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Recommended Citation

Hassan, Rakib; Muller, R Dietmar; Gurnis, Michael; Williams, Simon E.; and Flament, Nicolas, "A rapid burst in hotspot motion through the interaction of tectonics and deep mantle flow" (2016). *Faculty of Science, Medicine and Health - Papers: part A*. 4193.
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A rapid burst in hotspot motion through the interaction of tectonics and deep mantle flow

Abstract

Volcanic hotspot tracks featuring linear progressions in the age of volcanism are typical surface expressions of plate tectonic movement on top of narrow plumes of hot material within Earth's mantle¹. Seismic imaging reveals that these plumes can be of deep origin²-probably rooted on thermochemical structures in the lower mantle^{3, 4, 5, 6}. Although palaeomagnetic and radiometric age data suggest that mantle flow can advect plume conduits laterally^{7, 8}, the flow dynamics underlying the formation of the sharp bend occurring only in the Hawaiian-Emperor hotspot track in the Pacific Ocean remains enigmatic. Here we present palaeogeographically constrained numerical models of thermochemical convection and demonstrate that flow in the deep lower mantle under the north Pacific was anomalously vigorous between 100 million years ago and 50 million years ago as a consequence of long-lasting subduction systems, unlike those in the south Pacific. These models show a sharp bend in the Hawaiian-Emperor hotspot track arising from the interplay of plume tilt and the lateral advection of plume sources. The different trajectories of the Hawaiian and Louisville hotspot tracks arise from asymmetric deformation of thermochemical structures under the Pacific between 100 million years ago and 50 million years ago. This asymmetric deformation waned just before the Hawaiian-Emperor bend developed, owing to flow in the deepest lower mantle associated with slab descent in the north and south Pacific.

Disciplines

Medicine and Health Sciences | Social and Behavioral Sciences

Publication Details

Hassan, R., Muller, R. Dietmar., Gurnis, M., Williams, S. E. & Flament, N. (2016). A rapid burst in hotspot motion through the interaction of tectonics and deep mantle flow. *Nature*, 533 (7602), 239-242.

A rapid burst in hotspot motion through the interaction of tectonics and deep mantle flow

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Age progressive volcanic hotspot tracks are typical surface expressions of plate tectonic movement atop narrow plumes of hot material within the Earth’s mantle¹. Seismic imaging reveals that these plumes can be of deep origin² — likely rooted on thermochemical structures in the lower mantle^{3–6}. Although paleomagnetic and radiometric age data suggest that mantle flow can advect plume conduits laterally^{7,8}, the underlying flow dynamics that resulted in a sharp bend exclusively in the Hawaiian-Emperor hotspot track in the Pacific remains enigmatic. Here we present paleogeographically constrained numerical models of thermochemical convection and demonstrate that flow in the deep lower mantle under the north Pacific was anomalously vigorous between 100–50 Ma as a consequence of long-lasting subduction systems, in contrast to those in the south Pacific. Deep flow in the lower mantle and plume trajectories found in these models show a sharp bend in the Hawaiian-Emperor hotspot track through the interplay of plume tilt and the lateral advection of plume sources. The different trajectories of the Hawaiian and Louisville hotspot tracks arise from asymmetric deformation of thermochemical structures under the Pacific between 100–50 Ma that waned just prior to the time of the Hawaiian-Emperor bend, owing to flow in the deepest lower mantle associated with slab descent in the north and south Pacific.

The sharp bend in the Hawaiian-Emperor (HE) hotspot track has been interpreted as the result of a sudden change in Pacific plate motion over a fixed plume. In contrast, its south Pacific analogue — the Louisville hotspot track — has a more gradual arc through the period of

the bend⁹. Paleomagnetic and radiometric age data suggest that the Hawaiian plume underwent a phase of rapid 11–15° southward motion between 81–47 Ma⁷, while the Louisville plume remained within 5–7° of present-day latitude between 70–50 Ma⁸, suggesting that they are moving independently⁸. Seismic imaging reveals that these are robust plumes deeply rooted in the mantle² and deep plumes are likely to originate from thermochemical structures^{3–6}. Moreover, the spatial correlation of hotspot locations with the edges of large low shear velocity provinces (LLSVPs)⁶ suggests that these plumes are likely to be rooted on thermochemical ridges^{3,10}. Numerical models of thermochemical structures suggest that their edges should be mobile^{11,12}, in contrast to the longer-term stability implied by the correlation of the edges of LLSVPs and reconstructed volcanic features^{13,14}.

