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The Role of Mantle Plumes in Deep Earth and Surface Processes

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Key Messages

- Thermochemical mantle plumes are an integral part of a dynamic Earth's interior.
- Many mantle plumes originate from the deepest regions in Earth's mantle.
- Mantle plumes influence surface processes, including Earth's environment and climate.

1. Abstract

The existence of mantle plumes to explain intra-plate, hotspot volcanism was proposed half a century ago, but their role in Earth's mantle dynamics has been debated. Since then, our understanding of mantle processes has been informed by progress in seismic imaging, modelling of mantle flow in numerical and laboratory experiments, plate tectonic reconstructions, as well as the collection and interpretation of isotopic and geochronological data in ocean island basalts (OIB) and continental hotspot tracks. While the fine-scale structure of mantle plumes has yet to come into focus, seismological evidence for their presence, rooted at the core-mantle boundary, as proposed by Morgan (1971), is mounting. The classical model of purely thermal mantle plumes with narrow (~200 km) stems and large heads, rising vertically through the mantle right underneath all hotspot volcanoes, has been refined. Improved models substantiate that some plumes are thermochemical and can attain complex and broader shapes, that plumes are often not stationary but deflected in the mantle wind, and that many—but not all—plumes are rooted in a dense basal layer, likely of different composition than bulk mantle, and often as part of large regions characterized by anomalously low seismic velocities. Here, we review the recent evolution in our understanding of the morphology and composition of plumes, their role in global mantle convection, as well as how mantle plumes contribute to the long-term evolution of the mantle, and how they may impact climate, ocean chemistry, global biosystem evolution, and continental break-up. Our understanding of the nature and impact of mantle plumes has increased markedly, but more work is required to arrive at a deeper understanding of Earth's dynamic interior as connected to a large range of surface processes.

2. Introduction

Originally proposed to explain the existence of linear chains of intra-plate volcanoes with ages increasing in the direction of plate motion (Wilson 1963), mantle plumes have been classically defined as narrow, thermal upwellings, with large plume heads, as expected in a temperature dependent viscosity fluid of uniform composition that is heated from below. These plumes could start as deep as the core-mantle boundary (Morgan 1971) at roughly 2,900 km depth, rising to the base of Earth's lithosphere over tens of millions of years (**Figure 1a**). Here, partial melting of the plume material may result in magmatism that creates large igneous provinces (LIPs) and age-progressive volcanic chains for the plume head and tail, respectively. This intra-plate volcanism, which typically forms away from and seemingly independent of major plate tectonic boundaries, may be the only means of directly sampling materials from the deepest mantle, and as such provides a unique 'window' into the workings and makeup of Earth's interior.

Mantle upwellings, as an integral part of a convecting Earth, were first invoked in the first half of the 20th century. Alfred Wegener's (1915) theory that continents drift over geological time required a driving mechanism that by some was imagined to involve whole mantle convection (Holmes 1928; 1931). The plate tectonics revolution in the 1960s provided an explanation for volcanism on plate boundaries, but not in the middle of plates. J. Tuzo Wilson (1963) and W. Jason Morgan (1971) hypothesized that the volcanic islands of Hawaii formed on a rigid tectonic plate moving over a 'hotspot' in the Earth's asthenosphere, centered above a vertical, narrow, hot mantle plume from the deep mantle. Coming full circle, the deep mantle sources of these plumes were later connected to the existence of recycled 'old' oceanic lithosphere (and sediments) that were conveyed to Earth's deep interior via subduction and, over the course of hundreds of millions to billions of years, returned to the surface (Hofmann and White 1982; White and Hofmann 1982; Zindler and Hart 1986). If mantle plumes exist, the studies of their volcanic surface products, therefore, provide key insights into mantle dynamics, as well as insights into the convective scales of Earth's very deep interior near the core-mantle boundary, and into the global geochemical cycle governed by downwelling at subduction zones and upwelling from a variety of large-scale chemical mantle reservoirs.

The dynamics of mantle upwellings and their relation with hotspots are still debated. It is unclear whether hotspot volcanism is sourced by broad upwellings, or narrower upwellings in the form of mantle plumes, or both. It is also debated whether some kinds of upwellings are confined to the upper mantle alone, above the seismic 660 km mantle discontinuity, or whether they are all sourced much deeper from near the core-mantle boundary (e.g. Richards *et al.* 1989; Davaille 1999; Steinberger 2000; Courtillot *et al.* 2003; Schubert *et al.* 2004; Koppers 2011; Anderson 2013; Anderson and Natland 2014; Konrad *et al.* 2018a). How many mantle plumes exist, their longevity, dynamic behavior, and chemical make-up, are still poorly understood. While advocates for an important role of plumes in the physical and chemical evolution of the planet argue for connections between large-scale Earth processes, plate tectonics and volcanism, seismic tomography has only recently (Nelson and Grand 2018) imaged the first conduit that is sufficiently narrow that it could be a purely thermal mantle plume as proposed by Wilson and Morgan. Opponents of the plume concept have argued that, alternatively, near-surface processes, limited to Earth's lithosphere, may explain the same observations (Foulger and Natland 2003; Anderson 2013).

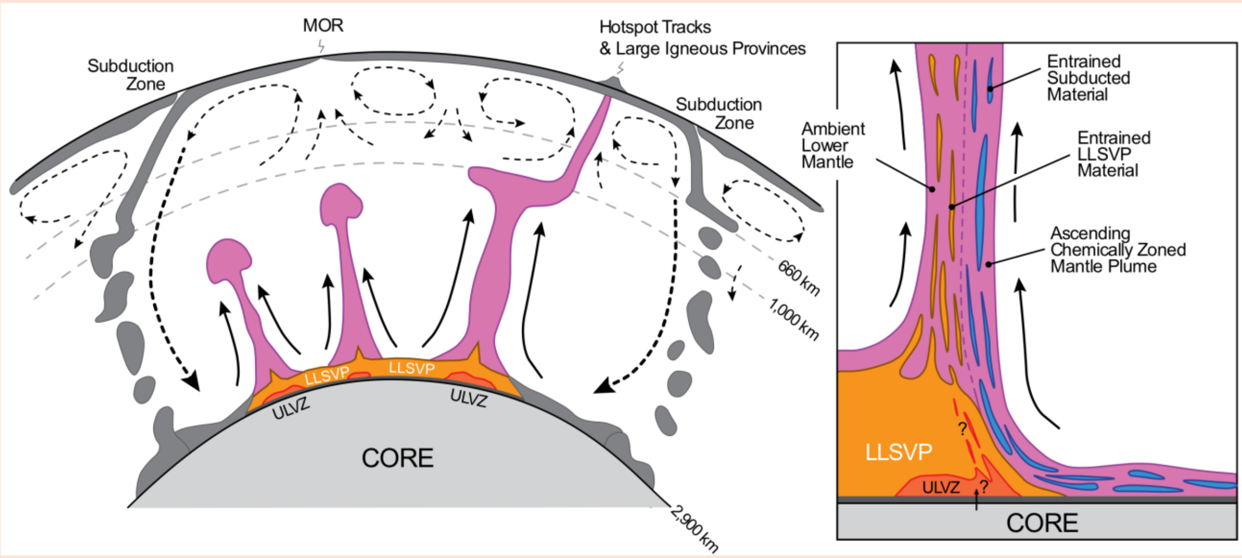


Figure 1: Dynamic Nature of Earth's Interior. **(left)** Schematic cross-section through Earth's interior, depicting the key components of plume generation and upwelling near, above, and along the edges of a Large Low Shear Velocity Province (LLSVP). Modified after Yuen and Romanowicz (2017). **(right)** Schematic cross-section of a plume root showing entrainment of possible mantle components at the edge of an LLSVP and centered above an Ultra-Low Velocity Zone (ULVZ) that may be a unique deep mantle locality containing partial melt. Modified after Torsvik *et al.* (2016).

Over the last decades, advanced computer modeling and laboratory experiments have shown how entrainment of chemical heterogeneity may change the physical and chemical characteristics of plume conduits, causing large variations in the behavior of the resulting 'thermochemical' plumes that are dependent on various plate tectonic and geophysical boundary conditions (e.g. Olson and Yuen 1982; Davaille 1999; Davaille *et al.* 2002; Jellinek and Manga 2004; Ballmer *et al.* 2011; Davies 2005; Deschamps *et al.* 2011; Tan *et al.* 2011; Bossmann and van Keken 2013; Hassan *et al.* 2015). At the same time, seismic tomography studies have pointed out the existence of two large anomalous domains in the deepest parts of the mantle that show lower seismic velocities than surrounding regions (e.g. Dziewonski and Woodhouse 1987; Li and Romanowicz 1996; Grand *et al.* 1997; van der Hilst *et al.* 1997; Su and Dziewonski 1997; Masters *et al.* 2000; Ritsema *et al.* 2011). These domains are now referred to as "large low shear velocity provinces" (LLSVPs) and some studies have proposed that they function as primary plume nurseries. Notably, recent mantle tomography studies have shown the existence of broader 'plume like' structures extending from the deep mantle in the vicinity of some hotspot volcanoes (Montelli *et al.* 2006; Boschi *et al.* 2008; French and Romanowicz 2015; Lei *et al.* 2020).

In this review, we discuss progress in imaging deep mantle structures, understanding their potential role in mantle plume formation, and modeling the shapes and behaviors of plumes depending on composition, rheology, and other boundary conditions. We discuss how the composition of oceanic island basalts increases our knowledge of deep mantle reservoirs and may provide geochemical evidence for plumes generated at the core-mantle boundary. We also discuss how paleomagnetic and age information in seamount chains inform past plate and plume motions, how and if true polar wander of our planet may happen given the observed overall mantle structure and behavior, and how continental break up and the formation of new ocean basins may occur when mantle plumes impinge on the base of Earth's lithosphere. Finally, we discuss how mantle plumes can influence the state of Earth's climate and ocean health on geological timescales.

3. Deep Mantle Superstructures

Based on today's global seismic tomography models we can recognize two large-scale features with anomalously low seismic velocities in the deepest parts of the mantle. Although the general outlines and locations underneath the Pacific and below Africa are generally agreed upon, the makeup and origin of these features (continuous piles versus plume bundles) are still debated as well as their role in the generation of plumes and hotspots (plume nurseries).

3.1 Continuous Piles versus Plume Bundles

The very first tomographic images of the Earth's lower mantle revealed the presence of a very long wavelength structure at the base of the mantle (degrees 2 and 3 in spherical harmonics expansion) anti-correlated with that observed in the gravity field (Dziewonski *et al.* 1977). The authors proposed two possible explanations for the unexpected sign of this correlation. The first is a dynamic interpretation in terms of thermal anomalies and core-mantle boundary (CMB) deflections due to mantle-wide convection; and the second involves lateral variations in composition due to the presence of eclogite-rich material in regions of past subduction and/or chemical plumes originating near the CMB. While these suggestions were quite speculative at the time, they are still actively pursued. The correlation, at large scale, of seismic structure in the deep mantle with anomalies in Earth's geoid and subduction zone configuration on the one hand (e.g. Hager *et al.* 1985; Dziewonski and Woodhouse 1987) and the distribution of hotspots and superswells on the other (Anderson 1982; Davies 1988; Richards *et al.* 1988; Richards and Engebretson 1992; Larson 1991a; Larson 1991b; McNutt 1998) was established soon thereafter.

