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The planet beyond the plume hypothesis

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Abstract

Acceptance of the theory of plate tectonics was accompanied by the rise of the mantle plume/hotspot concept which has come to dominate geodynamics from its use both as an explanation for the origin of intraplate volcanism and as a reference frame for plate motions. However, even with a large degree of flexibility permitted in plume composition, temperature, size, and depth of origin, adoption of any limited number of hotspots means the plume model cannot account for all occurrences of the type of volcanism it was devised to explain. While scientific protocol would normally demand that an alternative explanation be sought, there have been few challenges to ‘‘plume theory’’ on account of a series of intricate controls set up by the plume model which makes plumes seem to be an essential feature of the Earth. The hotspot frame acts not only as a reference but also controls plate tectonics. Accommodating plumes relegates mantle convection to a weak, sluggish effect such that basal drag appears as a minor, resisting force, with plates having to move themselves by boundary forces and continents having to be rifted by plumes. Correspondingly, the geochemical evolution of the mantle is controlled by the requirement to isolate subducted crust into plume sources which limits potential buffers on the composition of the MORB-source to plume- or lower mantle material. Crustal growth and Precambrian tectonics are controlled by interpretations of greenstone belts as oceanic plateaus generated by plumes. Challenges to any aspect of the plume model are thus liable to be dismissed unless a counter explanation is offered across the geodynamic spectrum influenced by ‘‘plume theory’’. Nonetheless, an alternative synthesis can be made based on longstanding petrological evidence for derivation of intraplate volcanism from volatile-bearing sources (wetspots) in conjunction with concepts dismissed for being incompatible or superfluous to ‘‘plume theory’’. In the alternative Earth, the sources for intraplate volcanism evolve from the source residues of arc volcanism located along sutures in the continental mantle. Continental rifting and the lateral distribution of intraplate sources in the asthenosphere are controlled by Earth rotation. Shear induced on the base of the asthenosphere from the mesosphere as the Earth rotates is transmitted to the lithosphere as basal drag. Attenuation of the drag due to the low viscosity of the asthenosphere, in conjunction with plate motions from boundary forces, results in a rotation differential of up to 5 cm yr^{-1} between the lithosphere and mesosphere manifest as westward plate lag/eastward mantle flow. Continental rifting results from basal drag supplemented by local convection induced by lithospheric architecture. Large continental igneous provinces are generated by convective melting, with passive margin volcanic sequences following the axis of rifting and flood basalts overlying the intersection of sutures in the continental mantle. As rifting progresses, the convection cells expand, cycling continental mantle from sutures perpendicular to the rift axis to generate intraplate tracks in the ocean basin. Continental mantle not melted on rifting, or delaminated on continental collision, becomes displaced to the east of the continent by differential rotation, which also sets up a means for tapping the material to give fixed melting anomalies. When

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plates move counter to the Earth's rotation, as in the example of the Pacific plate, asthenospheric flow is characterised by a counterflow regime with a zero velocity layer at depths within the stability field for volatile-bearing minerals. Intraplate volcanism results when melts are tapped from this stationary layer along lithospheric stress trajectories induced by stressing of the plate from variations in the subduction geometry around the margins of the plate. Plate boundary forces acting in the same direction as Earth rotation, as for the Nazca plate, produce fast plate velocities but not counterflow, though convergent margin geometry may still induce propagating fractures which set up melting anomalies. Lateral migration of asthenospheric domains allows the sources of Pacific intraplate volcanism to be traced back to continental mantle eroded during the breakup of Gondwana and the amalgamation of Asia in the Paleozoic. Intraplate volcanism in the South Pacific therefore has a common Gondwanan origin to intraplate volcanism in the South Atlantic and Indian Oceans, hence the DUPAL anomaly is entirely of shallow origin. Such domains constitute a second order geochemical heterogeneity superimposed on a streaky/marble-cake structure arising from remixing of subducted crust with the convecting mantle. During the Proterozoic and Phanerozoic, remixing of slabs has buffered the evolution of the depleted mantle to a rate of 2.2 εNd units Ga^{-1}, with fractionation of Lu from Hf in the sediment component imparting the large range in ^{176}Hf/^{177}Hf relative to ^{143}Nd/^{144}Nd observed in MORB. Only the high εNd values of some Archean komatiites are compatible with derivation from unbuffered mantle. The existence of a very depleted reservoir is attributed to stabilisation of a large early continental crust through either obduction tectonics or slab melting regimes which reduced the efficiency of crustal recycling back into the mantle. Generation of komatiite is therefore a consequence of mantle composition, and is permitted in ocean ridge environments and/or under hydrous melting conditions. Correspondingly, as intraplate volcanism depends on survival of volatile-bearing sources, its appearance in the Middle Proterozoic corresponds to the time in the Earth's thermal evolution at which minerals such as phlogopite and amphibole could survive in off-ridge environments in the shallow asthenosphere. The geodynamic evolution of the Earth was thus determined at convergent margins, not by plumes and hotspots, with the decline in thermal regime causing both a reduction in size of crust and continental mantle roots, the latter becoming a source for intraplate volcanism while the crust was incorporated into the convecting mantle.

Keywords: intraplate volcanism; plate tectonics; plumes; earth rotation; stress fields; mantle evolution

1. Introduction

During the first part of the 20th century there was a great debate exemplified by the views of Taylor (1910), Wegener (1920) and Jeffreys (1924) over the concept of continental drift. The views of those in favour of maintaining the status quo triumphed, and it was not until a half century later that the observations necessary to confirm the concept of continents moving over the Earth's surface were able to be made. The acceptance of plate tectonics was accompanied by the rise of a second major geological theory, the mantle plume concept, which now dominates mantle geodynamics from its use both as an explanation for intraplate volcanism and a reference frame by which plate motions are modelled. Like the concept of continental drift, the plume model is either right or wrong. But in contrast to the situation earlier this century, several new geochemical and geophysical tools such as the Sm–Nd, Lu–Hf, Re–Os isotopic systems and seismic tomography, have become available since the plume model was proposed. Despite these new methods, geodynamic concepts of the Earth have changed little over the past two decades (Figs. 1 and 2). A proponent of the plume model would argue that this reflects a fundamentally sound basis such that the new observations merely added details to pre-existing concepts as statements that the plume model has been confirmed by Os isotopes (Hauri and Hart, 1993; Shirey, 1994) or that plumes have been observed by seismic tomography (Nataf and VanDecar, 1993; Wolfe et al., 1997; Helmberger et al., 1998; Ji and Nataf, 1998; Bijwaard and Spakman, 1999) would lead to believe. However, it is not only the concepts that have remained unchanged, but also the difficulties associated with these, such that the complex picture of the Earth that emerges in the plume model is compounded by uncertainties in the number of hotspots, the depth of origin of plumes, whether hotspots are

