P and S wave delays caused by thermal plumes

Ross Maguire\textsuperscript{1*}, Jeroen Ritsema\textsuperscript{1}, Peter E. van Keken\textsuperscript{1}, Andreas Fichtner\textsuperscript{2}, and Saskia Goes\textsuperscript{3}

\textsuperscript{1} Department of Earth and Environmental Sciences, University of Michigan, Ann Arbor, MI, USA 48109-1005
\textsuperscript{2} Institute of Geophysics, ETH Zürich, Zürich, Switzerland
\textsuperscript{3} Department of Earth Science and Engineering, Imperial College London, UK

\textbf{SUMMARY}

Many studies have sought to seismically image plumes rising from the deep mantle in order to settle the debate about their presence and role in mantle dynamics, yet the predicted seismic signature of realistic plumes remains poorly understood. By combining numerical simulations of flow, mineral-physics constraints on the relationships between thermal anomalies and wave speeds, and spectral-element method based computations of seismograms, we estimate the delay times of teleseismic S and P waves caused by thermal plumes. Wavefront healing is incomplete for seismic periods ranging from 10 s (relevant in traveltime tomography) to 40 s (relevant in waveform tomography). We estimate P wave delays to be immeasurably small ($< 0.3$ s). S wave delays are larger than 0.4 s even for S waves crossing the conduits of the thinnest thermal plumes in our geodynamic models. At longer periods ($> 20$ s), measurements of instantaneous phase misfit may be more useful in resolving narrow plume conduits. To detect S wave delays of 0.4–0.8 s and the diagnostic frequency dependence imparted by plumes, it is key to minimize the influence of the heterogeneous crust and upper mantle. We argue that seismic imaging of plumes will advance
significantly if data from wide-aperture ocean-bottom networks were available since, compared to continents, the oceanic crust and upper mantle is relatively simple.

1 INTRODUCTION

Hotspots (Wilson 1963) and mantle plumes (Morgan 1971) have been important concepts in global geophysical and geochemical research (Ballmer and van Keken 2015) for more than half a century. In the classical view, plumes begin as thermal instabilities at the core-mantle boundary (CMB) and rise rapidly through the mantle. The voluminous plume head erupts as flood basalts to form large igneous provinces. The narrow plume tail is a persistent source of volcanism with a relatively fixed mantle position.

While plumes explain broad topographic swells, geoid highs, and the distinct geochemistry of basalts at hotspots, the plume hypothesis has not been universally accepted. A number of hotspots may not require a deep mantle origin (e.g., King & Ritsema 2000), and plumes in a heterogeneous mantle with chemical and phase changes are predicted to be more complex than plumes in the classical models (e.g., Samuel & Farnetani 2003; Ballmer et al. 2013; Lin & van Keken 2006). Seismic images (e.g., Bijwaard & Spakman 1999; Montelli et al. 2004; Allen et al. 2002; Wolfe et al. 2009; Styles et al. 2011; French & Romanowicz 2015) and statistical analyses (Boschi et al. 2007, 2008) suggest that several low-velocity anomalies are continuous from the top to the bottom of the mantle. In addition, mantle transition zone thinning has been observed beneath a number of hotspots (e.g., Schmandt et al. 2012; Shen et al. 1998), potentially indicating a thermal anomaly extending to the lower mantle. Yet, it remains controversial to associate the complex seismic observations and models uniquely to thermal plumes.

Several factors complicate the resolution of the seismic structure of the mantle beneath hotspots. First, most regional seismic networks, and especially those covering oceanic hotspots (i.e., Hawaii and Iceland), have limited aperture. The sparse wave-path coverage in the lower mantle leads to overwhelming seismic modeling artifacts. Second, the deceleration of waves traversing a plume tail may not be recorded at seismic stations on the surface due to the destructive interference of direct and diffracted waves (i.e., wavefront healing) (Nolet & Dahlen 2000; Malcolm & Trampert 2011). These diffractions, recorded in the coda of P and S waves (Rickers et al. 2012), are weaker than the coda signals produced by scattering in the crust.

