

The fate of subducted oceanic crust: a mineral segregation model

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Plate subduction and mantle plumes are two of the most important material transport processes of the silicate Earth. Currently, a debate exists over whether the subducted oceanic crust is recycled back to the Earth's surface through mantle plumes, and can explain their derivation and major characteristics. It is also puzzling as to why plume heads have huge melting capacities and differ dramatically from plume tails both in size and chemical composition. We present data showing that both ocean island basalt and mid-ocean ridge basalt have identical supra-primitive mantle mean Nb/U values of ~46.7, significantly larger than that of the primitive mantle value. From a mass balance calculation based on Nb/U we have determined that nearly the whole mantle has evolved by plate subduction-induced crustal recycling during formation of the continental crust. This mixing back of subducted oceanic crust, however, is not straightforward, because it generally would be denser than the surrounding mantle, both in solid and liquid states. A mineral segregation model is proposed here to reconcile different lines of observation. First of all, subducted oceanic crustal sections are denser than the surrounding mantle, such that they can stay in the lower mantle, for billions of years as implied by isotopic data. Parts of subducted oceanic crust may eventually lose a large proportion of their heavy minerals, magnesian-silicate-perovskite and calcium-silicate-perovskite, through density segregation in ultra-low-velocity zones as well as in very-low-velocity provinces at the core-mantle boundary due to low viscosity. The remaining minerals would thus become lighter than the surrounding mantle, and could rise, trapping mantle materials, and forming mantle plumes. Mineral segregation progressively increases the SiO₂ content of the ascending oceanic crust, which enhances flux melting, and results in giant Si-enriched plume heads followed by dramatically abridged plume tails. Therefore, ancient mineral-segregated subducted oceanic crust is likely to be a major trigger and driving force for the formation of mantle plumes.

Keywords: Nb/U; subducted oceanic crust; mantle plume; mineral segregation; flux melting; plume head; ultra-low-velocity zone; core-mantle boundary

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Introduction

There are two types of major mantle upwelling in the Earth, which form isotopically and chemically distinct basalts – namely ocean island basalt (OIB) and mid-ocean ridge basalt (MORB) (Jacobsen and Wasserburg 1979; Sun *et al.* 1979; Sun 1980; Zindler and Hart 1986; Sun and McDonough 1989; Hofmann 1997; Allegre 2002). OIB gives rise to one of the clearest surface expressions of mantle plumes originating from the lower mantle, whereas MORB is an important component of oceanic plates, which only samples the depleted upper mantle. Compared to MORB, OIB is usually more enriched in incompatible elements, such as U, Nb, Ta, Th, and K. Its isotopic composition is typically more radiogenic (with the exception Nd) and far more variable than MORB (e.g. Jacobsen and Wasserburg 1979; Sun 1980; Zindler and Hart 1986; Hofmann 1997; Allegre 2002; Frey *et al.* 2002; Abouchami *et al.* 2005; Frey *et al.* 2005; Huang *et al.* 2005; Sobolev *et al.* 2007). The distinctively different geochemical characteristics of MORB and OIB led to the development of layered-mantle-convection models, which were once widely accepted by geochemists. In these models, MORB came from the well-stirred, depleted upper layer above the 660-km discontinuity of the Earth's mantle, whereas OIB came from a much more heterogeneous and enriched lower layer (Jacobsen and Wasserburg 1979; Zindler and Hart 1986; Hofmann 1997; Allegre 2002) with major contributions from recycled oceanic crust (Hofmann and White 1980; Chase 1981; Hofmann and White 1982; Ren *et al.* 2005; Sobolev *et al.* 2005, 2007; Sun *et al.* 2008).

Interestingly, OIB is not derived from a primitive mantle, but rather from a source that has close connections to the MORB mantle as shown by Nb/U (Hofmann 1997; Sun *et al.* 2008). It has been proposed that the recycling of subducted oceanic crust, after long term residence (~1–2 Ga) in the lower mantle, is responsible for the geochemical characteristics of the OIB (Hofmann and White 1980, 1982; Chase 1981; Christensen and Hofmann 1994; Hofmann 2003). This hypothesis is seemingly supported by seismic tomographic images, which suggest that cold subducting slabs apparently can descend across the 660-km discontinuity and reach the core–mantle boundary (CMB) (van der Hilst *et al.* 1997). The descent of subducted slabs across the 660-km discontinuity would require a counterbalancing upward flow of lower mantle materials to the upper mantle.

