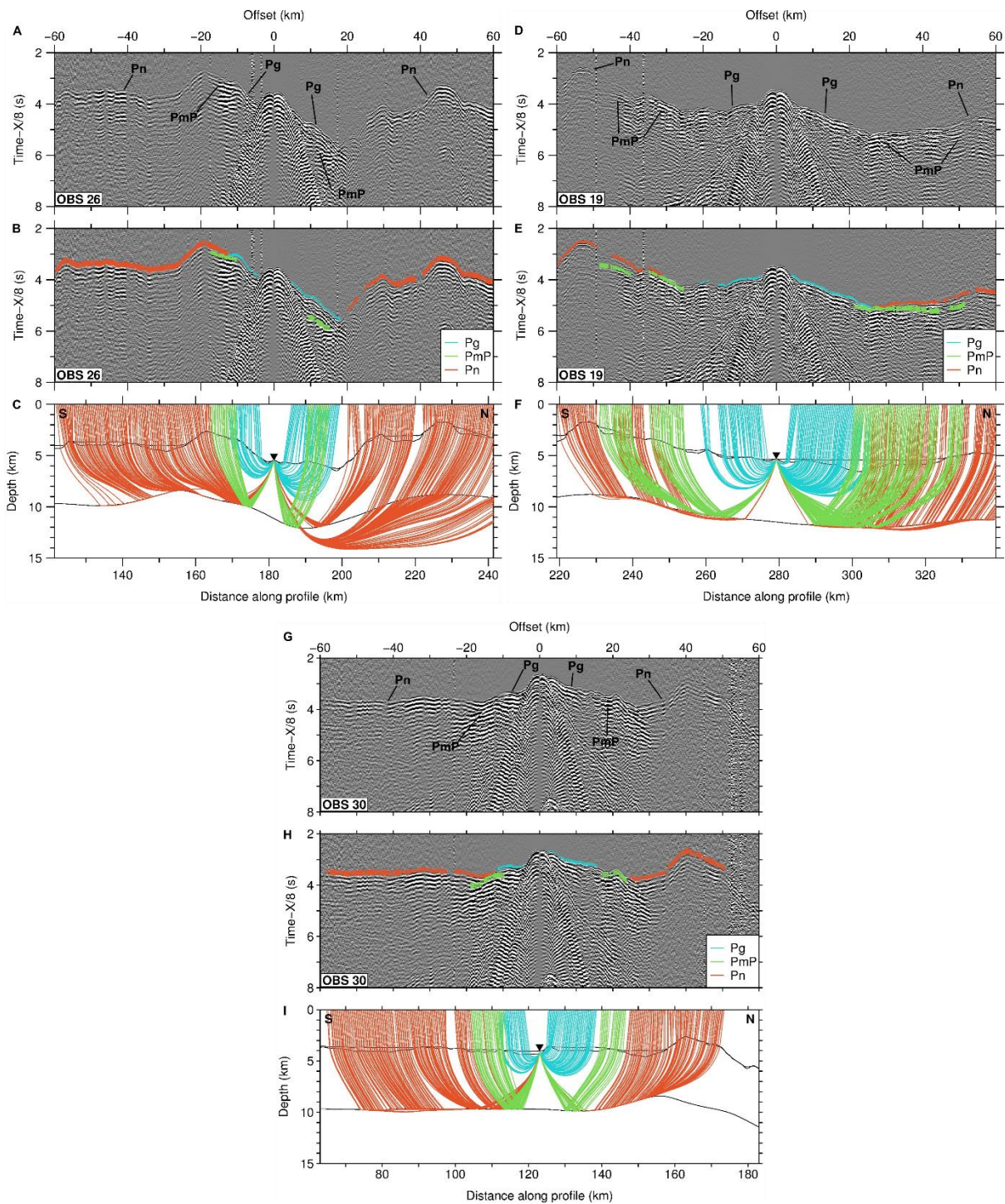
239

240

241

242

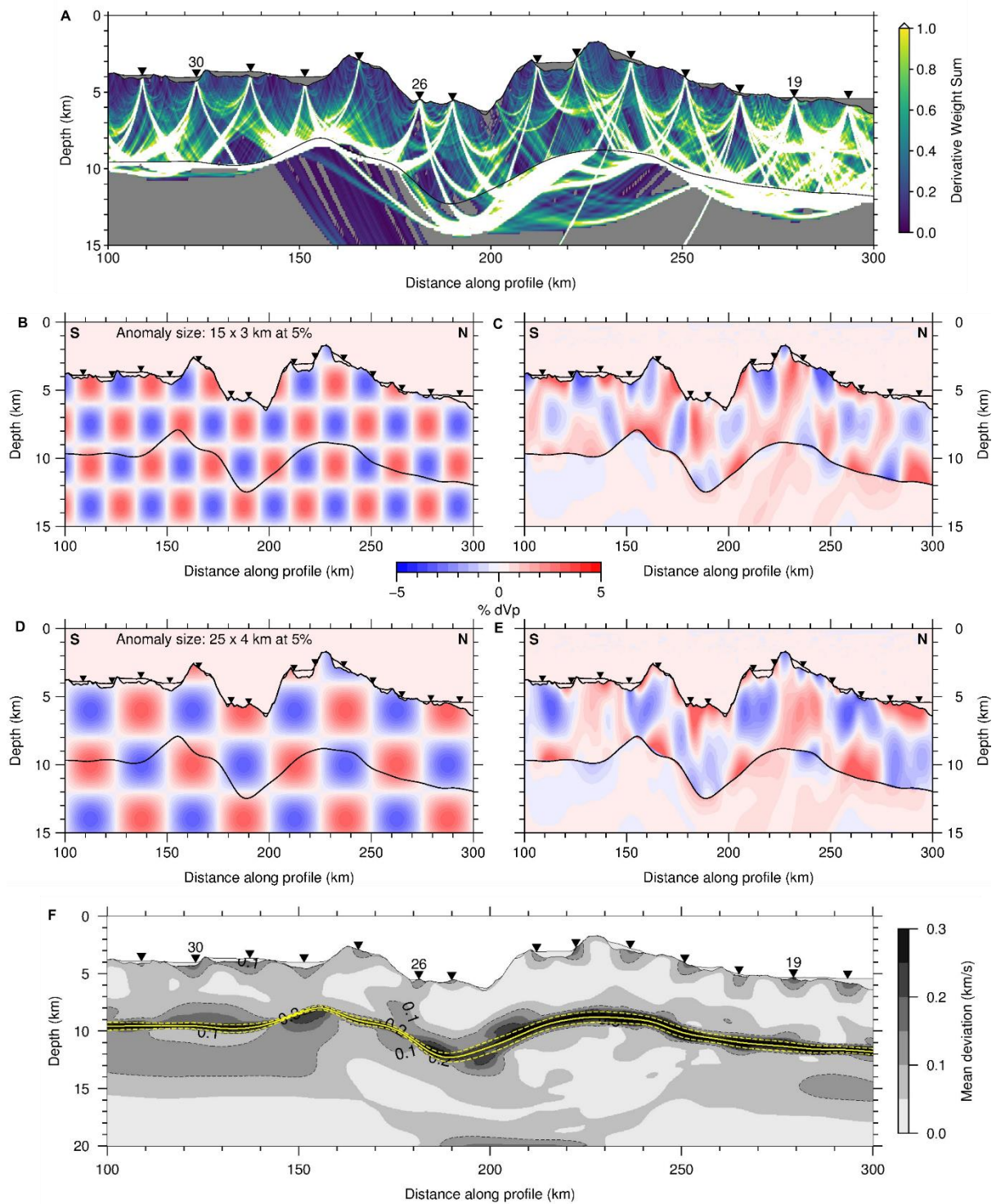
243



245

246 **Supplementary figure S1:** OBS records and ray tracing. A. Record section for OBS26, located within
 247 Romanche transform valley, with identified phases. B. Travel time picks, width of line indicates pick
 248 uncertainty. C. Ray tracing through final velocity model, with ray colour matching the picks in B. D, E
 249 and F show the same for OBS19, as an example over normal magmatic crust to the north of Romanche,
 250 and G, H and I for OBS30, as an example over thinner, tectonic crust to the south of Romanche.