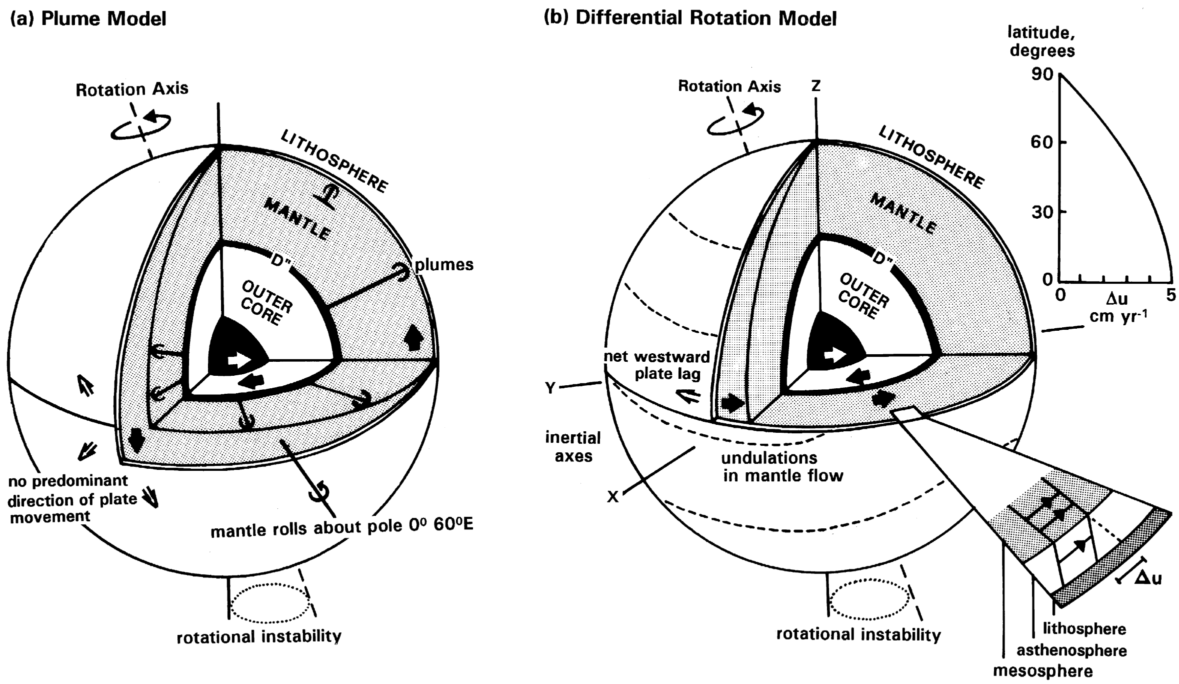
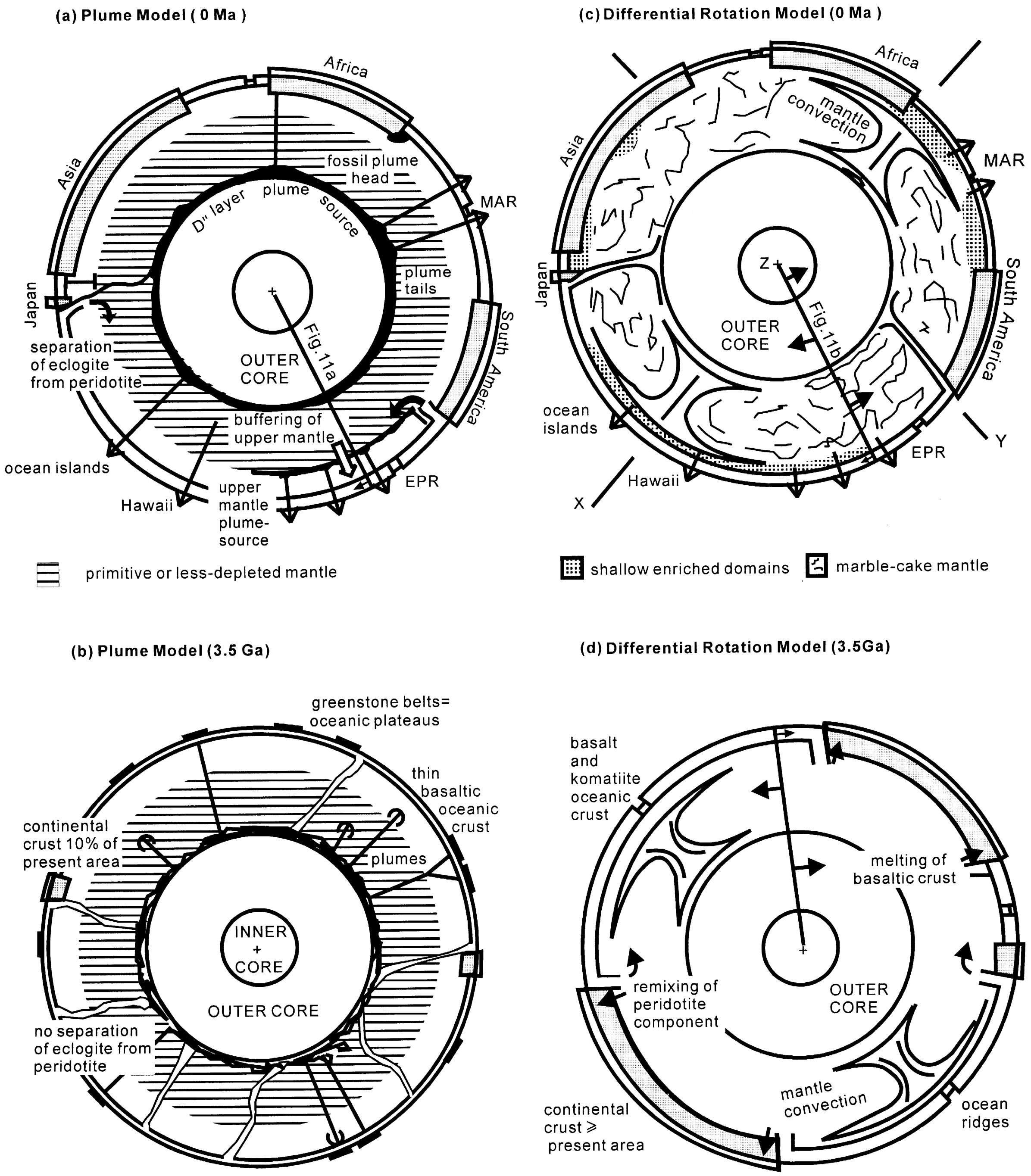


Fig. 1. The Earth as following: (a) The mantle plume model. While the inner core migrates eastwards and convection in the outer core shows a net westward drift relative to the Earth's rotation axis, the mantle behaves as a solid shell rolling about an equatorial axis, and the lithospheric plates move in no particular direction. (b) The differential rotation model. The inner core, outer core, mantle and lithosphere all revolve perpendicular to the Earth's rotation axis. Transmission of stresses through the low viscosity asthenosphere in conjunction with the action of plate boundary forces causes a net westward lag of lithospheric plates relative to the mantle. This lag constitutes a differential rotation of lithosphere and mantle (Δu) which can alternatively be seen as eastward mantle flow. The magnitude of the velocity differential ranges up to 5 cm yr^{-1}, and decreases toward the poles. Undulations in the mantle flow are caused by fluctuations in the position of the rotation axis (polar wander) in response to plate movements changing the mass distribution at the Earth's surface (after Doglioni, 1990; Smith and Lewis, 1999b).

fixed, the composition of plumes, the amount of melting in plumes, and the relationship of plumes to mantle convection (Smith and Lewis, 1999a).

How then, did the plume model come to dominate geodynamics? Maintenance of the status quo is often the hallmark of scientific endeavour, and the more effort that goes into expounding an idea, the more the belief increases that new observations will only refine details to the model, which belies other reasons as to why concepts have changed so little. While the ideas of Wegener and Jeffreys stirred a debate, the ideas of Wilson (1963), Morgan (1971; 1972), Chase (1981), Hofmann and White (1982) and others have led to a virtual monopoly on geodynamic thinking which has perpetuated a myth that the plume model is the only viable option. Under such circumstances, opposition to an established belief is often eliminated (Armstrong, 1991; Keith, 1993; Sheth, 1999a) while observations are made to fit the model and fundamental flaws are treated only as aberrations to be explained by future variations. For example, more than a decade after Okal and Batiza (1987) remarked that the evolution of scientific thinking might have been different had the Cook–Austral Islands been considered before Hawaii, Sleep (1997) and McNutt et al. (1997) are still at the stage of discussing whether or not these islands represent an exception to ''plume theory''. The plume model was of course a new concept once itself, but its introduction occurred in the wake of an upheaval in geological thinking caused by the introduction of the inherently more testable plate tectonic model. While the plume model did receive some initial criticism (e.g., Runcorn, 1974), the number of papers which explored it as an option stands in stark contrast to the lack of alternatives which subsequently accompanied plume models. Alternatives if considered, are now relegated to a subordinate role where they do not conflict with ''plume theory''. More direct challenges are liable to be dismissed as



''throwing the baby out with the bathwater'' (Sheridan, 1994 vs. Anderson, 1994a), even though the plume model is now more than a quarter century old and it would not be unreasonable to expect a fundamentally sound concept to be moving towards a synthesis.

In seeking a unified model for intraplate volcanism, the questions that therefore need to be addressed concern not only how or if the plume model could be developed into a comprehensive theory, but also whether ''proofs'' constructed only with the confines of thinking used to support the ruling theory can be considered valid tests of the model. The justification for maintaining the plume model by adding further variations every time it fails to account for a particular example of intraplate volcanism should also be examined. If the plume model cannot account for all examples of intraplate volcanism or yield a coherent picture of the Earth which is not beset by paradoxes and contradictions, then the time to seek a different picture of the Earth outside of ''plume theory'' is long overdue. To begin to address these issues, in Section 2 we present a brief synthesis of the evolution of some of the critical concepts of ''plume theory'' to outline the principal problems associated with the assumptions made in the model. Considering the number of variations in the plume model, it would not be feasible to do justice to all, but the aim is to demonstrate that sufficient uncertainty has always existed to mandate consideration of other concepts. We then examine in Section 3 whether the alternatives which could have been evaluated but were not, or were evaluated only within the auspices of the plume model, can yield a coherent model for the structure and evolution of the Earth on their own merit. That such an alternative synthesis can be made suggests that there is a largely unexplored Earth beyond ''plume theory'' and that is has not been a lack of scientific information which has held back debate over the workings of the Earth as was the case with the evolution of thought on continental drift earlier this century.

2. Foundations of a ruling theory

2.1. Requirement of a thermal anomaly

Several terms, ''intraplate'', ''hotspot'', and ''plume'' have all been used, often interchangeably, in referring to the subject of basaltic volcanism in continental, ocean island and oceanic plateau settings. None aptly summarises the type of volcanism concerned. Strictly speaking, ''intraplate'' should be used only for those examples of volcanism located in within-plate settings, and its use is therefore problematic when applied to oceanic plateaus likely to have formed in conjunction with ridge processes. Nonetheless, use of ''intraplate'' irrespective of tectonic setting is preferred here over the thermal connotations of ''hotspot''. The latter term was used by Carey (1958) to describe long-lasting centres of volcanism characterised by high heat flow, but became synonymous with the plume model after the statement of Wilson (1963) that the sources of ocean island basalts (OIB) might be localised thermal anomalies. The term ''plume'' arises from the elaboration of the concept of Wilson (1963) by Morgan (1971; 1972) who suggested that hotspots were manifestations of plumes emanating from lower mantle convection.