Nonetheless, a controversy arises over whether the subducted slabs have been mixed back into the mantle sources of OIB (Hofmann and White 1982; Hofmann 1997) and MORB (Hofmann 1997), or have been totally retained in the lower mantle (Niu and O'Hara 2003), that is, 'lower mantle density trap', because of the negative buoyancy of the bulk oceanic crusts in both solid (Kesson *et al.* 1998; Ono *et al.* 2001) and liquid states (Agee 1998; Ohtani and Maeda 2001) in the lower mantle.

Different models have been proposed, including moderate- to small-scale chemical heterogeneities (Helffrich and Wood 2001), poor mixing (van KeKen *et al.* 2002), deeper layering (Kellogg *et al.* 2002), a post-Archaeon global 'transition-zone water filter' (Bercovici and Karato 2003), subducted early crust 'isolation' at the CMB (Tolstikhin and Hofmann 2005), and serpentinite segregation (Lee and Chen 2007), to reconcile with geochemical and geophysical observations. Although these models managed to explain the chemical difference between the MORB and OIB sources, none of them could plausibly account for the 'lower mantle density trap'. Other models argue that the subduction of oceanic slabs probably existed ever since the Early Archaeon, while tomographic studies only show the current situation. If the descent of oceanic slabs penetrating into the lower mantle is only a relatively recent phenomenon, then it does not disprove the layered mantle convection model (Allegre 2002). These arguments raise questions about the physical and chemical

evolution of the Earth's mantle and the fate of the subducted oceanic crusts, as well as potential linkage between subducted oceanic crusts and mantle plumes.

Linkage between subducted oceanic crust and mantle plumes

The linkage of subducted oceanic crusts to mantle plumes is investigated here from the prospective of Nb/U values, one of the most useful geochemical tracers for subduction induced recycling (Hofmann *et al.* 1986; Hofmann 1997). Nb and U are similarly incompatible during MORB and OIB petrogeneses, such that they do not fractionate from each other and thus Nb/U values of basalts are identical to their corresponding mantle sources (Hofmann *et al.* 1986; Sun and McDonough 1989; Hofmann 1997; Sun *et al.* 2008). Uranium, however, is much more mobile than Nb during plate subduction (McCulloch and Gamble 1991; Pearce and Peate 1995), resulting in considerably lower Nb/U values in arc magmas and consequently in the continental crust (~6) (Pearce and Peate 1995; Rudnick and Fountain 1995; Barth *et al.* 2000), and correspondingly higher Nb/U values in the subducted residual crust than for the primitive mantle.

Available data show that the Nb/U values for OIB and MORB both have close to Gaussian distributions with small skewnesses of 1.9 and 1.4 for MORB and OIB, respectively, indicating a high degree of symmetry and implying that these mean Nb/U values are statistically valid (Figure 1). Remarkably, the mean Nb/U values for MORB (46.7 ± 0.3) and OIB (46.7 ± 0.5) are identical to each other, and are both considerably higher than the Nb/U value of ~30 for the primitive mantle (Hofmann *et al.* 1986; McDonough and Sun 1995; Hofmann 1997) and ~6 for the continental crust (Rudnick and Fountain 1995; Barth *et al.* 2000), confirming previous observations based on a smaller data set (Hofmann *et al.* 1986). This agreement in Nb/U strongly suggests that the sources of MORB and OIB are linked to each other (Hofmann 1997). Given that the formation of the continental crust is the only major process so far recognized in the silicate Earth that dramatically fractionates Nb from U, it indicates that both MORB and OIB sources were responsible for and/or involved in the formation of the continental crust (Hofmann 1997).

The amount of mantle source involved in the formation of the continental crust can be estimated by using a mass balance equation and the Nb, U abundances of the different reservoirs. Assuming the average Nb/U value for mantle sources that were involved in the formation of continental crust is 46.7, then nearly all the Earth's mantle needs to be depleted to produce the continental crust (Figure 2). Even if a considerable amount of Nb (e.g. ~10%) has been stored in the core as has been speculated (Wade and Wood 2001; Kamber *et al.* 2003), >80% of the mantle still needs to be depleted (Figure 1), a volume significantly larger than the upper mantle (above the 660 km discontinuity) and the depleted MORB mantle source estimated using isotope (Jacobsen and Wasserburg 1979; O'Nions *et al.* 1979; DePaolo 1980) or trace element data (e.g. McDonough 1991; Workman and Hart 2005).

These results indicate that a larger portion of the subducted slabs has been well mixed back into both the upper and lower mantle, and that both MORB and OIB have components from the subducted oceanic crust.

