

# 1 X-discontinuity and transition zone structure beneath 2 Hawaii suggests a heterogeneous plume

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## 7 **Abstract**

8 The Hawaiian Island chain in the middle of the Pacific Ocean is a well-  
9 studied example of hotspot volcanism caused by an underlying upwelling  
10 mantle plume. The thermal and compositional nature of the plume alters  
11 the mantle phase transitions, which can be seen in the depth and amplitude  
12 of seismic discontinuities. This study utilises > 5000 high quality receiver  
13 functions from Hawaiian island stations to detect P-to-s converted phases  
14 to image seismic discontinuities between 200 to 800 km depth. Common-  
15 conversion point stacks of the data are used to map out lateral variations in  
16 converted phase observations, while slowness stacks allow differentiation be-  
17 tween true conversions from discontinuities and multiples. We find that the  
18 410 discontinuity is depressed by 20 km throughout our study region, while  
19 the main 660 is around average depth throughout most of the area. To the  
20 southwest of the Big Island we observe splitting of the 660, with a major peak  
21 at 630 km, and a minor peak appearing at 675 km depth. This is inferred  
22 to represent the position of the hot plume at depth, with the upper disconti-

23 nuity caused by an olivine phase transition and the lower by a garnet phase  
 24 transition. In the upper mantle, a discontinuity is found across the region at  
 25 depths varying between 290 to 350 km. Identifying multiples from this depth  
 26 confirms the presence of a so-called X-discontinuity. To the east of the Big  
 27 Island the X-discontinuity lies around 336 km and the associated multiple  
 28 is particularly coherent and strong in amplitude. Strikingly, the discontinu-  
 29 ity around 410 km disappears in this area. Synthetic modelling reveals that  
 30 such observations can be explained by a silica phase transition from coesite  
 31 to stishovite, consistent with widespread ponding of silica-saturated material  
 32 at these depths around the plume. This material could represent eclogite en-  
 33 riched material, which is relatively silica-rich compared to pyrolite, spreading  
 34 out from the plume to the east as a deep eclogite pool, a hypothesis which  
 35 is consistent with dynamical models of thermochemical plumes. Therefore  
 36 these results support the presence of a significant garnet and eclogite com-  
 37 ponent within the Hawaiian mantle plume.

38 *Keywords:* Mantle, Discontinuities, Conversions, Hawaii, Eclogite

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## 39 **1. Introduction**

40 The mantle plume hypothesis can account for many key features of the hot  
 41 spot volcanism that has formed the Hawaiian-Emperor Seamount chain (e.g.  
 42 Wilson, 1963). Originally, mantle plumes were thought to be purely ther-  
 43 mal upwellings, but over the past few decades new evidence suggests that  
 44 mantle plumes carry a compositional component that is anomalous to the

background mantle. For the Hawaiian plume, this evidence comes from geo-  
chemical analysis which suggests its basalts have been derived from anomalous mineralogy in its source and contain traces of recycled oceanic crust and marine sediments (e.g. Hofmann and White, 1982; Hauri, 1996; Eiler et al., 1996; Sobolev et al., 2005). Geographical variations in the geochemistry of Hawaiian basalts have been linked to spatial variation in the proportion of non-peridotitic material in the mantle source regions (e.g. Sobolev et al., 2005; Frey et al., 2016; Herzberg, 2010; Weis et al., 2011). The basalts on the southwest end of the Big Island are part of the most recent expression of a chain of volcanoes with a distinctive chemistry that is possibly linked to the enhanced contribution of recycled basaltic material to their source regions (the so-called ‘Loa’ chain). While basalt in the northeast appears to have a greater contribution from melts of peridotitic mantle (the ‘Kea’ chain) (Sobolev et al., 2005).

Additionally, global seismic tomography shows the Hawaiian plume is broader in the lower mantle than expected for a purely thermal plume (French and Romanowicz, 2015). The regional seismic tomographic model of Cheng et al. (2015) shows a broad, low velocity zone across the upper mantle, which could be explained by ponding of the plume. At the core-mantle boundary, there is seismic evidence of compositional heterogeneity, which could represent the source or anchor of the Hawaiian plume (e.g. Garnero et al., 2016).

Dynamical models show how a plume composition enriched in recycled eclogite causes ponding of plume material above 410-km, creating a so-called

68 Deep Eclogitic Pool (DEP, Ballmer et al., 2013, 2015; Dannberg and Sobolev,  
69 2015).

70 One way to elucidate the thermo-chemical nature of the plume and its  
71 dynamics across the upper mantle is to image the mantle’s seismic discon-  
72 tinuities. These discontinuities are sharp changes in wave speed caused by  
73 changes in mantle material properties. The two main seismic discontinuities  
74 are around depths of 410-km and 660-km, and are associated with the phase  
75 transition of olivine to wadsleyite (Katsura and Ito, 1989) and dissociation of  
76 ringwoodite (Ito and Takahashi, 1989), respectively, in an olivine-dominated  
77 mantle. We will refer to these transitions as the 410 and the 660, and the  
78 region between them as the Mantle Transition Zone (MTZ).

79 The olivine phase transitions associated with the 410 and 660 discontinu-  
80 ities have opposite Clapeyron slopes in temperature-pressure space. In hot  
81 regions the 410 becomes depressed and the 660 is uplifted, leading to a thin  
82 MTZ; in cold regions the 410 is uplifted and the 660 is depressed, leading  
83 to a thick MTZ. Therefore, if mantle plumes were purely thermal features,  
84 mapping the MTZ thinning beneath a plume could be used as a thermometer  
85 for mantle temperature.

86 However, complications arise around 660-km depth, where, in addition  
87 to the olivine phase transition, there is a transition in majorite garnet with  
88 an opposite sign, a positive Clapeyron slope (e.g. Hirose, 2002; Liu et al.,  
89 2018). The discontinuity caused by this phase transition can dominate the  
90 seismic image if garnet is stable (Yu et al., 2011). Garnet stability occurs



91 at higher temperatures or in basalt enriched compositions (e.g. Xu et al.,  
 92 2008; Stixrude and Lithgow-Bertelloni, 2011). A discontinuity created by this  
 93 phase transition would be depressed in hot regions. Several seismic studies  
 94 have observed two discontinuities around 660-km, a so-called splitting of  
 95 the 660, suggesting both phase transitions are occurring (e.g. Andrews and  
 96 Deuss, 2008), while other studies observe a single deeper discontinuity in  
 97 plume regions (e.g. Jenkins et al., 2016).

98 Previous studies of the Hawaiian region, based either on MTZ thinning  
 99 or mapping slow velocity zones in tomographic models, show remarkable in-  
 100 consistency in their estimates of the position of the plume. In terms of P-to-s  
 101 converted phase studies, some (Li et al., 2000; Shen et al., 2003; Wölbern  
 102 et al., 2006) find maximum MTZ thinning to the south and southwest of  
 103 the Big Island, whereas Huckfeldt et al. (2013) find maximum thinning to-  
 104 wards the southeast and Agius et al. (2017) find thinning under north-central  
 105 Hawaii.

106 Seismic discontinuities that are not observed globally can indicate the  
 107 presence of compositional heterogeneity. For example, at around 300-km  
 108 depth in some regions around the globe, a discontinuity - named the X dis-  
 109 continuity - is present. A discontinuity around this depth beneath Hawaii has  
 110 been observed with ScS reverberations (Courtier et al., 2007) and SS precur-  
 111 sors (Deuss and Woodhouse, 2002; Schmerr et al., 2013; Schmerr, 2015). This  
 112 has been associated with various phenomena, including: a phase transition  
 113 in silica from coesite to stishovite, a crystallographic change in orthopyrox-

ene, the formation of hydrous phase A - a dense magnesium silicate - or the reaction of forsterite + periclase into anhydrous-phase B. It follows that if the X-discontinuity is detected then this has implications for the composition and dynamics of the mantle.

In this study we image both the MTZ and upper mantle structure beneath Hawaii using over 5000 P-to-s converted phases or receiver functions (RFs). We interpret our observations in relation to mineral physics and geodynamics which further highlights the thermochemical nature of the Hawaiian mantle plume.

## 2. Data and Methods

### 2.1. Data Acquisition

Seismic data are obtained from the publicly available IRIS (Incorporated Research Institutions for Seismology) data centre for stations across the Hawaiian Islands. Recordings are selected for stations located between  $15^{\circ}/25^{\circ}$  latitude and  $-165^{\circ}/-150^{\circ}$  longitude during the time period of 1990-2017. Data is collected for events with magnitude (Mw) 5.5-8, at epicentral distances between  $30^{\circ}$ - $90^{\circ}$ . This results in over 100,000 recordings from 77 stations across eight networks (Figure 1).

