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Title: Dynamical links between small- and large-scale mantle heterogeneity: seismological evidence

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Abstract

We identify PKP•PKP scattered waves (also known as P′•P′) from earthquakes recorded at small-aperture seismic arrays at distances less than 65°. P′•P′ energy travels as a PKP wave through the core, up into the mantle, then scatters back down through the core to the receiver as a second PKP. P′•P′ waves are unique in that they allow scattering heterogeneities throughout the mantle to be imaged. We use array-processing methods to amplify low amplitude, coherent scattered energy signals and resolve their incoming direction. We deterministically map scattering heterogeneity locations from the core-mantle boundary to the surface. We use an extensive dataset with sensitivity to a large volume of the mantle and a location method allowing us to resolve and map more heterogeneities than have previously been possible, representing a significant increase in our understanding of small-scale structure within the mantle. Our results demonstrate that the distribution of scattering heterogeneities varies both radially and laterally. Scattering is most abundant in the uppermost and lowermost mantle, and a minimum in the mid-mantle, resembling the radial distribution of tomographically derived whole-mantle velocity heterogeneity. We investigate the spatial correlation of scattering heterogeneities with large-scale tomographic velocities, lateral velocity gradients, the locations of deep-seated hotspots and subducted slabs. In the lowermost 1500 km of the mantle, small-scale heterogeneities correlate with regions of low seismic velocity, high lateral seismic gradient, and proximity to hotspots. In the upper 1000 km of the mantle there is no significant correlation between scattering heterogeneity location and subducted slabs. Between 600 and 900 km depth, scattering heterogeneities are more common in the regions most remote from slabs, and close to hotspots. Scattering heterogeneities show an affinity for regions close to slabs within the upper 200 km of the mantle. The similarity between the distribution of large-scale and small-scale mantle structures suggests a dynamic connection across scales, whereby mantle heterogeneities of all sizes may be directed in similar ways by large-scale convective currents.

Keywords: seismology; deep Earth; scattering; mantle structure; mantle dynamics; seismic arrays

1. Introduction

The high frequency (~1 Hz) seismic wavefield provides evidence of kilometre scale structure within the Earth [Cleary and Haddon, 1972]. Seismic energy that is not explained by wave propagation in smoothly varying velocity models of the Earth has been attributed to reflections and scattering from sharply contrasting volumetric heterogeneities and roughness on interfaces [Chang and Cleary, 1981]. The interaction of the wavefield with discrete, small-scale variations in elastic properties and/or density can divert seismic energy onto new paths, often generating precursors or postcursors (coda) to the main seismic phases that travel in the great circle plane. The size of the scatterers that can be imaged is dependent upon the wavelength that is analysed; for the teleseismic high-frequency P-wavefield above 1 Hz they are typically on the order of 1 to 10 km.

Global imaging of Earth's small-scale heterogeneities is difficult due to the uneven distribution of earthquake sources and seismic receivers, and the low amplitude of the scattered signals involved. Scattering can be studied using single stations, but with this approach the location of the scattering heterogeneity can be ambiguous [Wen, 2000]. Alternatively, seismic arrays, i.e., 3 or more closely located sensors, can resolve the incoming direction of scattered waves, thus it is possible to deterministically locate heterogeneities [Thomas et al., 1999; Rost and Earle, 2010; Frost et al., 2013]. In the last few decades a number of studies have started to unravel the distribution of small-scale heterogeneities of Earth's mantle. Hedlin et al. [1997], and later Mancinelli and Shearer [2013, 2016] studied the depth distribution of heterogeneity within the mantle through analysis of PKP pre- and postcursors recorded at single stations. Using a stochastic Rayleigh-Born scattering approach, Mancinelli and Shearer [2013, 2016] developed a global model of scattering heterogeneity containing 0.1% root-mean-square velocity variations in the deepest 1200 km of the mantle with heterogeneity scale sizes ranging from 2 to 30 km.

This work is complemented by studies that deterministically map small-scale scattering heterogeneity within the upper and lower mantle. These studies have noted lateral variations in heterogeneity distribution, as well as variations in amplitudes of scattered waves. Scattered P-to-P (P•P, where the "•" represents the location of scattering) and P-to-S (P•S) waves are sensitive to heterogeneities in the upper half of Earth's mantle; they have been used to map scattering heterogeneity in regions influenced by recent subduction [Kaneshima and Helffrich, 1998; Bentham and Rost, 2014]. Scattering in the lowermost mantle has also been observed to vary laterally [Waszek et al., 2015]. Strong scattering has been observed in regions beneath mantle hotspots [Wen, 2000], near small, regional ultra-low velocity zone (ULVZ) structures [Yao and Wen, 2014], beneath subduction zones [Miller and Niu, 2008], and near the edges of LLSVPs [Frost et al., 2013]. A near-global study of PK•KP – a PKP wave that is back-scattered in the lower mantle onto a second PKP path – suggests a spatial correlation between scattering and LLSVP edges in the lowermost 300 km of the mantle [Rost and Earle, 2010; Frost et al., 2017].

The volume of the mantle that can be investigated for scattering heterogeneity is controlled by the specifics of the seismic probe. PK•KP can be used to investigate the lower mantle close to the CMB [Chang and Cleary, 1981; Rost and Earle, 2010; Frost et al., 2017]. The direct wave PKPPKP (also called P′P′) results from a PKP wave (P′) reflecting from the underside of the surface, back into the Earth as a second PKP wave, along the great-circle path (GCP). This phase can be preceded by scattered energy called PKP•PKP (P′•P′), caused by back-scattering of PKP at any depth in the mantle [Rost et al., 2015]. Like PK•KP, P′•P′ has an unusual scattering geometry (Fig. 1) and can scatter from locations off the GCP, and the P′ segments need not be symmetric to each other. P′•P′ is the continuation of PK•KP towards the surface, thus this phase is able to sample the whole mantle from CMB to crust (Fig. 2). We extend our earlier work and investigate the mantle upwards from the CMB to the surface to deterministically map the vertical and lateral distribution of scattering heterogeneities throughout the mantle. In contrast to other scattering probes, the unusual (and versatile) raypath geometry of P′•P′ allows the study of previously unsampled regions of the Earth.

The internal structure of the Earth and the nature of mantle convection are inherently connected across scales [e.g. Tackley 2015]. The distribution of large-scale mantle structure as imaged by seismic tomography has been investigated using thermo-chemical geodynamic models, which indicate that downwelling of cold, dense slabs at subduction zones moves and shapes the hot, convecting piles of seismically slow material at the CMB, forming the Large Low Shear Velocity Provinces (LLSVPs) [McNamara and Zhong, 2005; Li et al., 2014; Domeier et al, 2016]. The LLSVPs, if compositionally distinct, may modulate mantle dynamics through thermal instabilities that result in mantle plumes that rise up causing hotspot volcanism [Thorne et al., 2004; French and Romanowicz, 2015]. Furthermore, calculations suggest that mantle plumes may be spatially correlated with the LLSVPs [Thorne et al., 2004; Doubrovine et al., 2016]. Geodynamic modelling of thermo-chemical structures in the deep mantle indicates that small-scale heterogeneities (as small as kilometre-sized) can be passively transported in the large-scale flow [Brandenburg and van Keken, 2007; Li et al., 2014, Mulyukova et al., 2015]. Furthermore, geochemical analysis of intraplate volcanism suggests that heterogeneities situated in the deep Earth may be transported to the surface by entrainment in mantle convection [Williams et al., 2015]. Therefore, there is compelling evidence that the distribution of small-scale seismic structure in the mantle is linked to the large-scale structures.

Here we use a global collection of earthquakes recorded at seismic arrays to identify P′•P′ and deterministically locate the position of the causative volumetric scattering heterogeneity within the mantle. We investigate the relationship between scattering heterogeneity and other seismologically imaged structures in the mantle. We use our observations to understand the distribution of small-scale heterogeneities throughout the whole of the mantle, and the connection with dynamic processes.