A combination of plume motions derived from backward advection of mantle flow with a model of plate motions that explicitly incorporated a major change in relative plate motions between 52 and 43 Ma was proposed to reproduce the observed HE track¹⁵. These earlier flow models result in a significant misfit of the predicted trail prior to 65 Ma¹⁵, and predict a gradual slowdown of the Hawaiian plume motion compared to the abrupt slowdown around HE bend time, as indicated by paleomagnetic data. Moreover, Pacific-Farallon relative plate motion changes contemporaneous to the HE bend were insignificant and gradual^{7,16}, suggesting that the bend is primarily a consequence of rapid southward migration of the Hawaiian plume^{7,17,18}.

A capture-release mechanism was also proposed¹⁷, where the Hawaiian plume may have been captured by the fast spreading Kula-Pacific ridge system, situated to the north of Detroit seamount during its construction. In this scenario, vigorous shallow mantle flow induced a strong plume tilt starting at mid-mantle depth, before cessation of spreading between 56–47 Ma resulted in a change in mantle flow regime. Consequently, the plume conduit rapidly returned to its original position, resulting in the HE bend¹⁷. However, an assumption implicit in this scenario is that the

base of the plume remains fixed, relative to lower mantle structures, during the change in flow regime. Here we advance the hypothesis that between ≈ 100 –50 Ma deep mantle flow beneath the north Pacific was anomalously vigorous, a consequence of robust and long-lasting subduction systems, compared to those in the south. This asymmetric flow in the deepest lower mantle under the north and south Pacific explains the rapid southward motion of the Hawaiian plume and the contrasting sluggish latitudinal motion of the Louisville plume, resulting in sharp differences in the geometry of hotspot tracks.

We follow the trajectory and tilt of plumes over the last 230 Myr with paleogeographically constrained, spherical numerical models of convection^{10,11,19}. We impose kinematic boundary conditions, consistent with plate motions, and assimilate thermal models of shallow slabs, at convergent plate margins (Methods). Descending slab material deforms a basal layer of anomalous density^{20,21} (Methods), covering the core mantle boundary (CMB), where slab descent rates observed in models¹⁰ are consistent with estimates inferred from global mantle tomography²². Viscosity in the deepest lower mantle, below the viscosity-peak^{23,24}, is low due to elevated temperatures in the thermal boundary layer above the CMB²¹ and thus flow velocities there are expected to be higher compared to those in the ambient lower mantle. This is a consistent feature of our models, in agreement with observations from earlier flow models²⁵. Consequently, within the first 50 Myr of model time, the dense material above the CMB deforms, ultimately resembling the seismologically-observed, large low shear velocity provinces (LLSVPs) in the lower mantle (Methods, Extended Data Fig. 4). We refer interchangeably to these dense thermochemical structures above the CMB in our models as LLSVPs. Fully dynamic plumes nucleate primarily around the LLSVPs and plume eruption locations are highly correlated to reconstructed eruption locations of large igneous provinces (LIPs)¹⁰. The models are characterized roughly by as many plumes at present-day as can be inferred to be of deep origin from tomographic studies^{2,26}.

Subduction history has been suggested to play a first-order role in shaping the geometry of LLSVPs^{10–12,27}. Here we compute poloidal flow, associated with the buoyancy resulting from subduction, in the deep lower mantle to evaluate the temporal evolution of the edges of LLSVPs. We compute mean poloidal flow within a 300 km thick shell above the CMB and subsequently compute a time-average of this evolving mean poloidal flow over the last 140 Myr of modeled time (Methods), highlighting regions that undergo strong coherent poloidal flow relative to regions where it is comparatively weak and diffuse. Over the last 140 Myr, subduction zones bounding the north Pacific migrate oceanwards (Fig. 1a), so the deep mantle here is dominated by a strong coherent flow. By contrast, flow in the south Pacific is diffuse and the darker region under the central Pacific marks the mean areal extent of the Pacific LLSVP where mean poloidal flow is weakest (Fig. 1a).

