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Origin of the LLSVPs at the base of the mantle is a consequence of plate tectonics – A petrological and geochemical perspective

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ABSTRACT

In studying the petrogenesis of intra-plate ocean island basalts (OIB) associated with hotspots or mantle plumes, we hypothesized that the two LLSVPs at the base of the mantle beneath the Pacific (Jason) and Africa (Tuzo) are piles of subducted ocean crust (SOC) accumulated over Earth's history. This hypothesis was formulated using petrology, geochemistry and mineral physics in the context of plate tectonics and mantle circulation. Because the current debate on the origin of the LLSVPs is limited to the geophysical community and modelling discipline and because it is apparent that such debate cannot be resolved without considering relevant petrological and geochemical information, it is my motivation here to objectively discuss such

information in a readily accessible manner with new perspectives in light of most recent discoveries. The hypothesis has the following elements: (1) subduction of the ocean crust of basaltic composition to the lower mantle is irreversible because (2) SOC is denser than the ambience of peridotitic composition under lower mantle conditions in both solid state and liquid form; (3) this understanding differs from the widespread view that OIB come from ancient SOC that returns from the lower mantle by mantle plumes, but is fully consistent with the understanding that OIB is not derived from SOC because SOC is chemically and isotopically too depleted to meet the requirement for any known OIB suite on Earth; (4) SOC is thus the best candidate for the LLSVPs, which are, in turn, the permanent graveyard of SOC; (5) the LLSVPs act as thermal insulators, making core-heating induced mantle diapirs or plumes initiated at their edges, which explains why the large igneous provinces (LIPs) develop at the edges of the LLSVPs; (6) the antipodal positioning of Jason and Tuzo represents the optimal momentum of inertia, which explains why the LLSVPs are stable in the spinning Earth.

Keywords: LLSVPs, OIB petrogenesis, Subducted ocean crust, Subducted oceanic mantle lithosphere, Mantle plumes, Mantle chemical differentiation, Optimal momentum of inertia

1. The history and context

The plate tectonics theory explains plate boundary zone processes (e.g., volcanism, metamorphism and earthquakes) in simple clarity, but cannot explain intraplate volcanism such as Hawaii in the interior of the Pacific plate. Wilson (1963) called the intraplate volcanic activities like Hawaii as ‘‘hotspots” of mantle melting anomalies derived from relatively fixed deep sources. Morgan (1971) expounded that the hotspots are surface expressions of mantle plumes coming from the core-mantle boundary (CMB). This conceptual connection between intraplate ocean island volcanism and mantle plumes becomes clearer by noting the geochemical distinction between mid-ocean ridge basalts (MORB) of shallow, plate-tectonics origin and intraplate ocean island basalts (OIB) of deep, mantle plume origin. The incompatible element depleted nature of MORB is interpreted as inheriting from the shallow mantle whose depletion resulted from continental crust extraction in Earth's early history, whereas the deep mantle source for OIB, not affected by the crustal extraction event, is incompatible-element enriched (e.g., Morgan, 1972; Hofmann, 1988). Studies over the last ~ 50 years have shown, however, that OIB sources are not just simply enriched, but chemically and isotopically highly heterogeneous on all scales (e.g., Gast, 1968; Sun et al., 1975; Schilling et al., 1983; Hart, 1984; Zindler et al., 1984; White, 1985, 2010, 2015; Zindler and Hart, 1986; Castillo, 1988; Hart et al., 1992; Mahoney et al., 1994; Hanan and Graham, 1996; Hofmann, 1997; Willbold and Stracke, 2006) (Fig. 1). They are overall much more enriched in both abundances and ratios of incompatible elements than the *primitive mantle* (e.g., Niu and O'Hara, 2003; Niu, 2009; Niu et al., 2011, 2012) (Fig. 2).

The primitive mantle (PM) is a hypothetical construct representing the silicate portion of the bulk Earth (Sun and McDonough, 1989; McDonough and Sun, 1995; Palme and O'Neill, 2014). Whether any material of this composition remains within the mantle is conjectural, but it has been implicitly assumed in all models and discussions that the PM was compositionally uniform from its inception upon the core separation in the early Earth. With this assumption kept in mind, and by assuming the depleted MORB mantle depletion indeed resulted from continental crust extraction (e.g., Armstrong, 1968; Hofmann, 1988), we can endeavour to understand the origin of OIB source compositional heterogeneity. Because elemental fractionation in solid state is unlikely in the deep mantle due to extremely slow diffusion (e.g., Hofmann and Hart, 1978), we can reason that processes known to occur in the upper mantle and crust (e.g., magmatism, metamorphism, weathering, transport and sedimentation) are the likely causes of elemental fractionation, and hence isotopic variability (Fig. 1). Subduction of such shallow/near-surface processed materials into the deep mantle over Earth's history would explain the mantle compositional heterogeneity. It follows that mantle compositional heterogeneity is a consequence of plate tectonics (see Niu, 2009; Niu and O'Hara, 2003). "Mantle plumes [OIB] from ancient oceanic crust" (Hofmann and White, 1982) is one of the earliest to invoke plate tectonics as mechanism to cause OIB source compositional heterogeneity, stating "oceanic crust is returned to the mantle during subduction . . . Eventually, it becomes unstable as a consequence of internal heating [in the region above the core-mantle boundary], and the resulting diapirs become the source plumes of oceanic island basalts (OIB) and hot spot volcanism."

This interpretation has been widely accepted since then and has gained broader acceptance because of the recognition of the two large-low-shear-wave-velocity provinces (LLSVPs) at the base of the mantle beneath the Pacific and Africa (e.g., Dziewonski, 1984; Grand et al., 1997; Su and Dziewonski, 1997; Ritsema et al., 1999; Kellogg et al., 1999; Mégnin and Romanowicz, 2000). The LLSVPs have been interpreted to be excessively hot mantle domains representing the locations of "superplumes" (e.g., Romanowicz and Gung, 2002; Ni et al., 2002; Burke and Torsvik, 2004; Thorne et al., 2004; Burke et al., 2008; Burke, 2011; Deschamps, 2014; Li et al., 2014; Torsvik et al., 2014), to be physical evidence for mantle plumes derived from the core-mantle boundary, to be responsible for the geographically associated surface geoid highs or "superswells" (McNutt, 1988), and probably also for mantle isotopic anomalies identified from volcanism on these highs and their peripheral regions (e.g., Hart, 1984; Castillo, 1988; White, 2015). Recent studies suggest that the LLSVPs have sharp boundaries with, and higher density than, the ambient mantle, indicating that they are chemically distinct from the surrounding mantle (Ni et al., 2002; Becker and Boschi, 2002; Ni and Helmberger, 2003; Wang and Wen, 2004; Toh et al., 2005; Ford et al., 2006; Garnero et al., 2007; Cottaar and Lekic, 2016; Koelemeijer, 2017; Lau et al., 2017). Their origin remains to be understood and has been the subject of intense modelling efforts in recent years (e.g., McNamara and Zhong, 2004; Lassak et al., 2010; Li et al., 2014; Deschamps, 2014; Torsvik et al., 2014; Huang et al., 2015; Mulyukova et al., 2015; White, 2015; Cottaar and Lekic, 2016; Doubrovine et al., 2016; Garnero et al., 2016; Torsvik and Domeier, 2017; Romanowicz, 2017; Flament et al., 2017). They could be residual Fe-rich material during core formation or subducted ocean crust (e.g., Garnero et al., 2007; Hirose and Lay, 2008; Niu et al., 2012; Deschamps, 2014; Li et al., 2014; Mulyukova et al., 2015; Garnero et al., 2016; White, 2016; Lau et al., 2017).

In this paper, I focus on the hypothesis that the LLSVPs are of plate tectonics origin and are piles of subducted ocean crust (SOC) cumulated over Earth's history (Niu et al., 2012). I will show step by step how a petrological and geochemical study of oceanic basalts has led to this hypothesis that needs attention by the geophysical community and modelling discipline towards a genuine understanding of the nature and origin of the LLSVPs in the context of mantle dynamics and chemical differentiation of the Earth.

2. Why OIB are not from SOC?

The conclusive statement that mantle compositional heterogeneity is a consequence of plate tectonics is readily justified as shown by the geochemistry of oceanic basalts (e.g., Zindler and Hart, 1986; Niu and Batiza, 1997; Hofmann, 1997; Niu et al., 2002; Andersen et al., 2015; Castillo, 2015, 2016; White, 2016; Jackson et al., 2018) although the uniform PM assumption needs verification (Niu, 2009) and the geochemical consequences of subduction-zone metamorphism need further investigations (Niu, 2009; Xiao et al., 2012, 2013). For example, the isotopic endmembers such as EM-1, EM-2 and HIMU (Fig. 1) can indeed be explained by subduction of shallow/near-surface processed materials. Subducted terrigenous sediments with high Rb/Sr and low Sm/Nd can contribute to the EM-2 OIB with high $^{87}Sr/^{86}Sr$ and low $^{143}Nd/^{144}Nd$. Given the understanding that subduction-zone dehydration metamorphism preferentially transfers more Pb over U to the mantle wedge, it is expected that the subducted slab materials with high U/Pb can contribute to the HIMU OIB with high $^{206}Pb/^{204}Pb$, whereas the mantle wedge materials transported to mantle source regions with low U/Pb can contribute to the EM-1 OIB with low $^{206}Pb/^{204}Pb$.

However, OIB with such isotopic endmember compositions are less abundant, and more than ~ 85% OIB suites are in the domain around "FOZO" or "C" (Fig. 1). Hence, our efforts to understand the petrogenesis of OIB need focussing on such majority as well as the endmembers. In this context, we need to clearly and correctly inform the community that the standard view of "mantle plumes [OIB] from ancient oceanic crust" needs total reconsideration as repeatedly demonstrated (Niu and O'Hara, 2003; Niu, 2009; Niu et al., 2011, 2012) because this standard view is inconsistent with observations and contradicts the straightforward physical, petrological and geochemical principles.

2.1 Melting of SOC cannot produce OIB

The bulk ocean crust has basaltic composition with no more than ~ 13 wt% MgO (Niu, 1997; Niu and O'Hara, 2003), whose partial melting would produce melt with MgO < 13 wt%, whose total melting (physically unlikely) would give melt with 13 wt% and will not in any way produce melt with MgO > 15 wt% (Niu, 2005). However, the primitive Hawaiian OIB melts (quenched glasses) have MgO > 15 wt% (Clague et al., 1991), which cannot be produced by melting of present-day ocean crust, nor SOC formed in the geological history unless the mantle potential temperature had been 100's of degree hotter beneath ocean ridges, which has been proposed, but has no convincing evidence. One could argue that the Hawaiian picrite melts were indeed produced by SOC melting, and their elevated MgO resulted from interaction with the ambient peridotitic mantle. The latter is possible, but it is this possibility confirms precisely that melting of SOC alone cannot produce primitive OIB such as the Hawaiian picrite (Niu and O'Hara, 2003). Hence, in terms of straightforward petrology, SOC cannot be the source of OIB.

2.2 SOC is chemically too depleted to be the source of any known OIB suite on Earth

It has been repeatedly demonstrated that the ocean crust (OC), compositionally approximated by MORB, is depleted in incompatible elements and is more depleted in the progressively more incompatible elements as the result of inheritance from incompatible element depleted upper mantle called depleted MORB mantle (DMM; see above) with $[La/Sm]_{PM} < 1$ (~ 0.649), $[Rb/Sr]_{PM} < 1$ (~ 0.170) and $[U/Pb]_{PM} < 1$ (~ 0.442) (Fig. 2a). Melting of SOC with such incompatible element depleted composition cannot in any way produce OIB with highly enriched composition as manifested in Figure 2a (Niu et al., 2012). Relative to the OC (relative abundances about the unity), OIB are enriched in all the incompatible elements and are more enriched in the progressively more incompatible elements (towards left; Fig. 2a) with $[La/Sm]_{PM} > 1$ (~ 3.461), $[Rb/Sr]_{PM} > 1$ (~ 1.771) and $[U/Pb]_{PM} > 1$ (~ 1.116). Hence, in terms of abundances and ratios of incompatible elements, SOC is too depleted to be the source of OIB.