Ocean Bottom Seismometers (OBSs) were deployed around the Hawaiian Islands in the PLUME experiment from 2007-2009 (Laske et al., 2009). However, we find that the data from these stations is excessively noisy. While we do observe significant, but weak, arrivals for the 410 and 660 in the OBS

136 data, including this data decreases our ability to detect generally weaker up-  
 137 per mantle signals, and interpret the observed amplitudes with confidence.  
 138 Audet (2016) describes the challenges of creating OBS teleseismic receiver  
 139 functions caused by the water column and marine sediments. For these rea-  
 140 sons OBS data is not included in this study.

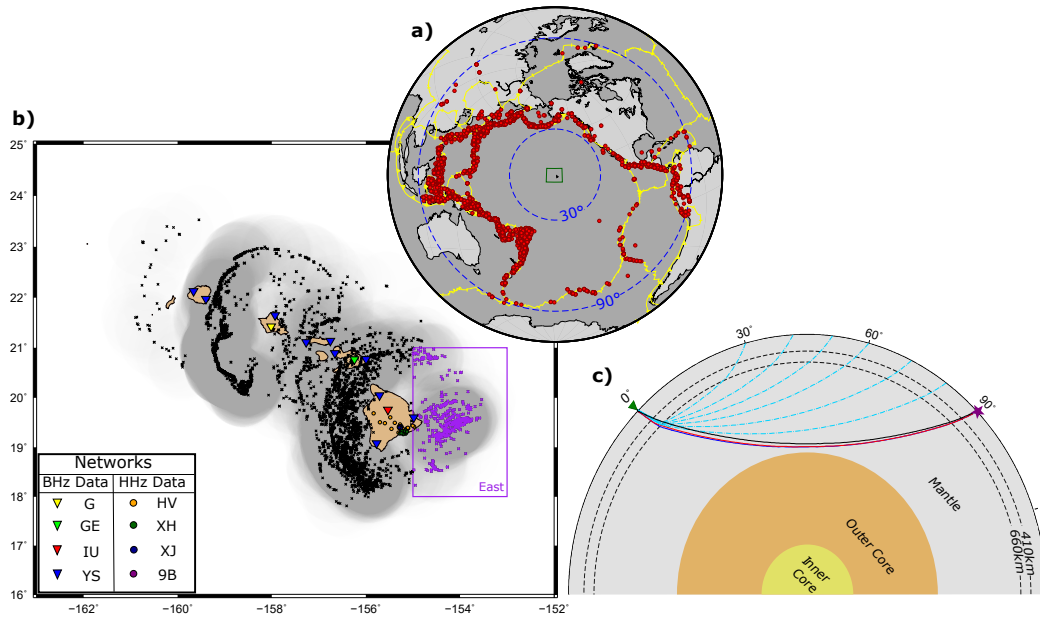


Figure 1: a) Global map showing earthquake epicentres (red circles) of the 5132 high quality RFs, epicentral distances 30° and 90° (blue dashed circles) from the centre of our study region (green square), and plate boundaries (yellow lines). b) Map of study area around the Hawaiian island chain showing seismic stations as inverted triangles (for BHZ data - sample rate 10-80 Hz) and circles (for HHZ data - sample rate 80-250 Hz), coloured by network. Pierce-points for P410s at 410 km are shown by black and purple (for region E) crosses, and half-width Fresnel zones of P410s at 410 km by 99% transparent grey circles. c) Schematic cross-section showing example ray paths of P660s phases at various distances (dashed cyan lines), and one set including the direct P (black), P410s (red) and P660s (blue) from source (pink star) to receiver (green triangle) to illustrate similarity in ray paths. (Adapted from Jenkins et al. (2016) Figures 3 and 4)

## 141 2.2. Receiver Functions and Quality Control

142 When P waves interact with seismic discontinuities, some of the energy  
143 can be converted into S waves, producing P-to-s or Pds phases (where d is the  
144 depth of the discontinuity in kilometres, e.g.: P410s and P660s). To observe  
145 converted phases, which have a relatively low amplitude, data from many  
146 events need to be stacked. To do this the source component is deconvolved  
147 from the data, creating so-called receiver functions (RFs). These are cre-  
148 ated by removing the source-time function, instrument response and source  
149 side effects from the converted phases, leaving a direct representation of the  
150 Earth's structure beneath the receiver along the incoming ray path. The  
151 vertical component (Z) of ground motion preferentially records the direct P  
152 arrival, which we assume is a good representation of the source signal, and  
153 the horizontal radial component (R) preferentially records the Pds converted  
154 waves. Initially we cut a time window of 25-seconds before, to 150-seconds  
155 after the main P arrival on the Z and R components. The Z component is  
156 then deconvolved from the R component using the Iterative Time Domain  
157 Deconvolution Method (Ligorra and Ammon, 1999) (Figure 2a).

158 Iterative Deconvolution uses Gaussian pulses to construct RFs in the time  
159 domain. Starting with an empty RF trace, we iterate between evaluating the  
160 misfit between the convolved RF and vertical component, and the radial  
161 component, adding a Gaussian peak where the misfit is largest (scaled by  
162 the misfit amplitude). The iteration stops when the misfit improves less  
163 than 0.01% or when 200 peaks are added. Here we construct two sets of RFs

164 using a ‘wide’ Gaussian half-width of 2.5-s and a ‘narrow’ Gaussian half-width  
 165 of 1.0-s. We refer to the two cases as ‘low-frequency’ and ‘high-frequency’  
 166 RFs. For both cases the data are pre-filtered using high-pass filter of 0.01 Hz  
 167 and a Gaussian filter related to the Gaussian peak width, respectively. The  
 168 vertical resolution of the low- and high-frequency RFs are around 23-km and  
 169 9-km, respectively, around 300-km depth, illustrating the importance of the  
 170 high-frequency RFs to distinguish nearby multiples and split arrivals (even  
 171 though amplitudes are weaker for high-frequency RFs).

172 Automatic and manual quality checks are applied to the RFs, removing  
 173 over 90% of the traces. Details are given in Supplementary Section 1.1.

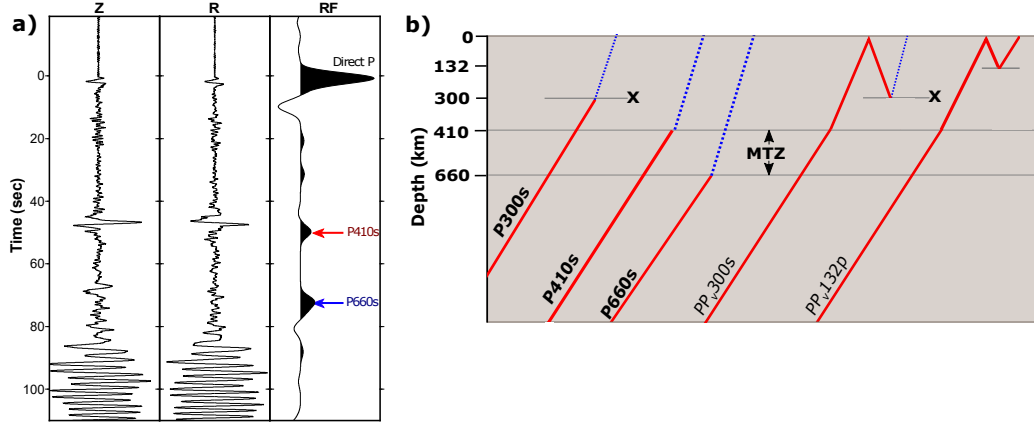


Figure 2: a) Examples of vertical (Z) and radial (R) components of ground motion for an event recorded at IRIS station KIP in Hawaii and the resulting RF obtained by deconvolving the vertical from the radial components (a ‘low-frequency’ RF). Direct P (black), P410s (red) and P660s (blue) phase arrivals indicated on the RF. b) Ray paths of P-to-s converted phases this study focuses on (labelled in bold in the form Pds where d is the depth of conversion) and surface multiples (labelled in italic in the form PPvds or PPvdp, where v denotes a topside reflection). P waves - solid red line, S waves - dotted blue line. (Adapted from Jenkins et al. (2017) Figure 3)

### 174 2.3. *Stacking Methods and Time to Depth Conversion*

175 The amplitude of the coherent Pds converted phases are small compared  
176 to the incoherent noise. Stacking the RFs (using a variety of methods) in-  
177 creases the signal-to-noise ratio, enhancing the Pds arrivals.

