Chapter 3

THE FATE OF SUBDUCTED OCEANIC CRUST AND THE ORIGIN OF INTRAPLATE VOLCANISM

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Abstract

Standard geodynamic models envisage subducted oceanic crust to be part isolated in thermal boundary layers and part remixed into the depleted mantle. The two fates for such material result from combining remixing models based on observation from orogenic lherzolites, with theoretical models for the generation of intraplate volcanism by mantle plumes. The concepts were combined because high ^{3}He/^{4}He ratios in intraplate basalts were interpreted to require a primitive source component and the convecting mantle was considered unable to retain ^{3}He on melting. However, high ^{3}He/^{4}He ratios may reflect low U+Th sources, and differences in ^{3}He/^{4}He between MORB and OIB can be explained by sampling of the convecting mantle. Interpretations of high ^{186}Os/^{188}Os in intraplate lavas as evidence for interaction with the core are likewise tenuous, as the signatures can be explained by pyroxenites or mantle sulphides. Instead, the remixing models should have been combined with models for the tapping of shallow mantle sources by plate tectonic processes, to give an explanation for the origin of intralate volcanism from the convecting mantle without plumes. Pyroxenitic sources for intraplate volcanism may be generated at convergent margins if subducted oceanic crust undergoes melting in the back-arc region, or along the flanks of convective upwellings beneath ocean ridge systems as melts from altered eclogite or sediment components of recycled crust react with peridotites of the depleted mantle. Generation of intraplate melts occurs in off-axis regions as a result of fluxing of the pyroxenite-veined mantle with fluids derived from dehydration or decarbonation of later generations of subducted slabs in the shallow mantle.

Introduction

The development of plate tectonic models in the 1960's led to the question of what is the fate of subducted oceanic crust? Early models (Armstrong 1968) suggested such material is recycled back toward mid ocean ridge systems by convection within the upper mantle.
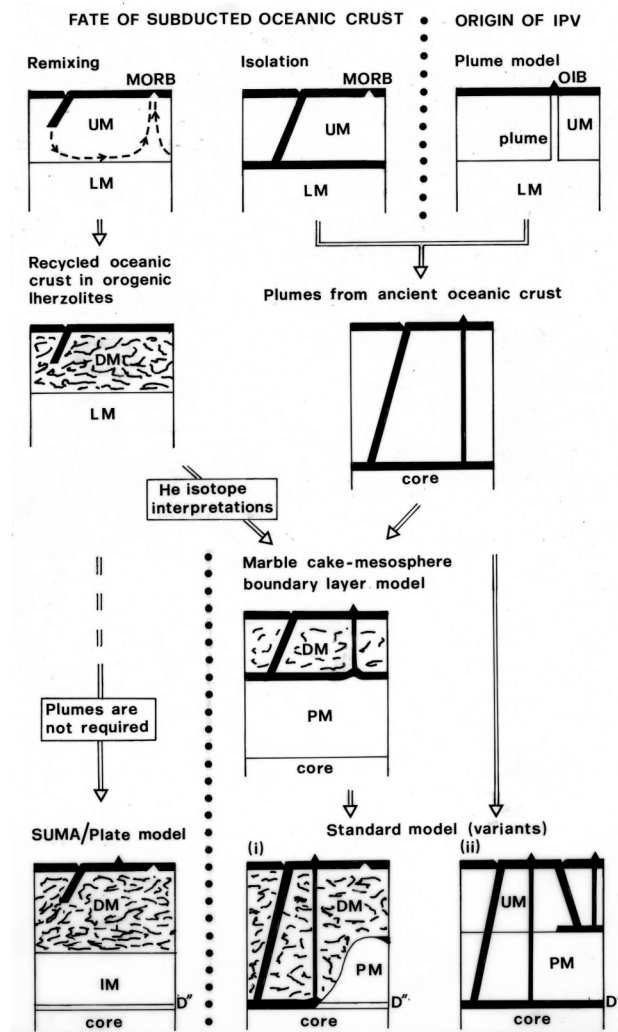
Figure 1. Evolution of concepts concerning the fate of subducted oceanic crust. Remixing (Armstrong 1968) and isolation (Dickinson and Luth 1971) of subducted material were proposed in parallel with development of the mantle plume model for the origin of intraplate volcanism (IPV) (Morgan 1971). The remixing model was supported by interpretation of pyroxenites in orogenic lherzolite massifs as recycled oceanic crust (Polvé and Allègre 1980), whereas the concept of isolating subducted crust was incorporated into the plume model of Hofmann and White (1982). The remixing and plume concepts were subsequently combined in the mesosphere boundary layer and marble-cake models of Allègre and Turcotte (1985, 1986), to give the standard geodynamic model on the basis of helium isotope interpretations. Several variants of the standard model exist (e.g. (i) Kellogg et al. 1999, (ii) Courtillot et al. 2003) because of difficulties in reconciling the requirements for a deep primitive mantle layer with geophysical models of mantle convection. Doubts about the existence of plumes and the ability of the plume model to provide a comprehensive explanation for intraplate volcanism, have resulted in the remixing model being revisited in the SUMA model of Meibom and Anderson (2004), which now forms an important component of the plate model where plumes are not required for the generation of intraplate volcanism. Abbreviations: DM = depleted mantle MORB source, IM = isolated mantle (may be of primitive composition or contain deeply subducted slabs which can not be entrained by mantle convection), LM = lower mantle, PM primitive mantle, UM = upper mantle.