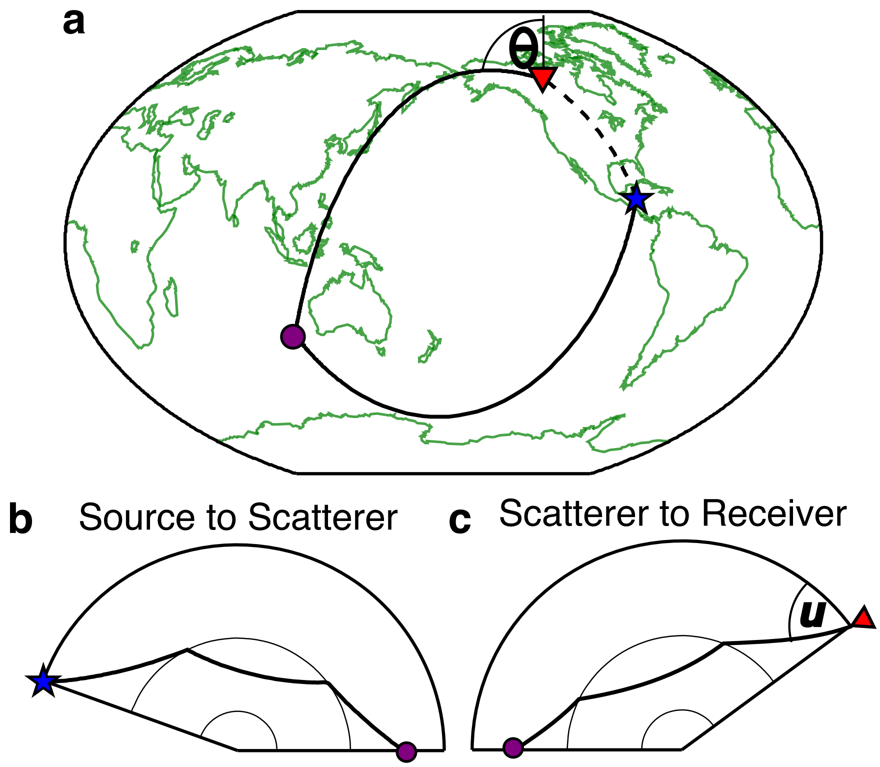

Figure 1: PKP•PKP (P′•P′) example path. **(a)** A P′•P′ path from the source (star) to a scattering point in the mantle (circle) and then to the receiver (triangle). P′•P′ travels along two great-circle paths (solid lines) to and from the scattering point, off the great-circle path between the source and receiver (dashed line). PKP ray paths from **(b)** source to scatterer (PKP_{ab}) and **(c)** scatterer to receiver (PKP_{bc}). The two PKP legs may be symmetric or asymmetric (as in this case) and can scatter from any depth in the mantle from the CMB to the surface. Rays observed at the surface arrive from a specific direction known as the back-azimuth, θ, measured relative to North, or the relative back-azimuth measured from the GCP, and from a vertical incidence angle, referred to as the slowness, *u*.

[SINGLE OR 1.5 COLUMN FIGURE]

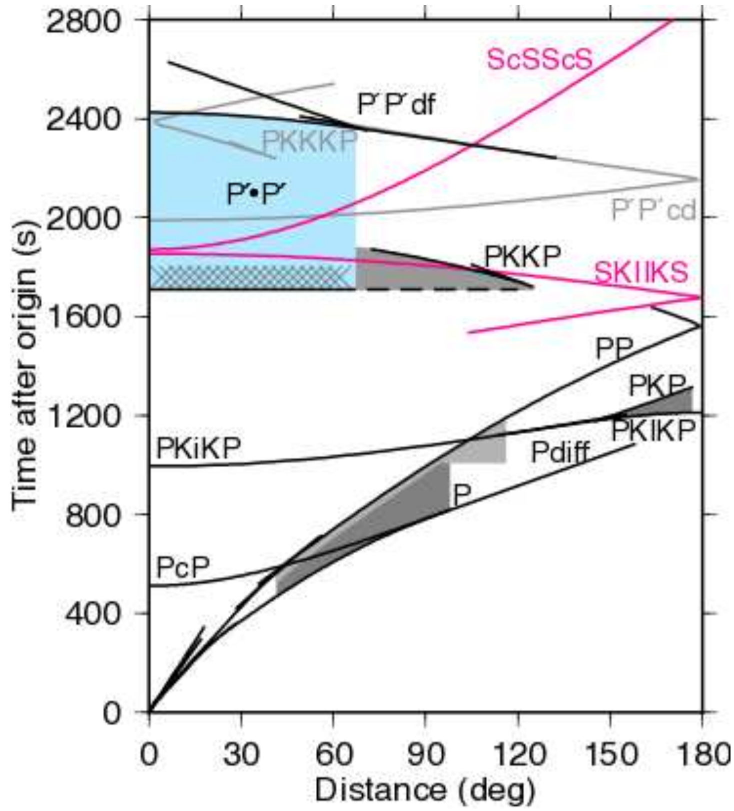

Figure 2: Travel-time curve displaying P′•P′ and other scattered phases in the high-frequency seismic wavefield. Black lines mark major P-wave phases. The blue region marks the time and distance region investigated for P′•P′ waves in this study. Hatched region marks time and distance region investigated for PK•KP in Frost et al., [2017]. Grey and pink lines mark the P- and S-waves, respectively, that may contaminate the P′•P′ study region. Other P- and S-waves are not shown for clarity. Differently shaded grey regions denote time and distance regions previously investigated for other scattered waves. Adapted from Rost et al., [2015].

[SINGLE COLUMN FIGURE]

2. Data

We collect data from 643 earthquakes at any depth with magnitudes M≥6 recorded at up to 12 small and medium aperture arrays within 65° of any event (Fig. 3). The arrays contain a mixture of short period and broadband instruments; we use only the most common instrument type in each array. These arrays were designed to determine the directivity of short-period P-waves thus are ideally suited for analysis of high-frequency scattered waves. The aperture of an array controls its directivity resolution, thus we select only arrays with apertures of 10km to 30 km to ensure that we are able to resolve well the incoming direction of waves.

Each event-array pair has a specific geographical volume of the mantle from which possible P′•P′ scattered waves can be detected (Fig. 3). The size and shape of the sampling region at any given scattering depth is dependent on the event-array distance. Using estimations of the potential scattering volumes combined for all source-array pairs, we develop a "potential sampling density map" of our dataset for different depths, which represents the abundance of scatterers we would detect if the actual distribution of scattering in the Earth distribution were uniform. The potential scatterer sampling distribution of the dataset is uneven, but in contrast to other probes, the southern hemisphere is well covered throughout the depth of the mantle, allowing investigation of the relationship between scattering heterogeneities and the South and Central American subduction zones, and the African and Pacific LLSVPs.

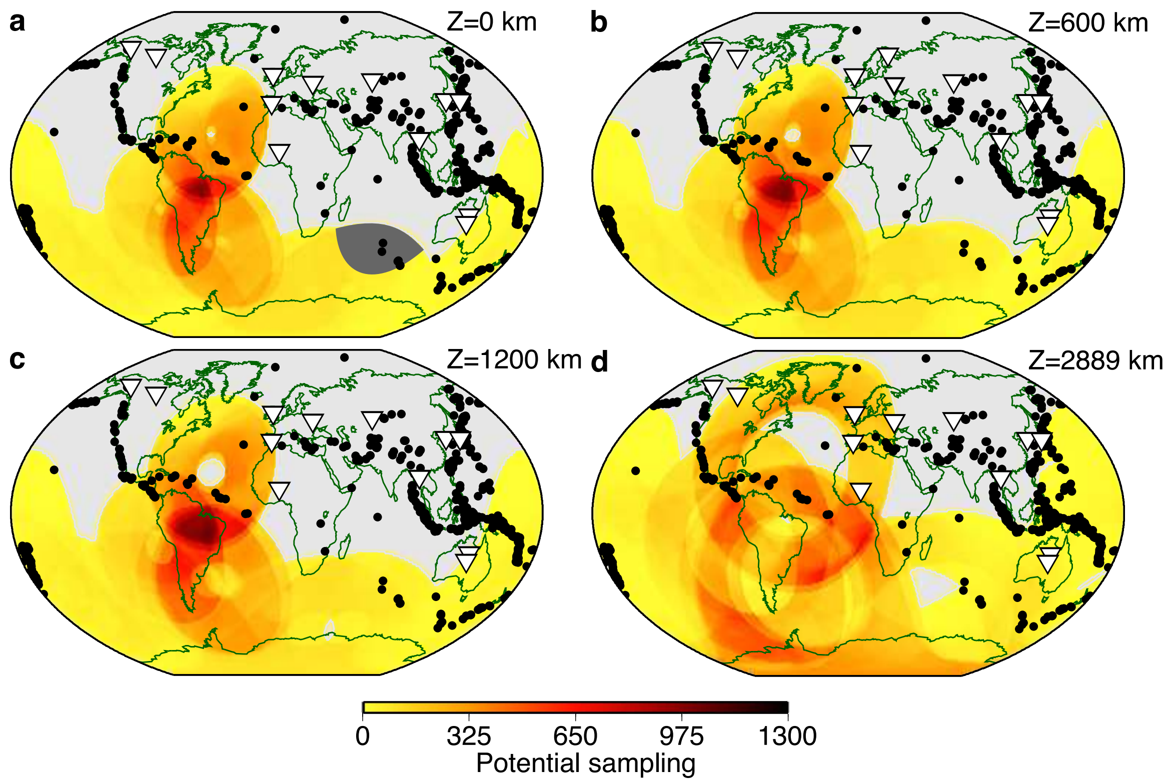

Figure 3: Earthquakes (dots) and arrays (triangles) in our dataset and resultant potential P′•P′ scattering sampling. The 643 events and up to 12 arrays yielded 1715 event-array pairs. Global sampling distributions are constructed by summation of the potential scattering sampling for all source-array pairs at: **(a)** the surface (0 km depth); **(b)** transition zone (600 km depth); **(c)** mid-mantle (1200 km depth); and **(d)** the Core-Mantle Boundary (2889 km depth). Sampling is densest in the mid-mantle and most geographically extensive in the lowermost mantle. Grey wedge in **(a)** displays an example of the potential scattering regions for a single event-array pair.