178 Some stacking methods are applied in the depth domain, requiring RFs  
179 to first be converted from time to depth. The RFs are converted using  
180 a combination of the crustal model, Crust1.0 (Laske et al., 2013) and the  
181 regional tomographic model of Cheng et al. (2015), which gives regional  
182 relative shear wave velocities with no fixed 1D reference model. The relative  
183 velocities are given to an unknown mean absolute velocity, which in the  
184 Hawaiian case is probably slower than the global mean, causing spurious  
185 fast anomalies around the plume in the Cheng et al. (2015) tomographic  
186 model (Bastow, 2012). To subdue this issue, we shift to relative velocities  
187 0.5% slower before converting to absolute velocities, using the 1D PREM  
188 (Dziewonski and Anderson, 1981). Using these conversions puts the 660  
189 discontinuity at approximately global averaged depths. The relative P wave  
190 velocities are scaled down by a factor 2 at the top, linear increasing to 2.35  
191 at 1000 km. For each depth, the predicted Pds - P differential time and  
192 conversion point for the Pds ray path are computed by back-tracing from the  
193 station towards the event. We account for the 3D velocities in the station-  
194 event plane, but use a 1D predicted incident angle at the station to start  
195 tracing the ray.

196 We show results for three different stacking techniques. ‘Depth’ and ‘com-

mon conversion point (CCP)’ stacks are used to identify the depth and location of conversions, while ‘slowness’ stacks are used to identify if arrivals are converted phases from depth or are surface multiples.

- Depth stacking (Figures 3a,c and 6a,c): RFs are averaged together in the depth domain for the entire study region as well as a subset of data sampling a region to the East of the Big Island (‘region E’, defined in Figure 1b). Each peak represents an arrival which could be a Pds converted phase, or a multiple. If the peak is found to be a converted phase, then the depth of the maxima represents the conversion depth and hence the discontinuity depth.
- Common Conversion Point (CCP) stacking (Figure 5): A 3-dimensional volume beneath Hawaii is discretized every 0.2-degrees in latitude and longitude and 2-km in depth. For each grid point the horizontal distance to each RF’s predicted Pds conversion pierce point within the 3D model is computed. The amplitude of the RF at that point is added to the grid point multiplied by a weighting factor dependent on the ratio of the distance and the Fresnel zone half-width for a 10-s S wave at the given depth (i.e. 116-km for P410s phase at 410-km and 162-km for P660s phase at 660-km). The weighting factor introduces smoothing by reducing to zero at twice the Fresnel zone half-width along a normalized cubic spline (see details in Cottaar and Deuss, 2016). We also track the standard error at each grid point using the difference

of the RF amplitude with the running average. We show amplitudes above twice the standard error in plotted cross-sections. Discontinuity topography maps (Figure 4) are extracted by picking and interpolating the maximum amplitude peaks found within a specified depth range.

- Slowness stacking (Figures 3b,d and 6b,d): Energy from surface multiples can interfere with conversions as they arrive at similar times (possible interfering phases are shown in Figure 2b). Slowness stacks are used to distinguish between them, as conversions come in with negative slowness relative to the direct P wave (equivalent to a steeper incoming angle), while surface multiples come in with positive slowness relative to the direct P wave (or shallower incoming angle). These stacks are created by shifting all the RFs in time to a common epicentral distance of  $60^\circ$  using relative slowness values between 1 and -1 compared to the direct P wave slowness, and then stacking the shifted RFs for each of these slowness values. A ‘bullseye’ pattern shows positive and negative coherent amplitude arrivals in slowness-time space (the term ‘bullseye’ is used to indicate a peak in coherent amplitude throughout this study even if it appears streaked). Predicted lines and positions in slowness-time space for predicted Pds converted phases and multiples are computed for PREM (Dziewonski and Anderson, 1981) and shown for reference. Note that at earlier times, and therefore increasingly at shallower depths, the predicted lines for phases converge, making it harder to distinguish between conversions and multiples.



## 242 2.4. *Synthetics*

243 To test interpretations of the observations, synthetic data are computed  
244 using reflectivity synthetics (CRFL, Fuchs and Müller, 1971). The processing  
245 of the synthetics largely follows the same procedure as the observations, and  
246 further details are given in Supplementary Section 1.2.

## 247 3. Results

248 We image the upper mantle and Mantle Transition Zone (MTZ) structure  
249 using 5132 high quality RFs around the Hawaiian islands. We first create  
250 depth and slowness stacks of the entire dataset to show the average depths  
251 of possible discontinuities across the region. The full dataset low-frequency  
252 depth stack (Figure 3a) shows four clear peaks above error at depths of 167,  
253 289, 434 and 656 km. We use the slowness stack (Figure 3b) to discern  
254 whether these peaks are true converted phases from depth or multiples. The  
255 bullseyes for the 434 km and 656 km arrive at correct slownesses to be depth  
256 converted phases. Both stacks confirm that on average the 410 is deeper (at  
257 an average of 434 km) while the 660 is only slightly shallower than expected.  
258 The MTZ thickness is on average 222 km, significantly thinner than the  
259 global average of 242.0-250.8 km (Lawrence and Shearer, 2006; Andrews and  
260 Deuss, 2008). Discussion on the lack of an observation around 520 km can  
261 be found in Supplementary Section 2.1.

262 For arrivals in the upper mantle it is more difficult to distinguish the  
263 slownesses of direct arrivals and multiples. The bullseyes in the slowness

264 stack at lower frequency (Figure 3b) corresponding to the arrivals at 167 and  
 265 289 km are of too limited resolution in the slowness domain to unequivocally  
 266 say they are arrivals from depth. Analysing the higher frequency stacks  
 267 (Figure 3c and d) helps to further distinguish between conversions and mul-  
 268 tiples. In Figure 3d, there are two clear bullseyes on the slowness stack that  
 269 correspond to peaks around 300 km in the depth stack: one strong arrival  
 270 centred around the predicted multiple lines, and one weaker arrival on the  
 271 direct conversion line. This indicates that a P300s likely interferes with a  
 272 multiple generated from shallower structure, which could be the PPv132p,  
 273 PPv84s or PSv68s. PPv132p is the closest predicted phase in both time and  
 274 slowness to observations and could result from the positive velocity jump  
 275 seen around 110-155 km beneath Hawaii in previous studies (Rychert et al.,  
 276 2013). Further evidence for the presence of a discontinuity around 300 km is a  
 277 bullseye (relative amplitude of 1.4%) arriving 90-100 seconds in the slowness  
 278 stack. This is very close to the predicted position for a PPv300s multiple,  
 279 providing further evidence that there is indeed a discontinuity around 300  
 280 km. Hereafter we will refer to this feature as the X-discontinuity (Schmerr,  
 281 2015) and the related phases PXs and PPvXs. The X-discontinuity has a  
 282 variable appearance across the region, and is difficult to observe due to in-  
 283 coming multiples from shallower structure, thus we could not produce a clear  
 284 map of X-discontinuity topography.

285 The bullseye for the arrival at 167 km is too shallow to distinguish between  
 286 the conversion and multiple line and there are no clear multiples coming from

287 this depth. For these reasons it is not investigated further in this study.

