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Plumes from the heterogeneous Earth's mantle

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The spectrum of geochemical compositions of Oceanic Island Basalts (OIBs) and their systematic differences from Mid-Ocean Ridge Basalts (MORBs) reveal that the Earth's mantle is chemically and isotopically heterogeneous. Two main processes, both related to plate tectonics, contribute to the creation of mantle heterogeneities: (1) partial melting generates melts enriched in incompatible elements and leaves a depleted residual rock; and (2) subduction of the oceanic lithosphere injects heterogeneous material at depth, in particular, altered oceanic crust and continental/oceanic sediments. Moreover, delamination and foundering of metasomatized subcontinental lithospheric mantle might have been important in the early Earth history, when plate tectonics did not operate as today. The fate of the subducted plate is still a matter of debate; presumably some of it is stirred by convection and some may segregate at the base of the mantle, in particular the oceanic crust, which is compositionally denser than the pyrolitic mantle. The view of the lower mantle as a "graveyard" of subducted crust prevailed for decades and was supported by the Hofmann and White (1982) observation that the geochemical fingerprint of most OIB reveals the presence of ancient recycled crust. However, recent geochemical data on short-lived systems (*e.g.* $^{182}Hf \rightarrow {}^{182}W$ has a half-life of 8.9 My) showed that some hotspots, namely Hawaii, Samoa, Iceland and Galápagos, have a negative $\mu^{182}W$ anomaly. This discovery prompted a change in our view of the deep mantle because anomalies in short-lived systems require additional processes, which include, but are not limited to, the preservation of 'pockets' of melt from a primordial magma ocean, and/or chemical reactions between the metallic core and the silicate mantle. Exchanges at the core-mantle boundary would cause a negative $\mu^{182}W$ anomaly, and might also add ^3He to mantle material later entrained by plumes. It is now clear that some plumes probe the deepest mantle and are highly heterogeneous, as revealed by isotope ratios from long-lived radiogenic systems, noble gases and short-lived isotope systems. Here I will focus on the dynamics of plumes carrying compositional and rheological heterogeneities. This contribution attempts to be pedagogic and multi-disciplinary, spanning from seismology to geochemistry and geodynamics.

1. Hotspots and mantle plumes

Hotspots are regions of intraplate volcanism, or of excessive volcanism along a portion of a spreading ridge (Koppers *et al.*, 2021). Their long-lasting magmatic activity can form an age-progressive volcanic chain, the best example being the Hawaiian hotspot: volcanoes younger than 1 Ma are still active (*i.e.* Kilauea and Loihi), whereas older, extinct, volcanoes form the 6000 km-long Hawaiian–Emperor chain (Fig. 1). The age of the oldest volcano (80 Ma) attests to the long-lived activity of the hotspot, whereas any information about previous volcanism is missing since the Pacific plate subducts at the

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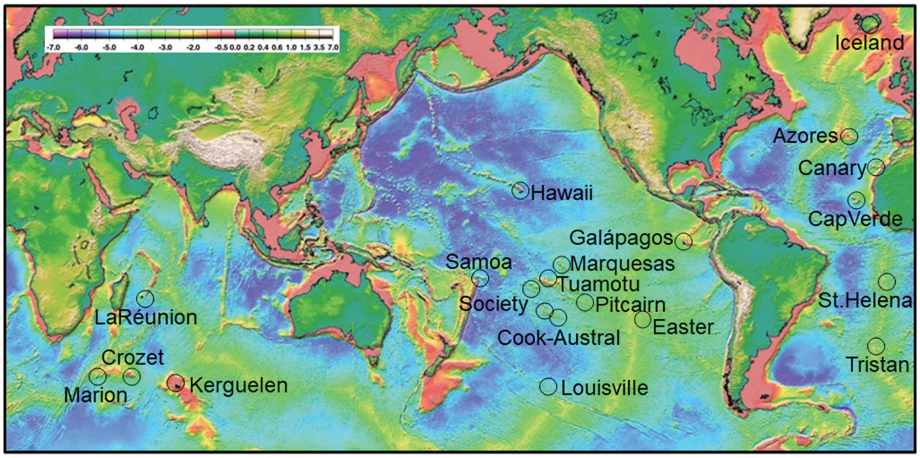

Fig. 1. Global map of sea-floor depth (Smith and Sandwell, 1997) and continental elevation. Only some hotspots are indicated; for a map of all hotspots see Ito and van Keken (2007).

Aleutians subduction zone. Ito and van Keken (2007) indicated that hotspots younger than 100 Ma are generally active, although their vigour may vary considerably, whereas older hotspots are either waning or inactive.