Scores of seismic studies have since then confirmed the presence of LLSVPs and interpreted the seismically fast areas elsewhere as the remnants of the downgoing slabs from present and past subduction zones (Ricard *et al.* 1993; Lithgow-Bertelloni and Richards 1998), with a time-depth progressive correlation demonstrated for the last 130 million years (Domeier *et al.* 2016). Recent tomographic models show considerable agreement on the extent of the LLSVPs and the shape of their borders when filtered up to spherical harmonics degree 16, or roughly 2,500 km wavelengths (Becker and Boschi 2002; Lekic *et al.* 2012). While the low seismic velocities within the LLSVPs indicate they are likely hot, the anti-correlation of bulk-sound and shear velocity within them (Su and Dziewonski 1997; Kennett *et al.* 1998; Masters *et al.* 2000) as well as the sharpness of their borders (e.g. Ni *et al.* 2002; Wang and Wen 2004; To *et al.* 2005) suggests that they may include a compositional and denser component, although the necessity of the latter has been questioned (Schuberth *et al.* 2009; Davies *et al.* 2012). If parts of the LLSVPs are denser than their surroundings (Ishii and Tromp 2004; Simmons *et al.* 2010) this would help resist entrainment in mantle convection and mixing with the overlying mantle and perhaps facilitate their survival over hundreds of millions of years (e.g. Mulyukova *et al.* 2015), and perhaps throughout most of Earth history, if they, at least partly represent primordial reservoirs (e.g. Ballmer *et al.* 2016). However, resolving the density structure in LLSVPs is challenging (Romanowicz 2001; Trampert *et al.* 2004) and different inferences on their effective density (Lau *et al.* 2017; Koelemeijer *et al.* 2017) might be due to thermal effects offsetting intrinsic density contrasts depending on the scale (e.g. McNamara 2019).

The distinct composition of LLSVPs could be due to the presence at the base of the mantle of a reservoir of primitive material, as might be suggested by studies that relate high $^3He/^4He$ hotspots to plumes originating near the CMB (e.g. Macpherson *et al.* 1998; Williams *et al.* 2019; Mundl *et al.* 2017; Mundl-Petermeier *et al.* 2019; 2020) or to an accumulation of eclogite from subducted crust (e.g. Mulyukova *et al.* 2015), or some depth-dependent stratification of both (e.g. Ballmer *et al.* 2016). But it remains unclear whether the LLSVPs are compact, continuous piles of compositionally distinct material (e.g. McNamara and Zhong 2005; Steinberger and Torsvik 2012; Li and Zhong 2017) where mantle plumes are generated across their tops and along their edges, or whether LLSVPs represent two bundles of closely-spaced thermo-chemical plumes (Schubert *et al.* 2004; Davaille and Romanowicz 2020) separated by a ring of downwellings constrained by the geometry of tectonic plates at the surface (Zhang *et al.* 2010; Bull *et al.* 2014). This remains a subject of debate (e.g. Garnero *et al.* 2016; McNamara 2019; Heron *et al.* 2019; Davaille and Romanowicz 2020).

3.2 Mantle Plume Nurseries

Also debated are whether or not plumes in general, or those leading to LIPs, originate primarily along the edges of LLSVPs (Thorne *et al.* 2004; Boschi *et al.* 2007; Tan *et al.* 2011; Steinberger and Torsvik 2012; Davies *et al.* 2015a; Austermann *et al.* 2014; Hassan *et al.* 2015; Doubrovine *et al.* 2016). As the LLSVP edges are defined by sharp vertical and horizontal gradients in seismic shear-wave velocities (Ni *et al.* 2002; To *et al.* 2005; Frost and Rost 2014) and plumes and reconstructed LIP locations tend to occur close to vertically above them, they are proposed by some to act as 'plume generation' zones (**Figure 1b**) for all major hotspots and LIPs that have been active and remained approximately in place over at least the last 200 million years (Burke and Torsvik 2004; Torsvik *et al.* 2006; Burke *et al.* 2008). However, this is particularly difficult to determine in the case of the rather narrow African LLSVP (**Figure 2**). Interestingly, the roots of at least some of the LLSVP-rooted plumes do seem to contain unusually large ultra-low-velocity zones (ULVZs) (Hawaii: Cottaar and Romanowicz 2012; Samoa: Thorne *et al.* 2013; Iceland: Yuan and Romanowicz 2017; Marquesas: Kim *et al.* 2020) that are indicative of a compositionally different, likely denser component (Rost *et al.* 2005), possibly due to the presence of partial melt (Williams and Garnero 1996) or at least iron enrichment (Mao *et al.* 2006).

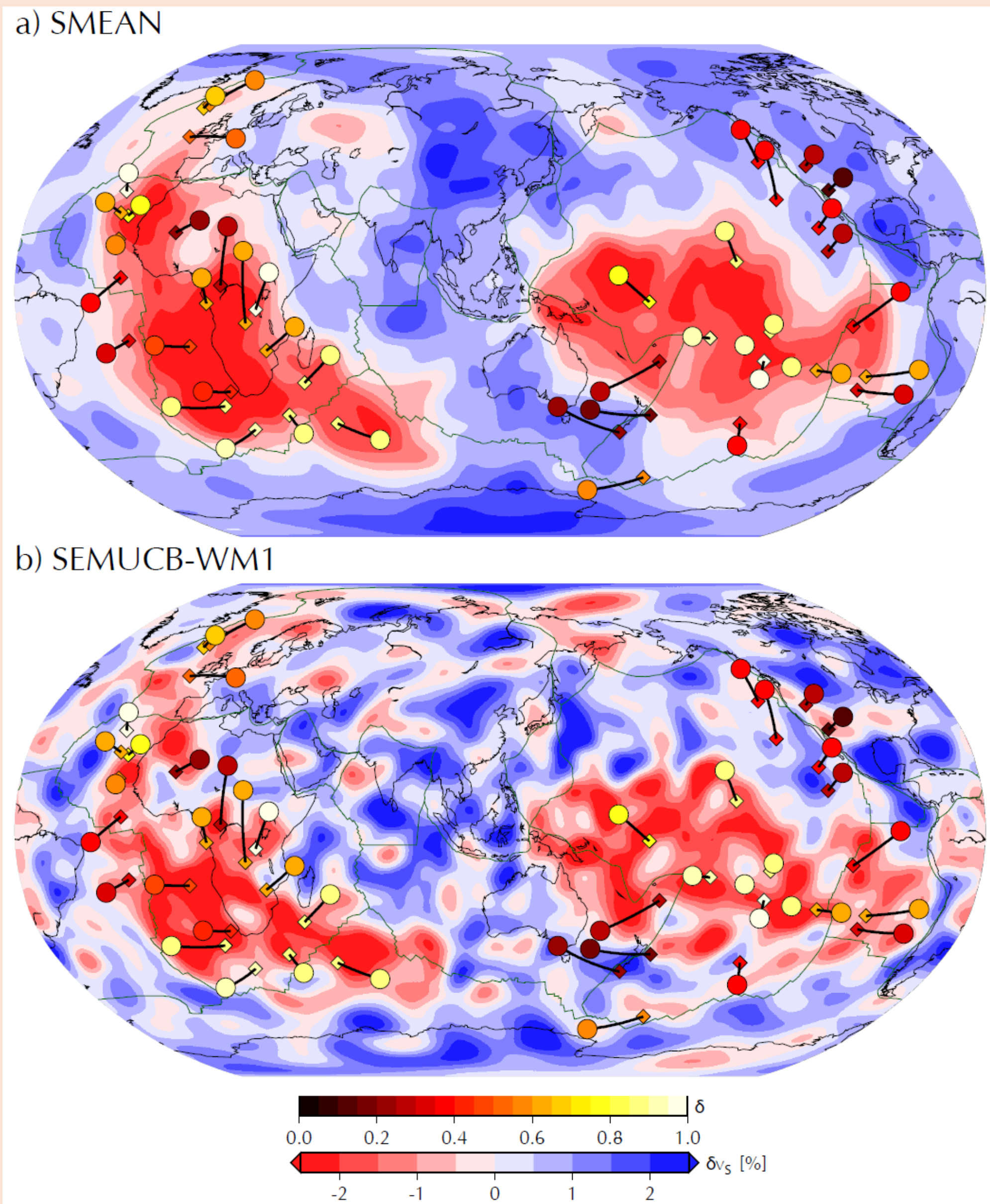


Figure 2: Mantle Plume Locations and LLSVPs. Global maps linking surface hotspots (circles) with depth-projected bottom source locations (diamonds) of mantle plumes (modified from Boschi *et al.* 2007). Background coloring is shear wave anomaly at 2,875 km depth depicting where the two LLSVPs are located beneath Africa and the mid Pacific. The SMEAN composite (Becker and Boschi 2002) and SEMUCB-WM1 (French and Romanowicz 2014; 2015) models illustrate evolving tomographic views of the LLSVPs with the latter model supporting the view that LLSVPs are more granular, potentially indicating that LLSVPs are bundles of closely-spaced plumes instead of large piles. The maps suggest that most plumes originate above LLSVPs with a smaller group (Bowie, Cobb, Guadelupe and Socorro in the eastern Pacific) forming away from these lower mantle anomalous regions. The δ metric in the upper color bar is used to color hotspots and inferred plume base locations; it is the normalized conduit length as identified in tomography where zero means that none and unity means that 100% of the plume length are mapped. The δ_{Vs} in the lower color bar is the amplitude of the seismic shear velocity anomaly.

4. Mantle Plume Characteristics

In this section, we review the key characteristics of mantle plumes. We will discuss how we know that plumes exist (imaging plumes) and what is the expected makeup and behavior of plume heads and tails (plume generation and ascent). From this a distinct picture emerges in which plumes are persistent features given the deep mantle conditions that occur in Earth today and, at least, over the last few hundreds of millions of years.

4.1 Imaging Mantle Plumes

Where seismic tomography more easily picks up large-scale LLSVP superstructures and faster subducting slabs, imaging seismically slower ‘tubular’ mantle plumes in the ocean domain is challenging (Ritsema and Allen 2003; Sleep 2006). Within the framework of purely thermal convection with temperature-dependent viscosity in a fluid heated from below, a ‘thermal’ mantle plume conduit may only be 100-200 km in diameter in the upper mantle but increase to more than 400 km in diameter in the lower half of the mantle that has a significantly higher viscosity (e.g. Steinberger and Antretter 2006). Plume detection is difficult because of a limited resolution in seismic tomography, especially due to a lack of earthquake sources and receiver stations in the ocean basins, and wavefront healing effects that hide low velocity domains of small diameter when classical travel time tomography is applied. The Iceland plume at first could only be seen in the upper mantle with an upwelling broader than expected (Wolfe *et al.* 1997; Allen *et al.* 2002). Similarly, imaging of the Hawaii plume originally resulted in the detection of a broad, inclined upwelling, disappearing from view below 1,500 km mantle depths (Montelli *et al.* 2004; Montelli *et al.* 2006; Wolfe *et al.* 2009). Generally, teleseismic travel time tomography has found plume-like conduits of at least 400 km diameter in earlier studies (Nolet *et al.* 2005; Bijwaard and Spakman 1999). In more recent, higher resolution models, such broader plume-like conduits are found beneath many major hotspots, appear to be rooted near the CMB and rise all the way through the lower mantle, reaching upper mantle depths in the vicinity of hotspots (Boschi *et al.* 2008; French and Romanowicz 2015; Lei *et al.* 2020). Their larger than ~500 km diameter (accounting for smearing due to the inversion process) indicates that they are likely thermochemical rather than purely thermal (French and Romanowicz 2015; Davaille *et al.* 2018) and most of these broad conduits are observed over the LLSVPs.