251

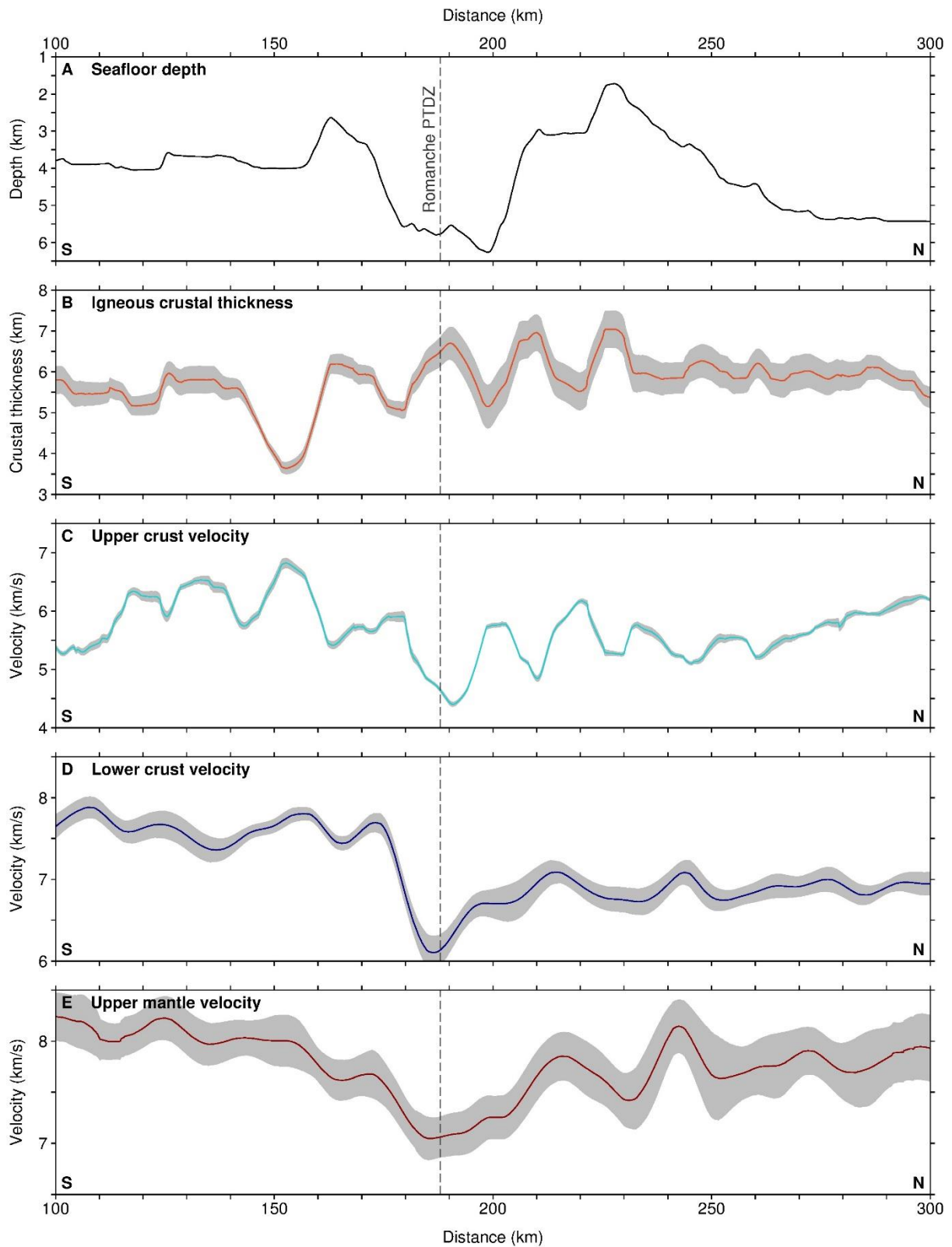


252

253 **Supplementary figure S2:** A. Derivative Weight Sum plot of final tomography model, showing good
 254 crustal ray coverage. B-E. Checkerboard test input checkerboards (B, D) and recovered checkerboards
 255 (C, E). F. Mean standard deviation from the average final velocity model. The average Moho is shown
 256 in yellow solid line, with its mean deviation shown in yellow dashed line.

