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1 **Title: Seismic evidence for Earth’s crusty deep mantle**

2

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16

17 **Abstract**

18 Seismic tomography resolves anomalies interpreted as oceanic lithosphere subducted

19 deep into Earth’s lower mantle. However, the fate of the compositionally distinct oceanic

20 crust that is part of the lithosphere is poorly constrained but provides important

21 constraints on mixing processes and the recycling process in the deep Earth. We present

22 high-resolution seismic array analyses of anomalous P-waves sampling the deep mantle,

23 and deterministically locate heterogeneities in the lowermost 300 km of the mantle.

24 Spectral analysis indicates that the dominant scale length of the heterogeneity is 4 to 7
25 km. The heterogeneity distribution varies laterally and radially and heterogeneities are
26 more abundant near the margins of the lowermost mantle Large Low Velocity Provinces
27 (LLVPs), consistent with mantle convection simulations that show elevated
28 accumulations of deeply advected crustal material near the boundaries of thermo-
29 chemical piles. The size and distribution of the observed heterogeneities is consistent
30 with that expected for subducted oceanic crust. These results thus suggest the deep
31 mantle contains an imprint of continued subduction of oceanic crust, stirred by mantle
32 convection and modulated by long lasting thermo-chemical structures. The preferred
33 location of the heterogeneity in the lowermost mantle is consistent with a thermo-
34 chemical origin of the LLVPs. Our observations relate to the mixing behaviour of small
35 length-scale heterogeneity in the deep Earth and indicate that compositional
36 heterogeneities from the subduction process can survive for extended times in the
37 lowermost mantle.

38

39 **1. Introduction**

40 Seismological analyses reveal heterogeneities in Earth's mantle from the surface to the
41 core-mantle boundary (CMB) spanning a wide range of scales. In the upper mantle,
42 seismic tomography shows oceanic lithosphere, on the order of 100s km thick,
43 subducting into the Earth (Grand et al., 1997). The oceanic crust component of the
44 lithosphere is subducted into the mantle at a rate of $\sim 20 \text{ km}^3$ per year at present-day (Li et
45 al., 2016). The crust has been modelled to advect into the lower mantle (Christensen and
46 Hofmann, 1994), and may represent up to 10% of the mass of the mantle from subduction

47 through Earth's history (Hofmann and White, 1982). Geochemical anomalies in ocean
48 island basalts sourced from the deep Earth suggest that oceanic crust is incompletely
49 mixed into the mantle (Stracke et al., 2003). Meanwhile, tomographic images of the
50 lowermost mantle are dominated by two large, 1000s km scale-length, nearly equatorial
51 and antipodal, structures of reduced seismic velocities (e.g. Dziewonski, 1984), both in S-
52 and P-wave velocity (V_s and V_p , respectively); these are surrounded by zones of higher
53 seismic velocities, which are commonly attributed to cooler subduction-related
54 downwellings. The nature and origin of these LLVPs remains enigmatic but may be
55 related to dense thermo-chemical piles (Garnero and McNamara, 2008) possibly
56 consisting of primordial material (Labrosse et al., 2007), products of chemical reactions
57 with the outer core (Knittle and Jeanloz, 1991), or accumulation of subducted oceanic
58 crust (Christensen and Hofmann, 1994). A purely thermal origin of LLVPs has also been
59 advocated (Davies et al., 2015). Geodynamic models indicate that subduction-related
60 currents shape the thermochemical structures into piles that internally convect
61 (McNamara and Zhong, 2005).

62

63 Significantly smaller scale heterogeneity has been inferred from high-frequency (~ 1 Hz)
64 seismic energy trailing (coda) or preceding (precursors) some seismic waves (Shearer,
65 2007), due to scattering from volumetric heterogeneities with scales similar to the
66 dominant seismic wavelength (Cleary and Haddon, 1972) (e.g. of order 10 km in the
67 lowermost mantle for 1 Hz waves). While seismic probes differ in their sensitivities to
68 the scale and depth of scattering heterogeneity, scattered waves help to characterize fine
69 scale mantle heterogeneity (Shearer, 2007). The radial dependence of scattering has been

70 investigated with PKP waves (P waves that go through Earth's core), which indicate the
71 presence of weak (e.g., V_p perturbations, dV_p , of 0.1% RMS), small-scale (6-8 km)
72 heterogeneity distributed throughout the mantle (Mancinelli and Shearer, 2013). These
73 studies present global, radially averaged statistically viable scattering populations but are
74 not able to deterministically locate scattering heterogeneities. Upper mantle regional
75 studies of scattered PP and SS waves (Fig. 1), have deterministically mapped scatterers in
76 subduction zones in the upper and mid-mantle, relating the heterogeneities to subduction
77 processes (Kaneshima and Helffrich, 1998). Lower mantle regional studies using PKP
78 have demonstrated regional deep mantle scattering (Frost et al., 2013, Ma et al., 2016)
79 with strong lateral variations. These lowermost mantle heterogeneities have been
80 attributed to a variety of processes including subduction, plumes, melt processes, and
81 phase transitions. Regional geodynamic models of the upper mantle have demonstrated
82 the role of large-scale convection in generating and manipulating heterogeneity across
83 length-scales (e.g. Korenaga, 2004). While scattering scale-lengths have been previously
84 inferred, the data used are typically band-limited or filtered to high frequency, thereby
85 restricting the constraint on the range of scale lengths that can be deduced. Here we
86 present a seismic probe and method for precise location of scattering heterogeneities near
87 the CMB and simultaneous determination of their dominant scale lengths over a wide
88 spectrum of possibilities.

89

90 We use a scattered form of PKKP which first propagates as a P-wave through the mantle,
91 into the core, and back up to the lower mantle (as a normal PKP wave), and then is
92 scattered in the lowermost mantle depth shell back into the core, and then travels through

93 the mantle to the receiver (Rost and Earle, 2010). We refer to this scattered path as
94 PK•KP, where the dot “•” represents the small portion of the path travelled as P-wave
95 from the CMB up into the lower mantle to the scattering location, then back to the CMB.
96 PK•KP may involve out-of-great circle plane scattering and travel along asymmetric
97 source and receiver paths (Fig. 1). This probe is especially suited for studying lower
98 mantle heterogeneities because PK•KP arrives in a quiet time-distance window for
99 teleseismic data (Fig. 2), it avoids the source-receiver CMB location ambiguity of PKP
100 scattering (Fig. 1d), it allows deterministic identification of the heterogeneity location,
101 and it allows sampling of an extensive volume of the Earth’s mantle (Fig. 3).

102

103 **2. Data and array-processing methods**

104 While past work introduced the feasibility of this phase for deep mantle heterogeneity
105 detection, analysis was limited geographically and in depth (Rost and Earle, 2010). We
106 collect earthquakes with magnitudes larger than 6.0 occurring in a 17-year period (1995-
107 2012) within 0-to-60° epicentral distance from 13 International Monitoring System (IMS)
108 arrays and the Gauribidanur array in India (Fig. 3). While PK•KP could feasibly be
109 observed at greater distances, the upper distance limit was imposed to avoid possible
110 contamination from other core waves (Fig. 2). Some earthquakes were detected by
111 multiple arrays, resulting in 2355 earthquake-array pairs from 1095 earthquakes. Every
112 source-array pair allows detection of energy related to multiple scattering heterogeneities.
113 This dataset samples 78% of the surface area of the CMB and lowermost mantle (Fig.
114 3e), including the southern hemisphere, which was poorly sampled in previous studies
115 and provides us with the to-date best sampling for lower mantle heterogeneity.