Anomalies in the travel times of seismic core waves, recorded by the dense USArray seismic network in North America, now also reveal a lower-mantle plume beneath the Yellowstone hotspot (Nelson and Grand 2018) that is probably unrelated to the Pacific LLSVP. This is a unique case study as Yellowstone is so far the only plume where the predicted tilted conduit shape can be matched in detail with a tomographic conduit image (Steinberger *et al.* 2019b). In the absence of individual conduits resolved at that time for most hotspots, Boschi *et al.* (2007) found that modeled plume conduits that take into account the effects of advection, and the associated displacement of plume sources at the base of the mantle, agree better with tomographic results than vertical conduits. The correlation of negative anomalies in seismic tomography with predicted plume conduits is indeed statistically highly significant (Boschi *et al.* 2008) and provides the counterpart to the correlation between mantle tomography and forward models of subduction (e.g. Steinberger *et al.* 2012). In addition, and different from geodynamically modelled conduits that tend to be tilted throughout the entire mantle (e.g. Steinberger 2000; Steinberger and Antretter 2006), French and Romanowicz (2015) imaged plumes that are nearly vertical in much of the lower mantle, but some are strongly tilted above 1,000 km depth. This difference between geodynamic models and seismic observations should provide important constraints on mantle rheology. Such a strong tilt above ~1,000 km is also found for the Yellowstone plume (Nelson and Grand 2018). This may indicate that current mantle flow models are overestimating overall flow speeds below ~1,000 km depth relative to rising speeds of plume heads and tails, and underestimating flow speeds above this horizon.

Recent global seismic tomography indicates that both downgoing slabs (e.g. Fukao and Obayashi 2013) and upwellings, as manifested by broad plume conduits (e.g. French and Romanowicz 2015; Lei *et al.* 2020) appear to be deflected horizontally not only around 660 km depth—where we know there is a seismic discontinuity that corresponds to the phase transition from ringwoodite to bridgmanite and oxides—but also around 1,000 km depth. This is accompanied by a decorrelation between the longest wavelength seismic structures in the extended transition zone (400-1,000 km) and the deeper mantle (e.g. Rudolph *et al.* 2015), which is visible even in lower resolution seismic mantle models and indicated by seismic data suggesting vertical decorrelation at ~800 km depth (Boschi and Becker 2011). This implies a change of material properties that may not coincide with the 660 km discontinuity, and could be explained by an increase in mantle viscosity somewhat deeper than traditionally considered, given the rather non-unique constraints provided by geoid and post-glacial rebound data (e.g. King and Masters 1992; Mitrovica and Forte 2004).

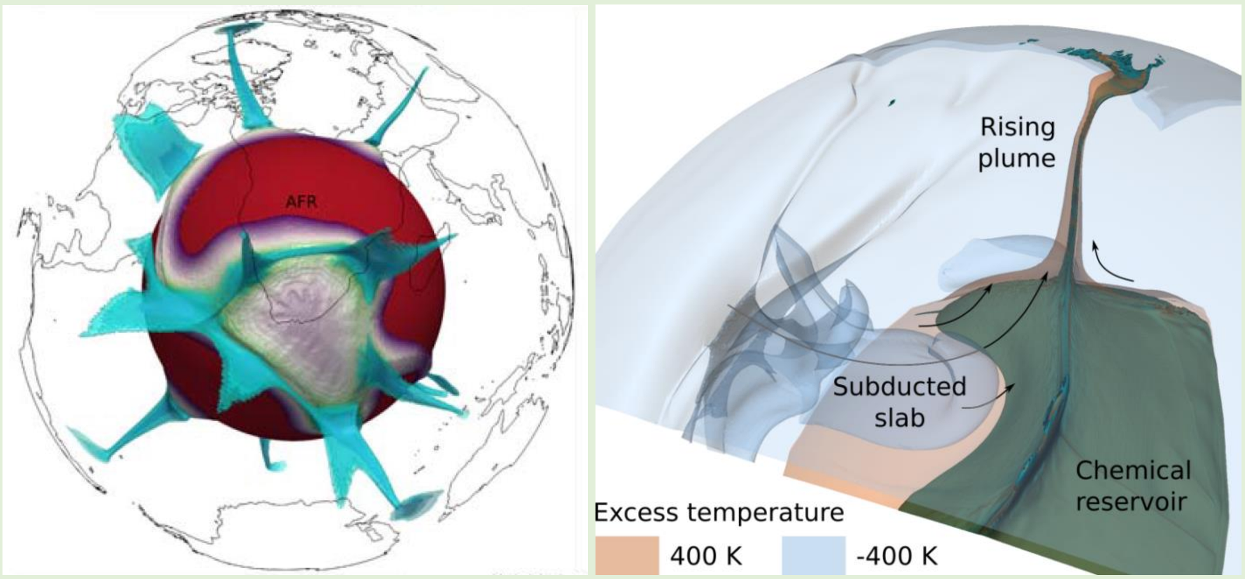
In summary, while there is still progress to be made in resolving the details of plume-like conduits, and in particular their actual diameter, seismic tomography has already shown the presence of mantle plumes below major hotspots, rooted at the core-mantle boundary, although the roots are generally shifted horizontally from the location of the corresponding hotspots due to plume conduit tilting in the upper third of the mantle. Most importantly, they are typically not the simple, purely thermal plumes that geophysicists have been hunting for. Such classical thermal plumes with narrow tails ~200 km in diameter (Farnetani 1997) but possibly reaching ~400 km in parts of the lower mantle are not ruled out, but they are probably mostly below the resolution of current global seismic tomography. Moreover, the broad plumes currently detected do not rise vertically from the CMB to the corresponding hotspot, they show significant deflections in primarily the upper 1,000 km of the mantle, likely due to mantle wind as predicted by geodynamic computations (Steinberger and O'Connell 1998; Steinberger 2000).

4.2 Models of Plume Generation and Ascent

Localized hot upwelling plumes are expected in any terrestrial-type planet mantle, where convection operates with some amount of bottom heating, or where other domains with concentrated heat production can sustain a thermal boundary layer (**Box 1**). Differences in temperature and/or composition will cause variations in density that may result in thermochemical instabilities near boundary layers as the beginning of an upwelling and, when it persists, a rising mantle plume (e.g. Jellinek and Manga 2004).

Box 1: Dynamic Simulations of Plume Behavior in 'Earth-like' Planets. Although thermal instabilities at the core-mantle boundary are inevitable, conditions in Earth's mantle are such that they cannot grow undisturbed. Rather, mantle flow driven by slabs, many of which are subducted to the deep lower mantle and piled up into 'slab graveyards' (Ricard *et al.* 1993; Lithgow-Bertelloni and Richards 1998; Steinberger *et al.* 2012; Mulyukova *et al.* 2015; van der Meer *et al.* 2018; Domeier *et al.* 2016) will trigger plumes, characteristically above Large Low Shear Velocity Provinces (LLSVP) and along their edges (Tan *et al.* 2002; Steinberger and Torsvik 2012; Hassan *et al.* 2015; Li and Zhong 2017). The left-hand panel shows the results of a 3D and time-dependent numerical model with plumes (blue) rising from a LLSVP situated below southern Africa (Hassan *et al.* 2015). Rising from LLSVP margins, plumes in this model are significantly hotter than surrounding mantle, starting out with large temperature anomalies of about 500° K in the lowermost mantle, as shown in the right-hand panel (Dannberg and Gassmöller

2018). The plumes will slowly cool down when they rise, with smaller plumes losing a larger fraction of their heat on the way up through the mantle. However, larger plumes will retain their heat better (Albers and Christensen 1996; Zhong 2006; Leng and Zhong 2008) with estimated excess plume temperatures of 200-300 °K at asthenospheric depths for Hawaii (Moore *et al.* 1998; Ribe and Christensen 1999; Schubert *et al.* 2001).



Typical ‘thermal’ mantle plumes in Earth will have a broad plume head, up to roughly thousand kilometers in diameter, followed by a narrow plume tail, not wider than a couple of hundred kilometers (e.g. Richards *et al.* 1989; Sleep 1990; Coffin and Eldholm 1994; Davaille 1999; Campbell 2007). These plumes can persist over long geological times, but only if the thermal boundary layer from which they arise is also maintained for hundreds of millions of years (Sleep 2003; Tan *et al.* 2002; Jellinek and Manga 2004; Burke *et al.* 2008). In fact, for plumes to keep rising through the entire mantle, a lower limit of ~500-1000 kg/s of anomalous mass flux is required to sustain the plume tails (Albers and Christensen 1996). Rising plumes from near the LLSVPs are expected to be buoyant due to their hotter temperatures, but if they entrain chemically distinct materials from the LLSVPs that are heavier than surrounding mantle (Lau *et al.* 2017) then their buoyancies may be substantially reduced (Lin and van Keken 2006a; Mulyukova *et al.* 2015). The anomalous density of LLSVPs (Lau *et al.* 2017; Koelemeijer *et al.* 2017) and the nature of the entrainment of other materials during plume ascent (e.g. Farnetani 1997; McNamara and Zhong 2004) are still debated. Because of time-dependent and variable amounts of entrainment, highly complex plume behavior and shapes (**Figure 3**) may result that substantially differ from the classical head-and-tail structure (Farnetani and Samuel 2005; Lin and van Keken 2006b; Kumagai *et al.* 2008). For example, negative buoyancy of (denser) material entrained from LLSVPs in plume heads may cause material to sink back into the ascending plume (Ballmer *et al.* 2013; Dannberg and Sobolev 2015) leading to broader plume conduits that are a few hundred kilometers wider than typical thermal plumes.

How long it takes for a plume head to traverse the mantle after forming at the core-mantle boundary is difficult to estimate. It primarily depends on its buoyancy, arising from a density contrast of about 30 kg/m^3 for thermal plumes, but which could be much less for thermochemical plumes, and the average viscosity of the surrounding mantle. Widely discrepant estimates exist for mantle viscosity, as it may be (locally) controlled by variations in temperature and stress that may render global average viscosity estimates not applicable for plumes (e.g. Larsen and Yuen 1997). One traversal time estimate can be made because reconstructed LIP eruptions are correlated with LLSVP margins and therefore LIPs are hypothesized to erupt from those margins. But in order for that correlation to be maintained during their rise through the mantle, plume heads must rise up from the lower mantle rather fast, probably within 30 million year or less, to avoid large lateral deflections, consistent with numerical models (e.g. Hassan *et al.* 2015). However, for smaller plume heads, such as for the Yellowstone plume—which is not associated with the Pacific LLSVP and is located in a region dominated by subduction—it is estimated that its plume head rose more slowly, taking 80 million years or longer (Steinberger *et al.* 2019b). Plumes rise therefore considerably faster than slab sink, at estimated speeds of 1-2 cm/yr, such that slabs require ~150-200 million years to reach the bottom of the mantle (Steinberger *et al.* 2012; Hassan *et al.* 2015; Butterworth *et al.* 2014; Van der Meer *et al.* 2018). In addition, during ascent, the even more slowly rising plume conduits are predicted to become increasingly tilted with time, as their roots become shifted towards large-scale upwellings, likely above the two LLSVPs (Steinberger and O'Connell 1998; Steinberger 2000).