Mineral segregation model

General consensus is that, during the subduction of oceanic crust, the descending basalt transforms first to blueschist, then to eclogite in the upper mantle, and finally to assemblages of high pressure minerals in the lower mantle. Subducted oceanic crust (in eclogite phases) is about 2–4% denser than peridotite in the upper mantle, which provides a major

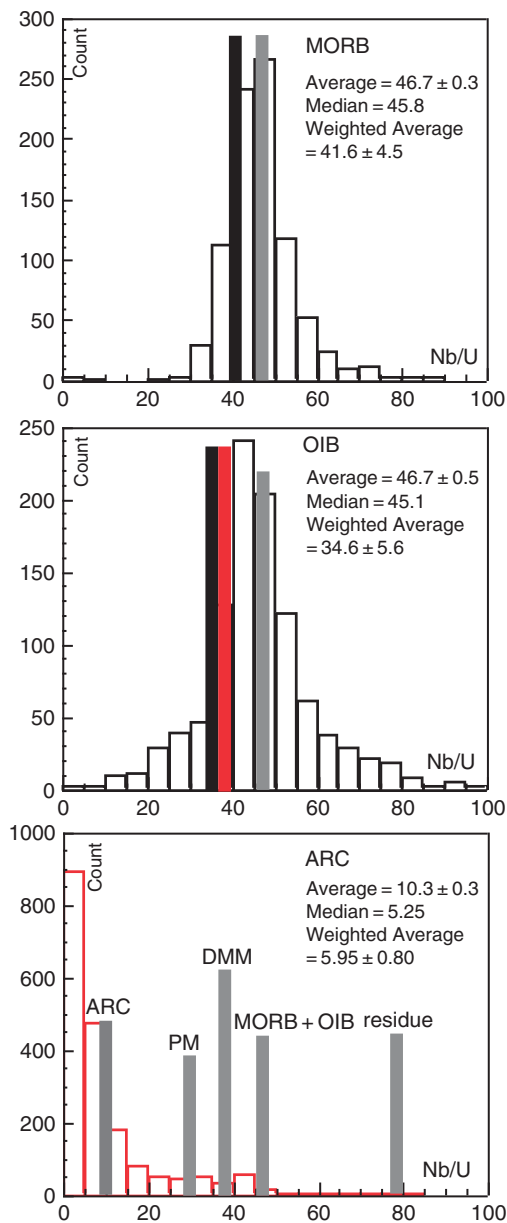


Figure 1. Nb/U distributions for (a) OIB and (b) MORB. Data and references can be found in the databases GEOROC (<http://georoc.mpch-mainz.mpg.de>) and PETDB (<http://petdb.ldeo.columbia.edu/petdb>) and (Sun *et al.* 2003). For MORB samples, only glasses were plotted concerning alteration effects on the whole rocks, otherwise the data were not filtered. OIB appears to be more diverse than MORB. This may be partly because MORB samples were restricted to glasses, which presumably means better data quality. It, however, may also reflect more diversity in the OIB sources (Zindler and Hart 1986; Hofmann 1997; Allegre 2002). The skewnesses are 1.9 and 1.4 for MORB and OIB, respectively, indicating a high degree of symmetry and statistically good mean values. The errors shown are 1σ standard errors.

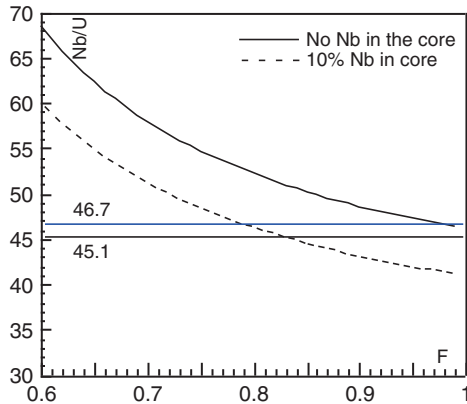


Figure 2. Mass balance calculation showing the proportion of the depleted mantle with Nb/U value of 46.7 required to account for the continental crust. (a) When there is no Nb in the core, ~98% of the mantle need to be depleted. (b) When 10% of the total Nb is stored in the core, then ~80% of the mantle may have been kept primitive and isolated from the MORB and OIB sources (have not been sampled). Data are: continental crust, U, 1.4 ppm (Rudnick and Fountain 1995), Nb, 8 ppm (Barth *et al.* 2000), total mass 2.09×10^{25} g (Kamber *et al.* 2003); primitive mantle, U, 0.0203 ppm, Nb, 0.658 ppm (McDonough and Sun 1995), total mass 4.019×10^{27} g (Kamber *et al.* 2003).