Evidence for such remixing came from the interpretation of pyroxenite bands in orogenic lherzolite massifs as subducted oceanic crust that had been stretched out and thinned by several orders of magnitude, by convection within the mantle over periods of several hundred million years (Polvé and Allègre 1980). The structure of the centimetre-metre thick bands within a predominantly lherzolite matrix found in such massifs was labelled as 'marble-cake mantle' by Allègre and Turcotte (1986). Other models, however, suggested that subducted oceanic crust was isolated at depth rather than being remixed with the mantle (Dickinson and Luth 1971). This latter suggestion became an integral part of models for the origin of intraplate volcanism which had been developed in parallel with models for the fate of subducted oceanic crust, with the suggestion by Hofmann and White (1982) that subducted material collected in a thermal boundary layer at the base of the mantle before becoming buoyant and rising as mantle plumes. Both the concepts of remixing and isolating subducted oceanic crust are now part of the standard geodynamic model for the Earth, where it is envisaged that remixed oceanic crust is responsible for geochemical heterogeneity in the convecting mantle, whilst oceanic crust stored in thermal boundary layers serves as the source of intraplate volcanism (Fig. 1).

The number of studies based on the standard model, however, belies the number of unresolved problems with the model. These include uncertainties as to whether oceanic crust should be stored at the core-mantle boundary or base of the upper mantle, and the number of plumes (e.g. Anderson 2005a). Advocates of the plume model have argued that the uncertainties reflect a theory in its early stages of development (Sheridan 1994; Sleep 2007). However, it remains that plume model cannot readily account for intraplate volcanism away from where plumes impact beneath the lithosphere (e.g. Natland and Winterer 2005, Hirano et al. 2006), such that the standard model must involve two distinct origins for intraplate volcanism. Rather than seeking ad hoc mechanisms to make the plume model fit, it should be asked whether the duplicate origins for intraplate volcanism are a consequence of unnecessary incorporation of concepts during development of the standard model. The reasoning for combining two opposing models for the fate of subducted oceanic crust should therefore be re-evaluated. The argument presented in this study is that the problems with the standard model result from inclusion of the plume model, and that remixing subducted oceanic crust with the convecting mantle as in the plate model (Foulger 2002, 2007), can adequately explain the heterogeneity found in ocean-ridge (MORB) and intraplate (OIB) basalts.

Development of the Standard Model

Why Were Two Contrasting Models for the Fate of Subducted Oceanic Crust Included?

The concepts of remixing and isolating subducted oceanic crust were combined in the geodynamic models of Allègre and Turcotte (1985, 1986) on the basis of the interpretation of Allègre et al. (1983) regarding He isotope signatures in mantle-derived rocks. The isotope ^{3}He is primordial, whereas ^{4}He is primordial and radiogenic from the decay of ^{238}U, ^{235}U, ^{232}Th. The isotope ratios of helium are expressed as R/Ra where Ra is the atmospheric ratio of 1.38×10^{-6}. Early analyses of rock and gas samples indicated higher ^{3}He/^{4}He ratios in Hawaiian samples compared to MORB (e.g. Craig and Lupton 1976, Kurz et al. 1983).

Although it was noted that low ratios could reflect high U+Th in the source region, the principal focus was on interpretation of high ^{3}He/^{4}He as indicating an excess of primordial ^{3}He. The concept of mantle beneath the 660 km seismic discontinuity having a primitive composition had been suggested from crust-mantle evolution models in which the upper mantle MORB-source was considered the residue from formation of the continental crust (e.g. Allègre 1982). A primitive lower mantle had also been suggested from the chondritic Nd-Sr isotope compositions measured in continental basalts (Wasserburg and DePaolo 1979), and a deep primitive reservoir was suggested by Allègre et al. (1983) as the source of apparent excess primordial ^{3}He in Hawaiian basalts.

Although ^{3}He/^{4}He ratios ratios appeared to be higher in the source of intraplate volcanic rocks, the ratios in MORB indicated the convecting mantle was not devoid of ^{3}He. The latter feature was considered incompatible with degassing at ocean ridges, as it was considered that re-mixing with subducted oceanic crust would not be able to replenish the ^{3}He budget (Allègre and Turcotte 1985). Subducted oceanic crust, along with delaminated continental lithosphere following the model of McKenzie and O'Nions (1983), was thus suggested to collect at the base the upper mantle with ^{3}He introduced into the upper mantle by diffusion through, or perturbation of, the mesospheric layer of subducted material (Allègre and Turcotte 1985). Allègre and Turcotte (1986) elaborated on recycling subducted crust into the convecting mantle following Polvé and Allègre (1980), but as in Allègre and Turcotte (1985), allowed collection of a proportion of subducted oceanic crust in plume sources at the base of the upper mantle.

Recycling crust into the convecting mantle was resisted in some models on the basis that trace element signatures in Atlantic and Pacific MORB did not indicate the involvement of a crustal component (e.g. Rehkämper and Hofmann 1997, Hofmann 1997). However, other studies embraced the concept of isolating a fraction of subducted oceanic crust, estimated at 13% by Christensen and Hofmann (1994), in plume sources, whilst remixing the larger fraction with the convecting mantle as an explanation for heterogeneity in the depleted mantle MORB-source (e.g. Saunders et al. 1988). The notion of high ^{3}He/^{4}He ratios requiring a primordial component has remained one of the cornerstones of the plume model (e.g. Kellogg and Turcotte 1990a, van Keken et al. 2002), but the problem of reconciling the location of such a layer with geophysical evidence against layered mantle convection has been a persistent problem which has lead to a multitude of variations in the distribution of plume sources (e.g. Hofmann 1997, Albarède and van der Hilst 1999, Kellogg et al. 1999, Courtillot et al. 2003) (Fig. 1). Despite the variations, this model is referred to here as the standard model because of the widespread, largely unquestioned adoption of the concepts concerned.