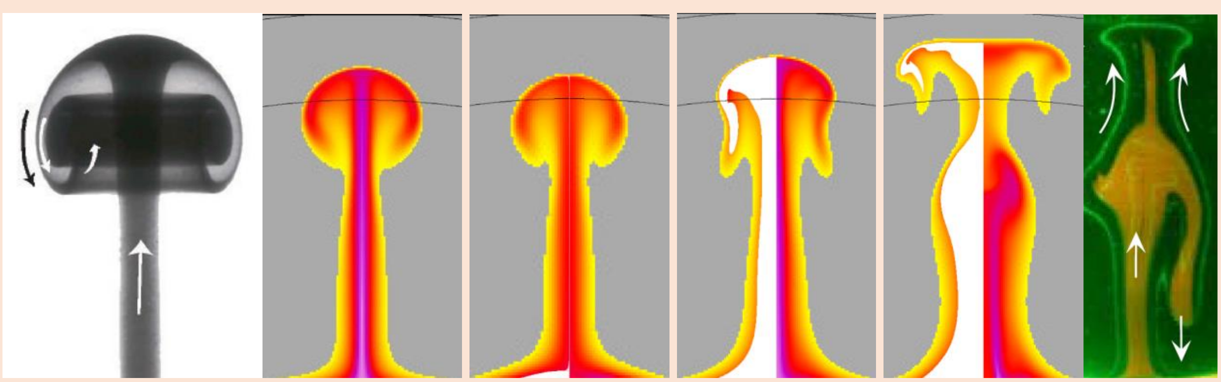


Figure 3: A Gallery of Rising Thermal and Thermochemical Plumes. **From left to right:** Laboratory model of rising thermal plume head, connected by a stem to the source region (Campbell *et al.* 1989); four numerical models of one axisymmetric thermal and three thermochemical plumes (Lin and van Keken 2006a) with colors representing temperature (magenta=hottest; yellow=coolest) and entrained material (white) with variable chemical excess densities of 200, 50 and 60 (from left to right) kg/m^3 in left

half of the visualized plume conduits; and a laboratory syrup tank model of a plume that partially fails when entrained materials (that are too heavy) collapse in the top of the conduit (Kumagai *et al.* 2008).

Eventually, when the plume head reaches the lithosphere, a LIP may form over short amounts of geological time, likely in less than a million years (Coffin and Eldhom 1994). Long-lived hotspot tracks may form after that and persist in some cases for 80 to 120 million years (Peate 1997; Koppers *et al.* 2012; Konrad *et al.* 2018a) or even longer (e.g. Hoernle *et al.* 2015). Several hotspot tracks are associated with flood basalts at one end (e.g. Tristan, Reunion) and thus may correspond to typical thermal plumes with a tail following the head (Richards *et al.* 1989; Ernst and Buchan 2003). Although the typical lifetime of a deep mantle plume is likely long, on the order of ~100 million years, some of the older Pacific hotspot tracks were only active in the Cretaceous and do not correspond to any currently active hotspot (e.g. Hess and Shatsky Rise; Koppers *et al.* 2001; Tejada *et al.* 2016). Evidence from the bathymetric record in the Pacific Ocean also indicates that many hotspot trails are characterized by shorter seamount trails, apparently only active for up to 30 million years (Clouard and Bonneville 2001; Koppers *et al.* 2003). These surface expressions, however, do not have to mean that the mantle sources are short-lived per se, but rather that instabilities may develop in rising mantle plumes, for example, once conduits are tilted more than 60° from the vertical, at which stage they may break apart (Whitehead 1982; Steinberger and O'Connell 1998). Plumes may also go extinct if their buoyancy fluxes are too high and they cut themselves off from a supply of hot mantle material. Plumes may become internally unstable and collapse due to insufficient buoyancy, they may appear to switch on/off if rising plumes are pulsating or boudinaging, and in some cases they may begin without flood basalts. All possible explanations for such intermittent plume behaviors and plumes without heads are still being mapped out and debated.

5. Mantle Plumes Illuminating Deep Mantle Heterogeneities

Plumes provide a unique view into Earth's mantle, revealing its intricate chemical makeup and evolution over billions of years. Although seamount trails and LIPs have complex construction histories, variations in their chemical compositions provide clues about early Earth (enduring ancient signatures) and the number and composition of distinct chemical endmembers that reside in Earth's interior (mantle heterogeneities). These clues lead to intriguing debates on how deep, and at what length and time scales, those endmembers are manifested in the mantle (location of primordial and recycled reservoirs) and how hotspot trails may 'mimic' these heterogeneities (striped plumes).

5.1 Enduring Ancient Signatures

Extinct short-lived isotope systems (such as ^{129}I-^{129}Xe [with a half-live of $t_{1/2}$=15.7 Ma], ^{182}Hf-^{182}W [$t_{1/2}$=8.9 Ma] and ^{146}Sm-^{142}Nd [$t_{1/2}$=103 Ma]) provide unmatched insights in the processes happening in the earliest approximately 50 (for ^{182}W) to 500 million years (for ^{142}Nd) of Earth's history (Halliday 2004; Carlson and Boyet 2009; Mundl *et al.* 2017; Mukhopadhyay and Parai 2019). Because radioactive decay in these isotope systems is rather rapid on planetary timescales, resolvable changes in these isotopic systems ($^{129}Xe/^{130}Xe$, $^{182}W/^{184}W$ and $^{142}Nd/^{144}Nd$) are restricted to the Hadean, the opening eon in Earth's history, ending 4 billion years ago. The presence of such ancient isotopic anomalies in mantle plume source regions, therefore, suggest that some primordial mantle reservoirs are still present in the Earth's interior despite extensive convective mixing (Mukhopadhyay 2012; Peters *et al.* 2018; Williams *et al.* 2019; Mundl-Petermeier *et al.* 2020). For example, the ^{129}I-^{129}Xe system shows a marked difference in $^{129}Xe/^{130}Xe$ between Earth's mantle and atmosphere (Allegre *et al.* 1983), and heterogeneous $^{129}Xe/^{130}Xe$ is also preserved in the mantle since the Hadean (Mukhopadhyay 2012). Application of the ^{182}Hf-^{182}W and ^{146}Sm-^{142}Nd systems confirms the survival of Hadean-generated signatures in the modern mantle, with resolvable ^{182}W and ^{142}Nd anomalies in OIB present in mantle plumes (Mundl *et al.* 2017; Peters *et al.* 2018; Horan *et al.* 2018), but the discovery of anomalous ^{182}W in plume-head-generated flood basalts (Rizo *et al.* 2016) remains controversial (Kruijer and Kleine 2018).

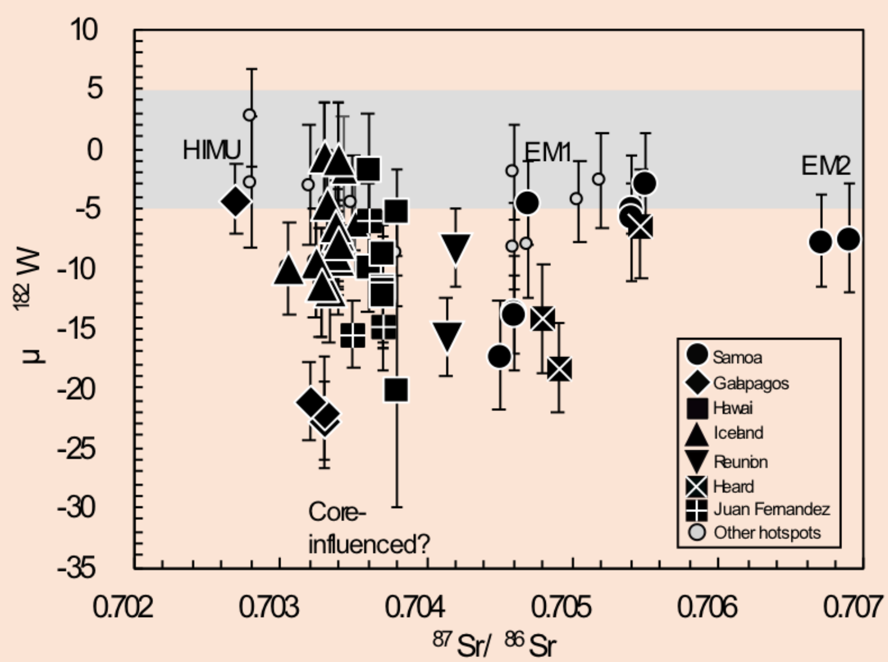


Figure 4: Isotope Systematics in Global Plume Volcanics. Negative $\mu^{182}W$ anomalies appear in some plume-generated OIB volcanoes, with no relation to long-lived $^{87}Sr/^{86}Sr$ signatures in the HIMU, EM1

and EM2 mantle sources. One interpretation of these observations is that that some deeply sourced mantle plumes with strongly negative $\mu^{182}W$ anomalies have inherited a W-isotopic signature of Earth's core (Rizo *et al.* 2019; Mundl-Petermeier *et al.* 2020). In this diagram the grey bar represents the 2σ reproducibility of the standard. Estimated core $\mu^{182}W$ is -220 (Touboul *et al.* 2012).

More recent work, however, established that ^{182}W anomalies in plume-formed OIB do not exhibit straightforward relationships (**Figure 4**) with long-lived heavy radiogenic isotopes, such as $^{87}Sr/^{86}Sr$ ($t_{1/2}$ = 49 billion years). Instead, most OIB that host negative ^{182}W anomalies also appear to be associated with high $^{3}He/^{4}He$ lavas (Mundl *et al.* 2017; Mundl-Petermeier *et al.* 2019; 2020) that in turn are typically interpreted to sample deep primordial mantle signatures (e.g. Kurz *et al.* 1982) present in only the hottest and most buoyant plumes (Jackson *et al.* 2017). It is hypothesized that these OIB ^{182}W anomalies reflect a contribution from Earth's core, which has preserved a low $\mu^{182}W$ value (the deviation of $^{182}W/^{184}W$ from the terrestrial standard in parts per million) because tungsten is a moderately siderophile element that, during core formation, became enriched in the Earth's core relative to the short-lived, lithophile radioactive parent (^{182}Hf), which remained in the mantle (Rizo *et al.* 2019; Mundl-Petermeier *et al.* 2020). It is possible that this core material is partitioned back into the mantle at the base of mantle plumes, aided by silicate melting (Mundl-Petermeier *et al.* 2020), possibly in the ultra-low seismic velocity zones at the core-mantle boundary (**Figure 1b**).

5.2 Mantle Heterogeneities Inferred from Hotspots

Lower-mass stable isotopes (such as oxygen, sulfur, calcium, and iron) provide distinguishing 'surface' isotopic signatures because low temperature alteration, biological processes, and other shallow-level mechanisms modify their isotopic ratios. The discovery of these 'surficial' signatures in mantle plume source regions, therefore, provides first order evidence that Earth is recycling its lithospheric plates in subduction zones at a global scale and is resupplying the deep source regions of mantle plumes with various 'crustal' materials (Eiler *et al.* 1996; Huang *et al.* 2011a; Cabral *et al.* 2013; Konter *et al.* 2016; Nebel *et al.* 2019; Gleeson *et al.* 2020). High-mass radiogenic isotopes (such as $^{87}Sr/^{86}Sr$, $^{143}Nd/^{144}Nd$, $^{206}Pb/^{204}Pb$ and $^{176}Hf/^{177}Hf$) paint an even more complex picture of global mantle heterogeneity, providing insight into which, how many, and in what way mantle end members are involved in the chemical dynamics of mantle plume formation (Allegre 1982; Hofmann and White 1982; Zindler and Hart 1986; White 2010; Jackson *et al.* 2018b). Many of the global hotspot systems have two, three, or more, distinct components in their mantle plume source regions (Harpp and White 2001; Hoernle *et al.* 2000; Jackson *et al.* 2014).