[2 COLUMN FIGURE]

3. Methods

We investigate energy associated with mantle scattering in a time window from the first possible arrival of P′•P′ at ~1700s after the earthquake origin time (for a surface focus event) corresponding to scattering at the CMB, up to the first possible arrival of the direct wave $P'{\bullet}P'_{df}$ at ~2400s, which is the earliest $P'{\bullet}P'$ GCP phase, reflecting from the underside of the surface on the antipodal side (Fig. 2). $P'{\bullet}P'$ scattering related to interactions with small-scale mantle heterogeneity is feasible for any time and distance in this window (blue shaded region in Fig. 2).

We de-trend the data and discard any discontinuous traces i.e. gaps in the recording. The remaining traces are filtered with a 2nd order bandpass between 0.5 and 2 Hz to enhance the frequencies most associated with small-scale scattering in past studies that investigated frequency content [Mancinelli et al., 2016; Frost et al., 2017]. We search for scattered signals within the wavefield data using fk-analysis (frequency-wavenumber), which performs a grid-search over incoming directions to maximise coherence (the similarity of two or more signals in the frequency domain) of the signal stacked across the array, calculated in the frequency domain [Capon, et al., 1967]. We search over slownesses from 0 to 8 sec/deg and back-azimuths between -180° to 180° relative to the GCP. By selecting signals with the highest coherence we determine the best fitting slowness vector (a combination of the back-azimuth, θ, and the horizontal slowness, $\underline{u}$) of the incoming signals in the scattering search time-window (1700s to 2400s after earthquake origin). To improve the resolution of the slowness vector of incoming signals, as well as to further increase the prominence of signals above the noise, we apply the F-statistic to the fk-analysis (Fig. 4) [Blandford, 1974; Selby, 2011]. The F-statistic calculates the ratio of the amplitude of the stacked signal to the sum of the differences between the stack and each trace used to form the stack. The F-trace has the effect of penalising stacks that differ from individual input traces i.e. signals that are incoherent across the array. Thus, the best fitting slowness and back-azimuth from the grid-search are those that produce the most representative stack of the individual array traces. By performing these calculations in the frequency domain we increase efficiency by reducing the number of transformations required between the time and frequency domains. However, the fk approach returns a single value of coherence from each slowness vector averaged across the whole time window, thus collapsing the time axis. Combining the F-statistic with traditional fk-analysis results in much-improved slowness vector resolution, even for the small-aperture arrays used here [Frost et al., 2013]. Thus the origin of the scattered energy can be more precisely estimated.

We measure the slowness and back-azimuth of the most coherent signals received at the array in consecutive 50 s long time windows (Fig. 4). This window length gives depth resolution comparable to that obtained in global tomography models, and is sufficient to identify broad-scale trends in scattering distribution, both laterally and with depth. We assume the arrival time of a signal to be the middle of the 50 s time window, and given that scattering of P′•P′ from a range of depths can arrive at the array with similar travel-times, each 50 s time window that we investigate is sensitive to scattering from a 50 to 200 km thickness of the mantle. The thickness of the scattering region that each time window is sensitive to decreases with scattering depth hence, at shallower depths, there is overlap in depth sensitivity between windows – adjacent time windows can contain energy scattered from the same depth (albeit from different locations).

Mantle scattered P′•P′ waves are expected to arrive with slownesses between 2.1 and 4.4 s/deg. The range of directions from which P′•P′ waves can be observed is dependent upon the event-array distance and scattering depth. Array analyses permit recognition and omission of contaminating waves by determination of the incoming direction of energy, compared with the directions possible for P′•P′. We compute the expected arrival times for possible contaminating waves: direct phases, depth phases, and multiple reflections of both P- and S-waves. We do not calculate multiples reflecting off upper mantle discontinuities (i.e. a downgoing wave reflecting off the 660 km discontinuity, then reflecting back down from the 410 km discontinuity). Contaminating waves would likely be detected along the GCP (we take both minor and major arc arrivals into account). In contrast, P′•P′ scattered energy most commonly arrives off the GCP, allowing clear identification of the scattered arrivals. However, at short event-array distances it is possible for P′•P′ to arrive along the GCP; these situations can be predicted and extra care is taken to exclude contaminating phases. As there are few phases that can arrive within the P′•P′ window (Fig 2), we would expect few time windows to be contaminated by other seismic phases. Nonetheless, we discard any time window where we both observe a signal within 20 degrees of the GCP (in major or minor arc directions) and any known seismic wave is predicted to arrive in the same time window and along the same backazimuth (i.e. minor or major great-circle path) (e.g Fig. 4d). Of all identified signals, only 2% match the time and direction predicted for known seismic phases, and thus are discarded.

The wavefield may also be contaminated by foreshocks or aftershocks to the analysed events, thus we exclude from further analysis any scattered signals where any magnitude ≥6 earthquake occurs within two hours of the origin time of the studied earthquake (11% of identified scattered signals). As a further test we remove any scattered signal that could be contaminated by a magnitude ≥5 event but find no systematic difference in the distribution of scattering heterogeneity. Our focus on core wave arrivals with slownesses from 2.1 to 4.4 s/deg helps to exclude contamination from smaller, closer events, which have higher slownesses associated with more horizontal incoming energy (and the discarding of GCP signals further minimizes energy from small local events contributing to data we analyse). Therefore, we are certain that our data selection prevents any contamination of the results by local and regional events. Lithospheric scattering directly beneath the array may redirect high slowness contaminating energy to lower slownesses typical of mantle scattering that we consider here. However, the direct contaminating wave would arrive in the same time window as the lithospheric scattered energy, and would likely be more coherent with an obviously inappropriate slowness. This allows a straightforward identification (and removal) of energy scattered from lithospheric structure.

After contaminated time windows have been removed, scattered signals are identified. We pick time windows containing energy prominently above the background noise level in f-k space and consistent with the directivity criteria for P′•P′ scattering (e.g. Fig. 4b). We identify the slowness and back-azimuth of the scattered signal, and select the time at the middle of the 50 s long window as scattered travel time; therefore, we only identify one scattered signal per 50 s window. If multiple P′•P′ signals are observed in the same time window we retain the signal with the highest coherence, as this will be the best spatially resolved. Multiple waves arriving at a similar time, either scattered or direct, may interfere causing the apparent arrival direction of energy at the array to be incorrect. The apparent signal would likely appear blurred across directions, thus we only select signals with tightly resolved slowness and back-azimuth (within the capabilities of the array).

The back-azimuth, slowness, and time information for each scattered signal are used to calculate a scattering location in the mantle. The back-azimuth of a signal indicates the horizontal direction along which the wave travelled while the slowness defines a discrete path for a 1D Earth model, and the travel-time relates to the scattering depth (Fig. S1). Thus there is a trade-off between the distance and depth of a scattered path, hence we attempt to fit both slowness and travel-time simultaneously with a grid search. We ray trace backwards from the array along the observed back-azimuth to a range of possible scattering depths and distances, and then ray-trace from these scattering locations to the source. Possible scattering locations are spaced every 0.01° in distance between the minimum and maximum possible path lengths of PKP along the resolved back-azimuth and 50 km in depth from the CMB to the surface. We model the scattering location by minimising the misfit between the calculated slowness and time for each potential scattering location and the observed values. Mapped scattering heterogeneity locations are discarded if traced rays to the solution location do not well fit the observed slowness and time: if the squared slowness misfit (observed minus predicted) plus twice the squared time misfit is greater than 10, i.e. a weighting factor of 2 is used for travel time misfit and therefore we favour fitting scattering locations with small travel-time misfits. The misfit value selected fits signals within the slowness resolution limit of the arrays. Overall, of the original 4319 identified scattered signals, we discard signals contaminated by other events (11% of the original population), other phases (2%), and poorly fit signals (44%), leaving 1876 mapped scattering heterogeneities.