Fig. 2. Opposing models of the Earth at present (0 Ma) and in the Early Archean (3.5 Ga) resulting from (a,b) following the mantle plume model (c,d) considering non-plume concepts. Each diagram depicts an equatorial section of Fig. 1 looking down along the rotation axis (+). The mantle plume model envisages a near-static mantle penetrated radially by plume conduits which introduce source material for intraplate volcanism from a range of depths in the mantle. In the present-day scenario, eclogite must be largely stripped from the peridotite component of subducted lithosphere, whereas in the Archean both eclogite and large volumes of peridotite must be incorporated into the plume source. The principal expression of plume activity also changes through time from the production of oceanic plateaus (now greenstone belts) to ocean islands. In the differential rotation model, subducted oceanic crust is remixed into the convecting mantle. Large-scale mantle convection (cells depicted following Bercovici et al., 1989) and plate configurations are arranged about the axes of a triaxial Earth. Erosion of continental mantle and its eastward displacement as a result of westward plate lag, superimposes volatile-rich enriched domains on the marble-cake asthenosphere. In the Early Archean, high geothermal gradients in subduction zones stabilise a large early crust by slab melting, while insulation from the crust in conjunction with less efficient cooling of the mantle by slabs leads to an elevated mantle thermal regime producing komatiites along ocean ridges.

Introduction of anomalously hot material to asthenospheric depths by plumes would be expected to produce an uplift of the lithosphere which should be easily testable. Topographic swells such as those around the Hawaiian islands hence became evidence of the existence of plumes (e.g., Crough, 1978). Unfortunately for the hotspot model, measurements along the axis of the Hawaiian swell suggested an increase in heat flow with distance away from the supposed site of the plume (von Herzen et al., 1982). A subsequent study by von Herzen et al. (1989) across the strike of the swell found no evidence of any excess thermal anomaly, and suggested the swell may not be dynamically supported, but merely represents a thick section of basalt. A similar interpretation, though of thickened peridotite predating Hawaiian volcanism, was made by Woods et al. (1991). Either of the latter two explanations would be consistent with the observations by Davies (1992) that the size of the swell does not decline along the Hawaiian chain, and there is no corresponding swell associated with the Emperor chain. While Hawaii is thought to be the strongest currently active plume (Sleep, 1990), its effects are small compared to those expected during the mid Cretaceous superplume event invoked for the formation of many of the intraplate edifices on the Pacific seafloor (Larson, 1991a,b; Vaughan, 1995). Uplift associated with a Cretaceous event has been observed, but fails to reach the amounts predicted in the plume model (Hardebeck and Anderson, 1996; Heller et al., 1996). Similarly, the largest continental flood basalt province, the Siberian Traps, would be expected to be associated with the greatest lithospheric uplift in continental regions. But a detailed analysis of the geological record (Czamanske et al., 1998) has shown that emplacement of the basalts was not accompanied by any more uplift than could be accounted for by local asthenospheric convection. If predictions of the plume model with regard to supposedly the strongest plume, the most recent superplume event, and the largest continental igneous province all fail, little confidence can be put in explanations of the model for less pronounced intraplate features.

Even though a thermal anomaly has never been proven, the temperature structure of plumes has been modelled in detail (e.g., Sleep, 1990; McKenzie and O'Nions, 1991; Schilling, 1991; Saunders et al., 1992; White and McKenzie, 1995). In all cases the temperature estimates are based on the circular argument that plumes exist and are capable of explaining certain geological phenomenon. Hence by assuming that plumes are responsible for continental rifting, a potential mantle temperature of +200 to +300°C greater than the surrounding asthenosphere can be estimated for the material in the core of a plume. Several problems arise, such as the temperatures for modern plumes being equivalent to geotherms depicted for komatiite genesis in the Archean (Fig. 3) and the petrological quirk whereby moving to greater melting temperatures and pressures relative to those required for the generation of MORB, produces melt types as different as undersaturated alkaline basalts and highly magnesian komatiites. The only Phanerozoic komatiites are the Cretaceous examples on Gorgona Island (Echeverria, 1980), so to account for the virtual absence of komatiite compared to the number of supposed hotspots, the zone of melting in a plume has to be confined by rapid movement of the plate (White and McKenzie, 1995). Otherwise using a candle-stick analogy, the plume would ''burn'' through the plate and generate large melt fractions. While the confinement mechanism may work for fast moving plates such as the Pacific plate, it does not explain the melt products of plumes impacting on slow moving plates, as for example, the Cape Verde islands where the thermal structure (von Herzen et al., 1989) and its cause is supposed to be the same as for the Hawaiian islands. At the other extreme of the plume thermal spectrum are eclogite-rich plumes with temperatures only 100°C higher than the surrounding mantle (Takahashi et al., 1996; Cordery et al., 1997). But as temperature decreases, viscosity increases, slowing the ascent rate of the plume, such that it unclear how such plumes could avoid entrainment by mantle flow and ever reach the shallow mantle.

2.2. Hotspot reference frame

The potential of a series of fixed, deep-seated plumes as an absolute reference frame for plate motions was realised by Morgan (1971) and subsequently embraced in many tectonic models. But whether the sources of ocean islands are fixed or not has been a chronic problem throughout the history of

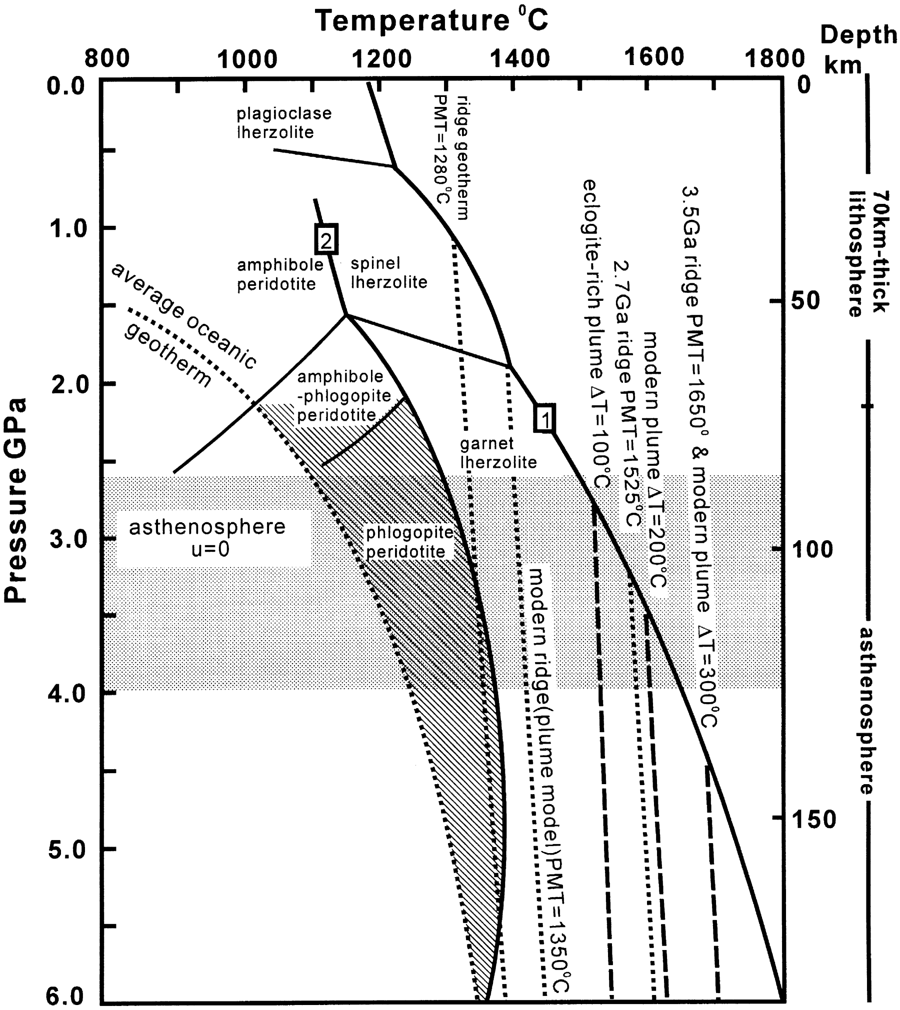


Fig. 3. Mantle thermal regimes in P–T space for melting of anhydrous- and volatile-rich peridotite. The anhydrous peridotite solidus (1) and plume geotherms are from Herzberg (1995). The solidus for amphibole and phlogopite peridotite (2) for ''wetspot'' melting has been constructed from Menner and Dunn (1995) and Sato et al. (1997). Average oceanic and ridge (PMT = 1280°C) geotherms are from Turcotte and Schubert (1982). If the thermal profile away from ridges is approximated by the average oceanic geotherm, hydrous peridotite may exist (hatched pattern) in the asthenosphere to depths of around 170 km. In plume models (e.g., Herzberg, 1995), the ridge geotherm must be hotter (PMT = 1350°C) to initiate melting within the garnet lherzolite facies to satisfy interpretations of Hf isotopic data. The shaded region shows the potential depths of the asthenospheric layer where the flow velocity (u) is zero in the counterflow regime in Fig. 4e; the height of this layer varies depending on latitude, lithospheric thickness, and plate velocity.