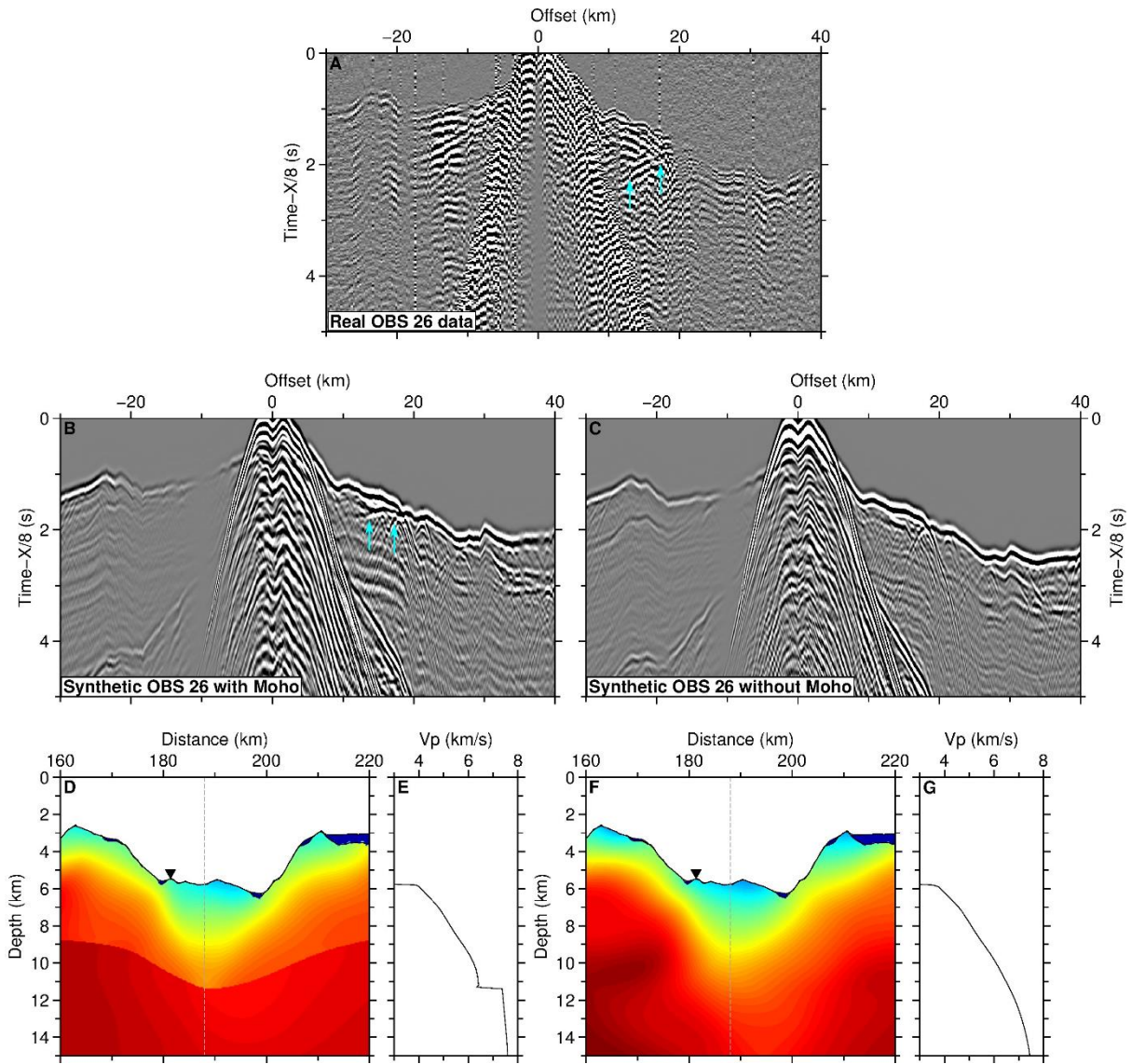
257

258



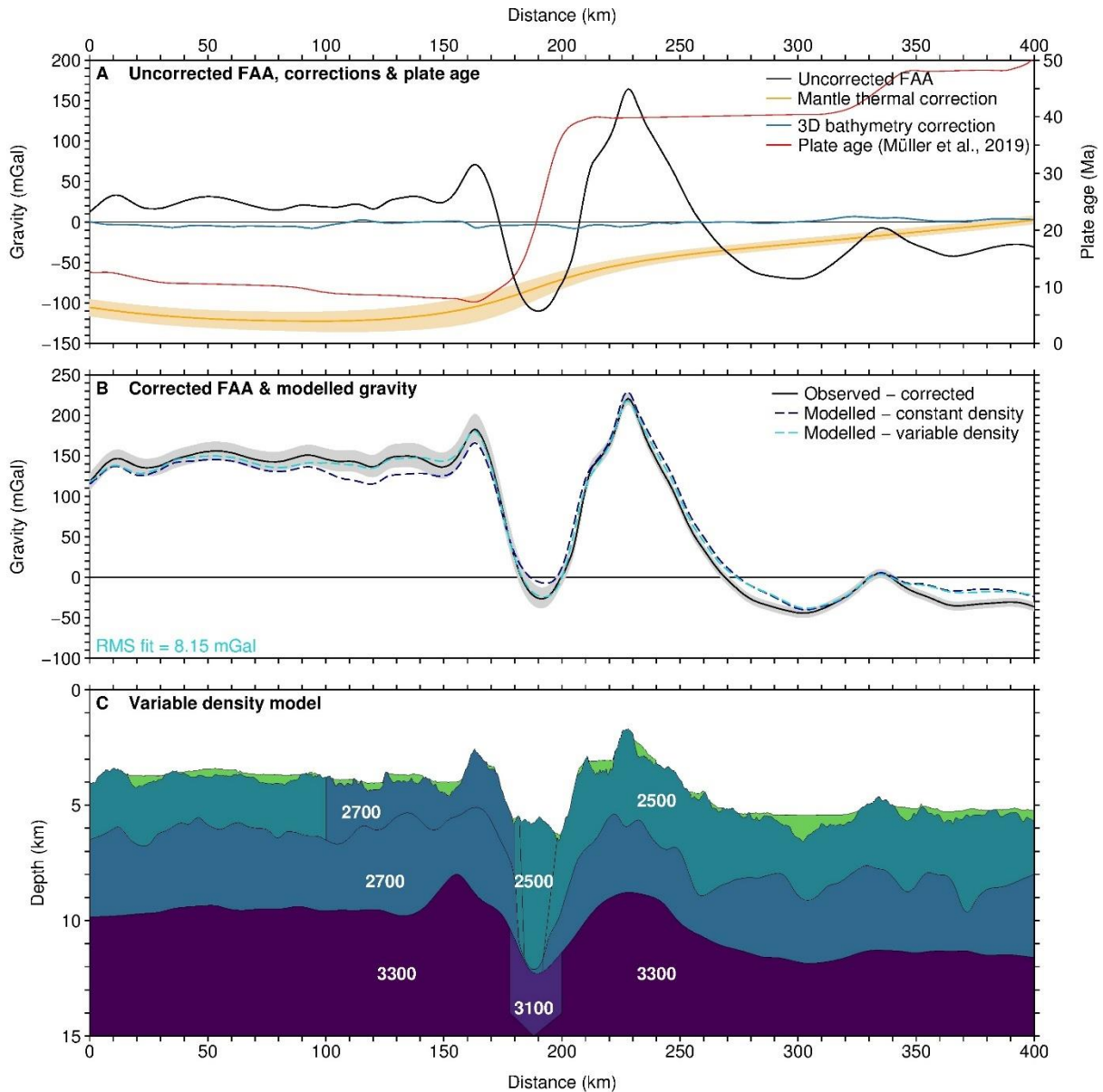
259

260 **Supplementary figure S3:** Key characteristics from the final velocity model. A. Seafloor depth from
 261 swath bathymetry. B. Crustal thickness from top basement to Moho. C. Average upper crust velocity
 262 (top 3 km of crust). D. Average lower crust velocity (2 km of crust above Moho). E. Average upper
 263 mantle velocity (top 0.5 km of mantle). Standard deviation from the average final model is shown by
 264 the grey bands.