Morgan (1971) first suggested that hotspots are the surface expression of plumes rising from the deep mantle. Ever since, the existence of plumes and their depth of origin have been hotly debated. Courtillot *et al*. (2003) proposed five criteria to assess whether a plume has a deep origin: (1) a hotspot fed by deep, relatively fixed, plume should form an age progressive volcanic chain the linear trend of which is consistent with the direction of plate motion. According to Ito and van Keken (2007) only 13 long-lived (> 50 Ma) and eight short-lived (<20 Ma) volcanic chains are age-progressive, whereas all other volcanic chains lack a clear age progression. (2) The onset of hotspot magmatism should be marked by the emplacement of a Large Igneous Province (LIP), a term including continental flood basalts and oceanic plateaus, formed in continental and oceanic settings, respectively. Note that both criteria are associated with the plume shape constrained by pioneering laboratory experiments (Whitehead and Luther, 1975), namely a large "head", the partial melting of which generates a LIP, and a narrow, long-lasting, plume conduit or "tail", the melting of which generates a volcanic chain (Richards *et al*., 1989). (3) A deep plume is expected to have a large buoyancy flux. This parameter is evaluated from the topographic swell of a hotspot. The swell is a region with an anomalously high topography (swell heights range from 500 to 1200 m) over lateral widths of 1000–1500 km, in the direction perpendicular to the volcanic chain (Crough, 1983; Wessel, 1993). Sleep (1990) estimated the buoyancy flux of Hawaii, $B_{Hawaii} = 8.7 \times 10^3$ kg s^{-1} and the total buoyancy flux of all hotspots $B_{Total} = 55 \times 10^3$ kg s^{-1}. Recent estimates propose $B_{Total} = 22 \times 10^3$ kg s^{-1} (King and Adam, 2014) and $B_{Total} = 46 \times 10^3$ kg s^{-1} (Hoggard *et al*., 2020). The last two criteria, proposed by Courtillot *et al*. (2003), rely on geochemical observations, namely large $^{3}He/^{4}He$ for hotspot lavas fed by a deep mantle plume, and on seismological observations, namely a low shear wave velocity zone in the underlying mantle. Today there is a general agreement that 18 mantle plumes originate from the deepest regions of the Earth's mantle (Koppers *et al*., 2021).

One important parameter not mentioned above is the plume excess temperature with respect to the surrounding mantle. Following Putirka (2005) and Bao *et al*. (2022), the plume excess temperature for the most vigorous hotspots ranges between 100 and 250°C. This excess temperature enables plumes to melt even below a thick lithosphere (*e.g.* Hawaii) or to generate excess magmatism along some parts of a spreading ridge (*e.g.* Iceland). Moreover, the excess temperature provides the buoyancy force necessary to ascend from the deep mantle. Last but not least, the excess temperature poses a first-order constraint on the location of the root zone of plumes (Stacey and Loper, 1983), likely to be the core–mantle boundary region, at 2700–2900 km depth, where the estimated temperature increment is 800–1000°C (Jeanloz and Morris, 1986).

2. Plumes and the Earth's mantle structure: seismic observations

Ever since the plume concept was proposed, the questions addressed to seismologists have been: "Can seismic tomography detect mantle plumes?" "Is there a deep seismic velocity anomaly beneath hotspots?"

Because of the uneven distribution of earthquakes and of seismic stations, and because of wavefront healing effects (Nolet and Dahlen, 2000) it is difficult to resolve narrow, low-velocity structures (Goes *et al*., 2004). Using finite frequency tomography, Montelli *et al*. (2006) showed that zones with a negative shear-wave velocity anomaly were broader (*i.e.* 600–800 km diameter) than the plume conduits hypothesized by Morgan (1971). Moreover, conduits had a variable depth extent and sometimes lacked a clear depth continuity. This is in agreement with resolution analysis, indicating that plumes would not be visible if their diameter was $<\sim$600 km (Montelli *et al*., 2006; French and Romanowicz, 2015). By using travel times of seismic core waves recorded by the dense USArray seismic network in North America, Nelson and Grand (2018) were able to detect a narrow, cylindrically shaped, slow-velocity anomaly, ~350 km in diameter, interpreted as the deep mantle plume beneath the Yellowstone hotspot.

Global tomography models have shown the existence of two low-velocity structures in the lower mantle below Africa and the central Pacific. The Large Low Shear Velocity Provinces (LLSVP) observed in both shear- and compressional-wave tomography, cover 25–30% of the CMB surface and rise 500–1000 km above it (Ritsema *et al*., 1999; Masters *et al*., 2000; Romanowicz and Gung, 2002). LLSVPs are frequently associated with mantle plumes beneath active hotspots (French and Romanowicz, 2015). Moreover, the reconstructed LIP position, obtained by calculating the plate motion back in time to the location where the LIP erupted, often corresponds to the edges of the LLSVPs (Burke and Torsvik, 2004). These reconstructions, together with other geophysical observations (*e.g.* Dziewonski *et al*., 2010), suggest that the LLSVPs might have been stable during the last 250 My. LLSVPs are probably hotter than the surrounding mantle, but there is still a debate about their nature, either purely thermal or thermo-chemical. To better constrain their nature, Su and Dziewonski (1997) looked at relative perturbations in the shear velocity ($V_s = (\mu/\rho)^{1/2}$, where μ is the modulus of rigidity, ρ is density) and in bulk sound velocity ($V_\Phi = (K/\rho)^{1/2}$, where K is the bulk modulus) and found that in the lowermost mantle the two are negatively correlated. Interestingly, thermal variations alone do not generate anti-correlation; therefore, both Su and Dziewonski (1997) and Masters *et al.* (2000) suggested that compositional anomalies are required to explain seismic observations. Although the chemical composition of LLSVPs is unknown, the hypothesis of ancient subducted oceanic crust was proposed decades ago. This is not surprising because the subducted crust probably represents 10–15% the mass of the mantle (Stracke *et al.*, 2003) and because the oceanic crust remains compositionally denser even in the lowermost mantle (Ricolleau *et al.*, 2010; Stixrude and Lithgow-Bertelloni, 2012). Geodynamic models (Christensen and Hofmann, 1996; Brandenburg *et al.*, 2008) confirmed that dense recycled crust piles up at the base of the mantle, forming LLSVP-like structures. However, Deschamps *et al.* (2012) proposed that LLSVPs have a distinct, more primitive, composition enriched in iron by 3.0% and in (Mg,Fe)-perovskite by 20%, compared to regular mantle. The last ten years have seen the bloom of thermo-chemical geodynamic models, combining recycled crust and relatively primitive material (Li *et al.*, 2014; Nakagawa and Tackley, 2014; Tucker *et al.*, 2020; Jones at al., 2021). The results show that the LLSVPs may indeed be compositionally heterogeneous and that plumes may be fed by LLSVP material (see Fig. 2 for a schematic representation). However, even the latest geodynamic models do not capture the complex shape observed by the most recent tomographic studies, *e.g.* Tsekhmistrenko *et al.* (2021) found that mantle upwellings of the African LLSVP are arranged in a tree-like structure: from a central, compact