The southward trajectory of the Hawaiian plume is nearly perpendicular to the strike of the northern edge of the Pacific LLSVP, marked by zones of quasi-stagnation (Fig. 1a). Southeastward flow beneath the northwest Pacific converges with southwestward flow from the northeast Pacific, resulting in a strong net southward flow that reaches a maximum along a corridor coincident with the location of present-day Hawaiian plume. The edge of the Pacific LLSVP recedes southward along this corridor (Fig. 1a) by $\approx 10^\circ$ over the last 100 Myr (Extended Data Fig. 2), making it a unique geodynamical setting. The pulse of southward poloidal flow in the north Pacific between 100–50 Ma is counterbalanced by the northeasterly poloidal flow under the southwest Pacific that strengthens after ≈ 85 Ma (Extended Data Fig. 2) — resulting in the transition from a phase of rapid southward motion of H_m to a much slower regime at ≈ 50 Ma (Fig. 1b, Extended Data Fig. 3).

A southward velocity of around 3 cm yr^{-1} in the deep lower mantle at 90 Ma (Fig. 2a, bottom row) translates to over 5 cm yr^{-1} near the surface, consistent with paleomagnetic estimates⁷. Over the same period, the edge of the LLSVP in the south Pacific — marked by the trough in velocity

(Figs 2a-d, bottom row) — remains comparatively stable. The Louisville plume is thus isolated from a rapid burst and slowdown of motion, similar to the one causing the sharp HE bend.

A dynamical explanation for the sharp bend in the HE track requires an abrupt slowdown of the plume at the time of the bend¹⁵, which cannot be accounted for by deep lower mantle flow alone. Ascent speed within plume conduits is low around the viscosity-peak in the lower mantle and is comparable to rates of horizontal advection, whereas in the shallow lower mantle and above, ascent dominates the motion of a plume conduit. Consequently, lateral advection of plume sources rooted on chemical ridges in the deep lower mantle is the primary factor controlling plume tilt. Between ≈ 81 –47 Ma, the surface location of H_m moves southward by $\approx 10^\circ$, followed by a sharp slowdown — while during this period, the base of the plume moves only $\approx 6^\circ$ (Fig. 1b). The remaining $\approx 4^\circ$ of motion near the surface arises from the slowdown of southward motion of the base of H_m . Consequently, the latitudinally tilted plume rapidly straightens, resulting in an abrupt slowdown in hotspot motion near the surface at ≈ 47 Ma (Fig. 1c, Extended Data Fig. 3). Predicted hotspot tracks are nearly identical to the observed HE track (Fig. 3a) and latitudinal motion of H_m corresponds to the much faster southward motion of Hawaiian plume prior to the HE bend (Fig. 3b), consistent with paleomagnetic data⁷.

Numerical experiments computed over a range of uncertain parameters (Methods, Extended Data Table 2) demonstrate a consistent pattern of strong time-averaged poloidal flow in the north Pacific compared to that in the south (Extended Data Fig. 1). Rapid southward migration of H_m is observed when the density contrast of the LLSVP material ranges between ≈ 2 –2.5%; for larger density contrasts and when the LLSVP material is intrinsically more viscous (Extended Data Figs 1f and 1e, respectively) the areal extent of the Pacific LLSVP is greater and plume motion smaller. Plume mobility could be a criterion to constrain the density contrast of LLSVP material.

We demonstrate that for a range of reasonable parameter values, the Pacific LLSVP deforms

asymmetrically between 100–50 Ma and that predicted hotspot tracks explain the sharp HE bend, without a major change in plate motion. We note that our present models do not capture the potential influences of a waning ridge system in the vicinity of a plume and that more sophisticated convection models — featuring self-consistent ridges and transform boundaries — are necessary to test the dynamic plausibility of a capture-release mechanism¹⁷ for plumes rooted on mobile lower mantle structures. Our results imply that the Hawaiian plume may have been active since ≈ 140 Ma, longer than previously recognized, and that the remnants of this activity would have been subducted beneath Kamchatka²⁸. Even though there is robust evidence for the overall stability of the African LLSVP for times while Africa was relatively stationary and distal from migrating subduction zones²⁹, our results demonstrate that a tectonic setting of fast moving oceanic plates surrounded by a migrating, dynamic set of subduction zones can lead to significant LLSVP deformation and hotspot mobility.