To make meaningful interpretations of seismic data and models, it is important to understand the expected imprint of plumes in waveforms. In this paper we estimate the delay
times of S and P waves, which are principal observations used in traveltime tomography. We develop seismic models of plumes by combining numerical simulations of flow and mineral-physics constraints on the relationships between thermal anomalies and wave speeds. We use 3D spectral-element method (SEM) computations to synthesize teleseismic S- and P wave propagation through plumes at frequencies up to 0.1 Hz, which are relevant in body-wave analyses. This work builds on analyses of uniform cylindrical anomalies (Rickers et al. 2012) and on analyses of 2.5D axisymmetric synthetics (Hwang et al. 2011).

Our seismic models of axisymmetric plumes are based on plume ascent in a compressible mantle with an isochemical pyrolitic composition and with phase changes. We vary the strength and width of plumes by varying the dynamic parameters. The plume buoyancy fluxes are about $2 \times 10^3$ kg/s in the lower mantle, which are within the range of the fluxes inferred from hotspot swell topography (Sleep 1990). It is likely that plumes in the Earth deviate substantially from our idealized numerical models due to, for example, entrainment of compositionally distinct material (e.g., Lin & van Keken 2006; Samuel & Bercovici 2006; Kumagai et al. 2008) and the shearing by the overriding plate (Ballmer and van Keken 2015). However, we focus primarily on the structure of the narrow, vertical plume conduit, whose seismic resolvability in the lower mantle is uncertain. Any deviation from an idealized vertical conduit due to large scale flow is unlikely to significantly alter the width of the plume tail, or the amplitude of the seismic anomaly.

2 NUMERICAL SIMULATIONS

2.1 Thermal plumes in a compressible mantle

We simulate plumes closely following Bossmann & van Keken (2013), in which we solve the equations governing conservation of mass, momentum, and energy as defined by the anelastic liquid approximation (Jarvis & McKenzie 1980). The equations are discretized via the finite-element method and solved in an axisymmetric spherical shell. We use a staggered grid refinement scheme to optimize the computations. The smallest grid spacing of 2.85 km is necessary to resolve the high temperature gradients in the plume head. The bottom boundary is a free-slip surface with a fixed temperature of 3270 K. The side wall is insulating, and the top boundary is fixed at 273.15 K. Rigid boundaries at the top and side of the domain limit large-scale horizontal flow and emphasize plume development.

The initial adiabatic temperature profile (Figure 1) is determined using the Adams-Williamson equation of state. Estimates of the temperature increase across the superadiabatic
thermal boundary layer above the CMB range from 500 K to 1,800 K (Lay & Buffett 2008). We vary the temperature contrast, $\Delta T_{\text{CMB}}$, across the basal thermal boundary layer between 550 and 750 K. The depth-dependent thermal expansivity, $\alpha$, decreases by a factor of three from the Earth’s surface to the CMB. A surface value of thermal expansivity, $\alpha_0 = 3 \times 10^5$ K$^{-1}$ and specific heat, $c_p = 1250$ J/kgK yield a dissipation number $D_i = 0.679$ for all plume models.

Viscosity, $\eta$, varies as a function of temperature $T$ and depth $z$

$$\eta(T, z) = \eta(z) e^{-b(T-\bar{T})}.$$  \hspace{1cm} (1)

A linearization of the Arrhenius viscosity law for diffusion creep with $E = 300$ kJ mol$^{-1}$ (e.g., Karato & Wu 1993) yields a viscosity reduction between one and three orders of magnitude over the range of plume excess temperatures we consider (i.e., $b = \ln(10)$ to $b = \ln(1000)$). The depth dependent viscosity prefactor $\eta(z)$ is given by three layers (Figure 1c). The viscosity in the lower mantle, $\eta_3$, is $10^{22}$ Pa s. A stiff upper layer of $\eta_1 = 100 \eta_{\text{LM}}$ simulates a 120-km thick lithosphere. The viscosity $\eta_2$ in the upper mantle is either $\eta_3$ or $\eta_3/30$.