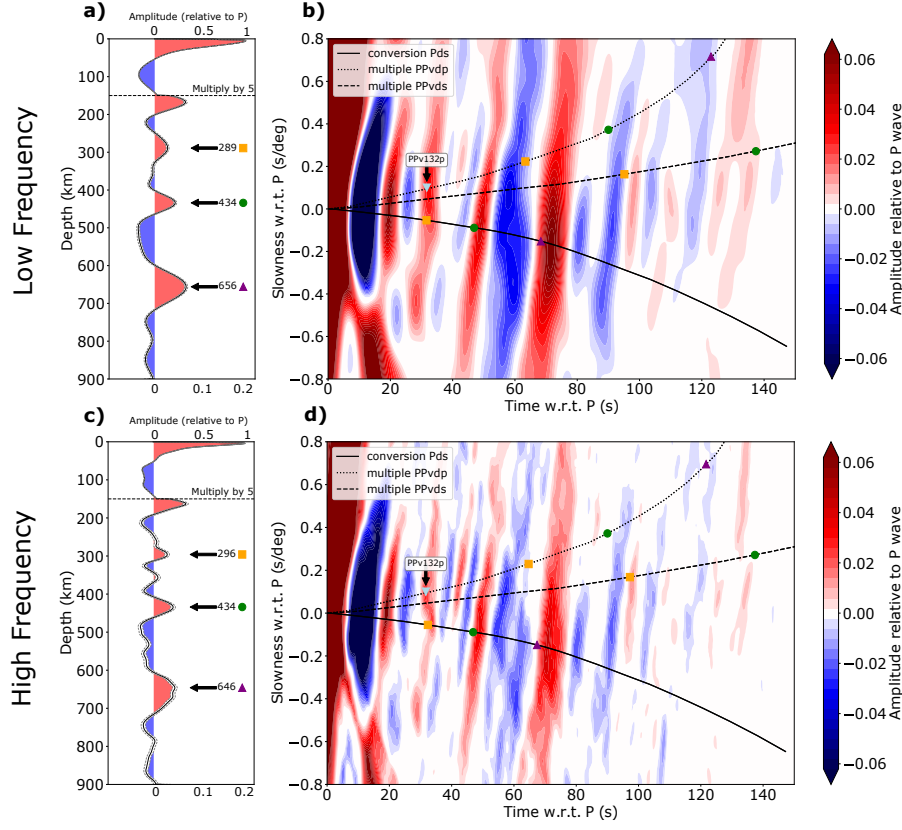


Figure 3: Depth (a and c) and slowness (b and d) stacks of all 5132 RFs used in this study at ‘low-frequency’ in a and b and ‘high-frequency’ in c and d. Depth stacks (a and c): The average amplitude at depth (solid black line) is plotted along with the lines reflecting 2 Standard Error (dashed black line). Arrows indicate the significant positive peaks, with the depth in kilometres and individual symbols. Beneath 150 km the stack is multiplied by 5 to bring out the lower amplitude peaks. Slowness stacks (b and d): Relative amplitudes ( $>2$  SE) shown as a function of time and slowness. Predicted lines for the conversion and multiple phases in slowness/time space using PREM are shown as: Pds (direct conversion) - solid line, PPvds (multiple) - dashed line, PPvdp (multiple) - dotted line. The symbols indicate predicted arrivals for the corresponding peaks in the depth stacks : 289/296 km - orange square, 434/434 km - green circle, 656/646 km - purple triangle for the low/high-frequency stacks. The predicted PPv132s phase is indicated with an light blue upturned triangle; this is the predicted phase that interferes with PXs. Note that the depth stacks use a 3D model to convert from time to depth, while the predictions for the slowness stack use 1D PREM.

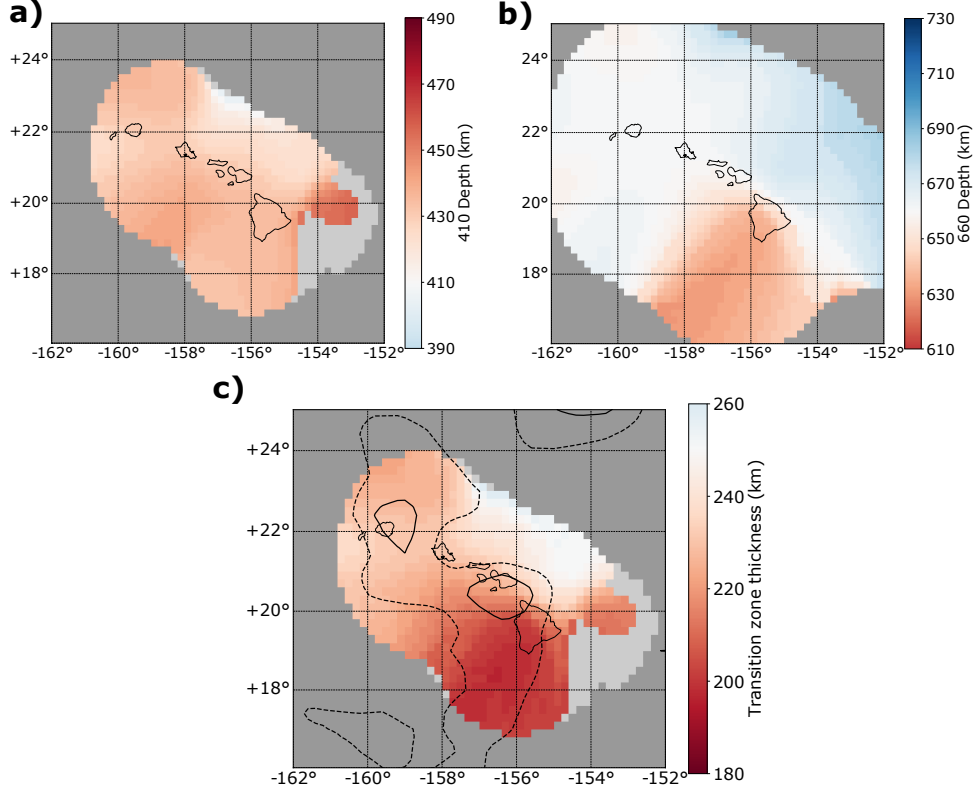


Figure 4: a) Map of topography of significant 410 arrivals from the low-frequency CCP stack. 410 km is defined as white and shallower regions are blue, deeper red. b) Map of topography of significant 660 arrivals. 660 km is defined as white and shallower regions are red, deeper blue (opposite from 410 to reflect the opposite Clapeyron slope commonly associated with 660). c) Map of MTZ thickness, i.e. the difference between a and b. 250 km is defined as white, thinner is red and thicker is blue. Contours of a slice at 410 km through the Cheng et al. (2015) regional tomographic model shown as solid (-2%) and dashed (-1%) black lines. In all three plots dark grey represents no data and light grey represents points where no significant positive arrival is observed within the depth range.

### 288 3.1. MTZ thinning southwest of the Big Island

289 Depths for significant peaks around 410 km (Figure 4a) and 660 km (Fig-  
 290 ure 4b) are extracted from our regional CCP stack and the difference is plot-  
 291 ted as a map of MTZ thickness (Figure 4c). The 410 appears deep across

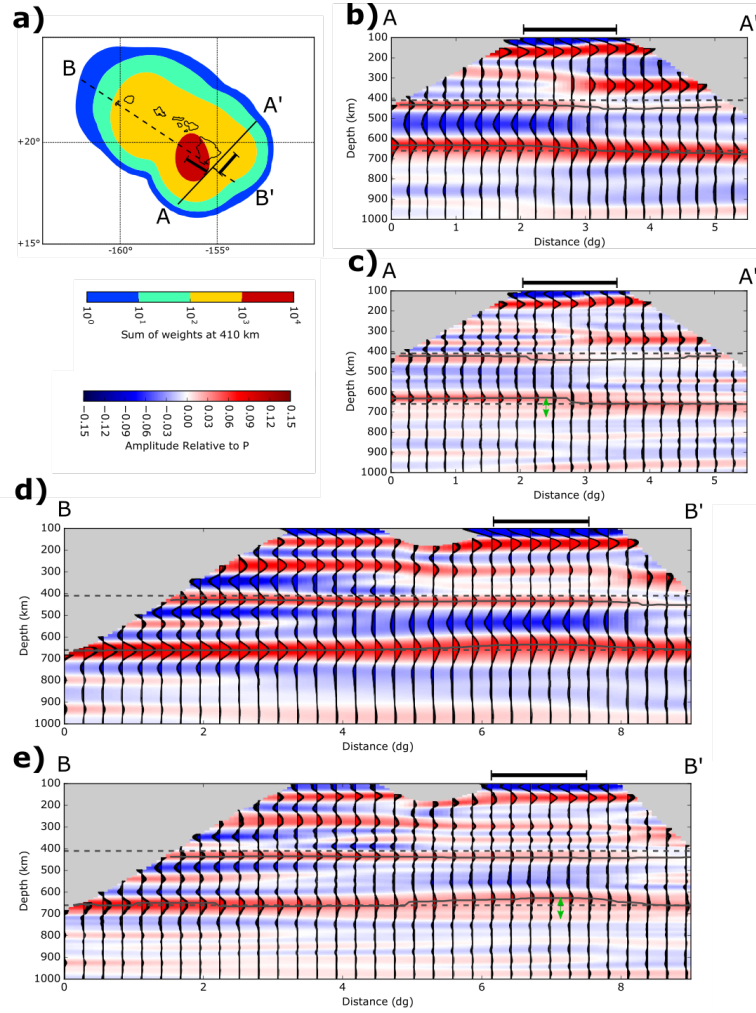


Figure 5: Cross-sections of CCP stacks. Grey is regions where the sum of the weights is less than 50. The two cross-sections run from SW-NE (A-A' ) b-c and from NW-SE (B-B' ) d-e. a) The map of Hawaii shows the summed weights in the CCP stack at 410 km depth and cross-section lines (A-A' solid and B-B' dashed). The background and regularly spaced profiles are interpolated from the CCP grid and show red for positive and blue for negative peaks ( $>2$  SE). The grey dashed lines mark out 410 and 660 km depths and solid grey lines track the observed peaks around 410 and 660 km. b and d are stacks for low-frequency RFs, c and e are stacks for high-frequency RFs. A black bar above each cross-section indicates the position of the Big Island. Green double-headed arrows (in c and e) indicate the peak-splitting at  $\sim 660$  km.

the entire area. The MTZ thickness map shows the thinnest MTZ of 200 km occurs to the southwest of the Big Island,  $\sim 50$  km thinner than the global average (Lawrence and Shearer, 2006), mainly due to uplifting of the 660 to  $\sim 630$  km.