116

117 Small aperture seismic arrays are effective for deep Earth studies, enhancing low
118 amplitude, yet coherent energy, relative to incoherent background noise. We analyse a
119 110 s time window following the first theoretical PK•KP arrival, allowing investigation
120 of scattering within the lowermost 320 km of the mantle. To identify scattered signals
121 and determine the slowness (u), back-azimuth (θ , measured relative to the great circle
122 path between the earthquake and the array), and arrival time of incoming signals we
123 analyse data using a frequency-wavenumber (fk) approach (Capon et al., 1967), in
124 addition to using the F-statistic (described below). While other high-resolution processing
125 schemes are available, fk-analysis was selected for its increased computational speed
126 over traditional beam-forming and Vespa (velocity spectral analysis) approaches (Davies
127 et al., 1971). Also, the F-statistic effectively suppresses aliasing and is applicable to a
128 wide variety of array configurations (Selby, 2008). The data are windowed in time,
129 transformed into the frequency domain, and filtered between 0.5 and 2 Hz. The power
130 spectral density, $S(\omega)$, is then calculated within a given range of incoming directions,
131 which are combined into a single wavenumber vector, k . The fk method collapses the
132 time information to a single point around which the data were windowed. Thus, we adopt
133 a sliding window approach, selecting 10 sec long windows of data (starting from 10 sec
134 prior to the predicted first arrival of PK•KP), applying a cosine taper, and shifting the
135 window in 1 sec steps through the whole PK•KP time range (blue box in Fig. 2), to
136 measure the power of incoming energy from different directions through time.

137

138 To increase the resolution of the measured slowness and back-azimuth of PK•KP signals
 139 received at the arrays, we apply the F-statistic (Equation 1. Blandford, 1974). This
 140 method involves first beaming data on a specific slowness and back-azimuth, dividing the
 141 sum of the differences between the beam, b , and each trace in the beam, x_i , within a given
 142 time window, M , by the amplitude of the beam within the same window, and then
 143 weighting the output by the number of traces, N . This produces a dimensionless number,
 144 F , representing the signal coherence by measuring the cumulative difference from the
 145 beam at a given time.

$$F = \frac{N - 1}{N} \frac{\sum_{t=1}^M b(t)^2}{\sum_{t=1}^M \sum_{i=1}^N (x_i(t) - b(t))^2}$$

146 Equation 1

147

148 The F-statistic is efficient method of determining the best-fitting slowness and back-
 149 azimuth for a given signal in a beam. The method assesses the difference between the
 150 waveshape of each trace forming the beam and the beam itself, thus the method heavily
 151 penalises beams that differ from the traces used to form that beam. Effectively, this
 152 suppresses signals misaligned due to deviations from signal slowness and backazimuth
 153 (Selby, 2008). This leads to an increase of the slowness and back-azimuth resolution
 154 using small-aperture arrays. This calculation is repeated in a sliding time window to
 155 create an F-trace, which displays the coherence along the beam.

156

157 **3. Results**

158 We detect 1989 signals, each associated with an individual scattered wave. Fig. 4
159 presents an example earthquake showing energy from several heterogeneities. The
160 amplitude of scattered PK•KP waves is very low, often close to the noise level, thus array
161 processing is necessary to clearly resolve these signals. We are unable to determine the
162 relative amplitude of the PK•KP waves, as there is no suitable reference phase for each
163 scatterer with a similar path.

164

165 Using the geometrical direction of scattered arrival information (u, θ) to constrain the
166 path from scatterer to receiver (•KP) in combination with the timing information (t),
167 which indicates the length of the whole path, we determine the locations of the scattering
168 heterogeneity in terms of latitude, longitude, and depth, by ray-tracing from the seismic
169 array through a 1-dimensional Earth model (IASP91, (Kennett and Engdahl, 1991)). A 2-
170 dimensional grid of distance from the array and height above the CMB is constructed
171 representing possible scattering locations of PK•KP, oriented in the direction of wave
172 propagation along the back-azimuth observed at the array. The whole PK•KP path can
173 comprise any combination of either the ab or bc branches of PKP. We ray-trace to and
174 from each point in the grid along PKd and dKP paths, respectively, where the “d”
175 represents the depth of scattering in the lower mantle. For each distance and each
176 scattering depth in the grid, all permissible combinations of down-going PK_{ab}d or PK_{bc}d
177 legs and up-going dKP_{ab} or dKP_{bc} legs are considered.

178

179 Each of the PK_{ab}d and PK_{bc}d branches has different, yet partially overlapping, distance
180 ranges (which vary as a function of depth, d , of the scattering heterogeneity). The grid is

181 formed for the distance range of the $PK_{ab}d$ and $PK_{bc}d$ paths, and heights from the CMB to
182 300 km above into the mantle. To maximise processing speed and accuracy, a coarsely
183 spaced grid is first searched (constructed with 40 km depth intervals and 4° distance
184 intervals) between the CMB and 300 km above, followed by the finer grid projected ± 50
185 km from the best-fitting point in the coarse grid with 10 km depth intervals and 0.5°
186 distance intervals, thus allowing scatterers to be located up to 350 km above the CMB.

187

188 Each complete path from the source to each point in the scattering grid and up to the
189 surface is associated with a travel-time (the sum of the down-going and up-going legs)
190 and slowness (of the up-going leg to the seismic array). From these possible paths, we
191 find the best fitting scattering distance, depth, and complete path branch combination by
192 minimising the residual between slowness and travel-time of the traced ray and the
193 observed slowness and travel-time. The minimisation process is weighted in favour of
194 picking the smallest possible slowness residual (relative to the observed value) as it
195 controls scattering location more strongly than does travel-time, meaning that for
196 anomalies of similar magnitudes (e.g. 1 s/deg slowness and 1 s travel-time) the slowness
197 anomaly affects scattering location more significantly than does the travel-time anomaly.

198

199 Of the signals initially identified in the fk analysis, those with slownesses or back-
200 azimuths incapable of producing viable $PK \bullet KP$ paths are discarded (508 signals). For
201 signals whose observed slowness and time can be matched to that of a $PK \bullet KP$ scattering
202 location, we retain the best fitting scattering point identified in the second grid iteration.
203 The misfit between the slowness and time of the best fitting ray and that observed in the

204 data (with the f_k F-statistic) is used to assess the quality of the scatterer solution location:
205 scatterers are disregarded if either the slowness misfit is greater than 1.5 s/deg, the travel-
206 time misfit is greater than 2.5 s, or if the F-amplitude (coherence) of the scattered signal
207 is more than twice that of the P-wave signal for the same event. For the latter, we assume
208 the direct P-wave measured on the minor arc should be considerably more coherent than
209 the scattered PK•KP energy. These mislocation ranges were determined from
210 analysis of synthetic signals calculated for the geometry of each array in this study.
211 This resulted in discarding 534 additional signals. This analysis does not include
212 consideration of the sub-array structure, but the influence of this is likely small for
213 the majority of the arrays in our study (Bondar et al., 1991). Therefore, we twice
214 discard data, firstly if the observed back-azimuth and slowness cannot trace a viable
215 PK•KP path, and secondly if the misfits between the traced and observed slowness and
216 time or the F-amplitude are above a threshold (1042 total signals discarded), leaving 947
217 signals traced to scattering heterogeneities in the deep mantle. The combined
218 uncertainties in u , θ , and t yield a mislocation range in the mapped scattering
219 heterogeneity locations up to ± 150 km laterally and ± 30 km vertically.

220

221 **4. Frequency analysis**

222 Six of the arrays used contain broadband sensors (Table S1). For these data (113 signals)
223 we investigate the spectral character of scattered energy by calculating the scattering
224 strength in seven different octave wide band-pass filters from 0.25 to 32 Hz. We calculate
225 the signal-to-noise ratio (SNR) by comparing the amplitude of the linear beam of
226 scattered energy (created using the observed slowness and back-azimuth of that signal) to

227 the noise (from the same beam) that precedes the predicted first possible PK•KP arrival.
228 Across these events, we observe a spectrum of scattered energy between 1 and 32 Hz
229 (indicating heterogeneity scales at the base of the mantle spanning 0.4-14 km) with the
230 strongest scattered energy most commonly observed in the 2 to 4 Hz frequency band
231 (Fig. 5), equivalent to 4 to 7 km heterogeneity size and consistent with heterogeneity
232 scales found recently by modelling PKP precursors (Mancinelli et al., 2016). Scattered
233 energy is found across the spectrum indicating that heterogeneities of a large range of
234 sizes exist in the lowermost mantle. While beam forming may act as a mild low-pass
235 filter, the overall spectral content of our beamed and unbeamed data show little
236 difference. Thus, beam-forming can be effectively used to align and sum the coherent
237 PK•KP signals resulting in their emergence from the background incoherent noise. Any
238 small changes to the spectral content caused by the beam-forming process will affect both
239 the scattered signals and the noise to which it is measured relative. Plots of the spectra of
240 the scattered energy calculated from individual records do not reveal the scattered signals
241 due to their very small amplitude relative to the noise level.