A Core-Signature in Intraplate Volcanism?

The Re-Os isotopic system (^{187}Re decays to ^{187}Os) was refined as a tool for studying mantle evolution in the early 1990's. Suprachondritic ^{187}Os/^{188}Os ratios found in OIB were interpreted within the framework of the standard model to result from recycling of oceanic crust into plume sources (e.g. Hauri and Hart 1993). However, it was also suggested that the Os isotopic signatures could be generated by interaction of plume sources with the outer core (Walker et al. 1995). Support for the latter model, and hence isolation of plume sources at the core-mantle boundary, came with development of the Pt-Os isotope system (^{190}Pt decays to

^{186}Os) and the demonstration of coupled enrichments in ^{186}Os/^{188}Os – ^{187}Os/^{188}Os in picrites from Hawaii and komatiites from Gorgona Island (Brandon et al. 1999, 2003). Because ^{190}Pt is a low abundance isotope with a long half life, and recycled crust has low Os content, such material is unsuitable for explaining elevated ^{186}Os/^{188}Os signatures, which instead were used to infer the presence of up to 1% outer core material in plumes (Brandon et al. 1998).

The core interaction model was supported by Humayun et al. (2004) who suggested that high Fe contents of intraplate basalts relative to MORB, also resulted from interaction with the core. Coupled with suggestions that ^{3}He might be incorporated into plume sources from the core (Porcelli and Halliday 2001), placement of plume-sources at the core-mantle boundary appeared to present a solution to the paradox that whilst ^{3}He/^{4}He ratios could be interpreted as a primordial mantle signature, there was no evidence (e.g. Hofmann et al. 1986) for mixing with a primitive mantle reservoir in trace element signatures in intraplate lavas.

Melting Regimes in Plumes

The plume model explains the large volumes of some intraplate volcanic provinces by virtue of the thermal anomaly inherent in invoking a source deep in the mantle. The thermal anomalies allowed for plumes are typically 100-200^{0}C above the convecting mantle adiabat. However, low temperature plumes have been suggested (Cordery et al. 1997, Takahashi et al. 1998) and the possibility of a temperature range of 200^{0}C within the convecting mantle has been interpreted against there being any thermal anomaly involved in the generation of intraplate volcanism (Anderson 2000). Melting models have considered plumes to be composed of fertile peridotite or eclogite fragments embedded within a depleted peridotite matrix (e.g. Cordery et al. 1997, Kogiso et al. 1998, Kogiso and Hirschmann 2006). The amount of eclogitic in a plume has typically been estimated at 10-30% (e.g. Sobolev et al. 2007), although some models based on Os isotope calculations have required up to 90% recycled crust in a plume (Becker et al. 2000). Pyroxenitic sources have recently been suggested as more suitable than olivine-bearing sources in accounting for the high Ni and Si contents of OIB (Sobolev et al. 2005).
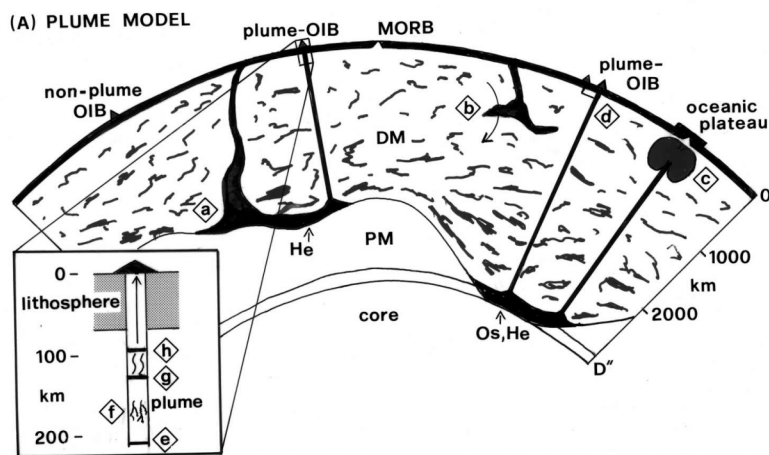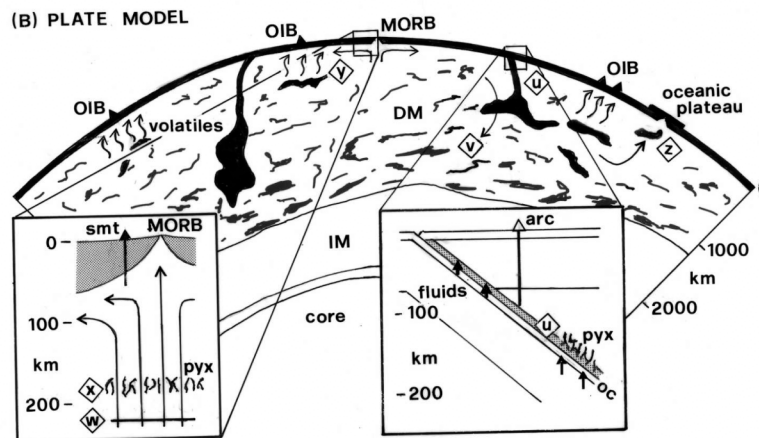

Figure 2. Continued on next page.