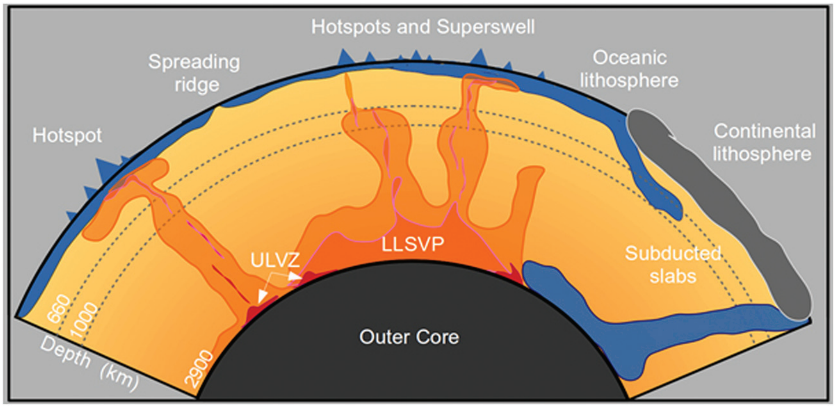

Fig. 2. Cartoon showing, in a simplistic way, various plume morphologies. For the LLSVP and the ULVZ the thickness is exaggerated for graphical reasons. Complexities such as plume–ridge interactions are not shown. The fate of subducted lithosphere is illustrated schematically. Refer to Fukao and Obayashi (2013) for tomographic images of slabs and to French and Romanowicz (2015) for tomographic images of LLSVP.

trunk below ∼1500 km depth, three branches tilt outwards and up towards various Indo-Austral hotspots. Tomographic images by Wamba *et al*. (2023) of the African LLSVP in the Indian ocean, form the Comores, La Réunion and Crozet hotspots and indicate the presence of slow-velocity anomalies in the lower mantle, forming broad conduit-like structures with a diameter of ∼900 km at 1500 km depth. At 1000 km depth the vertical continuity is disrupted and horizontal spreading seems to prevail between 1000 and 660 km depth. For the Pacific LLSVP, French and Romanowicz (2015) and Davaille and Romanowicz (2020) showed the presence of well-separated, low-velocity conduits that extend vertically throughout most of the lower mantle.

Seismologists have also discovered the Ultra Low Velocity Zones (ULVZ), patches on the core-mantle boundary, that exhibit reductions of S-wave velocity by as much as 30% and of P-wave velocity by up to 10% (*e.g.* Garnero and Helmberger, 1996; Rost and Revenaugh, 2003; Thorne and Garnero, 2004). ULVZs are 10–40 km thick, have a lateral extent of hundreds of km, and up to 1000 km for the Mega-ULVZs (Cottaar and Romanowicz, 2012; Kim *et al*., 2020). Their density increase may be up to ∼10% (Rost *et al*., 2005). Several mechanisms have been proposed to explain their high density; just two are mentioned here: core–mantle reaction; and the presence of dense silicate melts or of iron-enriched oxides (see the review by McNamara (2019) and references therein).

3. Plumes and the Earth's mantle structure: geochemical observations

Oceanic island basalts have a different chemical composition in both trace and major elements with respect to Mid-Ocean-Basalts (Schilling, 1973). These differences are due, in part, to a lower degree of melting in plumes than at spreading ridges; however, the isotopic differences between OIBs and MORBs suggest an isotopically distinct source (Hart *et al*., 1973; Hofmann and White, 1982; Hofmann, 2003; White, 2010, 2015b). The present study will focus on key isotopic ratios, starting with long-lived isotopes and moving on to short-lived isotopes.

3.1. The message from the long-lived radioactive decay systems

Isotopic systems with long-lived radioactive parent elements (*i.e.* half lives of billions of years) shed light on mantle processes that occurred over long time-scales, *e.g.*: (1) partial melting and melt extraction, leading to a residual mantle depleted in incompatible elements; (2) the generation of the crust, enriched in incompatible elements; (c) recycling of the oceanic crust and of portions of continental crust, in the form of marine sediments and/or delaminated lower crust; and (4) dynamic instabilities of the continental lithosphere (Cottrell *et al*., 2004; Fourel *et al*., 2013) leading to foundering of the sub-continental lithospheric mantle. All of these compositionally and isotopically distinct materials, possibly with distinct density and viscosity, are variably stirred by mantle convection and/or partially segregate in the deep mantle over long time-scales (1–2 Gy), before being entrained in mantle plumes.