242

243 To confirm the assumption that scattered energy in the 2 to 4 Hz frequency range is
244 indicative of heterogeneities from 4 to 7 km in size, we model the scattered PK•KP
245 energy generated in different frequency bands for point heterogeneities with an 8 km
246 correlation length using an exponential representation of the heterogeneity spectrum. We
247 model the scattered energy using a Monte Carlo, Phonon Scattering code (Shearer and
248 Earle, 2004) that maps the passage of particles representing seismic energy, or phonons,
249 from a source through a 1-dimensional velocity structure along a range of ray parameters

250 (take-off angles). The synthetic seismic wavefield is constructed by accumulation of
251 phonons across distance and through time. The 1-dimensional velocity structure can be
252 augmented with addition of layers of heterogeneity, which may cause waves passing
253 through the layer to scatter. The velocity and scattering structure is 1-dimensional but the
254 algorithm takes off-Great Circle Path scattering into account, thus this method can model
255 PK•KP paths. However, as the scattering structure is uniform laterally, this method
256 cannot be used to accurately model the observed scattered signals that are generated by
257 unevenly distributed structure. Nonetheless, the method is an effective way to assess the
258 effects on the scattered wavefield of varying the depth distribution, elastic parameters, or
259 scale-length of scattering heterogeneity, as well as the incident seismic wave frequencies
260 (Mancinelli and Shearer, 2013; Mancinelli et al., 2016). Using a different autocorrelation
261 function to describe the heterogeneity spectrum would likely have a small influence on
262 the resultant scattered energy.

263

264 We utilize the above method to construct a model with scattering heterogeneities in the
265 lowermost ~300 km of the mantle, consistent with the depth range that we are able to
266 study with our data. Heterogeneities have an RMS velocity perturbation of 0.5%, and a
267 scale-length of 8 km. We simulate the scattered wavefield at different incident
268 frequencies (effectively bandpass filtering the scattered signals): 0.25, 0.5, 1, 2, 4, 8, 16,
269 and 32 Hz. We find that at lower frequencies up to 1 Hz, scattering in the PK•KP time
270 window is sporadic and weak (Fig. 6). At 1 Hz, scattering becomes more prominent.
271 However, scattering is strongest and most consistent at frequencies from 2 to 4 Hz
272 indicating that scattering from heterogeneities 8 km in size would be most strongly

273 observed at these higher frequencies. At 8 Hz and above scattered signals very rapidly
274 become weak, seven orders of magnitude lower amplitude than at 4 Hz. Above 8 Hz
275 scattered waves are absent from the synthetics. The relatively smooth envelope of the
276 scattered energy results from the 1-D nature of the model. The amplitude of scattered
277 energy increases through time as the PK•KP wave interacts with a greater lateral and
278 vertical extent of the mantle, expanding from a scattering point on the CMB at 1710 s
279 after the origin.

280

281

282 **5.1 Distribution of scattering heterogeneities and mantle structures**

283 We investigate the relationship between the abundance of scattering heterogeneities and
284 the large-scale seismic structures in the lower mantle as resolved by seismic tomography.
285 To remove sampling bias from our scattering population due to uneven event-array
286 distributions, we normalise the scattering population by the geographic distribution of
287 scattering (Fig. 3e). The resulting normalised scattering distribution indicates laterally
288 uneven heterogeneity distributions, which cannot be explained by the sampling
289 (Supplementary Fig. 1). We analyse the spatial correlation between the locations of
290 scattering heterogeneities and regions of (1) high or low tomographic velocity anomalies
291 and (2) high or low lateral velocity gradients across the CMB for seven recent S-wave
292 tomographic models: GyPSuM (Simmons et al., 2010), HMSL-S06 (Houser et al., 2008),
293 savani (Auer et al., 2014), SEMUCB-WM1 (French and Romanowicz, 2014),
294 S362WMANI+M (Moulik and Ekstrom, 2014), S40RTS (Ritsema et al., 2011), and
295 TX2011 (Grand, 2002) (Figs. 7 and Supplementary Fig. 2) and for 4 recent P-wave

296 tomographic models: GAP_P4 (Obayashi et al., 2013), GyPSuM_P (Simmons et al.,
297 2010), HMSL_P06 (Houser et al., 2008), and MIT-P08 (Li et al., 2008) (Fig. 8 and
298 Supplementary Fig. 2). While the scattering heterogeneities are observed using P-waves,
299 we compare our results to S-wave tomography models from which LLVPs are defined.
300

301 We first distinguish regions related to subduction (positive dVs), and the LLVPs
302 (negative dVs). For each model, we calculate lateral velocity gradients ($\nabla(\text{dVs})$)
303 measured over a horizontal distance of 10° . This gradient distance results in regions of
304 high velocity gradient consistent with the outline of the LLVPs as observed by high-
305 resolution forward modelling studies (He and Wen, 2009).
306

307 We compare the distribution of velocity perturbations and gradients against the
308 normalised scattering heterogeneity distribution. As the magnitude, range, and precise
309 pattern of velocity anomalies varies between tomography models, we are unable to select
310 single values of dVs or $\nabla(\text{dVs})$ that are suitable for defining LLVP boundaries for all
311 models. To standardise comparisons of our scattering heterogeneity distribution and the
312 different tomographically derived LLVPs, we compare to specific percentage areas of the
313 CMB occupied by high or low velocity or gradient (Supplementary Fig. 5). We calculate
314 the number of scattering heterogeneities, as a proportion of the total population of
315 scatterers, within a range of values of velocity anomaly, calculated by the CMB area (5,
316 10, 15, 20, 25, 30, 40, 50, 60, 70, 80, 85, 90, and 95% area of highest or lowest velocity
317 or gradient anomalies).
318

319 We assess the statistical significance of these correlations by comparing them with the
320 spatial correlation between randomly rotated tomographic models and the scattering
321 dataset (Supplementary Figs 3 and 4). Each tomography model is rotated (through
322 random co-latitude and longitude angles) and the correlations with scattering
323 heterogeneities are recomputed. This is repeated 100 times for each model. The
324 significance of correlations between scatterer locations and actual locations of
325 tomography features is determined by comparison with the distribution of correlations
326 with the randomly rotated tomography models. When correlation of the scattering
327 distribution with the original, unrotated models are fully or dominantly outside of the
328 range calculated for with randomly rotated models (defined by 1 standard deviation,
329 shown as grey shaded regions in Figs. 7 and 8), we conclude the relationship to be
330 statistically significant. We observe a statistically robust increased concentration of
331 scattering heterogeneities in regions of the highest lateral velocity gradients (Fig. 7c) and
332 a weak correlation between scattering heterogeneities and low velocity anomalies (Fig.
333 7e), the former are associated with the edges of the LLVPs (Thorne et al., 2004).

334

335 When reviewed more closely, we find that scattering heterogeneity shows the strongest
336 correlation with moderately low velocities (~20-40% of CMB area occupied by lowest
337 velocities). This correlation is most prominent with P-wave models (Supplementary Figs.
338 2 and 3). Meanwhile, the correlation with lateral gradients is highest for the very
339 strongest gradients. Both moderately reduced velocities and high gradients characterise
340 the boundaries of the LLVPs.