driving force for the subduction; it then becomes less dense than the surrounding mantle at depths of 650–750 km (Ringwood 1991; Ono *et al.* 2001). Seismic tomographic images, however, indicate that cold subducting slabs can descend across the 660-km discontinuity and reach the CMB (van der Hilst *et al.* 1997). Once subducted slabs do penetrate the 660-km discontinuity and descend into the lower mantle, the bulk density of the basaltic oceanic crusts is again higher than the surrounding mantle (Kesson *et al.* 1998; Ono *et al.* 2001). Importantly, both basalt and komatiite melts are also denser than the surrounding mantle at lower mantle depths (Agee 1998; Ohtani and Maeda 2001). Such a ‘lower-mantle-density-trap’ was used to argue that oceanic crust subducted to the lower mantle cannot come back (Niu and O’Hara 2003), and thus the lower mantle becomes a graveyard of subducted oceanic crusts.

We propose that subducted oceanic crust has been recycled back towards the surface through mantle plumes after mineral segregation at the CMB (Figure 3). According to the mineral segregation model, subducted oceanic crust is denser than the surrounding mantle, and it stays in the lower mantle for billions of years, as implied by isotopic data (Zindler and Hart 1986; Hofmann 1997; Hofmann 2003). In the lower mantle, the subducted oceanic crust transforms into an assemblage of magnesian-silicate-perovskite (MgPv), calcium-silicate-perovskite (CaPv), stishovite and Ca-ferrite (Table 1) (Kesson *et al.* 1998; Ono *et al.* 2001). The first two minerals are considerably denser than the others, and are the reason for the ‘lower mantle density trap’ (Table 1). Therefore, the most straightforward and practical way to surmount the ‘lower mantle density trap’ would be that the subducted oceanic crust eventually loses some of its heavy minerals, MgPv and CaPv, through mineral segregation at the CMB, thus becoming lighter than the bulk mantle and capable of rising up to the upper mantle in the form of mantle plumes.

Viscosity, density differences and mineral grain size are the key factors that control mineral segregation. Geophysical observations have detected ultra-low-velocity zones (ULVZs) of different thicknesses (Williams and Garnero 1996; Wen and Helmberger

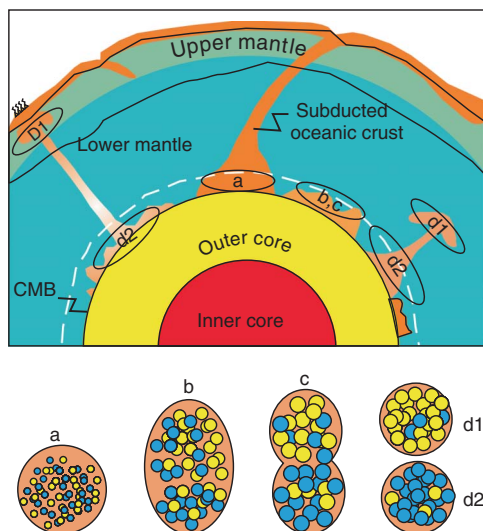


Figure 3. Illustration of the mineral segregation model. (a) Minerals in newly subducted oceanic crust are small, with evenly distributed heavy and light minerals; (b) and (c) minerals in subducted oceanic crust grow with time and may start to segregate from each other where the viscosity is low enough. After segregation, the light mineral enriched (D_1) portion can be lighter than the surrounding mantle and float up, whereas the heavy mineral enriched portion (D_2) remains in the CMB.

Table 1. Mineral densities and proportions in the lower mantle.

| | At 1100 km (g cm^{-3}) | At CMB (g cm^{-3}) | wt% in pyrolite | wt% in MORB | wt% in MORB ^a | wt% in MORB ^b | wt% in MORB ^c |
|---|--------------------------------------|----------------------------------|--------------------|----------------|-----------------------------|-----------------------------|-----------------------------|
| Mw (Mg# 83.5) | 4.62 | 5.68 | 22.00 | | | | |
| MgPv (5% Al_2O_3 ; Mg#92.7) | 4.63 | 5.50 | 69.00 | | | | |
| MgPv (20% Al_2O_3 ; Mg#71.0) | 4.71 | 5.60 | | 36.00 | 32.5 | 26.67 | 20.83 |
| CaPv | 4.73 | 5.56 | 9.00 | 31.00 | 26.25 | 18.33 | 10.42 |
| Stishovite | 4.61 | 5.28 | | 14.00 | 17.5 | 23.33 | 29.17 |
| Ca-ferrite (Mg, Fe, Na)AlSiO ₄ ss | 4.60 | 5.48 | | 19.00 | 23.75 | 31.67 | 39.58 |
| PREM | 4.64 | 5.56 | | | | | |
| Pyrolite | 4.64 | 5.55 | | | | | |
| MORB | 4.68 | 5.52 | | | | | |
| MORB ^a | 4.67 | 5.51 | | | | | |
| MORB ^b | 4.66 | 5.48 | | | | | |
| MORB ^c | 4.64 | 5.46 | | | | | |