Since the studies of Zindler *et al*. (1982) and Hart *et al*. (1992), radiogenic isotope ratios such as $^{143}Nd/^{144}Nd$, $^{87}Sr/^{86}Sr$, $^{206}Pb/^{204}Pb$ have been the 'backbone' of the definition of mantle end-member isotopic compositions (*i.e.* the Depleted MORB Mantle (DMM), the Enriched Mantle (EM-1 and EM-2), the High-μ (HIMU), where μ is the $^{238}U/^{204}Pb$ ratio). Clearly, these end-members should not be taken as physical entities that actually exist in the mantle. For a thorough discussion, which goes beyond the classical isotopic variability in 2D isotope ratio, see Stracke (2021) and Stracke *et al*. (2022). Here I will follow a more traditional, albeit simplistic, approach. In Fig. 3 is a pedagogic way to understand how, over time, the residual mantle and the crust develop distinct $^{143}Nd/^{144}Nd$ and $^{87}Sr/^{86}Sr$ ratios. Two aspects are noteworthy: first, strontium isotope ratios are negatively correlated with neodymium. Second, there is a complementarity between the residual mantle and the crust. Figure 3c shows that partial melting does not modify an isotopic ratio; this is important because, by measuring the isotopic ratios in basalts we obtain information about the source rock, assumed to be isotopically homogeneous.

MORB and OIB are isotopically variable, and the isotopic variation is non-random. For example, OIBs are systematically more radiogenic in strontium and less radiogenic in neodymium than MORBs, and they extend to more extreme values of $^{206}Pb/^{204}Pb$. OIBs do not have the isotopic composition of a pure end-member but form arrays between end-members. Detailed discussions of geochemical end-members can be found in many articles (Zindler *et al*., 1982; Hart *et al*., 1992; Hofmann, 1997; Stracke *et al*., 2005; White 2015a, 2015b, and references therein); here, just a brief description is provided.

The HIMU end-member has a very radiogenic Pb isotopic composition (*i.e.* high $^{206}Pb/^{204}Pb$) which requires a mantle source with high U/Pb for time periods of the

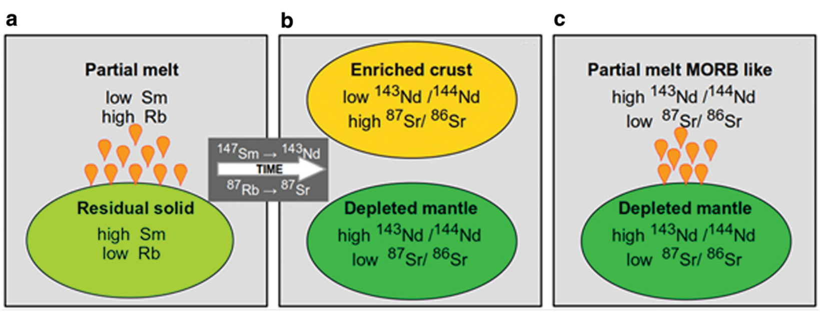

Fig. 3. (***a***) Partial melting in geologic times. For the system $^{147}Sm \rightarrow {}^{143}Nd$, the parent element, samarium, is more compatible than neodymium. Thus, the residual solid has a high Sm (*i.e.* high Sm/Nd ratio) while the liquid magma has a low Sm. For the system $^{87}Rb \rightarrow {}^{87}Sr$, the parent element, rubidium, is less compatible than strontium. Thus, the residual solid has a low Rb (*i.e.* low Rb/Sr ratio), while the liquid magma has a high Rb. (***b***) Over time, the parent element decays into the daughter element (^{147}Sm has a half-life of 106 Gy; ^{87}Sr has a half-life of 48.8 Gy). The residual solid, assimilated to present day depleted mantle, has a high $^{143}Nd/^{144}Nd$ and low $^{87}Sr/^{86}Sr$. The crust, formed from the crystallization of magmas, has low $^{143}Nd/^{144}Nd$ and high $^{87}Sr/^{86}Sr$. (***c***) Partial melting does not change an isotopic ratio.

order of ~1.5–2 Gy since the parent isotope ^{238}U has a long half-life (4.469 Gy). Only a few hotspots have a strong HIMU component, namely St. Helena in the Atlantic, the Cook-Austral islands (Polynesia) in the South Pacific, and some of the Galápagos islands (Harpp and Weis, 2020). However, Homrighausen *et al*. (2018) proposed that the HIMU end-member has a more global distribution than previously thought. What is the origin of the HIMU component? The classical view is that HIMU reflects the presence of recycled oceanic crust enriched in continental ^{238}U carried by sediments; alternatively, it might reflect the presence of recycled basaltic crust that lost fluid-mobile trace elements, such as Pb, because of hydrothermal alteration at ridges or dehydration during subduction (Chauvel *et al*., 1992; Hofmann, 1997). Anomalous sulphur isotope signatures, indicating mass-independent fractionation (MIF), were found by Cabral *et al*. (2013) in sulfides from Mangaia's OIBs (Cook Islands). This is important, because MIF processes require photo-chemical reactions that occurred during the Archaean, when the atmosphere was oxygen-poor and relatively transparent to solar ultraviolet radiation. Cabral *et al*. (2013) proposed that the source of Mangaia's lavas carries ancient (>2.45 Gy old), hydrothermally altered oceanic crust. There is also a long-standing debate about the role of metasomatized lithosphere in HIMU basalts. Weiss *et al*. (2016) pointed out that olivine phenocrysts in HIMU lavas do not support melting of recycled crust but indicate melting of peridotite metasomatized by carbonatite fluids. The metasomatism probably occurred during the Archaean and affected subcontinental lithospheric mantle (Weiss *et al*., 2016; Homrighausen *et al*., 2018). According to this model, ancient, metasomatized subcontinental lithospheric mantle delaminated and cycled in the deep mantle before being entrained in plumes.