341

342 **5.2 Distribution of scattering heterogeneity and geodynamic models**

343 We explore geodynamical calculations of subducted oceanic crust entrained into mantle
344 flow in a model where the lower mantle contains chemically distinct thermochemical
345 piles. We model flows in the mantle with a 2-dimensional Cartesian numerical
346 convection calculation, which solves the conservation equations of mass, momentum, and
347 energy in Boussinesq approximation (Li et al., 2014) using the CitCom code (Moresi and
348 Gurnis, 1996). Our model contains three compositional components, including ambient
349 mantle, denser pile material, and oceanic crust. The density for each compositional
350 component is the same as the reference case of Li et al. (2014): the oceanic crust and pile
351 material have the same non-dimensional buoyancy number of 0.8 ($\sim 2\text{-}3\%$ denser than
352 the ambient mantle). The Rayleigh number is $Ra=1\times 10^7$. The temperature dependent
353 viscosity is given by $\eta_T = \exp [A(0.5 - T)]$, with activation coefficient $A=9.21$, resulting
354 in four orders of magnitude variations of viscosity due to changes of temperature. We
355 employ a viscosity increase by a factor of 50 from upper mantle to lower mantle at 660
356 km depth. In addition, a phase transition from bridgmanite to post-Perovskite acts to
357 reduce the viscosity in colder regions of the lowermost mantle. All boundaries are free-
358 slip. The upper and lower surfaces are isothermal while the sides are insulating. The
359 advection of the composition field is simulated by tracers. A 6 km thick oceanic crust is
360 introduced on the surface of the model by setting the identity of tracers shallower than 6
361 km depth to crust, and the oceanic crust is later subducted into the deep mantle. We
362 analyze the evolving distribution of crustal tracers with respect to the dense primordial
363 piles in the lowermost part of the model close to the CMB (Fig. 9). While some studies of
364 the dynamic behaviour of heterogeneities are able to analyse the evolution of their shape

365 (e.g. Olson et al., 1984), by treating heterogeneities as tracers we are unable to determine
366 the nature of the mechanical mixing (stretching and folding) of initially planar collections
367 of tracers that represent the crust. For a full description of the modeling procedure
368 employed here see Li et al., (2014) and supplementary information therein.

369

370 In this model, subducted crust has an elevated concentration near the boundaries of the
371 thermochemical piles due to the change in flow direction from lateral (between piles) to
372 radial (vertically along pile margins). The edges of the modelled thermochemical piles
373 display high temperatures, and also high thermal gradients across the margins (Li et al.,
374 2014). The high concentration of oceanic crust along the pile edges in these simulations
375 (Fig. 9) thus matches seismic observations showing increased concentration of scatterers
376 in regions of low seismic velocity and high velocity gradients (Fig. 7). The distribution of
377 crustal remnants near pile margins is time dependent, due to the subduction-related flow
378 patterns, but agrees well with our observations over many time snapshots and different
379 locations (Supplementary Figs. 5 and 6). In this model there is initially no crust and the
380 mantle, but the amount of crust increases through time, dependent on the rates and
381 densities chosen. As such, the model cannot be used to compare relative quantities of the
382 three compositions. We find that the general patterns of crust distribution in the models
383 qualitatively match our observations. In addition, the oceanic crust can be entrained in to
384 thermochemical piles (Li et al., 2014), which may cause seismic scattering within
385 LLVPs.

386

387

388 **6. Discussion and conclusions**

389 Past work has investigated how shallow structure beneath seismic arrays can affect the
390 resolved direction of incoming seismic waves (Bondar et al., 1999). In our analysis, we
391 dominantly use arrays that have previously been demonstrated to have minor or
392 insignificant sub-surface structure. Not all arrays used here have calculated slowness-
393 azimuth station corrections, and calculating these for all arrays is beyond the scope of this
394 paper. Nonetheless, when we remove the data from Chiang Mai array, which shows the
395 largest station corrections (which accounts for of the 20% of the well-resolved scattering
396 locations, there is no notable change in the correlations with the tomographic structures,
397 thus the observed correlations are defined by results from other arrays with less
398 influential sub-surface structure.

399

400 The dominance of 2 to 4 Hz energy in the scattering spectrum (Figs. 5) indicates the
401 presence of dominant heterogeneity scale lengths in the deepest mantle of ~4 to 7 km.
402 Relatively high thermal conductivity in the lowermost mantle (Hofmeister, 1999) should
403 lead to fast thermal equilibration of small-scale heterogeneities (Olson et al., 1984)
404 suggesting a compositionally distinct origin from the background mantle. The 4 to 7 km
405 scale size agrees well with the thickness of oceanic crust (~6 km) and the spacing of
406 normal faults related to slab bending at oceanic trenches (Masson, 1991). The fine-scale
407 heterogeneity mapped here, therefore, may be related to convectively driven
408 segmentation of the formerly contiguous layer of subducted oceanic crust and shows little
409 evidence for strong deformation through viscous forces in the mantle convection
410 (Tackley, 2011).

411
412 Other sources of heterogeneity within the lowermost mantle can feasibly produce
413 scattering. LLVP thermochemical pile material can be entrained into mantle flow
414 (Williams et al., 2015), however this does not explain scattering observed within LLVPs.
415 Small scale CMB topography could scatter waves (Doornbos, 1978), but generating wide
416 out-of-great-circle plane PK•KP paths by underside CMB reflections would require
417 unrealistically steep and high amplitude topography at CMB scattering locations to get
418 energy back to the array. Additionally, topography could not explain travel times of
419 signals used to map scattering higher above the CMB. Heterogeneities undergoing the
420 phase transition from bridgmanite (perovskite) to post-perovskite (pPv) (Murakami et al.,
421 2004; Oganov and Ono, 2004) may have locally sharp velocity contrasts. However, the
422 range of meta-stability, where pPv would coexist and thus contrast with bridgmanite, is
423 measured to be 70-600 km thick (Grocholski et al., 2012), below which pPv would be
424 ubiquitous and not provide an elastic contrast, and above which it would not exist, thus
425 this hypothesis would struggle to explain the observed depth range of scattering.
426 Furthermore, pPv is not expected to be stable within the relatively hot LLVPs, thus may
427 not be responsible for scattering in these regions, although there have been limited
428 observations of discontinuities within the LLVPs (Lay et al., 2006). Ultra low velocity
429 zones (ULVZs) are observed to have distinct velocities and densities and may comprise
430 partial melt or products of chemical reactions between mantle and core material
431 (Williams and Garnero, 1996; Knittle and Jeanloz, 1991). While this material can provide
432 a sufficiently sharp elastic contrast, ULVZs, with thicknesses of only a few 10s of km,
433 are unable to explain scatterer locations away from the CMB. The increased density of

434 ULVZ material (up to 10%) might prohibit entrainment of the material to the observed
435 scattering heights (Bower et al., 2011). In the vicinity of the LLVP boundaries, it is
436 possible that scattering may occur at the transition between the LLVP material and the
437 ambient mantle, however, the mechanism by which this would occur has not been fully
438 investigated. It is unclear how any of the discussed mechanisms could explain the
439 observed lateral and vertical distribution of small-scale heterogeneities, both inside and
440 outside of LLVPs, as well as the heterogeneity size, as simply as subducted oceanic crust.
441 However, a combination of these sources may be expected, and might explain aspects of
442 the observed scattering.

443

444 The distribution of small-scale heterogeneity near LLVP margins is consistent with a
445 thermo-chemical origin of LLVPs (McNamara and Zhong, 2005) with strong radial flow
446 near pile margins that entrains crust up off the CMB (Li et al., 2014) (Supplementary
447 Figs. 7 and 8). Nevertheless, scattering heterogeneities are observed throughout the lower
448 mantle, in regions away from the LLVPs, but in lower abundance. This suggests recycled
449 oceanic crust may be stirred into the mantle, and can retain a seismic signature distinct
450 from background deep mantle material with little apparent deformation over long time
451 periods of 100s of Myrs and greater. These results suggest a connection between small-
452 and large-scale structures through dynamics processes within the lower mantle. Further
453 analysis of scattering heterogeneities throughout the mantle will help to resolve the extent
454 of this connection.