Plumes are initiated by applying a cosine perturbation to the basal thermal boundary layer. The peak perturbation at the symmetry axis is equal to half of the temperature difference between the surface and the CMB. We vary the structure of plumes by modifying the radial viscosity profile, the temperature-dependence of rheology, the temperature contrast across the basal thermal boundary layer, and phase changes in the mantle transition zone. Plumes are broad in a mantle with relatively high viscosity and localized when rheology is strongly temperature dependent. The endothermic phase transition at the base of the mantle transition zone can inhibit plume ascent.

Phase functions describe the relative fraction of each mineral phase as a function of excess pressure at the 410-km and 660-km phase transitions. For models considering phase changes, we assume that the Clapeyron slopes of the olivine-wadsleyite phase change ($\Gamma_{110}$) and the ringwoodite-perovskite phase change ($\Gamma_{660}$) are +3.8 MPa/K and -2.5 MPa/K, respectively. The effects of latent heat on temperature are included in the reference temperature profile.

Plume excess temperature in the upper mantle inferred from OIB major-element geochemistry is expected to be in the range of 100–300 K (e.g., Courtier et al. 2007). Here, plume models slightly exceed this range ($\sim 350–500$ K) for the upper mantle. This is potentially due to a lack of lateral motion imposed by the axisymmetry constraint. It may also be an indication that plumes in the Earth carry a fraction of the heat from the CMB to the surface, potentially due to chemical stratification of a dense layer above the core (e.g., Farnetani 1997).
Stagnation just below the transition zone may also promote cooling (e.g., Ballmer and van Keken 2015). Our models neglect latent heat loss due to melting, but this is likely a second order effect.

We calculate the buoyancy flux $B$ by integrating the product of mass flux and thermal expansion due to the plume excess temperature over a spherical surface $S$ with a radius of 10 degrees, centered on the plume axis.

$$B = \int \rho \alpha w(T - T_A) \, dS$$  \hspace{1cm} (2)

where $\rho$ is density, $\alpha$ is thermal expansivity, $w$ is the upward velocity, $T_A$ is the adiabatic reference temperature. Further details are provided in Bossmann & van Keken (2013).

### 2.2 Seismic velocity conversion

The P-wave ($V_P$) and S-wave ($V_S$) velocity structure of the plume is determined using thermodynamic first principles. We assume a pyrolite mantle composition (Workman & Hart 2005) and the elastic parameter database described in Stixrude & Lithgow-Bertelloni (2011) for the six oxides SiO$_2$, MgO, FeO, CaO, Al$_2$O$_3$, Na$_2$O. Mineral parameters in the database are calculated for a third-order, finite-strain equation of state with Mie-Grüneisen temperature correction. We use the PerPleX software (Connolly 2005) to compute an equilibrium mineral assemblage at each point in $P$ and $T$. $V_P$ and $V_S$ of a bulk mineral assemblage is determined by Voigt-Reuss-Hill averaging of the velocities of each constituent phase (Figure 2a and 2b). The effects of anelasticity on shear wave velocity are incorporated using a model for the S-wave quality factor $Q_S$ that varies with temperature $T$ and depth $z$ as $Q_S(z,T) = Q_0 \omega^a \exp \left( \frac{\alpha T_m}{T} \right)$, where $\omega$ is frequency, $a$ is exponential frequency dependence, $\xi$ is a depth scaling factor, and $T_m$ is the dry solidus melting temperature. Our anelasticity model, $Q_7g$ uses values of $Q_0 = 0.1, 0.5, 1.5$ for the upper mantle, transition zone, and lower mantle respectively. $\xi$ in these intervals is 38, 30, and 26. The frequency $\omega$ is assumed to be 1/20 Hz, and $a$ is assumed to be 0.15. We use the dry solidus calculated in Herzberg et al. (2000) for the upper mantle and Zerr & Diegeler (1998) for the lower mantle. The calculation of $V_P$ and $V_S$ is not well defined where partial melt is present, which may occur in the shallow plume head. We estimate reductions in seismic velocity at temperatures above the dry solidus by linear extrapolation using the local temperature derivatives at the solidus.