Due to the uncertainty in travel-time (from using the middle of the 50 s time window) and the uncertainty in slowness (due to the ability of the arrays to resolve the incoming direction) we determine the dimensions of the region that contains the heterogeneity based on these limitations. We calculate scattering locations for signals arriving at the start and end of the 50s time window, and with slowness variation of ±0.3 s/deg relative to that measured at the array (estimated from the slowness spacing of the grid-search). This defines a region around the best fitting heterogeneity location that is, on average, ±100 km laterally and vertically. For mid-mantle scattering at high slowness values (~1000-1800 km depth), the error regions can occasionally grow to values as large as ±800 km laterally and ±500 km vertically but this larger misfit is only relevant for around 5% of the solution scattering locations, thus the majority of the scattering heterogeneities identified in our dataset are located to within ±100 km vertically and laterally.

Sub-surface structure beneath the majority of the arrays used in this study has been demonstrated to have an insignificant effect on the resolved slowness and back-azimuth (Bondar et al., 1999). Nevertheless, removing scattering heterogeneities observed at Chiang Mai array, which is most affected by sub-surface structure, dominantly reduces scattering in the upper 200 km of the mantle and causes no significant change in our conclusions on the relationship with lower mantle structure.

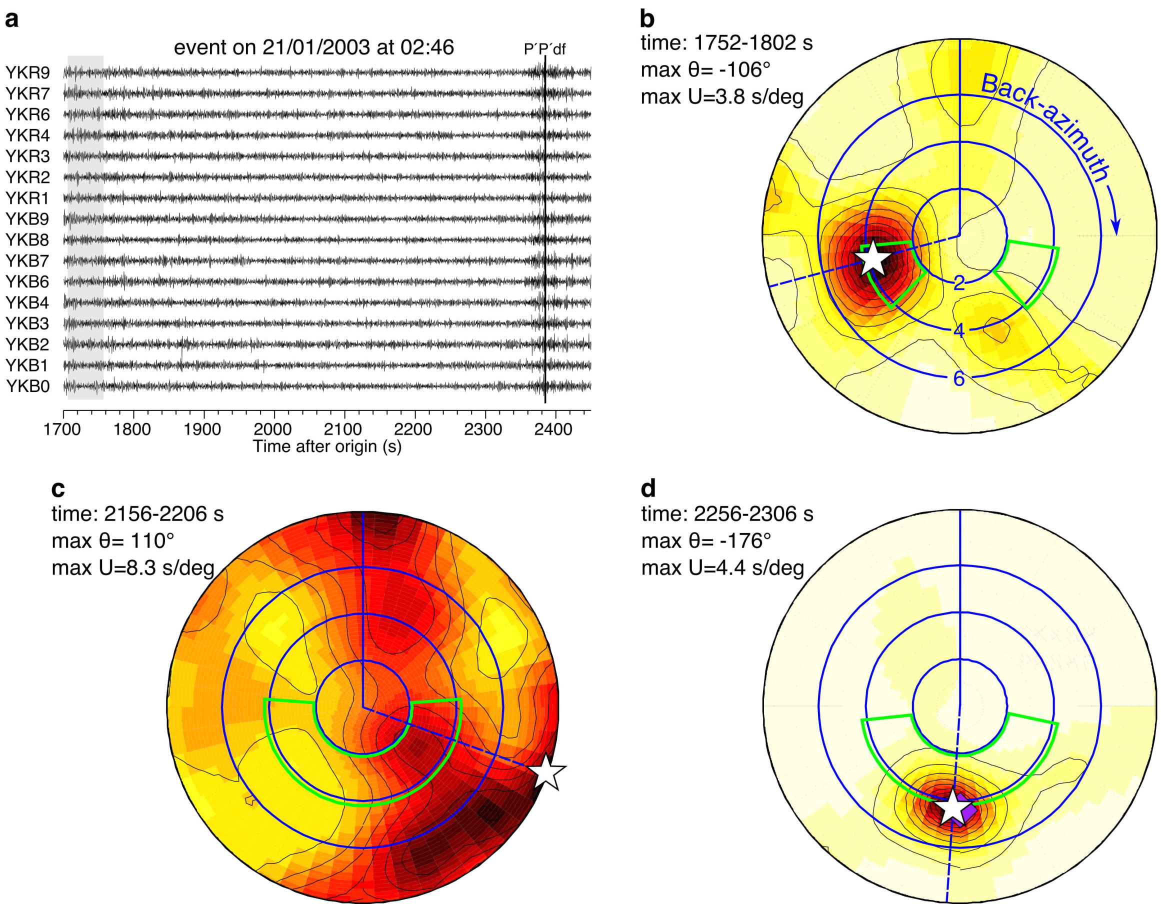

Figure 4: Array data are shown for (**a** and **b**) a magnitude 6.5 event, 24 km depth, 52° away from Yellowknife array, and (**c** and **d**) a magnitude 7.8 event, 0 km depth, 37° away from Warramunga array. **(a)** The time window for P′•P′ scattering (1700-2450 sec for this event, blue region in Fig. 2). The predicted time of the direct phase, P′P′$_{df}$, is shown by the vertical line, marking the end of the scattering window used here. Data are filtered between 0.5 and 2.0 Hz. The grey shaded time window

corresponds to information shown in **(b)**. **(b)** f-k processing of the 50 sec time window shown grey in **(a)**, displayed in terms of back-azimuth (θ , azimuthal axis) and slowness (*u*, radial axis outwards from 0 to 8 s/deg with rings marking 2 to 6 s/deg). Back-azimuth is measured relative to the great-circle path (vertical blue line). The white star shows the maximum coherence in the f-k analysis, arriving with relative back azimuth = -106° (blue dashed line). The 90% coherence contour is roughly ±10° wide in back-azimuth and ±0.5 s/deg in slowness around the maximum. The green regions show the range of possible slownesses and back-azimuths for P′•P′ waves scattering at this distance and the median depth of scattering for this time window (from the shape of the potential scattering regions, grey regions in Fig. 3). **(c)** f-k processing of a time window showing no clear P′•P′ waves. (**d**) f-k processing of a time window that is likely contaminated by the direct phase PKKKP (predicted slowness and back-azimuth marked by purple diamond). Time windows (**c**) and (**d**) are not picked for further processing.

[2 COLUMN FIGURE]

4. Results

The mapped locations of scattering heterogeneities are unevenly distributed in the mantle, both laterally and with depth. This is not unexpected given that the potential sampling capacity of our dataset also varies in location and depth (Fig. 3). We divide the number of mapped scatterers by the potential sampling density (Fig. 3) in order to compare relative scattering density for different regions. This normalised scattering population shows that heterogeneities are distributed throughout the mantle, but more abundant scattering heterogeneity is present in the uppermost and lowermost mantle (Fig. 5). The radial scatterer distribution also shows a small increase in scattering heterogeneity between 600 and 900 km depth, just below the transition zone, and a minimum in the mid-mantle between 1400 to 1800 km depth.

We find that the radial abundance of small-scale scattering heterogeneity matches the RMS amplitude of large-scale tomographic velocities (Fig. 5): scattering is most common and the RMS variation of tomographic velocities is highest in the uppermost and lowermost mantle. This correlation holds roughly for all tomographic models (Fig. S2).

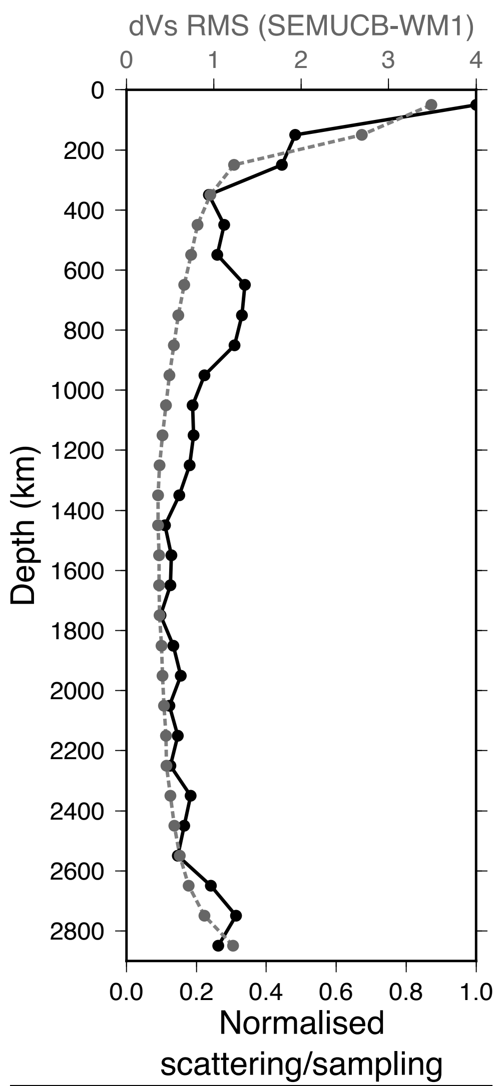

Figure 5: Normalised scattering heterogeneity density with depth (number of scatterers divided by number of samples in each 100 km thick layer) for the complete dataset (black, lower x-axis) and RMS of the shear velocity perturbations from the global tomographic model SEMUCB-WM1 (grey and dashed, upper x-axis) [French and Romanowicz, 2014]. The depth distribution of small-scale scattering heterogeneity roughly correlates with the RMS of long-wavelength dVs perturbations. Both lines are scaled to fit the same axis.