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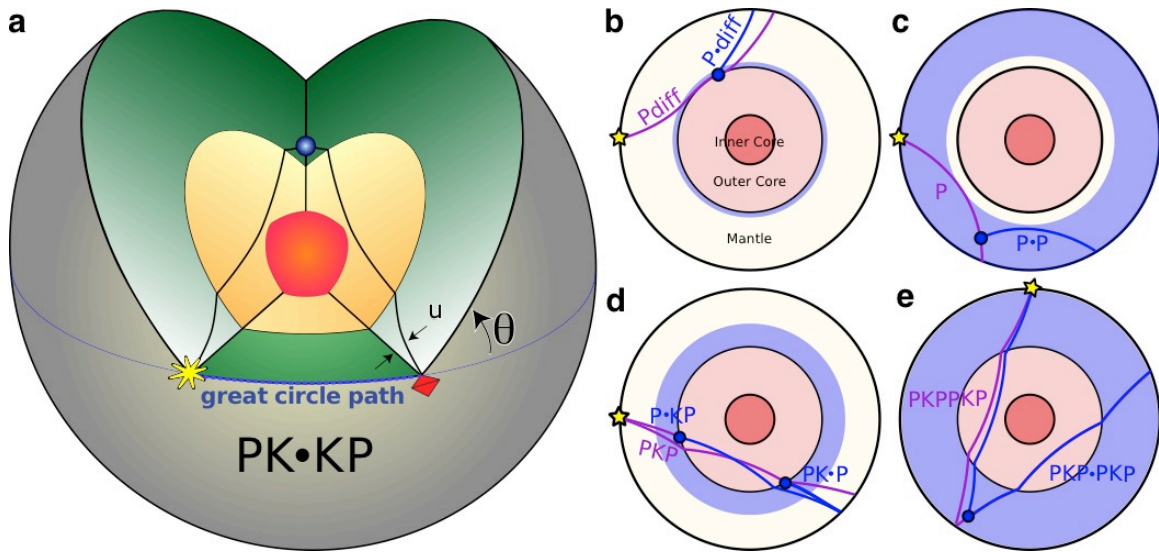
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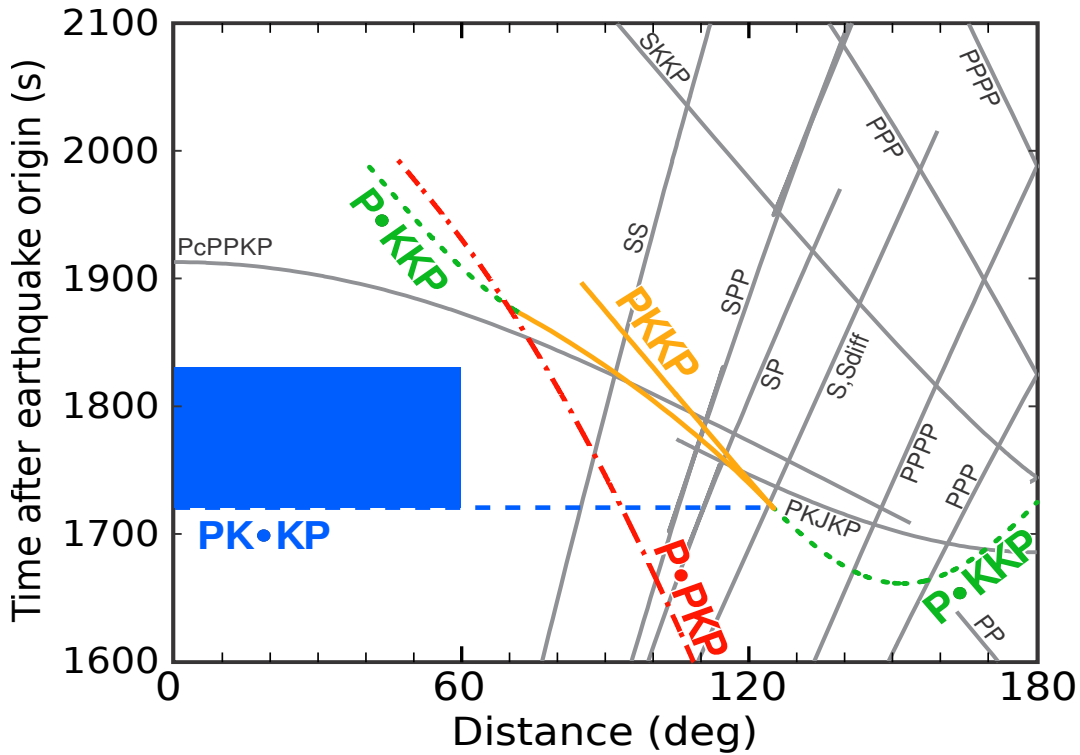
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639

640 **Fig. 1:** Raypaths of direct and scattered waves used to map small-scale mantle
 641 heterogeneity. (a) PK•KP path from an earthquake (star) to a scattering heterogeneity at
 642 or near the CMB (blue dot), and to the array (pyramid). The scattered wave travels out of
 643 the great-circle path. Other scattered waves used to study mantle structure include (b)
 644 P_{diff} : a P-wave diffracted along the core-mantle boundary (CMB) and back up to the
 645 surface. The wave can be scattered at some distance along the diffracted portion, denoted
 646 $P_{\text{diff}}\bullet$. Direct and scattered paths (purple and blue lines, respectively), and example
 647 scattering location (blue circle) are shown along with the depth range (blue shaded
 648 region) that can be studied with this probe. (c) P: a direct P-wave, which can be scattered
 649 some depth, indicated by \bullet , to another P wave (or S wave, then called P•S) which travels
 650 back to the surface, (d) PKP: a P-wave which travels through the mantle, outer core, and
 651 back up to through the mantle the surface. This wave can be scattered within the mantle,
 652 the location of which is indicated by \bullet , either before the wave enters the outer core
 653 (P•KP), or as the wave exits the outer core (PK•P), (e) PKPPKP: a wave which travels
 654 down through the mantle and outer core, up through the mantle and reflects off the
 655 underside of the CMB, and returns down through the mantle and up through the outer

656 core and mantle to the surface. A similar wave can be scattered back down through the
657 mantle on the antipodal side and back up to the surface, which can be written as
658 PKP•PKP. [TWO-COLUMN FIGURE]