The enriched EM-1 end-member is present in hotspots such as Kerguelen, Tristan, Hawaii and Pitcairn (Eisele *et al*., 2003; Weis *et al*., 2011 and references therein). This component indicates the contribution of recycled ancient pelagic sediments and/or of recycled delaminated subcontinental lithosphere. The enriched EM-2 end-member is observed in hotspots such as Society, Marquesas, Samoa (White and Hofmann, 1982; Jackson *et al*., 2007; Chauvel *et al*., 2012 and references therein) and is associated with a small fraction (~2%) of subducted terrigenous-continental sediments. Globally, the enriched components (*i.e.* high $^{87}Sr/^{86}Sr$, low $^{143}Nd/^{144}Nd$, unradiogenic Pb isotopic ratios) indicate that plumes carry subducted material, either continental sediments or oceanic crust, as first suggested by Hofmann and White (1982).

3.2. The message from the short-lived radioactive decay systems

During the last ten years, geochemists were able to define heterogeneities by measuring isotopic ratios (*e.g.* $^{182}W/^{184}W$, $^{129}Xe/^{136}Xe$, $^{142}Nd/^{144}Nd$) from short-lived radioactive parent elements (Touboul *et al*., 2012; Horan *et al*., 2018; Rizo *et al*., 2019). The parent element has such a short half-life that it became extinct soon after the formation of the solar system. In other words, only fractionation processes that occurred very early in the Earth's history could form these isotopic anomalies. The hafnium-tungsten systematics offer a clear example: ^{182}Hf has a half-life of 8.9 My, so that after 50 My no more daughter isotope ^{182}W could be produced. Today the observed $\mu^{182}W$ (*i.e.* ppm deviation relative terrestrial standard) in MORBs and most OIBs is close to zero, but hotspots such as Hawaii, Samoa, Iceland and Galápagos have negative $\mu^{182}W$, often associated with high $^3He/^4He$ (Mundl *et al.*, 2017; Jackson *et al.*, 2020; Mundl-Petermeier *et al.*, 2020; Peters *et al.*, 2021).

Short-lived isotope systems prompt us to consider processes that occurred very early in the Earth's history. The physical conditions that prevailed during the first 50 My of the Earth's life are still debated, but there is a general consensus that this period was highly energetic. Impacts of planetesimals and embryos generated enormous amounts of heat and the giant Moon-forming impact probably melted the entire Earth, forming a global silicate magma ocean (Stevenson, 1987; Canup, 2008; Elkins-Tanton, 2012; Bolrão *et al.*, 2021 and references therein). The possibility that heterogeneous domains could form during crystallization of the magma ocean is supported by first-principles molecular dynamics (Deng and Stixrude, 2021) showing that Hf is more compatible than W (see Fig. 4a). Even though it remains unclear if the solidification of a magma ocean proceeded "bottom up" (Andrault *et al.*, 2011) or from mid-depth (Stixrude *et al.*, 2009; Boukaré *et al.*, 2015), the residual liquids of a magma ocean are progressively enriched in incompatible elements; therefore, they are expected to have a high iron content and a sub-chondritic Hf/W (Brown *et al.*, 2014). This 'Enriched Magma-Ocean Residue', if preserved from 4 Gy of convective stirring, would still have a slightly negative $\mu^{182}W$ (see Fig. 4c, magenta colour).

Alternatively, the metallic core could be a reservoir with a strongly negative $\mu^{182}W$ (Touboul *et al.*, 2012; Rizo *et al.*, 2019), as W is a moderately siderophile element, whereas Hf is lithophile (see Fig. 4b,c). A key issue with this model is how to transfer a 'core signature' to the lowermost mantle. High-pressure experiments by Yoshino *et al.* (2020) suggest that is possible to transfer W from the core to the mantle because grain boundary diffusion of W is a fast and strongly temperature-dependent process. This supports the idea that a fraction of a lowermost mantle might be thermally and chemically equilibrated with the core. Mundl-Petermeier *et al.* (2020) proposed the existence of a "Core-Mantle equilibrated reservoir", with a core-like negative $\mu^{182}W$, and show that