2.3 SOC passing through subduction-zone dehydration has elemental systematics not suitable as source for OIB

The widely accepted standard model states that the island arc basalts (IAB) result from subducting slab dehydration induced mantle wedge melting. That is, water transferred from the subducting ocean crust to the mantle wedge lowers the solidus and causes mantle wedge melting for IAB. Along with the transfer of water are water-soluble (mobile) elements such as Ba, Rb, U, K, Pb and Sr whereas water-insoluble (immobile) elements such as high field strength elements (HFSEs, e.g., Nb, Ta, Zr, Hf, Ti) are left behind in the residual SOC. This explains why IAB are not only enriched in water and water-soluble elements but also depleted in HFSEs as illustrated by the average IAB such as IAB-E03 (Fig. 2a). Consequently, it is expected that the residual SOC passing through the subducting dehydration should have elemental systematics complimentary to the IAB as qualitatively shown in Figure 2b. How can melting of such SOC produce melts with incompatible element patterns resembling OIB (Fig. 2a)? The answer is not possible.

2.4 SOC is isotopically too depleted to be the source of any known OIB suite

As illustrated in Figure 2a above, SOC must have low Rb/Sr and U/Pb, high Sm/Nd and Lu/Hf, and will, with time, produce depleted isotopes with low radiogenic $^{87}Sr/^{86}Sr$ and $^{206}Pb/^{204}Pb$, high radiogenic $^{143}Nd/^{144}Nd$ and $^{176}Hf/^{177}Hf$. This reasoning is manifested in Figure 3. Compared with the correlated trends of the global OIB, the present-day MORB (blue meshed ovals) have the most depleted isotopic compositions with low $^{87}Sr/^{86}Sr$, and high $^{143}Nd/^{144}Nd$ and $^{176}Hf/^{177}Hf$. Importantly, the present-day isotope compositions of SOC (the red meshed rectangles) subducted in the geological histories (2.5 – 1.0 Ga) are also highly depleted (see Niu and O'Hara, 2003 for straightforward calculations). Hence, SOC is isotopically too depleted as source suitable for global OIB.

2.5 OIB source materials recorded no subduction-zone dehydration signature

From section 2.3 above, we understand that it is the subduction-zone slab dehydration (metamorphism) that is ultimately responsible for the petrogenesis of IAB (e.g., Niu, 2009; Xiao et al., 2012). Along with the slab dehydration is the preferential transfer of progressively more mobile elements to the mantle wedge, resulting in these elements being progressively depleted in the residual ocean crust (i.e., SOC; Fig. 2b). The order of decreasing mobility for elements of interest here are Rb (most mobile) > Sr > Nd ≥ Sm > Lu ≥ Hf. In terms of the relevant radiogenic isotope elements, the relative mobility is 187% (Sr) > 118% (Nd) > 100% (Hf) (see Niu and O'Hara, 2003). This analysis states explicitly that the subduction-zone dehydration has the least effect on Lu/Hf (hence ε_{Hf}), moderate effect on Sm/Nd (hence ε_{Nd}) and highest effect on Rb/Sr (hence ε_{Sr}). That is, subduction-zone slab dehydration will cause uncorrelated Rb/Sr, Sm/Nd and Lu/Hf elemental and ratio fractionations. Consequently, if SOC that has passed through subduction-zone dehydration processes played an important part in OIB sources, we should not see the significantly correlated ε_{Hf}-ε_{Nd}-ε_{Sr} isotopic ratio variation in Figure 3. In other word, OIB source materials have not undergone subduction zone dehydration processes in their histories. That is, OIB preserve no evidence of their sources ever associated with subduction-zone processed materials. On the other hand, Sr, Nd, and Hf are similarly incompatible elements during mantle melting and magmatic differentiation (e.g., Niu and Batiza, 1997), suggesting the correlated isotopic variations in Figure 3 result from magmatic processes (see below).

2.6 What is the nature of OIB source materials?

This is not the focus of the paper, but the discussion is relevant and essential. The above straightforward demonstrations (see more below) deny SOC as primary source rocks for the global OIB. Then, what is the nature of OIB source rocks? We can answer this question by considering the following:

(1) OIB source rocks must be peridotitic in composition, whose melting can produce high MgO picritic melts parental to OIB (e.g., Hawaiian picrite with MgO > 15 wt%; see above);

(2) In terms of incompatible elements, the OIB source rocks are more enriched than the PM (see Fig. 3 of Niu et al., 2012), suggesting that the OIB source rocks, whether they are PM-like or previously depleted, must have been re-fertilized or re-enriched prior to mantle melting for OIB;

(3) Figure 2a shows clearly that OIB are not only enriched in incompatible elements, but are more enriched in the progressively more incompatible elements manifested by the leftward increase in the relative abundances, meaning that the OIB source re-fertilization or re-enrichment processes are magmatic (Niu and O'Hara, 2003);

(4) Such re-enrichment or re-fertilization processes are commonly termed as mantle metasomatism (e.g., Frey and Green, 1974; Sun and Hanson, 1975; Menzies and Hawkesworth, 1987; O'Reilly and Griffin, 1988) accomplished through infiltration of low-degree (low-F) melt enriched in volatiles and incompatible elements;

(5) Such metasomatism may take place in many possible locations in the mantle (Menzies and Hawkesworth, 1987; O'Reilly and Griffin, 1988; Halliday et al., 1995; Niu et al., 1996;

Donnelly et al., 2004), but with all the scenarios considered, the lithosphere-asthenosphere boundary (LAB) is the most likely locations for mantle metasomatism, which takes place beneath ocean basins and continents, and happens today and in Earth's history (Niu and O'Hara, 2003, 2009; Niu, 2008, 2009; Humphrey and Niu, 2009; Niu et al., 2011, 2012; Pilet et al., 2005, 2008; Niu and Green, 2014).

(6) The petrological evidence for mantle metasomatism is the incompatible element-enriched dike or vein lithologies (hornblendite, amphibole-, phlogopite-bearing peridotites, garnet pyroxenite and pyroxenite) carried to the surface as xenoliths by alkali and kimberlitic volcanism or exposed in massif peridotites (e.g., Frey and Green, 1974; Frey et al., 1978; Frey, 1980; Menzies and Murphy, 1980; Frey et al., 1985; Menzies and Hawkesworth, 1987; O'Reilly and Griffin, 1988; Hirschmann and Stolper, 1996; Takazawa et al., 2000; Pilet et al., 2005).

Hence, OIB source rocks are low-F melt metasomatized peridotites enriched in volatiles and progressively more incompatible elements (e.g., McKenzie and O'Nions, 1995; Niu and O'Hara, 2003; Niu, 2008; Niu et al., 2011, 2012).

2.7 The role of pyroxenites ≠ SOC

Pyroxenites are volumetrically minor but widespread rocks as dikes or veins in the mantle peridotite host (e.g., Hirschmann and Stolper, 1996) and are of metasomatic origin because of their incompatible element enriched characteristics as elaborated above. Therefore, the pyroxenites of mantle metasomatic origin can indeed contribute to the incompatible element enriched nature of OIB (see section 2.6 above).

Pyroxenites have been widely invoked in the recent literature to explain the petrogenesis of intra-plate basaltic magmatism. However, caution must be exercised because there have been widespread confusions by equating SOC with pyroxenites in defence of the idea "mantle plumes [OIB] from ancient oceanic crust" (Sobolev et al., 2000, 2005, 2007). SOC is of basaltic composition and depleted in incompatible elements (Fig. 2), whereas the pyroxenites are of peridotitic composition of metasomatic origin with elevated abundances of incompatible elements (e.g., Niu and O'Hara, 2003; Pilet et al., 2005, 2008). These authors (Sobolev et al., 2005, 2007) invoke SOC indirectly to explain the high Ni content in olivine phenocrysts of the Hawaiian OIB by proposing melting of SOC to produce SiO_2-rich melt that reacts with the harzburgite and consumes all the olivine, producing olivine-free pyroxenites as the source of Hawaiian OIB. This complex multi-stage process seemed to have convinced many, but readers should realize the fact that both SOC-derived melt (if indeed formed in the deeper asthenosphere) and harzburgite remain too depleted to explain the highly enriched geochemical nature of OIB in general and Hawaiian OIB in particular. In fact, the Ni content in the global OIB (and in the olivine phenocrysts) is largely controlled by the final depth of OIB melt extraction, i.e., the depth of the lithosphere-asthenosphere boundary (LAB) at the time of OIB eruption, which is the "lid-effect" (Niu et al., 2011). The "lid-effect" interpretation well explains the global OIB olivine Ni data and is consistent with P-T controlled Ni partitioning between olivine and melt (Matzen et al., 2013, 2017) without the need of olivine-free pyroxenites in mantle source regions of global OIB (Niu et al., 2011).

2.8 How about EM1, EM2 and HIMU OIB?

We should note that globally the volumetrically unimportant EM2 and HIMU OIB suites do not plot along the array defined by the majority of the OIB but forms a linear correlation at a high angle with the array (Fig. 3a), suggesting a possible EM2-HIMU link that may be associated with subducted materials (see above). Subducted terrigenous sediments would contribute to EM2 OIB (e.g., Jackson et al., 2007) and subducted carbonate may contribute HIMU OIB (Castillo, 2015), but neither of the two candidate materials is relevant to SOC of basaltic composition under discussion. We should also note that despite between suite variability in the abundances and ratios of incompatible elements, all OIB show similarly enriched incompatible element abundance patterns (Willbold and Strake, 2006). That is, whatever subducted materials may be for each of these isotopic endmembers, they share one thing in common, i.e. the prior source enrichment by metasomatism (Niu et al., 2012).

3. What is the fate of SOC?

Since the work by Forsyth and Uyeda (1975), it has been repeatedly tested that subducting *slab pull* is the primary driving force for plate motion and plate tectonics, and other forces such as *ridge push* is one order of magnitude less important (see Niu, 2014 for review). Slab subduction results from its enhanced density and negative buoyancy because of seafloor heat loss and thickening (thermal contraction). While it is a basic concept that the slab has the basaltic ocean crust atop the harzburgitic mantle lithosphere, many have implicitly treated the slab as the ocean crust only in their petrological and geochemical studies. Correctly, the subducting/subducted slab has compositionally distinct two lithologies, the SOC of basaltic composition and the SML (subducted mantle lithosphere) of peridotitic composition. The positive Clapeyron slope at the 410 km seismic discontinuity ($dT/dP > 0$) enhances the slab pull, whereas the negative Clapeyron slope at the 660 km seismic discontinuity (660-D; $dT/dP < 0$) hampers slab subduction (see Niu, 2014 for review), as indicated by slab stagnation above the 660-D in a number of places as a result of fast trench retreat like Pacific plate stagnation beneath eastern China (e.g., Niu, 2014). Nevertheless, seismic studies have shown that in most cases, subducting slabs can enter the lower mantle, and approach the base of the mantle (e.g., van der Hilst et al., 1997; Grand et al., 1997; Karason and van der Hilst, 2000; Li et al., 2008), offering convincing evidence in support of the whole-mantle convection scenario (e.g., Davies and Richards, 1992) and the SOC-SML segregation hypothesis at the core-mantle boundary (CMB) (Hofmann and White, 1982; Christensen and Hofmann, 1994).

Figure 4a illustrates that the down-going slabs into the lower mantle through subduction zones must be mass-balanced by lower-to-upper mantle mass transfer. However, the nature of the material and the style of the upward mass transfer need understanding. Mantle plumes are thought to be the very mechanism of such mass transfer as illustrated in Figure 4b. The seismic difficulty in detecting rising plumes (e.g., Foulger, 2005; Julian, 2005) cannot be used to argue against mantle plumes because of the present-day technique limitation in resolving small thermal contrast with the ambience and narrow size of the plumes. The mass balance can be achieved through multiple small plumes (Fig. 4b). On the other hand, for conceptual understanding of the lower-to-upper mantle mass balance, mantle plumes are not required because the mass balance can be accomplished through large scale regional "swells" without the need of mantle plumes. This is possible simply because the nature of the 660-D is the ringwoodite/perovskite phase change and its depth of 660 km remains constant regardless of large scale material movement across the 660-D (Fig. 4c). While the scenario in Figure 4c (see Niu and O'Hara, 2003) is generally not considered, it is useful for testing the mantle plume hypothesis and for discussing mantle convection (e.g., Davies, 2005; Foulger, 2005; Niu, 2005; Niu et al., 2017).