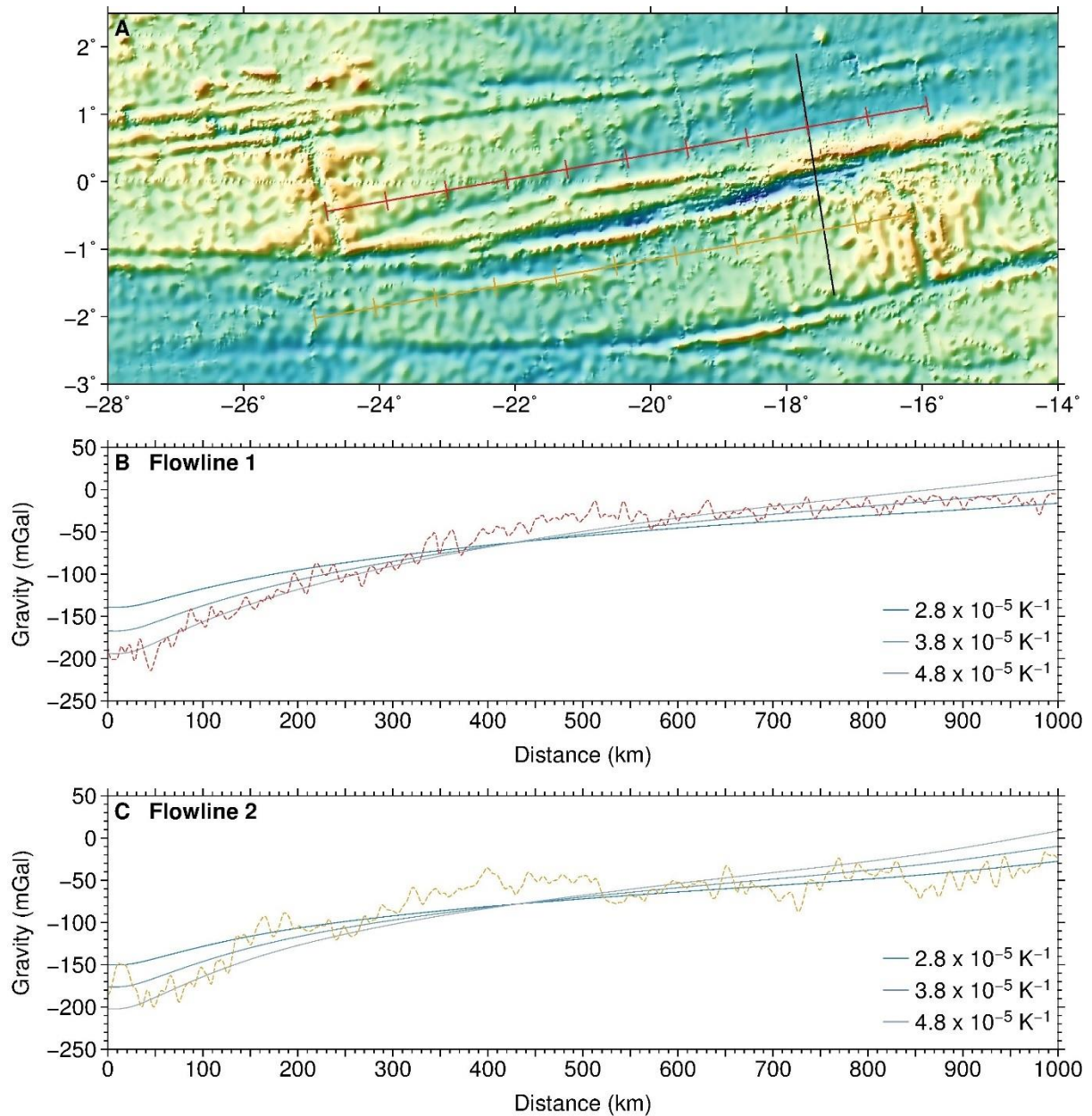
265

266 **Supplementary figure S4:** Synthetic modeling of seismograms, illustrating that a model with a velocity
 267 contrast at the Moho produces a seismogram containing PmP arrivals, consistent with the observed data.
 268 A. Observed data from OBS 26, PmP reflections marked by blue arrows. B. Synthetic seismogram of
 269 OBS 26 produced through finite difference modeling using a model containing a Moho (see D., E.).
 270 PmP arrivals marked by blue arrows. C. Synthetic seismogram produced using a model without a Moho,
 271 with gradually increasing velocity with depth (see F., G.). D. and F. Tomographic models used to
 272 produce the seismograms in B. and C., respectively. The grey dashed lines show the location of the 1D
 273 velocity-depth profiles shown in E. and G. Black triangle is the position of OBS 26. Static shifts have
 274 been applied to the OBS data shown here to remove the effect of seafloor topography.



275

276 **Supplementary figure S5:** Gravity modelling. A. Observed, uncorrected Free-Air Anomaly (FAA -
 277 black) from shipboard gravimeter, alongside the corrections to compensate for the mantle thermal
 278 gravity anomaly (orange – uncertainty bounds shown in pale orange), and the effect of bathymetric
 279 features not on the 2D profile (blue). Also shown is the plate age used for the mantle thermal anomaly
 280 calculation (red), extracted along the profile. B. Observed, corrected FAA gravity from shipboard
 281 gravimeter (black – uncertainty bounds shown in grey), corrected for the thermal effect of lithospheric
 282 cooling from the age contrast across the transform fault and the effects of 3D bathymetry. Calculated
 283 gravity (dark blue dashed) using a constant density crust, with thickness defined by the velocity model.
 284 Calculated gravity (light blue dashed) using a variable density model containing anomalies (see C). C.
 285 The variable density anomaly produced to match the observed gravity, with the upper/lower crust
 286 boundary defined using the 6.6 km/s contour from the velocity model. There are three density
 287 anomalies: 1. The whole crust inside the valley is at upper crust density (2500 kg/m³), which is -200
 288 kg/m³ from lower crust density; 2. Mantle anomaly below the valley is -200 kg/m³ from the rest of the
 289 mantle; 3. Upper crust to the south of Romanche is +200 kg/m³ as compared to the normal upper crust.



290

291 **Supplementary figure S6:** Analysis of the mantle thermal expansion coefficient parameter using
292 mantle bouguer anomalies. A. Bathymetric map of the study area showing the location of the two
293 flowlines used (red and orange lines, every 100 km marked with a tick), in relation to the seismic profile
294 (black line). B. Mantle bouguer anomaly calculated along flowline 1 (red dashed line; red line in A),
295 using a 6 km-thick crust and densities of 1035 kg/m^3 , 2700 kg/m^3 and 3300 kg/m^3 for water, crust and
296 mantle, respectively. The mantle thermal gravity anomalies calculated using different thermal
297 expansion coefficients are shown in blue. C. Same as in B for flowline 2 (orange dashed line; orange
298 line in A).