Lava flows from LIPs and hotspot-related OIBs thus sample a diverse array of long-lived radiogenic isotopic compositions. These chemical trends have been classified into four primary "species" or endmembers (**Figure 4**) based on isotopic composition: **EM1** (enriched mantle I) characterized by low moderately-high $^{87}Sr/^{86}Sr$ and low $^{206}Pb/^{204}Pb$; **EM2** (enriched mantle II) characterized by high $^{87}Sr/^{86}Sr$ and intermediate $^{206}Pb/^{204}Pb$; **HIMU** (or high "μ", where $\mu = ^{238}U/^{204}Pb$) with low $^{87}Sr/^{86}Sr$ and high $^{206}Pb/^{204}Pb$; and a geochemically-depleted component variously referred to as **PREMA** (Prevalent Mantle; Zindler and Hart 1986), **FOZO** (Focus Zone; Hart *et al.* 1992), or **C** (Common; Hanan and Graham 1996).

The EM2 mantle domain almost certainly relates to recycling of terrigenous sediments derived from ancient upper continental crust, and is manifested mostly clearly in Samoan hotspot lavas (Jackson *et al.* 2007; Workman *et al.* 2008), where the lavas exhibit clear radiogenic isotopic and trace element signature fingerprints seen only in continental crust. There is less certainty about the origin of the HIMU component in hotspots, defined by their highly radiogenic $^{206}Pb/^{204}Pb$, a signature that has been attributed to sampling of ancient subducted oceanic crust (Hofmann and White 1982) or marine carbonates (Castillo 2015), metasomatism of the underlying oceanic lithospheric mantle (Pilet *et al.* 2008) or the subcontinental lithospheric mantle (Weiss *et al.* 2016), and possibly could originate from a reservoir in the transition zone (Nebel *et al.* 2013; Mazza *et al.* 2019; Huang *et al.* 2020) or lower mantle (Collerson *et al.* 2010). The presence of mass independently fractionated sulfur (MIF-S) isotopes in endmember HIMU lavas from Mangaia Island supports an oceanic crustal recycling model (Cabral *et al.* 2013) as these MIF-S are interpreted to represent a unique fingerprint from Archean atmosphere that become associated with oceanic crust (Farquhar *et al.* 2002). The origin of EM1 lavas is the least certain among the mantle domains, as both deep and shallow mantle metasomatic processes have been invoked, as well as recycling of a variety of different lithospheric protoliths (see Garapic *et al.* 2015 and references therein). However, the discovery of MIF-S in EM1 lavas from Pitcairn supports models advocating recycling of shallow crustal protoliths (Delavault *et al.* 2016), and radiogenic isotopic and trace element signatures are consistent with a continental crustal protolith in the EM1 mantle (Eisele *et al.* 2002).

Nonetheless, extreme EM1-EM2 and HIMU compositions are relatively uncommon: the bulk of volcanism at hotspots and oceanic LIPs (Zindler and Hart 1986) is geochemically depleted relative to a chondrite-based bulk silicate Earth—the long-accepted compositional model for the formation of our planet (McDonough and Sun 1995). This suggests that, like the upper mantle, the lower mantle also must have experienced a strong depletion of incompatible elements by ancient prolonged crustal extraction (Hart *et al.* 1992). Consistent with this observation, studies based on the short-lived ^{146}Sm-^{142}Nd system suggested that the accessible silicate Earth has elevated $^{142}Nd/^{144}Nd$ and is the product of a global early-Hadean mantle depletion event (Boyet and Carlson, 2005; Jackson *et al.* 2010; Caro and Bourdon, 2010). This model requires a complementary early-formed enriched reservoir (now hidden) with low $^{142}Nd/^{144}Nd$ that resides in the deep mantle or was lost to space (Jellinek and Jackson 2015; and references therein).

While this model was questioned (Bouvier and Boyet 2016; Burkhardt *et al.* 2016) more recent work may support a non-chondritic composition for the silicate Earth caused by a massive Hadean depletion event (Debaille *et al.* 2019). In this debate geoneutrinos present an opportunity to map out the spatial distribution of geochemical reservoirs in the Earth's deep interior, particularly with respect to the geochemically-important radioactive heat-producing elements (U, Th, K), which would be elevated in a putative hidden early enriched reservoir. This mapping of geoneutrinos may help resolve the debate over the bulk composition of the planet (Šrámek *et al.* 2013).

5.3 Location of Primordial and Recycled Reservoirs

Mechanisms for the long-term preservation of the Hadean geochemical anomalies are imperfectly understood (Gülcher *et al.* 2020), but storage in dense, viscous domains of the deep mantle may isolate Hadean-formed reservoirs from convective motions of the mantle (Samuel and Farnetani 2003; Lin and van Keken 2006a; Deschamps *et al.* 2011). The two LLSVP regions at the core-mantle boundary are attractive locations in this regard (e.g. Tackley 1998; Macpherson *et al.* 1998; Mukhopadhyay 2012), though alternatives have been suggested (e.g. Allegre *et al.* 1984; Becker *et al.* 1999; Ballmer *et al.* 2017). Hotspots with the highest (most primordial) $^{3}He/^{4}He$ ratios do appear to be positioned over the LLSVPs (Macpherson *et al.* 1998; Garapic *et al.* 2015; Williams *et al.* 2019) and are consistent with an ancient source for these seismic features. However, to date ^{129}Xe, ^{142}Nd, and ^{182}W OIB short-lived isotope datasets are still statistically too small to conclusively link the Hadean isotopic anomalies to the LLSVPs. Much larger Sr-Nd-Pb isotope datasets confirm that EM1-EM2 hotspots are geographically linked to both of the LLSVPs (Castillo 1988; Jackson *et al.* 2018a). However, a recent study suggests that only the Atlantic LLSVP hosts EM signatures (Doucet *et al.* 2020), but such arguments hinge on the selection of plumes (e.g. Jackson *et al.* 2018a) and more work is needed to better identify reservoir geometries and dynamics. HIMU hotspots seemingly are not linked to the LLSVPs (Jackson *et al.* 2018b), leaving the location of this scarce domain uncertain. In fact, the EM and HIMU domains appear to be spatially decoupled in Earth’s mantle, which is unexpected given that both likely formed following subduction and recycling as part of the plate tectonic cycle (Jackson *et al.* 2018b).

While the ULVZs are argued to be compositionally distinct (Li *et al.* 2017) it is not yet possible to evaluate whether ULVZs sample geochemical reservoirs different from LLSVPs. Where some plume conduits are clearly associated with ULVZs (Cottaar and Romanowicz 2012; Thorne *et al.* 2013; Yuan and Romanowicz 2017; Kim *et al.* 2020), other plumes may not be (or, they are, but the ULVZs have not yet been seismically imaged at their plume roots). Until the distribution of ULVZs has been conclusively “mapped out” it will not possible to determine whether ULVZ-related plumes have a different composition than plumes that are not, and this currently limits what can be inferred about the isotope geochemistry of the ULVZs. Ultimately, the evolutions of these geochemical reservoir are linked to plate tectonic processes, including recycling of oceanic and continental crust and transport by plumes, and these processes control the composition, location, size, and longevity of geochemical reservoirs in the Earth's interior.

5.4 Striped Plume Expressions

Surface expressions of plume-fed volcanism in the oceanic realm are highly varied and complex. A most notable complexity is the formation of 'double' track volcanic hotspot trails that are geochemically 'striped' over millions of years of plume history. These kind of surface expressions may be governed by the makeup of the plumes themselves (Abouchami *et al.* 2005; Farnetani and Hofmann 2010; Huang *et al.* 2011b; Weis *et al.* 2011; Hofmann and Farnetani 2013; Hoernle *et al.* 2015; Dannberg and Gassmöller 2018), and by their interactions with the overriding tectonic plates and any changes in plate motion and direction (Moore *et al.* 1998; Hieronymus and Bercovici 1999; Davies *et al.* 2015b; Jones *et al.* 2017). The presence of these geochemically-resolved dual volcanic trends in ocean island systems was first noticed for the Hawaiian Islands by Tatsumoto (1978) and follow the 'Loa' and 'Kea' volcanic tracks (**Figure 5a**) as described by Dana (1849). Since then, *en echelon* trends have also been observed for the Easter, Foundation, Galapagos, Marquesas, Samoan, Society, Tristan-Gough and Rurutu hotspot tracks (Hoernle *et al.* 2000, 2015; Werner *et al.* 2003; Huang *et al.* 2011b; Payne *et al.* 2013; Harpp *et al.* 2014; Chauvel *et al.* 2012; Koppers *et al.* 2011a; Finlayson *et al.* 2019).

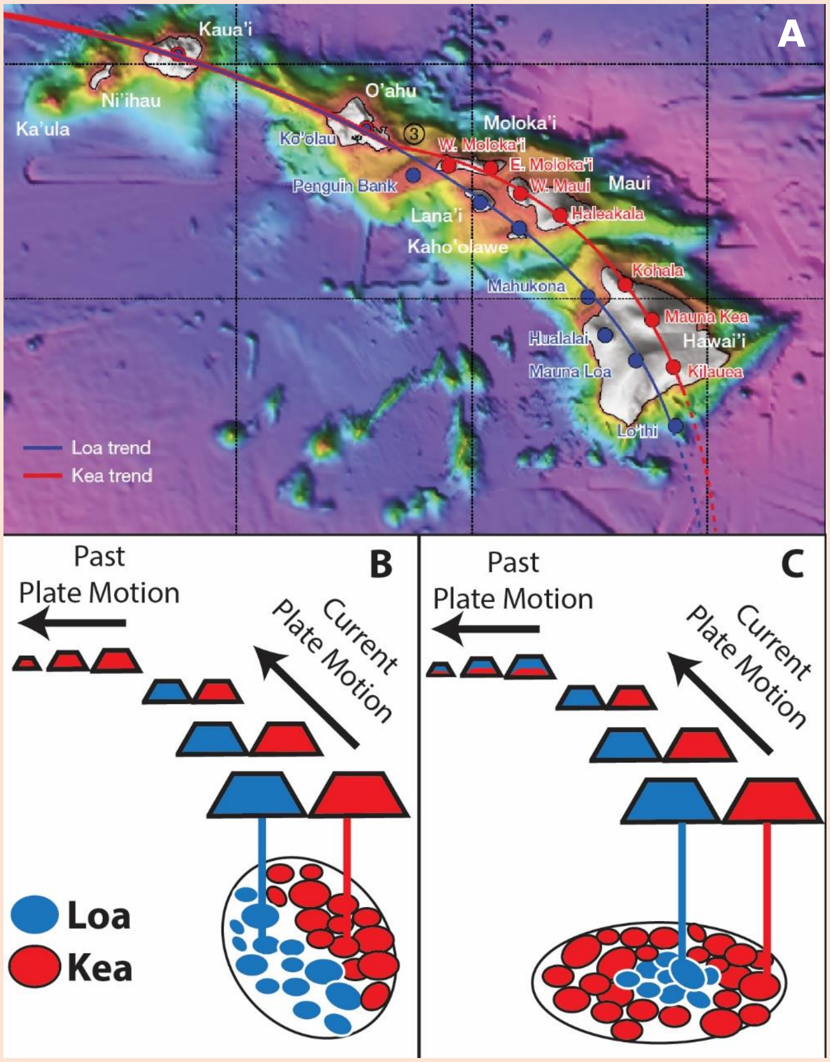


Figure 5: Explaining the Hawaiian Striped Plume. A schematic illustration comparing models for a bilaterally versus concentrically zoned Hawaiian mantle plume. **(A)** The location of the Loa (Blue) and Kea (Red) isotopic trends superimposed on a map of the Hawaiian Islands from Jones *et al.* (2017). **(B)** A top-down view of a bilaterally zoned plume, which has become aligned with the direction of plate motion within the last 5 Myrs (e.g. Farnetani *et al.* 2012). The older seamounts are shown as primarily Kea-types based on the model of Harrison *et al.* (2017). **(C)** A concentrically zoned plume, which currently displays bimodal volcanism due to the recent change in plate motion contrasting with the prior orientation of the plume (e.g. Jones *et al.* 2017). A prediction of this model is that the older seamounts would have Kea-type as their bases, with Loa-type lava capping those in their later volcanic stages.