Figure 2. Comparison of the fate of subducted oceanic lithosphere in the plume (variant of Kellogg et al. 1999 illustrated) and plate models. In the plume model, subducted material is both isolated in thermal boundary layer plume sources [a] and remixed with the convecting mantle [b]. Material in plume sources becomes buoyant after 1-2 Gyr, rising as plumes. In the oceanic domain, plateau volcanism is considered to result from the impact of plume heads (albeit in conjunction with ocean ridge systems) [c], whereas plume tails form ocean island chains [d]. Detail inset illustrates melting regimes within a plume: [e] melting of eclogite [f] formation of pyroxenites, [g] melting of pyroxenites, [h] melting of peridotite (see also figure 3a). In the plate model, pyroxenites (pyx) may be formed as part of a layer of refertilized peridotite (shaded, inset) in the hanging wall of the mantle wedge at convergent margins [u], before remixing of slabs with the convecting mantle [v]. In the forearc region, the refertilized peridotite is generated from fluids/melts from the crustal layer (oc) of the slab, whereas in the back arc region, fluids/melts may be generated as a result of serpentinite dehydration in the slab as indicated by arrows. Pyroxenites may also be formed along the flanks of ocean ridge upwellings [x] as a result of melting of altered recycled eclogite [w]. Pyroxenites in the latter region may then be incorporated into the shallow sub-lithospheric mantle where they may undergo melting as a result of lowering of the solidus by invasion of CO_2- or H_2O- rich fluids [y] from thermal equilibration of younger subducted slabs at depths of 300-400 km in the mantle (see also figure 3b). Voluminous melting, such as involved in the generation of oceanic plateaus, is explained by entrainment of large fragments of subducted oceanic crust into ocean ridge upwelling [z].

The standard model is thus summarised in Figure 2 based on the model of Kellogg et al. (1999) that arranges primitive mantle and plume sources to reconcile interpretations of He isotope systematics with tomographic studies of the mantle. Some subducted slabs are deflected at the base of the upper mantle and undergo remixing with the convecting mantle, whereas others sink into the deep mantle to be isolated as plume sources at the core-mantle boundary or around 1600 km depth where geophysical models have indicated an increase in viscosity suggesting a change in mantle composition. Subducted material in the thermal boundary layers is usually envisaged to become buoyant after 1 to 2 Gyrs, with oceanic plateaus or continental flood basalt provinces formed by plume heads and ocean island chains formed by plume tails (Hofmann and White 1982, Hofmann 1997). Melting in a plume begins at approximately 200 km depth on crossing of the eclogite solidus (Sobolev et al. 2005) (Fig. 3a). The resulting melts react with the overlying mantle peridotite to form pyroxenites which subsequently undergo melting at shallower depths accompanied to varying extents by the surrounding peridotite, to form a range of intraplate melt compositions (Sobolev et al. 2005).
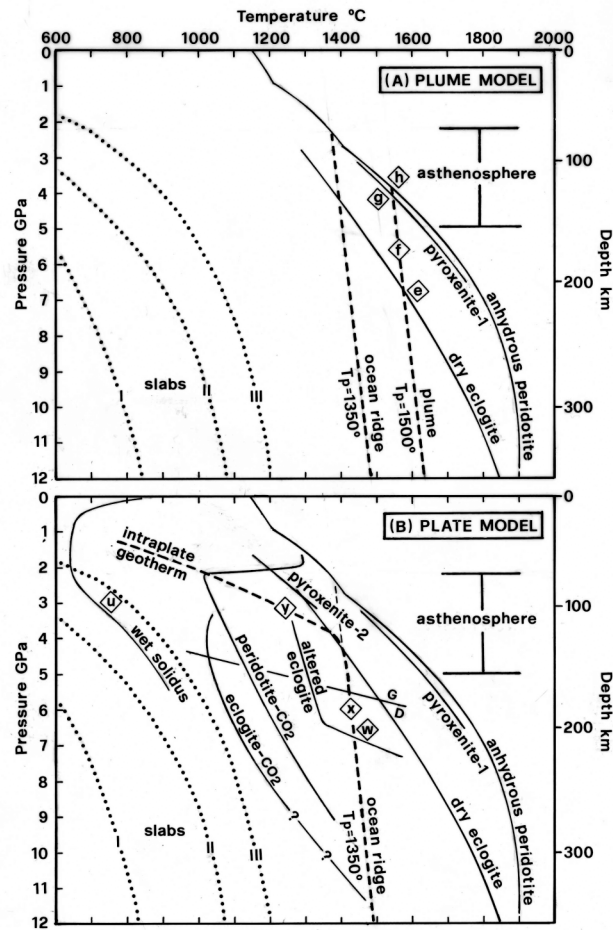
Figure 3. Generation of intraplate melts in P,T space according to the mantle plume and plate models. Stages outlined [e-h] and [w-y] correspond to the geodynamic models in figure 2. In the plume model, the eclogite component of a rising diapir (illustrated with a potential mantle temperature T_p of 1500^0C) begins to melt at [e], generating compositions that react with the overlying peridotite to form pyroxenites in the region [f]. The pyroxenite and peridotite solidii are exceeded at [g] and [h], respectively, generating a range of intraplate melts. In the plate model, upwelling results from mantle convection, and is illustrated for a potential mantle temperature of 1350^0C. The solidus for altered eclogite is reached at [w], generating melts which react with the overlying peridotite in the region [x]. Along the flanks of an ocean ridge system, the solidus would not be reached, allow incorporation of pyroxenites into the shallow sub-lithospheric mantle. However, generation of intraplate melts could be achieved along the intraplate geotherm at [y] by fluxing with volatiles such as CO_2 from the devolatilization of slabs at depth (note intersection of eclogite-CO_2 solidus with adiabat at ~ 11 GPa), in which case the pyroxenite solidus should be suppressed similar to the solidus of perdiotite-CO_2 and eclogite-CO_2 relative to volatile-free compositions. Solidii: anhydrous peridotite (Green and Falloon 2005), pyroxenite-1 (MIX1G of Hirshmann et al. 2003, Kogiso et al. 2003), pyroxenite-2 (MORB-like pyroxenite of Pertermann and Hirschmann 2003), dry eclogite (Yasuda et al. 1994), altered eclogite (Spandler et al. 2007), peridotite-CO_2 (Presnall and Gudfinnsson 2005, Dasgupta et al. 2007), eclogite-CO_2 (Dasgupta et al. 2004), wet solidus (basalt, greywacke and pelite compositions; Schmidt et al. 2004). P,T profiles for top of subducting slabs: I = old lithosphere, fast subduction, II = old lithosphere, slow subduction, III = young lithosphere, slow subduction, are from Kincaid and Sacks (1997). G-D is the graphite-diamond transition.

