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1 **Seismic evidence for partial melt below tectonic plates**

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5

6 **The seismic low velocity zone (LVZ) of the upper mantle is generally**
7 **associated with a low-viscosity asthenosphere that plays a key role for the**
8 **dynamics of plate tectonics¹. However, its origin remains enigmatic, some**
9 **authors attributing the reduction in seismic velocity to a small amount of**
10 **partial melt^{2,3}, others invoking solid-state mechanisms near the solidus⁴⁻⁶, or**
11 **the effect of volatile contents⁶. Observations of shear attenuation provide**
12 **additional constraints to unravel the origin of the LVZ⁷. Here, we report the**
13 **discovery of partial melt within the LVZ from the simultaneous**
14 **interpretation of global 3D shear attenuation and velocity models. We**
15 **observe that partial melting down to 150-200 km depth beneath mid-ocean**
16 **ridges, major hotspots and back-arc regions feeds the asthenosphere. A small**
17 **part of this melt (<0.30%) remains trapped within the oceanic LVZ. The**
18 **amount of melt is related to plate velocities and increases significantly**
19 **between 3 and 5 cm yr⁻¹, similar to previous observations of mantle crystal**
20 **alignment underneath tectonic plates⁸. Our observations suggest that by**
21 **reducing viscosity⁹, melt facilitates plate motion and large-scale crystal**
22 **alignment in the asthenosphere. Melt is absent under most of the continents.**

23 Our finding results from the simultaneous analysis of two upper mantle
24 tomographic models of shear wave velocity (V_s) and attenuation (parameterised

25 with Q_s the quality factor, a measure of energy dissipation). Until now, most global
26 tomographic studies of the upper mantle and their thermochemical interpretation
27 have focused on shear velocity^{4,7}. Recent experiments on olivine suggest that wave
28 speed and attenuation are insensitive to water, which would imply that elevated
29 water contents are not responsible for the LVZ¹⁰. However, V_s is sensitive to
30 temperature, composition and melt content, and deciphering the causes of its
31 variations represents a strongly non-unique inverse problem, severely limiting our
32 understanding of the Earth's interior. Shear attenuation has a different sensitivity
33 to these quantities, and therefore provides complementary constraints on the
34 origin of seismic heterogeneities¹¹. Shear attenuation is negligibly dependent on
35 major element chemistry¹², and exponentially dependent on temperature¹³. The
36 relation between attenuation and melt is debated, with some authors arguing for a
37 weak dependence based on experiments, models and seismological observations¹²,
38 while others suggest a larger effect³. Measuring shear attenuation is nevertheless
39 difficult. For this reason, only a few global Q_s models have been published in the
40 last 20 years¹⁴, and the only recent joint interpretation of 3D Q_s and V_s
41 tomographic models at global scale is based on models built from different
42 datasets and modeling approaches¹⁵.

43 The novelty of our study is to simultaneously interpret two recent global V_s and Q_s
44 models that are consistent as obtained from the same Rayleigh wave dataset, at the
45 same resolution and using the same modelling approach. These V_s (DR2020s) and
46 Q_s (Q_s ASR17) models are displayed in **Figure 1**. Details of our tomographic
47 procedure can be found in Methods. Before interpreting these two models in the
48 light of laboratory experiments, a few words are needed to emphasize in what
49 temperature and pressure range our interpretation is pertinent. The attenuation
50 models derived by mineral physicists^{4,13} are valid for temperatures T larger than
51 900°C, which correspond to the base of the lithosphere and to the asthenosphere.

52 They consider thermally activated processes varying exponentially with $1/T$ that
53 would imply quality factor in excess of 2,000 in the upper 100 km of the
54 lithosphere and reaching several million near the surface (see Methods). However,
55 finite Q_s is observed in the crust and in the cold mantle lithosphere where
56 attenuation is most likely related to non-thermal processes such as scattering or
57 fluid-fracture interactions¹⁴. Furthermore, the attenuation observed in seismology
58 accumulates along the seismic ray and the observation of Q_s is only possible when
59 the amplitude of a wave is measurably smaller than in a pure elastic model. Given
60 the uncertainties on amplitude data, it appears impossible to resolve quality
61 factors larger than $\sim 2,000$ with long period Rayleigh waves (at periods of 100s,
62 assuming velocities of 4.5 km s^{-1} and ray lengths of 10,000 km, the amplitude
63 reduction with respect to a pure elastic model, would be less than 3.5%). Our
64 inversion leads indeed to a Q_s model with strong lateral variations, by two orders
65 of magnitudes, but with a maximum Q_s of $\sim 1,750$. Our attenuation model is
66 therefore mostly adapted to depths greater or equal to ~ 100 km, where Q_s values
67 between fifty and a few hundreds are expected¹⁵. Therefore, our interpretation
68 applies to the oceanic asthenosphere and the mantle structure at depths greater or
69 equal to ~ 100 km, where our tomographic models are accurate and where
70 conditions similar to those used in laboratory experiments prevail.

71 **Figure 2** displays Q_s as a function of V_s for each pixel of the maps, at different
72 depths in the upper mantle. The curves in dark and light blue represent the
73 theoretical relations due to temperature variations for a meltless pyrolytic
74 mantle¹⁶, given by two anelasticity models^{4,13} based on laboratory experiments,
75 appropriate for asthenospheric conditions (see Methods for details). These models
76 only explain a limited part of the velocities and attenuations of our dataset. We
77 consider that the first theoretical curve⁴ is compatible with a given V_s - Q_s
78 observation if it falls within the typical uncertainties of the V_s and $\ln(Q_s)$

79 observations, 1% and 10% respectively (ivory colour). We show in Methods and
80 supplementary **Figures S1-10** that reconciling the low values of V_s with Q_s cannot
81 be done by invoking radial anisotropy, elastically accommodated grain boundary
82 sliding, or the effect of composition or water. However, this can be done by adding
83 partial melt, thus reducing V_s , and shifting the theoretical curves to the left in
84 **Figure 2**, since as discussed below, adding partial melt reduces V_s but has little effect
85 on Q_s . Warm colours in **Figure 2** indicate the amount of melt ($<0.7\%$) needed to
86 reconcile observations with the first theoretical model⁴ (dark blue curve), which
87 requires the smallest amount of melt. On the right side of the ivory region, points in
88 grey are those for which V_s is too high compared to the theoretical value. They are
89 associated with the lithospheric depletion of cratonic roots. In this case, the grey
90 intensity quantifies the departure in percent from the theoretical curve assuming a
91 meltless pyrolitic mantle. Results using the second theoretical model¹³ are shown
92 in **Figures S11-12**. They lead essentially to the same conclusions, but require
93 larger melt fractions (up to 1%), which are more difficult to reconcile with the very
94 small melt fraction ($\sim 0.1\%$) suggested by geochemical studies⁴.

95 The effect of melt on V_s has been estimated to 7.9% reduction per percent of melt
96 based on model calculations¹⁷. Recent experiments³ are in qualitative agreement
97 but require a slightly larger V_s reduction (**Figure S13**). The effect of melt on
98 attenuation is not well constrained and depends on the mechanisms of
99 attenuation. We show in Methods that melt may have a large effect on Q_s at short
100 period (1 s), but not in the period range of surface waves (50-250s). We therefore
101 neglect the effect of melt on Q_s and we model its effect on V_s based on recent
102 experiments³. **Figure 2** shows that the slowest shear velocities require less than
103 0.7 % of melt.

104 **Figure 3** displays global maps of melt content at different depths, using the same
105 colour-coding as in **Figure 2**. The associated mantle temperatures, derived by our
106 approach are on average slightly above the solidus in oceanic regions, between
107 100 and 200 km depth (**Figure S14, panel g**). The differences between the maps
108 of temperature (**Figure S14 panels a-f**) and those of melt content are attributed to
109 a variable amount of volatile⁹. A higher amount of volatiles above subduction
110 zones may lower the solidus and favor melting, while a dryer mantle in other
111 regions may impede melting. The heterogeneities in **Figure 3** display a strong
112 correlation with surface tectonics. Regions where V_s is too high (in grey) are
113 located beneath continents down to 150 km depth. The discrepancy is likely due to
114 our assumption of a homogeneous and pyrolitic mantle. This assumption is
115 reasonable in the well-mixed convective mantle. However, beneath cratons,
116 compositional heterogeneities and depletion of incompatible elements are
117 contributing to the high seismic velocities. Our observations under cratons, at 150
118 km depth, are on average 2.4% faster than a pyrolitic mantle which is compatible
119 with compositional effects¹⁸.

120 At 100 km depth, melt is required below mid-ocean ridges, some hotspots near the
121 Atlantic ridge and in the south Pacific Ocean, back-arc basins around the Pacific
122 Ocean including the eastern margin of Asia and some other active tectonic regions
123 (Afar, Tibet, West of North-America and Southwest Pacific, including the North
124 Fidji basin and the Northfolk ridge south of the Vanuatu arc). In these regions, the
125 amount of melt exceeds 0.3% and can reach 0.7%. Melt is not required beneath
126 most of the remaining oceanic and continental Phanerozoic lithosphere where
127 temperature variations alone explain our observations.

128 The depth range 150-200 km corresponds to the oceanic LVZ, where a small
129 amount of melt (<0.3%) is required over broad regions. The largest amount is

130 beneath hotspots, ridges, and volatile-rich back-arc regions and can extend deeper
131 than the LVZ, suggesting that these deep regions feed the asthenosphere with
132 partial melt. For example, partial melt is observed down to 250 km beneath Hawaii
133 and 300 km beneath the Afar and East-African rift, the hotspots located on the
134 western part of North-America, the region of the Balleny Islands in Antarctica, the
135 western Pacific and the Indian ocean near the Ninety-East ridge. A melting
136 anomaly near 300 km on the eastern part of the Tibetan plateau is also observed.

137 The depth extent and amount of melt beneath oceanic regions are consistent with
138 local studies. The NoMELT experiment¹⁹ was performed beneath a location
139 relatively far away from Pacific hotspots (blue star in **Figure 3**) where we too
140 confirm the absence of melt in what could be a volatile-poor mantle. We observe
141 melting beneath the East Pacific Rise (green star in **Figure 3**) down to 250 km
142 depth, in agreement with the MELT experiment²⁰. The depth extent of melting
143 beneath Hawaii down to 250 km is supported by measurements of water
144 abundance in the deep region that feeds the plume²¹. Our results agree with
145 observations of significant melting beneath the Philippine Sea plate and the
146 Western Pacific². Finally, melting down to at least 300 km beneath the Balleny
147 Islands hotspot is also consistent with previous observations²².

148 The origin of partial melt within most of the oceanic LVZ is uncertain. Melting at
149 mid ocean-ridges can exceed 1% in the depth range 40-80 km^{20,23}, but smaller
150 volume melting may occur down to depths of 150-250 km²³. After melting at mid-
151 ocean ridges, a small amount of melt may remain unextractable from the mantle
152 peridotite^{9,24,25}. Our observations suggest that decompression melting with at least
153 0.3 to 0.7% of melt occurs also beneath some hotspots and back arc basins down to
154 about 200 km. Most of the melt produced in these regions is extracted and
155 incorporated to the oceanic crust, but a small amount remains in the oceanic LVZ

156 as it ages. The quantity of melt, if any, decreases close to continents below 0.1%,
157 where a simple model without melt explains most of our observations. This may be
158 due to the difficulty of melting the depleted continental lithosphere. However under
159 tectonically active regions, beneath the Afar and East-African rifts, Tibet, Western
160 North-America and Transantarctic Mountains, the asthenosphere contains a small
161 amount of melt (**Figure 3**).