Figure 5 shows cross-sections through CCP stacks from southwest to northeast of the Big Island (A-A') and northwest to southeast of the Big Island (B-B'). B-B' (Figure 5d,e) shows the transition from the more average TZ in the northwest, with a deep 410, to the anomalous TZ in the southwest, with a shallow 660. The cross-section shows that the 660 also becomes wider and more diffuse (Figure 5d). In the high-frequency cross-section (Figure 5e) this diffuse 660 splits into two distinct peaks, one that upwells and the other lower amplitude peak that slightly deepens. In the lower frequency CCP stacks, the shallower larger amplitude 660 peak controls the observed discontinuity topography in Figure 4b.

The deepening of the 410 across the area suggests it is affected by a widespread thermal anomaly above the 410, while the 660 is only locally affected. However, this image can depend on how we apply the time-to-depth conversion. We use the relative velocity model of Cheng et al. (2015) shifted slower by 0.5% and converted to absolute velocities using PREM. Compared to a CCP stack using the 1D PREM for time-to-depth conversion, the 3D model shifts the average 410 depth 10.0 km shallower and the average 660 depth 17.24 km shallower. Instead, if we were to shift the velocities even slower (suggesting the velocity anomalies associated with the mantle plume

are under-resolved or the background mean is slower than our assumption), two alternative scenarios can be created. In the first scenario, using a negative velocity shift both discontinuities would move upwards. In this case, the 660 is shallower and the 410 only slightly deeper than global averages and the thermo(-chemical) anomaly would be interpreted to affect both phase transitions across the whole imaged region. In the second scenario, if the velocity model were shifted even slower, the discontinuities become even shallower, causing an uplift of the 660 by 20-30 km and a 410 appearing at average global depth. This scenario suggests an anomaly that only affects the 660 discontinuity which could be explained by widespread ponding of hot plume material or a harzburgitic component (Yu et al., 2018) beneath the 660. Here we favour the interpretation of a widespread temperature anomaly above the MTZ affecting the depth of the 410, produced with the Cheng et al. (2015) model shifted slower by only 0.5%, as the most realistic scenario. We discuss the the potential cause of such an anomaly in Section 4.

### 3.2. *Anomalous Eastern Region*

Cross-section A-A' in Figure 5b,c shows the transition between the southwest of the Big Island and region to the East. The 660, which is anomalous to the southwest, appears at an average depth to the east, where the 410 arrival weakens in amplitude.

Figure 6 shows separate depth and slowness stacks for the region to the east of the Big Island, hereafter named region E (location shown in Figure

1). The most striking observation for region E, is a very weak arrival from the 410, which falls below the 2 standard error significance level in the high-frequency stacks (Figures 6c and d).

Region E also shows a strong amplitude X-discontinuity. We note the direct arrival of the X-discontinuity again interferes with a multiple (potentially the PPv163p phase). The X-discontinuity's multiple (PPvXs) however, appears as a more coherent arrival compared to the entire region (Figure 3b) and has an average relative amplitude of 2.0% across both filters. As well as being particularly strong in amplitude here, the X-discontinuity is also slightly deeper (336 km) than the regional average (296 km, Figure 3b). This deepening is seen as a delay to the PXs arrival (from 35 to 40 seconds) and a corresponding delay for the PPvXs multiple (from 108 to 113 seconds). An artefact of the multiple is also observed in the cross-sections around 900-1000 km (Figure 5), mirroring the topography seen on the shallower X-discontinuity arrival. We note that there appears to be a strong negative arrival before the positive arrival of the X-discontinuity in these stacks, with suggestions of a corresponding negative multiple.

In the Supplementary Section 2.2, we show how the variation in MTZ around the Big Island can be illustrated by stacking by back-azimuth of incoming events.



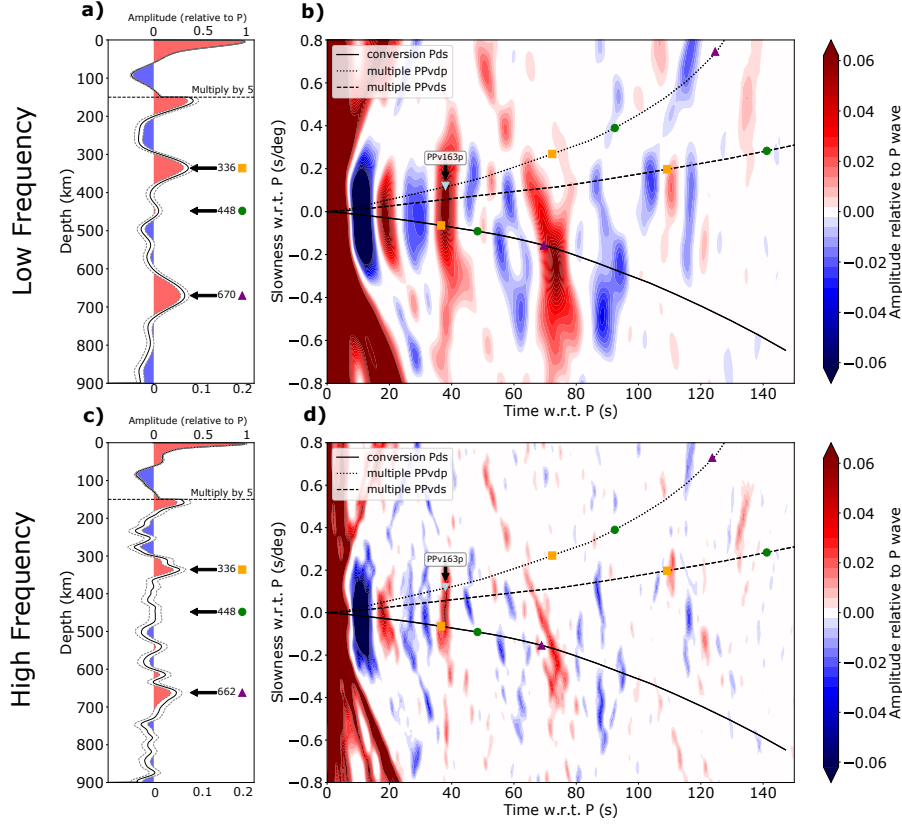


Figure 6: Depth (a and c) and slowness (b and d) stacks of low-frequency (a and b) and high-frequency (c and d) RFs for the 594 RFs to the East of the Big Island (region E, see Figure 1). See Figure 3 for explanation of the different stacks, lines and markers. PPv163p indicated as possible phase interfering with PXs.

#### 357 4. Discussion

358 This study maps converted phases from the upper mantle and MTZ be-  
 359 neath the Hawaiian Islands. Notable observations include a significantly  
 360 thinned MTZ to the SW of the Big Island combined with an observation  
 361 of a double peak on the 660 in the area of maximum thinning. The X-  
 362 discontinuity is observed throughout the region and appears particularly

363 strong towards the E of the Big Island, where the 410 conversion almost  
364 disappears. Here we interpret the appearance and form of these discontinu-  
365 ities in terms of potential thermal and compositional properties of mantle  
366 material.

#### 367 *4.1. Plume signature across the transition zone*

##### 368 *4.1.1. Discontinuity topography*

369 The 410 appears deep across the area of study by about 20 km, leading  
370 to generally thinned MTZ. To the SW of the Hawaiian Big Island, there is  
371 an area that has an even thinner MTZ than the average for Hawaii,  $\sim 50$   
372 km thinner than the global average. The additional thinning is mainly due  
373 to the 660 in this area shallowing to  $\sim 630$  km. We interpret this to be the  
374 position of the upwelling mantle plume across the MTZ.

375 The plume location based on the 410 topography is less clear, as its  
376 depression is quite consistent, suggesting a potential widespread thermal  
377 anomaly affects the 410. The interpretation of a widespread thermal anomaly  
378 above the 410 correlates with widespread low shear velocities at the bottom  
379 of the upper mantle in the velocity model of Cheng et al. (2015). We will  
380 discuss further in Section 4.2 how this interpretation is also supported by  
381 our observation of the X-discontinuity and the missing 410, as well as re-  
382 sults from recent geodynamic modelling. Temperature estimates based on  
383 the discontinuity topography are discussed in Supplementary Section 2.3.

384 The regional tomographic model of Cheng et al. (2015), shows a  $< -2\%$

385 velocity structure to the NW of the Hawaiian islands at around 400 km depth  
 386 (Figure 4c), further north than where our study predicts the plume position.  
 387 The location of the plume to the SW of the Big Island is consistent with  
 388 multiple other P-to-s conversion studies (Li et al., 2000; Shen et al., 2003;  
 389 Wölbern et al., 2006). However, more recent P-to-s studies come to different  
 390 conclusions. Huckfeldt et al. (2013) finds the strongest thinning towards the  
 391 southeast, while Agius et al. (2017) finds thinning of 13 km under north-  
 392 central Hawaii. The study of Huckfeldt et al. (2013) also observes a deep  
 393 410 across the region, while in Agius et al. (2017) both the 410 and 660 are  
 394 at average depths. There are clearly large discrepancies in predicted plume  
 395 location between studies. However we are confident in our interpretation  
 396 that the plume stem is located SW of the Big Island, due to the additional  
 397 observation of a split 660 in this region.