By accepting the mantle plume hypothesis, the question then is what would be the mantle plume material transferred from the lower mantle (Fig. 4)? Is it SOC or preferentially the SOC-dominated material? We have demonstrated above that SOC cannot be the major source material for OIB in terms of straightforward petrology, trace element systematics and isotope geochemistry (Figs. 2,3). The next question then is what may have happened to SOC in the context of plate tectonics and mantle circulation over Earth's history. If ocean crust of MORB composition can indeed sink to the lower mantle as illustrated by global seismic tomography studies (e.g., van der Hilst et al., 1997; Grand et al., 1997; Karason and van der Hilst, 2000; Li et al., 2008), SOC will not return in bulk to the upper mantle source regions of oceanic basalts because it is denser than the ambient mantle at all depths greater than ~ 700 km (Fig. 5; Ono et al., 2001; Niu and O'Hara, 2003; Hirose et al., 2005; Niu et al., 2012). For example, at the depth of 780 km, SOC of MORB composition will be > 2.3 % denser than the ambient peridotitic mantle. This density difference is rather significant as this is equivalent to temperature effect of

~ 700K for a volume thermal expansion coefficient of ~ 3 x 10^{-5} K^{-1}. One could argue that SOC with lower solidus may melt in the lower mantle and the SOC-derived melt maybe adequately buoyant to rise as mantle plumes to the upper mantle source regions feeding OIB, but this cannot be true because such melt of MORB composition has far greater density than that of the solid SOC and any other material towards the deep lower mantle conditions, by ~ 15% denser than the ambient mantle of peridotite composition (e.g., the PREM; Fig. 5b). These mineral physics data led Niu and O'Hara (2003) to conclude that subduction of SOC to the lower mantle is irreversible:

> "*… … oceanic crust subducted into the lower mantle will not return in bulk to the upper mantle because of the negative buoyancy in both solid and liquid states. Transfer of basaltic crust to the lower mantle would be an irreversible process. This supports the argument for a hidden component deep in the lower mantle that has not been sampled by known volcanism … … and would also lead to chemical stratification of the mantle with the mean composition of the lower mantle becoming progressively enriched in residual ocean crust lithologies (i.e., compositionally lower in Ca/Al, and higher in Fe/Mg, Si/Mg, Al, and water insoluble incompatible elements such as Ti, Nb, Ta, Zr, and Hf). If subduction of oceanic crusts into the lower mantle has continued for some time, then a large compositional contrast in terms of these elements must exist between the upper and lower mantle.*"

Hence, it is reasonable to consider that once the oceanic lithosphere subducted into the lower mantle, its SOC portion will not *in bulk* return to the upper mantle, and thus contribute to the lower-versus-upper mantle compositional differentiation. We should note, however, that small SOC debris could be entrained in the mantle upwellings, but this is physically insignificant. Therefore, the fate of the SOC is largely its irreversible storage in the lower mantle.

4. The origin of LLSVPs

It follows that SOC accumulated over earth's history is the best candidate for the two LLSVPs (Niu et al., 2012). Seismic velocity variation and waveform analysis indicate that the LLSVPs have sharp boundaries with, and higher density (~ 2-5%) than, the ambient mantle, indicating that they are chemically distinct from the surrounding mantle (Ni et al., 2002; Becker and Boschi, 2002; Ni and Helmberger, 2003; Wang and Wen, 2004; Toh et al., 2005; Ford et al., 2006; Garnero et al., 2007). This is consistent with SOC having distinct composition, lowered seismic velocity and high density relative to the ambient mantle of peridotite composition (e.g., PREM; Fig. 5). The recent tidal tomography study (Lau et al., 2017) confirms the results of seismic tomography that the LLSVPs are chemical anomalies denser than ambient mantle, but argues for only 0.5% denser, which is less than previous estimates, suggesting large uncertainties exist for further improvements. Nevertheless, the LLSVPs are not excessively hot thermal anomalies, questioning the argument for their being sources of superplumes. Some studies suggest the possibility that the LLSVPs could be Fe-rich materials from the core or residues of the core separation in Earth's early history because of the high density (e.g., Garnero et al., 2007; Hirose and Lay, 2008; Garnero et al., 2016; Lau et al., 2017), but how to test this hypothesis may be forever challenging. In this context, it is important to emphasize that the proposal of "superplumes" in 1980s came from the interpretation of the LLSVPs as huge hot thermal anomalies, but as we know now that LLSVPs are not thermal anomalies, the term "superplumes" no longer has significance. Furthermore, super upwellings would require super downward flows, but the latter is not observed nor physically straightforward to understand.

In the following, I focus on several key questions concerning the SOC hypothesis for the LLSVPs.

4.1 Can SOC be volumetrically adequate for the LLSVPs?

Figure 6, adapted with minor modification from Torsvik et al. (2017), is the best illustration of the LLSVPs at the base of the mantle in the context of mantle dynamics (Fig. 6a) and their surface projections together with the locations of known large igneous provinces (LIPs; Fig. 6b). Burke et al. (2008) estimated the LLSVPs to be ~ 2.0 wt% of the mantle. We can conduct a simple exercise in terms of SOC accumulated over Earth's history by assuming (1) plate tectonics began at ~ 4.0 Ga (unknown; see Stern, 2007), (2) the oldest ocean crust may not be older than 200 Myrs before subduction as is the case at present, (3) oceanic crust has always occupied 65% of the Earth's surface area (vs. 35% area for the continental crust), (4) the average thickness of the ocean crust is about 6 km with a mean density of 3.0 g cm^{-3} (vs. 3.3 g cm^{-3} for the mantle), and (5) all the SOC entered the lower mantle irreversibly, then the amount of SOC since 4.0 Ga would be ~ 3.0 wt% of the entire mantle. That is, there is 4 Gyrs/200 Myrs = 20 times the present-day ocean crustal mass stored in the deep mantle. The SOC stored in the deep mantle would be ~ 2.6 wt% or ~ 2.2 wt% of the mantle mass if the plate tectonics began 3.5Ga and 3 Ga, respectively. There have been several alternative estimates involving more complex assumptions that give variably greater values. For example, Helffrich and Wood (2001) estimated that ~ 5 wt% SOC is randomly distributed throughout the entire mantle.

Given the likely uncertainties in these estimates I do not wish to overstate the significance of my calculations, but they do suggest that the LLSVPs can indeed be explained by SOC in terms of both volumes and masses in the mantle because of plate tectonics. They also offer an independent line of reasoning in support of the concept that mantle plumes do not originate from SOC that is too dense to return in bulk to the upper mantle source regions of oceanic basalts (Niu & O'Hara, 2003).

4.2 The "buoyant" superswells overlie the dense LLSVPs - a contradiction or natural consequence?

The geoid highs, superswells (Fig. 6a), and reduced seismic velocity in the lower mantle are all consistent with mantle "superplumes" initiated close to the CMB or lower mantle beneath the Pacific and Africa. However, because the LLSVPs at the base of the mantle beneath these two regions are *not buoyant*, but rather dense (> 0.5-2% denser than the ambient mantle), it is difficult to understand how they could be the source of the "superplumes" (see above). This apparent contradiction points to a physical problem – how could this relationship be possible in terms of basic physics? Models that apply an imposed plate history can result in focusing of the subducted materials into piles at the core-mantle boundary beneath Africa and the Pacific, corresponding to the two LLSVPs (e.g., Garnero et al., 2007). These models are thought to be able to explain the locations of the two LLSVPs, but still cannot explain the observation that the "massive upwellings" inferred from the superswells are underlain by regions of dense materials of huge negative buoyancy. The "superplume" models suggest that the negative buoyancy of the dense materials can be overcome by thermal buoyancy (cf. Garnero et al., 2007), thus leading to the upwelling of the "superplumes". However, it seems unlikely that thermal buoyancy can be sufficient to overcome the ~ 0.5-2% density contrast (Niu and Batiza, 1991b,c), especially because of the very low thermal expansion coefficients at the deep lower mantle conditions (Birch, 1952; Anderson, 2004).

Figure 5b shows that under lower mantle conditions, the ocean crust with MORB composition is significantly denser (~ 3.6%) than the ambient mantle (PREM) and will tend to sink. By contrast, the mantle lithosphere portion must be similar to or even less dense than the ambient mantle (PREM) because it contains: (1) a thick section (< 30 to 60 km) of MORB melting residues beneath the crust that are less dense (Niu, 1997), (2) water because of incomplete subduction-dehydration (Niu, 2004), which is hosted in serpentines before subduction (developed near ridges and at the trench-outer rise), in DHMS phases during subduction, in perovskite in the lower mantle, in ringwoodite/wadsleyite in the Transition Zone, and in olivine in the upper mantle (or a vapour phase, causing incipient melting), which reduces both the bulk-rock density and the elastic moduli (e.g., Frost, 1999; Litasov et al., 2003; Mao et al., 2008a.b; Ye et al., 2009; Li et al., 2009; Jacobsen et al., 2008), and (3) re-fertilized volatile-rich deep portions of the oceanic lithosphere (Niu and O'Hara, 2003, 2009; Niu, 2008, 2009; Niu et al., 2011, 2012). This means that the subducted slab will likely separate into the dense crust (SOC) that sinks to the base of the mantle to form the LLSVPs and the buoyant mantle lithosphere of the slab (SML) that remains at shallower mantle depths. It is thus possible that the dense LLSVPs (SOC) could be overlain by such buoyant SML (Fig. 7). Because the SML is ~ 15 times the mass of the SOC, it can occupy much of the lower and perhaps part of the upper mantle, and because it is buoyant relative to the ambient mantle (e.g., PREM), it will result in a surface manifestation – the geoid highs and superswells in the Pacific and the otherwise "poorly-understood high elevation" of the African continent.

The above scenario of the buoyant SML overlying the dense SOC is hypothetical and speculative but does explain the observations. Hence, we must not ignore this possibility, but give adequate attention in modelling the scenario in terms of mantle convection. One observation that needs consideration is the fact that the two LLSVPs beneath the Pacific (Jason) and Africa (Tuzo) are antipodal (Fig. 6; Torsvik et al., 2014), and their centres of mass (or centres of gravity) may also be antipodal and equatorial (?). This may offer clues for better understanding their distribution in the context of mantle circulation and Earth's rotation despite the potentially very slow movement in response.

4.3 Why are the positions of the LLSVPs stable?

Burke and Torsvik (2004) found that 24 active hotspot volcanoes can be projected to the ΔVs = -1% contours along the edges of the LLSVPs (Fig. 6b; also see Thorne et al., 2004; Garnero et al., 2007; Torsvik et al., 2014). Burke et al. (2008) further showed that all LIP (large igneous province) eruption sites with ages < 300 Ma lie above the ΔVs = -1% contours at the LLSVP edges, which they called plume generation zones (PGZs), suggesting that the LLSVPs may have remained unchanged for at least the past 300 Myrs, which is confirmed recently (Torsvik et al., 2014). To emphasize the stable nature of the two LLSVPs, Burke (2011) named the LLSVP beneath Africa as Tuzo (abbreviated form **T**he **U**nmoved **Z**one **O**f Earth's deep mantle) in honour of Tuzo Wilson and the LLSVP beneath the Pacific as Jason (abbreviated from **J**ust **A**s **S**table **ON** the opposite meridian) in honour of Jason Morgan. I reason that the stability of the two LLSVPs is expected:

(1) if they are indeed piles of SOC because ocean crust of MORB composition is much denser than the ambient materials at the base of the mantle in both solid state and liquid form (see Fig 5; also Fig. 7 of Niu and O'Hara, 2003), and if my calculations are reasonable that by 300 Ma, the LLSVPs would already have ~ 95% of their present mass/volume.

