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SUPPLEMENTARY MATERIALS

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 Materials and Methods
 Figs. S1 to S9
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 MDAR Reproducibility Checklist
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REPORTS

GEOPHYSICS

On the relative temperatures of Earth's volcanic hotspots and mid-ocean ridges

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Volcanic hotspots are thought to be fed by hot, active upwellings from the deep mantle, with excess temperatures (T_{ex}) ~100° to 300°C higher than those of mid-ocean ridges. However, T_{ex} estimates are limited in geographical coverage and often inconsistent for individual hotspots. We infer the temperature of oceanic hotspots and ridges simultaneously by converting seismic velocity to temperature. We show that while ~45% of plume-fed hotspots are hot ($T_{\text{ex}} \geq 155^\circ\text{C}$), ~15% are cold ($T_{\text{ex}} \leq 36^\circ\text{C}$) and ~40% are not hot enough to actively upwell ($50^\circ\text{C} \leq T_{\text{ex}} \leq 136^\circ\text{C}$). Hot hotspots have an extremely high helium-3/helium-4 ratio and buoyancy flux, but cold hotspots do not. The latter may originate at upper mantle depths. Alternatively, the deep plumes that feed them may be entrained and cooled by small-scale convection.

On Earth's surface, two types of volcanism are observed. The dominant type, observed at tectonic plate boundaries, manifests the large-scale global circulation in Earth's mantle. In contrast, isolated intraplate volcanoes, such as those of Hawaii, Iceland, or the Galápagos, do not fit with classical plate tectonics theory and are thought to reflect dynamical processes rooted in the deep mantle. The lavas found at these “hotspot” volcanoes provide a singular window into the thermochemical dynamics of Earth's interior. These hotspots usually appear as chains of intraplate volcanoes that may reflect relative movement between plates and the underlying mantle (1). The presence of active, hot, upwelling plumes rising from the core-mantle boundary (CMB) to the bottom of the lithosphere under hotspot volcanoes has been intensely debated (2, 3) since it was

proposed 50 years ago (1). Nonetheless, the idea of hot mantle plumes originating in the deep mantle that sample sources distinct from those that give rise to mid-ocean ridge volcanism reconciles many geophysical and geochemical observations. For example, seismic studies showing an expected thinner mantle transition zone under hotspots (4, 5) and low-velocity columns in tomographic models that extend from the surface to the CMB beneath most hotspots (6) suggest that plumes can indeed extend well into the deep mantle. Geochemically distinct signals can be observed between ocean island basalts (OIBs) at hotspots and mid-ocean ridge basalts (MORBs), reflecting their source regions. MORBs have relatively uniform $^3\text{He}/^4\text{He}$ ratios, $\sim 8 \pm 1$ Ra (atmospheric ratio) (7) (1 σ), whereas OIBs have a much wider range, with ratios up to 43 Ra in Iceland (8) [and up to 50 Ra at the Baffin Island large igneous province (9)]. The high $^3\text{He}/^4\text{He}$ at hotspots might reflect a deep reservoir that preserves ancient Hadean material (10). Hotspots with high $^3\text{He}/^4\text{He}$ are also those with the highest buoyancy flux, a measure of plume strength (11). High $^3\text{He}/^4\text{He}$ signals and high buoyancy flux suggest a deep origin and active

upwellings, which are further confirmed by the correlation between low shear wave velocities (V_s) at 200 km and high $^3\text{He}/^4\text{He}$ signals at hotspots (12). If the low seismic velocities are dominantly thermal, it implies that hotspots with higher $^3\text{He}/^4\text{He}$ anomalies are hotter and sufficiently buoyant to entrain the possibly denser high $^3\text{He}/^4\text{He}$ domain (12, 13) from the deep mantle.