the plume model. There are essentially two aspects to the problem, motion between groups of hotspots and motions of individual hotspots within a particular group. Individual hotspots have been generally considered fixed to within 1 cm yr^{-1} (e.g., Duncan and Richards, 1991), though other models have argued for movement up to several centimetres per year, comparable to rates for lithospheric plates (Burke et al., 1973; LePichon et al., 1973; Molnar and Stock, 1987). Not even the type example of a plume, Hawaii, may have followed Morgan's vision of hotspots. The fixed nature of this hotspot has been challenged both from a lack of evidence for a change in Pacific plate motion direction at 43 Ma and from paleomagnetic evidence, such that it is now thought that the plume was not fixed and may even have moved at an oblique angle to the plate during formation of the Emperor chain (Norton, 1995; Tarduno and Cottrell, 1997). The problem of such mobility is rooted in the relationship of individual plumes to mantle convection, which is somewhat ironic in consideration of Morgan's original suggestion of plumes being the manifestation of convection. Ascent velocities estimated for plumes have ranged from 200 cm yr^{-1} (Morgan, 1971), through 60 cm yr^{-1} (Loper, 1991), to 10 cm yr^{-1} or less (Christiansen, 1984; Richards et al., 1989; Duncan and Richards, 1991). The lower estimates can be regarded as the more reasonable since to avoid plumes ''burning'' through the plate and generating large melt fractions, the ascent velocities of the plume material must be of the same magnitude or less than plate velocities. A delicate balance then has to be achieved between plume, plate, and mantle velocities, because if the latter approach or exceed plume ascent rates, the result would be significant motion between hotspots or the mixing of plumes into the convecting mantle. Accommodating the plume model therefore constrains mantle convection to be a sluggish effect with velocities around 1 to 5 cm yr^{-1}, while allowing the existence of ''fossil'' plume conduits (Olson and Singer, 1985; Loper, 1991) which can be re-utilised by later plume eruptions would essentially require a motionless mantle.

Motions between groups of hotspots en masse, are indicated from differences between the polar wander curve relative to the hotspot frame and to the geomagnetic axis, and have been explained by true polar wander whereby the mantle containing the hotspot frame moves relative to the rotation axis (Hargraves and Duncan, 1973; McElhinny, 1973; Harrison and Lindh, 1982; Andrews, 1985; Duncan and Richards, 1991). The rate of motion varies over time, corresponding to 12° for the period 40 to 0 Ma, and 20° for the period 180 to 40 Ma, with Atlantic hotspots appearing to have moved northwards and Pacific hotspots southwards about an equatorial pole at 60°E (Courtillot and Besse, 1987). Such behaviour of the mantle rolling like a ball within the lithospheric shell (Fig. 1a) was described as ''curious'' by Hargraves and Duncan (1973). Again, investigations focused on mechanisms for explaining such behaviour rather than on the possibility, noted by McElhinny (1973), that the unlikely nature of the effect meant that the plume model was incorrect. Alterations to the Earth's moment of inertia by convection and subduction have been proposed for why the mantle should roll within the Earth. The mantle roll rates correspond to 3.3 cm yr^{-1} for the Eocene to Recent and should also result in displacements in the position of arcs from the pinning of subducted slabs in the mantle. Furthermore, as the roll would be perpendicular to convergence at many subduction zones, motion should also be imparted to lithospheric plates. No such systematic changes have been documented. Rather, long-term subduction patterns have been used in support of correlations between hotspots and the geoid as outlined in Section 2.4. Indeed, the control on geodynamic thinking by the plume model has been such to initiate a search for undocumented boundaries within the Antarctic plate (e.g., Duncan and Richards, 1991; Cande et al., 1995) in order to make plate motion circuits more consistent with the hotspot frame rather than consider the possibility that the true polar wander effect implied a serious flaw in the plume model.

2.3. Controls on plate tectonics

The controls on plate tectonics which result from having a series of radial conduits penetrating a near-static mantle are quite insidious. A key assumption has always been that use of different reference frames will not affect relative plate motions. It should also have been asked whether use of different reference frames would affect the relative importance of plate driving forces. When plate driving mechanisms were first considered it became apparent that there are two categories, drag driven by mantle flow acting on the base of the plate, and boundary forces acting along the edges of plates (e.g., Forsyth and Uyeda, 1975; Chapple and Tullis, 1977; Turcotte and Schubert, 1982; Bott, 1984). To move plates by drag alone would require mantle flow rates of 10 to 20 cm yr^{-1} (Bott, 1984), so if the mantle is required to be nearly static, drag cannot be an active driving force. Plates must then move themselves by boundary forces, so that it is not surprising that basal drag becomes a minor and resisting force. Rather, in conventional plate tectonic models, it is the motion of the plates which induces flow in the asthenosphere (Fig. 4a). Counterflow (Fig. 4d) can only arise from a combination of Couette flow induced by drag from the lithosphere and pressure–gradient flow, and so requires either return flow of mantle to the ridge, or asthenosphere displaced by a subducting slab. Imposing a fixed lower boundary (mesosphere) also causes steep velocity gradients and hence high shear stresses through the counterflow profile. Even a moderate plate velocity of 5 cm yr^{-1} would generate a shear stress of 2.2 MPa on the base of the lithosphere

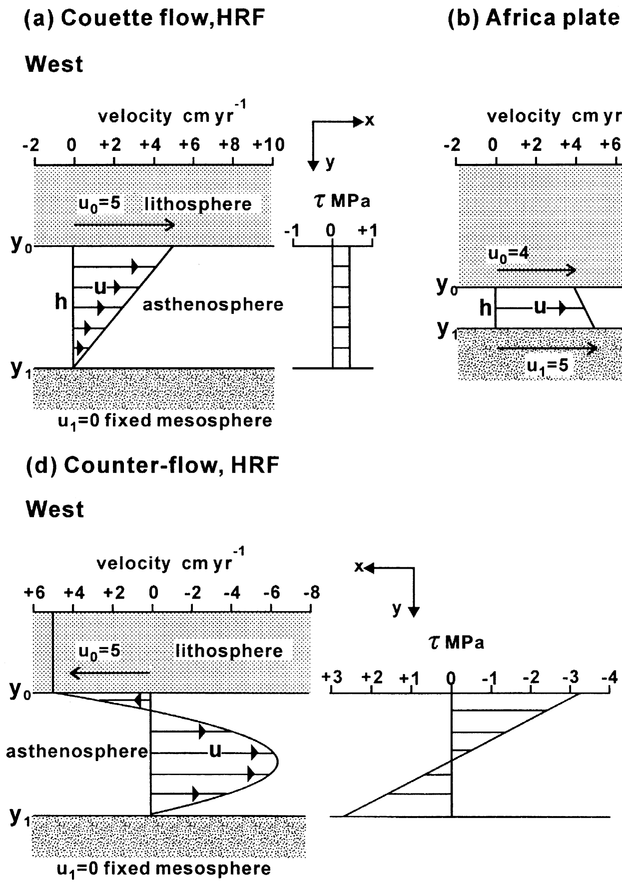


Fig. 4. Possible flow velocity (u) regimes and shear stress (τ) profiles through the asthenosphere in conventional plate tectonic models using the hotspot reference frame (HRF) and in the differential rotation model where the rotation axis is used as a reference (RAR: rotation axis reference). Flow equations are given in Smith and Lewis (1999b). For simplicity, the asthenosphere is depicted (following Turcotte and Schubert, 1982; Montagner and Tanimoto, 1991) as a channel where decoupling takes place between the base of the lithosphere (y_0) and the top of the mesosphere (y_1). (a) Couette flow, hotspot reference frame with fixed mesosphere. Movement of plates by boundary forces exerts a drag on the top of the asthenosphere. (b) Couette flow, differential rotation model. Eastward movement of the mesosphere imparts a drag on the base of the asthenosphere as the Earth rotates. The flow induced in the asthenosphere causes a drag on the base of the lithosphere. (c) Plate boundary forces supplementing drag induced from the mesosphere, as in the example of the Nazca plate. When boundary forces predominate, the result is a Couette flow profile with diminished velocity gradient. (d) Counter flow, hotspot reference frame. Movement of plates by boundary forces is balanced by asthenospheric flow in the opposite direction. The latter may be equated with either return flow to a ridge or asthenosphere displaced by a subducting slab. (e) Counterflow, differential rotation model. The flow regime is the product of plate boundary forces and drag induced from the mesosphere acting in opposing directions. Note the development of a stationary ($u = 0$) layer at shallow asthenospheric depths.