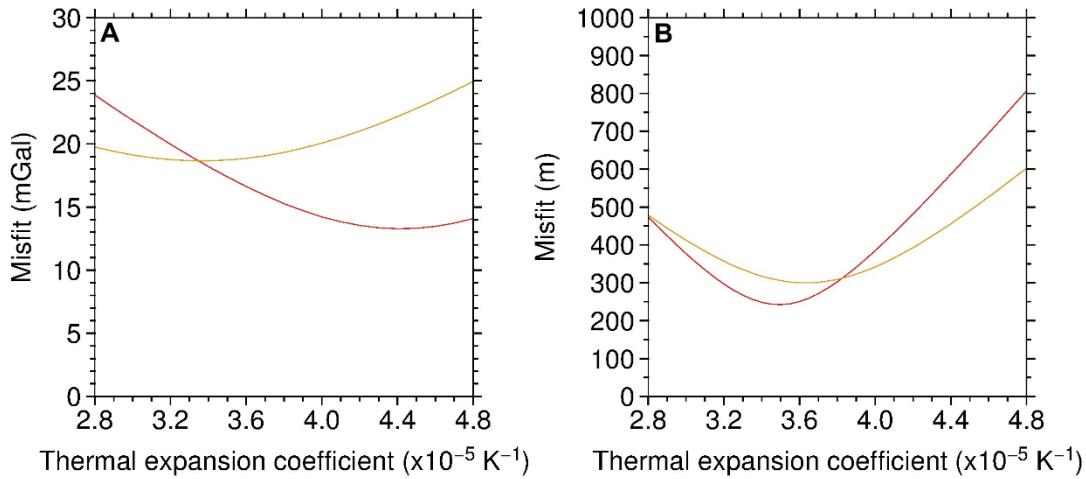
299

300

301

302

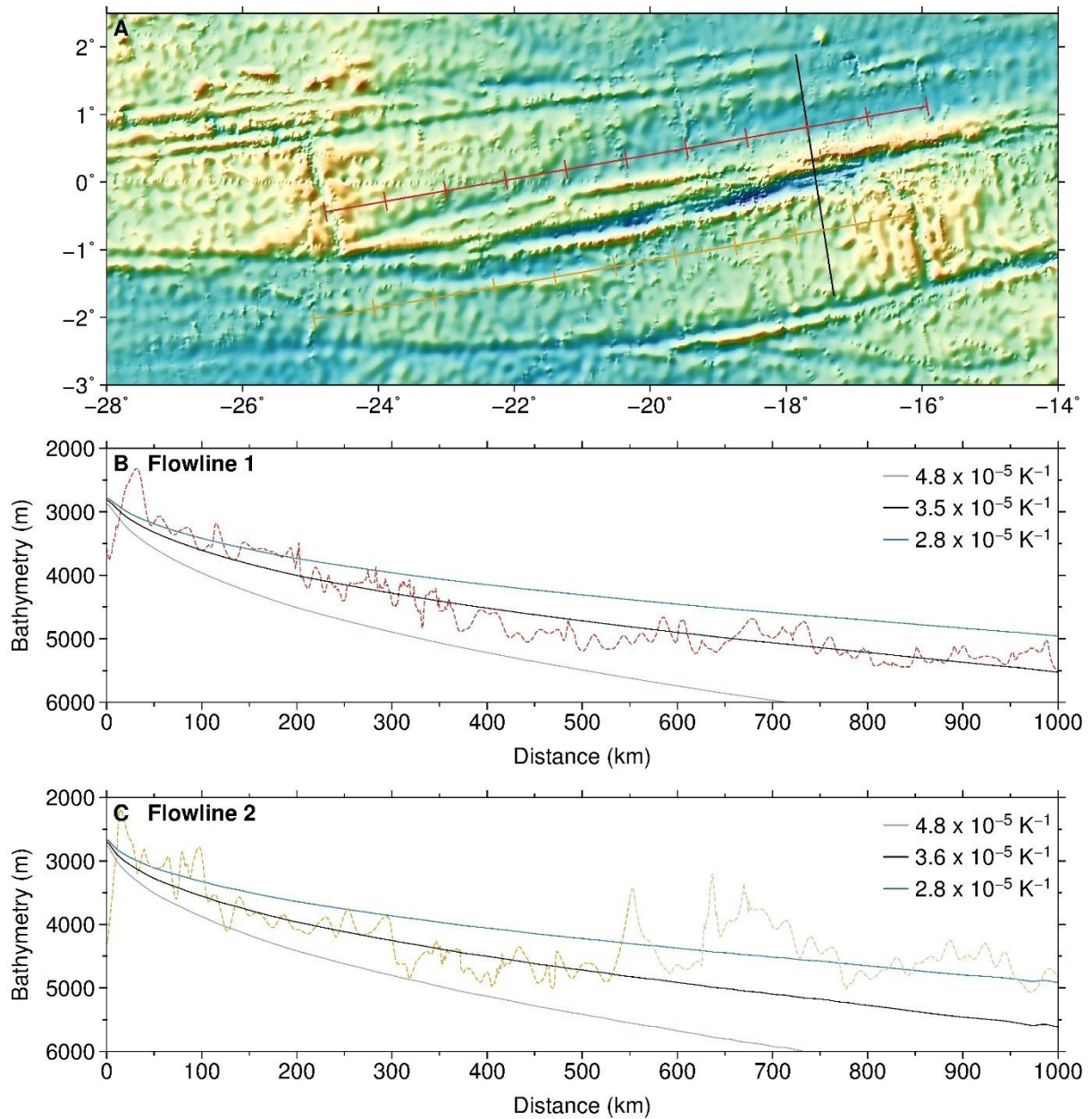
303



304

305 **Supplementary figure S7:** Misfit analysis of using different thermal expansion coefficients. A. Misfit
306 between mantle bouguer anomaly and calculated mantle thermal gravity correction using different
307 values for the coefficient for flowline 1 (red line) and flowline 2 (orange line). B. Misfit between
308 observed bathymetry and calculated seafloor depth using a half-space cooling model and different
309 values for the coefficient for flowline 1 (red line) and flowline 2 (orange line).