#### 398 *4.1.2. A double peak at 660 km*

399 The splitting of the 660 in the SW region offers possible insights into  
 400 both thermal and compositional heterogeneities in the mantle. In the high-  
 401 frequency cross-section of the CCP stack (Fig 5e), the SW region shows two  
 402 peaks at around 660 km: one that appears shallower (by  $\sim 30$  km) and one  
 403 that appears deeper (by  $\sim 50$  km). The upper peak is likely to correspond  
 404 to the dissociation of ringwoodite to bridgmanite and magnesiowstite. The  
 405 deeper peak appears to have the opposite Clapeyron slope, becoming deeper  
 406 as the upper peak shallows. We interpret this to represent a discontinuity

407 caused by the phase transition of majorite garnet to bridgmanite that also  
408 is predicted to occur around this depth at relatively higher temperatures,  
409 and has a positive Clapeyron slope (Liu et al., 2018). If this is the case the  
410 location of the split at 660 represents the location of highest temperature  
411 anomaly, further supporting our interpretation of the plume being located to  
412 the SW of the Big Island.

413 The garnet transition is generally predicted to be more gradational with  
414 depth than the ringwoodite transition as majorite garnet can co-exist with  
415 bridgmanite over a large range of pressures and temperatures (e.g. Yu et al.,  
416 2011). Additionally, compositional effects from inclusion of mafic compo-  
417 nents, such as recycled basalt, would broaden the majorite stability field (Xu  
418 et al., 2008). This could explain the smaller amplitudes of the deeper of the  
419 two peaks we observe, since broader discontinuities produce lower amplitude  
420 converted arrivals.

421 The presence of both phase transitions occurring together is predicted  
422 to happen over a very specific temperature range, approximately 200–300 K  
423 above global average (Hirose, 2002; Stixrude and Lithgow-Bertelloni, 2011).  
424 The garnet-controlled phase transition has been suggested to dominate obser-  
425 vations of depressed 660 topography beneath Iceland (Jenkins et al., 2016),  
426 where the olivine-controlled phase transition is not observed. This could in-  
427 dicate that Icelandic plume stem is hotter and/or carries more garnet than  
428 the Hawaiian plume.

## 429 4.2. *Heterogeneous signals in the upper mantle: X-discontinuity*

### 430 4.2.1. *Comparison to previous observations*

431 Various studies have reported the presence of the X-discontinuity be-  
432 neath the Pacific and specifically Hawaii using ScS reverberations (Courtier  
433 et al., 2007) and SS precursors (Deuss and Woodhouse, 2002; Schmerr et al.,  
434 2013; Schmerr, 2015). Schmerr (2015) observes the X-discontinuity across  
435 the Pacific at  $293 \pm 65$  km which is consistent with our observations. SS  
436 precursor bounce points have much broader coverage across the Pacific, but  
437 also average over an order of magnitude wider Fresnel zone compared to RFs  
438 (1000s km for precursors versus 100s km for RFs). Schmerr (2015) observes  
439 a weak ( $< 2\%$  impedance contrast) presence of the X-discontinuity beneath  
440 Hawaii; this could be due to strong topography on the discontinuity causing  
441 incoherent reflections and stacking.

### 442 4.2.2. *Proposed causes*

443 Various mineral and physical processes have been proposed to explain the  
444 X-discontinuity, but these hypotheses do not always apply to mantle plume  
445 settings:

- 446 • Formation of hydrous phase A - a dense magnesium silicate - (e.g.  
447 Akaogi and Akimoto, 1980): Stability of this phase requires relatively  
448 low temperatures and high water content conditions as found in sub-  
449 duction zone settings.

- 450 • The reaction of forsterite + periclase into Anhydrous-phase B: This  
 451 mechanism requires substantial periclase enrichment (Chen et al., 2015),  
 452 which could occur in hydrated mantle/subduction zone settings (Gan-  
 453 guly and Frost, 2006).
- 454 • A crystallographic transition in pyroxene (clinoenstatite) from orthorhom-  
 455 bic to monoclinic structure (Woodland, 1998): This transition has a  
 456 strong positive Clapeyron slope and a weak impedance contrast ( $< 2\%$ ),  
 457 which further weakens at higher temperatures (e.g. Xu et al., 2008;  
 458 Schmerr, 2015), making its visibility unlikely.
- 459 • A phase transition in silica from coesite to stishovite with a positive  
 460 Clapeyron slope (e.g. Akaogi et al., 1995): These silica phases are ex-  
 461 pected to be present in mafic material with basaltic bulk compositions,  
 462 potentially brought up in mantle plumes.

463 From here on we will explore the potential for the coesite-stishovite phase  
 464 transition to explain our observations. This model can account for precur-  
 465 sor observations of the X-discontinuity across the broader Pacific (Schmerr,  
 466 2015), and is consistent with a high-temperature plume setting. The pres-  
 467 ence of this transition is easier to invoke in regions where basalt is subducted  
 468 (Williams and Revenaugh, 2005), but recycled oceanic basalt has been sug-  
 469 gested to be present in the Hawaiian plume (Hofmann and White, 1982;  
 470 Sobolev et al., 2005; Herzberg, 2010).

#### 471 4.2.3. *Coesite-stishovite transition*

472 Average mantle is thought to have a pyrolitic composition (McDonough  
473 and Sun, 1995), which contains a large modal proportion of olivine at low  
474 pressure, leading to the generation of the globally observed olivine-wadsleyite  
475 and ringwoodite-bridgmanite+magnesiowstite phase transitions at 410 and  
476 660 km depth respectively (Katsura and Ito, 1989; Ito and Takahashi, 1989).  
477 No silica phases are present in a pyrolitic composition, and therefore we do  
478 not expect a globally observed seismic discontinuity at the predicted depth  
479 of 300 km. However, the presence of unequilibrated mafic material, in a me-  
480 chanical mixture of different compositions, allows for the presence of silica  
481 phases (e.g. Xu et al., 2008). A possible source of such compositional het-  
482 erogeneity is from the presence of recycled basaltic material in the plume  
483 source. Recycled basalt compositions are expected to be stable as an eclog-  
484 ite containing pyroxene, garnet and a free silica phase in the P-T conditions  
485 of the upper mantle under Hawaii (Jennings and Holland, 2015). Composi-  
486 tional characteristics of Hawaiian basalts have been linked to the presence of  
487 recycled material in their mantle source regions (e.g. Hauri, 1996; Eiler et al.,  
488 1996; Frey et al., 2016). The major element compositions of the Loa-trend of  
489 volcanoes provide some of the strongest evidence for the presence of recycled  
490 basalt (e.g. Sobolev et al., 2005; Herzberg, 2010).

491 The strong impedance contrast of the co-st transition indicates that only  
492 a small % of free-silica is required to explain X-discontinuity observations  
493 (Chen et al., 2017). However the potential for a reduction of free-silica in re-

494 cycled basalts after dehydration and alteration processes during subduction,  
495 has called into question whether enough silica would be present to produce  
496 X-discontinuity observations (Knapp et al., 2015). In the context of Hawaiian  
497 magmatism, however, it is important to note that geochemical studies have  
498 concluded that mafic lithologies that have not lost substantial SiO<sub>2</sub> during  
499 subduction processes are present in the mantle source regions (Jackson et al.,  
500 2012).

501 Ballmer et al. (2013, 2015) explore the dynamical effects of a plume en-  
502 riched by dense eclogitic compositions. Their study suggests that such ma-  
503 terial may only be transported to the upper mantle in the central and there-  
504 fore hottest part of the plume. In the region between the coesite-stishovite  
505 transition and the olivine-wadsleyite transition (300-410 km), the eclogitic  
506 component is negatively buoyant, which causes ponding in this depth range,  
507 forming a so-called Deep Eclogitic Pool (DEP). When the eclogitic material  
508 crosses the stishovite-to-coesite phase transition, the material becomes posi-  
509 tively buoyant again. The presence of hot material ponding in a DEP could  
510 explain the broad low velocity anomalies around these depths in the tomo-  
511 graphic model of Cheng et al. (2015). The numerical study by Dannberg  
512 and Sobolev (2015) also finds that mantle plumes containing up to 15-20%  
513 recycled oceanic crust as eclogite cause broad-scale ponding in the upper  
514 mantle.