Consequence: Duplication of Mechanisms

The trend over the last two decades has been to assume a plume origin for intraplate volcanism, but despite its widespread application, significant weaknesses have been pointed out in the plume model. Few examples of intraplate volcanism conform to the predictions of the plume model (e.g. Sheth 1999, Anderson 2005a), but perhaps the greatest problem is the plurality of mechanisms, which applies not only to the fate of subducted oceanic crust, but also to the origin of the intraplate volcanic rocks the model was devised to explain in the first place. The plume model was originally developed as an explanation for linear age-progressive ocean island chains (Morgan 1971). However, much of the intraplate volcanism in the Pacific basin is neither linear nor age progressive (Natland and Winterer 2005). Likewise, much of the intraplate volcanism in regions such as Africa and Asia shows little temporal or spatial variation that could be linked to plumes, but shows a strong relationship with lithospheric architecture and can be readily explained by plate tectonic processes (Flower et al. 1998, Smith 1998, Bailey and Woolley 2005, Liégeois et al. 2005).

To explain all examples of intraplate volcanism by the plume model would also require a large number of plumes. Estimates of the number of plumes have ranged up to 5240 (Malamud and Turcotte 1999), but most are between 7 (Courtillot et al. 2003) and 42 (Crough and Jurdy 1980), which would require a non-plume origin for most examples of intraplate volcanism. Proponents of the plume model have appealed to "flow" of plume material for thousands of kilometres through the asthenosphere (e.g. Ebbinger and Sleep 1998, Niu et al. 2002) or considered the upper mantle to be entirely composed of plume residues (e.g. Morgan et al. 1995). But even with such modifications it remains that non-plume processes would still be required to tap the migrating plume material or residues.

What if the Marble-Cake and Plume Models Had not Been Combined?

A Path Followed Two Decades Later

Non-plume mechanisms which have recently been applied to various examples of intraplate volcanism include lithospheric loading (Hieronymus and Bercovici 2000, Got et al. 2008), shallow mantle convection (Ballmer et al. 2007), and propagating fractures (Stuart et al. 2007). While such mechanism could be used to alleviate difficulties with the plume model, it should be remembered that these are modern versions of concepts which were suggested as explanations for intraplate volcanism (e.g. Jackson and Shaw 1975, Bonatti and Harrison 1976, Walcott 1976) before geodynamic thinking became dominated by the plume model. Had the emphasis not switched to exploring the plume model, it is tempting to postulate that derivation of both MORB and OIB from the convecting mantle could have been explored in conjunction with such models in the 1980's. At that time, preferential melting of fertile heterogeneities in the ocean ridge environment had been proposed for the origin of alkalic axial seamounts which show many similar geochemical features to OIB (Zindler et al. 1984). Generation of OIB from a veined mantle without plume influence had also been explored (Fitton and James 1986).

The possibility of generating MORB and OIB from a common reservoir was also suggested by the study of Hamelin and Allègre (1988), which showed the pyroxenite layers in orogenic lherzolite massifs to cover the range of Pb isotopic composition found in both MORB and OIB. Petrological studies in the 1990's on the orogenic lherzolites, however, indicated two population of pyroxenite to be present in the massifs, with only a small proportion of the pyroxenite layers having the geochemical characteristics of recycled oceanic crust (Kornprobst et al. 1990). The majority of pyroxenites were interpreted to have formed as high pressure cumulates of melts that had intruded the peridotites, and a lack of relationship between radiometric ages and thickness of the layers was cited against the marble-cake concept (Pearson et al. 1993, Pearson and Nowell 2004). However, subducted materials are likely to undergo melting at some stage during remixing with the mantle (Anderson 2006), and the complexity observed in the lherzolite massifs can be considered evidence that such a process has happened for a large proportion of the recycled material.