[SINGLE COLUMN FIGURE]

We investigate possible spatial correlation between the resolved scattering heterogeneities and large-scale mantle features, which may be interpreted as proxies for dynamic processes, as in Frost et al., [2017]. We compare the location of scattering heterogeneity to geographical regions beneath hotspots, subducted slabs, regions of high and low tomographic velocities, and regions of high and low lateral tomographic velocity gradients. The high/low velocities and gradients from tomographic models likely relate to the locations of LLSVPs and subducted slabs in the mantle. The spatial locations of scattering heterogeneities are shown, normalised by sampling, in the Supplementary Material, while the absolute latitude, longitude, and depth information for each scattering heterogeneity is shown in Supplementary Table 1

4.1 Relationship between scattering heterogeneities and mantle structure

We compare the distribution of scattering heterogeneity with S-wave tomographic models, both because they are the basis for the definition of the Large

Low Shear Velocity Provinces, and also show consistency between models [Garnero et al., 2016]. We use several tomographic models: GyPSuM [Simmons et al., 2010], SEMUCB-WM1 [French and Romanowicz, 2014], S40RTS [Ritsema et al., 2011], and TX2011 [Grand, 2002]. Additional comparisons with P-wave models are shown in the supplementary material (Figs. S7-9). We calculate lateral velocity gradients from tomographic models, revealing abrupt changes in mantle structure, which thus serve as a proxy for boundaries of the LLSVPs [Thorne et al., 2004, Garnero et al., 2016]. We calculate gradients over a distance of 10° as the resulting gradients well replicate the margins of the LLSVPs found in forward modelling studies [Garnero et al., 2016 and references therein].

We use hotspots from the study of Courtillot et al. [2003]. French and Romanowicz [2015] analysed the tomographic model SEMUCB-WM1 [French and Romanowicz, 2014] and characterised hotspots based on associated tomographic velocity anomalies. We use the 20 hotspots that were labelled as either "primary" or "clear" meaning that the hotspot overlies a column of low velocities from the CMB to 1000 km depth with dVs less than -1.5% or less than -0.5 %, respectively.

We use slab locations from the Regionalized Upper Mantle (RUM) model, which locates slabs at depth using intra-slab seismicity [Gudmundsson and Sambridge, 1998]. When comparing with scattering heterogeneity locations we use slab locations at the surface (zero depth). Slabs move only a small amount laterally as they subduct (≤5° relative to the plate boundary at the surface [Steinberger et al., 2012]), which is unlikely to strongly influence our correlations.

To account for differences in the magnitude, range, and pattern of velocity anomalies and velocity anomaly gradients between tomographic models, and differences in the number of locations in of maps of hotspot and slab locations, we convert maps of tomographic velocities to maps of percentage cumulative area on a sphere sorted by decreasing velocity anomaly (from fast to slow). For example, the 20% area corresponds to the area of the 20% highest tomography velocities of a given depth shell (Fig. 6). We only consider the regions of the tomographic models that match the regions sampled by the P′•P′ dataset at each depth. In this way, the highest and lowest velocities in several tomographic models with inherently differing amplitudes of velocity variation can be directly compared. We establish geographical area percentages associated with the locations of hotspots and subducted slabs by computing the cumulative area surrounding the features within specific distances from them (within the area sampled by the P′•P′ dataset). For example, the first 20% area for slabs corresponds to the region closest to slabs that adds up to 20% of the Earth’s surface area; conversely the last 20% area indicates that amount of surface area furthest from slabs.

To estimate correlations between the abundance of small-scale scattering and subducted slabs, hotspots, tomographic velocities and gradients, we compare the location of these features to the distribution of scattering heterogeneity. For each 100 km depth shell, we count scattering heterogeneities in each 20% area division from the feature of interest. To account for the variability in sampling coverage of our dataset (Fig. 3), we count our estimation of potential scatterers (afforded by our event-array distributions) in the same 20% area regions, and calculate the ratio of the number of observed-to-potential heterogeneities. This allows us to construct a map of normalised scattering prevalence, thus effectively removing the bias of our uneven sampling.

The first set of comparisons is displayed in Fig. 7 as a cumulative histogram as a function of depth. In the upper 200 km of the mantle, scattering heterogeneities are most common in regions of high velocity (Fig. 7a), which is evident from the horizontal width of the light and dark blue shading being greater than the width of the light and dark red shading over the same depth range. In the lower mantle, especially in the deepest 500 km or so, the opposite is true: scattering heterogeneities are more abundant in low velocity regions (as evident by wider red shading). Regions of the lowermost mantle with high seismic velocities show virtually no correlation with scattering heterogeneities. Scattering is slightly more common in regions of high seismic velocity between 600 and 900 km depth.

In the deepest 200 km of the mantle, scattering heterogeneities are more common in regions of high lateral seismic velocity gradients (Fig. 7b: the width of the black and dark blue shading is significantly greater than the light green colors). In the lowest ~1000 km of the mantle, scattering heterogeneities are in greater abundance in the 20% area around hotspots than in any other bin; there is also a slight increase in mapped heterogeneities beneath hotspots in the mid-mantle between 600-900 km depth (see the wide red colors shading, Fig. 7c). Our mapped scattering heterogeneities show little correlation with regions surrounding the surface location of slabs, except in areas furthest from slabs in the 600-900 km depth range (indicated by the wide orange-yellow shading, Fig. 7d). In the upper

200 km of the mantle, scattering strongly correlates with high seismic velocities and proximity to slabs (Figs. 7a and 7d, blue and yellow shading, respectively), which, at these shallow depths is most closely related to the location of continents. While the precise locations of the heterogeneities is different, the heterogeneities resolved with P′•P′ show a very similar distribution in the lowermost 300 km of the mantle to those heterogeneities resolved with PK•KP in an earlier study [Frost et al., 2017].

To test the robustness of these correlations we determine how likely they are to have been produced by chance. We rotate the tomographic models (of velocity and lateral gradient), and hotspot and slab locations by a random angle about a randomly located pole of rotation. We then recompute the correlations between the rotated geographical features and the distribution of the unrotated scattering heterogeneities. The random rotation is repeated 200 times for each tomography model, as well as for the hotspot and slab locations, to calculate the range of possible correlations. The mean and standard deviation of the range of correlations at each depth is computed, assuming Gaussian statistics. We compare this with the original, unrotated data in Fig. 8, and consider any correlation to be statistically significant if the correlation value between scatterers and regions of the unrotated phenomena plots outside one standard deviation from the mean correlation of the rotated phenomena (demonstrating that at least 84% of the random correlations are a lower value). When we do not assume a distribution and instead calculate the proportion of samples above and below one standard deviation of the data, we find very similar patterns of significant observations. Using this metric, we define the following correlations as significant and unlikely the product of chance:

(1) An increased correlation with scatterers in regions of low velocity at depths greater than 1800 km (solid red line in the left panel of Fig. 8a)

(2) An increased abundance of scattering heterogeneities in regions of high velocity gradient in the deepest few hundred km of the mantle, as well as between 1600-2000 km depth (solid red line in the right panel of Fig. 8b).

(3) An increased abundance of heterogeneities close to surface hotspot locations at depths greater than 2100 km depth (solid red line in left panel of Fig. 8c).

(4) A decreased abundance of heterogeneities far from surface hotspot locations at almost all depths greater than 800 km depth (solid red line in right panel of Fig. 8c).

There is no significant correlation seen between scatterer locations and slab locations, except an increase in correlation between heterogeneities and large distances from slabs between 600 and 900 km depth, which matches the depth range of the increased correlation with low velocity gradients (solid red lines in left panel of Fig. 8b and right panel of 8d), and an increased correlation between heterogeneities and large distances from slabs throughout much of the lower mantle (which is what one expects if correlations are strong for low velocities).