162 The quantity of melt needed to reconcile Vs and Qs under large swaths of the
163 oceanic LVZ is larger than usual estimates of unextractable melt, which range from
164 very small values⁹ to a maximum of 0.1%²⁴. Although melt is likely connected and
165 able to percolate even at very small porosity⁹, surface tension resists the phase
166 separation^{9,26}. The ability of the melt to rise depends on various parameters,
167 surface tension, buoyancy, permeability, melt and matrix viscosities, which are all
168 known with large uncertainties²⁵. The melt fractions greater than 0.1% that we
169 obtain are therefore plausible. This melt concentration is in overall agreement
170 with the range of estimates derived from electromagnetic studies²⁷ of the LVZ,
171 which often propose even larger values (see Methods). A structure made of
172 magma-rich sills embedded in a meltless mantle might also be mapped by
173 tomography as an average medium with moderate melt content. This partially
174 molten layered model has been proposed for the northwest Pacific and Philippine
175 plates², but it could extend more generally to the entire oceanic LVZ. This layering
176 would explain both the radial anisotropy observed within the oceanic LVZ²⁸ and
177 the sharp velocity and viscosity contrasts at the lithosphere-asthenosphere
178 boundary².

179 Finally, the amount of melt exhibits a very peculiar relation with the plate
180 velocity²⁹ expressed in a no-net-rotation reference frame (**Figure 4**). The melt
181 fraction in the asthenosphere, abruptly increases by a factor close to 2 when the

182 velocity is larger than 4 cm yr⁻¹. This variation is very similar to the observed
 183 variation of azimuthal anisotropy with present-day plate motion⁸. Our results
 184 suggest that plate-scale crystal alignment beneath fast-moving plates is associated
 185 with a greater amount of melt. This requires either that melt facilitates
 186 deformation⁹ or that deformation favours melt retention in the LVZ², or both. In
 187 any case, the small amount of melt observed beneath large swaths of the oceanic
 188 LVZ is likely to significantly decrease viscosity (by one to two orders of
 189 magnitude⁹, see Methods) and to play a significant role in the decoupling of
 190 tectonic plates from the mantle.

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270

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 272 paper. E.D. wrote the codes for the interpretation of the seismic models and wrote
 273 the drafts of the manuscript. E.D. wrote the tomography code for V_s , Y.R. adapted
 274 this code for Q_s . T.B. contributed to the design of the figures and to the writing of
 275 the manuscript. Y.R. developed preliminary codes for interpreting the seismic
 276 models, contributed to all mineralogical aspects and to the writing of the
 277 manuscript. S.D realized the tests of the effect of composition and contributed to
 278 the writing of the revised manuscript.

279

280 **Fig. 1: Shear velocity and attenuation in the upper mantle.** Panels a-c-e-g:
 281 perturbations in shear wave velocities from DR2020s shown in percent with respect to a
 282 mean value V_{ref} given in km s^{-1} above the colour scale. Panels b-d-f-h: maps of our Q_s
 283 tomographic model QsADR17 at different depths in the upper mantle. Q_s is plotted with a
 284 logarithmic scale. Its geometric average is given above the colour scale. Hotspot locations
 285 are shown with black triangles.

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289 **Fig. 2: Scatter plot depicting the observed shear attenuation as a function of shear**
 290 **velocity compared with theoretical predictions.** The velocities (from DR2020s) and
 291 attenuation values (from QsADR17) are plotted at 100 km (panel a), 150 km (panel b) and
 292 200 km depth (panel c). Each dot corresponds to a geographical location. The dark blue
 293 and light blue curves are the theoretical predictions assuming a pyrolitic composition³⁴ in
 294 the absence of melt using the anelasticity models of Takei⁴ and Jackson et al.¹³,
 295 respectively. The upper colour scale indicates the amount of melt in percent required to
 296 explain our observations using Takei⁴'s model. Adding melt enables to lower the predicted
 297 velocity without changing Q_s . The lower grey scale indicates the misfit in percent between
 298 theory and observations, in regions where V_s is too high and cannot be reconciled with
 299 model predictions assuming a pyrolitic mantle.

300

301

302 **Fig. 3: Melt content at different depths in the upper mantle.** This melt content is
 303 derived from the joint interpretation of QsADR17 and DR2020s. The colour coding is
 304 identical to Figure 2. Melt content in percent is indicated with warm colour from ivory
 305 (0% melt) to brown (0.4 to 0.7% melt). The grey scale indicates the misfit in percent
 306 between theory and observations, in regions where V_s is too high compared with
 307 predictions. Hotspots locations are indicated with black triangles. The blue and green stars
 308 indicate the location of the NoMELT and MELT experiments, respectively.

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312 **Fig. 4: Percentage of melt at different depths as a function of absolute plate velocity.**
313 (The velocities²⁹ are expressed in a no-net-rotation reference frame. The percentage of
314 melt is averaged for all geographical points with similar plate velocities, using a sliding
315 window of ± 2 cm yr⁻¹. The amount of melt increases significantly in the asthenosphere
316 (100-200 km) for plate velocities between 3 and 5 cm yr⁻¹. This result links with previous
317 observations that only plates moving faster than 4 cm yr⁻¹ can organize the flow at large
318 scale in the underlying asthenosphere⁸, suggesting that melt reduces viscosity⁹ and
319 facilitates large-scale crystal alignment.
320

321 **Methods**

322 **Qs and Vs tomographic models.** Figure 1 presents maps of DR2020s, a new
323 global Vs model and QsADR17 a recent global Qs model³⁰. Both models are built
324 from the same massive Rayleigh wave measurements³¹, and are obtained from a
325 similar tomographic procedure. The first step is an automated waveform inversion
326 approach which was applied to approximately 375,000 Rayleigh seismograms³¹.
327 From a single surface wave seismogram, the waveform inversion extracts
328 simultaneously a path-average depth-dependent shear velocity profile, Vs, and
329 quality factor, Qs. By jointly interpreting the amplitude and phase of each
330 waveform, we ensure that the interplay between Vs and Qs is accounted for, and
331 that the shear quality factor and velocity profiles are constrained within the same
332 period range. The waveform analysis is performed in the period range 50-250 s
333 and accounts for the fundamental and up to the 5th higher mode of Rayleigh
334 waves, thus ensuring a good depth resolution for Vs and Qs from 50 km depth
335 down to the transition zone. It is a non-linear iterative process, which also
336 produces frequency-dependent phase velocity and attenuation curves compatible
337 with the recorded waveform. The effect of physical dispersion due to attenuation is
338 accounted for in the modelling.

339 The second step is a regionalization of the 1D path-average models. DR2020s is
340 obtained from the regionalization at each depth of the path-average Vs models
341 using a continuous regionalization approach³¹. This tomographic inversion yields
342 3D absolute velocities. QADR17, a global model of Rayleigh-wave attenuation was
343 obtained after adapting the same regionalization approach to our dataset of
344 Rayleigh wave attenuation curves, parameterized as $\ln(Q)$ ³². The regionalisation of
345 the path-average attenuation curves accounts for frequency-dependent effects like
346 focussing-defocussing, which can have important effects on the amplitude of

347 Rayleigh waves. The logarithmic parameterization brings the distribution of the
348 quality factor dataset close to a Gaussian, allows the large variations of Q
349 documented by local seismic studies and guarantees to avoid negative attenuation
350 values in the inverted model. The horizontal smoothing in DR2020s and QADR17
351 is determined by a Gaussian a priori covariance function controlled by an angular
352 correlation length. Adenis et al.³² chose a conservative value of 10° (meaning that
353 the Q model is resolved accurately up to spherical harmonic 12). Debayle and
354 Ricard³¹ used a shorter correlation length of 3.6° for their V_s model DR2012. We
355 re-inverted their dataset using a correlation length of 10° in order to obtain a S_v -
356 wave tomographic model at the same horizontal resolution and vertical smoothing
357 than QsADR17. To minimize biases due to un-modelled radial anisotropy, we
358 computed the isotropic Voigt average of our S_v model and the S_h model obtained
359 by adding the radial anisotropy of PREM to our S_v model. The resulting isotropic
360 model DR2020s is plotted in **Figure 1**. Using other 1D or 3D models of radial
361 anisotropy does not affect our conclusions, as discussed below. QsADR17 was
362 obtained from the inversion at depth of QADR17³⁰, using the same vertical
363 smoothing as for DR2020s. DR2020s and QsADR17 are therefore consistent since
364 they are derived from the simultaneous inversion of the same waveforms and
365 inverted using the same regionalization approach with the same Gaussian filtering.

366 **Prediction of Q_s and V_s using a temperature-dependent model**⁴. The dark blue
367 curve in **Figure 2** is obtained by predicting V_s and Q_s at each depth for a range of
368 temperatures, using a temperature-dependent anelasticity model⁴ and assuming a
369 pyrolitic composition¹⁶. This model parameterizes the relaxation spectrum with a
370 small number of variables determined from experimental data. It accounts for a
371 monotonic background spectrum plus a broad temperature-dependent absorption
372 peak in the seismic frequency band, whose amplitude and width increase below
373 the solidus. This produces a significant enhancement of polycrystal anelasticity

374 before melting. This pre-melting effect induces large Vs reductions under the
 375 solidus. We choose the solidus according to Hirschman³³. We use the model
 376 parameters given in Tables 1 and 2 of Takei⁴, except for the unrelaxed shear
 377 modulus μ_U . We estimate μ_U for a pyrolitic mantle using the mineralogic phase
 378 diagram computed by the Perple X software³⁴. In **Figures S15-S16** we show the
 379 results obtained with the shear modulus of Takei⁴, which reduces the amount of
 380 melt in oceanic regions but fails to explain our $\ln(Q_s)$ -Vs observations beneath old
 381 oceanic basins at 100 km depth. It also significantly increases the misfit beneath
 382 continents.

383 **Prediction of Q_s and Vs from the experimental results of Jackson et al.**¹³. The
 384 light blue curve in **Figure 2** is obtained by a two-step process. First, we use
 385 experimental results on melt-free polycrystalline olivine to predict a theoretical
 386 quality factor Q_t . Computations are performed in the temperature range 800-1800
 387 K and the pressure range 1.49-12.99 GPa, corresponding to the depth range 50-
 388 390 km. We use the following relation¹³ :

$$389 \quad Q_t^{-1}(T_0, T, P, d) = A (T_0 d^{-1} \exp(-(E + PV)/RT))^{\alpha} \quad \text{Eq. A.1}$$

390 In this formulation, Q_t^{-1} at period T_0 depends on the temperature T , the pressure P
 391 and to a lesser extent the grain size d . R is the gas constant, E is the activation
 392 energy and V is the activation volume. **Table S1** summarizes the values of the
 393 different parameters of Eq. A.1. In the range of values compatible with
 394 experiments¹³, we choose $T_0=100$ s, the average period of our long period Rayleigh
 395 waves and $d=0.01$ m. The theoretical curves corresponding to Eq. A1 are shown in
 396 **Figure S9**, with the effect of changing grain size.