Methods

The numerical model. We consider models of thermochemical convection within the Earth’s mantle under the extended-Boussinesq approximation³⁰ in a spherical shell, with depth-dependent thermodynamic properties and temperature- and depth-dependent rheology. The equations for the conservation of mass, momentum and energy are solved using CitcomS³¹ as a finite element problem, which has been modified to allow for progressive assimilation of surface plate motion and inferred slab material based on global plate reconstructions¹⁹. Our choice of model parameters and underlying assumptions have been outlined in earlier work^{10,11}. Extended Data Table 1 lists parameters held fixed across all model cases and additional details can be found in¹⁰.

Model setup. The spherical shell representing the Earth’s mantle is decomposed into ≈ 12.6 million mesh elements. Radial mesh refinement provides a vertical resolution of ≈ 15 km and ≈ 27 km near the top and bottom boundary layers, respectively. We assume an *a priori* mantle adiabat

augmented by thermal boundary layers (TBLs). The top and the bottom TBLs each encompass a temperature drop of 1225 K and the initial adiabatic temperature profile has a potential temperature of 1525 K. A non-dimensional internal heat generation rate of 100 is applied in all model cases. We compute a reference profile for thermal expansion based on analytical parameterizations of $\alpha(T, z)^{32}$, using the *a priori* mantle temperature profile.

We use piecewise Arrhenius laws to describe the variation of viscosity with temperature, depth and composition in the Earth’s mantle, which takes the following non-dimensional form:

$$\eta(T, r) = A(r)\eta_c \exp\left(\frac{E_a(r) + (1 - r)V_a(r)}{T + T_{off}} - \frac{E_a(r) + (1 - r_{inner})V_a(r)}{1 + T_{off}}\right), \quad (1)$$

where r , A , η_c , E_a , V_a and T_{off} are the radius, pre-exponential parameter, intrinsic composition-dependent pre-factor, activation energy, activation volume and temperature offset, respectively. The second term within the exponential ensures that it reduces to 1, when $T = 1$ and $r = r_{inner}$. For the lower mantle, we use a dimensional activation energy of 320 KJ mol⁻¹ and activation volume of 6.7E-6 m³ mol⁻¹, corresponding to non-dimensional units of 11 and 26, respectively, comparable to estimates in³³. However such viscosity parameterizations lead to large viscosity variations, causing numerical difficulties. In order to limit the viscosity contrast to 3 orders of magnitude, we adjust the pre-exponential parameter $A(r)$ and the temperature offset T_{off} ³⁴. The resulting viscosity profile is similar to the preferred viscosity profiles of²³. Additional details on model setup can be found in¹⁰.

Initial and boundary conditions. We apply kinematic surface boundary conditions that are a function both of the relative plate motion (RPM) and absolute plate motion (APM) model. We use the RPM from Seton et al.³⁵, updated for circum-Arctic regions as described in Shephard et al.³⁶. In most model cases we use the APM model of Torsvik et al.²⁹ between 0 and 70 Ma, the APM

model of Steinberger and Torsvik³⁷ between 105 and 200 Ma with an interpolation between 70 and 105 Ma. We refer to this hybrid APM model as APM1 in Extended Data Table 2. We also test a kinematic scenario using the same RPM model combined with an APM model constrained by lower mantle slab remnants²², which we refer to as APM2 in Extended Data Table 2. Surface velocities derived from these global plate tectonic reconstructions are obtained in one million year intervals with a linear interpolation in between.