4.2 Dependence upon choice of model

When comparing small-scale scattering locations with tomographically derived high or low velocities, the results may depend upon the choice of the tomography model. In our previous analysis, we compared the distribution of scattering heterogeneities to tomography model SEMUCB-WM1 [French and Romanowicz, 2014]. We further explore the relationship between our mapped finescale scattering heterogeneities with large-scale structures in other tomography models: GyPSuM, S40RTS, and TX2011 (Figs. S4-6 and S7-9 for P-wave models). We find small differences in precise depths and magnitudes of correlations with different models, but the correlation between scattering and low velocities at depths below 1600 km, and with high velocities at depths of 200 km and shallower and the robustness of these correlations are consistent between models.

To test the dependence of correlation on the pattern of hotspots, in addition to comparing with rotated hotspot locations, we create a population of randomly located hotspots, equal in number to the primary and clear hotspots from Courtillot et al., [2003] and French and Romanowicz [2015]. We find that a synthetic population generates no preferential spatial correlation with the scattering heterogeneities (Fig. S10). Furthermore, when the population of random hotspot locations is rotated to test the robustness of the correlation, the correlation of the random population very often falls well within the one standard deviation range of the rotated data (Fig. S12). This implies that the observed correlation between hotspot locations and scattering heterogeneities in the lower mantle is caused by the specific distribution of hotspots.

We test the influence of our decision to use only the surface slab locations of the RUM model. We calculate the spatial correlation between scattering heterogeneities and slab locations as described above, but use slab locations at the depth of the heterogeneity. When considering scattering heterogeneities at depths greater than that which the slab is mapped to we use the location of the slab at the last mapped depth and project this position vertically down to the CMB. This method of vertical extrapolation likely still misrepresents the locations of slabs: some amount of lateral movement at greater depths is evident in tomographic and geodynamic models but is typically on the order of a few degrees [e.g. French and Romanowicz, 2014 and Steinberger et al., 2012]. Nonetheless, we find no significant difference in the correlations between using the surface slab location and slab locations with depth (Fig. 7 and Figs. S11 and S12).

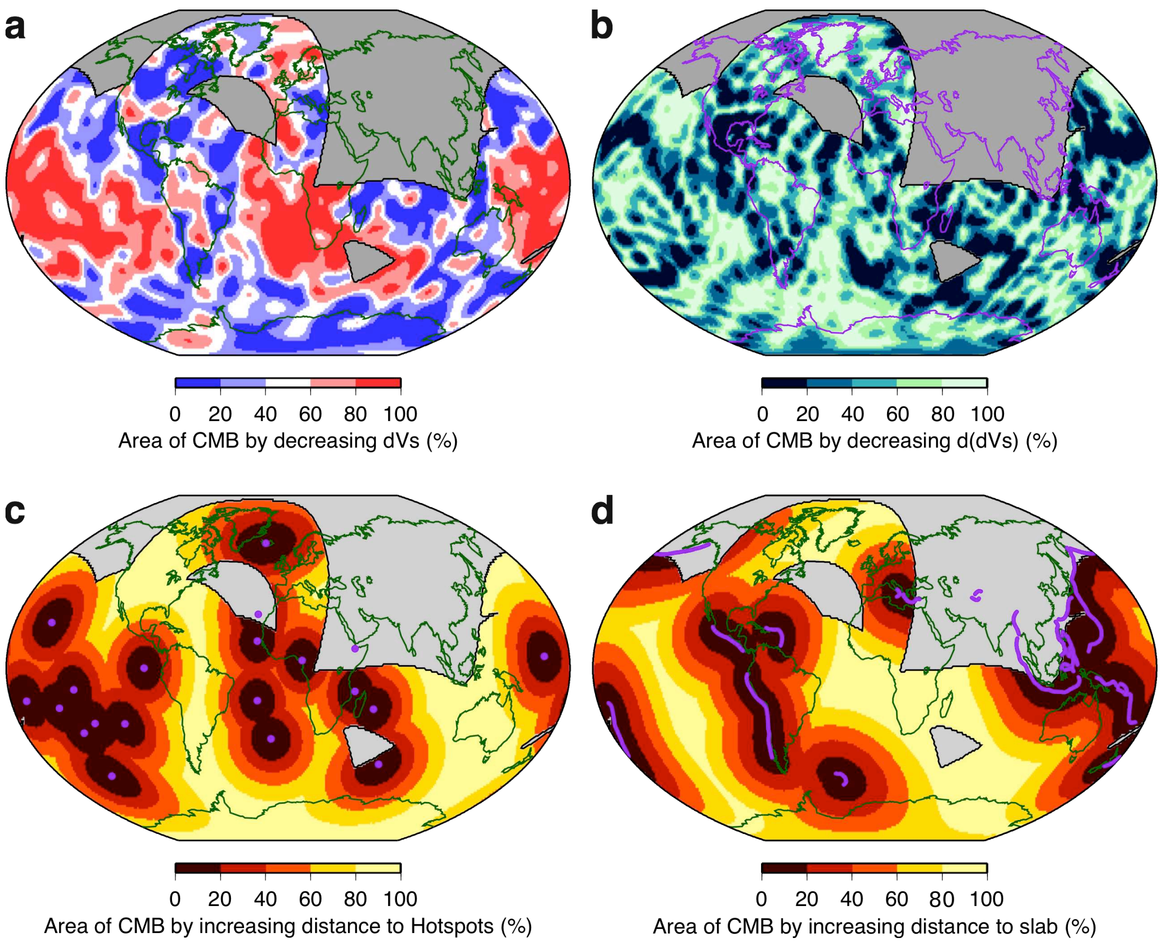

Figure 6: Tomography and distance from subduction zones and hotspots by percentage area calculated for sampling at the CMB. (a) The magnitude of the velocity anomalies in SEMUCB-WM1 [French and Romanowicz, 2014] in the area sampled by our dataset at the CMB displayed by decreasing anomaly (from fast blue

areas to slow red areas) in regions occupying 20% of the area of the CMB. (b) The magnitude of the lateral velocity gradient decreasing from high to low in 20% area regions. (c) Distance from hotspots (connected to plumes identified as either primary or clear in the analysis of French and Romanowicz [2015]). (d) Distance from slabs (at zero depth slice in RUM [Gudmundsson and Sambridge, 1998]. Black line marks the extent of the sampled area (as in Figure 3d).

[2 COLUMN FIGURE]

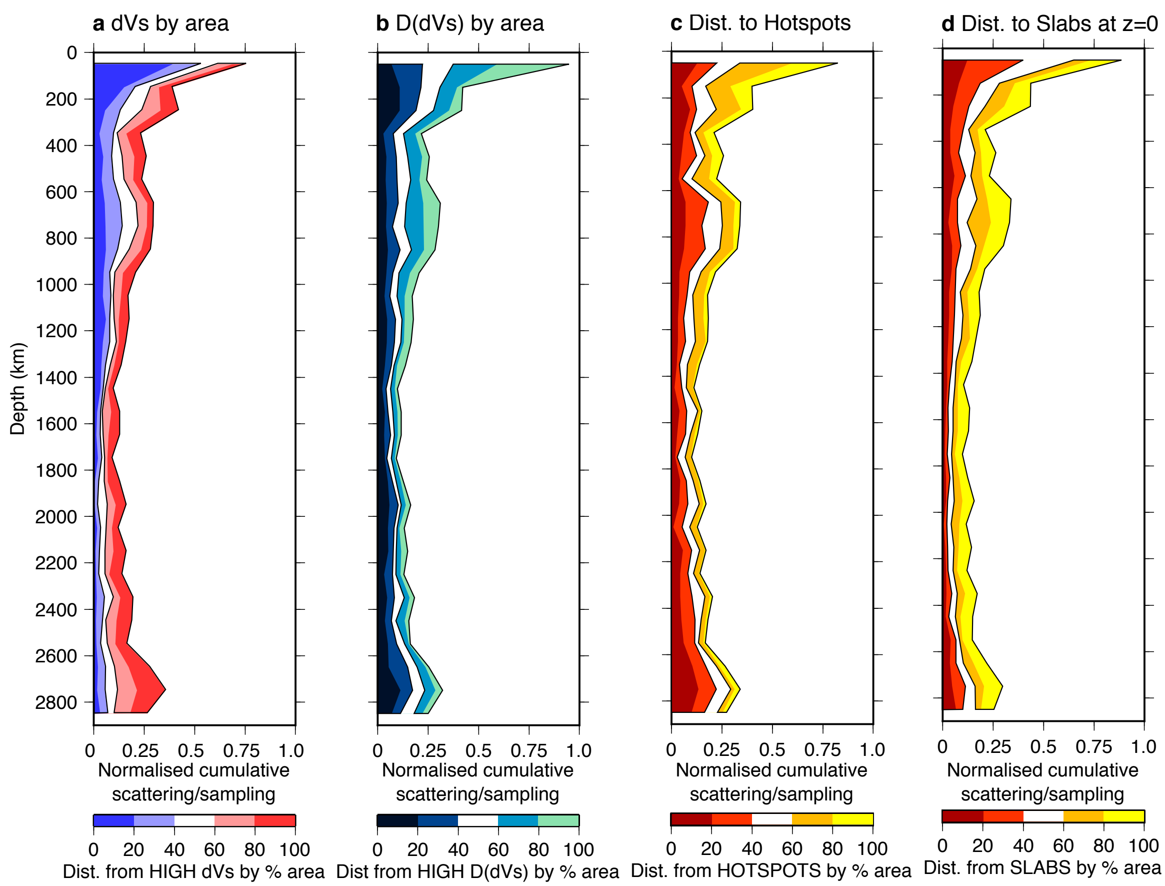

Figure 7: Scattering abundance (bar width) with depth compared with the distribution of large-scale heterogeneity throughout the mantle (colour scale). Scattering abundance is calculated cumulatively across all areas, is divided by