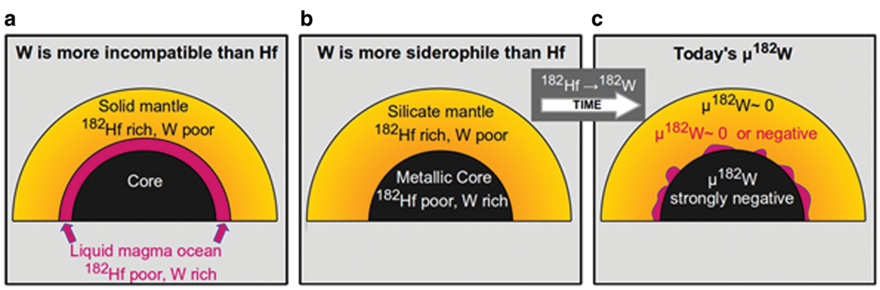

Fig. 4. (***a–b***) At the time when the parent element ^{182}Hf was extant. (***a***) W is more incompatible than Hf, thus a residual liquid from a magma ocean is expected to be W rich and Hf poor. (***b***) W is more siderophile than Hf, thus the metallic core is W rich, whereas the silicate mantle is Hf rich. (***c***) By definition, $\mu^{182}W$ indicates the deviation of $^{182}W/^{182}W$ of the sample from the $^{182}W/^{182}W$ of a standard. The mantle has $\mu^{182}W \approx 0$, relics of the magma ocean are expected to have a slightly negative $\mu^{182}W$ and the core has a strongly negative $\mu^{182}W$.

only 0.3% of its entrainment in plumes would be sufficient to explain the most negative μ^{182}W observed in OIBs. Although highly speculative, it is tempting to associate a "Core-Mantle equilibrated reservoir" with the ULVZs, now detected beneath several hotspots.

4. Fluid dynamics of mantle plumes

Our view of mantle plumes has evolved considerably over time, moving from the 'classical' thermal plume in Newtonian rheology to the more complex thermo-chemical plumes.

4.1. Purely thermal plumes

Purely thermal plumes have one source of buoyancy, namely, their excess temperature. The morphology of a thermal plume in a Newtonian fluid depends on the viscosity contrast between the hot plume and the colder ambient mantle. For strongly temperature-dependent viscosity, the plume has a 'mushroom-shape' with a large head and a narrow tail, whereas for constant viscosity the plume has a 'spout' morphology, with a roughly constant diameter (Whitehead and Luther, 1975). The upwelling flow within the conduit is controlled by the viscosity contrast between the hot axial part and the colder periphery: the vertical velocity (Vz) is maximum at the plume axis and decreases exponentially with the square of the radial distance (Olson *et al.*, 1993). This velocity profile (Fig. 5) has two profound implications. First, Vz decreases with radial distance more rapidly than the excess temperature ΔT. Material with a low ΔT, either surrounding mantle conductively heated by the plume, or the most peripheral parts of the plume, have such a negligible upwelling velocity that they contribute very little to the plume buoyancy flux (Fig. 5) and may never reach the melting zone in the plume head. Thus, the key parameter to define entrained material should not be the excess temperature (*e.g.* Hauri *et al.*, 1994, used a ΔT value which is only 1% of the axial ΔT) but the upwelling velocity. Second, this velocity profile generates zones with high strain rates within the conduit so that passive (*i.e.* not affecting the flow) geochemical heterogeneities are stretched readily into filaments (Fig. 6) as they ascend in the plume tail (Farnetani and Hofmann, 2009). The existence of geochemically distinct 'streaks' in the Hawaiian plume was first proposed by Abouchami *et al.* (2005) in order to explain the spatio-temporal geochemical variability of volcanoes belonging to the Kea-trend. Numerical simulations later showed that a filament crosses the Hawaiian melting zone over timescales of the order of 1 My (Farnetani and Hofmann, 2010) so that two volcanoes can indeed sample the same heterogeneity. Obviously this model is very simplified, as it ignores mixing of partial melts *en route* to the surface and/or ponding of melts in a magma chamber. However, it shows that the laminar flow within the plume conduit does not induce toroidal stirring.

4.2. Thermo-chemical plumes

Thermo-chemical plumes have two sources of buoyancy: thermal and compositional. The nature of compositionally denser material is still elusive. Ancient recycled eclogitic

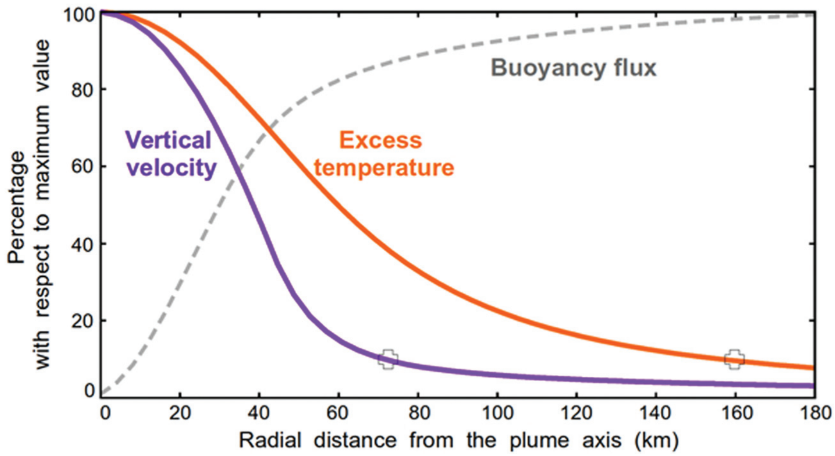

Fig. 5. Profiles of the vertical velocity, Vz, of the excess temperature ΔT, and of the buoyancy flux across the plume conduit. Both Vz_{max} and ΔT_{max} occur at the plume axis. Because of temperature-dependent viscosity, Vz decreases more rapidly than ΔT with radial distance. The open crosses indicate that Vz is 10% of Vz_{max} at 73 km distance, whereas ΔT is 10% of ΔT_{max} at 160 km distance. This implies that low excess temperature material (either from the plume periphery or entrained mantle conductively heated) cannot upwell efficiently and will probably never reach the melting zone. Note also that at the radial distance where Vz is 10% of Vz_{max}, the buoyancy flux $B \approx 87\% B_{max}$ meaning that the low-velocity plume periphery contributes very little to the total buoyancy flux. The buoyancy flux $B = \int \rho\ \alpha\ \Delta T\ Vz\ dS$, where the integral is over the horizontal surface S of the conduit, ρ is density and α the thermal expansion coefficient.