A thermal model of slabs is constructed based on the reconstructed location and age of oceanic lithosphere at convergent plate margins. These slabs are initially inserted from the surface to a depth of 1200 km. Slabs that appear during modelled geological time are progressively inserted into the upper mantle based on their age of appearance. This imposed thermal structure is blended with the dynamically evolving temperature field at each timestep — see Bower et al.^{11,19} for a more detailed description of the progressive data assimilation method. This simple approach captures the essential aspects of subduction by injecting slabs with realistic thickness and mass flux, without requiring complex rheological laws to model the physics of plate boundaries in our models.

In order to evaluate the influence of parameter values and initial conditions we have computed a range of models where we varied the density contrast of the dense chemical layer above the CMB, the intrinsic viscosity contrast of the dense material, the initial geometry of the dense layer and the plate reconstruction that dictates surface plate velocities and the location of subduction zones. Extended Data Table 2 lists the model cases presented in this study. In most model cases, an initially uniform ≈ 100 km thick layer of anomalously dense material covers the CMB. However, prior to the Mesozoic, the dense LLSVP material may have already been displaced and deformed by slabs^{11,38}. To test the influence of an already thickened dense layer on model outcomes, we present a model case (M5 in Extended Data Table 2) where the dense material is confined to two slightly elongated domes — resembling present-day LLSVPs — in the initial condition. These

191 domes are ≈ 1000 K hotter than the ambient mantle¹¹ and have a relief of ≈ 450 km, with their
 192 combined volume comparable to estimates of the volume of LLSVPs¹³.

Poloidal flow in the lower mantle. The velocity field, $\vec{u}$, in our global models is divergence free,

$$\nabla \cdot \vec{u} = 0. \quad (2)$$

193 This allows us to decompose the tangential velocity field on each radial shell of the spherical
 194 mesh into poloidal and toroidal components. The poloidal flow originates from buoyancy driven
 195 convergent or divergent flows, such as at spreading centres and subduction zones. Following from
 196 Helmholtz's theorem on spheres³⁹,

$$\vec{V}_s = \nabla_s \Phi + \nabla_s \times (\Psi \vec{r}), \quad (3)$$

197 where $\vec{V}_s$, ∇_s , Φ , Ψ and $\vec{r}$ are the tangential velocity on a radial shell of the computational domain,
 198 gradient operator on a spherical shell, poloidal potential, toroidal potential and radial unit vector.
 199 See Backus³⁹ for more details.

The poloidal potential, Φ , is obtained as

$$\Phi = \nabla_s^{-2}(\nabla_s \cdot \vec{V}_s). \quad (4)$$

200 Subsequently, the poloidal flow field, V_p , on a radial shell is obtained as

$$\vec{V}_p = \nabla_s \Phi. \quad (5)$$

Model hotspot tracks and plume tilt. A detailed account of a scheme for detecting plumes in global mantle convection models is given in Hassan et al.¹⁰. We process model outputs at 5 Myr intervals, since the transit time of a plume head from nucleation to eruption under the base of the lithosphere is always > 5 Myr. We compute a set of extant plumes conduit locations, $S_{t_j} = \{P_i, \dots, P_n\}$, at a depth of 350 km based on the plume detection scheme, where subscript n denotes the total number of plumes detected at a given model time, t_j . Age progressive model hotspot tracks in the mantle frame of reference are subsequently identified by binning conduit locations, P_i , from the ensemble set $S = \{S_{t_j}, \dots, S_{t_m}\}$ based on spatiotemporal proximity, where t_m denotes model time corresponding to present-day.

A fourth-order Runge-Kutta scheme is used to compute streamlines, passing through each P_i and spanning mantle depths of 350–1500 km, which mark the topological skeletons of extant plume conduits at a given model time. We limit the depth-extent of these streamlines to 1500 km, since plumes are rooted on chemical ridges that can reach more than ≈ 1000 km above the CMB. Plume tilt is then computed based on P_i and its corresponding location at 1500 km depth, relative to the radial normal.

Predicted hotspot tracks. The motion of a model plume in the mantle frame of reference can be represented implicitly by a sequence of finite rotations on the sphere, decoupling it from explicit spatial locations. By applying such motion trajectories obtained from convection models — such as H_m (Fig. 1a) — to the corresponding observed plume (Hawaii), a direct comparison can be made between predicted and the observed hotspot tracks.