There are currently two debated models on the origin of these dual isotopic signatures and the inferred chemical structure of mantle plumes: the bilaterally zoned plume (Abouchami *et al.* 2005; Farnetani and Hofmann 2010; Huang *et al.* 2011b, Weis *et al.* 2011; Hoernle *et al.* 2015; Chauvel *et al.* 2012; Dannberg and Gassmöller 2018) and concentrically zoned plume model (Frey and Rhodes, 1996; Kurz *et al.* 1996; DePaolo *et al.* 2001; Jones *et al.* 2017; Konrad *et al.* 2018b). The bilaterally zoned plume model postulates that the geochemical stripes observed among most ocean island chains originate from a mantle plume structure that is divided into two distinct chemical reservoirs (**Figure 5b**). The plume could contain a bilaterally continuous structure (Hofmann and Farnetani 2013), two zones consisting of vertically continuous filaments with some spacing in-between (Abouchami *et al.* 2005), or a partly ordered structure with some mixing between the zones (Ren *et al.* 2005). As some mantle plumes appear to be rooted on the boundary between LLSVPs and ambient lower mantle, it is possible that bilaterally zoned plumes sample both LLSVP (± ULVZ) and the ambient lower mantle materials (Huang *et al.* 2011b; Weis *et al.* 2011; Hofmann and Farnetani 2013).

In the bilaterally zoned plume model for Hawaii, the southern ‘Loa’ component (EM1) of the Hawaiian plume would represent incorporation of LLSVP material, while the northern ‘Kea’ component (PREMA) would represent ambient lower mantle (Weis *et al.* 2011). However, to explain the general absence of a Loa component in the Hawaiian plume prior to ~5 million years ago, Harrison *et al.* (2017) argue that the LLSVP-derived Loa component only became entrained in the plume conduit intermittently between ~47 and 6.5 Ma, and then consistently from 6.5 Ma to the present day. Similar geographical and temporal trends are seen within the Marquesas and Samoan Islands (Huang *et al.* 2011b; Chauvel *et al.* 2012), showing the potential ability to link surficial geochemical signatures to the lowermost mantle geophysical domains.

The concentrically zoned plume model argues that plumes concentrate the hottest and densest materials in their centers (**Figure 5c**) during ascent from the core-mantle boundary (Jones *et al.* 2016). This model explains surficial isotopic stripes as being derived from sampling of the plume core versus its outer rim (DePaolo *et al.* 2001). The concentric model has been supported by noble gas studies that indicate that the most chemically ‘primitive’ lava flows with high ^{3}He/^{4}He ratios are typically found within the central regions of ocean islands (Kurz *et al.* 1996; DePaolo *et al.* 2001; Konrad *et al.* 2018b). An alternative concentric plume model argues for the melting of differing lithologies at differential depths in the plume as a function of plume-plate interaction (Jones *et al.* 2017). However, an unrelated (and controversial) change in Pacific plate motion at ~2.5 Ma is required to force a depth shift in mantle melting that initially would have produced Kea-type lava flows (from a deeper peridotite melting region) followed by Loa-type lava flows (from a shallow pyroxenite source) after the plate motion change (Jones *et al.* 2017). This model currently does not explain the complex zonation patterns observed along the Tristan-Gough track on the African plate (Hoernle *et al.* 2015).

6. Mantle Plumes and Plate Tectonics

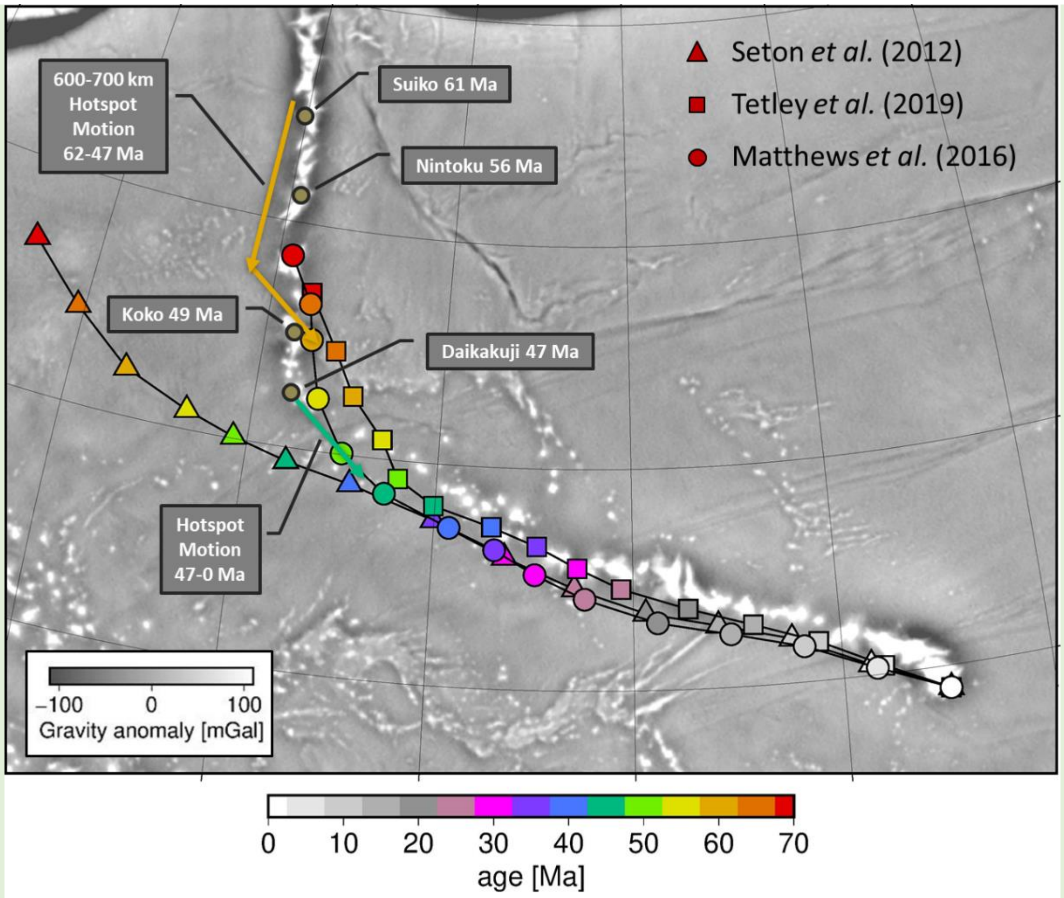
Understanding if and why certain plumes move, in which directions, in unison or not, and how fast, is an ongoing debate. In this section, we show that plumes can move independently and at speeds typically less than plate tectonic movements (distinguishing plume motions). We also discuss how plumes provide insights into possible reorientations of the entire Earth relative to its spin axis (true polar wander) and how they are considered powerful initial agents in the global plate tectonic cycle (breaking up continents).

6.1 Distinguishing Plume Motions

The presumed stability of 'thermal' mantle plumes initially allowed scientists to use the shapes and age progressions of seamount trails to derive directions and speeds of past plate motions, and with those models in hand, to chart out the positions of tectonic plates back through geological time (e.g. Wilson 1963; Morgan 1971; Minster and Jordan 1974; Duncan and Clague 1985; Müller *et al.* 1993; Koppers *et al.* 2001; Wessel and Kroenke 2008; Torsvik *et al.* 2014; Wessel and Conrad 2019). In these models, fixed mantle plumes were presumed to persist over more than one hundred million years. Expected would be that all seamount trails forming on a particular tectonic plate would record the same history of plate rotations around the same Euler poles and with the same angular velocities; and when plate motion changes occurred, the timing of the 'bends' (or turns) that would form in each seamount trail, would be contemporaneous (**Box 2**). In other words, it would be expected that the geometries between different seamount trails would be fixed and their chronologies identical.

Paleomagnetic inclination data from seamount trails indicate that plumes are in fact not stationary with respect to the spin axis (Tarduno *et al.* 2003, Koppers *et al.* 2012; Tarduno and Koppers 2019). Distance comparisons between coeval seamount trails show that plumes are moving away or closing in to each other (O'Connor *et al.* 2013; Konrad *et al.* 2018a) and the $^{40}Ar/^{39}Ar$ dating of the Louisville and Rurutu hotspot tracks on the Pacific plate show that the most acute parts in their bends (not as pronounced or clearly visible as for Hawaii) appear to occur about 3 million earlier than the Hawaii-Emperor Bend around 47 million years ago (Sharp and Clague 2006; Koppers *et al.* 2011b; Finlayson *et al.* 2018; Konrad *et al.* 2018a) but their timing does coincide with the start of the Hawaii-Emperor bend around 50 million years ago. For these Cretaceous and early Cenozoic times, the most recent analysis by Bono *et al.* (2019) of related paleomagnetic data and age dates concluded that the Hawaiian hotspot moved at 48 ± 8 mm/yr between 63 and 52 Ma, while other Pacific hotspots have moved much more slowly. Similarly, based on age progressions along tracks and changing distances between tracks, Konrad *et al.* (2018a) found a total relative motion of 53 ± 21 mm/yr between Hawaii and Louisville and of 57 ± 27 mm/yr between Hawaii and Rurutu, most likely due to a large individual Hawaiian hotspot motion from 60 to 48 Ma. These observations are not unexpected in a convecting mantle, as computer simulations show that rising mantle plumes will be advected in Earth's overall mantle circulation regime (Steinberger and O'Connell 1998; Steinberger 2000; Hassan *et al.* 2016; Arnould *et al.* 2019) causing the locations of hotspots on the Earth's surface to wander over geological time.