Data in “italics” are calculated in this study, others are from Kesson *et al.* (1998). MORB = subducted basaltic oceanic crust.

^aMORB lost 10% MgPv and CaPv.

^bLost 20% MgPv and CaPv.

^cLost 26% MgPv and CaPv due to mineral segregation.

1998; Williams *et al.* 1998; Ni and Helmberger 2001a, 2001b; Rost and Revenaugh 2003; Rost *et al.* 2005), ranging from several km up to more than 100 km. The ULVZ has been interpreted as being partially melted based on seismic observations (Williams and Garnero 1996; Vidale and Hedlin 1998; Rost *et al.* 2005). In addition to ULVZs, there are also

very-low-velocity provinces (VLVPs) near the CMB, which are much larger in volume with slightly higher velocity than ULVZs. Detailed studies of a VLVP beneath the western Pacific clearly show melting features (He *et al.* 2006b). Further studies suggest that this kind of VLVP is closely related to mantle plumes (Wen 2006). Partial melting presumably should result in a reduced viscosity over that for the unmelted solid state, allowing separation of crust from the rest of the subducted slab (Christensen and Hofmann 1994). We propose that if the individual mineral grains are large enough, a segregation of heavy minerals from lighter ones will also occur.

At the CMB, the density of CaPv from the subducted basaltic oceanic crust is approximately the same as bulk pyrolite and the preliminary reference Earth model (PREM), while the density of MgPv is ~0.7% higher than bulk pyrolite. By contrast, stishovite and Ca-ferrite are about 5 and 1.5% lighter than bulk pyrolite, respectively (Table 1) (Kesson *et al.* 1998). Given the big density differences between these minerals and the low viscosity in ULVZs and VLVPs at the CMB, stishovite and Ca-ferrite may gradually segregate from the other two minerals in the ULVZs and VLVPs if they become large enough.

The size of minerals at the CMB cannot be directly determined at present. For the following reasons, we believe that minerals at the CMB, particularly in ULVZs and possibly also in VLVPs, may be large enough to cause a significant chemical differentiation. First, geophysical observation shows that mineral grain size increases with increasing depth in the mantle. For example, recent study suggests that the grain size in the upper mantle increases from millimetre size in the shallow upper mantle to centimetre size above the transition zone (Faul and Jackson 2005). Second, in principle, the grain size of a mineral in its stable P-T condition depends on time and diffusion rates (availability of materials). Because subducted oceanic crust apparently resides in the mantle for ~1–2 Ga before returning to the surface as estimated from radiogenic isotopic data (Zindler and Hart 1986; Hofmann 1997), major mineral phases in subducted oceanic crust are likely to have had enough time to grow large enough to initiate mineral segregation. Meanwhile, the diffusion rates increase with increasing temperature and, particularly, in the presence of fluids (melts). In the ULVZs and VLVPs, the temperature is known to be high, and partial melts almost certainly do exist (Williams and Garnero 1996; Wen and Helmberger 1998; Rost *et al.* 2005; He *et al.* 2006b), both conditions of which lead to significantly higher diffusion rates than those in the upper mantle. These conditions make it very likely that the grain size of minerals in the ULVZ as well as in VLVPs can be reasonably large.

Discussion

Plate subduction and mantle plumes are two important matter transfer processes of the silicate Earth, which are still not well understood. Among the major remaining questions are: whether subducted oceanic crust has been recycled back through mantle plumes and can account for the major chemical and physical characteristics of plumes; what is the main driving force for the origin of plumes; and why plume heads have huge melting capacities and are dramatically different from plume tails both in size and composition. These questions can be plausibly answered using a mineral segregation model.

The lower mantle density trap and the initiation of plumes

The lower mantle density trap was used to argue against recycling of subducted oceanic crust through mantle plumes (Niu and O'Hara 2003). We believe the higher density of subducted oceanic crust compared to the surrounding mantle is an essential feature linking

it to mantle plumes. Without high density, subducted oceanic crusts cannot stay in the lower mantle for $\sim 1\text{--}2$ Ga.