397 Second, we use Perple X³⁴ to estimate an isotropic Vs for a pyrolitic model¹⁶. We
 398 compute Vs for the same temperatures and pressures as Q_t . Perple_X produces

399 unrelaxed S-wave velocities at infinite frequency, above the absorption band of
 400 seismic attenuation, while our long period (>50 s) seismic surface waves are likely
 401 to see relaxed velocities. We therefore correct for the effect of anelasticity on V_s^{35} :

$$402 \quad V_s^t = V_{s\infty} \left(1 - \frac{Q_t^{-1}(T_0, T, P, d)}{2 \tan(\pi\alpha/2)} \right) \quad \text{Eq. A.2}$$

403 where $V_{s\infty}$ is the unrelaxed velocity and V_s^t is the velocity corrected from the effect
 404 of attenuation. According to Eq. A2, the stronger the attenuation, the stronger the
 405 velocity reduction is. The light blue curve in **Figure 2** is finally obtained by plotting
 406 at a given depth the obtained $\ln(Q_t)$ as a function of V_s^t for each temperature.

407 **Comparison between our observations and the theoretical relations.** At each
 408 geographical location and depth, we extract the observed value of Q_s in QsADR17
 409 and V_s in DR2020s and we compare our observations with the chosen theoretical
 410 relation. We consider that a theoretical curve is compatible with a given V_s - Q_s
 411 observation if it falls within $\pm 1\%$ and $\pm 10\%$ of the V_s and $\ln(Q_s)$ observations,
 412 respectively (**Figure 2**). These errors are on the conservative side of recent
 413 estimates^{36,37}. If the theoretical relation cannot explain our observations, two
 414 situations can arise: 1) the observed V_s is lower than the theoretical V_s^t . In this case
 415 melt can be added to reduce V_s^t . This is done assuming melt has no effect on Q_s ; 2)
 416 the observed V_s is higher than V_s^t . This is mostly the case under continents and can
 417 be explained by the depletion of the pyrolitic mantle increasing the theoretical
 418 V_s^{t18} .

419 **Effect of grain size.** Grain sizes in the shallow upper mantle are likely 1-20 mm as
 420 observed in upper mantle xenoliths, harzburgite and dunite bodies¹³. The effect of
 421 grain size on Eq. A1 is shown in **Figure S9**. The blue areas around the theoretical
 422 curves cover the influence of grain size from 1 mm (bottom dot curve) to an upper
 423 bound of 100 mm (upper dashed curve). For a given value of Q_s , increasing the

424 grain size increases the temperature and therefore decreases the velocity. As melt
425 is required for explaining slow velocities, increasing grain size will decrease the
426 amount of melt required to explain our observations. Using the equation A1
427 proposed by Jackson, we show in **Figure S12** melt distributions obtained for
428 $d=0.01$ m and in **Figure S17** those for the extreme value $d=0.1$ m, which minimizes
429 the amount of melt required in the LVZ. Maps in **Figure S17** are similar to those of
430 **Figure S12** with a smaller amount of melt. However, in both cases, amounts of
431 melt are larger than for our preferred model using the equations of Takei⁴ (**Figure**
432 **3**).

433 **Amounts of melt.** Although the amount of melt present in the LVZ depends on the
434 choice of the anelasticity model and on the effect of melt on V_s , we show in **Figures**
435 **S12, S16, S17, S18** that different choices of parameters yields similar maps and
436 that melt fractions away from source regions always stay below 1%, in agreement
437 with the partially molten asthenosphere model². We note however that our
438 preferred anelasticity model⁴ generally explains our V_s and $\ln(Q_s)$ observations
439 without the need of partial melt above and below the asthenosphere (**Figure 3**).
440 This is therefore a conservative choice, able to reconcile our V_s and $\ln(Q_s)$
441 observations in most of the upper mantle, while minimizing the amount of melt
442 required in the LVZ.

443 **Effect of radial anisotropy.** We used the anisotropic parameter $\xi=(V_{sh}/V_{sv})^2$ of
444 PREM to convert our S_v observations into an isotropic model, DR2020s. We
445 checked that 3D models of ξ obtained from anisotropic models like SEMUM2³⁸ or
446 S362ANI³⁹ would not affect our main conclusions. As an example, **Figure S1** is
447 similar to **Figure 3** but based on our S_v observations corrected with the
448 anisotropic ξ parameters of SEMUM2 (**Figure S2**). Only minor changes are
449 observed (for example in the Pacific or Indian oceans at 150 and 200 km), but

450 none that would change the interpretations of this paper. The lateral variations of ξ
 451 in SEMUM2 (**Figure S2**) are too small to affect our results. To confirm this point,
 452 we computed the anisotropy ξ that would be needed to account for the discrepancy
 453 between observed Qs and Vs, without invoking mantle depletion or partial melt
 454 (**Figure S3**). Comparison of this figure with **Figure S2** shows that neither the
 455 patterns nor the amplitude would fit (a 3 to 4 times larger anisotropy than in
 456 SEMUM2 would be needed to explain our observations). The anisotropy therefore
 457 does not affect our conclusions.

458 **Elastically accommodated grain boundary sliding (EAGBS) hypothesis.**

459 Anelastic relaxation caused by EAGBS can produce a sharp velocity reduction and
 460 may explain the LAB and the mid-lithosphere discontinuity observed beneath
 461 continents⁴⁰. We tested the EAGBS hypothesis as an alternative explanation to
 462 partial melting for reconciling Vs and Qs in the LVZ. EAGBS is characterized by a
 463 sharp attenuation peak at a characteristic frequency ω_{EAGBS} , followed by a diffused
 464 absorption band at lower frequencies. We first compute the characteristic
 465 frequency for the transition between unrelaxed and relaxed shear moduli. It is
 466 given by Karato et al.,⁴⁰ :

$$467 \quad \omega_{EAGBS} = A \cdot d^{-1} \cdot \exp\left(-\frac{E^* + PV^*}{R.T}\right) \cdot \left[1 + \left(\frac{C_w}{C_{w0}}\right)^r\right] \quad \text{Eq. A3}$$

468 where C_w is the water content in wt%, $A=2.3 \times 10^{-14} \text{ m s}^{-1}$, $C_{w0}=10^{-4} \text{ wt}\%$, $r=1$,
 469 $E^*=350 \text{ kJ mol}^{-1}$, $V^*=10^{-6} \text{ m}^3$, $d=5 \times 10^{-3} \text{ m}$. Following Cobden et al.⁷ we assume a dry
 470 mantle with $C_w=10^{-6} \text{ wt}\%$.

471 Surface waves analysed at frequencies lower than ω_{EAGBS} should see relaxed
 472 velocities within the absorption band (see Fig. SI-2 of Karato et al.⁴⁰), providing
 473 depth and temperature are within the activation ranges of EAGBS. EAGBS is
 474 activated between 60 and 160 km depth in the temperature range 900-1,350 K. We

475 consider that EAGBS may affect the interpretations of our long period (>40 s) Qs
 476 and Vs maps when $\omega_{EAGBS} > 0.025$ Hz and, from figure 4 of Karato et al.⁴⁰, in the
 477 following temperature ranges for the depths of our tomographic inversion:

- 478 • 70 km: 930-1,140 K
- 479 • 90 km: 980-1,180 K
- 480 • 100 km: 1,010-1,220 K
- 481 • 125 km: 1,060-1,270 K
- 482 • 150 km: 1,130-1,330 K

483 Regions where EAGBS can apply are displayed in green in **Figure S4**. These
 484 regions correspond mostly to continental areas. Assuming that EAGBS correction
 485 applies, Karato et al.⁴⁰ list three alternative formulations for the velocity reduction.
 486 The Ghahremani⁴¹ equation, which produces the smallest velocity reduction is :

$$487 \quad V_s^{eagbs} = V_s^t \cdot \sqrt{\frac{0.86-0.83\nu}{1-\nu} \cdot \frac{1+\nu}{1.14+0.83\nu}} \quad \text{Eq. A.4}$$

488 where ν is the Poisson's ratio (~ 0.3 for upper mantle minerals). Regions where
 489 this can reconcile our Vs and Qs observations are displayed in red in **Figure S4**.
 490 The two other formulations give too large velocity reduction. **Figure S4** shows that
 491 EAGBS fails at reconciling our long period observations, excepted in a few localized
 492 regions (in red). This does not preclude EAGBS to play a significant role in
 493 explaining the sharp velocity reduction associated with the LAB, which cannot be
 494 resolved by long period surface waves. However, the strong velocity reduction
 495 observed within the LVZ requires another mechanism.

496 **Effect of composition.** Compositional heterogeneities exist in the upper mantle,
 497 likely in the domain limited by pyrolite, harzburgite and pyroxenite⁴². Here, we
 498 check the effect of changing the composition from pyrolite to these end-members
 499 using *Perple X*³⁴ to compute the velocities. Pyroxenites have variable bulk

500 compositions⁴³ and we consider both a silica-deficient composition (see, Stixrude
 501 and Lithgow-Bertelloni⁴⁴, Table 1) and a silica-excess composition similar to that
 502 of a subducted basalt (see, Xu et al.,¹⁶, Table 1). A silica-excess pyroxenite would
 503 only be faster than pyrolite after eclogitization, below 150 km depth (**Figure S5**,
 504 orange curve). At 100 km depth, a mantle made entirely of silica-excess pyroxenite
 505 would be too slow to explain the observed velocities (**Figure S6**). At larger depth,
 506 after eclogitization, it would lead to large melt fractions (**Figure S6**). The silica-
 507 deficient pyroxenite and the harzburgite are characterized at asthenospheric
 508 conditions by much faster or comparable velocities than pyrolite (**Figure S5**) and
 509 hence cannot explain the low velocities that we attribute to the presence of melt.
 510 Harzburgite would give melt fractions and distribution similar to pyrolite (**Figure**
 511 **S7**), while the silica-deficient pyroxenite would lead to higher melt fractions (up
 512 to 1.7%) and a ubiquitous melt presence below 150 km depth (**Figure S8**).
 513 Pyroxenite, whether enriched or depleted in silica, should only be present in a
 514 small proportion (~5%)⁴² that hardly affects the seismic velocities of the
 515 peridotitic mantle. At any rate, large-scale chemical heterogeneities (> 1000 km)
 516 are not likely to exist in the LVZ, where viscosity is small (below 10²⁰ Pa s) and the
 517 convective mantle should be well stirred and mixed. Therefore, our observations
 518 cannot be explained by reasonable variations of compositions in the LVZ.

519 **Possible effect of water.** Although, it has been recently shown that in olivine,
 520 wave-speed and attenuation are insensitive to water¹⁰, some previous studies had
 521 suggested an effect of water quantified following the equation⁴⁵:

$$522 \quad Q_t^{-1}(T_0, T, P, C_{OH}, d) = A \left(T_0 d^{-1} \left(\frac{C_{OH}}{C_{OH(ref)}} \right)^r \exp(-(E + PV)/RT) \right)^\alpha \quad \text{Eq. A.5}$$

523 where C_{OH} is the water concentration, r is a dimensionless constant and all other
 524 parameters are as in Eq. A1. The average H₂O content of Earth's upper mantle is
 525 estimated in the range 50-200 ppm⁴⁶. We test a value close to the upper bound

526 ($C_{OH}=1000H/10^6$ Si, corresponding to 125 ± 75 ppm by weight of water) with $r=1$, a
 527 standard value for moderate water content and $C_{OH(ref)}=50H/10^6$ SI⁷. The
 528 corresponding theoretical curves are shown in blue on **Figure S9**. At fixed
 529 temperature, water increases attenuation. However to keep Q constant, an
 530 increase of water content must be balanced by a decrease of temperature and
 531 therefore by a higher expected velocity V_s^t . Reconciling this higher V_s^t with our
 532 observations requires larger amounts of melt (**Figure S10**). Therefore, if there were
 533 an effect of water on Vs and Qs, it would increase the required amount of melt.