#### 515 4.2.4. *Synthetic exploration*

516 We apply a simplified synthetic test to explore if a coesite-stishovite phase  
517 transition in a DEP can explain the observations seen here, specifically those  
518 in the region E where we see a strong X-discontinuity and disappearance  
519 of the 410. We compute impedance contrasts for the coesite-stishovite and  
520 olivine-wadsleyite phases in different fractions using BurnMan - a Python li-  
521 brary used to calculate thermo-elastic properties of mantle minerals (Cottaar  
522 et al., 2014) - with the database of Stixrude and Lithgow-Bertelloni (2011)  
523 (which does not currently account for the possibility of silica reduction in  
524 basalt by dehydration processes (Knapp et al., 2015)). We create synthetic  
525 models by modifying the PREM velocity model (Dziewonski and Anderson,  
526 1981) to accommodate the computed velocities and density jumps at their  
527 observed depths beneath Hawaii, while removing the original 220 and 410  
528 discontinuities in PREM. With increasing basalt fraction, the impedance  
529 contrast for the X-discontinuity increases, while that for the 410 diminishes  
530 (Figure 7a). This is reflected in the synthetic RF depth stacks (Figure 7b and  
531 c) by a change in relative RF amplitudes for the different conversions. The  
532 stacks use the same distance distribution as the stacks for region E (Figure  
533 6), but for each distance are stacked over different event depths (see Section  
534 2.4). In the synthetics we see an increase in the amplitudes of the arrivals for  
535 higher frequencies, which is not reflected in the real data (Figure 6). This  
536 could be due to less coherent stacking of high-frequency arrivals in the real  
537 data, or the phase transitions occurring over a broader depth than has been

538 modelled here. In general, given more incoherent stacking and noise in the  
539 data, around 60-70% basalt accumulation can explain the disappearing 410.

540 Both for the real data and synthetics it is easier to compare the amplitude  
541 of the multiple phase (PPvXs) rather than the direct phase (PXs) in slowness  
542 stacks as there are fewer interfering phases  $\sim 100$  seconds after the P wave.  
543 Synthetic slowness stacks are shown for 20% and 50% basalt for both filters  
544 in Figure 8. The observed relative amplitude of PPvXs is 2.0% for region  
545 E (Figure 6). In the synthetics, such amplitudes are reached when 40-50%  
546 basalt is included. Thus less basalt accumulation (40-50%) can explain the  
547 observations from the X-discontinuity at the top of the DEP, while stronger  
548 accumulation of basalt (60-70%) at the bottom of the DEP in region E is  
549 needed to explain the disappearing 410.

550 It should be stressed that the basalt component of 40-70% required across  
551 the DEP to explain both the X-discontinuity and the 410 cannot be carried  
552 up by a plume. The plume could carry a basaltic component of up to 20%  
553 (Ballmer et al., 2013, 2015; Dannberg and Sobolev, 2015) which would have  
554 to accumulate within the DEP to create higher percentages. Dynamical  
555 models that allow for segregation and accumulation of components have not  
556 been tested to our knowledge. Additionally, dynamical models would have  
557 to test if the DEP can expand laterally and to shallower depths to allow the  
558 coesite-stishovite to be visible over a broad area.

559 We note that the arrivals from the X-discontinuity in the synthetic slow-  
560 ness stacks do not capture the negative swing before these arrivals observed

in data (e.g. around 30 seconds in Figure 6b). Creating synthetics with a negative velocity jump (i.e. the top of a lower velocity zone as invoked to explain similar observations in Huckfeldt et al. (2013)), did not recreate a strong amplitude multiple. We note that subtle changes in velocity model (i.e. a change in gradient) or broader discontinuities, can change the shape of the phase arrival in receiver functions. Exploring this space of velocity models is beyond the scope of this study.

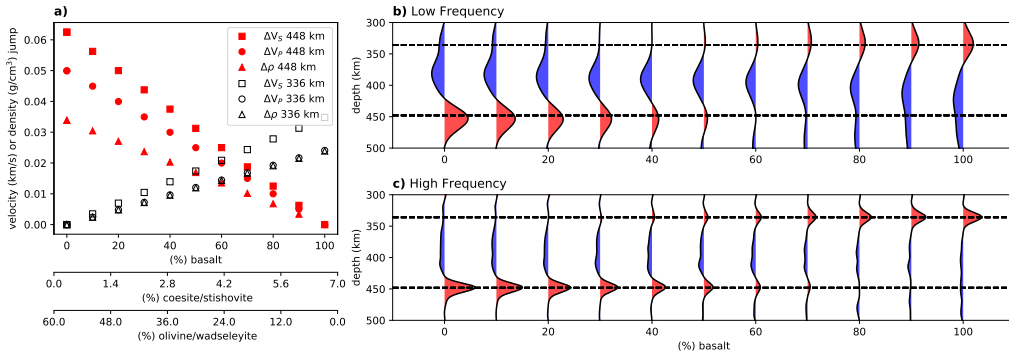


Figure 7: a) Predicted velocity and density jumps for a mechanical mixture of harzburgite and basalt. b) and c) Depths stacks for low-frequency (b) and high-frequency (c) synthetic receiver functions using predicted velocity and density jumps fixed at 336 km for the X-discontinuity and a gradual discontinuity from 443-453 km for the 410 for different basalt contents.

#### 4.3. Summary of the plume across the upper mantle

We suggest the plume stem crosses the MTZ to the southwest of the Big Island (see cartoon in Figure 9), where its hot temperatures (200–300 K) thin the MTZ and lead to splitting of the 660 due to the presence of both an olivine and a majorite garnet transition. The plume carries recycled basaltic material which may act to enhance the garnet transition. As the

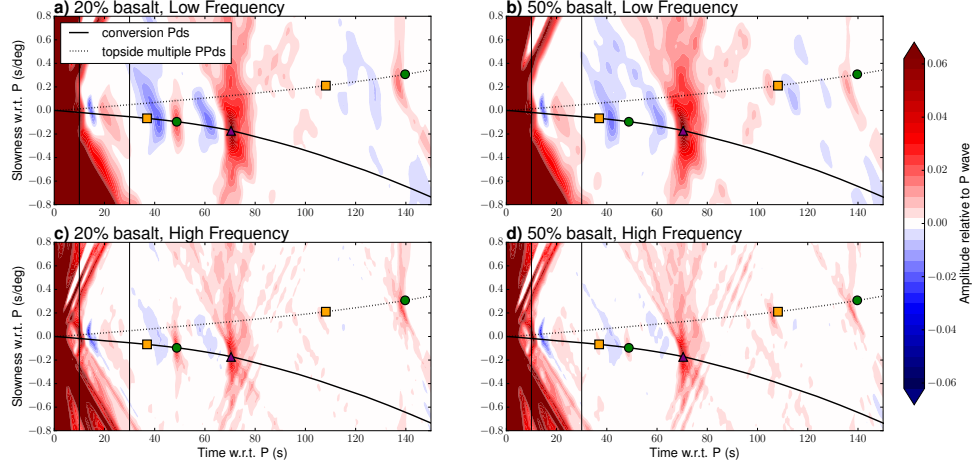


Figure 8: Slowness stacks ( $>2$  SE) for synthetic receiver functions for a model with 20% basalt (a and c) and 50% basalt (b and d) at the Low (a and b) and High (c and d) frequency. Predicted lines for the conversion (solid) and multiple phases (dotted) in slowness/time space using PREM are shown as: Pds (direct conversion) - solid line, PPvds (multiple) - dashed line. Markers are shown for arrival times of Pds and PPvds phases from depths of 336 km (orange square), 448 km (green circle) and 670 km (purple triangle).

574 plume material crosses the 410 phase transition, it becomes less buoyant and  
 575 starts to pond and spread out, creating a Deep Eclogitic Pool from 300-410  
 576 km (Ballmer et al., 2013). Spreading of the hot material in the DEP causes  
 577 the 410 to appear depressed over a wide region and correlates with wide-  
 578 spread slow velocities in the model of Cheng et al. (2015). The lack of an  
 579 olivine phase transition at 410 km to the east of the Big Island could result  
 580 from strong accumulation of basaltic material at the bottom of the DEP and  
 581 sinking of the material into the transition zone. The presence of widespread  
 582 basaltic material in the upper mantle is supported by the presence of the  
 583 X-discontinuity, which can be related to the coesite-to-stishovite transition.

584 *4.3.1. Connection to geochemical trends*

585 The geochemistry of Hawaiian basalts implies spatial variation in the pro-  
586 portion of non-peridotitic material in their mantle source regions, showing  
587 an enhanced contribution of recycled material in the Loa chain in the SW as  
588 opposed to the Kea chain in the NE (e.g. Sobolev et al., 2005; Frey et al.,  
589 2016; Herzberg, 2010; Weis et al., 2011). A straightforward explanation for  
590 the distribution of these chains is that the mantle under the Loa-chain volca-  
591 noes contains a greater proportion of recycled mafic material, which may be  
592 present as eclogite at depth and react with surrounding peridotite to form py-  
593 roxene beneath the SW Loa volcanoes. Our mapping of the X-discontinuity  
594 does not have the resolution to map variation in eclogite within the plume  
595 stem.