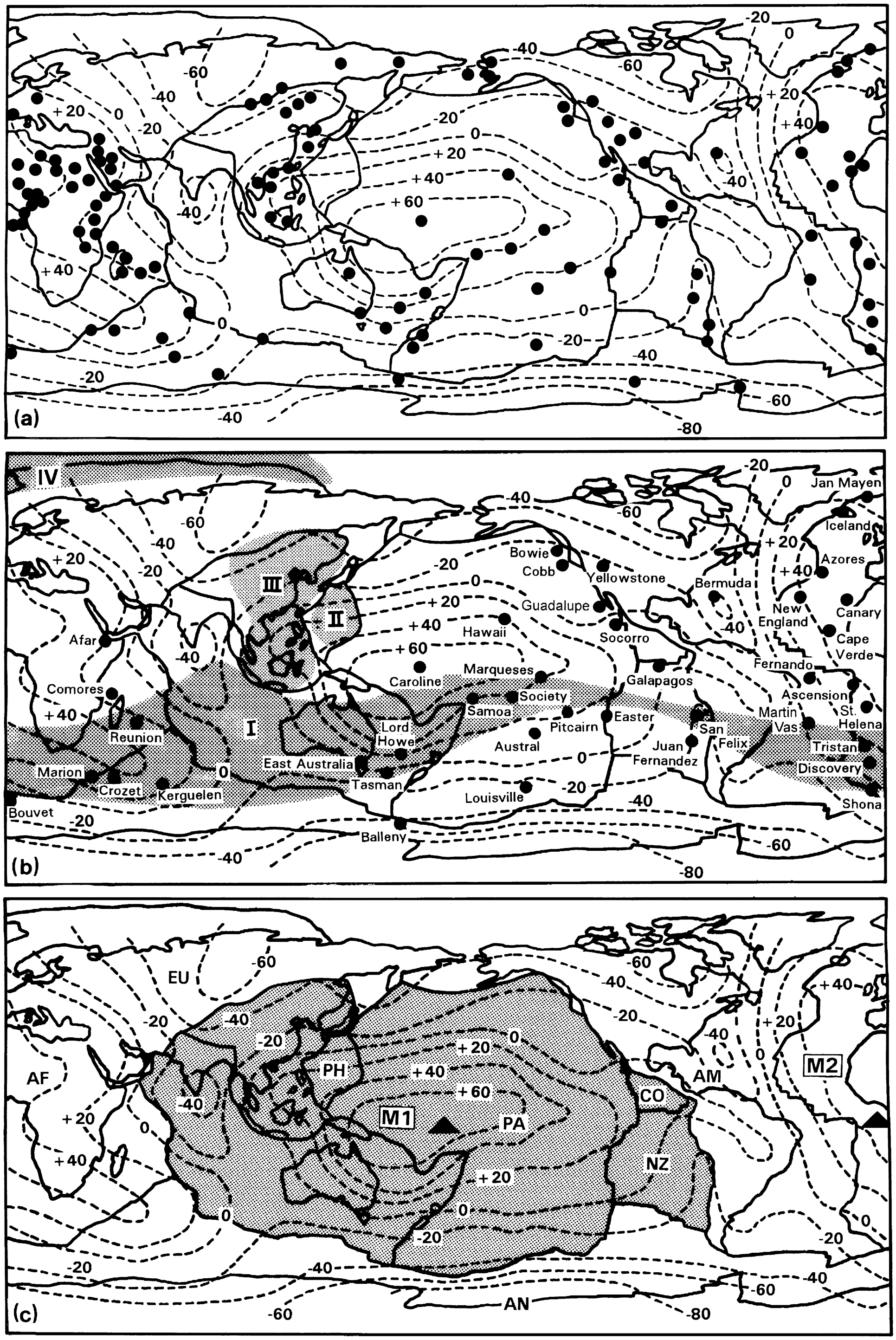
(Turcotte and Schubert, 1982) which is unrealistic, as for a Pacific-size plate, drag would then balance slab-pull causing the plate to be near stationary.

The ensuing problem is that plate movement by boundary forces alone cannot explain the plate tectonic record. The problem is aptly illustrated by considering continental rifting where the principal forces acting in the oceanic regime, slab pull and ridge push (Forsyth and Uyeda, 1975; Chapple and Tullis, 1977) would be ineffective. Either basal drag must be important for large continental masses (e.g., Ziegler, 1992; Stoddard and Abbott, 1996), in which case plate tectonics becomes a complex phenomenon with drag and boundary forces varying in importance at different stages in geological history, or an explanation has to be sought outside of plate tectonics. Conveniently, rifting can be attributed to deviatoric stresses arising from the interaction of plume heads with the continental lithosphere (Morgan, 1971, 1983; Turcotte and Emerman, 1983; Saunders et al., 1992; Zeyen et al., 1997), though as the distribution of stresses in a plume should be radial, a multitude of plumes, all fortuitously lined up along the axis of rifting, is required. The plumes may potentially be assisted by forces such as trench suction (Bott, 1992), however, it is their arrival which is generally regarded as the impetus for continental breakup and associated volcanism (White and McKenzie, 1989). A necessity for plumes is reinforced by treatments (e.g., Arndt and Christensen, 1992) of continental mantle as a cold, geochemically depleted, and structureless entity. Such concepts stem in part from equating continental mantle with lithosphere, and in part by using xenoliths from cratonic regions as representative of the entire continental mantle. Confidence in a model, however, should be based on its ability to solve a particular problem. The hotspot model appears to offer a solution to the problem of continental rifting, but the model itself is the cause of the problem, hence the solution should not be taken as evidence for the existence of plumes.

2.4. Fitting the geoid: a problem of how many hotspots?

When mantle convection is reduced to a sluggish, subordinate feature, it follows that large-scale geophysical features such as the Earth's residual geoid anomalies should also be interpreted in terms of plume dynamics. Positive residual geoid anomalies have therefore been correlated with hot regions where plumes are supposed to be actively upwelling, while negative geoid anomalies have been related to cold subducted slabs which have not yet been warmed to become plume sources (Jurdy, 1983; Richards et al., 1988; Davies and Richards, 1992; Richards and Engebretson, 1992). However, any correlation between hotspots and geoid is critically dependent on the number of hotspots, which of course involves the problem of how to define a hotspot. Morgan (1971) depicted 16 but suggested at least 20 hotspots. Burke and Wilson (1976) calculated the number to be 117 by placing hotspots under all occurrences of intraplate volcanism (Fig. 5a), an approach which produced a random distribution which bore no relation to the geoid (Turcotte and Schubert, 1982; Stefanick and Jurdy, 1984), although it did at least imply a common origin for intraplate volcanism. To improve the correlation, Crough and Jurdy (1980) chose to define what should quintessentially be a thermal phenomenon by the ability of the boundary forces acting on the overlying plate to move it fast enough to produce a linear age progression. The result, by ignoring intraplate volcanism in large areas of the world such as Asia (Smith, 1998), was a set of 42 hotspots largely distributed within the Earth's major Pacific and Atlantic positive residual geoid anomalies (Fig. 5b). The correlation was further enhanced by Duncan and Richards (1991) by ignoring negative geoid anomalies and smoothing the contours used by Crough and Jurdy (1980) to include examples of intraplate volcanism in areas of marginal geoid anomaly. Thus, geoid contours around Juan Fernan-

Fig. 5. Correlation of residual geoid anomalies (from Crough and Jurdy, 1980) with: (a) Hotspots as placed under all examples of intraplate volcanism (after Stefanick and Jurdy, 1984). (b) Set of 42 hotspots defined by Crough and Jurdy (1980). Shaded regions indicate positions of DUPAL domains (I equatorial belt; II Philippine Sea; III Asia; IV Arctic). (c) Megaplate configuration (triangles indicate centres of configurations M1 and M2; Lewis and Smith, 1998) defined on the basis of lithospheric angular momentum. Plates: AF Africa, AM America, AN Antarctic, CO Cocos, EU Europe, IN India, NZ Nazca, PA Pacific, PH Philippine.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+40
+60
-80
(b)
IV
III
II
I
Jan Mayen
Iceland
Azores
Bowie
Cobb
Yellowstone
Bermuda
Guadalupe
New England
Canary
Cape Verde
Hawaii
Socorro
Afar
Fernando
Caroline
Marqueses
Galapagos
Ascension
Comores
Society
Samoa
Easter
Martin Vas
St. Helena
San Felix
Reunion
Lord Howe
Pitcairn
Austral
Juan Fernandez
Tristan
Discovery
Marion
Crozet
Kerguelen
East Australia
Tasman
Louisville
Shona
Bouvet
Balleny
(c)
EU
AF
PH
M1
PA
CO
AM
M2
NZ
AN

dez/San Felix are -20 m in Crough and Jurdy (1980) but $+20$ m in Duncan and Richards (1991).