(2) Importantly, if the centres of mass of Jason and Tuzo are antipodal (see above; Fig. 6), their present-day positions are expected to be stable and have been stable for some time to maintain the stable or uniform spinning of the Earth. In other words, the antipodal positioning means that the mass center of Jason and Tuzo combined (i.e., "Jason+Tuzo") must align with the axis of the spinning Earth. If this is indeed the case, both Jason and Tuzo would be permanently stable as this configuration represents the optimal momentum of inertia in the spinning Earth. If the antipodal configuration is proven to be equatorial (Fig. 6), the LLSVPs would be even more stable because the mass center of Jason and Tuzo combined would coincide with the center of the spinning Earth.

(3) From the above, I can further reason that (a) mantle convection will not be able to move the two LLSVPs; (b) future SOC would continue to contribute to both Jason and Tuzo regardless of where the shallow subduction zones may be; and (c) why the SML overlays the LLSVPs and cause the geoid highs or "superswells" (Fig. 7) may also result from the same effect of optimizing the momentum of inertia in the spinning Earth. All this need considering in modelling mantle circulation.

4.4 Thermal insulation of the LLSVPs defines the locations of mantle diapirs for LIPs

It would be puzzling why LIP eruption sites should correspond to the edges (vs. centres/interiors) of the LLSVPs if the LLSVPs were huge thermal anomalies feeding mantle plumes, but it is in fact not surprising at all because the LLSVPs are not thermal anomalies, but chemically distinct piles of the SOC. Burke et al. (2008) speculated that "hot material that has been heated by conduction from the core in the basal part of the slab graveyard may be driven toward the PGZs (i.e., the edges of the LLSVPs) by slabs or slab fragments acting like push brooms." My explanation is much simpler as shown in Figure 7. The LLSVPs of SOC construct act as thermal insulators preventing heat transfer from core to the mantle, making the core-to-mantle heat transfer effectively focussed at the edges of the LLSVPs, causing localized instability (or Rayleigh-Taylor instability) and initiation of mantle diapirs dominated by compositionally buoyant SML (vs. SOC) materials (Fig. 7). Water contained in the SML may facilitate partial melting in this hot setting, and because peridotite melt is ~ 12% less dense than the ambient mantle (e.g., PREM, see Fig 5). Such partially molten peridotitic packages will readily rise as plumes, explaining why major hotspot volcanoes and LIP eruption sites correspond to the edges (vs. the interiors) of the LLSVPs. Although rising plumes are thought to be in solid state (e.g., Campbell, 2005; Niu et al., 2017), melting of wet peridotitic compositions at such deep mantle conditions is possible and can help overcome the viscosity for ascent. Some "small" volcanoes do occur on the topographic highs or swells, for which my interpretation is given in Figure 7 (represented by "g"). Predictably, these volcanos would not be so hot compared with those derived from the hot thermal boundary layer at the CMB, i.e., LLSVP-edge derived LIPs or plumes.

The OIB source materials are thus likely peridotitic in composition and could include the recycled deep portions of ancient oceanic lithosphere of metasomatic origin (i.e., SML or portions of SML; see Figs. 5,7). The SOC is likely forever stored at the base of the mantle so long as the plate tectonics continues, even though we do not rule out the possibility that volumetrically insignificant debris of SOC could be incorporated in the rising diapirs.

4.5 The ULVZ are layers of SOC melt

The observation that the ultra-low velocity zones (ULVZ) are confined beneath, or in the vicinity of, the LLSVPs above the CMB (e.g., Garnero et al., 1998; Williams et al., 1998; Torsvik et al.,

2014) is consistent with the ULVZ being layers of SOC melt as the result of core heating. Such melt is too dense to rise (Figs. 5-7).

5. Summary

In this paper, I have elaborated my hypothesis on the origin of the LLSVPs at the base of the mantle beneath the Pacific and Africa as summarized below:

(1) The LLSVPs are SOC accumulated over Earth's history since the beginning of the plate tectonics and when the subducting slab began to enter the lower mantle across the 660-D.

(2) The statement (1) above is likely to be correct or at least reasonable because it is based on sound experimental observations that SOC has significantly greater density in both solid state and liquid form than the ambience under lower mantle conditions, and thus SOC will not in bulk return to the upper mantle.

(3) The statement (2) above means that subduction of SOC into the lower mantle is largely irreversible, leading to the irreversible upper-to-lower mantle compositional differentiation with the lower mantle progressively more enriched in the SOC chemical compositions.

(4) The LLSVPs act as thermal insulators, making the core-to-mantle heat transfer by conduction effectively focused at the edges of the LLSVPs, where thermal diapirs dominated by SML material develop and rise as plumes, explaining why LIPs are projected to the LLDVP edges.

(5) Because it cannot return in bulk to the upper mantle, SOC cannot physically be the source material for OIB. In fact, chemically and isotopically SOC is too depleted and cannot in any way meet the requirement of any known OIB suite on Earth.

(6) The LLSVPs have been stable over the past ~ 300 Myrs because (a) > 95% of their mass had already been accumulated by ~ 300 Ma, and (b) Jason and Tuzo are antipodal. The antipodal positioning means that the mass centre of both Jason and Tuzo combined ("Jason+Tuzo") aligns with the axis of the spinning Earth with the optimal momentum of inertia. This explains why the two LLSVPs stably stay where they are in the spinning Earth.

(7) The statement (6) above further means that the two LLSVPs will not be moved by mantle convection but could drift slowly in response to Earth's spinning characteristics if any over some long time. Future SOC would continue to contribute to both Jason and Tuzo regardless of where the shallow subduction zones may be.

(8) The scenario of buoyant SML overlying the dense SOC (LLSVPs) explains the observations of the high geoids above Jason in the Pacific and the elevated African continent above Tuzo. Such configuration may also result from the same effect of optimizing the momentum of inertia in the spinning Earth. Hence, all these observations and physical reasoning need considering in models of mantle dynamics and convection.

(9) I consider that the hypothesis presented here based on petrological, geochemical and experimental studies as well as physical reasoning and logical analysis is testable by using improved techniques to resolve whether the LLSVPs are indeed dominated by SOC. If the physics community can design experiments to verify the presence of gravitational waves and their interferences, I am optimistic for such testing with further advanced geophysical means.

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References

Agee, C.B., 1988. Crystal-liquid density inversions in terrestrial and lunar magmas: Physics of the Earth and Planetary Interiors 107, 63-74.

Albarede, F., 1995. Introduction to Geochemical Modeling, Cambridge University Press, New York, 1995.

Andersen, M.B., Elliott, T., Freymuth, H., Sims, K.W.W., Niu, Y.L., Kelley, K.A., 2015. The terrestrial uranium isotope cycle. Nature 517, 356-359.

Anderson, D.L., 2004, Simple scaling relations in geodynamics: the role of pressure in mantle convection and plume formation. Chinese Science Bulletin 49, 2017-2021.

Armstrong, R.L., 1968. A model for the evolution of strontium and lead isotopes in a dynamic Earth. Reviews of Geophysics 6, 175-200.

Becker, T.W., Boschi, L., 2002. A comparison of tomographic and geodynamic mantle models: Geochemistry, Geophysics, Geosystems 3, 2001GC000168.

Birch, F., 1952. Elasticity and constitution of the earth interior. Journal of Geophysical Research 57, 227-286.

Burke, K., 2011. Plate Tectonics, the Wilson Cycle, and Mantle Plumes: Geodynamics from the Top. Annual Review of Earth and Planetary Sciences. 39, 1-29.

Burke, K., Steinberger, B., Torsvik, T.H., Smethurst, M.A., 2008. Plume Generation Zones at the margins of Large Low Shear Velocity Provinces on the core–mantle boundary. Earth and Planetary Science Letters 265, 49-60.

Burke, K., Torsvik, T.H., 2004. Derivation of Large Igneous Provinces of the past 200 million years from long-term heterogeneities in the deep mantle. Earth and Planetary Science Letters 227, 531- 538.

Campbell, I.H., 2005. Large igneous provinces and the mantle plume hypothesis. Element 1, 265–270.

Castillo, P.R., 1988. The DUPAL anomaly as a trace of the upwelling lower mantle. Nature 336, 667-670.

Castillo, P.R., 2015. The recycling of marine carbonates and sources of HIMU and FOZO ocean island basalts. Lithos 216, 254–263.

Castillo, P.R., 2016. A proposed new approach and unified solution to old Pb paradox. Lithos 252/253, 32-40.

Castillo, P.R., 2017. An alternative explanation for the Hf–Nd mantle array. Science Bulletin 62, 974-975.

Christensen, U.R., Hofmann, A.W., 1994. Segregation of subducted oceanic crust in the convecting mantle. Journal of Geophysical Research 99, 19867-19884.

Clague, D., Weber, W.S., Dixon, J.E., 1991. Picrite glasses from Hawaii. Nature 353, 553-556.

Cottaar, S., Lekic, V., 2016. Morphology of seismic slow lower-mantle structures. Geophysical Journal International 2017, 1122-1136,

Davies, G.F, Richards, M.A., 1992. Mantle convection. Journal of Geology 100, 151–206.

Davies, G.F., 2006. A case for mantle plumes. Chinese Science Bulletin 50, 1541-1554.

Deschamps, F., 2014. Mantle plume chemical diversity. Nature Geoscience 7, 330-331.

Donnelly, K.E., Goldstein, S.L., Langmuir, C.H., Spiegelman, M., 2004. Origin of enriched ocean ridge basalts and implications for mantle dynamics. Earth and Planetary Science Letters 226, 347-366.

Doubrovine, P.V., Steinberger, B., Torsvik, T.H., 2016. A failure to reject: Testing the correlation between large igneous provinces and deep mantle structures with EDF statistics. Geochemistry, Geophysics, Geosystem 17, 1130-1163.

Dziewonski, A.M., 1984. Mapping the lower mantle - determination of lateral heterogeneity in P-velocity up to degree and order-6. Journal of Geophysical Research 89, 5929-5942.

Dziewonski, A.M., Anderson, D.L., 1981. Preliminary reference Earth model. Physics of the Earth and Planetary Interiors 25, 297-356.

Elliott, T., 2003. Tracers of the slab. American Geophysical Union Geophysical Monograph 238, 23-45.

Flament, N., Williams, S., Muller, R.D., Gurnis, M., Bower, D.J., 2017. Origin and evolution of the deep thermochemical structure beneath Eurasia. Nature Communications 8, 14164, doi: 10.1038/ncomms14164.

Ford, S.R., Garnero, E.J., McNamara, A.K., 2006. A strong lateral shear velocity gradient and anisotropy heterogeneity in the lowermost mantle beneath the southern Pacific. Journal of Geophysical Research 111, B03306, doi:10.1029/2004JB003574.

Forsyth, D., Uyeda, S., 1975. On the relative importance of the driving forces of plate motion. Geophysical Journal of International 43, 163–200.

Foulger, G.R., 2005. Mantle plumes: why the current skepticism? Chinese Science Bulletin 50, 1555-1560.

Foulger, G.R., 2010. Plates vs plumes: a geological controversy. Wiley-Blackwell, 328pp.

Frey, F.A., 1980. The origin of pyroxenites and garnet pyroxenites from Salt Lake Crater, Oahu, Hawaii, trace element evidence. American Journal of Science 80, 427-499.

Frey, F.A., and Green, D.H., 1974. The mineralogy, geochemistry, and origin of lherzolite. Geochimica et Cosmochimica Acta 38, 1023-1059.

Frey, F.A., Green, D.H., Roy, S.D., 1978. Integrated models of basalt petrogenesis: a study of quartz tholeiites to olivine melilitites from South eastern Australia utilizing geochemical and experimental petrological data. Journal of Petrology 19, 463-513.

Frey, F.A., Suen, C.J., Stockman, H., 1985. The Ronda high temperature peridotite: Geochemistry and petrogenesis. Geochimica et Cosmochimica Acta 49, 2469-2491.

Frost, D.J., 1999. The stability of dense hydrous magnesium silicates in earth's transition zone and lower mantle. Geochemical Society Special Publication 6, 283-297.

Garnero, E.J., Lay, T., McNamara, A., 2007, Implications of lower mantle structural heterogeneity for existence and nature of whole mantle plumes. Geological Society of America Special Paper 430, 79-102.

Garnero, E.J., McNamara, A.K., Shim, S.H, 2016. Continent-sized anomalous zones with low seismic velocity at the base of Earth's mantle. Nature Geoscience 9, 481-489.