310



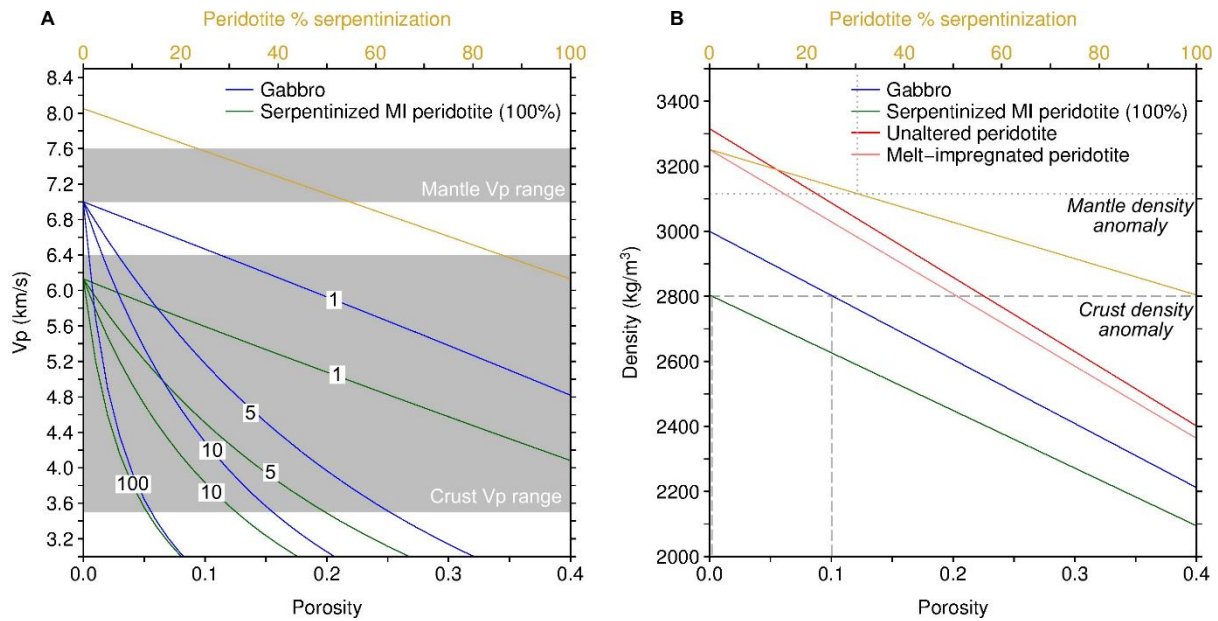
311

312 **Supplementary figure S8:** Analysis of the thermal expansion coefficient parameter using seafloor
 313 subsidence. A. Bathymetric map of the study area showing the location of the two flowlines used (red
 314 and orange lines, every 100 km marked with a tick), in relation to the seismic profile (black line). B.
 315 Bathymetric profile along flowline 1 (red dashed line; red line in A). Predictions of seafloor subsidence
 316 using a half-space cooling model and different thermal expansion coefficients are shown in blue
 317 (maximum and minimum values) and black (best-fit value). C. Same as in B for flowline 2 (orange
 318 dashed line; orange line in A). Here we only used the first 550 km (29 Myr) of the profile (shown in
 319 darker orange) as beyond this the flowline is disturbed by a propagator.

320

321

322



323

324 **Supplementary figure S9:** Effective medium and density analyses. A. Porosity-velocity results from
 325 differential effective medium analysis (Taylor & Singh, 2002), assuming aligned fluid-filled fractures
 326 in either a gabbro or 100% serpentinized melt-impregnated (MI) peridotite matrix, at a range of crack
 327 aspect ratios (indicated by numbered labels). Grey shading shows the Vp range of the transform valley
 328 crust and mantle in the velocity model. Fracturing within either gabbro or MI peridotite is necessary to
 329 explain the crustal Vp range, with slightly less maximum porosity required for MI peridotite (12% vs
 330 15% for aspect ratio of 10). A minimum porosity of 5% is required to explain the lowest velocities
 331 considering an aspect ratio of 100. Also shown in yellow is % serpentinization vs Vp for altered MI
 332 peridotite. Approximately 25-55% serpentinization is required to explain the mantle Vp range. B.
 333 Porosity-density relationships for gabbro, 100% and 0% serpentinized MI peridotite, and the unaltered,
 334 fertile lherzolite peridotite, alongside the % serpentinization-density relationship for MI peridotite.
 335 Shown in grey dashed are the transform valley crust and mantle density anomalies from gravity
 336 modelling. The anomalies (both -200 kg/m³), are shown relative to the background composition of
 337 undepleted lherzolite for the mantle and gabbro for the crust. The crustal density anomaly is consistent
 338 with ~10% porosity gabbro or 0% porosity serpentinized MI peridotite. The mantle density anomaly is
 339 consistent with ~30% serpentinized MI peridotite.

340