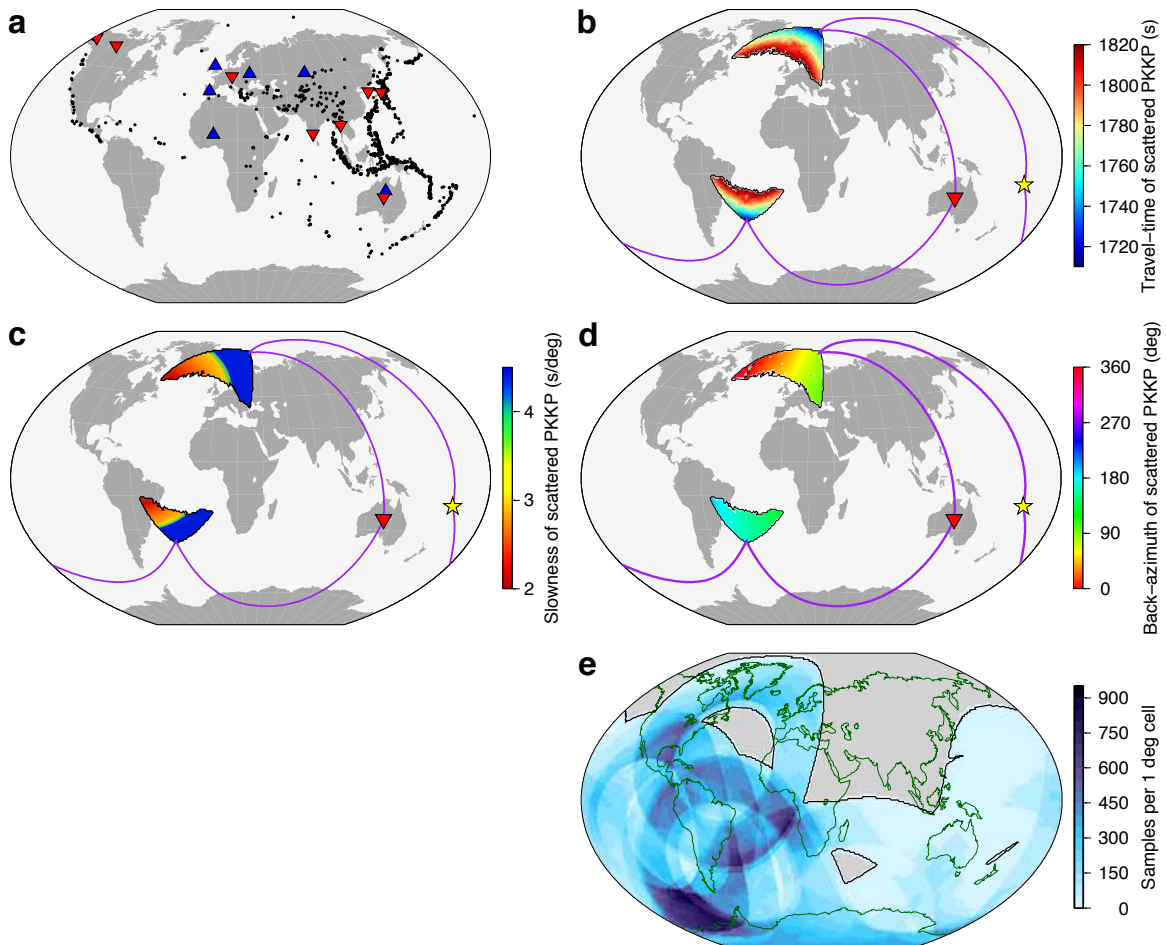


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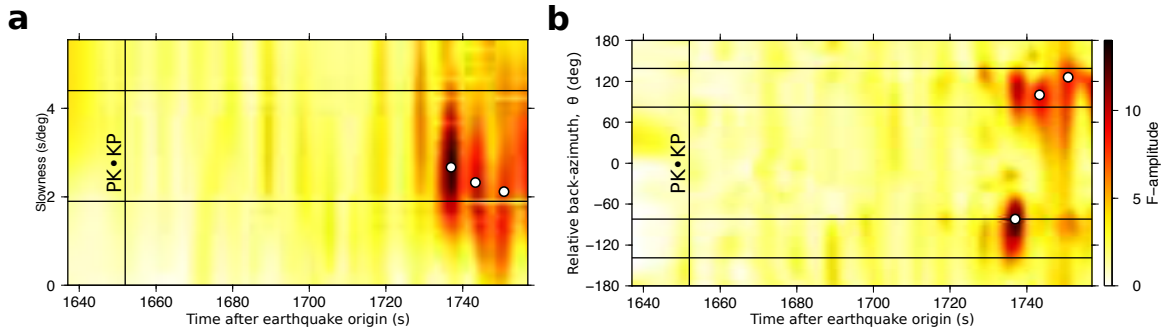
660 **Fig 2.** Travel-time curves of PK•KP and related scattered waves. In the study we use data
 661 with source-receiver distances between 0 and 60 degrees and from the first theoretically
 662 viable arrival of PK•KP to 110 s after (blue box). Times of other variants of the PKKP
 663 path (coloured lines) are also shown (See Earle, 2002 for discussion of other scattered
 664 phases). Times of non-scattered waves are shown as grey lines. The blue dashed line
 665 depicts PK•KP times at larger distances, which are not used here due to interference with
 666 other arrivals. All travel time curves are calculated using IASP91 (Kennett and Engdahl,
 667 1991) and a surface focus earthquake. Figure after Earle (2002). [SINGLE-COLUMN
 668 FIGURE]

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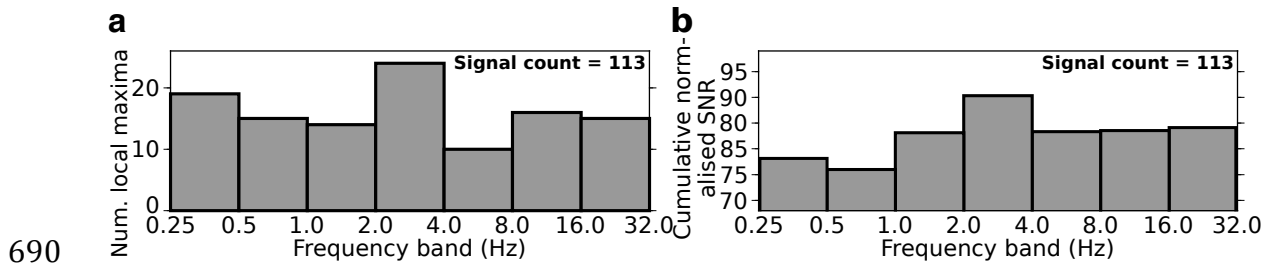


671 **Fig 3.** Data and resultant sampling of the lower mantle by PK•KP. (a) 1095 events (black
672 dots), and 6 broadband (blue triangles) and 8 short-period (red triangles) arrays used. For
673 a given source (yellow star) and receiver array (inverted red triangle), the regions of
674 possible scattering can be predicted along with the associated (a) travel-time, (b) back-
675 azimuth, and (c) slowness of a PK•KP wave scattered at the CMB at each point in the
676 sampled region. Example ray paths from the source to both scattering regions (on either
677 side of the inner core) and back to the array are predicted and displayed as purple lines.
678 (e) Sampling of CMB by PK•KP for the source and receiver distribution in (a) calculated
679 by super-position of potential scattering regions as in (c). Black line outlines region
680 sampled by this dataset. [TWO-COLUMN FIGURE]
681



682

683 **Fig. 4.** Array processed PK•KP waves. For a single example event, energy in the PK•KP
 684 time window, is fk processed, and displayed in terms of (a) slowness and time and (b)
 685 back-azimuth (relative to the great-circle path) and time, shown as F-amplitude. Three
 686 coherent and distinct PK•KP signals are identified (circles). Vertical line marks the first
 687 possible PKKP arrival (Fig. 2). Horizontal lines mark the possible range of PK•KP, (a)
 688 slownesses (Fig. 3c) and, (b) back-azimuths on both sides of the great-circle path (Fig.
 689 3d). [TWO-COLUMN FIGURE]