A maximum plume temperature anomaly of 750 K in the lower mantle corresponds to shear-velocity reductions of up to 4%. Although plume excess temperature is smaller in the
upper mantle, the shear-velocity reduction is, on average, about 10% along the plume axis. Maximum shear-velocity reductions of 15% near 410 and 660 km depth are due to perturbations of the phase-transition depth.

2.3 3D waveform computations

We compute the full 3D wavefield using the spectral element solver SES3D (Gokhberg & Fichtner 2015), which solves the integral form of the elastic wave equation in a heterogeneous media. We simulate signals with periods between 10–200 s using a bandpass filtered Heaviside source time function. The simulation of $T > 10$ s waveforms for 30-minute long seismograms is computed in approximately 30 minutes real time on 864 parallel compute nodes, each equipped with a GPU accelerator.

The computational domain is a spherical shell that spans 120 degrees in longitude, 70 degrees in latitude, and 2500 km in depth. The domain consists of $4.68 \times 10^7$ elements and $5.85 \times 10^9$ grid points to ensure that at least two elements sample the shortest wavelengths ($\approx 40$ km).

We place the earthquake source at 400 km depth in order to avoid interference of direct waves and depth phases. The plume axis is 45° from the source (Figure 3) so direct S and P waves simulated for teleseismic distances (30–80°) do not interact with the core. We calculate seismograms for the reference background model and each plume model at a grid of stations behind the plume axis (Figure 3b). This geometry represents a hypothetical seismic experiment in the NE Pacific designed to image the mantle structure beneath Hawaii using recordings of earthquakes in the Tonga-Fiji region.

Synthetic seismograms for plume model R1b (Figure 4) show delays and waveform complexity as a function of distance, which are typical for all plume models. Waveform complexity near $X = 4^\circ$ is due to multi-pathing as $S$ interacts with the strong wavespeed gradients at the margins of the plume head. For $X > 8^\circ$, when $S$ crosses the plume axis in the lower mantle, waveforms for plume R1b and the background model are nearly identical except for a time offset (i.e., the traveltime delay). For increasing $X$, the traveltime delay diminishes from about 3 s at $X = 10^\circ$ to about 0.5 s at $X = 40^\circ$. Wave diffraction around the plume conduits is not clearly visible in the $S$ coda at 10 s period.
3 DYNAMICAL PLUME MODELS

We simulate four seismic structures for three plumes with different morphologies by modifying the thermal Rayleigh number $Ra$, $\eta$, $\Delta T_{\text{CMB}}$, and by incorporating phase changes (Figure 5 and Table 1). We choose a limited set of plumes because of the computationally expensive waveform simulations. However, they represent the range of widths and strengths of plume conduits in the lower mantle and the range of expected P-wave and S-wave delays.

Plume models R1a and R1b represent two stages of evolution of the same plume. This plume ascends in a mantle with moderately temperature-dependent rheology ($b = \ln(10^2)$), a thermal Rayleigh number of $2 \times 10^6$, and with phase transitions. $\Delta T_{\text{CMB}} = 750$ K, which renders an excess temperature in the upper mantle of $\sim 450$ K. The viscosity in the upper mantle is $30\times$ smaller than the viscosity in the lower mantle, i.e., $\eta_2 = \eta_3 / 30$. At 45 Myr (model R1a), the broad plume head is still in the lower mantle. At 55 Myr (model R1b), the plume head has crossed the upper mantle transition zone and begins to spread beneath the lithosphere, and its conduit has narrowed due to the reduction in viscosity. Reductions of $V_S$ are as large as 15% in the plume head. The tail has a diameter of 200 km and $V_S$ has been reduced by up to 4%. We define plume tail width to be the point at which plume excess temperature diminishes to half of the maximum value at a given depth, as in Goes et al. (2004).