Instead, it was not until the statistical mantle assemblage (SUMA) model of Meibom and Anderson (2004) that generation of both MORB and OIB from oceanic crust remixed with the convecting mantle was re-considered. In the SUMA model, subducted oceanic crust is remixed entirely with the convecting mantle, with a greater proportion of recycled oceanic crust involved in the generation of OIB relative to MORB. Voluminous examples of intraplate volcanism are explained by invoking a greater size range in recycled materials than in the marble-cake model. MORB and OIB do not form simple endmembers in isotopic space, but this can be resolved by invoking either the recycling of additional materials such as delaminated lower crust or continental mantle (e.g. Anderson 2005b, Lustrino 2005, Ishikawa et al. 2007), or modifications to subducted material such as slab melting in addition to the slab dehydration processes usually considered (Smith 2005). Development of the SUMA model accompanied renewed interest in stress fields and lithospheric architecture as controls on intraplate volcanism (e.g. Favela and Anderson 2000, Smith, 2003a), and the concepts became integral parts of what is now the plate model where plumes are not required to explain any aspect of intraplate volcanism (Foulger 2002, 2007, Foulger and Natland 2003).

Helium Isotopes Re-visited

The standard model was founded on beliefs that the convecting mantle should have been degassed of ^{3}He which could not be replaced by remixing of oceanic crust, and that high ^{3}He/^{4}He ratios indicated an excess of ^{3}He. Experimental studies have now shown that He may reside in olivine and be more compatible on melting than U or Th as the bulk distribution coefficients for the latter decline rapidly as clinopyroxene and garnet are consumed (Parman et al. 2005). Harzburgitic and dunitic residues from melting at ocean ridges could therefore retain ^{3}He in addition to being characterised by high ^{3}He/^{4}He, and impart such signatures to the convecting mantle upon recycling. Introduction of ^{3}He into the convecting mantle by subduction of cosmogenic dust could further supplement the He budget of this reservoir (Anderson 1993). The interplanetary dust particles were suggested to retain some volatiles to temperatures of up to 950^0C, which would allow survival to depths of 400 km depending on slab geotherm. Such replenishment was generally dismissed following the work of Staudacher and Allègre (1988) that suggested subducting slabs are degassed of noble gases in the subduction zone, but recycled atmospheric noble gas signatures have been found in

Alpine ultramafic rocks of the Horoman peridotite (Matsumoto et al. 2001), indicating that noble gases can survive the subduction barrier.

The interpretation of high ^{3}He/^{4}He ratios indicating a primitive source is problematic because He contents are higher in MORB than OIB: the converse would be expected if OIB were derived from a primitive and therefore relatively undegassed source (Anderson 1998a,b). Rather than reflecting an abundance of ^{3}He, high ^{3}He/^{4}He ratios can be interpreted as a deficit in ^{4}He from a source with low U+Th (Anderson 1998a,b). If the latter model is correct, there should also be a correlation between low ^{206}Pb/^{204}Pb and high ^{3}He/^{4}He, as ^{206}Pb is the end product of one of the principal decay series producing ^{4}He. Although the inverse relationship, high ^{206}Pb/^{204}Pb with low ^{3}He/^{4}He has been demonstrated for high-μ (HIMU) OIB (e.g. Hanyu and Kaneoka, 1997), OIB with low ^{206}Pb/^{204}Pb have long been considered to show both high (Iceland) and low (Gough, Tristan da Cunha) ^{3}He/^{4}He (e.g. Zindler and Hart 1986). However, in a recent study by Class and Goldstein (2005), a significant correlation was noted between low ^{206}Pb/^{204}Pb and high ^{3}He/^{4}He when OIB with similar neodymium isotopic composition were compared.

Class and Goldstein (2005) interpreted their results relative to a convecting mantle that underwent a reduction in ^{3}He/^{4}He ratio through time as a result of recycling of relatively He-deficient, but U-rich recycled crust over time. High ^{3}He/^{4}He -low ^{206}Pb/^{204}Pb signatures were suggested to be maintained in ancient mantle that had been isolated from remixing with the convecting mantle by storage in plume sources. However, there is no reason, other than conformity with the standard model, why the older reservoir should be equated with plume sources instead of unmelted streaks of recycled oceanic crust in the convecting mantle. The range of ^{3}He/^{4}He ratios in MORB (9 ± 4 unfiltered; Anderson 2000) lies within the range shown by intraplate basalts (5 to >40). The correlations observed by Class and Goldstein (2005) are therefore compatible with sampling of a common source as proposed to explain the He isotopic variation in MORB and OIB by Anderson (2000) and Meibom et al. (2003). The narrow range in MORB may thus only reflect homogenisation during more extensive melting than involved in the generation of intraplate basalts.

Osmium Isotopes and Heterogeneity in the Convecting Mantle

The concept of a core Os signature in some intraplate lavas has been a key piece of evidence for an ultra-deep origin for the sources of such volcanism, but the interpretations rest on a series of assumptions, detailed counter-arguments to which are given in Meibom et al. (2004). Essentially, formation of the inner core has to take place shortly after accretion of the Earth, whereas thermal modelling (Labrosse et al. 2001) has indicated much later formation. Addition of outer core material should also impart tungsten isotopic variations that are not observed in intraplate basalts (Scherstén et al. 2004). The tungsten and osmium isotope signatures were suggested to have been decoupled by percolation of FeO-saturated melts through the lowermost mantle during the early history of the Earth (Humayun et al. 2004), but non-plume mechanisms involving remixing of sediments, and isotopic evolution in pyroxenites and sulphides have also been suggested as means of generating high ^{186}Os/^{188}Os in OIB.