Garnero, E.J., Revenaugh, J., Williams, Q., Lay, T., Kellogg, L.H., 1998. Ultralow velocity zone at the core-mantle boundary. American Geophysical Union Geodynamics Series 28, 319-334.

Gast, P.W., 1968. Trace element fractionation and the origin of tholeiitic and alkaline magma types, Geochimica et Cosmochimica Acta 32, 1055-1086.

Grand, S.P., van der Hilst, R.D., Widiyantoro, S., 1997. Global seismic tomography: A snapshot of convection in the Earth. GSA Today 7, 1-7.

Halliday, A.N., Lee, D.-C., Tommasini, S., Davies, G.R., Paslick, C.R., Fitton, J.G., James, D.E., 1995. Incompatible trace elements in OIB and MORB source enrichment in the sub-oceanic mantle. Earth and Planetary Science Letters 133, 379-395.

Hanan, B.B., Graham, D.W., 1996. Lead and helium isotope evidence from oceanic basalts for a common deep source of mantle plumes. Science 272, 991-995.

Hart, S.R., 1984. A large-scale isotope anomaly in the southern-hemisphere mantle. Nature 309, 753-757.

Hart, S.R., Hauri, E.H., Oschmann, L.A., Whitehead, J.A., 1992. Mantle plumes and entrainment: Isotopic evidence. Science 256, 517-520.

Helffrich, G.R., Wood, B.J., 2001. The Earth's mantle. Nature 412, 501–507.

Hirose, K., Lay, T., 2008, Discovery of post-perovskite and new views on the core-mantle boundary region. Elements 4, 183-189.

Hirose, K., Takafuji, N., Sata, N., Ohishi, Y., 2005. Phase transition and density of subducted MORB crust in the lower mantle. Earth and Planetary Science Letters 237, 239-251.

Hirschmann, M.M., Stolper, E.M., 1996, A possible role for garnet pyroxenite in the origin of the 'garnet signature' in MORB. Contributions to Mineralogy and Petrology 124, 185-208.

Hofmann, A.W., 1988. Chemical differentiation of the Earth: the relationship between mantle, continental crust, and oceanic crust. Earth and Planetary Science Letters 90, 297–314.

Hofmann, A.W., 1997, Mantle geochemistry: The message from oceanic volcanism: Nature 385, 219-229.

Hofmann, A.W., Hart, S.R., 1978. An assessment of local and regional isotopic equilibrium in the mantle. Earth and Planetary Science Letters 38, 44-62.

Hofmann, A.W., White, W.M., 1982. Mantle plumes from ancient oceanic crust. Earth and Planetary Science Letters 57, 421-436.

Huang, C., Leng, W., Wu, Z., 2015. Iron-spin transition controls structure and stability of LLSVPs. Earth and Planetary Science Letters 423, 173-181.

Humphreys, E.R., Niu, Y.L., 2009. On the composition of ocean island basalts (OIB): The effects of lithospheric thickness variation and mantle metasomatism. Lithos 112, 118-136.

Jackson, M.G., Hart, S.R., Koppers, A.A.P., Staudigel, H., Konter, J., Blusztajn, J., Kurz, M., Russell, M.A., 2007. The return of subducted continental crust in Samoan lavas. Nature 448, 684-687.

Jackson, M.G., T.W. Becker, T.W., Konter, J.G., 2018. Evidence for a deep mantle source for EM and HIMU domains from integrated geochemical and geophysical constraints. Earth and Planetary Science Letters 484, 154-167.

Jacobsen, S.D., Jiang, F., Mao, Z. Duffy, T.S., Smyth, J.R., 2008. Effects of hydration on the elastic properties of olivine. Geophysical Research Letters 35, doi:10.1029/2008GL034398.

Julian, B.R., 2005. What can seismology say about hotspots? Geological Society of America Special 388, 155-170.

Karason, H., Van der Hilst, R.D., 2000. Constraints on mantle convection from seismic tomography. American Geophysical Union Geophysical Monograph 121, 277-288.

Kellogg, L.H., Hager, B.H., van der Hilst, R.D., 1999. Compositional stratification in the deep mantle: Science 283, 1881-1884.

Kennett, B.L.N., Engdahl, E.R., Buland, R., 1995. Constraints on seismic velocities in the Earth from travel times. Geophysicala Journal International 122, 108-124.

Koelemeijer, P., Deuss, A., Ritsema, J., 2017. Density structure of Earth's lowermost mantle from Stoneley mode splitting observations. Nature Communications 8, DOI: 10.1038/ncomms15241.

Lassak, T.M., McNamara, A.K., Garnero, E.J., Zhong, S., 2010. Core–mantle boundary topography as a possible constraint on lower mantle chemistry and dynamics. Earth and Planetary Science Letters 289, 232-241.

Lau, H.C.P., Mitrovica, J.X., Davis, J.L., Tromp, J., Yang, H.-Y., Al-Attar, D., 2017. Tidal tomography constrains Earth's deep-mantle buoyancy. Nature 551, 321-326.

Li, C., van der Hilst, R., Engdahl, E.R., Burdick, S., 2008. A new global model for P wave speed variation in Earth's mantle. Geochemistry, Geophysics, Geosystem 9: doi:10.1029/2007GC001806.

Li, L., Brodholt, J., Alfe, D., 2009. Structure and elasticity of hydrous ringwoodite: A first principle investigation. Physics of the Earth and Planetary Interiors 177, 103-115.

Li, M., McNamara, A.K., Garnero, E.J., 2014. Chemical complexity of hotspots caused by cycling oceanic crust through mantle reservoirs. Nature Geoscience 7, 366-370.

Litasov, K., Eiji Ohtani, E., Langenhorst, F., Yurimoto, H., Kubo, T., Kondo, T., 2003. Water solubility in Mg-perovskites and water storage capacity in the lower mantle. Earth and Planetary Science Letters 211, 189-203.

Mahoney, J.J., Sinton, J.M., Kurz, D.M., Macdougall, J.D., Spencer, K.J., Lugmair, G.W., 1994. Isotope and trace element characteristics of a superfast spreading ridge: East Pacific Rise, 13-23°S. Earth and Planetary Science Letters 121, 173–193.

Mao, Z., Jacobsen, S.D., Jiang, F., Smyth, J.R. Holl, C.M., Duffy, T.S., 2008a. Elasticity of hydrous wadsleyite to 12 GPa: Implications for Earth's transition zone. Geophysical Research Letters 35, L21305, DOI: 10.1029/2008GL035618.

Mao, Z., Jacobsen, S.D., Jiang, F.M., Smyth, J.R., Holl, C.M., Frost, D.J., Duffy, T.S., 2008b. Single-crystal elasticity of wadsleyites, β-Mg2SiO4, containing 0.37–1.66 wt.% H_2O: Earth and Planetary Science Letters 268, 540–549.

Matzen, A.K., Baker, M.B., Beckeyy, J.R., Stolper, E.M., 2013. The temperature and pressure dependence of nickel partitioning between olivine and silicate melt. Journal of Petrology 54, 2521-2545.

Matzen, A.K., Wood, B.J., Stolper, E.M., 2017. The roles of pyroxenite and peridotite in mantle sources of oceanic basals. Nature Geoscience 10, 530-535.

McDonough, W.F., Sun, S.-s., 1995. The composition of the Earth. Chemical Geology 120, 223-253.

McKenzie, D., O'Nions, R.K., 1995. The source regions of oceanic island basalts. Journal of Petrology 36, 133-159.

McNamara, A.K., Zhong, S., 2004. Thermochemical structures within a spherical mantle: Superplumes or piles? Journal of Geophysical Research 109, doi:10.1029/2003JB002847.

McNutt, M.K., 1988. Superswells. Reviews of Geophysics 36, 211-244.

Mégnin, C., Romanowicz, B., 2000. The shear velocity structure of the mantle from the inversion of body, surface, and higher modes waveforms. Geophysical Journal International 143, 709–728.

Menzies, M., Murthy, V.R., 1980. Mantle metasomatism as a precursor to the genesis of alkaline magmas – isotopic evidence. American Journal of Science 280, 622-638.

Menzies, M.A., Hawkesworth, C.J. (eds), 1987, Mantle Metasomatism: Academic Press, London, Geology Series.

Montelli, R., Nolet, G., Dahlen, F. et al., 2004. Finite-frequency tomography reveals a variety of plumes in the mantle. Science 303, 338-343.

Morgan, W.J., 1971. Convection plumes in the lower mantle. Nature 230,42-43.

Morgan, W.J., 1972, Plate motions and deep mantle convection. Geological Society of America Memoir 132, 7-22.

Mulyukova, E.B. Steinberger, B., Dabrowski, M., Sobolev, S.V., 2015. Survival of LLSVPs for billions of years in a vigorously convecting mantle: Replenishment and destruction of chemical anomaly. Journal of Geophysical Research 120, 3824-3847.

Ni, S., Helmberger, D.V., 2003. Seismological constraints on the South African superplume; could be the oldest distinct structure on Earth. Earth and Planetary Science Letters 206, 119-131.

Ni, S., Tan, E., Gurnis, M., Helmberger, D.V., 2002. Sharp sides to the African superplume. Science 296, 1850-1852.

Niu, Y.L., 1997. Mantle melting and melt extraction processes beneath ocean ridges: Evidence from abyssal peridotites. Journal of Petrology 38, 1047-1074.

Niu, Y.L., 2004. Bulk rock major and trace element compositions of abyssal peridotites: Implications for mantle melting, melt extraction and post, melting processes beneath ocean ridges. Journal of Petrology 45, 2423-2458.

Niu, Y.L., 2005. Generation and evolution of basaltic magmas: Some basic concepts and a hypothesis for the origin of the Mesozoic, Cenozoic volcanism in eastern China. Geological Journal of China Universities 11, 9-46.

Niu, Y.L., 2005. On the great mantle plume debate. Chinese Science Bulletin 50, 1537-1540.

Niu, Y.L., 2008. The origin of alkaline lavas. Science 320, 883-884.

Niu, Y.L., 2009. Some basic concepts and problems on the petrogenesis of intra-plate ocean island basalts. Chinese Science Bulletin 54, 4148-4160.

Niu, Y.L., 2014. Geological understanding of plate tectonics: Basic concepts, illustrations, examples and new perspectives. Global Tectonics and Metallogeny 10, 23-46.

Niu, Y.L., Batiza, R., 1991a. In-situ densities of silicate melts and minerals as a function of temperature, pressure, and composition. Journal of Geology 99, 767-775.

Niu, Y.L., Batiza, R., 1991b. DENSCAL: A program for calculating the densities of sili-cate melts and minerals as a function of temperature, pressure, and composition in magma generation environment. Computers and Geosciences 17, 679-687.

Niu, Y.L., Batiza, R., 1997. Trace element evidence from seamounts for recycled oceanic crust in the eastern Pacific mantle. Earth and Planetary Science Letters 148, 471-483.

Niu, Y.L., Collerson, K.D., Batiza, R., Wendt, J.I., Regelous, M., 1999. The origin of E-Type MORB at ridges far from mantle plumes: The East Pacific Rise at 11°20'. Journal of Geophysical Research 104, 7067-7087.

Niu, Y.L., Green, D.H., 2014. The nature of the lithosphere-asthenosphere boundary (LAB) beneath ocean basins. Goldschmidt Conference Abstract, 1819.

Niu, Y.L., O'Hara, M.J., 2003. Origin of ocean island basalts: A new perspective from petrology, geochemistry and mineral physics considerations. Journal of Geophysical Research 108, 2209, doi:10.1029/2002JB002048.

Niu, Y.L., Regelous, M., Wendt, J. I., Batiza, R., O'Hara, M.J., 2002, Geochemistry of near-EPR seamounts: Importance of source vs. process and the origin of enriched mantle component. Earth and Planetary Science Letters 199, 327-345.

Niu, Y.L., Shi, X.F., Li, T.G., Wu, S.G., Sun, W.D., Zhu, R.X., 2017. Testing the mantle plume hypothesis: An IODP effort to drill into the Kamchatka-Okhotsk Sea basement. Science Bulletin 62, 1464-1472.