Unfortunately, dating has shown many of the age progressions in ocean island chains to be distinctly non-linear (Jackson, 1976; Turner and Jarrard, 1982; Okal and Batiza, 1987; McNutt et al., 1997; Dickinson, 1998). To fit the local expression of volcanism in many regions requires the addition of multitudes of plumes (7 for the Line Islands, 3 for the Cook–Austral Islands, 2 for the Kodiak–Bowie chain and so on) of the type Crough and Jurdy (1980) sought to eliminate, all fortuitously lined up along the axis of volcanism. Other plume models have resorted to a variety of ad hoc plumbing arrangements for melt to reach the surface including swaying or deflection of plumes, channelling of plume material by lithospheric structure, multiple conduits from a single plume head (e.g., Morgan, 1978; McNutt and Fischer, 1987; Granet et al., 1995; Sleep, 1997), or have devised special shapes such as the tabular plumes of Sleep (1992). A further option is plume splitting, which has been invoked for multiple episodes of volcanism on oceanic plateaus (Bercovici and Mahoney, 1994). Multiple episodes of volcanism are also found in continental provinces (e.g., Bailey, 1992), but the events are often separated in time by hundreds of millions of years such that plume splitting would not be a feasible explanation. Plumes would either have to hit the same spot repeatedly under continents, or be channelled (e.g., Thompson and Gibson, 1991) to the appropriate position. Even with such variation, the plume model still does not fit all examples of intraplate volcanism, and what seems to have been greatly overlooked is that the adoption of any limited set of hotspots is an admission that the plume model provides only a partial explanation. In effect, the plume model requires two categories of intraplate volcanism, plume-related and non-plume-related, producing the same petrological products from thermal regimes differing by up to 300°C.

2.5. Geochemical constraints, or lack of?

A significant stage in the development of the plume model came in the late 1970s and early 1980s with refinements to analytical methods in mass spectrometry which allowed evidence on the origin of OIB from Pb and Nd isotopic measurements. Until this stage, plume models had competed with concepts of propagating fractures (Jackson and Wright, 1970; Green, 1971; Anquita and Hernan, 1975), stress fields (Shaw, 1969, 1973; Jackson and Shaw, 1975; Pilger, 1982), and shallow mantle convection (Richter, 1973; Bonatti and Harrison, 1976) as causes of linear intraplate tracks. Both plume and shallow-source models had emphasised differences in melting regimes as a small melt fraction from the MORB-source can show similar trace-element features to OIB (Gast, 1968; Fitton and James, 1986). However, isotopic data (e.g., Sun, 1980) indicated a distinct and isotopically more ancient source for OIB compared to MORB, which favoured the plume model from its emphasis on introducing material from depth in the mantle. No attempt was made to examine whether non-plume concepts could be compatible with distinct sources; the differential rotation mechanism (see Section 3.3) which could have allowed for lateral introduction of source material was dismissed because it did not appear to agree with plate motions in the hotspot frame.

Some plume models (e.g., Wasserburg and DePaolo, 1979) followed the ideas of Morgan (1971) and derived plumes from primitive or relatively undepleted material in the lower mantle on account of a clustering of εNd values around zero in some examples of intraplate volcanism. High $^{3}He/^{4}He$ ratios have also been consistently interpreted to mandate a plume origin from primitive material even though no mantle-derived melt has trace-element characteristics compatible with derivation from such a reservoir (Hofmann et al., 1986). The concept of a large-scale equatorial plume belt (Fig. 5b), characterised by high $^{208}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$, and $^{87}Sr/^{86}Sr$ signatures, termed the ''DUPAL'' anomaly, can also be considered in the primitive or less-depleted category. Plumes of DUPAL composition were envisaged to form either as the result of large-scale equatorial upwelling or as lower mantle was displaced by subducting slabs (Hart, 1984; Castillo, 1989). Subsequent discovery of the same geochemical signature in intraplate volcanism in Asia led to another variation that a DUPAL plume-source could be sediment-contaminated mantle (e.g., Basu et al., 1991; Tatsumoto and Nakamura, 1991; Tu et al., 1991). Duplication of sources with identical isotopic signatures at opposite ends of the mantle was avoided by invoking delamination of sediment-contaminated mantle into the plume-source following the mechanism proposed by McKenzie and O'Nions (1983). However, neither the original model nor the variations provide a satisfactory explanation for the presence of DUPAL signatures in MORB on the Nansen–Gakkel ridge in the Arctic (Mühe et al., 1993), nor is the concept of equatorial upwelling supported by results from seismic tomography.

Other plume models interpreted linear arrays in OIB Pb isotopic compositions as pseudo-isochrons giving the ages of OIB sources (e.g., Chase, 1981). The concept was expanded into the crustal recycling type of plume model by Hofmann and White (1982) who envisaged that plumes were derived from ancient subducted basaltic oceanic crust. From density considerations, and also to avoid the problem of excessive amounts of subducted lithosphere accumulating at depth in the mantle, basalt would have to be largely separated from the peridotite component of the slab on conversion to eclogite, although the difficulty of deriving an OIB from a MORB parent would require the retention of at least 10–20% peridotite in the plume source. The ages of 1.0 to 2.0 Ga suggested by OIB Pb arrays corresponded to the time required for subducted crust to evolve to HIMU isotopic compositions given favourable increases in U/Pb ratio by seafloor alteration and subduction zone processes, and hence were interpreted as the time required for plume-source material to heat-up and become buoyant. However, whether the corresponding change in Rb/Sr ratios would be suitable to produce the low $^{87}Sr/^{86}Sr$ ratios which also characterise HIMU is uncertain (Kogiso et al., 1997). The number of OIB, except for basalts on the islands of St. Helena and Mangaia (Sun, 1980; Palacz and Saunders, 1986; Zindler and Hart, 1986), displaying pure HIMU signatures is also small (Fig. 6). Hofmann et al. (1986) precluded significant amounts of sediment in the plume source from Nb/U, Ce/Pb ratios. Later variations of the plume model (e.g., Weaver, 1986) added subducted sediment to account for the EM isotope signatures which predominate in most examples of OIB, with blatant disregard for Hf isotopic evidence (Fig. 7a) which had placed strict limitations on any role for such material by showing that only a specific sediment composition would be suitable as an endmember for generating the OIB array. No modern arc subducts the appropriate sediment mixture, such that mixing has to take place within the plume-source layer (Patchett et al., 1984). But mixing would require convection, an inherently difficult prospect in a thin ($<$ 200 km-thick) thermal boundary layer, which if equated with the D″ layer, also has to separate outer core and mantle differing in temperature by at least 1000°C (Williams, 1998), for up to 2 Gyr.

With the inclusion of the Kodiak–Bowie and Cobb–Eichelberg chains, where basalts are characterised by DM isotopic signatures (Hegner and Tatsumoto, 1989), in the commonly used hotspot sets (e.g., Crough and Jurdy, 1980; Sleep, 1990), every major radiogenic isotope endmember (BE, DM, HIMU, EM1, EM2; Zindler and Hart, 1986) has now been equated with a plume. Such diversity has been embraced (e.g., Kerr et al., 1995) rather than questioned. But even with the flexibility of any lithosphere or mantle component being allowed to characterise a plume-source, the model still fails to account for evidence (e.g., Clague and Frey, 1982; Bonatti, 1990; Hoernle and Schmincke, 1993; Francis, 1995; Francis and Ludden, 1995; Hasse et al., 1996) for hydrous minerals in the source of most OIB. Volatiles in plumes have to come from unspecified sources in the deeper mantle, or be ascribed to contamination of the plume with lithosphere at shallow level (e.g., Wyllie, 1988; Schiano et al., 1994; Class and Goldstein, 1997). Such explanations might succeed in deferring the problem, except that the volatile-bearing minerals are often required to remain as residual phases, which would be impossible as the thermal regime of even the coldest eclogite-rich plume would be at least 150°C above the hydrous peridotite solidus (Fig. 3).