Most crustal materials have too low Os content and Pt/Os ratio to produce the high ^{186}Os/^{188}Os ratios found in Hawaiian lavas. However, metalliferous sediments are an

exception and it has been suggested that recycling of a few percent of such material into the source of intraplate melts could explain the $^{186}Os/^{188}Os$ – $^{187}Os/^{188}Os$ compositions of Hawaiian lavas (Ravizza et al. 2001, Baker and Jensen 2004, Scherstén et al. 2004). Nielsen et al. (2006) also noted that remixing of ferromanganese sediments into the mantle could explain the Tl-isotope systematics of Hawaiian lavas, but as there was no correlation between Tl and $^{186}Os/^{188}Os$, these authors preferred the plume model to explain Os isotope variations and the sediment recycling model to explain Tl isotopic variations. Humayun et al. (2004) discounted a ferro-manganese sediment component for generating the Os isotope variation, however, on the basis of there being no complementary enrichment in Mn in the Gorgona komatiites.

Combined Pt, Os and Re analyses on pyroxenites are rare, but samples from ophiolite suites (Bay of Islands and Urals belt; Edwards 1990; Garuti et al. 1997) have high Pt/Os and Re/Os ratios that are expected to result in generation of supra-chondritic $^{186}Os/^{188}Os$ – $^{187}Os/^{188}Os$ ratios (Smith 2003b). High Pt/Os ratios have also been reported for pyroxenites from the Beni Bousera orogenic lherzolite massif by Luguet et al. (2008), who estimated the samples to show a comparable range of $^{186}Os/^{188}Os$ ratios to Hawaiian lavas by calculating their isotopic composition from an assumed age and depleted mantle initial ratio. The calculated signatures were supported by two measured $^{186}Os/^{188}Os$ ratios that bracketed the variation found in the Hawaiian lavas. The Pt, Os abundances in the Beni Bousera pyroxenites reported by Luguet et al. (2008) are also similar to those of Hawaiian picrites analysed by Brandon et al. (1998), which would be consistent with derivation of the lavas by large degrees of melting of a pyroxenitic source. Nonetheless, Os contents in pyroxenites are still around a thousand times lower than in mantle sulphides, suggesting the latter are more likely to control Os isotope compositions in the mantle (Luguet et al. 2008). Sulphides from ophiolite suites display a much wider range in $^{186}Os/^{188}Os$ than the Hawaiian lavas, and the isotopic composition of a melt could be controlled by only a small amount of such minerals in its source (Meibom et al. 2004). Such interpretations suggest the range of $^{186}Os/^{188}Os$ estimated for the depleted mantle by Brandon et al. (1999) was too restricted, and that sufficient heterogeneity exists within the convecting mantle to explain the $^{186}Os/^{188}Os$ signatures of intraplate volcanic rocks without need to invoke interaction with the core (Luguet et al. 2008, Meibom 2008).

Production of Pyroxenitic Sources in the Convecting Mantle

The crustal recycling process envisaged in the plate model is outlined in figure 2b. After termination of subduction, slabs are entrained and thinned by mantle convection as in the marble-cake model, although inefficient remixing or stagnation of deeply subducted slabs in the lower mantle (Hirose et al. 1999) will not affect the following model as plume processes are not invoked. The time for thermal equilibration of subducted material has been estimated at 100-200 Myrs (Stein and Stein 1997), hence slabs will be thermally equilibrated with the mantle within timeframes for convective stirring (240-960 Myrs depending on layered or whole mantle convection; Kellogg and Turcotte, 1990b). The ongoing subduction of oceanic lithosphere will result in recycled crust with varying thickness and state of thermal equilibration being present within the mantle (Anderson 2006). Such material is unlikely to be homogeneously distributed, as localisation of subduction regimes for tens of millions of years

may produce large-scale mantle domains characterised by hydrous or enriched mantle interspersed with large accumulations of slabs as proposed to underlie the western Pacific region (Komiya and Maruyama 2007).

At convergent margins, the subducted material first undergoes modification as a result of dehydration or melting of the slab (Poli and Schmidt 2002, Smith 2005). The wet solidus, which is similar for basalt and sediment compositions, is intersected by the geotherm for hot slabs at depths of 80-100 km, suggesting that slow subduction of young slabs will result in melting (Schmidt et al. 2004) (Fig. 3b). Slab melting may have been more common in the Precambrian, but the majority of modern slabs likely undergo dehydration. The latter process takes place in the fore-arc region, such that the basaltic and sedimentary layers of the slab are essentially anhydrous by the time the slab lies beneath the arc region (e.g. Tatsumi and Eggins 1995). Water driven off the slab produces volatile-bearing minerals in the hanging wall of the mantle wedge. Convection induced in the mantle wedge by descent of the slab, drags the volatile-bearing assemblages to greater depth, whereupon the breakdown of amphibole to phlogopite, followed by phlogopite to K-richterite, releases fluids that lead to the generation of arc volcanism (e.g. Tatsumi and Eggins 1995). K-richterite is stable to depths of 300-400 km above cold slabs and formation of this mineral is a possible means of introduction of water into the deep mantle (Konzett et al. 2000).