534 **Effect of melt on attenuation.** A small amount of melt might have a large effect on
 535 attenuation⁴⁷ in the case of grain boundary sliding, where a broad dissipation peak
 536 is observed in the seismic frequency band. Body wave studies have reported low
 537 Qs values at short period (1s) with $Q_s \leq 25$ beneath the Juan de Fuca and Gorda
 538 ridges⁴⁸ and $25 \leq Q_s \leq 80$ in the back-arc mantle of Central America, the Marianas
 539 and the Lau Basin⁴⁹. A recent experimental study suggests that 0.2% of melt
 540 produces $Q_s=48$ ³. However such a large effect of melt on Qs is difficult to reconcile
 541 with long period (>30 s) surface waves observations of $Q_s=80$ beneath the fast
 542 spreading southern East Pacific Rise²⁰ where up to 1-2% of melt is expected⁵⁰. If
 543 the attenuation mechanism is melt squirt, then the dissipation peak may lie outside
 544 the seismic frequency band and cause little attenuation. Model calculations⁵¹, long-
 545 period surface waves²⁰ and experiments¹² favour such interpretation. This is also
 546 supported by our long period (100 s) global surface wave tomographic models,
 547 which suggest that strong Vs and Qs reductions are not necessarily correlated^{30,32}.
 548 For these reasons, we assume that in the period range of surface waves (50-250s),
 549 melt does not significantly affect attenuation and we neglect its effect on Qs.

550 **Temperature in the upper mantle.** Our Qs and Vs seismic models are interpreted
 551 using mineral physics results^{4,13}. These experimental results are valid at temperatures

552 larger than 900°C or 1000°C, which are reached at the base of the lithosphere and in the
553 asthenosphere. The predictions of our model can be checked with respect to the well-
554 known thermal behaviour of the cooling oceanic lithosphere^{52,53} although this is
555 certainly pushing the model outside its applicability range. **Figure S19 (panel a)**,
556 displays the temperature variations beneath the Pacific as a function of sea-floor age
557 predicted from our seismic models using experimental results⁴. It demonstrates that
558 even at depths shallower than 100 km, we retrieve the well-known age-dependence of
559 temperature in oceanic regions (we also pick up the cold signal of the West Pacific
560 subductions at old ages). However, our temperature variations between the ridge and an
561 old lithosphere (~100°) are lower than the predictions of the plate-cooling model (250°
562 at 75 km depth, **panel b**). **Panel c** displays the quality factors that would be deduced
563 from the plate-cooling model using the experimental results⁴ used in this paper. In the
564 oceanic lithosphere, Q_s values much greater than 2000 ($\ln(Q_s) > 7.6$) would be
565 predicted, that cannot be retrieved by long period seismology. The limitations of
566 experimental data at low temperatures together with the inability of surface wave
567 seismology to quantify precisely negligible attenuations, explain why we do not
568 interpret results at depths shallower than 100 km. Below the oceanic plates, **Figure S14**,
569 **panel g** shows that our predicted temperatures beneath oceans are perfectly compatible
570 with geodynamic and petrologic expectations. In agreement with our findings, under
571 oceans, the average 1D temperature profile is above the solidus⁴ and the adiabat⁵⁴
572 in the depth range 100-200 km, where we predict partial melting. This overshoot
573 of the temperature, above the adiabat, is indeed found in all numerical simulations
574 of mantle convection⁵⁵. Under continents, the temperature appears to be below the
575 solidus⁴ and the adiabat⁵⁴. It reaches the adiabat around 250 km depth (**Figure**
576 **S14, panel g**).

577 **Compatibility of our results with electrical conductivity.** Our results are
578 consistent with the interpretation of a number of recent studies of electrical

579 conductivity. In the depth range 100-150 km, the joint interpretation of electrical
580 and seismic data⁵⁶ requires 0.3-2.5% of melt beneath the mid-Atlantic ridge, 1% or
581 less melt beneath Hawaiï and less than 5% of melt beneath the East-pacific rise, in
582 good agreement with **Figure 3**. We also confirm the absence of melt in the region
583 of the NoMELT experiment¹⁹, in agreement with conductivity data in this region²⁵.
584 Previous experimental results suggest that between 0.3 and 2% of hydrous
585 basaltic melt can account for the observed electrical conductivity in the LVZ²⁷. A
586 more recent study⁵⁷ has refined these results by simultaneously measuring wave
587 velocity and electrical conductivity on a simplified partial melt analogue. They
588 conclude that the low velocity zone away from spreading ridges can be explained
589 by 0.3-0.8% volatile-bearing melt, the upper bound of our observations.

590 **Implications for viscosity.** The variation of viscosity as a function of melt content
591 occurs in two steps⁹. The onset of melting brings already a significant effect on
592 viscosity when a connected network of melt tubules is formed. The viscosity η is
593 expected to decrease by one or two orders of magnitude before the melt fraction ϕ
594 reaches 0.1%. For larger melt content, η decreases further⁹ with $d\ln(\eta)/d\phi = -26$
595 but this effect is minor for the low melt content that we observe. Under oceanic
596 plates where ϕ is around 0.3%, η should be one or two orders of magnitude lower
597 than under continents.

598 **Data availability:** The dataset generated during this study (3D Vs, Qs models and
599 melt fraction models) is available as IRIS data products at
600 <https://doi.org/10.17611/dp/emc.2020.dbrdnature.1>.

601 **Code availability:** Numerical modelling codes related with this paper can be
602 downloaded from <https://doi.org/10.17611/dp/emc.2020.dbrdnature.1>. Requests
603 about the numerical modelling codes associated with this paper should be sent to

604 Eric.Debayle@ens-lyon.fr. Most figures were created with open software GMT

605 4.5.13.

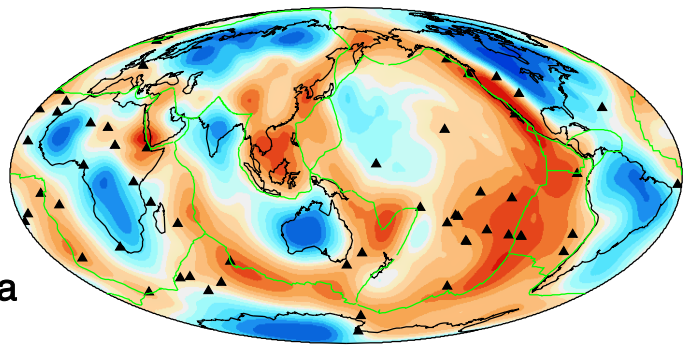
606 **Competing interest:** The authors declare no competing interests.

607 **References from Methods:**

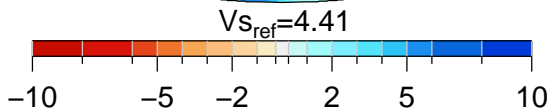
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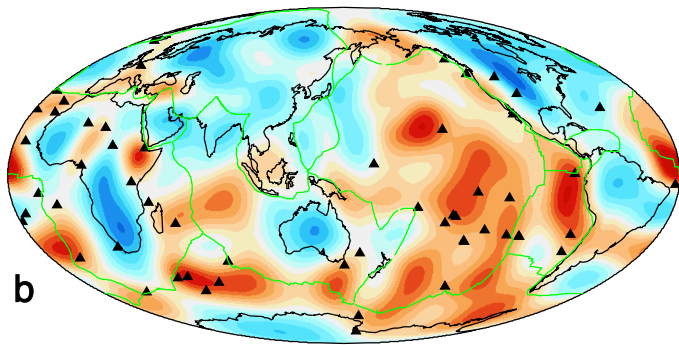
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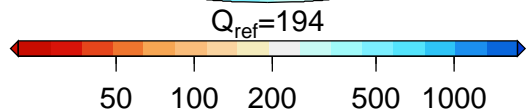
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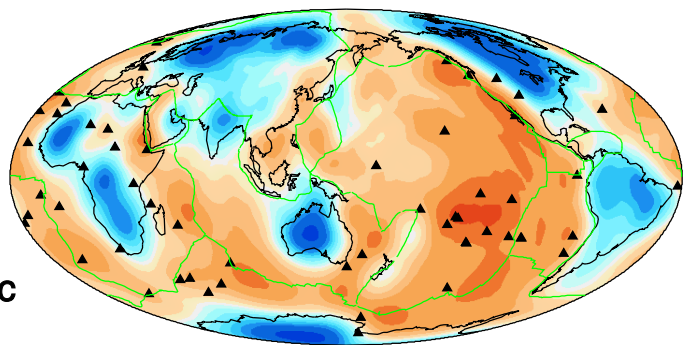
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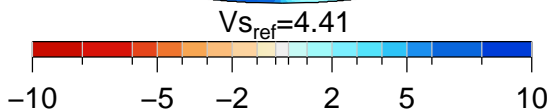
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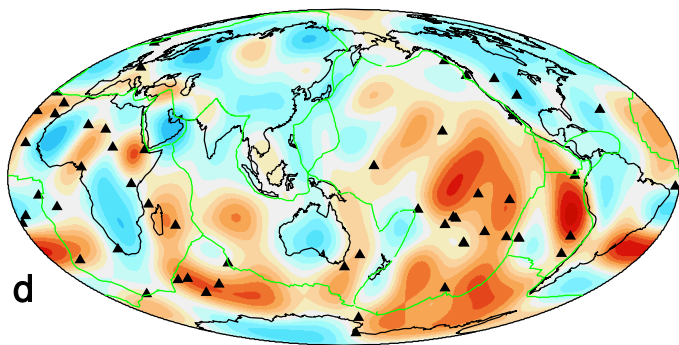
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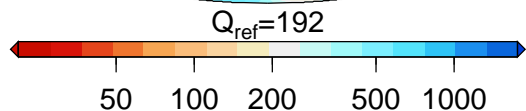
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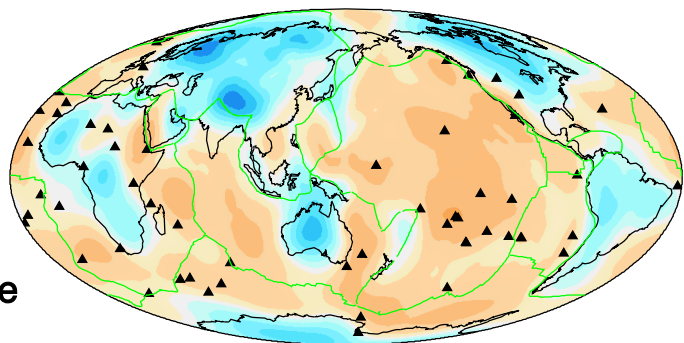
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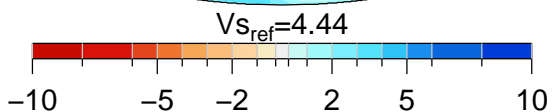
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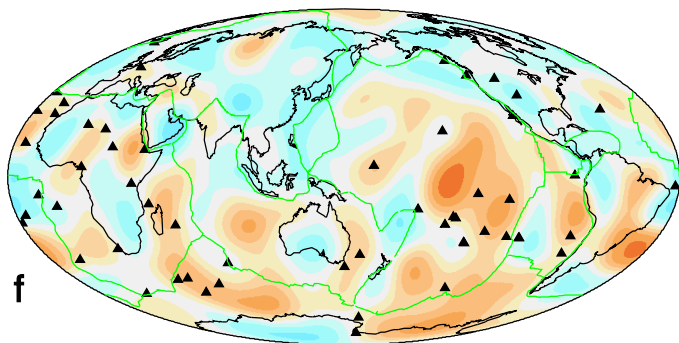
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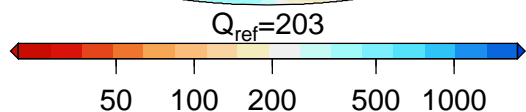
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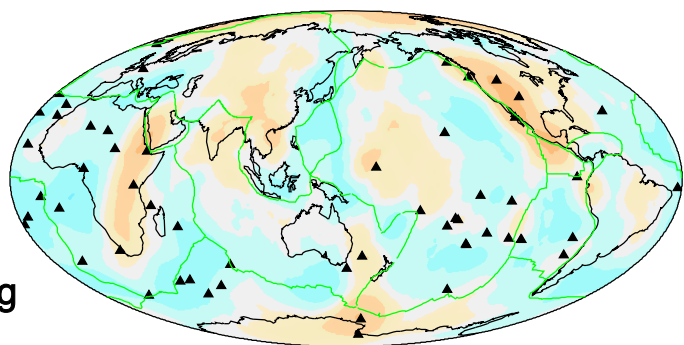
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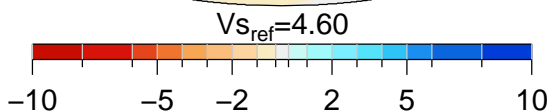
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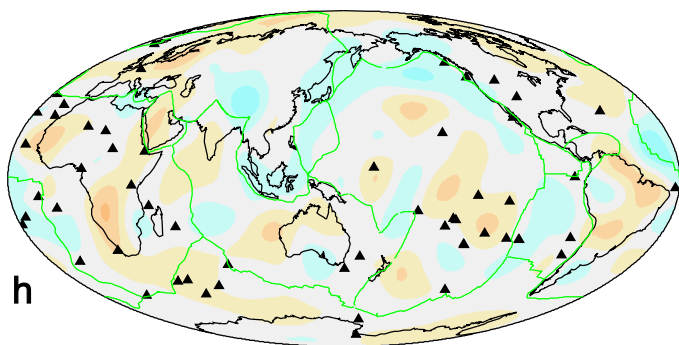
300 km