Niu, Y.L., Waggoner, G., Sinton, J.M., Mahoney, J.J., 1996. Mantle source heterogeneity and melting processes beneath seafloor spreading centers: The East Pacific Rise 18°-19°S. Journal of Geophysical Research 101, 27,711-27,733.

Niu, Y.L., Wilson, M., Humphreys, E.R., O'Hara, M.J., 2011. The origin of intra-plate ocean island basalts (OIB): The lid effect and its geodynamic implications. Journal of Petrology 52, 1443-1468.

Niu, Y.L., Wilson, M., Humphreys, E.R., O'Hara, M.J., 2012. A trace element perspective on the source of ocean island basalts (OIB) and fate of subducted ocean crust (SOC) and mantle lithosphere (SML). Episodes 35, 310-327.

Ohtani, E., Maeda, M., 2001. Density of basaltic melt at high pressure and stability of the melt at the base aof the lower mantle. Earth and Planetary Science Letters 193, 69-75.

Ono, S., Ito, E., and Katsura, T., 2001, Mineralogy of subducted basaltic crust (MORB) from 25 to 37 GPa, and chemical heterogeneity of the lower mantle. Earth and Planetary Science Letters 190, 57-63.

O'Reilly, Y.S., Griffin, W.L., 1988, Mantle metasomatism beneath western Victoria, Australia: I, Metasomatic processes in Cr-diopside lherzolites. Geochimica et Cosmochimica Acta 52, 433-447.

Palme, H., O'Neill, H. St. C., 2014. Cosmochemical estimates of mantle composition. Treatise on Geochemistry Chapter 3.1, 1-39.

Pilet, S., Baker, M.B., Stolper, E.M., 2008., Metasomatized lithosphere and the origin of alkaline lavas. Science-320, 916-919.

Pilet, S., Hernandez, J., Sylvester, P., Poujol, M., 2005. The metasomatic alternative for ocean island basalt chemical heterogeneity. Earth and Planetary Science Letters 236, 148-166.

Regelous, M., Niu, Y.L., Abouchami, W., 2009. Shallow origin for South Atlantic Dupal Anomaly from lower continental crust: Geochemical evidence from the Mid-Atlantic Ridge at 26°S. Lithos 112, 57-72.

Regelous, M., Niu, Y.L., Wendt, J.I. et al., 1999. An 800 Ka record of the geochemistry of magmatism on the East Pacific Rise at 10°30'N. Insights into magma chamber processes beneath a fast-spreading ocean ridge. Earth and Planetary Science Letters 168, 45-63.

Ritsema, H.J., van Heijst, J.H., Woodhouse, J.H., 1999. Complex shear velocity structure beneath Africa and Iceland. Science 286, 1925-1928.

Romanowicz, B., Gung, Y., 2002. Superplumes from the core-mantle boundary to the lithosphere: Implications for heat flux. Science 296, 513-516.

Romanowicz, B., The buoyancy of Earth's deep mantle. Nature 551, 308-309.

Salters, V. J. M., White, W. M., 1998. Hf isotope constraints on mantle evolution. Chemical Geology 145, 447– 460.

Schilling, J.-G., Zajac, M., Evans, R., Jonson, T., White, W., Devine, J.D., Kingsley, T., 1983. Petrological and geochemical variations along the mid-Atlantic Ridge from 29°N to 73°N. American Journal of Science 283, 510–586.

Sobolev A.V., Hofmann, A.W., Nikogosian, I.K., 2000. Recycled oceanic crust observed in "ghost plagioclase" within the source of Mauna Loa lavas. Nature 404, 986-990.

Sobolev, A.V., 19 others, 2007, The amount of recycled crust in sources of mantle-derived melts. Science 316, 412-417.

Sobolev, A.V., Hofmann, A.W., Sobolev, S.V., Nikogosian, I.K., 2005. A olivine-free mantle source of Hawaii shield basalts. Nature 434, 590-597.

Steinberger, B., Torsvik, T.H., 2017. A geodynamic model of plumes from the margins of Large Low Shear Velocity Provinces. Geochemistry, Geophysics, Geosystems 13, Q01W09, doi:10.1029/2011GC003808.

Stern, R.J., 2007. When and how did plate tectonics begin? Theoretical and empirical considerations. Chinese Science Bulletin 52, 578-591.

Stracke, A., Zindler, A., Salters, V.J.M. et al. 2003. Theistareykir revisited. Geochemistry, Geophysics, Geosystem 4, doi: 10.1029/2001GC000201

Su, W., Dziewonski, A.M., 1997. Simultaneous inversion for 3-D variations in shear and bulk velocity in the mantle. Physics of the Earth and Planetary Interiors 100, 135-156.

Sun, S.-s, Tatsumoto, M., Schilling, J.-G., 1975. Mantle plume mixing along the Reykjanes ridge: Lead isotope evidence. Science 190, 143-147.

Sun, S.-s., Hanson, G.N., 1975. Origin of Ross Island basanitoids and limitations upon the heterogeneity of mantle sources for alkali basalts and nephelinites: Contributions to Mineralogy and Petrology 52, 77-106.

Sun, S.-s., McDonough, W.F., 1989. Chemical and isotopic systematics in ocean basalt: Implication for mantle composition and processes. Geological Society Special Publication 42, 313-345.

Takazawa, E., Frey, F.A., Shimizu, N., Obata, M., 2000. Whole rock compositional variations in an upper mantle peridotite (Horoman, Hokkaido, Japan). Are they consistent with a partial melting process? Geochimica et Cosmochimica Acta 64, 695-716.

Thorne, M., Garnero, E.J., Grand, S., 2004, Geographic correlation between hot spots and deep mantle lateral shear-wave velocity gradients. Physics of Earth and Planetary Interiors 146, 47-63.

Toh, A., Romanowicz, B., Capdeville, Y., Takeuchi, N., 2005. 3D effects of sharp boundaries at the borders of the African and Pacific Superplumes: observation and modeling. Earth and Planetary Science Letters 233, 137-153.

Torsvik, T.H., Burke, K., Steinberger, B., Webb, S.J., Ashwal, L.D., 2010. Diamonds sampled by plumes from the core-mantle boundary. Nature 466, 352–355

Torsvik, T.H., Domeier, M., 2017. Correspondence: Numerical modelling of the PERM anomaly and the Emeishan large igneous province. Nature Communications 8, 821, doi: 10.1038/s41467-017-00125-2.

Torsvik, T.H., van der Voo, R., Doubrovine, P.V., Burke, K., Steinberger, B., Ashwald, L.D., Trønnes, R.G., Webb, S.J., Bulla, A.L., 2014. Deep mantle structure as a reference frame for movements in and on the Earth. PNAS 111, 8735-8740.

van der Hilst, R.D., Widiyantoro, S., Engdahl, E.R., 1997. Evidence for deep mantle circulation from global tomography. Nature 386, 578–584.

Wang, Y., Wen, L., 2004. Mapping the geometry and geographic distribution of a very-low velocity province at the base of the Earth's mantle. Journal of Geophysical Research 109, B10305, doi:10.1029/2003JB002674.

Wendt, J.I., Regelous, M., Niu, Y.L., et al., 1999. Geochemistry of lavas from the Garrett Transform Fault: insights into mantle heterogeneity beneath the eastern Pacific. Earth and Planetary Science Letters 173, 271—284.

White, W.M., 1985. Sources of oceanic basalts: Radiogenic isotopic evidence. Geology 13, 115- 118, 1985

White, W.M., 2010. Ocean island basalts and mantle plumes - the geochemical perspective. Annual Review of Earth and Planetary Sciences 38, 133-160.

White, W.M., 2015. Isotopes, LLSVPs, and Adekantavada. Chemnical Geology 419, 10-28.

White, W.M., 2016. Probing the Earth's deep interior through geochemistry. Geochemical perspectives 4, 96-218.

Willbold, M., Stracke, A., 2006. Trace element composition of mantle endmembers: Implications for recycling of oceanic and upper and lower continental crust. Geochemistry, Geophysics, Geosystems 7, doi:10.1029/2005GC001005.

Williams, Q., Revenaugh, J., Garnero, E., 1998. A correlation between ultra-low basal velocities in the mantle and hot spots. Science 281, 546-549.

Wilson, J.T., 1963. A possible origin of the Hawaiian Islands. Canadian Journal of Physsics 41,863-870.

Xiao, Y.Y., Lavis, S., Niu, Y.L., Pearce, J.A., Li, H.K., Wang, H.C., Davidson, J., 2012. Trace element transport during subduction-zone ultrahigh pressure metamorphism: Evidence from Western Tianshan, China. Geological Society of America Bulletin 124, 1113-1129.

Xiao, Y.Y., Niu, Y.L., Song, S.G., Davidson, J., Liu, X.M., 2013. Elemental responses to subduction-zone metamorphism: Constraints from the North Qilian Mountain, NW China. Lithos 160/161, 55-67.

Ye, Y., Schwering, R.A, Smyth, J.R., 2009. Effects of hydration on thermal expansion of forsterite, wadsleyite, and ringwoodite at ambient pressure. American Mineralogist 94, 899-904.

Zhao, D., 2004. Global tomographic images of mantle plumes and subducting slabs: Insights into deep Earth dynamics. Phys Earth Planet Interiors 146, 3-34.

Zindler, A., Hart, S.R., 1986. Chemical geodynamics. Annual Review of Earth and Planetary Sciences 14, 493-571.

Zindler, A., Staudigel, H., Batiza, R., 1984. Isotope and trace element geochemistry of young Pacific seamounts: Implications for the scale of upper mantle heterogeneity. Earth and Planetary Science Letters 70, 175-195.