Plume R2 ascends in a mantle with weakly temperature-dependent viscosity ($b = \ln(10)$), a thermal Rayleigh number $Ra = 10^6$ and $\Delta T_{\text{CMB}} = 750$ K. We omit the effect of phase changes and the viscosity $\eta_2 = \eta_3 = 10^{22}$ Pa s in both the upper and lower mantle, which leads to a more simple lower mantle structure in contrast to R1b. Plume R2 is sluggish because of the weak temperature-dependence of viscosity and the low Ra. Its rise time is $\approx 200$ Myr as opposed to $\approx 50$ Myr for R1b. The plume tail has a diameter of about 400 km in the lower mantle. Without the viscosity reduction in the upper mantle, the plume conduit remains broad after crossing the 660-km phase transition. Although the viscosity structure and Rayleigh number differ significantly, the temperature and $V_S$ along the plume axis in models R1b and R2 are similar because the excess temperature in the upper mantle is primarily controlled by the dissipation number $Di$ and $\Delta T_{\text{CMB}}$ (Albers & Christensen 1996).

The plume in model R3 has developed in a relatively weak thermal boundary layer ($\Delta T_{\text{CMB}} = 550$ K), and it has a relatively small excess temperature ($\approx 350$ K) at the base of the lithosphere. The mantle has a rheology with strong temperature dependence ($b = \ln(10^3)$). As in R1a and R1b, we incorporate a viscosity reduction by a factor of 30 in the upper mantle and the effects of phase changes in the transition zone. The thermal Rayleigh number $Ra =$
The plume tail is narrow (150 km in diameter) and $V_S$ reductions are smaller than in R1a, R1b, and R2. The thin and weak tail of R3 may be the most challenging to image seismically. As plumes R1b and R3 evolve further in time the tails broaden slightly, as they no longer feel the pull of the buoyant plume head.

Figure 6 shows the depth-dependence of buoyancy flux $B$ for each plume model. The variations of $B$ with depth primarily reflects the position of the buoyant plume head. Buoyancy flux also increases from the lower mantle to the upper mantle if the viscosity in the upper mantle is lower. The plume in model R1b has the largest upper mantle buoyancy flux of $B = 20 \times 10^3$ kg/s, which is more than twice as large as $B$ estimated for Hawaii. R1b is not in steady state as the plume head is still rising in the upper mantle. We expect a transient reduction of $B$ for a long-lived plume supported swell (such as Hawaii) after the plume head spreads and cools beneath the lithosphere. The values of $B$ in Table 1 apply to the plume tail at 2500 km depth. These estimates may be most comparable to estimates of buoyancy flux beneath hotspots (Goes et al. 2004).

4 SEISMIC WAVE PROPAGATION THROUGH PLUME MODELS

We analyze the waveform differences between the background model and the four seismic models using two approaches. In the first approach, we measure S and P wave delay time by cross-correlation. In the second approach, we measure instantaneous phase differences (Bozdag et al. 2011), which allow for small amplitude diffracted arrivals to be analyzed.

4.1 Traveltime delays from waveform correlation

The $P$ and $S$ delay times ($\Delta T_P$ and $\Delta T_S$) (Figure 7) are defined by waveform cross-correlation functions. We determine the delay using the first upswing and maximum, which correspond to about an 8 s long wave segment. Given the slight differences between the waveforms for the background and plume models, we modify the cross-correlation window to determine how the delay times vary as a function of window length. From the variability we estimate that our measurements of delay times have uncertainties of about $\pm 0.1$ s at a period of 10 s. The uncertainties are slightly higher at longer periods.