sampling, and is normalised to unity, representing the maximum scattering abundance at any depth. **(a)** Scattering heterogeneity and tomographic velocity anomalies (from SEMUCB-WM1 [French and Romanowicz, 2014]) sorted from highest (blue) to lowest (red) measured as a function of surface area in 20% area bins. **(b)** Scattering heterogeneity and lateral tomographic velocity gradient sorted from highest (dark blue) to lowest (light green). **(c)** Scattering heterogeneity and distance from hotspots from low to high (red and yellow, respectively). **(d)** Scattering heterogeneity and distance from slabs from low to high (red and yellow, respectively). Scattering heterogeneity in the lower mantle shows an affinity for both low seismic velocities and hotspots. Black lines encapsulate the highest and lowest 40% area regions.

[2 COLUMN FIGURE]

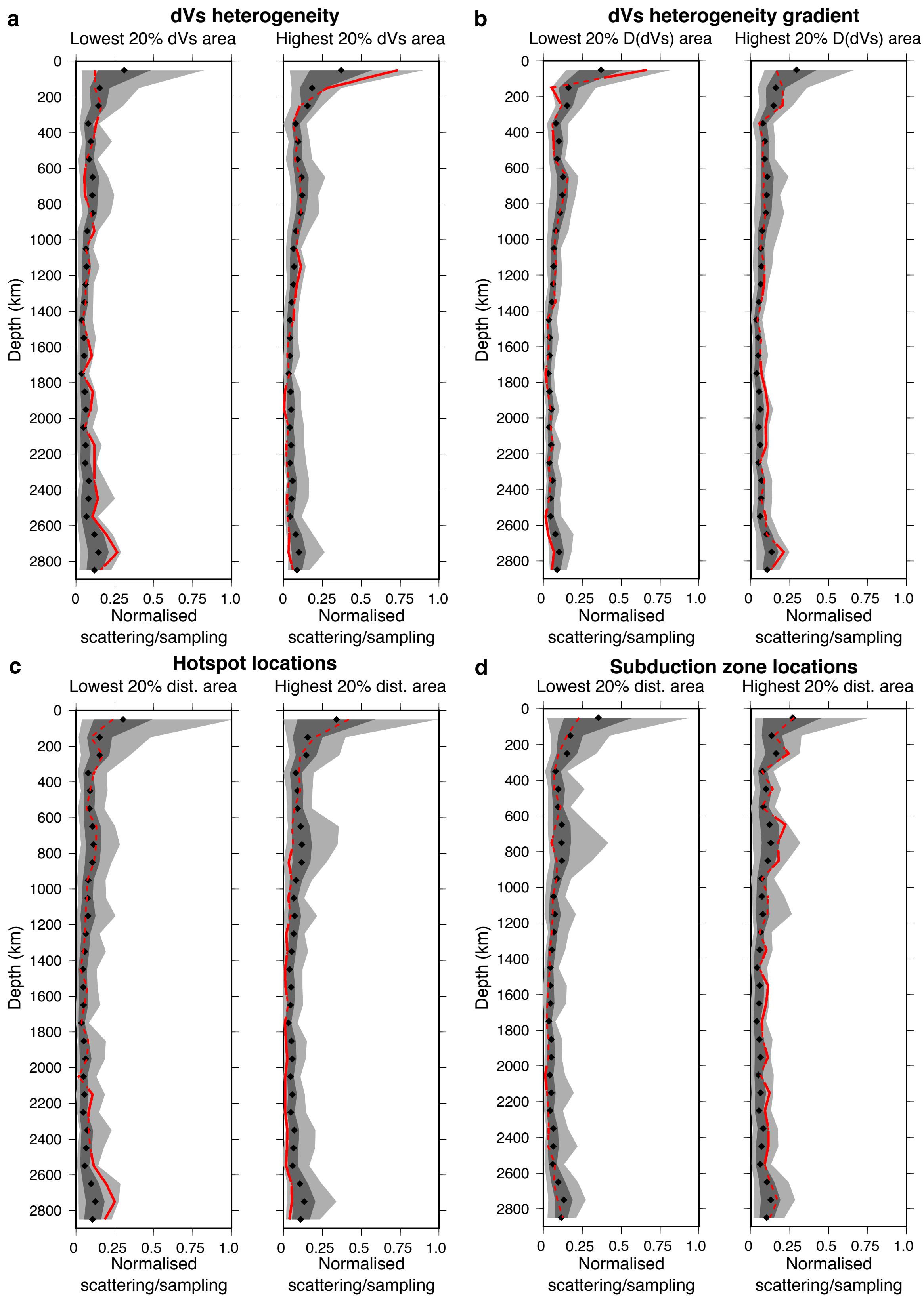

Figure 8: Scattering abundance with depth, divided by sampling, showing the unrotated model (red line) compared with rotated models (grey). The unrotated

model (red line) is dashed when within one standard deviation (dark gray shading) of the mean of the spatial correlations (black diamonds) with the rotated models, and solid when outside this level. The lighter shaded region marks the range of all correlations with the randomly rotated phenomena. Comparisons are shown for: **(a)** tomographically derived velocity heterogeneity from SEMUCB-WM1 for the 20% area corresponding to the lowest (left panel) and highest velocities (right panel). Correlation between increased scattering abundance and low velocities appears robust in the deepest mantle, and correlation to high velocities is robust in the shallowest 200 km of the mantle, as well as around 1200 km depth. **(b)** As in **(a)** except correlations are between observed scattering and rotated shear velocity gradients in model SEMUCB-WM1. Correlations are most significant for the strongest gradients (right panel) at the base of the mantle. **(c)** As in **(a)** except correlations are between scatterers and distance to rotated hotspot regions. Correlations are most significant in the deepest mantle in close proximity to being beneath hotspots (left panel). **(d)** as in **(c)** except correlations are between scatterers and distance to rotated slab regions. Our random rotation test shows no significant correlation between scatterers and proximity to slabs.

[2 COLUMN FIGURE]

5. Discussion

In this study we mapped scattering heterogeneities and explored their geographical relationship to tomographic velocities and gradients, as well as hotspots and slabs. Our results may be interpreted in terms of the distribution of mantle heterogeneity, which we will discuss here.

5.1 Possible origins of scattering heterogeneity in the mantle

We observe scattering from small-scale heterogeneity throughout the mantle, with increased heterogeneity at the top and bottom of the mantle. Seismic waves can be scattered by volumetric heterogeneity with sharp impedance contrasts, when the heterogeneity has a minimum scale length comparable to the wavelength of the incident wave. Our method is not capable of resolving the precise partitioning of the incident wavefield into scattered versus transmitted energy, since we do not have a consistent reference phase to compare to the amplitude of the scattered wave. Thus we are unable to constrain the properties of the heterogeneities (e.g. impedance contrast). Nonetheless, the frequencies of waves that we study (between 0.5 and 2.0 Hz) imply that observed scattering heterogeneities have a minimum scale length of one to tens of km.