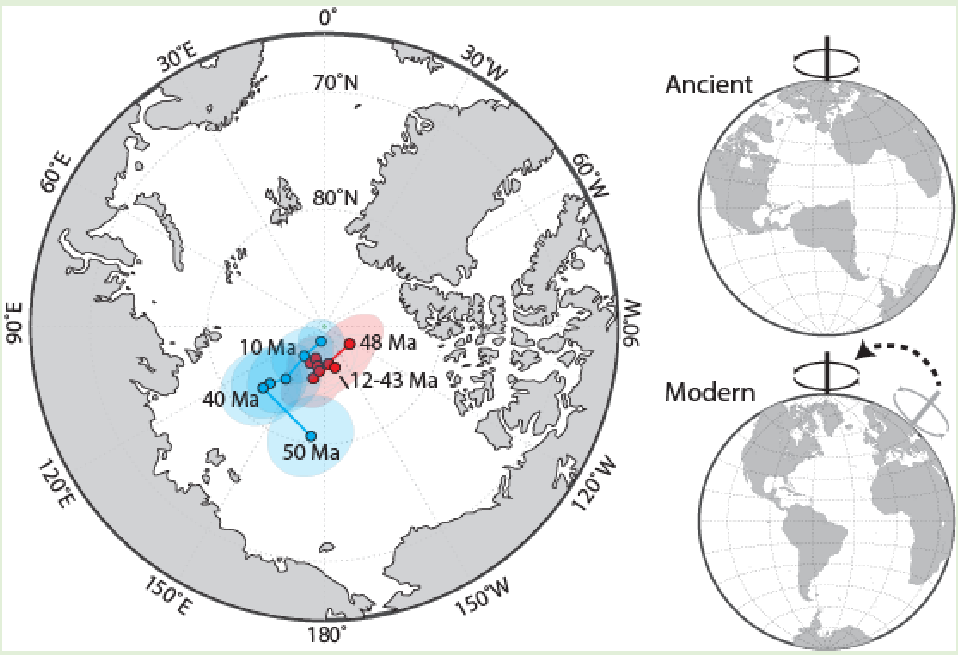
Box 2: Scenarios for the HEB Formation. The prominent 120° bend in the Hawaii-Emperor seamount trail (HEB) was first interpreted as representative of a major change in Pacific plate motion (Duncan and Clague 1985) occurring around 47 million years ago (Sharp and Clague 2006). However, an absence of geological evidence for a change in Pacific plate motion at that time, and recognition that hotspots are mobile, led to the proposal that the HEB may rather represent Hawaiian hotspot motion, which came to a rather abrupt stop at the time of the bend (Norton 1995). This hypothesis was tested through the analyses of magnetic paleolatitudes in scientific ocean drilling cores recovered from Emperor seamounts, which support a significant southward hotspot motion between ~80 and 47 Ma (Tarduno and Cottrell 1997; Tarduno *et al.* 2003; Bono *et al.* 2019). Konrad *et al.* (2018a) compared the relative distances of Pacific hotspots using age progressions along three tracks (Hawaii, Louisville, Rurutu). The results show the Hawaiian hotspot moving southward, with most of the motion occurring between 62 and 47 million years ago, but with a maximum ~600-700 km of motion detected that is insufficient to explain the entire, more than 2,000 km length of the Emperor chain. There remains an active debate on the relative contribution of hotspot drift and plate motion changes to the shape of the 120° bend, with scenarios ranging from the HEB being entirely caused by a ~60° change in plate motion (Steinberger *et al.* 2004; Torsvik *et al.* 2017) to being mostly caused by hotspot drift (Bono *et al.* 2019; Hassan *et al.* 2016). The map shows modeled 0-70 Ma tracks for a ***fixed*** Hawaii hotspot and three plate reconstructions that all use the Antarctic plate circuit. Matthews *et al.* (2016) and Tetley *et al.* (2019) use relative motions in Zealandia as additional constraints, whereas Seton *et al.* (2012) do not. A large body of work is suggesting a major plate reorganization at ~50 Ma centered around the Pacific Plate (e.g. Cosca *et al.* 1998; Meffre *et al.* 2012; Reagan *et al.* 2019), yet these plate motion models account for at most 35° of the expected 60° absolute plate motion change and cannot accurately predict the location and age of seamounts in the Hawaii-Emperor track prior to 30 million years ago. The mismatch between the predicted 47 Ma hotspot locations and the actual location of the bend (Daikakuji Seamount) is consistent with ~1 cm/yr SE hotspot motion (green arrow) from 47-0 Ma obtained by numerical models (Steinberger *et al.* 2004; Hassan *et al.* 2016). The mismatch between the predicted 61 Ma hotspot locations and Suiko Guyot can be amended by additional rapid southward hotspot motion (long orange arrow) that is consistent with age progressions along the hotspot tracks (Konrad *et al.* 2018a) and numerical modeling (Hassan *et al.* 2016). In this debate, however, it appears that a large portion of the length of the Emperor chain only can be explained by southward Hawaii hotspot drift.



For the most recent 5 million years in Earth's history, the rate of motion of major hotspots was recently computed by Wang *et al.* (2018) using a maximum likelihood optimization that incorporated present-day plate motion models and compared those to the azimuths and age progressions of hotspot tracks (Morgan and Phipps Morgan 2007). They obtained highly variable hotpot motion rates between 2.5 mm/yr (Afar) and 49.4 mm/yr (Caroline) with Hawaii moving at 11.6 mm/yr. In this model, the Pacific hotspots were found to move at speeds between 10 and 50 mm/yr, while Atlantic and Indian Ocean hotspots would move more slowly, below 20 mm/yr. An alternative kinematics-based approach was recently developed by Tetley *et al.* (2019) to determine the motions of major hotspots for the last 80 Ma by estimating misfits while fitting hotspot tracks and treating all hotspots as fixed. The resulting hotspot trail misfits represent a robust estimate of hotspot motion, with rates generally below 40 mm/yr for all major hotspots other than Hawaii (Tetley *et al.* 2019). However, using other modeling approaches, the absolute plate motion reference frame for the present-day yields similar net rotations (i.e. wholesale lithospheric spin of all tectonic plates with respect to a fixed lower mantle) for fixed or moving hotspot assumptions (Becker *et al.* 2015). Likewise, other authors have used similar data, but only considering the azimuth of hotspot trends, to conclude that there is no requirement for hotspots at the present day to move at all, and that a fixed hotspot reference frame may fit the data better (Wang *et al.* 2019). These contrasting papers exemplify the extent of the ongoing controversy about hotspot motion.

Importantly, these scenarios now also can be compared to geodynamic models of hotspot motion. Steinberger and O'Connell (1998) pioneered an approach based on large-scale mantle flow models (see also Steinberger 2000; Doubrovine *et al.* 2012) in which mantle flow is computed based on density models inferred from 3D seismic tomography, assumed whole mantle viscosity profiles, and time-dependent plate motions as surface boundary conditions. The approach is completed by inserting initially vertical plume conduits that then are advected while buoyantly rising through the convecting mantle. In these geodynamic models the computed hotspot motions are typically slower than plate motions, mostly less than 1 cm/yr. This slow motion foremost is a consequence of plumes being anchored in, and rising from the lower mantle, which is only sluggishly convecting because of its high viscosity up to about 10^{23} Pa s. This geodynamic modeling approach is most reliable in the Cenozoic, but less so for earlier times, as backward-advection of mantle density anomalies becomes increasingly unreliable. The gap is now being filled with geodynamic forward models that can capture both the tilt of plume stems in the mantle as well as the motion of the plume generation zone (Hassan *et al.* 2015; 2016).

Box 3: True Polar Wander. Motion of the magnetic north pole as seen from a specific tectonic plate may be caused by plate motion and/or by re-orientation of the entire Earth relative to the poles. The former is termed apparent polar wander, the latter true polar wander. For approximately the last 120 million years, plate motion can be determined from hotspot tracks, allowing a “true polar wander path” to be reconstructed by rotating a single plate back to its past locations in the hotspot reference frame, and transferring the corresponding paleomagnetic pole positions with it. The remainder trace of past pole positions then in theory equals the true polar wander path. In practice, true polar wander paths for different plates differ somewhat, due to non-dipolar components of the magnetic field and other uncertainties (e.g. Besse and Courtillot 2002; Torsvik *et al.* 2012), but they can be combined to devise a global true polar wander path. Two modeled paths with uncertainty ellipses representing 95% confidence by Doubrovine *et al.* 2012 (blue) and Woodworth and Gordon 2018 (red) are shown for the last 50 million years relative to respectively global and Pacific hotspot reference frames, showing only minor true polar wander in this time frame. As true polar wander is expected from significant shifts in Earth’s mass distribution (Gold, 1955) on geological timescales, it is likely to result from reconfigurations in large-scale mantle structures (Steinberger and O’Connell 1997). The two LLSVP superstructures are thought to be stable for hundreds of millions of years (Burke and Torsvik 2004; Torsvik *et al.* 2006; Burke *et al.* 2008; Dziewonski *et al.* 2010) and we can infer that the two corresponding antipodal geoid highs are also stable for the same amount of time. Since the spin axis remains aligned with the minimum point of the spherical harmonic degree two geoid component—in between those two geoid highs—it is expected that true polar wander occurs in a “ring” around the Earth that runs in between the two LLSVPs and that it is mostly driven by the sinking of “heavy” lithospheric slabs subducting for a large part near that ring (Steinberger *et al.* 2017). The largest amount of true polar wander since 120 Ma that can be determined occurred at speeds in the order of 10 mm/yr, mostly between 100 and 110 million years ago, and closely following a direction going around that ring. True polar wander in any other period is less expected and less prominent.



6.2 Breaking Up Continents

Mantle plumes may play an active role in the breakup of continents and/or oceanic lithosphere, when they impinge those from beneath, providing a starting point for the Wilson Cycle in plate tectonics and a likely important initiation mechanism for far-field major plate reorganizations. Continental breakup occurs when enough extension occurs to split continental lithosphere and form a new ocean basin. Many of the continental flood basalt provinces in the Mesozoic and Cenozoic are closely related in time and space with continental breakup (Morgan 1971; Storey 1995; White and McKenzie 1989; Courtillot *et al.* 1999). When reconstructed back to their original plate tectonic configuration, flood basalt provinces are often found along continental margins, and dike swarms have more or less radial patterns and terminate at the margins of continental cratons (Burke and Dewey 1973; Fahrig and Schwarz 1973).

These observations led to the active continental rift hypothesis, where mantle plumes are thought to drive continental breakup. In this model, continental rifting is actively driven by mantle plume processes, including arrival of a plume head at the base of the lithosphere, heating and erosion of the lithosphere, and heating, uplift and tensional failure of the mechanical lithosphere (Courtillot *et al.* 1999). Alternatively, continental breakup may be explained by the passive rifting model, where continental rifting is driven by far-field tectonic forces and interaction with an existing plume is co-incidental. In both scenarios, the combination of lithospheric extension and upwelling plume material causes abnormally large volumes of magma to erupt through passive decompression melting (White and McKenzie 1989). It also appears that most continental rifts without a plume failed, with only one third proceeding to break-up (Ziegler and Cloetingh 2004). The fact that continental rifting can extend from only a few million years up to around 100 million years before progressing to continental breakup, therefore, has called for a combination of passive and active plume head forces (Courtillot *et al.* 1999).