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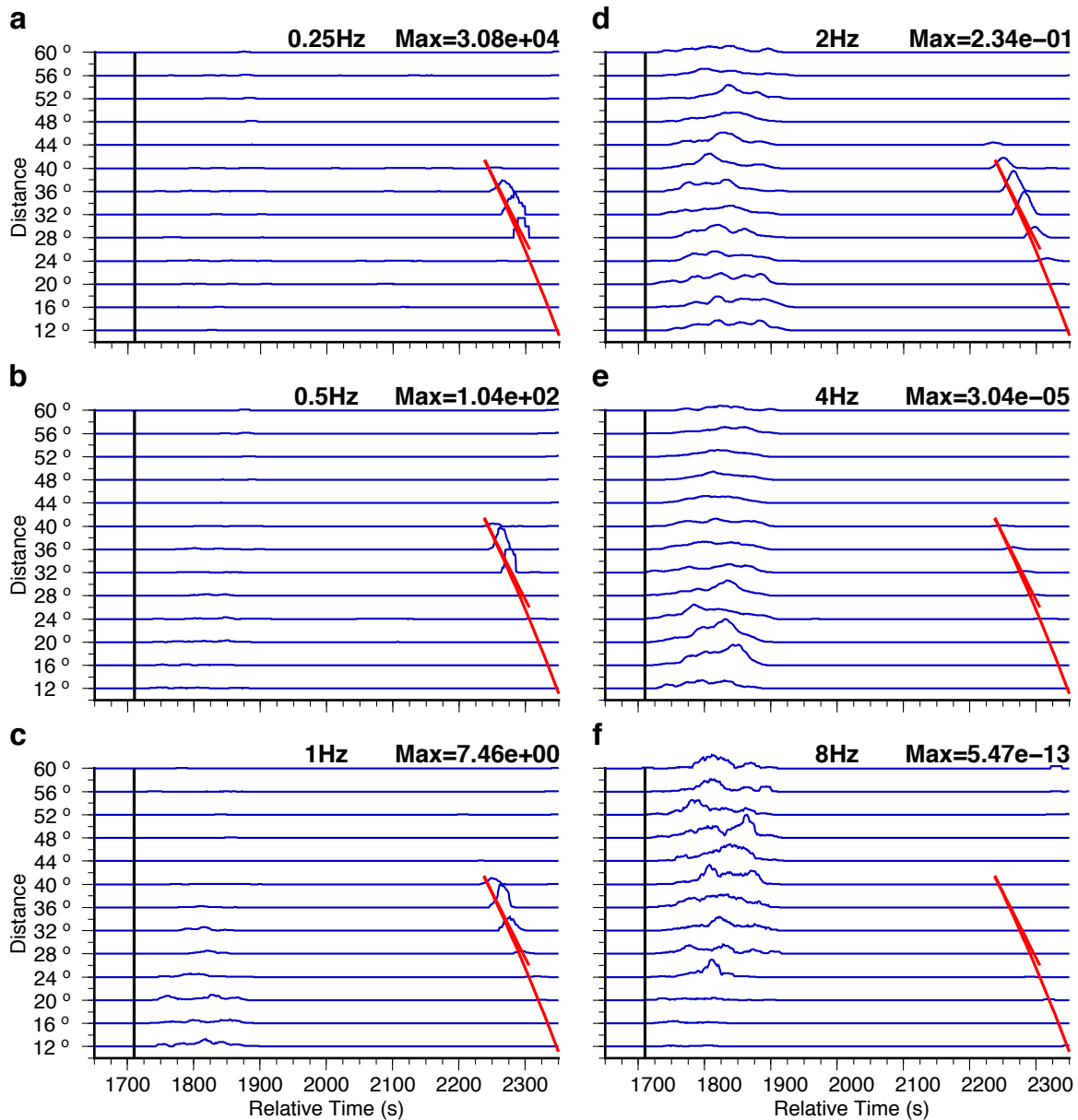
691 **Fig 5.** Frequency characteristics of 113 PK•KP waves observed at the 6 broadband arrays

692 displayed as, **(a)** strongest Signal-to-Noise-Ratio (SNR) (each signal contributes 1 to

693 count) and, **(b)** cumulative SNR (each signal normalized to one before summing). [TWO-

694 COLUMN FIGURE]

695



696

697 **Fig. 6.** Synthetic scattering model of a 1-dimensional Earth probed at different

698 frequencies. A Monte Carlo method (Shear and Earle, 2004) treating waves as packets of

699 energy (phonons) is used to build a synthetic wavefield including scattering at incident

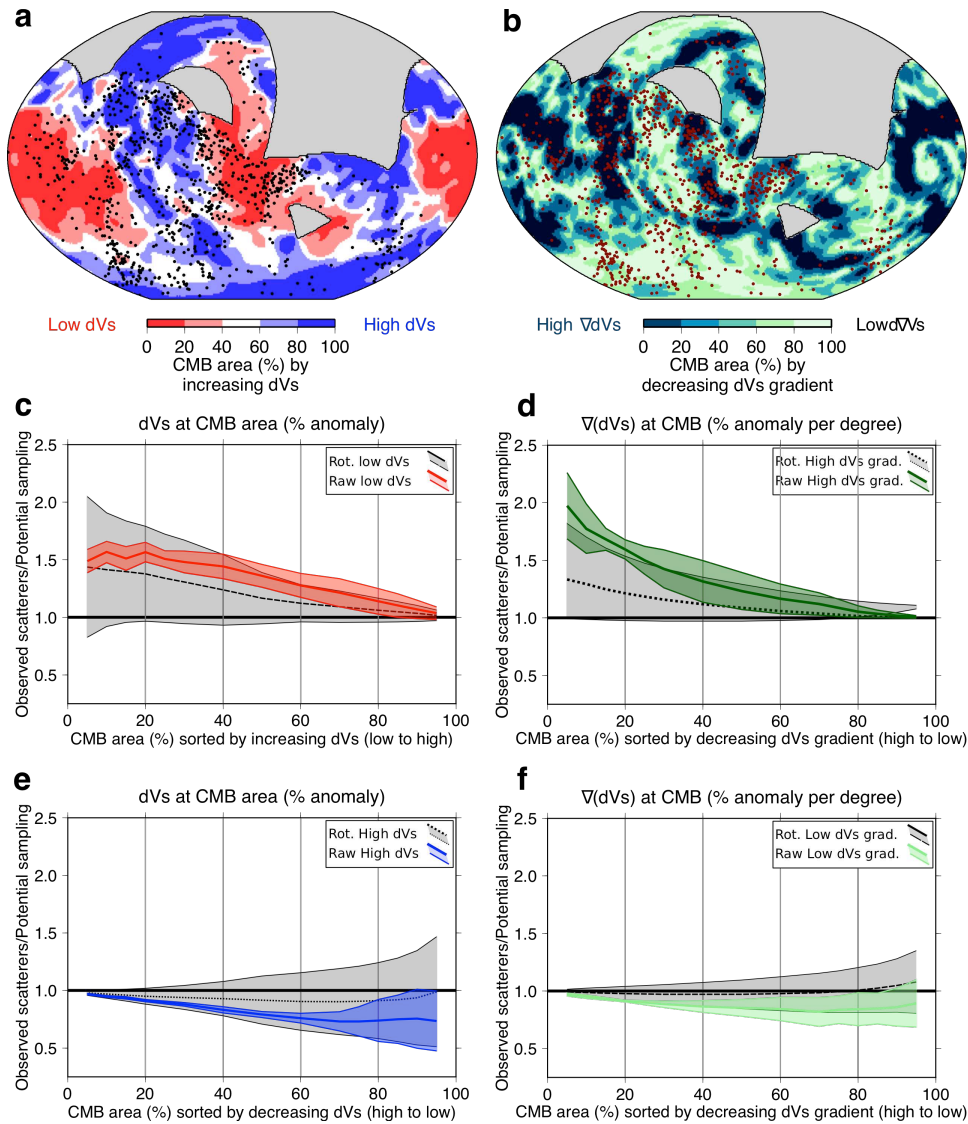
700 frequencies of (a) 0.25, (b) 0.5, (c) 1, (d) 2, (e) 4, (f) and 8 Hz. Phonons sample a 1-

701 dimensional Earth model with a 290 km thick layer in the lowermost mantle containing

702 scattering heterogeneities with an RMS velocity perturbation of 0.5%, with correlation

703 lengths of 8 km. We model a range of distances comparable to our data, and display

704 envelopes of the smoothed sum of wavefield profiles in a ± 2 degree distance range about
705 the central distance. For each frequency, all profiles are normalised to the maximum of
706 the profile at 32° (shown at the top right) to account for the amplitude at the P3KP
707 caustic, the arrival of which is marked in red. The black line marks the expected onset
708 time of PK•KP scattering. [TWO-COLUMN FIGURE]
709



710

711 **Fig. 7.** Correlation of scattering heterogeneities with large-scale S-wave seismic

712 structure. Percentage area of the CMB covered by (a) S-wave velocity anomalies

713 increasing from low to high, and (b) lateral velocity anomaly gradients decreasing from

714 high to low (S40RTS (Ritsema et al., 2011)). Scattering heterogeneities are shown as dots.