According to the mineral segregation model, the lower mantle density trap does not persist forever. The subducted oceanic crust can be released in portions that experience sufficient mineral segregation to form mantle plumes (Figure 4). In other words, the lower mantle is not a graveyard but rather a ‘holding reservoir’ for subducted oceanic crust.

The mineral segregation model proposes a genetic relationship between mantle plumes and ULVZs as well as VLVPs at the CMB, which are believed to be the roots of plumes because of their distribution correlation with hot spots (Williams *et al.* 1998; Burke *et al.* 2008) and unstable physical structures (Rost *et al.* 2005; Wen 2006). According to our mineral segregation model, at the early stage of formation of a plume (i.e. in the ULVZ at the CMB), it is mainly a chemical plume (controlled by mineral densities), which changes gradually to a thermal plume during ascent. Such a transition is compatible with seismic observations. For example, studies on the African superplume structures suggest that the superplume is not likely to be purely thermal, but rather a chemical plume (Ni *et al.* 2002; Ni and Helmberger 2003), similar to those observed by experimental models (Davaille 1999). Our mineral segregation model also implies that the formation of a plume results in a denser residue left behind it in ULVZs (and probably also VLVPs) at the CMB. This is again consistent with seismic observations, showing that the ULVZs are denser than the surrounding mantle (Rost *et al.* 2005).

Nb/U fractionation during mineral segregation

Our mineral segregation model requires that the subducted oceanic crust loses a portion of its Mg-Pv and Ca-Pv in the formation of plumes, such that the partition coefficient of Nb,

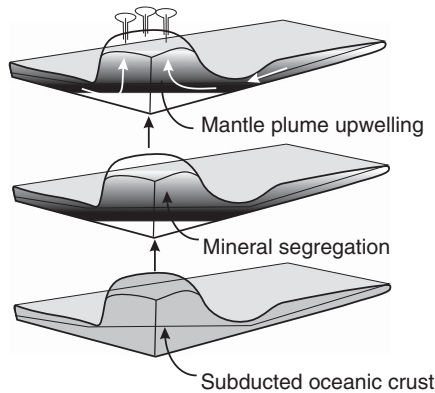


Figure 4. Illustration of the relationship between subducted oceanic crusts and mantle plumes. Subducted oceanic crusts are denser than the surrounding mantle in the lower mantle, and thus can be piled up and stay at the CMB for billions of years. The basaltic composition implies that subducted oceanic crusts start to melt at lower temperatures compared to mantle rocks. As the temperature increases because of thermal conduction as well as radioactivity, subducted oceanic crust may eventually be partially melted, forming ULVZs and VLVPs. Meanwhile, minerals in the subducted oceanic crust also grow with time. When minerals grow large enough, mineral segregation starts, such that the density within subducted oceanic crusts decreases from the bottom to the top. Once part of a subducted oceanic crust becomes lighter than the surrounding mantle, a buoyancy-controlled upwelling is initiated, which may eventually form a plume depending on the size and temperature differences.

U between different minerals might affect the Nb/U. There is as yet no partition coefficient data between Ca-Pv and Mg-Pv, Fe-Pv and other minerals at the P–T condition of the CMB. Nevertheless, recent experiments at considerably lower pressure (23–27 GPa) provide useful data for the partitioning between melt and Ca-Pv and Mg-Pv (Hirose *et al.* 2004; Walter *et al.* 2004; Corgne *et al.* 2005). The results suggest that most elements are incompatible in Mg-Pv, whereas Ca-Pv is the main carrier for most of the incompatible elements. For Nb, the partition coefficient between Mg-Pv and melt ranges from 0.011 to 0.23, while that for Ca-Pv and melt ranges from 0.244 to 2.15. The large variations are mainly due to different compositions used in the experiments. Based on results of experiment with coexisting Ca-Pv, Mg-Pv and melt (Hirose *et al.* 2004), the partition coefficient of Nb between Ca-Pv and Mg-Pv, $D_{(Ca/Mg)}(Nb)$ is about 7. For U, the partition coefficient between Mg-Pv and melt ranges from 0.022 to 0.16, whereas that for Ca-Pv and melt ranges from 7.3 to 15. Based on the same experiment (Hirose *et al.* 2004), the partition coefficient of U between Ca-Pv and Mg-Pv, $D_{(Ca/Mg)}(U)$ is about 25. For the subducted oceanic slab, the amounts of Ca-Pv and Mg-Pv are nearly equal (Table 1) (Kesson *et al.* 1998), such that most of the Nb and U are hosted by Ca-Pv, and the Nb/U value of Ca-Pv is only 10% lower than that of the bulk oceanic crust. Moreover, according to our mineral segregation model, the subducted crust loses roughly similar portions of Mg-Pv and Ca-Pv (Table 1), which would lead to minor fractionation between Nb and U. Therefore, the Nb/U characteristics of subducted oceanic crusts can be transferred to plumes.