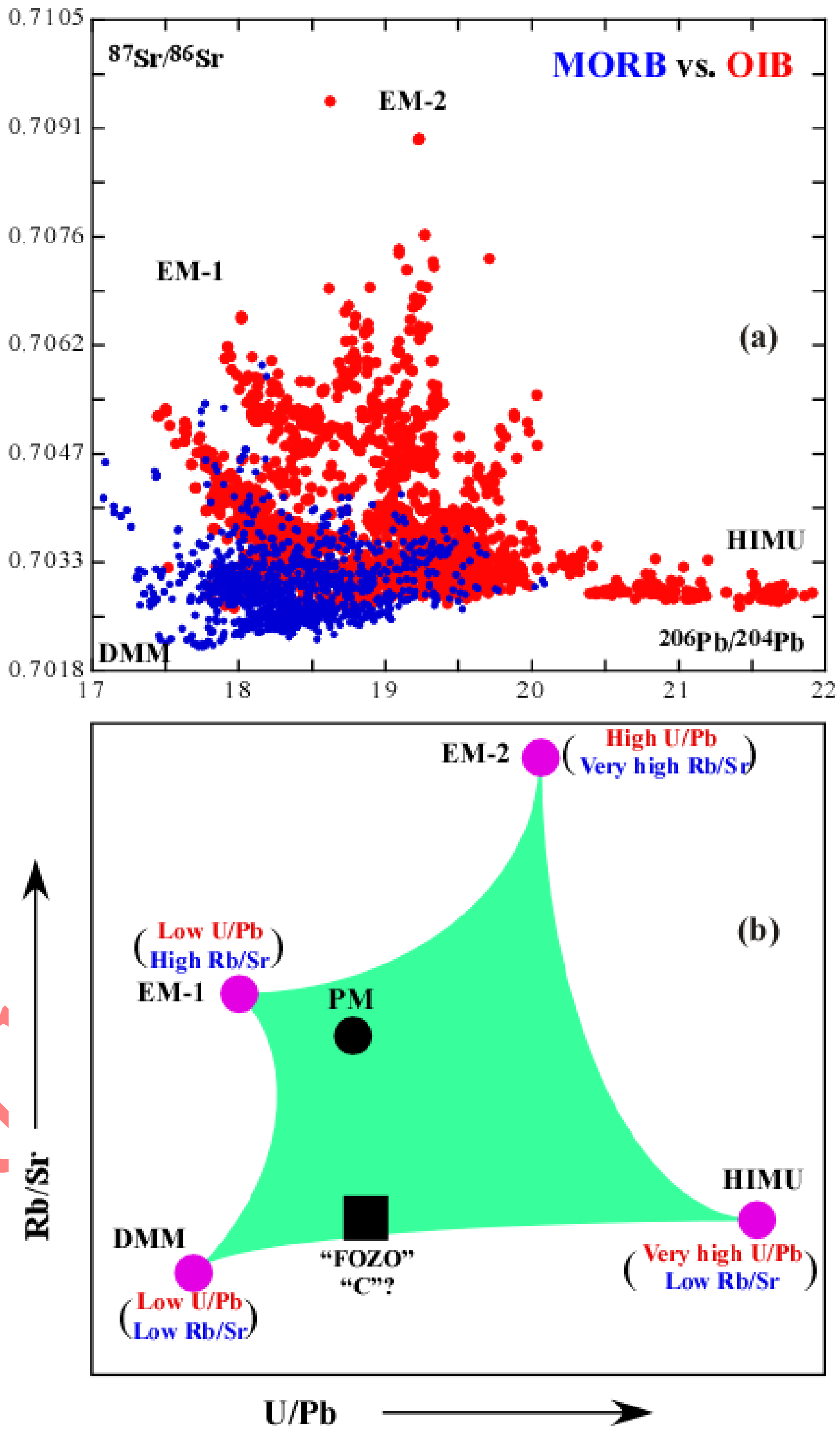


Figure 1. (**a**) Mantle isotopic heterogeneity shown as an example in $^{87}Sr/^{86}Sr$ vs. $^{206}Pb/^{204}Pb$ space using global OIB and MORB data. The data are from the compilation of Stracke et al. (2003) plus those reported by Regelous et al. (1999, 2009), Wendt et al. (1999) and Niu et al. (1999, 2002). The mantle isotopic endmembers DMM, EM-1, EM-2 and HIMU are from Zindler and Hart (1986). Mixing of these mantle isotopic endmembers in varying proportions is thought to cause the large mantle isotopic variability. Because the isotopic ratio (e.g., $^{87}Sr/^{86}Sr$, $^{206}Pb/^{204}Pb$) of a sample depends on the ratio of radioactive parent (*P*, e.g., ^{87}Rb, ^{238}U) to radiogenic daughter (*D*, e.g., ^{87}Sr, ^{206}Pb) of the source and the time elapsed since the isolation of the source and because most commonly used radioactive parents have a long half-life, the isotopic ratio variation in the source is largely controlled by the P/D ratio variation in the source (e.g., Rb/Sr, U/Pb). Hence, (**b**) illustrates this concept and explains that the process of understanding the origin of mantle isotopic heterogeneity and endmembers is equivalent to understanding the origin of P/D ratio variation in the mantle sources. PM is the primordial mantle, "FOZO" (Hart et al., 2992) and "C" (Hanan and Graham, 1996) appear to be the common "component" of all the isotopic endmember arrays in multiple isotope spaces (Castillo, 2015, 2016, 2017; Jackson et al., 2018). Note that more than 85% of studied OIB suites/samples are in the compositional regions around the "FOZO", and thus represent the first-order OIB characteristics for conceptual understanding OIB, which is the focus of this study.

Figure 2. (**a**) Model ocean crust (OC, Niu and O'Hara, 2003) normalized multi-element diagram showing two model estimates of average composition of the global OIB (Sun and McDonough, 1989; Wilbold and Stracke, 2006) and average composition of IAB (island arc basalts by Elliott, 2003). The values in the tables are primitive-mantle (Sun and McDonough, 1989) normalized OIB (WS06 used) and model OC to show that (1) OC is too depleted to be the possible source material for highly enriched OIB in terms of incompatible element abundances and ratios; (2) Melting of such OC (if subducted as SOC) will be isotopically too depleted to resemble OIB. (**b**) Qualitative illustration of trace element characteristics, complementary to IAB in (**a**), that SOC would have after passing through subduction-zone dehydration if we accept the slab-dehydration induced mantle wedge melting model for IAB; melting of such SOC will not produce melts in any way relevant to OIB in (**a**) (see Niu and O'Hara, 2003).

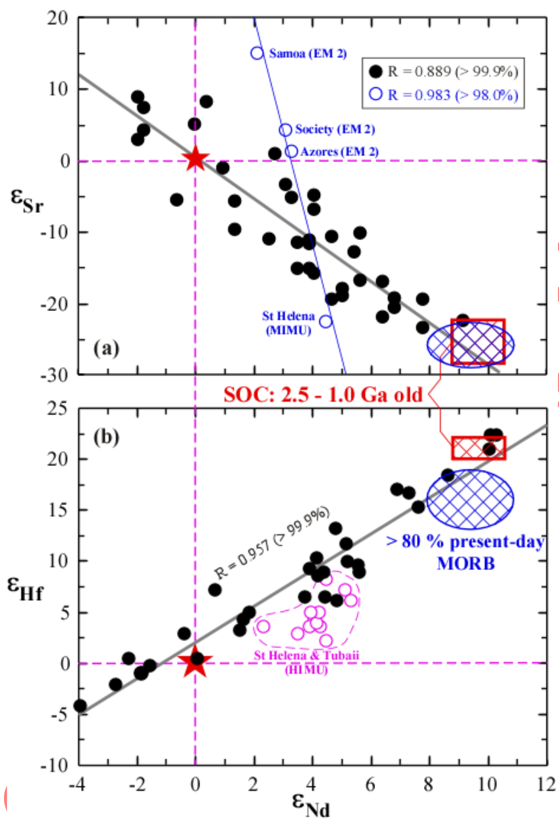
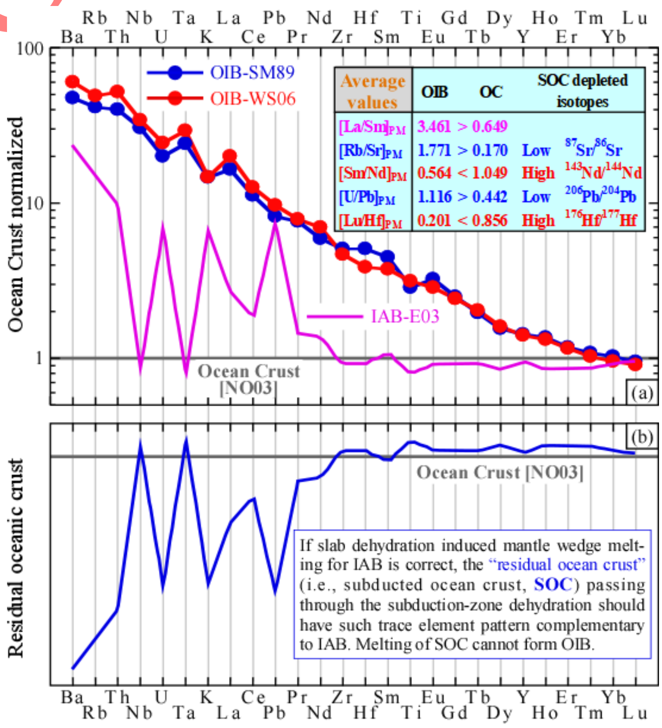


Figure 3. Plots of OIB in (**a**) εSr-εNd (data from Albarede, 1995) and (**b**) εHf-εNd (data from Salters and White, 1998) spaces modified after Niu and O'Hara (2003). In (**a**) each data point is average representing an ocean island suite, and in (**b**), the data are OIB suites from a number of representative ocean islands with global coverage. Note that except for three EM2 OIB suites and one HIMU suite in (**a**) and HIMU suites in (**b**) all the rest of the OIB data define significant linear trends coincident with the mantle arrays in these two spaces. The red stars are CHUR values. The linear trends suggest that the process or processes that have led to the correlated trends may be simple. The EM2 and HIMU OIB suites may be genetically different. Note also that the 3 EM2 and 1 HIMU OIB suites define a linear trend at a high angle to the mantle array in (**a**). The meshed red rectangles indicate the present-day isotope compositions of ancient (2.5 – 1.0 Ga) SOC stored in the mantle calculated in Niu and O'Hara (2003). They are too depleted to be possible sources of any OIB suite. The meshed ovals indicate the isotope compositions of > 80% of the present-day MORB (data from Stracke et al., 2003). Note, large global OIB isotope data are available (e.g., Stracke et al., 2003), but the global OIB averages used here capture the intrinsic properties of the global OIB data in simple clarity.



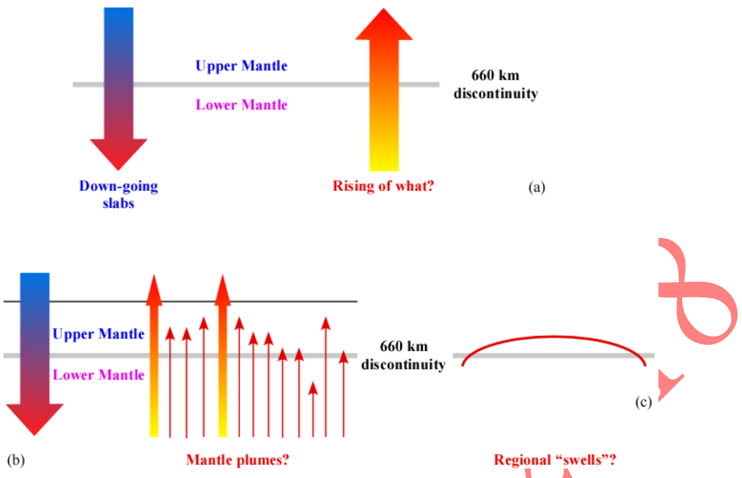


Figure 4. Cartoons illustrating the possible mass exchanges between the upper and lower mantle. (**a**) Despite the potential difficulty of subducting slab penetration into the lower mantle because of the negative Clapeyron slope at the 660-D (600 km seismic discontinuity) (see Niu, 2014 for a review), subducting slabs do penetrate the 660-D and enter the lower mantle as shown by many seismic tomography studies (e.g., van der Hilst et al., 1997; Kellogg et al., 1999; Karason and van der Hilst, 2000; Li et al., 2008). This requires same amount lower-to-upper mantle mass transfer to keep mass balanced, but the material that may rise from the lower mantle to upper mantle may be density filtered? (**b**) lower-to-upper mantle mass transfer is generally thought to take the form of mantle plumes, whose narrow diameters and small temperature contrast with the ambient mantle may make them seismically difficult to detect (see Niu et al., 2017 for a review; Foulger, 2010; Julian, 2005) despite great efforts (e.g., Montelli et al., 2004; Zhao, 2004). (**c**) It is possible that the upward mass transfer takes the form of regional "swells" across the 660-D without the need of mantle plumes as in (**b**).