Classical plume theory predicts focused thermal anomalies beneath hotspots. Directly measuring the excess temperature of the subhotspot mantle relative to the mantle upwelling beneath ridges may therefore allow us to constrain the origin and dynamics of plumes that feed them. Previous temperature estimates using petrological thermometers suggest that the subhotspot mantle is typically 100° to 300°C hotter than the subridge mantle (4, 14, 15). This implies an excess temperature (T_{ex}) for hotspots compared with ridges, where T_{ex} is the difference between the potential temperature (T_p)—the temperature a parcel of mantle will have at Earth's surface if extrapolated along an adiabat—beneath an individual hotspot and the average ridge temperature $\overline{T_p}$ over all ridge segments. Dynamically, a T_{ex} of between 100° and 150°C beneath the lithosphere is needed for pure thermal plumes of ~100 km radius to rise fast enough (~10 cm/year) in the upper mantle (16) to avoid cooling too much by diffusion, still create excess melt, and have enough buoyancy to continue rising (17). This estimate provides a dynamical limit for the minimum T_{ex} and is coincidentally the same as the lower bound of typical petrological estimates (i.e., $T_{\text{ex}} = 100^\circ\text{C}$). The question of whether all oceanic hotspots exceed the average ridge temperature and the minimum dynamic limit is hard to answer from petrological thermometers alone, given the limited geographical coverage and inconsistent estimates. Only a subset of hotspots (≤ 28) have petrological temperature estimates, and values vary substantially between studies (4, 14, 15). Estimates of the average temperature at mid-ocean ridges range from 1280° to 1400°C (4, 14, 15, 18, 19).

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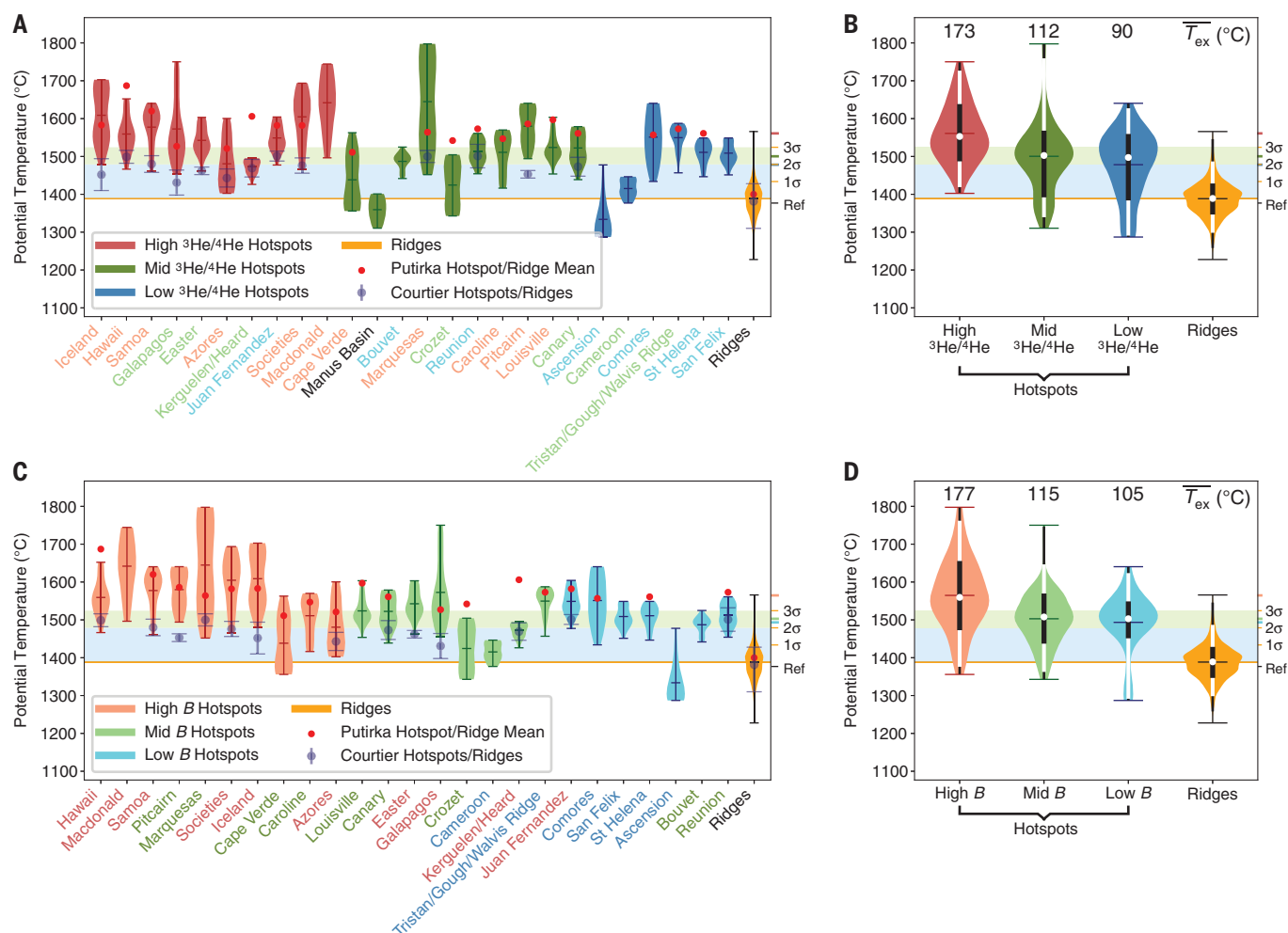