Plume head

The mineral segregation model provides a mechanism allowing the involvement of subducted oceanic crusts in plumes, which can plausibly interpret the chemical features of plumes (White and Hofmann 1982; Hofmann 1997). It can also explain the major physical features of plumes, for example, the different characteristics of plume heads and tails. The major difference between plume heads and tails is the amount of melt, that is, melting capacity.

The enormous outpourings from plume heads are best represented by large igneous provinces, some of which cover an area of more than 2 million km², with a volume of more than 40 million km³ (Ernst *et al.* 2005), whereas plume tails are much smaller. The formation of huge plume heads has been puzzling for a long time, particularly as to why the plume heads have such a huge melting capacity. If their large amount of melt were purely due to high temperature, we would also see large degrees of melting and severe consumption of the mantle and overlaying crust. The reality is that products of high degrees of melting (Si-poor, e.g. komatiite and picrite) are rare, and usually occur only at the centre of the plume head (Ernst *et al.* 2005). Interestingly, plume heads usually initially erupt a large amount of basalt, followed by a small amount of komatiite (Campbell *et al.* 1989). Also, plume heads can lead to kilometre scale elevation of the crust (Griffiths and Campbell 1990; Griffiths and Campbell 1991; Xu *et al.* 2004; Campbell and Davies 2006; He *et al.* 2006a). These phenomena can be readily interpreted by a mineral segregation model.

Stishovite (SiO₂) is considerably lighter than other minerals in subducted oceanic crusts in the lower mantle. Consequently, the tops of plumes derived from subducted oceanic crust should become progressively enriched in SiO₂ as a result of mineral segregation during their ascent. It is a general rule that rocks with higher Si contents have a lower melting temperature. Therefore, mineral segregated oceanic crust (Si-enriched) should melt preferentially to the surrounding pyrolytic mantle. Subsequently, the early-formed melts do react

with surrounding mantle, which considerably lowers the solidus temperature of adjacent mantle rocks and produces large amounts of melts (Yaxley and Green 1998; Yaxley 2000; Sobolev *et al.* 2005; Sobolev *et al.* 2007). Therefore, the ascending Si-rich crust acts as a melting flux, which greatly enhances the melting capacity of the plume heads. As a result, large amounts of basalts are produced leaving behind a more refractory residue. After a huge melt extraction (i.e. eruption of plume heads), further melting of the refractory residual mantle is limited to the hottest part, forming a much smaller amount of komatiites and/or other Si-poor magmas (a mixture of the hot, Si-poor portion of the recycling crust and the adjacent depleted mantle rocks).

An ascending crust that is thermally equilibrated with the surrounding mantle must lose large amounts of its heavy minerals (Table 1) to get through the 'lower-mantle-density-trap'. Fortunately, the bulk ascending plume is likely to be considerably hotter than the surrounding mantle as it ascends from the CMB to the base of the upper mantle, thus requiring less mineral segregation. Such a hot ascending plume probably also incorporates mantle material on its way up. Some of the ascending recycled crust may travel all the way to the Earth's surface, as well-defined plumes. Other masses may stay in the lower or upper mantle and consequently change the chemical composition of the mantle, depending on the degree of mineral segregation, thermal conditions, and the total mass involved.

Subducted oceanic lithosphere mantle

Subducted oceanic lithosphere mantle metasomatized by the infiltration of low-degree partial melts should be enriched in highly incompatible elements with high U/Pb, Th/Pb (Sun and McDonough 1989), and thus can account for isotopic characteristics of some OIB, e.g. HIMU (Hofmann 2003). It also has been proposed that metasomatized deep portions of oceanic lithosphere mantle are the most likely candidate source for OIB, based on chemical compositions and mineral physics (e.g. the lower mantle density trap) (Niu and O'Hara 2003). Metasomatized oceanic lithosphere has a high content of radioactive elements (U, Th, K), and because it has been infiltrated by low-degree partial melts is itself likely to be melted at a lower temperature than the adjacent mantle rocks. Therefore, its unique chemistry may indeed have contributed to the formation of all plumes, and even controlled the isotopic characteristics of some plumes. For example, it may be a key factor that controls partial melting in the ULVZs and VLVPs. It may also offer highly incompatible elements to plumes. The current classification of OIB may simply be due to different contributions from metasomatized oceanic lithosphere and subducted oceanic crusts. Nevertheless, metasomatized oceanic lithosphere cannot account for the higher-than-primitive mantle Nb/U, which requires that the subducted oceanic crust be mixed back into it.