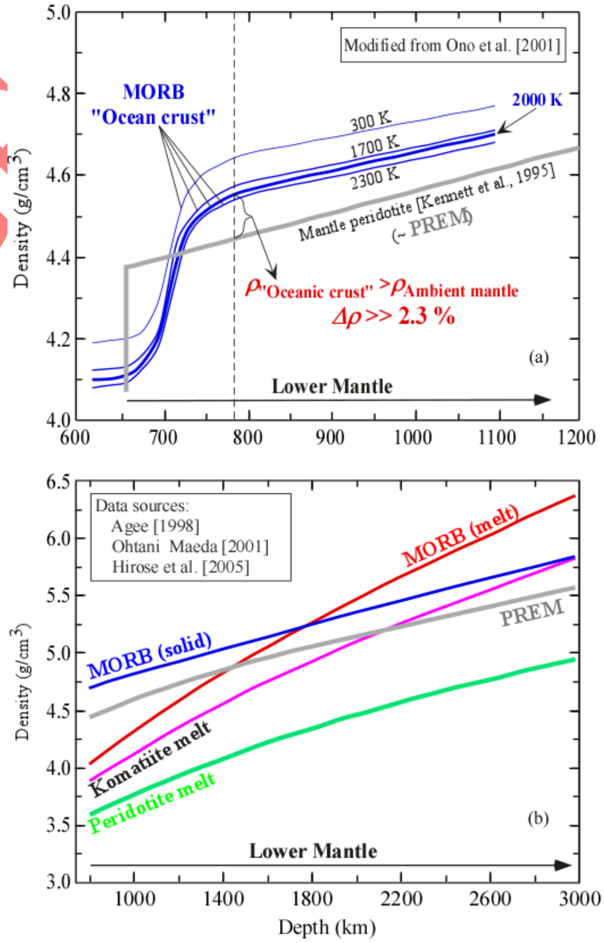


Figure 5. (a) Experimental data (Ono et al., 2001) showing that the ocean crust subducted into the upper portion of the lower mantle will be transformed to a high-pressure mineral assemblage whose bulk density is significantly greater than the ambient peridotite mantle (the PREM model of Dziewonski and Anderson, 1981; Kennett et al., 1995). Consider a whole mantle convection scenario, the mantle temperature at ~ 780 km depth would be ~ 2000 K. In this case, the oceanic crust, if subducted into the lower mantle, would be > 2.3% denser than the ambient mantle. Such huge negative buoyancy impedes the rise of subducted ocean crust into the upper mantle source regions of oceanic basalts (Niu and O'Hara, 2003; Niu et al., 2012). (**b**) Densities of various Earth materials under deeper lower mantle conditions (Niu and O'Hara, 2003; Niu et al., 2012) based on the experimental data of Agee (1998), Ohtani and Maeda (2001) and Hirose et al. (2005) to show that under lower mantle conditions, SOC of MORB composition [MORB (solid)] is significantly denser (~ 3.6%) than the ambient mantle represented by PREM (Dziewonski and Anderson, 1981). In the molten state, especially at deep lower mantle conditions, molten ocean crust [MORB (melt)] is denser than komatiite melt (by ~ 8.6%), and even denser than the ambient mantle (by ~ 13%). Importantly, peridotite melt under lower mantle conditions has the lowest density; it is ~ 12% less dense than the ambient mantle and is ~ 17% less dense than the solid ocean crust [MORB (solid)].

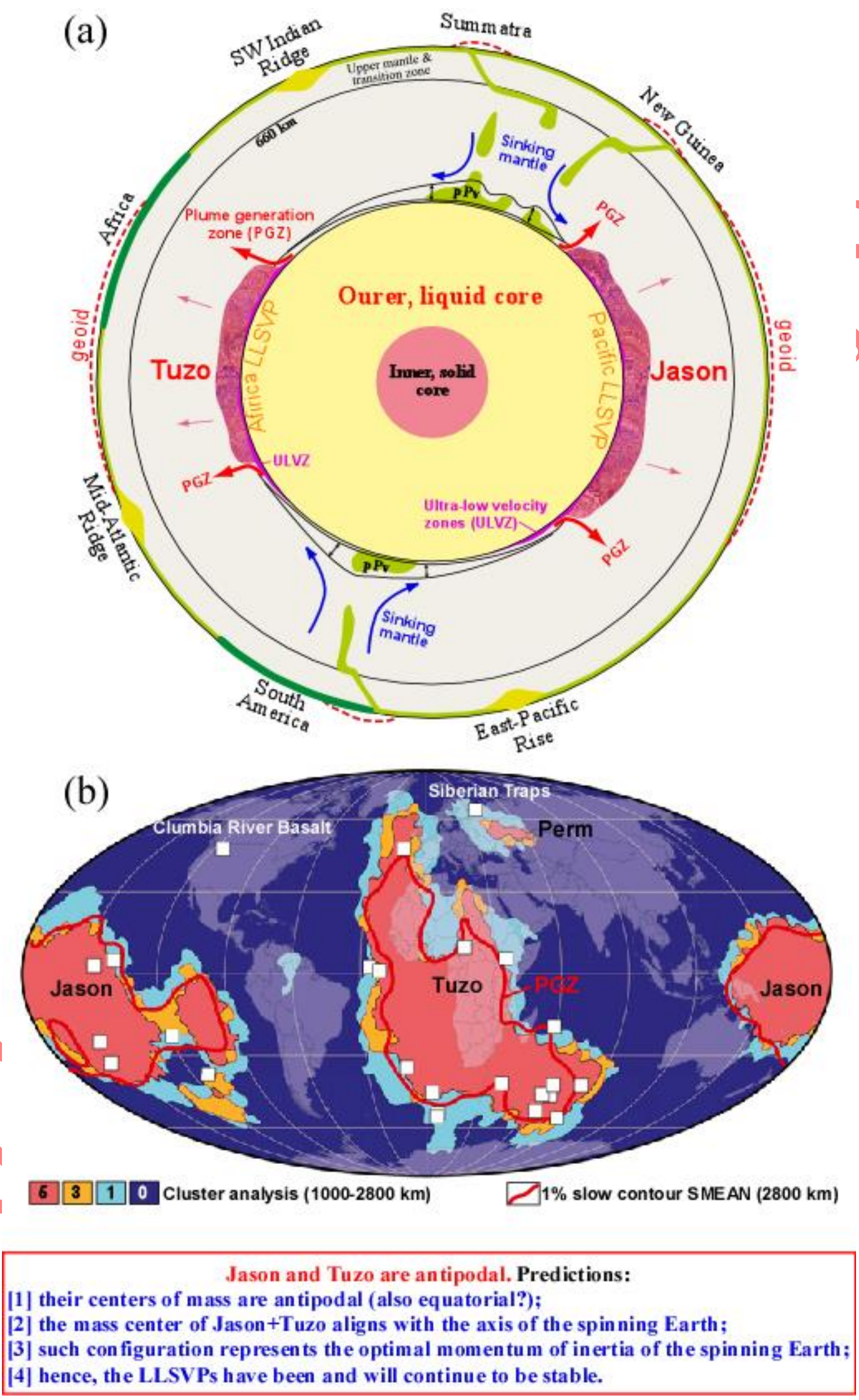


Figure 6. Adapted and slightly modified from Torsvik et al. (2014). (**a**) A cartoonish cross-section of the Earth viewing from the South Pole. The Earth's lower deep mantle has two antipodal large low shear-wave velocity provinces (LLSVPs) beneath Africa (Tuzo) and the Pacific (Jason), named by Burke (2011) in honour of Tuzo Wilson and Jason Morgan. These correspond to elevated regions of the residual geoid highs (dashed red lines) or buoyant "super swells" (McNutt, 1988) and were interpreted to represent super plumes (Romanowicz and Gung, 2002; Thorne et al., 2004; Torsvik et al., 2014). Their margins are called plume generation zones (PGZs) because they correspond to the large igneous provinces (LIPs) over the last ~ 300 Myrs (Burke and Torsvik, 2004; Burke et al., 2008). The green "pPv" patches are interpreted as postperovskite (Torsvik et al., 2014). (**b**) Shows large igneous provinces (LIPs) for the past 300 Myrs (Torsvik et al., 2010) are projected at margins of the LLSVPs, leading to the proposal that the LLSVPs have been "stable" or "unmoved" in the geological past, at least in the past ~ 300 Myrs. The 1% slow SMEAN (Becker and Boshi, 2002) contour (2,800 km depth) is used as a proxy for the plume generation zones (Torsvik et al., 2014). Contours 5, 3, and 1 define the Tuzo and Jason seismically slow regions. The Perm anomaly may be connected with Tuzo? The Columbia River Basalt (17 My) is the only anomalous large igneous province away from the LLSVPs. Jason and Tuzo are antipodal, and if their respective centres of mass are also antipodal (also equatorial?) with respect to the spinning Earth, this would explain why they are stable/unmoved because the mass centre of Jason and Tuzo combined ("Jason+Tuzo") would align with the axis of the spinning Earth.

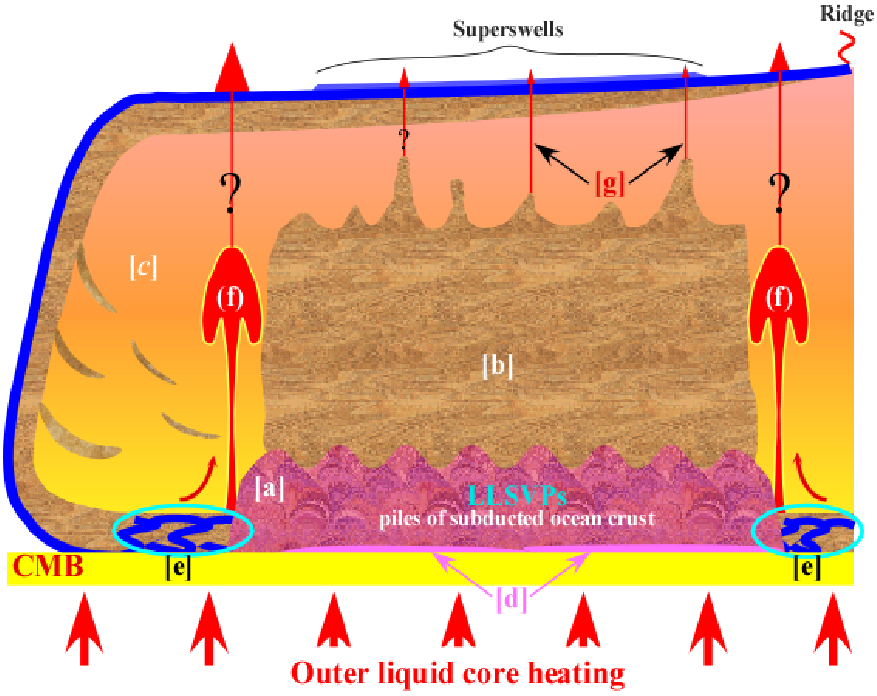


Figure 7. Cartoon (modified from Niu et al., 2012; not to scale, excluding complications due to phase changes etc.) illustrating my interpretations on the origin of the LLSVPs using the petrological and geochemical arguments given in the text and the seismic (in the deep mantle) and geological (at the surface) observations. [a], the LLSVPs, which are made up of piles of SOC accumulated over Earth's history, i.e., the LLSVPs are permanent 'graveyard' of the SOC. They also act as a thermal "insulator", preventing effective core-to-mantle heat transfer. [b], accumulated piles of SML that is less dense (see Fig. 5) than and separated from the dense [a]. [c], "normal mantle" like "PREM". [d], ultra-low velocity zone (ULVZ) represented by highly localized melt layers/pockets of highest density (Fig. 5,6). [e], edge regions of the LLSVPs at the CMB, where hot diapirs [f] of SML develop and rise. The hot diapirs may contain a peridotitic melt phase, and their ascent is driven by compositional (vs. thermal) buoyancy. They may feed volumetrically significant surface volcanism. [g], possible compositional diapirs originating from SML at shallow depths that may also feed some surface volcanism. Note the mean density relationships among these different constituents at a comparable depth: [d] > [a] > [c] > [b] > [g] ≈ [f] (see Fig. 5). Note that the lower density of [b] than [c] is largely due to compositional buoyancy contrast. The SML [b] contains (i) MORB melting residue, which is less dense, (ii) water probably hosted in DHMS phases (dense hydrous magnesium silicates) derived from incompletely subduction-dehydrated serpentine beneath the crust developed near ocean ridges, (iii) serpentine developed in the deep lithosphere at the trench-outer-rise, and (iv) re-fertilized volatile rich deep portions of the lithosphere (see Niu et al., 2011, 2012). Water can reduce both density and elastic moduli. Consequently, the surface "superswells" may be caused by the "density deficiency" of [b], which is not in contradiction to the dense [a]. The thermal insulation of [a] makes [b] not thermally buoyant, but at the edges of [a] the materials of the SML can be heated up and melt. The compositional buoyancy of the SML and its melt facilitates the development of diapirs and their ascent as "plumes". This explains why sites of major volcanoes and LIPs are associated with the edges, not the interiors of, the LLSVPs. Note also that the OIB source materials are NOT [a] (or SOC) that is permanently left at the base of the mantle, but are SML of peridotite compositions for logical and petrological reasons discussed in the text (also see Niu and O'Hara, 2003; Niu et al., 2012). [e], hot thermal boundary layer ≈ core-mantle boundary (CMB). Note that disintegration (crust-mantle lithosphere separation) of the subducted slabs may take place throughout the mantle during descent with the dense SOC sinking, adding to the LLSVPs at the base of the mantle and the compositionally buoyant SML in the mid mantle, which can also develop into "chemical plumes" (e.g., [g]). Relative to the dense SOC, the less dense SML is volumetrically 15 times of the SOC and may explain the surface super swells.