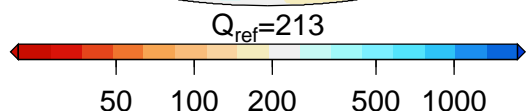
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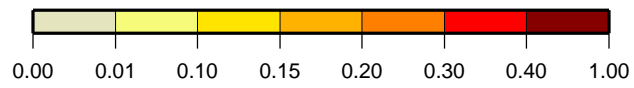
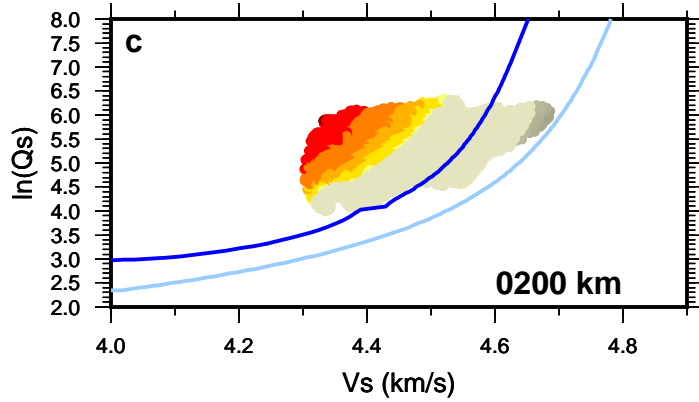
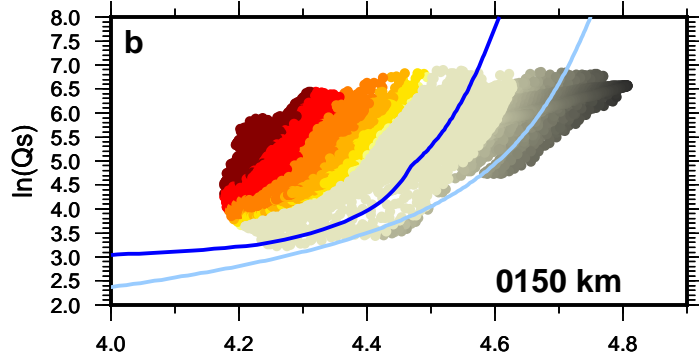
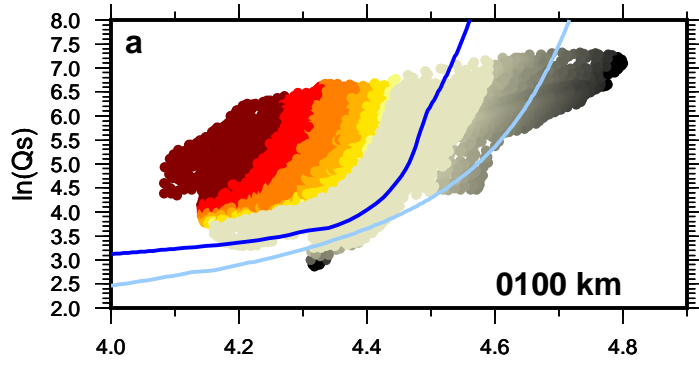


300 km

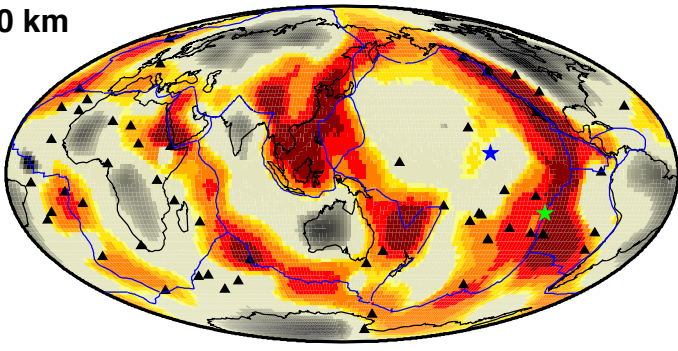


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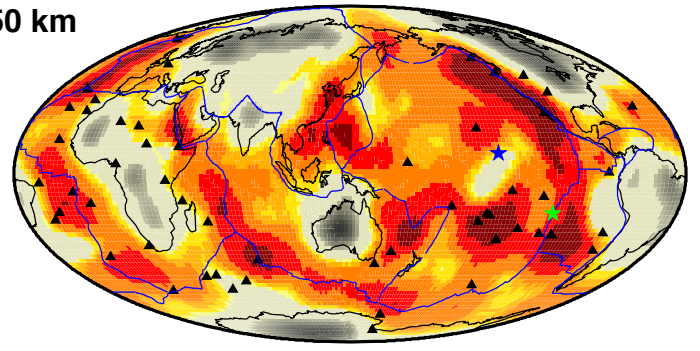




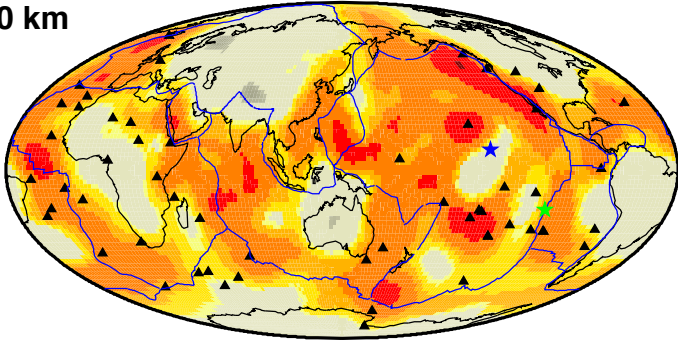
0100 km



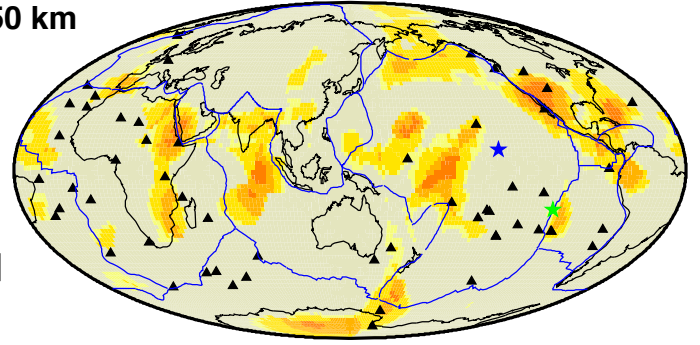
0150 km



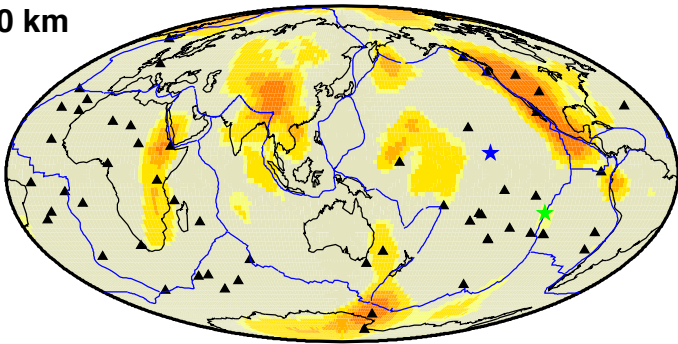
0200 km



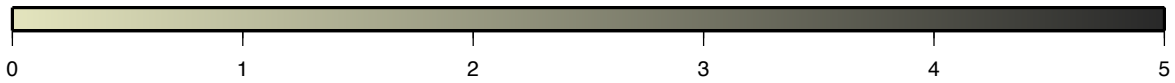
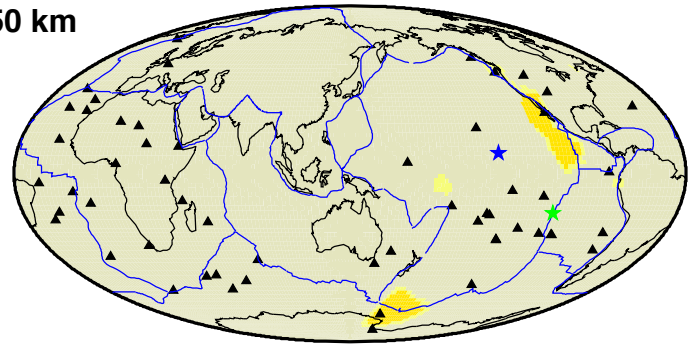
0250 km