Fig. 3. Potential temperature of hotspots with different $^3\text{He}/^4\text{He}$ ratios or geometrical buoyancy flux (B). Violin plots of the inferred T_p , for plume hotspots sorted by $^3\text{He}/^4\text{He}$ and B . Red, green, and blue ticks on the right are the $\overline{T}_{\text{ex}}$ for the corresponding group. Ticks and shaded regions as in Fig. 2. $\overline{T}_p$ of ridges in the yellow violins. (A) From left to right: dark red, high (>15.7 Ra); dark green, mid; and dark blue, low (≤ 9 Ra) $^3\text{He}/^4\text{He}$. Purple and red circles as in Fig. 2. Hotspot names colored after groups from B. (B) Hotspots stacked by high, mid, and low (red, green, and blue) $^3\text{He}/^4\text{He}$ and compared with

ridges (yellow). Black and white bars as in Fig. 2B. (C) From left to right: light red, high (>0.66 Mg/s); light green, mid; and light blue, low (≤ 0.19 Mg/s) B . Purple and red circles as in (A). Hotspot names colored after groups from $^3\text{He}/^4\text{He}$. (D) As in (B), but stacking B . For both $^3\text{He}/^4\text{He}$ and B , when comparing temperature of the “high” group and the “mid” plus “low” group hotspots (plume hotspots or all hotspots), Welch’s t test gives a P value of ≤ 0.04 . Note that 9 Ra corresponds to 1σ above MORB average (see text for more details).

a self-consistent thermodynamic model of the phase assemblage of the mantle and its physical properties (21). SEMUCB-WM1 (22) uses full waveform inversions and accurate forward wavefield computations and—as is crucial for inferring temperature—has some of the most robust estimates of absolute velocities in the upper mantle (24). SEMUCB-WM1 also shows 26 oceanic hotspots with well-resolved plumes (6) in our catalog. The reference adiabat, extracted velocities for ridges and hotspots, velocity-to-temperature conversions, and inferred temperatures are shown in fig. S1, and our compiled dataset with the temperature inferred from all seismic models (22, 25–28) can be found in data S1.

Whereas Dalton *et al.* (19) found temperature to be the primary driver of MORB major-

element compositional variability and seismic velocity variations, we went a step further and started with the null hypothesis that seismic anomalies in both the subridge and the sub-hotspot mantle are due to temperature alone. Further, we assumed that the major element source of mantle melting beneath hotspots and ridges is identical.