For continental rifts that do proceed to breakup, plumes appear to play a triggering, but not an essential role. The role for plumes is particularly striking when the timing of continental rifting, flood basalt eruption, and continental break up are compared, with continental rifting often extending for tens of millions of years and—after this prolonged period—ending with the voluminous eruption of flood basalts that are coevally, or closely followed by, continental breakup (Buiter and Torsvik 2014). Brune *et al.* (2013) tested the role of plumes as a trigger for continental breakup by modelling lithosphere under far-field extension, investigating the role of plume-related lithospheric erosion, finding that plume erosion decreases lithospheric strength and controls the timing or even occurrence of continental breakup.

Finally, mantle plumes, along with changes in plate boundary forces (Whittaker *et al.* 2016, Gaina *et al.* 2009) and “wrench” tectonics (Nemcok *et al.* 2016) also are thought to drive the formation of microcontinents, small continental fragments that become separated from their parent continental margin and eventually surrounded by oceanic lithosphere. In these smaller scale manifestations of plate tectonics, mantle plumes have been implicated in the rifting and separating of microcontinents from relatively young continental margins, less than 25 million years old, through one or multiple mid-ocean ridge relocations that are centred around the mantle plume, resulting in ocean spreading asymmetries and ridge propagation towards the locus of the plume (Müller *et al.* 2001).

7. Mantle Plumes Impacting Earth’s Environment

Mantle plumes are the source of a phenomenal volume of volcanic products on the Earth’s surface (Coffin and Eldholm 1994). When mantle plumes cause hotspot volcanism, outputs are particularly high during plume head volcanism that causes a rapid outpouring of voluminous LIPs, in the oceans referred to as oceanic plateaus and on land as flood basalts. Emplacement of these LIPs is hypothesized to cause extreme global environmental perturbations and mass extinctions (Courtillot *et al.* 1988; Duncan and Pyle 1988; Renne *et al.* 2013). According to these hypotheses, flood basalt volcanism introduces large quantities of volatiles such as sulfuric acid into the Earth's atmosphere (with SO_2 forming H_2SO_4) and carbon dioxide (CO_2) release may be increased significantly, resulting in greenhouse scenarios (Kerr 2005; Self 2006; Kerr 2014). In contrast, introduction of volcanic ash and aerosols in the Earth's atmosphere may cause shielding from solar radiation, leading to the cooling of global climate (Self 2006). Each of these processes may induce strong perturbations in the environment and trigger tipping points in Earth's climate state, with the increased extinction of species as one of the major outcomes. Geodynamic simulations also show that mantle plumes may have accumulated oxidized materials in the upper mantle and with that may have influenced the evolution of atmospheric oxygen (Gu *et al.* 2016).

However, it remains hard to undeniably link the causes and chronologies associated with global extinction events to mantle plume-induced eruptions. For example, emplacement of the Deccan Traps flood basalt seems to have started a few tens of thousands of years prior to the extinction at the K-Pg boundary around 66 million years ago, but it is now also apparent that more than 75% of Deccan Trap volcanism occurred following the Chicxulub meteorite impact (Renne *et al.* 2013; Schoene *et al.* 2019; Sprain *et al.* 2019). Similarly, emplacement of the Siberian Traps was extremely rapid, causing massive volcanism over a period of less than ten thousand years, but geochronological techniques lack resolution to definitely tie this flood basalt to the largest Earth extinction event at the Permo-Triassic boundary around 250 million years ago (Courtillot and Renne 2003; Reichow *et al.* 2009; Sobolev *et al.* 2011; Courtillot and Fluteau 2014). Even though plume-related volcanism resulting in voluminous flood basalts are likely to have direct impacts on global climate, from the Chicxulub impact event it appears that there are additional primary drivers that could tip over Earth's climate system and cause massive extinctions (Hull *et al.* 2020).

In the oceans, the impacts of large-scale mantle plume eruptions on the environment are different, yet allegedly they also cause widespread species extinction. These submarine eruptions may introduce toxic metals that poison marine life, or they may provide high levels of nutrients that cause planktonic blooms that then massively deprive the oceans of oxygen after these organisms die off and decompose (Sinton and Duncan 1997; Kerr 2005). These episodes are recognized as 'oceanic anoxic events' that are chronicled in marine sedimentary records through the deposition of black shales. The timing of ocean plateau formation relative to the formation of those globally dispersed black shales is still highly uncertain. Future scientific ocean drilling and modern-day geochronology is required to uniquely tie the formation of mantle plume-derived oceanic plateaus to these deadly global oceanic anoxic events.

8. Future Mantle Plume Research

Even though mantle plumes are sometimes considered independent of the key plate tectonic processes, such as plate spreading and subduction, mantle plumes in fact are an integral part of a dynamic Earth.

Mantle plumes are part of the overall 'rock cycle' in the Earth system, playing a key role in Earth's overall convectional regime, and in the continuous recycling of Earth's deep interior and surface materials. Many mantle plumes are originating from the deepest regions in Earth's mantle, often in association with LLSVPs and ULVZs, and plumes potentially even take in materials from the Earth's core itself. On the other end, at Earth's surface, we see clear evidence of plume activity, in intra-plate volcanism, continental break-up, and long-term effects on Earth's climate system. Yet, our knowledge of mantle plumes remains limited, largely because of the low-resolution view we have of their structures and behaviors through seismic mantle tomography studies, and because of the complexity of plume expressions in the volcanic structures (seamount trails, oceanic plateaus) on Earth's surface. Below we compile a short list of future research themes where we can improve our understanding of mantle plumes themselves and their effects on the interconnected Earth system.

Mantle Plume and LLSVP Imaging: Reconciling current seismic images of "mantle plumes" with the geodynamic modeling and surface observations is THE challenge for the future. Large-scale deployments of OBS (ocean bottom seismometers) with wider apertures are required to create detailed views of the deep portions of mantle plumes. For example, we need to improve on the geometries and placement of OBS instruments and design them to capture specific seismic phases that can better resolve structures at the roots of mantle plumes and shed light on their potential generation in the LLSVP and ULVZ regions.

Thermal versus Thermochemical Plumes: What is the nature of the plumes that now start to get resolved by tomography in the lower mantle? Are they purely thermal or thermo-chemical and how do they relate to a potential basal dense layer? Do we need to explore more complex rheologies to explain their morphologies and temporal behaviors? We also need to understand how these plumes may interact with mantle flow and get deflected primarily in the 660-1,000 km depth range.

Dynamic Topography: The upwelling of mantle plumes may affect the dynamic topography of oceanic lithosphere (Conrad *et al.* 2004; Li and Zhong 2009; Zhang *et al.* 2010; Poore *et al.* 2011; Parnell-Turner *et al.* 2014; Steinberger *et al.* 2019a). Where the presence of strong or weak mantle plumes may cause large or small hotspots swells (King and Adam 2014) and regional modifications in mantle viscosity profiles, the 'pulsating' behavior of mantle plumes may cause significant variations in local sea level that may change ocean circulation patterns and deep-sea sedimentation accumulation rates (Parnell-Turner *et al.* 2014). Gaps in understanding mantle viscosity profiles, in particular in the proximity of larger plume swells, are causing major uncertainties in modeling future sea level rise across the world (Müller 2010; Petersen *et al.* 2010; Rovere *et al.* 2015).

Oceanic LIPs: The formation and history of oceanic LIPs are still poorly understood compared to their extensively studied onshore counterparts. There is an intriguing association with spreading ridges and triple junctions, which now appears to be archetypal for plumes in the ocean basins. The recent work on Shatsky Rise (Sager *et al.* 2013; 2019) reinforces the idea that some LIPs might not be caused by plume head eruptions, which raises interesting questions whether Shatsky is a headless plume interacting with the overriding oceanic lithosphere.

Spherical Shell 3D Earth Modeling: An exciting and growing field of mantle geodynamics involves the generation of three-dimensional spherical shell numerical models of the Earth that solve the governing equations for appropriate physical parameters. These models can evolve through geological time allowing us to study the interactions between tectonic processes, large-scale mantle flow, and plumes (e.g. Zhong *et al.* 2000; Zhong 2006; Tackley 2008; Arnould *et al.* 2019; 2020). The simulation output can then be used to interpret complex surface and seismic tomographic features. With advancements in supercomputing, it is expected that this style of 3D modeling will continue to advance and provide important insights into Earth's inner workings.

Low-flux Hotspots: Given the geological importance of linking chemical components observed in ocean island lava flows to underlying mantle features, future improvements in our understanding of the variable makeup of the mantle require joint geochemistry-geodynamics-geophysical research at various global hotspots. It is important to focus on other hotspot systems besides Hawaii, as that plume displays an anomalously high buoyancy flux relative to the global average for hotspots. Its high buoyancy and related melt flux have the potential to obscure geographic trends in plume-derived lava chemistry, which might be more easily discernable at low-flux hotspots (Chauvel *et al.* 2012).

Enriched and Primordial Mantle Domains: Over the last half a century geochemical studies caused a paradigm shift in our understanding of the makeup of the mantle—which represents 84% of Earth's volume—from a homogenous material body to a complex mix of mantle domains each characterized by a different heritage and geological history. We still have limited ideas on the time and length scales of these domains, where they reside, and what is their long-term stability.

Coordinating Geochemical Efforts: In terms of geochemistry, a better effort must be made to coordinate geochemical and isotopic and geochronological analyses on the same samples, in particular for multiple novel short-lived isotopic systems. For example, we have no idea how ^{129}Xe anomalies (which track the degassing history of the planet) relate to ^{182}W anomalies (which track core formation) because exactly *one* rock has been characterized for both isotopic systems, but this sample was not characterized for ^{142}Nd (which tracks silicate Earth differentiation). Understanding the earliest history of the planet will require a multi-proxy effort with coordination across laboratories.

Cyclicities in Global Intra-Plate Volcanism and Plume Life Spans: Over the last 200 million years a waxing and waning of global intra-plate volcanism (Larson 1991b) and ocean crust formation (Müller *et* *al.* 2008) is observed, but drivers and potential links between these cyclicities are unknown. Future research needs to provide process-based understanding and whether plume formation through time can be linked to coupled plate tectonic and mantle evolution, and to cycles of plate aggregation, dispersal, and subduction. Related to this is figuring out why plumes have such a long range of life spans, anything from ~30 to 150 million years and longer, and how their roles changed as a function of a decreasing internal heat production over Earth's history.

Transition Zone Plumes: The modes of rising mantle plumes and from which depths they originate is still under debate. For example, we don't know if "transition zone plumes" exist, and if so, if they form by ponding of superplumes on 660 km mantle discontinuity (e.g. Tan *et al.* 2002), if they are deflected deep plumes, or if they are the natural consequence of asthenospheric convection related to the subduction process and slab induced hydration of the transition zone (e.g. Faccenna *et al.* 2010).

Global Plume Heat Flux: Another enigma relates to figuring out whether the overall global plume flux is consistent with the balance between the heat flux across the core-mantle boundary, internal heating, and the thermal evolution of Earth.

Solid Earth, Climate and Biosphere Interactions: Mantle plume activity results in extensive volcanism on Earth's surface with impacts for its environment, and all forms of life on this planet. A key challenge is to capture the complete chronologies that record the mantle plume events that produce flood basalts and oceanic plateaus, and to tie those to other geological records—such as sedimentary records on land and in subseafloor scientific ocean drilling cores—that contain information about Earth's environment, species extinction, ocean chemistry, ocean acidification, oxygenation, and more. We need those records to help figure out how relevant plumes are in solid Earth, climate and biosphere interactions, including their impacts on long-term climate change, mass extinctions, and biosystem resiliency.

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