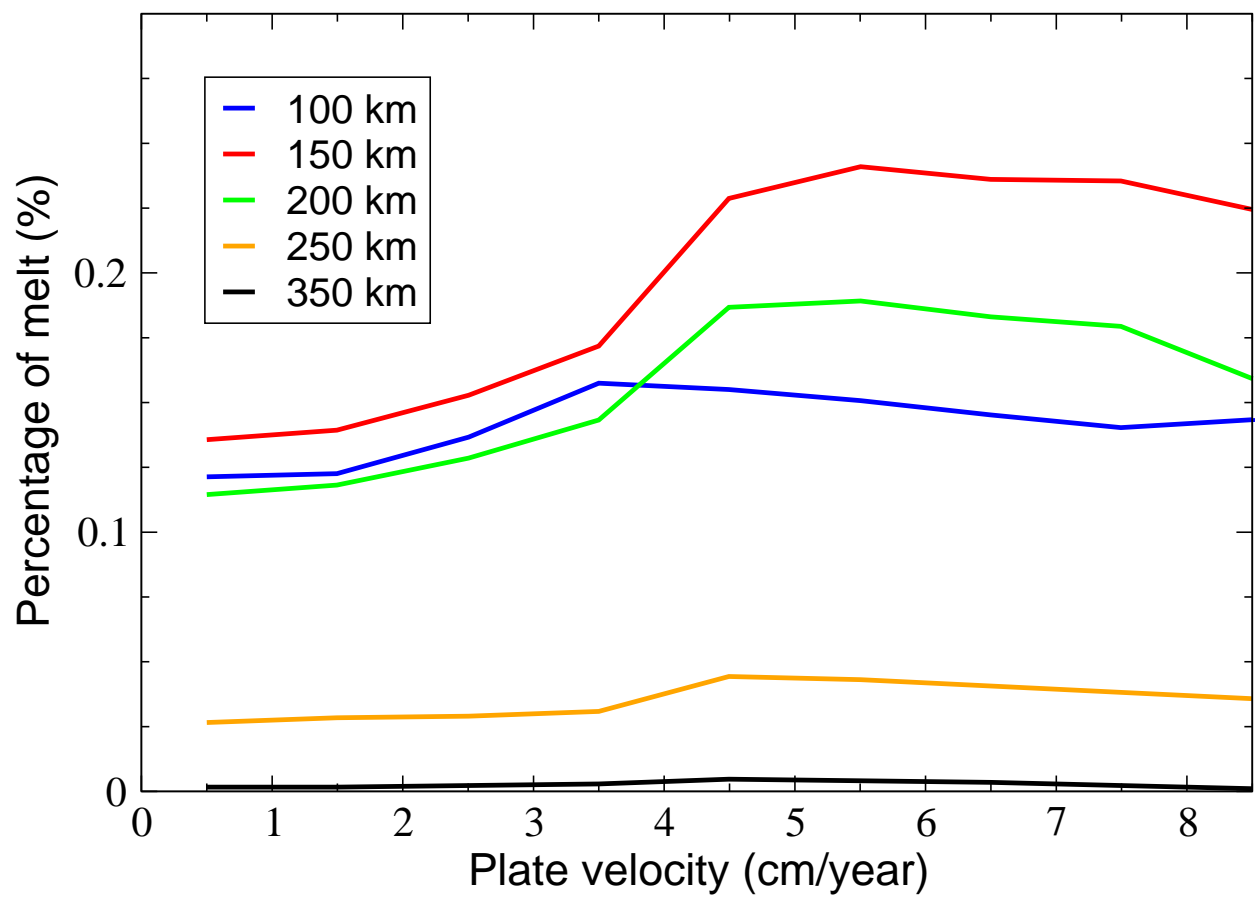


0300 km



0350 km





<i>Symbol</i>	<i>Description</i>	<i>Value</i>	<i>Units</i>
α	exponent	0.26	
A	prefactor	7.5×10^{-2}	$\text{s}^{-\alpha} \mu\text{m}^{\alpha}$
d	grain size	1-100	mm
T_0	period	100	s
E	activation energy	424	kJ/mol
V	activation volume	6×10^{-6}	m^3/mol

Table S1: Reference parameters for Eq.A.1 after Jackson et al.¹

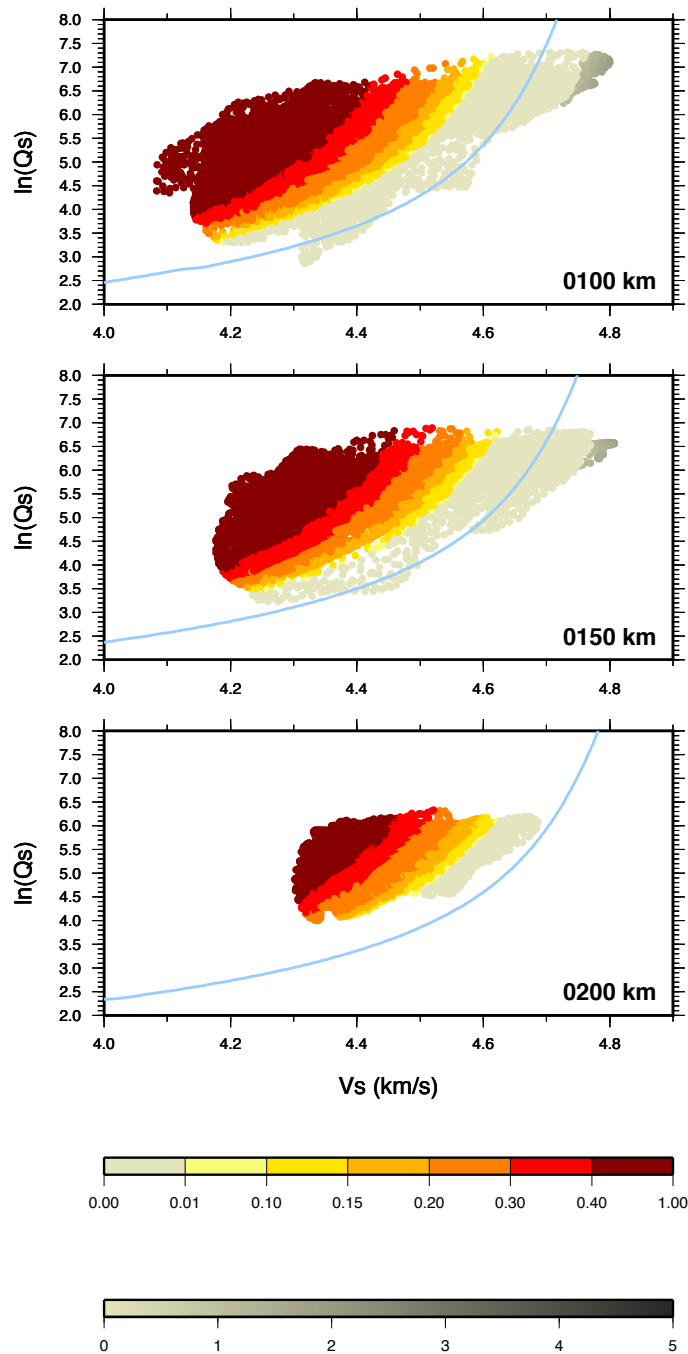


Fig. S1: same as Figure 2 but the color scales indicate the departure from the model of Jackson et al.¹ (light blue curve), for a grain size of 10 mm. The upper color scale indicates the amount of melt in percent required to explain our observations (ivory color from 0 to 0.01% of melt underlines data for which the model can reconcile our Q_s and V_s observations). The lower grey scale indicates the misfit in percent between theory and observations, in regions where V_s is too high and cannot be reconciled with model predictions.

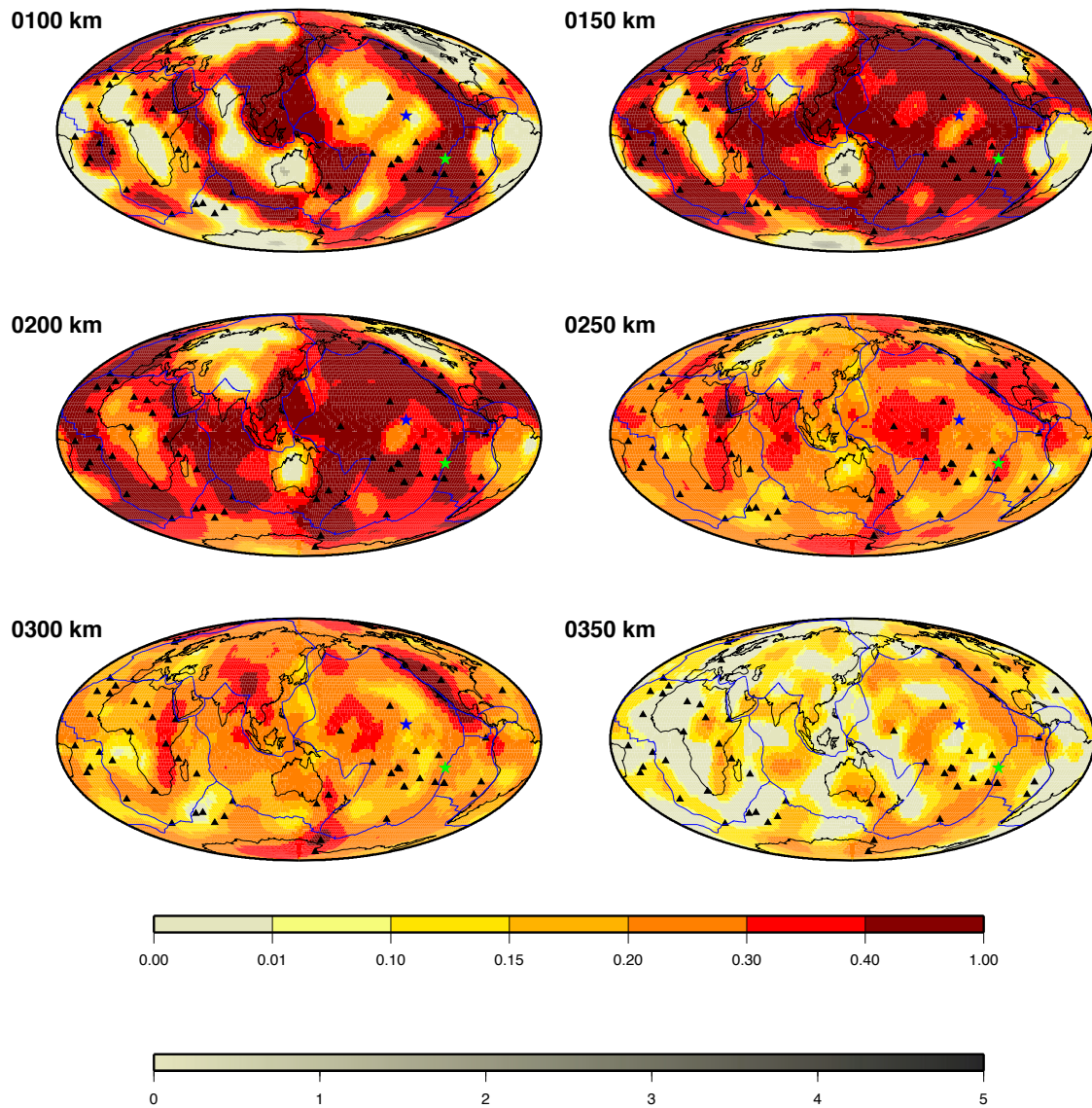


Fig. S2: same as Figure 3 but color scales indicate the departure from the model of Jackson et al.¹, for a grain size of 10 mm. Melt content in percent is indicated with warm colors (upper color scale) from ivory (0% melt) to brown (0.4 to 0.7% melt). The lower grey scale indicates the misfit in percent between the theory and observations, in regions where V_s is too high compared with predictions.