Our reference mantle composition was depleted MORB mantle (DMM) (29) with a reference adiabat of 1377°C (1650 K) (21), consistent with previous studies (4, 19) and MORB formation (30). At each hotspot, and every depth interval, we searched for the local minimum V_s anomaly ($\delta\ln V_s$), which translates to the highest mantle T_p , within a search radius R centered around the hotspot. We set R to 500 km at Earth’s surface to account for the

possible lateral deflection of the plume conduit by the “mantle wind” (Fig. 1B) (21). For ridges, we took the average seismic velocity as representative of the ambient mantle that melts to form MORB. We expect the average temperature to be more representative of their dynamical origin, as ridges are largely the result of passive return flow away from slabs and have no undisputed slow anomalies throughout the entire upper mantle (21). Hence, at each ridge segment, for every depth interval, we averaged all $\delta\ln V_s$ values in a disk of radius R , centered on the ridge segment. By using the local highest temperature for hotspots and the local average temperature beneath each ridge segment, we maximized the inferred T_{ex} between ridges and hotspots (21). Even in this conservative case, optimal for making hotspots

hotter than ridges, we found that >10% of oceanic hotspots [$\sim 15\%$ of plume-fed hotspots (6)] are not resolvably hotter than ridges ($T_{\text{ex}} < 50^\circ\text{C}$). Had we chosen the local highest temperature for ridges, as we did for hotspots, then $\sim 70\%$ of oceanic hotspots ($\sim 80\%$ of plume-fed hotspots) would not be hotter than ridges.

The effects of limited resolution, smoothing, and damping in seismic tomography models broaden and reduce the amplitude of $\delta \ln V_s$, leading to lower inferred temperatures. This may be especially true for the local minimum $\delta \ln V_s$ extracted beneath hotspots compared with the broad average $\delta \ln V_s$ beneath mid-ocean ridges. We attempted to correct for these effects beneath hotspots by scaling up $\delta \ln V_s$ by a constant scaling factor f (21). For SEMUCB-WM1 (22), three-dimensional (3D) synthetic resolution tests (6, 22) suggest a conservative scaling factor of $f = 2$ for plumes that have a core radius [cut off at half peak $\delta \ln V_s$ (21)] of 150 km in the upper mantle. To assess both the resolving power of SEMUCB-WM1 for narrower conduits and the appropriate corresponding f value, we performed two additional 3D synthetic resolution tests (fig. S2) for 100 km radius conduits under Ascension ($T_{\text{ex}} = -10^\circ\text{C}$) and Cameroon ($T_{\text{ex}} = 27^\circ\text{C}$), two cold hotspots. This complements existing resolution tests on Hawaii and Iceland (6). For narrower plumes, f may, depending on resolution, be greater (figs. S2 and S3) (21), but not by enough to alter our main conclusions.

In Fig. 2, we show violin plots (21) of the stacked inferred temperature for all oceanic hotspots and ridge segments at all depths (260 to 600 km). The width of each violin column represents the number density of T_p . The $\overline{T_p}$ for all ridges is $1388^\circ \pm 45^\circ\text{C}$ (1 σ), very close to the 1377°C reference adiabat. The $\overline{T_p}$ of all oceanic hotspots is $1527^\circ \pm 95^\circ\text{C}$, resulting in an average T_{ex} ($\overline{T_{\text{ex}}}$) of 139°C , which spreads over a large range of T_{ex} for individual hotspots. Cluster analysis (Fig. 2) (21) suggests three distinct groups of hotspots: hot ($\overline{T_{\text{ex}}} = 199^\circ\text{C}$), warm ($T_{\text{ex}} = 104^\circ\text{C}$), and cold ($T_{\text{ex}} = -10^\circ\text{C}$). For 26 oceanic hotspots associated with seismically resolved plumes [i.e., plume-fed hotspots as defined in (6), where we expect the inferred T_p to be most reliable, as the seismic anomaly is visible through much of the mantle], the $\overline{T_p}$ ($1519^\circ \pm 93^\circ$ versus $1527^\circ \pm 95^\circ\text{C}$) and $\overline{T_{\text{ex}}}$ (131° versus 139°C) are essentially unchanged compared with all oceanic hotspots. We thus focused only on plume-fed hotspots (Fig. 2A, black outlines).