As for the 'lower mantle density trap', as discussed in 4.1, it is essential for subducted oceanic crusts to stay in the lower mantle for ~1–2 Ga before recycling back through the mantle plume.

Chemical filter between MORB and OIB

The mineral segregation model alone can explain how subducted oceanic crust can recycle back to the surface as mantle plumes (overcoming the density trap) and how large plume heads are produced (huge melting capacities coupled with Si-enriched melts). It, however, cannot readily explain the chemical difference coupled with the identical Nb/U values for MORB and OIB sources. Considering that initiating mineral segregation takes ~1–2 Ga, while subduction is a continuing process, 'background upwellings' are

also likely occurrences, which are probably responsible for the chemical and isotopic difference between MORB and OIB through 'transition-zone chemical filters', similar to the so called 'transition-zone water filter' model (Bercovici and Karato 2003). Given that the formation of the continental crust was the main mechanism that fractionated Nb from U, and that the average age of the continental crust is ~ 2.0 Ga (Hofmann 1997), the main fractionations of Nb and U, the descent of oceanic slabs into the lower mantle, and the complementary upward flows of lower mantle materials to the upper mantle must have started long ago. Therefore, the transition-zone chemical filter should have appeared very early, giving enough time for the two mantle layers to develop different isotopic compositions in MORB and OIB mantle sources. Melting experiments suggest that basaltic components could start to melt in the uppermost lower mantle if the mantle was 300–400 K hotter than its present temperatures (Hirose and Fei 2002). If the Archaean mantle was hot enough to be melted in the transition zone, then the 'transition-zone water filter' (Bercovici and Karato 2003) can be better described as a 'transition-zone melt filter': melt generated below the density crossover (Agee 1998; Ohtani and Maeda 2001) would not be able to rise into the upper mantle, thus keeping incompatible elements in the lower mantle, while the residue travels upward through background upwelling. By adding this depleted residue to the upper mantle, it becomes even more depleted without fractionating U from Nb.

Subducted early crust

Iron is a moderately incompatible element, and as such its abundance in the depleted MORB mantle (the source of MORB) is lower than in the primitive mantle. More importantly, mantle melting would start at a deeper depth in the hotter early Earth, and thereby result in higher Fe contents (Langmuir *et al.* 1992). Therefore, oceanic crust formed at the early stage of the Earth (before major depletion of the mantle) must have a higher Fe content. Whereas, as previously proposed, 'the respective regions of the crust, loaded with chondrite-like debris' (Tolstikhin and Hofmann 2005), were probably even more significantly enriched in Fe, they might not necessarily have had enough light minerals to segregate. They would have therefore stayed at the CMB, serving as an important source of incompatible elements and noble gases as proposed (Tolstikhin and Hofmann 2005).

Conclusion

The mean Nb/U values for MORB and OIB (46.7 ± 0.3 and 46.7 ± 0.5 , respectively) are both higher than that of the primitive mantle, consistent with previous estimates. Given that Nb and U are similarly incompatible and that the subducted oceanic crust is the main proxy for fractionating Nb/U, the elevated Nb/U signatures for MORB, OIB, and their corresponding mantle sources indicate that a considerable amount of subducted oceanic crust has been recycled back into the mantle sources of MORB and OIB. A mass balance calculation based on Nb/U shows that nearly the whole mantle has evolved by plate-subduction induced crust recycling and the formation of the continental crust.

Given that subducted oceanic crust is denser than the surrounding mantle both in liquid and solid states, it can stay in the lower mantle for billions of years as implied by isotopic data. Considering the low viscosity in ULVZs and VLVs, as well as significant density differences among major minerals in subducted oceanic crust at the CMB, mineral segregation is likely to occur after major minerals grow to a large enough size during their long term residence at the CMB. Portions of subducted oceanic crust that have lost a large

proportion of their heavy minerals, magnesian-silicate-perovskite (MgPv) and calcium-silicate-perovskite (CaPv), through mineral segregation, become SiO₂ enriched and lighter than the surrounding mantle, rising up as giant Si-rich plume heads due to flux melting, and followed by dramatically abridged plume tails.

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