Comparison of model LLSVPs with tomography. Extended Data Fig. 4 shows a comparison of present-day model LLSVPs for case M3 (Extended Data Table 2) with the SAVANI tomography model⁴⁰. The first-order shapes of both the African and Pacific LLSVPs are largely in agreement with our model. The western extension of the Pacific LLSVP is much smaller in north-south

extent than in the center, consistent with our model. The Pacific LLSVP in the east is broken up into blobs, where the details are a little different compared to our model, but it shows that in the east this feature is more discontinuous. However, model LLSVP edges in the vicinity of the actual and model Hawaiian plumes agree well with the tomography (Extended Data Fig. 4).

In the African hemisphere, the tomography model shows the long east-west oriented eastern arm of our African model LLSVP in the southern central Indian Ocean, thickening to the west as it straddles South Africa, and a more discontinuous northern extension where we have a narrow arm stretching towards Iceland. The ‘Perm Anomaly’⁴¹ beneath present-day western Siberia appears as a rounded feature in the tomography model, compared to the corresponding feature that is more elongate in our model — though both are isolated and separated from the northern part of the African LLSVP. This comparison suggests that our models do a reasonable job in capturing lower mantle dynamics — considering their relative simplicity — and it is not expected that model LLSVP edges would match higher-order geometric features found in mantle tomography models.

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Acknowledgements M.G. was supported by the NSF (awards EAR1161046 and EAR1247022). R.D.M. and N.F. were supported by Statoil ASA through ARC IH130200012.

338 **Author Contributions** R.H. and R.D.M. conceived and collaborated in developing the concept of the
339 study. R.H. and M.G. designed the numerical experiments and collaborated in developing technical aspects
340 of the study. All authors contributed both intellectually and to the writing of the paper.

341 **Competing Interests** The authors declare that they have no competing financial interests.

342 **Correspondence** Correspondence should be addressed to R.H. (email: rakib.hassan@sydney.edu.au).

Extended Data Table 1: Physical Parameters and Constants

Parameter	Symbol	Value	Units
Rayleigh number	Ra	5×10^8	-
Earth radius	R_0	6371	km
Density	ρ_0	3930	kg m^{-3}
Thermal expansivity	α_0	1.42×10^{-5}	K^{-1}
Thermal diffusivity	κ_0	1×10^{-6}	$\text{m}^2 \text{s}^{-1}$
Specific heat capacity	C_p	1100	$\text{J kg}^{-1} \text{K}^{-1}$
Gravitational acceleration	g	10	m s^{-2}
Surface Temperature	T_s	300	K
Dissipation number	Di	0.8	-
Reference Viscosity	η_0	1×10^{21}	Pa s
Internal Heating	H	100	-

Extended Data Table 2: Model cases

Case	$\Delta\rho_{ch}\%$	APM Model	η_c	Geometry of Dense Layer
M1	2.125	APM1	1	Uniform
M2	2.25	APM1	1	Uniform
M3	2.5	APM1	1	Uniform
M4	2.5	APM2	1	Uniform
M5	2.5	APM1	1	Domed
M6	2.5	APM1	5	Uniform
M7	3.0	APM1	1	Uniform

Figure 1: Time evolution of model plume trajectories in a model of thermochemical convection (Extended Data Table 2, case M3). (a) Time-averaged mean poloidal flow (see text) is shown in the background and associated velocity vectors are shown in the foreground. Subduction zones over the last 140 Myrs are plotted at 20 Myr intervals. Red triangles indicate present-day location of observed plumes of deep origin at the surface. Corresponding model plume trajectories for Hawaii (H_m), Louisville (L_m), Easter (E_m), Kerguelen (K_m) and Tristan (T_m), at a depth of 350 km, are shown in multicolor that generally reflect the mean poloidal flow. Model plumes in the Indo-Atlantic are shorter-lived and are less mobile compared to those in the Pacific. See Fig. 2 for time-dependent profiles along PP' and QQ'. (b) Latitudinal deviation, $\delta\lambda$, of model plumes from their present-day locations at a depth of 350 km. Latitudinal offset between the locations of plume conduits at depths of 350 and 1500 km is shaded in corresponding colors. Between ≈ 81 –47 Ma (shaded rectangle), H_m rapidly moves southward by $\approx 10^\circ$. (c) Model plume tilt, as outlined in Methods. Between ≈ 81 –47 Ma, tilt accumulated by H_m over an earlier period of coherent southward motion decreases sharply.