The T_{ex} of the four hotspots ($\sim 15\%$ of the 26 hotspots considered) that are cold ($-54^\circ \leq T_{\text{ex}} \leq 36^\circ\text{C}$) all fall within 1 σ (45°C) of the ridge $\overline{T_p}$. For the 10 hotspots ($\sim 40\%$) that are warm ($50^\circ \leq T_{\text{ex}} \leq 136^\circ\text{C}$), the T_{ex} falls within the ridge 3 σ (135°C). Only 6 of the 10 warm hot-

spots have sufficiently high T_{ex} to match or exceed the minimum dynamical limit for mantle plumes with a 100 km radius to rise at 10 cm/year (138°C) (table S1) (21). Thus, even warm hotspots are barely hot and buoyant enough to actively upwell. The remaining 12 hotspots have $T_{\text{ex}} \geq 155^\circ\text{C}$, beyond the ridge 3 σ and well above the dynamical limit. Taken at face value, our results suggest that nearly a third ($N = 8$) of plume-fed hotspots are either not resolvably hotter than ridges or not beyond the minimum dynamical limit ($T_{\text{ex}} \leq 100^\circ\text{C}$). The presence of three classes of hotspots, including cold hotspots, is robust and independent of our choice of seismic model (fig. S4) or reference profile (fig. S5). It is of interest that the cutoff between hot and warm hotspots from the cluster analysis (136°C) matches the T_{ex} needed for a 100 km radius plume to rise at 10 cm/year (138°C) (21) (table S1). We use this plume radius and terminal speed in the main text for all calculations, unless otherwise noted.

We found that the $\overline{T_p}$ for ridges ($1388^\circ \pm 45^\circ\text{C}$) is consistent with Courtier *et al.*'s (4) and Putirka's (14) petrological estimates (1381°C and $1400^\circ \pm 35^\circ\text{C}$, respectively) and Dalton *et al.*'s (19) hybrid thermometer ($1385^\circ \pm 40^\circ\text{C}$). Our estimate of the T_{ex} for plume-fed hotspots ($131^\circ \pm 77^\circ\text{C}$) lies between those of Courtier *et al.* (4) and Putirka (14) ($91^\circ \pm 24^\circ\text{C}$ and $177^\circ \pm 57^\circ\text{C}$, respectively) (Fig. 2 and figs. S6 and S7) (21). We note that Courtier *et al.* (4) also found hotspots with T_{ex} as low as 50°C , below the typical lower bound of 100°C of other petrological studies (14, 31) but in agreement with our inferences.

We need to evaluate whether our null compositional hypothesis is valid by allowing for the presence of a substantial fraction of recycled crust in the plume source (32). We expect the addition of recycled crust (eclogite) to increase the inferred T_{ex} because eclogite is seismically faster than DMM (33). Mechanically mixing 10 and 25% normal N-MORB (normal MORB) (29) with DMM in the hotspot mantle source (21) increases the inferred plume-fed hotspot $\overline{T_p}$ by 14°C ($1533^\circ \pm 97^\circ\text{C}$) and 35°C ($1554^\circ \pm 107^\circ\text{C}$), respectively, compared with the eclogite-free results ($\overline{T_p}$ of $1519^\circ \pm 93^\circ\text{C}$). Even after adding 25% eclogite, 8% of plume-fed hotspots are still cold ($T_{\text{ex}} \leq 45^\circ\text{C}$) (fig. S8).

However, the addition of a recycled crustal component comes at a substantial dynamical cost. Any additional eclogitic fraction increases density substantially and requires higher T_{ex} to overcome the negative buoyancy. For example, plumes with 25% recycled crust would need a T_{ex} of 368°C (142°C for 10%) just to be neutrally buoyant in the upper mantle (table S1) (21), but they gain only an additional 35°C from the presence of eclogite in the seismic velocity-to-temperature conversion. A T_{ex} of 368°C is consistent with prior estimates (34)

for the upper temperature bound for eclogite-bearing plumes to remain neutrally buoyant at the bottom of the lithosphere. At such a T_{ex} threshold, all hotspots with 25% added eclogite have insufficient T_{ex} to remain neutrally buoyant, let alone rise actively. At 10% eclogitic fraction, >40% of all plume-fed hotspots remain below the neutral buoyancy threshold (142°C). Therefore, the addition of eclogite cannot explain cold hotspots.