crust is denser than the pyrolitic mantle (Hirose *et al.*, 1999; Ricolleau *et al.*, 2010) while deep-seated portions of relatively primitive mantle might be iron-enriched relative to the depleted surrounding mantle. The effect of compositional density is important, *e.g.* Dannenberg and Sobolev (2015) estimated that a plume with an excess temperature of 250°C can entrain ~20% of dense eclogite before becoming neutrally buoyant. For thermo-chemical plumes the subtle balance between positive thermal buoyancy and negative compositional buoyancy can induce an oscillatory behavior (Davaille, 1999) and a complex internal dynamics, where parts of the conduit sink while others well up (Kumagai *et al.*, 2008). Moreover, thermo-chemical plumes have irregular and non-symmetric morphologies (Tackley, 1998; Farnetani and Samuel, 2005; Nakagawa and Tackley, 2014; Li *et al.*, 2014; Limare *et al.*, 2019). Another important difference is that in a purely thermal mantle a large-scale isotopic zonation across the source region translates into an isotopic zonation of the plume conduit (Farnetani *et al.*, 2012), whereas this might not be true in a thermo-chemical mantle. As pointed out by Jones *et al.* (2016) compositionally denser material rises preferentially at the plume axis, thereby disrupting any 'geochemical mapping' between the conduit and the source. As mentioned above we also do not know to which extent partial melts mix within a complex melt drainage system. In any case, the geochemical zonation of many hotspots is an observational fact, first recognized at Hawaii, where volcanoes younger than 5 Ma form two parallel and geochemically distinct chains. The difference between lavas from Kea- and Loa-trend volcanoes, clearly expressed in radiogenic lead isotope ratios, has been explained by a bilateral zonation (Abouchami *et al*., 2005) of the Hawaiian plume. According to Weis *et al*. (2011) the southern Loa-side of the plume entrains geochemically enriched LLSVP material, whereas the northern Kea-side samples more depleted lower mantle. These geochemical considerations, taken alongside the results of Jones *et al*. (2016), indicate that the entrained LLSVP component is only slightly denser than the depleted mantle; otherwise the bilateral zonation could not be maintained. Interestingly, a bilateral zonation is observed in other hotspots, *e.g.* Samoa, Societies, Galápagos, Easter, and Marquesas Islands in the Pacific (Hoernle *et al*., 2000; Huang *et al*., 2011; Chauvel *et al*., 2012; Payne *et al*., 2012; Harpp and Weis, 2020) and the Tristan-Gough hotspot track in the Atlantic (Rohde *et al*., 2013; Hoernle *et al*., 2015), but see Stracke (2021) for a different interpretation.

4.3. The role of a distinct rheology

Most models consider that the mantle has a Newtonian rheology, while Davaille *et al*. (2018) explored the effect of a visco-plastic rheology. In such a case, the flow occurs only when the local deviatoric stress is greater than a critical yield stress. This rheology affects both the plume morphology, which becomes broader, and the vertical velocity profile across the conduit, which becomes smoother than shown in Fig. 5.

The Earth's mantle could also be rheologically heterogeneous, with volumes of rocks characterized by a distinct viscosity, caused by an increase in the mineral-grain size (Ammann *et al*., 2010), by an increase in the silica content (Yamazaki *et al*., 2000) or by variations in the water content (Hirth and Kohlstedt, 1996; Karato, 2010). Zones with an increased viscosity have been modelled at different length-scales, *e.g.* Ballmer *et al*. (2017) considered the hypothesis that the whole lower mantle might be more viscous because of a silica enrichement. Gülcher *et al*. (2020) explored a wide range of parameters (*i.e.* excess compositional density and/or viscosity) and mapped distinct regimes of thermo-chemical convection, ranging from efficient mixing to variable degrees of preservation of the heterogeneities forming piles, domes or isolated blobs. Farnetani *et al*. (2018) focused instead on mantle plumes carrying finite size (30–40 km radius) rheological heterogeneities 20–30 times more viscous than the surrounding rocks. Their numerical simulations showed that the heterogeneities do not stretch, they simply rotate as they well up in the plume conduit (Fig. 6). Intrinsically more viscous material, that resists stretching and mixing, is an ideal candidate to preserve a distinct isotopic fingerprint. In such a case, the presence of rheological heterogeneities crossing the plume melting zone, would induce 'pulses' of geochemically distinct material with a time-scale of $\sim$1 My.