596 We do observe that the plume stem across the MTZ lies towards the  
597 southwest of the Big Island on the Loa side of the chain. While the plume  
598 is offset to directly beneath the Big Island across the DEP, the plume flux  
599 is likely higher closer to its source across the transition zone and could thus  
600 entrain more eclogitic material on the Loa-side as is shown in asymmetrical  
601 plume models (e.g. Ballmer et al., 2015). Seismic studies of the lithosphere  
602 find slower velocities (Laske et al., 2011) and deeper onset of melting (Rychert  
603 et al., 2013) on the Loa-side.

604 We note that in our observations, the strongest evidence of eclogite pond-  
605 ing (DEP) lies to the east, where the 410 disappears, which is on the Kea-  
606 side. However, this observation does not have to have a direct relationship to

607 the zonation of the plume-derived melts at the surface. To the east the DEP  
608 appears so enriched in eclogitic material that strong accumulations (60–70%)  
609 are ponding and sinking into the transition zone (affecting the observations  
610 of 410 arrivals), hence the negative buoyancy of the accumulated material in  
611 the DEP might not allow entrainment of this material. The dynamical mod-  
612 els of Ballmer et al. (2015) also show sinking of enriched material through  
613 the 410 away from the main plume stem.

614 It has also been suggested that the geochemical zonation is inherited by  
615 different entrained compositions from the lowermost mantle (e.g. Farnetani  
616 and Hofmann, 2010; Weis et al., 2011). Ballmer et al. (2013, 2015) shows  
617 that such zonation is not retained in the presence of a DEP.

618 Ballmer et al. (2015) argue for an alternative explanation where thermal  
619 asymmetry resulting from a model with a DEP can cause the observed trends  
620 when melting behaviour of the different lithologies is included in the geody-  
621 namical models. In these models the greatest relative contribution of fusible  
622 lithologies such as eclogite or pyroxenite is greatest in the cooler parts of the  
623 plume. In higher temperature parts of the planform the relative contribution  
624 from refractory lithologies, such as peridotite, is increasingly important. As  
625 such, the melting of lithologically heterogeneous mantle, temperature and not  
626 the amount of eclogite fed from the plume causes the variations in enrichment  
627 of basalt compositions. Hotter temperatures resulting in less enriched melt  
628 would argue for the Kea-side to be hotter, which is inconsistent with our ob-  
629 servations of the plume crossing the MTZ closer to the Loa-side. Therefore,

630 we suggest that it is entrainment processes, rather than the thermal nature  
631 of the plume, that limits the eclogite sourced on both sides of the plume to  
632 cause pyroxenite melts.

633 While the geodynamical models of Ballmer et al. (2013, 2015) and Dannberg  
634 and Sobolev (2015) show eclogite enrichment in the plume source has a great  
635 impact on plume dynamics and creation of a DEP, we do note that their  
636 models are limited by tracking two fixed compositions: enriched plume ma-  
637 terial (up to 16% eclogite) and surrounding peridotite. This model does not  
638 allow for further accumulation during the ponding of eclogite in the DEP,  
639 which is required to explain our observations. Therefore the models may not  
640 reflect the full complexities of variable entrainment of eclogite material out  
641 of the DEP, and this could be a motivation for further research.

## 642 5. Conclusion

643 We use 5132 high quality RFs to detect P-to-s conversions and associated  
644 multiples in order to image seismic discontinuities in the mid-to-upper mantle  
645 beneath the Hawaiian Islands. The RFs are stacked in a variety of ways  
646 to increase the signal-to-noise ratio, including depth stacks to define the  
647 depth of possible discontinuities, slowness stacks to distinguish between true  
648 conversions and multiples, and CCP stacks to investigate lateral variability.

649 We find lateral variations on three distinct discontinuities:

- 650 • Across the region we find the presence of an X-discontinuity around  
651 290-350 km depth. While the direct arrival of this discontinuity (PXs)

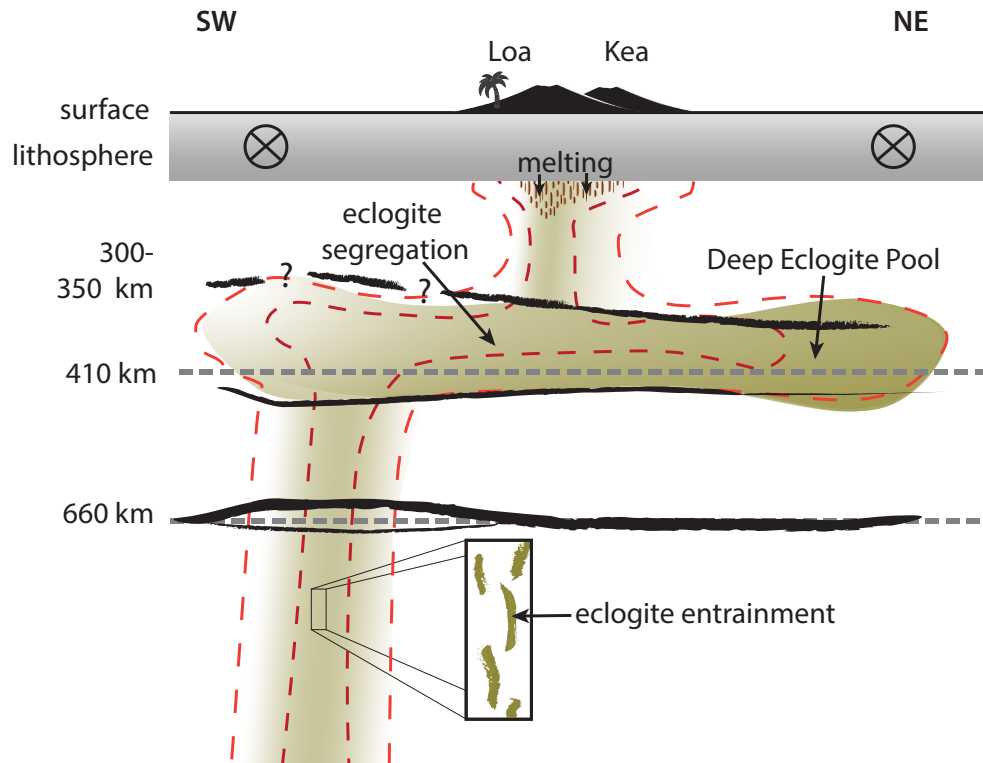


Figure 9: Summary cartoon. The hot mantle plume material (shown in a green-brown) upwells to 660 km in the SW where the 660 discontinuity appears split due to increased temperatures and basalt content. The 410 is depressed throughout the region below 410 km. There are patches of the X-discontinuity between 300-350 km throughout the region. The cartoon shows the hypothesised Deep Eclogite Pool which spreads above the 410 and could explain the presence of the X-discontinuity and the disappearance of the 410. The hot plume continues from the DEP up to the surface to cause the hotspot volcanism that created Hawaii.

interferes with a strong multiple from shallower depths, corroborative evidence of its presence comes from the observation of a multiple from this discontinuity (PPvXs).

- The 410 is depressed throughout the region by  $\sim 20$  km. Additionally,



656 the conversion from the 410 almost disappears to the east of the Big  
657 Island in conjunction with the strongest amplitude observations of the  
658 X-discontinuity.

- 659 • The 660 appears around 660 km depth across much of the region, except  
660 the area to the southwest of the Big Island. Here the 660 is split into  
661 a stronger arrival around 630 km and a weaker arrival around 700 km.

662 We hypothesise that southwest of the Big Island is the location where the  
663 hot, upwelling mantle plume crosses the 660. The high temperatures cause  
664 the dissociation reaction of ringwoodite to occur at shallower depths (thin-  
665 ning the MTZ), and garnet to be stable, causing a deeper garnet-controlled  
666 peak. More garnet can also be present due to an eclogitic component carried  
667 up in the plume.

668 In the upper mantle, we hypothesise ponding and accumulation of an  
669 eclogitic component. This would cause widespread hot temperatures, which  
670 would deepen the 410 across a wide area. Strong accumulation of eclogite  
671 on top of the 410 can also cause the observed disappearance of the 410 to  
672 the East of the Big Island. The stishovite component present in eclogite  
673 undergoes a conversion to coesite explaining the observed X-discontinuity  
674 across the region. The variability of geochemical trends observed in erupted  
675 lavas at the surface might be explained by increased entrainment of eclogite  
676 towards the SW where the plume may be hotter due to its proximity to the  
677 plume stem across the MTZ.

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