The existence of cold hotspots appears robust both from our results and recent work. A petrological study at Cameroon (35) and a seismological study at Ascension (36) show lower T_{ex} ($\sim 50^\circ\text{C}$ or less; T_p of 1400° to 1430°C), in agreement with our inferences for these hotspots. Low petrological T_{ex} values have been found for Juan Fernández (37) and Pitcairn (Gambier Island) (38), although these appear as hot hotspots in our estimates. This opens the possibility that, in some cases, we may have overestimated T_{ex} for some plumes, which would strengthen our conclusions.

Low T_{ex} values would seem to make it hard for plumes to rise rapidly enough without losing buoyancy. A minimum T_{ex} of $\sim 135^\circ\text{C}$ is needed for pure DMM plumes with a 100 km radius to rise at ~ 10 cm/year (16) in the upper mantle (21), consistent with the dynamical limit (17) and simulations (39). Plumes of the same radius containing eclogite require much higher T_{ex} to rise at the same velocity ($>300^\circ\text{C}$ for 10% fraction) (table S1). Envisioning cold, or even warm, hotspots as the same dynamical entities as hot hotspots is difficult.

One hypothesis proposed by Courtillot *et al.* (40) suggests that cold and warm hotspots are fed by passive upwellings or have a shallow source (3). A shallow source for pure DMM hotspots may be possible if they are generated by edge-driven or small-scale convection (41) or by sublithospheric drainage (42). Alternatively, cold and warm hotspots may still be fed by deep plumes that become trapped and cooled by small-scale convection in the upper mantle (43). In either case, these plumes will have small or even negligible T_{ex} in the deeper upper mantle. This could be the case for some cold hotspots without clear age progressions, such as Cameroon (42) ($T_{\text{ex}} = 27^\circ\text{C}$) near the West African passive margin. Or these plumes may be weak ($T_{\text{ex}} \sim 50^\circ\text{C}$), as suggested by geophysical (36) and geochemical (44) studies beneath Ascension. For narrower plumes (<50 km core radius), it may be that global tomography is unable to resolve them in the upper mantle. This may be the case for Cape Verde (45), which has a relatively high $^3\text{He}/^4\text{He}$ signal [15.7 Ra (46)] but low inferred T_{ex} in our model and is well resolved in the lower mantle (6). But it is not the case for Ascension and Cameroon, two cold plumes that are resolved even at core radii of 50 km (fig. S2) (21). A core radius of <35 km would be required for the hottest cold

plume ($T_{\text{ex}} = 36^\circ\text{C}$) to equal the $\overline{T_{\text{ex}}}$ for hot hotspots (199°C), as our synthetic resolution tests suggest (21). Narrower plumes, however, would need to rise faster, and thus be hotter than the minimum dynamical limit, so as not to lose their buoyancy and be buffeted by the mantle wind.

For hotspots with a small fraction of recycled oceanic crust (e.g., ~5%), a T_{ex} of $\sim 50^\circ\text{C}$ is enough to overcome the excess density of eclogite. Such low T_{ex} is enough to allow a thermochemical plume to rise with the help of broad passive upward return flow that occurs above large low shear wave velocity provinces (LLSVPs), a return flow that complements downward flow at global subduction zones (21). Broad passive flow can contribute half [~ 1.5 to 3 cm/year (47)] of the velocity needed by low T_{ex} plumes to rise fast enough to keep their buoyancy (21), the other half arising from their T_{ex} . We cannot rule out the possibility that the cold hotspots are wet, which lowers the melting temperature at shallow depths and leads to decompression melting (48). Whatever the explanation, these hotspots do not fit the classical plume model.