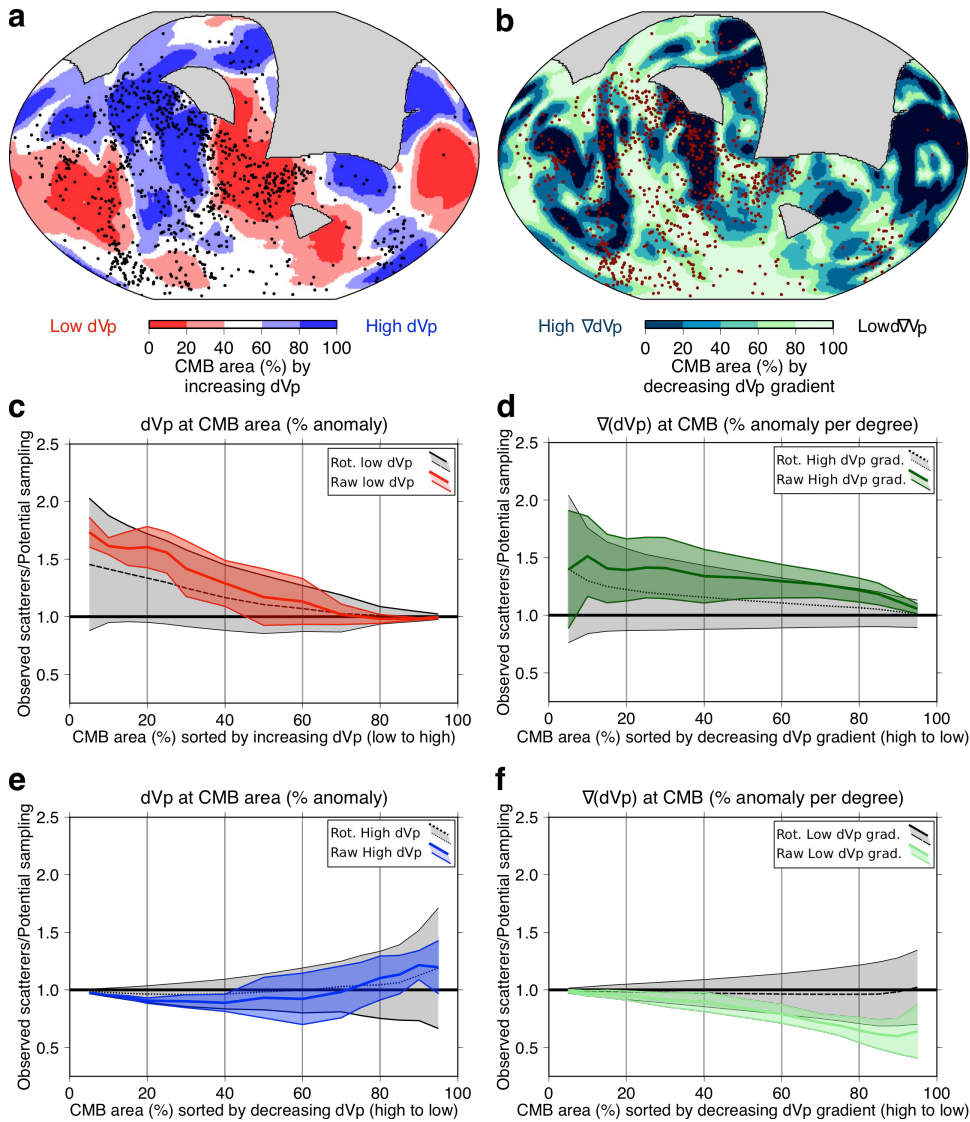
715 The highest gradients in (b) match well the edges of LLVPs seen in the tomography map,

716 (a) where the magnitude of the velocity anomaly changes rapidly. Ratio of observed to

717 potential scatterers against (c) increasing velocity anomaly (red lines), (d) decreasing

718 lateral velocity gradient (dark green lines), (e) decreasing velocity anomaly (blue lines),

719 and (f) increasing lateral velocity gradient (light green lines). The black horizontal line
720 indicates a 1-to-1 ratio of sampling-to-scattering, data above and below this line indicate
721 more or fewer heterogeneities than expected based on sampling, respectively. Displayed
722 is a compilation of the correlations with 7 tomography models shown by the mean and 1
723 standard deviation (thick coloured lines and shading, respectively) (Supplementary Fig. 3
724 shows individual analysis). Correlations are also calculated for random rotations of the 7
725 models (black lines and grey regions for mean and standard deviation, respectively).
726 Tomographic anomalies are displayed by CMB area, sorted by anomaly magnitude.
727 [TWO-COLUMN FIGURE]
728



729

730 **Fig. 8.** Correlation of scattering heterogeneities with large-scale P-wave seismic

731 structure. Percentage area of the CMB covered by (a) P-wave velocity anomalies

732 increasing from low to high, and (b) lateral velocity anomaly gradients decreasing from

733 high to low (GAP_P4 (Obayashi et al., 2013)). Scattering heterogeneities are shown as

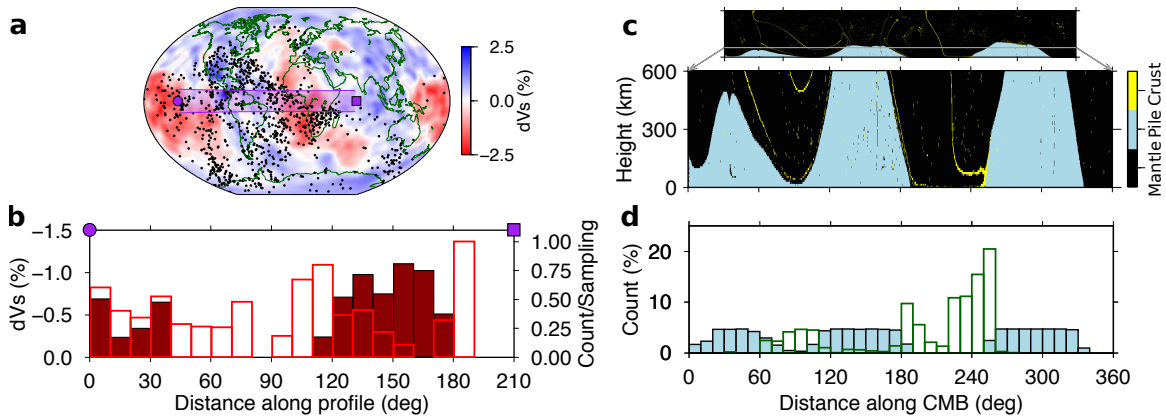
734 dots. Ratio of observed to potential scatterers against (c) increasing velocity anomaly (red

735 lines), (d) decreasing lateral velocity gradient (dark green lines), (e) decreasing velocity

736 anomaly (blue lines), and (f) increasing lateral velocity gradient (light green lines). The

737 horizontal line indicates a 1-to-1 ratio of sampling-to-scattering, data above and below

738 this line indicate more or fewer heterogeneities than expected based on sampling,
739 respectively. Displayed is a compilation of the correlations with 7 tomography models as
740 the mean and 1 standard deviation (thick coloured lines and shading, respectively)
741 (Supplementary Fig. 4 shows individual analysis). Correlations are also calculated for
742 random rotations of the 7 models (black lines and grey regions for mean and standard
743 deviation, respectively). Tomographic anomalies are displayed by CMB area, sorted by
744 anomaly magnitude. [TWO-COLUMN FIGURE]



745

746 **Fig. 9.** Scattering heterogeneities, crust, and thermochemical pile margins. **(a)** Velocity
 747 anomalies at CMB (S40RTS (Ritsema et al., 2011)), scattering heterogeneities (dots), and
 748 a 20° wide cross-section swath from 140° W, 0° N to 70° E, 0° N (shaded). **(b)**
 749 normalised ratio of observed to potential scatterers along the cross-section (open bars,
 750 right legend, maximum = 2.08×10^{-4}), and tomographic velocity reductions averaged
 751 across swath, indicating LLVPs (closed bars, left legend). **(c)** A time snapshot from
 752 numerical thermo-chemical convection calculation (Li et al., 2014) displaying
 753 distribution of tracers representing crustal (yellow), pile (blue), and background mantle
 754 (black) material in lowermost 600 km of mantle with 10-times vertical exaggeration and
 755 whole mantle to scale in upper panel. Tracers are discretised into 10×10 km cells and the
 756 dominant tracer defines the cell type (crust, pile, or mantle). **(d)** Lateral distribution of
 757 crust (green bars) and thermo-chemical piles (blue bars) in the lowermost 300 km for the
 758 time snapshot in **(c)**. [TWO-COLUMN FIGURE]

759