Pyroxenite-rich peridotites in the Solomon Islands (Berly et al. 2006) and Cabo Ortegal in Spain (Santos et al. 2002) have been suggested to form by fluid metasomatism of the shallow mantle, but their compositions are too low in Al_2O_3 to constitute a source for intraplate volcanism. Pyroxenites may also be formed at greater depth in the mantle wedge as a result of metasomatism by slab-derived fluids/melts generated by fluxing of the crustal layers of the slab by fluids from the breakdown of hydrous minerals in the peridotitic layers of the slab (Ringwood 1990) (Fig. 2b). Such a model has been suggested for formation of the sources of E-MORB (Donnelly et al. 2004), and a mantle wedge setting has been suggested for formation of pyroxenites in the Beni Bousera massif (Davies et al. 1993, Pearson et al. 1993). However, other studies (Sánchez-Rodriguez and Gebauer 2000) have suggested formation of pyroxenites in the Beni Bousera massif during mantle upwelling associated with the opening of the Tethys Ocean, which leads to an alternative mechanism of generation of pyroxenites along the flanks of ocean ridge upwellings.

The sequence of melting for upwelling convecting mantle containing streaks of recycled eclogite in a peridotite matrix has similarities with the plume model of Sobolev et al. (2005): the ocean ridge adiabat intersects the dry eclogite solidus around 120 km depth (Fig. 3b), generating melts which react with the overlying mantle to form pyroxenites which may subsequently play a role in the genesis of MORB (Hirschmann and Stolper, 1996). However, the eclogite solidus is dependent on the K content of the recycled material, and is suppressed between 2 and 6 GPa in altered eclogite compositions (Spandler et al. 2007). Recycled sediments are also likely to have high K content, and are postulated to show a similar melting behaviour to altered eclogite at such pressures. Melts generated from altered eclogite or recycled sediments could therefore react with mantle peridotites to form pyroxenites at around 180 km depth. It is postulated here that the pyroxenites thus formed would have a similar composition to the compositions reported by Pearson et al. (1993) which crystallized in the stability field for diamond and show isotopic evidence for a sediment component in their source. Such pyroxenites have MgO contents intermediate between the MORB-like and refractory compositions previously investigated in melting experiments (Hirschmann et al.

2003, Pertermann and Hirschmann 2003, Kogiso et al. 2003) and their solidii are predicted to lie between the curves for such compositions in figure 3b. The average intraplate geotherm would not intersect the solidus for such pyroxenites, which if formed along flanks of ocean ridge upwellings, could then be incorporated into the shallow mantle in the off-axis region (Fig. 2b). Because of the curvature of the intraplate geotherm away from the pyroxenite solidus at pressures around 4 GPa, this model is insensitive to the potential mantle temperature of the upwelling mantle.

The off-axis geotherm may be raised a few tens of degrees Celsius by shear heating in the asthenosphere (Smith and Lewis 1999, Doglioni et al. 2005), but the principal cause of melting is likely to be lowering of the solidus by the introduction of volatiles. Very cold slabs may be capable of transporting H_2O to the transition region (Ivanov and Litasov 2008). Similarly, CO_2 may be transported to the transition region in carbonated eclogite as the solidus for such material is not intersected by any of the potential slab geotherms (Dasgupta et al. 2004) (Fig. 3b). As such slabs undergo thermal equilibration with the mantle, however, they may release volatiles which would then migrate upwards following the intraplate geotherm in thermal regime (Dasgupta et al. 2004). The presence of volatiles markedly reduces the solidii for peridotite and eclogite compositions (e.g. Green and Falloon 2005, Presnall and Gudfinnsson 2005), and is expected to produce a similar effect on the pyroxenite solidus. Undersaturated intraplate melt compositions are thus suggested to be produced along the intraplate geotherm at temperatures of between 1200-1300^{0}C and approximately 3 GPa pressure in figure 3b, in agreement with the P,T estimates of Green and Falloon (1998) for the generation of intraplate melts and the observations of Frezzotti and Peccerillo (2007) for the presence of CO_2- rich fluids in the mantle beneath Hawaii. The composition of sources for intraplate volcanism may thus be controlled by processes at convergent margins, with generation of melts controlled by the volatile content of the shallow mantle.

Conclusion

Two competing models for the fate of subducted oceanic crust were proposed following the acceptance of plate tectonics: remixing with the convecting mantle and isolation at depth in the mantle. The former was supported by evidence from orogenic lherzolite massifs and became the marble-cake mantle model. The latter, which remains speculative, was developed into the mantle plume hypothesis for the origin of intraplate volcanism. The marble-cake model was combined with the plume model, thereby giving rise to modern geodynamic models, on the basis of interpretations of rare gas isotope systematics in the 1980's. The model that resulted was subsequently supported by interpretations of Os isotope systematics as indicating a component from the Earth's core in the source of intraplate volcanic rocks.

Helium isotope models developed over the last decade have shown the original interpretations to be non-unique. Likewise, Os isotope systematics in intraplate volcanic rocks can be interpreted as evidence for a pyroxenite or sulphide rather than core signature. Instead of incorporating both marble-cake and plume models, it is suggested a geodynamic model based on remixing subducted crust solely into the convecting mantle can account for the geochemical features of ocean ridge and intraplate volcanism. The duplication of mechanisms that must otherwise result to explain plume- and non-plume intraplate volcanism in the

standard model is, in effect, a consequence of unnecessary duplication of fates for subducted oceanic crust.

In the proposed model, pyroxenitic sources for intraplate volcanism are formed at depth in the mantle wedge at convergent margins, or along the flanks of ocean ridge upwellings as melts from subducted oceanic crust react with mantle peridotites. The pyroxenites are subsolidus in the P,T regimes of the shallow off-axis mantle, but may undergo melting to produce intraplate melts in the presence of volatiles, for example, if the shallow mantle is refluxed with CO_2 from the breakdown of subducted carbonates in the transition region. The control on the location of intraplate volcanism in the model is plate tectonic, not the result of deep-seated plumes.

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