2.6. Volumes, melting, and fluxes

The issue of whether intraplate volcanism could be derived from recycled crust or primitive/less depleted mantle plumes could potentially be resolved from the production rate of ocean crust relative to the volume of intraplate volcanism, if the composition and amount of partial melting in plumes was known. The volume of oceanic plateaus and islands produced over the past 150 Myr since the mid

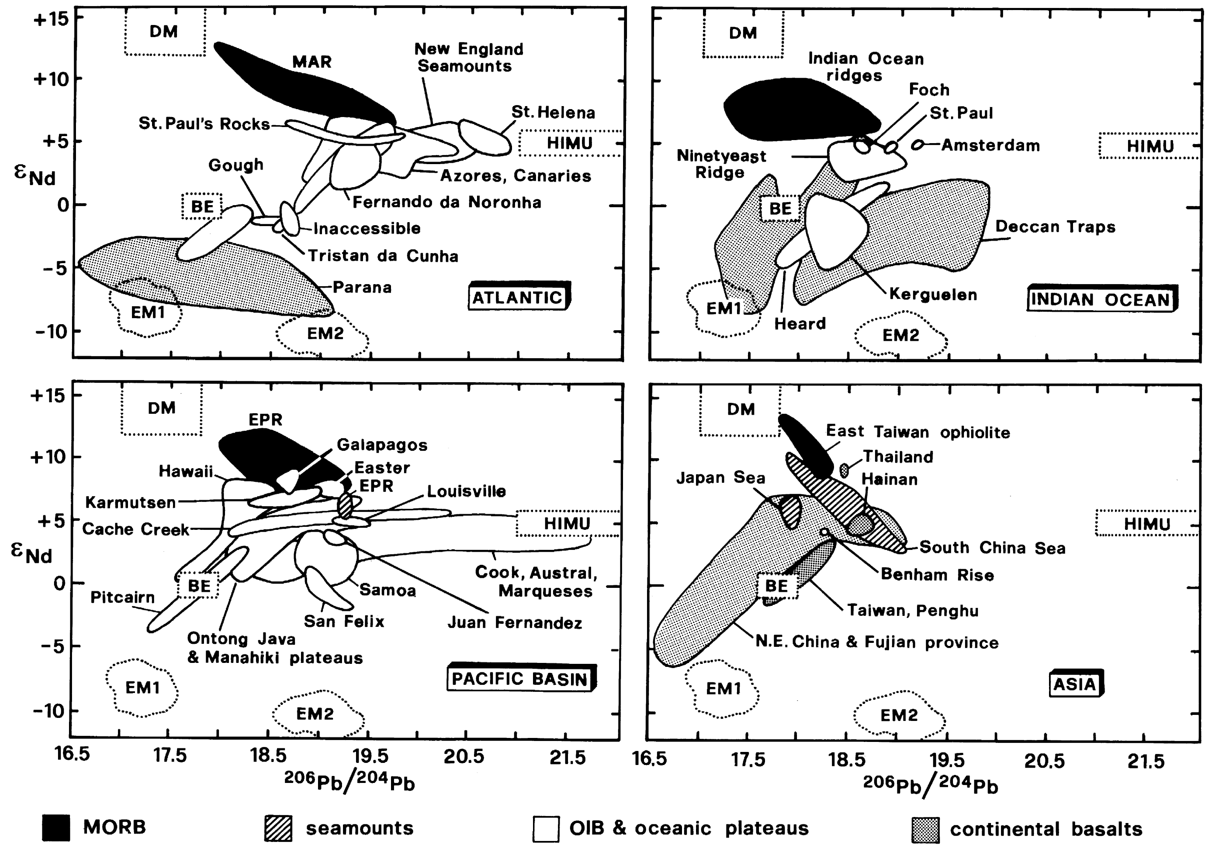


Fig. 6. Nd–Pb isotopic signatures of MORB, seamounts, and intraplate basalts in continental and oceanic settings in Asia and the Atlantic, Indian, and Pacific Ocean basins. The depleted mantle (DM), bulk Earth (BE), enriched mantle (EM1 and EM2), and high mu (HIMU) isotope endmembers of Zindler and Hart (1986) are shown for reference (after data compilations in Zindler and Hart, 1986; Mahoney and Spencer, 1991; Weis et al., 1992; Smith, 1998; Smith and Lambert, 1995).

Cretaceous is around 390×10^6 km^3 (Schubert and Sandwell, 1989; Larson, 1991a; Vaughan, 1995), which corresponds to only 11% of the volume of MORB produced over the same time period. Over 95% of oceanic intraplate volcanism is accounted for by plateaus, rises and aseismic ridges. A conservative estimate would be that at least 60% of the oceanic plateaus formed in conjunction with ridge activity. Taking a minimum estimate of 50% for the volume of melt generated from depleted mantle sources for such plateaus (Anderson, 1982a) gives a revised volume for oceanic intraplate volcanism corresponding to 8% of MORB production. Plume models have also emphasised that, at least for a short time, the volumes of basalt produced in continental provinces exceeds that along ridges. However, even taking the higher range of estimates for the volumes of continental basalt provinces (e.g., Kent, 1995) the total volume for the Parana, North Atlantic, Deccan, and Columbia River provinces over the last 150 million years is only 15×10^6 km^3. The formation of continental basalt provinces tends to be episodic from the association with continental break-up; however, even to include the Karoo, Ferrar, and Siberian Traps provinces back to the Permo-Triassic would only add an additional 9×10^6 km^3 to the total. The time-averaged basalt production rate for continental provinces is therefore less than 1% and likely around 0.5% of that for ocean ridges. However, even though large volumes and rapid eruption rates of some intraplate volcanism has been taken as unassailable evidence for a plume origin, the amount of melting

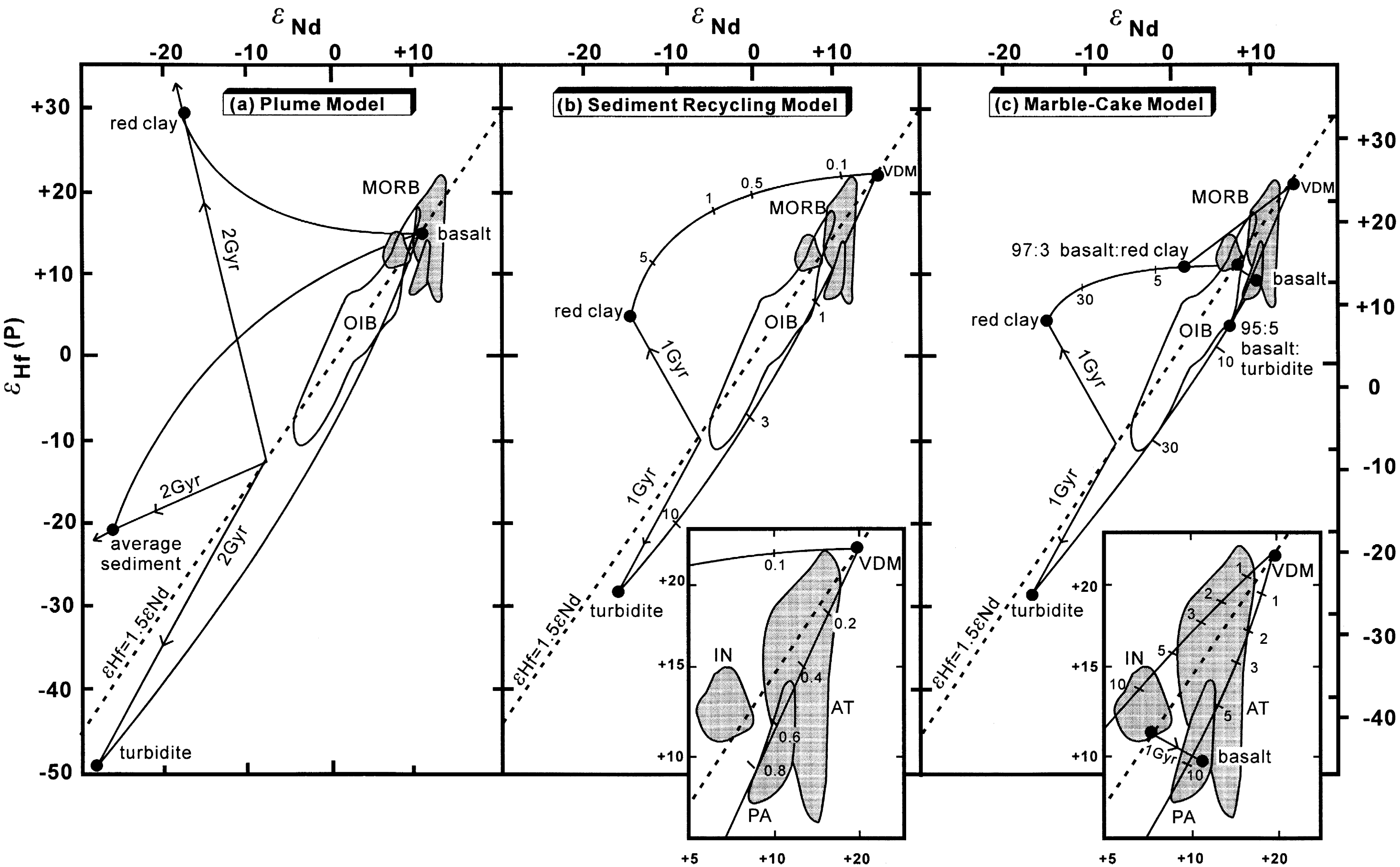


Fig. 7. The fate of subducted crust in plume and marble-cake mantle models as indicated from the Nd–Hf isotope variations in mantle derived rocks (MORB and OIB) (a) Plume model (Patchett et al., 1984). A specific ''average'' sediment mixture is required to generate the co-variation in OIB isotope ratios, while isotope ratios in MORB appear to be decoupled. (b) Sediment-only recycling model. Numbers indicate the percentages of sediment required as a buffer on the MORB-source. (c) Marble-cake model with recycling of basalt-sediment mixtures. Note: sediment-recycling and marble-cake models concern mixing crust with a very depleted mantle (VDM) reservoir to generate the depleted mantle (DM) MORB-source. Values for Nd and Hf concentration (ppm) and $^{176}Lu/^{177}Hf$, $^{147}Sm/^{144}Nd$ ratios are the same for endmembers in all three models and are based on estimates in Patchett et al. (1984) and Sun and McDonough (1989): red clay 3.80, 95.0, 0.0580, 0.1387; average turbidite 5.02, 15.1, 0.0075, 0.1200; VDM 0.114, 0.050, 0.0507, 0.2480; N-MORB 2.05, 7.30, 0.0315, 0.2200, respectively. The initial sediment composition corresponds to a crustal source with a Middle Proterozoic depleted mantle model age. Whereas the plume model extrapolates compositions into the future, the mixing calculations in the sediment recycling and marble-cake models are for the time period 1 Ga to the present (arrows indicate isotopic evolution of components). Fields for OIB and MORB (AT Atlantic; IN Indian; PA Pacific Oceans) are from Nowell et al. (1998) and Salters and White (1998). Hafnium isotopic compositions relative to the CHUR values used by Patchett et al. (1984) are shown on the left hand axis as εHf(P), whereas values relative to the CHUR estimate of Blichert-Toft and Albarède (1997) are shown on the right as εHf(BT).