On the other hand, hotspots with the highest T_{ex} do fit the classical model, as they are more than hot enough to rise actively. These hot plumes are also associated with many of the largest buoyancy fluxes and the highest $^3\text{He}/^4\text{He}$. Figure 3 shows a series of violin plots of the stacked T_p of plume-fed hotspots, colored and sorted by $^3\text{He}/^4\text{He}$ (12) (Fig. 3, A and B) or geometrical buoyancy flux B (5, 21) (Fig. 3, C and D). We use $^3\text{He}/^4\text{He} \leq 9$ Ra and $B \leq 0.19$ Mg/s as the threshold criterion for hotspots with low flux and He ratios, where 9 Ra is the 1σ deviation from the MORB mean (8 Ra) (7), and both correspond to the lowest 30th percentile. We further divide the remaining hotspots into high ($^3\text{He}/^4\text{He} > 15.7$ Ra; $B > 0.66$ Mg/s), the top 70th percentile, and intermediate ($9 < ^3\text{He}/^4\text{He} \leq 15.7$ Ra; $0.19 < B \leq 0.66$ Mg/s) categories. We find that the $\overline{T_{\text{ex}}}$ of hotspots reduces as a function of decreasing $^3\text{He}/^4\text{He}$ or B (Fig. 3 and figs. S9 and S10). These results directly confirm the relationships between higher plume temperature and extreme $^3\text{He}/^4\text{He}$ and buoyancy flux proposed by Jackson *et al.* (12) using seismic velocity at 200 km depth, and by Putirka (14) using petrologically derived temperatures.

For Iceland and Hawaii, the two hotspots with the highest $^3\text{He}/^4\text{He}$, we find a T_p of 1609°C ($T_{\text{ex}} = 221^\circ\text{C}$) and 1559°C ($T_{\text{ex}} = 171^\circ\text{C}$), respectively, firmly in the hot hotspot cluster (Fig. 2). Iceland's higher T_{ex} compared with Hawaii is more compatible with its higher $^3\text{He}/^4\text{He}$ signal [up to 47.5 Ra (49) versus 35.3 Ra (50)] and estimates of its buoyancy flux (21, 51) but is not compatible with the greater lithospheric thickness beneath Hawaii (48).

If high $^3\text{He}/^4\text{He}$ domains are associated with primordial and denser material, it may

be that only the more buoyant (hotter) plumes can entrain it and rise to the surface (12). High $^3\text{He}/^4\text{He}$ in OIBs is geographically correlated to the two LLSVPs (Fig. 1A) in the lowermost mantle (10). LLSVPs may be denser and chemically distinct (52), compatible with a reservoir of dense oceanic crust or primordial material with high $^3\text{He}/^4\text{He}$ (53). Our results suggest that hot hotspots are indeed thermochemical in nature but are hot and buoyant enough to entrain LLSVP material with high $^3\text{He}/^4\text{He}$.

We find a $\overline{T_{\text{ex}}}$ of 205°C for five hotspots overlying large ultra low velocity zones (“mega-ULVZs”) at the CMB: Iceland, Hawaii, Samoa, and Marquesas (54) as well as Galápagos (55) (fig. S11A). These hotspots also have the highest $^3\text{He}/^4\text{He}$, with the exception of Marquesas (54), which has a moderately-high $^3\text{He}/^4\text{He}$ but high B . If these mega-ULVZs are broad regions of partial melt they may provide roots for some hot, strong plumes (54). They may also represent a core-mantle interaction zone, which may be the source of ancient isotopic anomalies (56) in hotspot lavas. Two other classes of plumes appear to be cooler and less thermally buoyant and give rise to low-flux hotspots. They do not appear to entrain deep-seated primordial domains—possibly because they are too cold and thus insufficiently buoyant to entrain a deep and dense high $^3\text{He}/^4\text{He}$ domain (12)—but they may provide important clues to shallower mantle processes such as slow-rising plumes and small-scale convection.

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SUPPLEMENTARY MATERIALS

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Materials and Methods

Figs. S1 to S11

Table S1

References (57–89)

Data S1

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