A variety of structures could scatter the energy observed in our data. We can use the distribution and sizes of scattering heterogeneities to address the feasibility of possible causes. Material undergoing phase changes such as from bridgmanite to post-perovskite (pPv) in the lower mantle (or the back transformation) [Murakami et al., 2004; Oganov and Ono, 2004], as well as transitions of olivine to wadsleyite to ringwoodite to perovskite through the upper mantle transition zone could provide an impedance contrast with the ambient mantle. The bridgmanite to pPv phase transition is predicted to occur in the deepest few 100 km of the mantle, and only in relatively cold regions of the mantle for a standard pyrolitic composition, thus would not be appropriate to explain scattering at all depths and locations, unless mineralogical alterations are considered [Lay et al. 2006]. The phase transition is controlled by temperature, composition, and pressure. While pressure is assumed hydrostatic, local changes in composition, perhaps by contamination of the mantle by subducted mid-ocean ridge basalt (MORB), may influence the pPv transition [Grocholski et al., 2012], possibly causing the transition to occur locally in the vicinity of the MORB contamination. Metastability of phase transitions due to chemical heterogeneity [Catalli et al., 2009] could allow transformed minerals to persist outside of their expected stability range. High thermal conductivity in the lower mantle [Stackhouse et al., 2015] renders small-scale temperature changes an unlikely cause of spatially limited occurrence of the pPv transition. While many morphologies and scale lengths of pPv regions can be envisioned that could contribute to wave field scattering observed here, the details of such processes are not constrained. However, pPv should not be stable in the upper mantle, and thus cannot explain observed scattering there. Nonetheless, pPv remains a viable contributor to wavefield scattering in the deepest mantle.

The subduction process continuously introduces compositional heterogeneity into the mantle. Scattering has previously been mapped in the upper mantle and lower mantle in the proximity of subduction zones [Kaneshima and

Helffrich, 1998; Rost and Earle, 2010; Miller and Niu, 2008; Bentham and Rost, 2014]. We do not observe a robust preference of scattering heterogeneity in upper mantle regions of subduction over other regions. While we do observe slightly increased scattering in regions associated with subduction at around 600 to 900 km depth (Fig. 7a), this does not appear to be statistically significant (Fig. 8a, right panel). Nonetheless, the increased concentration of scattering heterogeneity between 600 and 900 km depth shows robust spatial correlation with regions away from subduction zones and areas of low amplitude lateral velocity gradient (Figs. 7 and 7). In some tomographic models subducting slabs are observed to flatten at a similar depth, between ~800-1200 km depth [e.g. French and Romanowicz, 2015].

Oceanic crust may be responsible for scattering throughout the mantle. Subducted oceanic crust may remain unmixed due to slow chemical diffusion rates [Olson et al., 1984] and is only homogenised into the mantle through mechanical stirring. If the observed scattering heterogeneities are oceanic crust then the dispersal of heterogeneities throughout the mantle must be faster than stirring and removal of heterogeneities since scattering heterogeneity is also observed in regions that have not been influenced by subduction for a long time.

The iron spin transition affects the velocity and density of iron-bearing mantle materials [Lin et al., 2005]. Recently, this has been observed to occur over a 60 GPa pressure range (~600 to 2000 km depth) [Holmstrom and Stixrude, 2015] and thus would likely not generate discrete heterogeneities capable of causing scattering.

Products of chemical reactions between core and mantle materials are predicted to have physical properties in contrast with the ambient mantle [Knittle and Jeanloz, 1989] thus may be capable of causing seismic scattering. Experiments demonstrate that such mantle material enriched in iron would likely be denser than the ambient mantle [Wicks et al., 2010]. An interesting possibility is the development of a reaction product layer that would inhibit further interaction with the core; for this case, products are likely to be constrained to a very limited thickness close to the CMB, on the order of a few meters to kilometers [Kanda and Stevenson, 2006]. However, flow in the deep mantle could generate thicker accumulations of reaction products [Mao et al., 2006], which could scatter waves. In addition, ULVZs are commonly imaged to have vastly reduced seismic velocities of up to -10% dVp and -30% dVs, and increased density of +10-20% relative to the surrounding mantle [e.g., McNamara et al., 2010]. Partial melt of mantle material has been proposed as an explanation of ULVZs [Williams and Garnero, 1996]. Partial melt may be denser than the solid state [Ohtani and Maeda, 2001] as well as having strongly reduced seismic velocities. While ULVZs and CMB reaction products could explain deeper scattering heterogeneities, simulations have suggested that dense material may also be entrained up to 200 km above the CMB, dependent on density, viscosity, and vigor of mantle flow [Bower et al., 2011]. CMB topography or roughness might cause scattering [Chang and Cleary, 1981; Mancinelli et al., 2016], but this could not explain heterogeneities we map up off of the CMB throughout the mantle. LLSVPs may be compositionally distinct from the surrounding mantle [e.g. Garnero et al., 2016], and dynamical flow models predict that the LLSVP material will be gradually entrained into mantle flow on small length scales [Li et al., 2014; Williams et al., 2015; Mulyukova et al., 2015]. Thus, depending on the LLSVP properties and entrained heterogeneity scale, this process might give rise to scattering. Geodynamic models also predict that surrounding ambient mantle material can be downward entrained into the LLSVPs, thus offering an origin of scattering within LLSVP regions.

5.2 Distribution of scattering heterogeneity

The distribution of small-scale volumetric heterogeneities is likely strongly dependent on the dynamic properties and processes within the Earth. In numerical simulations of mantle dynamics small-scale heterogeneity, particularly that derived from subducted oceanic crust, tends to be concentrated in regions of upwelling from the lower mantle around plumes and downwelling from the surface around subduction zones (Fig. 1 of Li et al., [2014]). The same focusing beneath upwellings is expected for basal heterogeneities [McNamara et al., 2010] (e.g., compositionally distinct ULVZ material, CMB reaction products, and entrained LLSVP material). Furthermore, large-scale mantle heterogeneity may influence radial small-scale heterogeneity distribution by modifying the convective flows in which the heterogeneities could be entrained [Li et al., 2014].

As wavelength at some fixed frequency is a function of the local velocity, which changes with depth, and the wavelength of scattering structure that can be resolved is dependent on the incident frequency, it follows that in band limited data, the resolvable scattering wavelength changes with depth. We filter all data between 0.5 and 2.0 Hz, therefore, we resolve scattering heterogeneity with wavelengths between about 7-28 km at the CMB, decreasing to about 3-12 km at the surface. Stirring of initially larger-scale heterogeneity is suggested to lead to a cascade of heterogeneity sizes, increasing in abundance with decreasing scale [Olson et al., 1984]. A previous study of the scale of scattering heterogeneities in the lowermost mantle found the most common scale-length to be 4-7 km, but other scales were also present [Frost et al., 2017]. Despite the limited frequency range used in this study, we are likely imaging heterogeneity of a similar size (around 7 km) throughout the mantle.

The similarity between scattering heterogeneity abundance and tomographic amplitude (Fig. 5) may arise from processes relating to convection and chemical differentiation that likely generate strong lateral velocity variations on continental scales and smaller through stirring and diffusion. Lower mantle anomalies manifest at a range of spatial scales (LLSVPs, ULVZs, D″, CMB reaction products), and stirring and entrainment may further decrease their size [Olson et al., 1984, Li et al., 2014], leading both to high-amplitude large-scale velocity anomalies and abundant small-scale scattering. Upper mantle heterogeneity related to subduction, magmatism, and convective processes are also likely to occur across scales. In addition to increased scattering at the top and bottom of the mantle, we also observe a slight but marked increase in scattering abundance from 600-900 km depth, independent of the tomographic velocity structure, which may relate to slab subduction processes or large-scale vertical viscosity changes [Rudolph et al., 2015].

6. Conclusion

Through analysis of the high-frequency seismic wavefield we map the distribution of small-scale seismic heterogeneity, on the order of ~1-10 km in size, throughout Earth's mantle. We deterministically locate vastly more scattering heterogeneities than has been done previously, significantly improving our understanding of small-scale mantle structure. The spatial distribution and scale-length of this scattering heterogeneity suggests it may be the product of several on-going processes in the mantle. These include oceanic crust disseminated throughout the mantle, entrainment of basal heterogeneities such as ULVZ material or core-mantle reaction product, and compositionally distinct LLSVP material swept into mantle flow. Subducted MORB may suitably explain all scattering observations without scattering contributions from other sources. However, we cannot rule out that scattering is caused by a mixture of heterogeneities with different origins in different regions and depths. While small-scale heterogeneity appears present in much of the mantle, we find increased scattering heterogeneity within the uppermost 200 km of the mantle and the lowermost 300 km of the mantle, similar to heterogeneity amplitudes seen in tomography models. We find no statistically significant correlation between scattering and subducting slabs in the upper 1000 km of the mantle. In the lower mantle (from around 1500 km depth down to the CMB), scattering is most common in regions related to the LLSVPs and close to deeply sourced mantle hotspots. Meanwhile, scattering is rare in regions far from deeply sourced mantle hotspots. This suggests that large-scale convective lower mantle structures may entrain and concentrate small-scale heterogeneity in regions of upwelling, downwelling, and stagnant flow.

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