For plumes R1b, R2, and R3, which have plume heads in the upper mantle, the form of $\Delta T_P$ and $\Delta T_S$ up to $X = 10^\circ$ is determined by the shape and width of the plume head in the upper mantle. The peak delay is recorded near the plume axis at about $X = 4^\circ$ when $P$ and $S$ propagate steeply through the center of the plume head. Model R2 produces the
largest ($\Delta T_S = 15$ s and $\Delta T_P = 4$ s) and broadest imprints of $\Delta T_P$ and $\Delta T_S$ because it has the strongest and widest plume head beneath the lithosphere. At $X > 10^\circ$, $\Delta T_P$ and $\Delta T_S$ decrease smoothly as $P$ and $S$ traverse the plume tail at progressively larger depth. For plume model R3, the weakest and thinnest plume, $S$ and $P$ delays are approximately 0.4 s and 0.1 s, respectively. The $P$ and $S$ delays due to plume R1a begin at about $X = 10^\circ$ behind the plume axis when $P$ and $S$ turn in the lower mantle. For all models, the $P$ delay is recorded slightly earlier than the $S$ delay because, at the same period, $Ps$ has a broader Fresnel zone than $S$.

Figure 8 shows $\Delta T_S$ as a function of $X$ along the axis through the event and plume (i.e. $Y = 0$) for periods of $T > 10$ s, $T > 20$ s, and $T > 40$ s, and computed using ray-theory. The ray-theoretical delay times and the delay times determined from waveform correlation have the same character. The maxima in $\Delta T_S$ at $X < 10^\circ$ reflect the complexity of the seismic structure of the plume head. Ray-theory predicts these maxima to be strongly peaked. The maxima in $\Delta T_S$ are smallest and smoothest when measured from the longest period waveforms because, at increasingly longer periods, the Fresnel zones widen.

At $X > 10^\circ$, when $S$ samples the plume tail in the lower mantle (for models R1b, R2, and R3), the ray-theoretical $\Delta T_S$ and $\Delta T_S$ determined by cross-correlation decrease monotonically with $X$ for three reasons. First, for increasing $X$, $S$ crosses the plume tail along a shorter path. Second, $\Delta V_S$ decreases with depth (see Figure 2), albeit slightly. Third, wave diffraction (i.e. wavefront healing) causes wave delays to diminish. This effect is strongest at the longest periods. Therefore, $\Delta T_S$ is smallest when determined from waves with the longest periods and the widest Fresnel zones.

For model R2, a relatively strong plume, $\Delta T_S$ at $X = 35^\circ$ is 1.2 s, 1.0 s, and 0.8 s at periods of 10 s, 20 s, and 40 s, respectively. For model R3, the weakest and narrowest plume, $\Delta T_S$ at $X = 35^\circ$ is 0.5 s, 0.4 s, and 0.3 s at periods of 10 s, 20 s, and 40 s, respectively. These delays are about half the delays predicted by ray theory.

### 4.2 Instantaneous phase misfit

The instantaneous phase analysis (Rickers et al. 2012, 2013) is useful to isolate signals of wave diffraction around the plume tail. These diffractions have low amplitudes relative to direct arrivals and would contribute insignificantly to waveform cross-correlation functions. Instantaneous phase differences are independent of signal amplitude and thus not dominated by the high-amplitude direct arrival. Instantaneous phase misfits, $\Delta \phi(t)$ are calculated for $S$ at periods $T > 20$ s and $T > 40$ s over an extended time window of 80 s to include signal
due to plume diffraction. At this relatively long period and with water-level stabilization, the analysis is not complicated by large phase mismatches. Measurements of instantaneous phase misfit are visually inspected to ensure that $\Delta \phi(t)$ is well behaved.

Figure 8 (right column) shows the L2 norm of the instantaneous phase misfit, $||\Delta \phi(t)||$, as a function of $X$ for the seismic models in Figure 5. The form of $||\Delta \phi(t)||$ resembles $\Delta T_S$ determined by cross correlation. However, $||\Delta \phi(t)||$ decays more slowly with distance because the small-amplitude diffractions contribute to the measurement of the delay. At periods of $T > 40$ s, cross-correlation measurements of R3 decay to less than 2% of their maximum value at $X = 10^\circ$. Equivalently large instantaneous phase misfits are still observed up to $X = 40^\circ$. In addition, non-zero instantaneous phase misfit values are recorded over a wider range in azimuth. These results demonstrate that, at low frequencies, the instantaneous phase measurement is more useful than time-domain waveform cross-correlation to resolve the narrow plume conduits in the lower mantle, in agreement with Rickers et al. (2012).