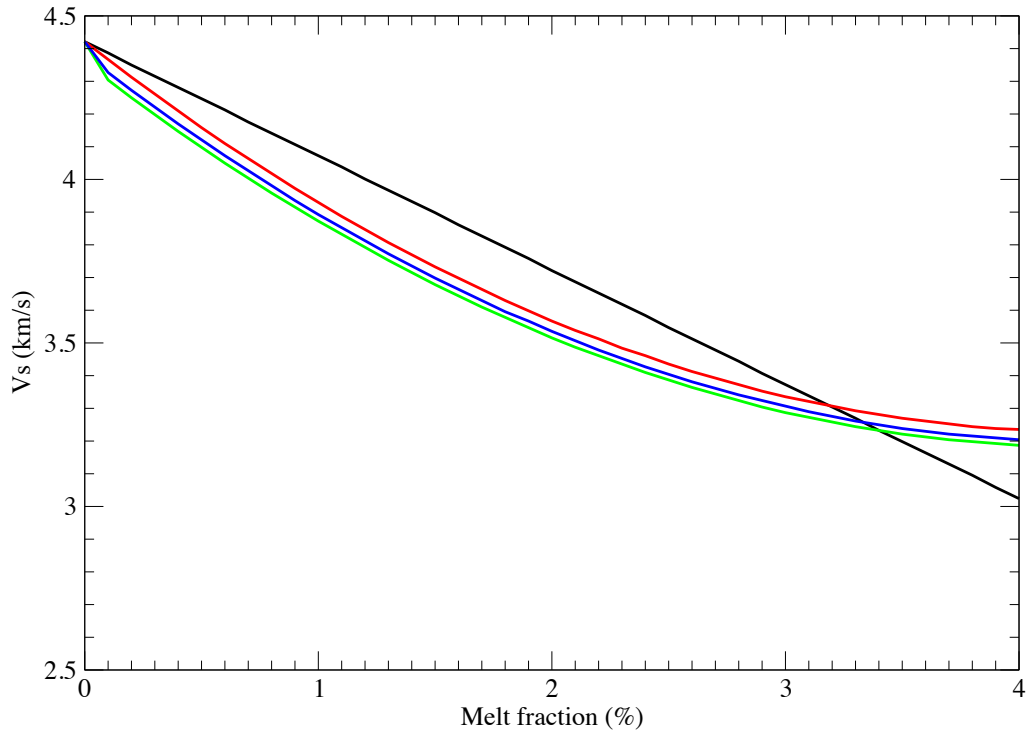


Fig. S3: Dependence of shear velocity on melt fraction φ . The linear V_s reduction of 7.9% per percent of melt² is shown in black from a reference velocity V_{ref} . The polynomial expression ($V_s = 0.065\varphi^2 - 0.5565\varphi + V_{ref}$) derived from experimental results³ is shown in red. The anelastic effect expected for seismic waves at high temperature⁴ is shown assuming $Q_s=80$, the value of PREM⁵ in the asthenosphere, for two values of α (blue curve for $\alpha=0.38$ and green curve for $\alpha=0.26$). For small melt fractions (<1%), a stronger V_s reduction is obtained using the polynomial expression and the choice of α has a small effect.

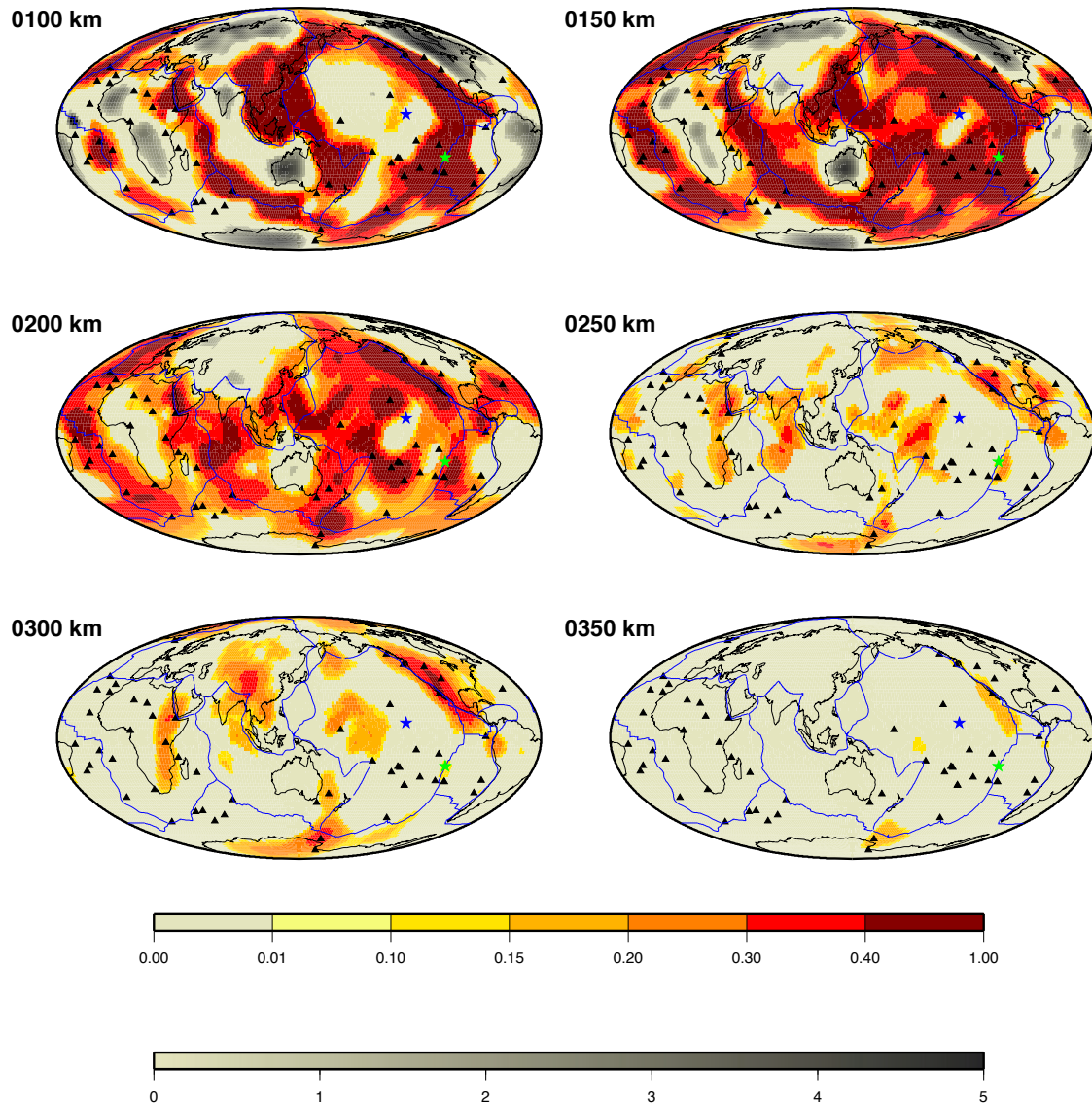


Fig. S4: same as Fig. 3 but using a linear 7.9% V_s reduction per percent of melt² instead of the polynomial expression derived from experimental results³. The upper color scale is slightly modified to allow melt content up to 1% (the maximum value at 100 km depth).

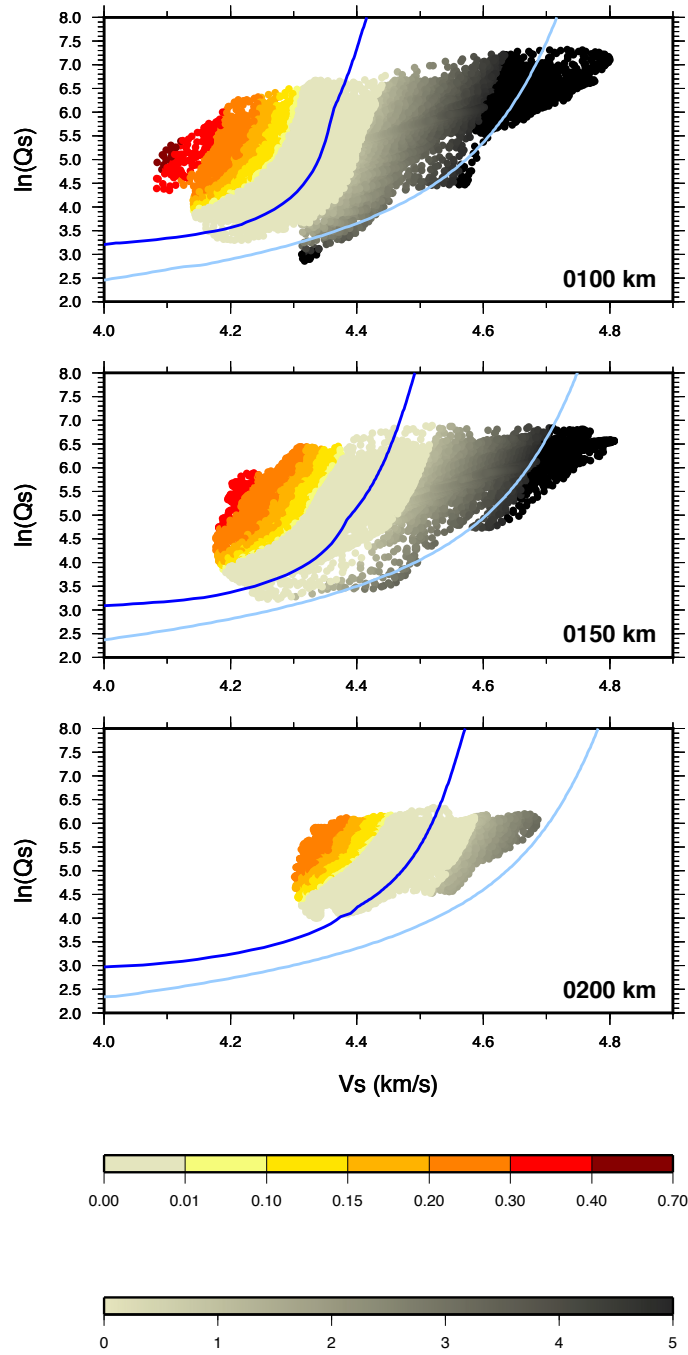


Fig. S5: same as Figure 2 but the unrelaxed shear modulus μ_U needed to compute the temperature-dependent model (dark blue curve) is calculated using parameters proposed for of the temperature model of the Pacific⁶, instead of those deduced from Perple X assuming a pyrolitic composition.