Figure 2: Longitudinal cross sections. Cross sections through H_m (see Fig. 1a) show its southward motion and the evolution of tilt that arises from the offset of the base of the plume relative to conduit location near the surface (see Figs 1b-c). Ascent speed within model plume conduits wane towards present-day as plumes mature. This is consistent with large volumes of lava being emplaced on the earth's surface when a plume-head erupts, followed by weaker magmatic activity associated with a waning plume-tail. (a) Temperature profile along a constant meridian, through H_m , at labelled age and between latitudinal extent shown. Temperature field away from layer averages in the panel is contoured by the depth-averaged standard deviation of layer temperature values to delineate the plume conduit in red. The thick black contour marks 75% anomalous chemical concentration. Velocity vectors outside of the plume conduit are shown in grey to avoid visual clutter. Intensity of mean poloidal flow in a 300 km thick shell above the CMB, at labelled age, along profiles PP' and QQ' (see Fig. 1a), are shown in the bottom panel. (b to d) The same as in (a), but for labelled ages.

Figure 3: Predicted hotspot tracks. (a) The thick multicolored trajectory represents the motion of Hawaiian plume, based on the motion of H_m (see Fig. 1). The thin multicolored trajectory represents the same from a similar model (Extended Data Table 2, case M4), where the plate reconstruction is based on a reference frame constrained by lower-mantle slab remnants (Methods). The thick solid and thin dashed black lines represent predicted surface hotspot tracks, computed based on the thick and thin multicolored trajectories of H_m , respectively (Methods). White crosses along these hotspot tracks mark 10 Myr intervals. The background is shaded by gravity anomaly to illuminate the Hawaiian-Emperor seamount chain. (b) The thick solid and thin dashed black lines show paleolatitudes corresponding to the thick and thin multicolored trajectories in (a), respectively. Red circles and error bars are inferred paleolatitudes for Emperor seamounts⁷.

Extended Data Figure 1: Inter-model comparisons. (a) For case M1 (Extended Data Table 2), the intensity of mean poloidal flow in a 300 km thick shell above the CMB, time-averaged over the last 140 Myrs is shown in the background shading and associated velocity vectors are shown in the foreground. Subduction zones over the last 140 Myrs are plotted at 20 Myr intervals. Model plume trajectory for Hawaii (H_m) at a depth of 350 kms is shown in multicolor. (b) For case M2. (c) For case M4. (d) For case M5. (e) For case M6. (f) For case M7.

Extended Data Figure 2: Evolution of mean poloidal flow. Animation (S2) showing the evolution of mean poloidal flow in a 300 km thick shell above the CMB for case M3 (Extended Data Table 2). The white rectangular region marks the extent of the three dimensional plot in Extended Data Fig. 3. Cross sections along profiles through the Pacific LLSVP show the evolution of its edges driven by subduction-induced flow.

Age :100 Ma

Age :80 Ma

Age :60 Ma

Age :40 Ma

Age :20 Ma

Age : 0 Ma

Extended Data Figure 3: Trajectory of model Hawaiian plume. Animation (S3) in three dimensions (Cartesian projection of spherical geometry) showing the southward motion and evolution of tilt for model plume (H_m) corresponding to Hawaii in case M3 (Extended Data Table 2). The black contour marks the 75% chemical concentration isosurface 100 km above the CMB.

Extended Data Figure 4: Comparison of model LLSVPs with tomography. SAVANI tomography model⁴⁰ at 2818 km depth. Contours of the 75% chemical concentration isosurface, at labeled heights above the CMB, show the present-day shapes of the model LLSVPs in case M3 (Extended Data Table 2). The red triangle and purple cross symbols mark the locations of the actual and model Hawaiian plumes at present-day, respectively.