5 DISCUSSION AND CONCLUSIONS

Models R1b and R3 (see Figure 5) are our end-member estimates of $V_P$ and $V_S$ reductions within the conduits in the lower mantle. The seismic structure of R1b includes up to 2% and 4% reductions in $V_P$ and $V_S$ which delay teleseismic $P$ and $S$ waves by about 0.15 s and 0.7 s. $V_P$ and $V_S$ in the tail of R3 are reduced by as much as 1.8% and 3.5% and lead to $P$ and $S$ delays of about 0.1 s and 0.45 s. The delay times depend on the chosen frequency band in which $P$ and $S$ are analyzed because Fresnel zones broaden with decreasing frequency. $S$ delays for model R1b (at a distance $X = 35^\circ$ in Figure 8) are 0.7 s for $T > 10$ s and 0.4 s for $T > 40$ s. $S$ delays for model R3 are 0.5 s for $T > 10$ s and 0.3 s for $T > 40$ s. These delays are up to 50% smaller than ray-theoretically predicted delays. To ensure that the presence of a plume head in the upper mantle is not biasing our results, in addition to the work shown, we separately modeled the upper and lower mantle expression of plume R1b (i.e., both just the head and just the tail). We find that delays induced by the plume head disappear entirely for distances larger than $X = 10^\circ$, and that any delay signal beyond this distance can be attributed entirely to the plume tail.

Given that plumes may have a distinct composition (Ballmer et al. 2013; Lin & van Keken 2005; Dannberg & Sobolev 2015) and that the conversion between temperature, composition, and wave speed structure is uncertain (Styles et al. 2011; Cobden et al. 2008), we estimate that the $P$ and $S$ wave reductions are uncertain by 30%. Waveform simulations indicate that the $P$ and $S$ delays depend linearly on the $V_P$ and $V_S$ reductions in agreement with previous
modeling (Mercerat & Nolet 2013). If $V_P$ and $V_S$ are enhanced or reduced by 30%, the delay times increase or decrease by 30% (Figure 9).

These delays are somewhat larger than the delays we have previously determined using 2D modeling (Hwang et al. 2011). Nevertheless, we remain skeptical that faint delays in $P$ ($< 0.3$ s) associated with thin thermal plumes are detectable in currently available seismic data sets. The 0.4–0.8 s delay of $S$ waves by lower mantle plume conduits and the diagnostic frequency dependence should be observable when the influence of the heterogeneous crust and uppermost mantle is small. For example, recordings of abundant earthquakes in the Tonga-Fiji and Kermadec regions by a wide-aperture network of ocean-bottom seismometers (OBS) in the northeast Pacific (as sketched in Figure 3b) would provide wave sampling of the lower mantle beneath Hawaii. Since the structure of the crust and lithosphere beneath the northeast Pacific is relatively simple, traveltime delays accrued in the lithosphere may be estimated using plate cooling models and from delay time measurements over a broad range of source azimuths. The traveltime dispersion due to reverberations in the crust (Hwang et al. 2011) is different than the dispersion due to plumes and can be be estimated from layered models of the oceanic crust. Ideally, such OBS network cover the ocean floor beyond the Hawaiian swell to ensure lower mantle sampling beneath Hawaii and to record the smooth and systematic decay of the traveltime delays.

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REFERENCES


Table 1. Physical parameters used in plume simulations.