5. Concluding remarks and future perspectives

In this brief review I tried to combine insights from seismology, geochemistry and fluid dynamics in order to constrain the existence, the depth of origin and the heterogeneous nature of plumes. I think it is now clear that some plumes have a deep origin and thus

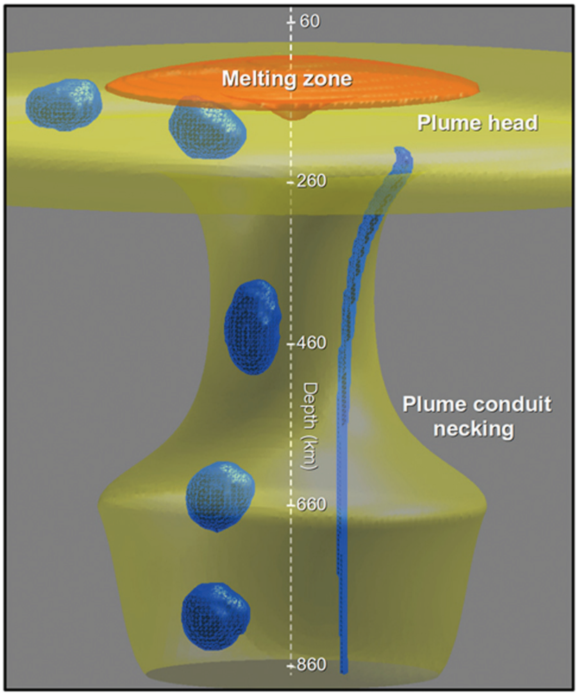

Fig. 6. A passive heterogeneity (right of the plume axis) is deformed readily into a filament. Instead, an active (*i.e.* affecting the flow) heterogeneity 20 times more viscous than the surroundings maintains a 'blob-like' shape, and does not deform. The figure shows the same 'blob' at various times. The plume conduit necking at 660 km depth is due to a 30 times viscosity jump between the lower and upper mantle.

represent a unique window into the lower mantle; however, deciphering the message carried by deep plumes has proved to be challenging. During the past decade there have been considerable advances in analytical geochemistry capabilities (*i.e.* increases in precision, sensitivity and resolving power of mass spectrometers) enabling us to detect small isotopic anomalies and to explore short-lived isotopic systems. In parallel, advances in whole-mantle seismic tomography techniques enabled us to have increasingly sharp images of the Earth's internal structure. In the previous paragraphs I focused on aspects that are generally accepted, albeit still debated. Here I will focus on aspects that still need to be elucidated; I did not not order them according to their importance, they are all important, but following a depth, top-down, criteria.

- Lithospheric and sub-lithospheric processes include, but are not limited to, partial melting of distinct lithologies carried by plumes, the subsequent melt transport within the lithosphere, and the hypothetical mixing of distinct batches of melt in a magma chamber. The key issue will be to quantitatively understand to which extent all these processes can modify the geochemical fingerprint of plume-derived magmas. We also need to understand how to explain geochemical variations occurring on short-time scales (*e.g.* 10^3 years or less), as models based on filaments/blobs carried by plumes only provide a framework to explain isotopic variations occurring on time-scales of order 10^6 years.
- Dynamic processes within an 'enlarged' mantle transition zone. For decades the mantle transition zone was bounded by two mineralogical solid-state phase transitions, at depths of 410 and 660 km, which led to density and viscosity variations. However, seismic tomography has clearly shown that both subducted slabs (Fukao and Obayashi, 2013) and mantle plumes (French and Romanowicz, 2015)

are 'perturbed' also at a depth of 1000 km. In particular, plumes become narrower and start to be more tilted by the global mantle circulation. A viscosity change (Rudolph *et al*., 2015) and/or a compositional change (Ballmer *et al*., 2017) have been invoked, but the exact nature of the 'transition' at a depth of 1000 km is still elusive.

- Linking seismic observations to the geochemical fingerprint of plumes. Plumes with anomalous μ^{182}W have large ULVZs at their base (*e.g.* Cottaar and Romanowicz, 2012; Kim *et al*., 2020; Li *et al*., 2022) something that is not yet explained but is an intriguing observation. Do ULVZs represent the 'Core-Mantle Equilibrated Reservoir' proposed by Mundl-Petermeier *et al*. (2020)? Is this consistent with the observation that plumes with a negative μ^{182}W, coupled to high ^{3}He/^{4}He ratio, also show a weak geochemical fingerprint of recycled material (Jackson *et al*., 2020)? How would this dense material be transported by plumes? LLSVPs have also been considered as a heterogeneous reservoir feeding mantle plumes (*e.g.* Weis *et al*., 2011; McNamara, 2019; Williams *et al*., 2019). Resolving uncertainties about the heterogeneous nature (recycled *vs.* more primitive materials) of LLSVPs and their internal structure might help to explain spatial differences in OIB compositions.
- Interactions between the core and the mantle. The metallic core is a major reservoir for siderophile elements (Carlson *et al*., 2014 and references therein). The core can also be a major reservoir for ^{3}He (Olson and Sharp, 2022) as it was preserved from degassing caused by impacts and by plate cycling. But, is it possible for elements to 'leak out' of the core? We clearly need more metal-silicate partitioning experiments to understand over which time-scales and length-scales such exchanges might occur. This will enable us to assess if core–mantle processes do explain the negative μ^{182}W and the high ^{3}He/^{4}He observed in some plumes.

I conclude by saying that understanding mantle plumes requires understanding large-scale planetary processes responsible for the chemical evolution of the Earth over billions of years.

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