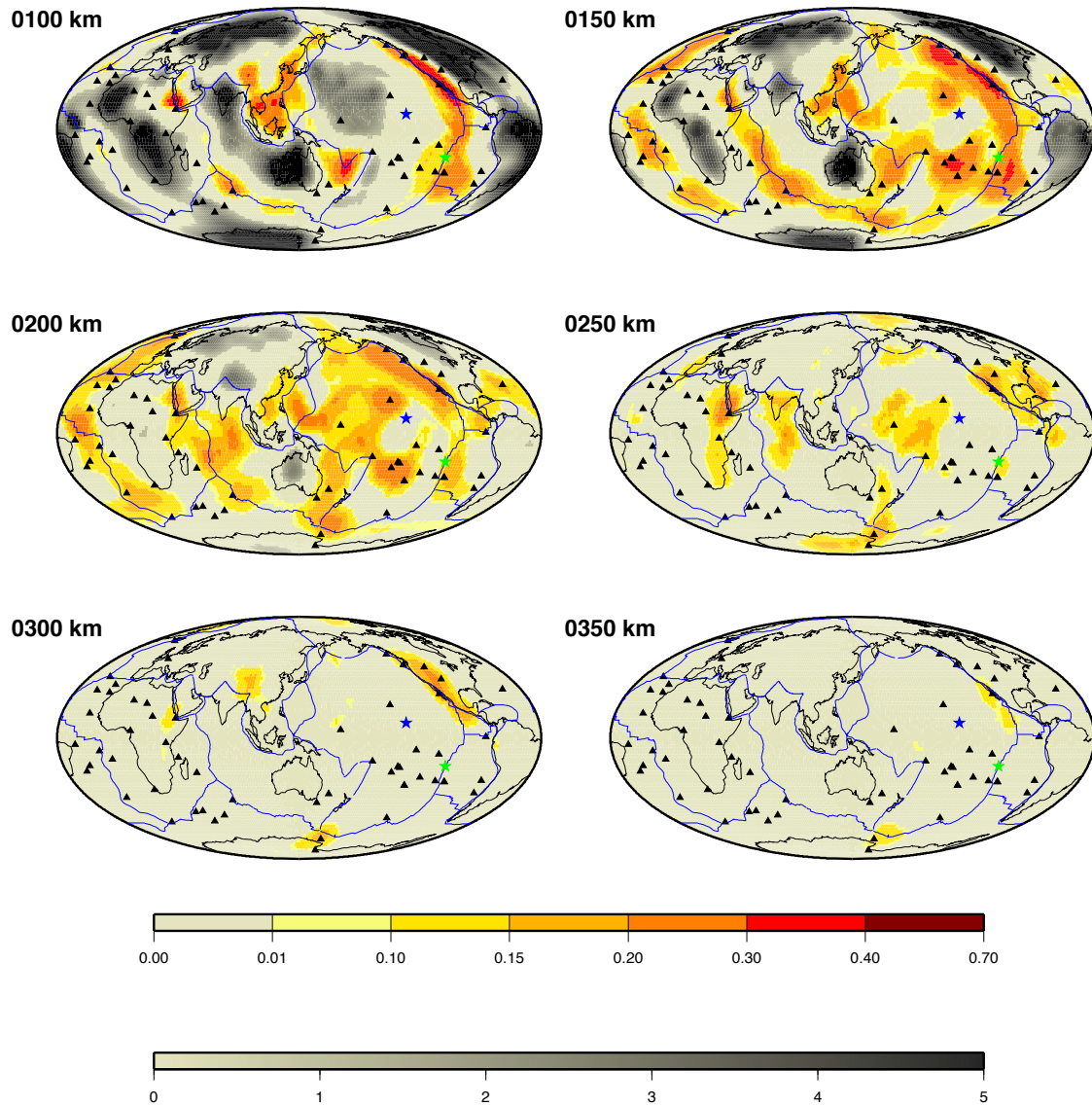


Fig. S6: same as Figure 3 but instead to estimate the unrelaxed shear modulus μ_U using Perple X and a pyrolitic model, we use fitting parameters of the temperature model for the Pacific⁶.

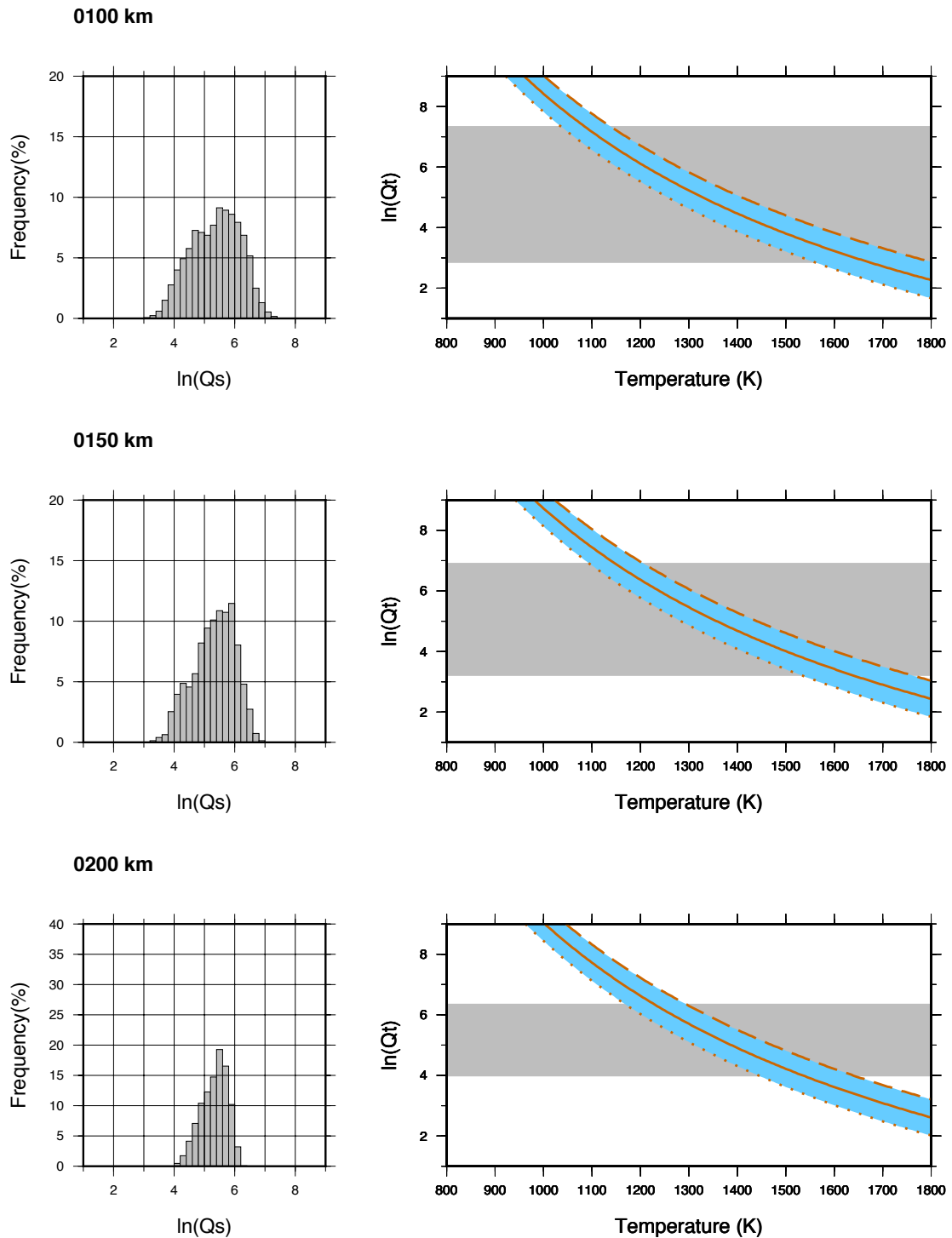


Fig. S7: Left column: Histograms of the distribution of $\ln(Q_s)$ values extracted at 100 (top), 150 (middle) and 200 (bottom) km depth in QsADR17⁷. Right column: at the same depths, theoretical relation between $\ln(Q_t)$ and temperature computed using Eq. A.1 for different grain sizes. The continuous lines in brown are the theoretical curves assuming a grain size of 10 mm. The light blue areas around this curve cover the influence of grain sizes from 1 mm (bottom dotted curve) to 100 mm (upper dashed curve). The shaded grey area shows the range of Q_s variations observed in QsADR17.

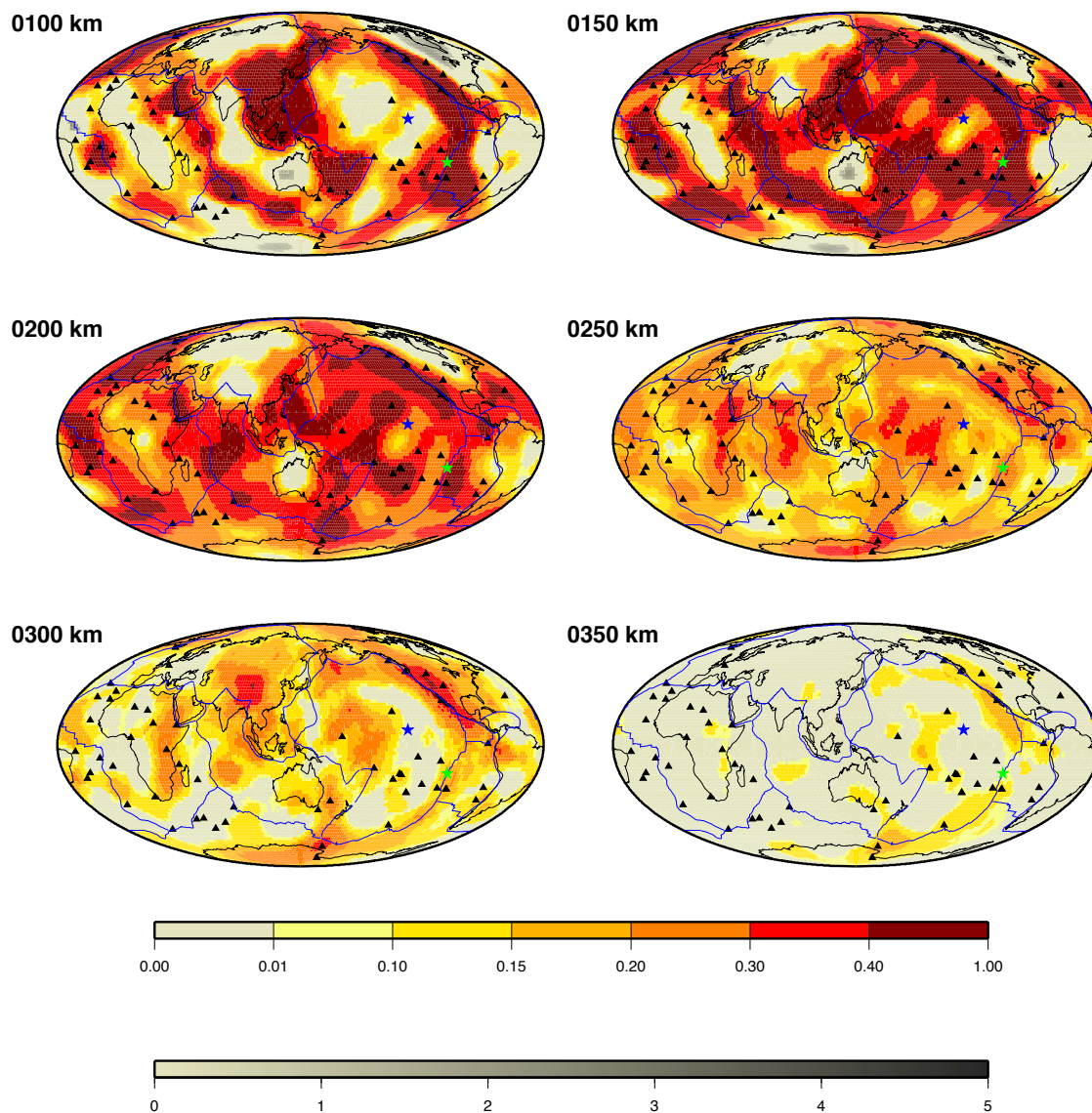


Fig. S8: same as Figure S2 but for a grain size of 100 mm.

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