<table>
<thead>
<tr>
<th>Ra</th>
<th>Time (Myr)</th>
<th>$\Delta T_{\text{CMB}}$ (K)</th>
<th>$b$</th>
<th>$B$ (Mg/s)</th>
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Figure 1. References profiles for (a) temperature and (b) $V_S$ (red line), $V_P$ (blue line), and density (black line). Reference values are calculated along the reference geotherm of the dynamic plume models. Note the anomalous changes in temperature, $V_S$, $V_P$, and density near the 410-km and 660-km phase transitions (dashed lines) due to latent heat effects. (c) Prefactors of three-layer viscosity profile $\eta(z)$: $\eta_3 = 10^{22}$ Pa s in the lower mantle, $\eta_1 = 100\eta_3$ in the lithosphere, $\eta_2 = \eta_3$ or $\eta_2 = \eta_3/30$ in the upper mantle.

Figure 2. (a) $V_P$ and (b) $V_S$ as a function of pressure and temperature. The blue line is the geotherm for the reference structure. The red line is the geotherm along the plume axis of model R1b. The dry solidus of pyrolite is shown as a bold dashed line. Seismic velocity contours are shown every 0.1 km/s. (c) Plume excess temperature, $\Delta T$, and (d) shear-velocity reduction along the axis of plume R1b. Peaks near 410 and 660 km depth are due to phase transitions.
Figure 3. (a) Vertical cross section of the geometry of the seismic model. Plume R1b is at X=0, the earthquake at X=−45° and a depth of 400 km. The black line are ray paths to illustrate the $P$ and $S$ propagation through the plume. (b) Map view representing the geometry of the model domain. The domain spans 120 degrees in X and 70 degrees in Y. The earthquake (yellow star), plume (yellow circle), and a grid of seismic stations (dashed line) represents a hypothetical seismic deployment designed to image the mantle beneath Hawaii using recordings of earthquakes in the Fiji-Tonga region at stations in the NE Pacific.
Figure 4. $S$ waveforms for the background model (black) and plume model R2 (red) as a function of distance $X$ from the plume axis. The waveforms have a minimum period $T = 10$ s. They have been aligned on the theoretical arrival time of the $S$ wave for the background model.

Figure 5. The temperature field (left half) and $S$-wave velocity perturbations $\Delta V_S$ (right half) of plume models R1a, R1b, R1c, R2, and R3. $\Delta V_S$ is relative to the reference shear velocity profile in shown in Figure 1. Temperature and $\Delta V_S$ contours are shown every 200 K and every 2%, respectively. The cross sections are $20^\circ$ wide and extend from the surface to the core mantle boundary.
Figure 6. Plume buoyancy flux as a function of depth for plumes R1a (red), R1b (blue), R2 (orange), and R3 (green).

Figure 7. Cross-correlation delay times for plume models R1a, R1b, R2, and R3 as a function of X and Y. The top half and bottom half of each map show P and S delays, respectively. The S wave delay time scale is 4 times wider than the P delay time scale. P wave contours are drawn every 0.125 s, starting at 0.1 s. S wave contours are drawn every 0.5 s, starting at 0.4 s.
Figure 8. (left column), $\Delta T_S$ as a function of $X$ along the earthquake-plume axis (i.e., $Y = 0$) for each of the four plume models. The dashed line shows calculated ray theoretical delays. The solid lines show $\Delta T_S$ determined by cross-correlation of waveforms with periods larger than (red) 10 s, (blue) 20 s, and (green) 40 s. At distances greater than $X = 20$, the vertical scale is exaggerated to show detail. (right column) Norm of instantaneous phase misfit measured along the earthquake-plume axis for periods larger than 20 s (blue) and periods larger than 40 s (green).
Figure 9. $P$ (left) and $S$ (right) delay times as a function of distance $X$ behind the plume for models R1b (a and b) and R3 (c and d), measured at periods larger than 10 s. The shaded regions indicates the measurement uncertainty of $\pm 0.1$ s. The black line shows delay times for models R1b and R3 (see Figure 5). The blue and red lines show measured delay times after multiplying the $P$ and $S$ velocity reductions by a factor 0.7, and 1.3 respectively. These represent the upper and lower bounds of the uncertainties associated with the temperature conversion.