within a plume is poorly constrained, with estimates generally ranging between 10 and 40% (e.g., Hofmann and White, 1982; Cordery et al., 1997). If the relative volumes of intraplate and ridge volcanism differ by an order of magnitude, the estimate of 10% melting in a plume would equate with a steady state flux between the volume of basalt subducted and the amount of intraplate volcanism produced at the surface.

Unfortunately, large amounts of melting in plumes agrees poorly with the trace element evidence used to justify the recycled plume model. Similar Nb/U and Ce/Pb trace element ratios between MORB and OIB have often been cited in support of the recycled crust type of plume model (Hofmann et al., 1986) as the bulk distribution coefficients for these elements in an anhydrous mineralogy would result in little fractionation of the ratios between melt and source. However, for subducted crust to reach HIMU isotopic compositions by increasing the U/Pb ratios by seafloor alteration alone would require a 300% increase in U content (e.g., McCulloch and Gamble, 1991). As it is unlikely that an entire 7-km-thick section of oceanic crust would be affected by such a process, Pb must also be lost to slab-derived fluids in the subduction zone (Brenan et al., 1995a,b; Chauvel et al., 1995). But estimates of Pb losses to slab fluids (Brenan et al., 1995a,b; Kogiso et al., 1997) would cause Ce/Pb ratios to increase to around 70 in subducted oceanic crust. The Ce/Pb ratios of less than 35 found in OIB could then only be attained with very low ($\sim 1\%$) degrees of melting, which would be difficult to reconcile with the presence of picritic basalts in some intraplate volcanic sequences which indicates large degrees of melting. Nor would low degrees of melting explain why material more geochemically fertile than the MORB-source should undergo an order of magnitude less melting when brought to the surface at temperatures hundreds of degrees Celsius hotter. Further contradictions are illustrated by the results of Cordery et al. (1997) who showed that generation of the melt volumes in continental flood basalt provinces would take tens of millions of years, much longer than the timescale of eruption, thereby defeating the basis for invoking plumes to explain large igneous provinces in the first place. Moreover, it was also shown by Cordery et al. (1997) that the amounts of melt can only be generated if lithospheric extension was already underway, which contradicts the whole category of models where plumes are a requisite for continental rifting.

2.7. MORB paradoxes

Despite the contradictions and chronic uncertainties with regard to plume composition and melting, the plume model has nonetheless managed to exert severe controls on other aspects of mantle geochemistry. The depleted mantle MORB-source is conventionally considered the geochemical complement of the continental crust, but its evolution, and that of the continental crust, becomes closely tied to plume sources in the hotspot model. Compounding the situation are paradoxes concerning the co-variation and evolution rate of Nd–Hf isotopic ratios within the MORB-source. Relative to other planetary bodies (Moon and Mars) the Earth's depleted mantle appears to have followed a very moderate rate of Nd isotopic evolution (Fig. 8) which does not appear to be supported by parent/daughter ratios calculated from standard melting models (DePaolo, 1988; Smith and Ludden, 1989; Salters and Hart, 1991). Furthermore, while Nd–Hf isotope ratios appear to correlated in OIB and crustal rocks, MORB show 45% of the Hf variation but only 22% of the Nd isotopic variation in mantle-derived rocks (Patchett and Tatsumoto, 1980; Salters and Hart, 1991; Nowell et al., 1998). The paradoxes may be partially resolved within the confines of the plume model by appealing to a very specific melt extraction regime (Salters and Hart, 1989). Although the model implies a relatively young MORB-source and the fit to the trace element and isotopic features is imperfect, it has nonetheless been construed as justification for re-adjusting the MORB-source potential mantle temperature of 1280°C calculated by McKenzie and Bickle (1988) to a higher temperature of 1350°C so that melting is initiated with the garnet lherzolite facies (Fig. 3).

Another option to explain the MORB paradox is to invoke buffering of the mantle with an external contaminant. As the distinctive feature of the Earth relative to Mars and the Moon is its active plate tectonic regime characterised by subduction of oceanic plates, the most straightforward explanation would be buffering the MORB-source with subducted crust. Banding in oceanic peridotites (Hame-

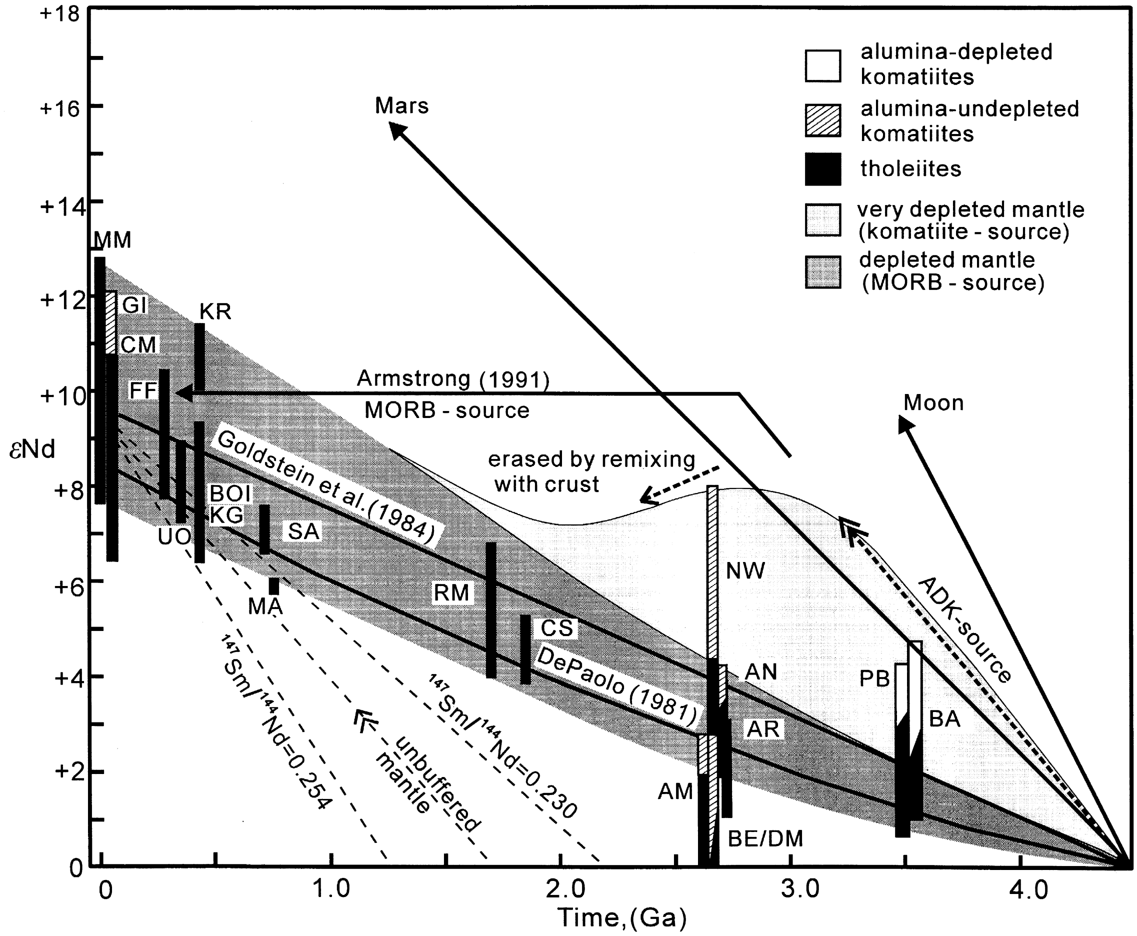


Fig. 8. Models for the Nd isotopic evolution of the mantle. Many studies use the gradual evolution curves proposed by DePaolo (1981) and Goldstein et al. (1984). Smith and Ludden (1989) argued that such curves were in error and proposed a much greater amount of isotopic variation in the Precambrian mantle. Mantle evolution rates through the Phanerozoic and Proterozoic can be modelled as the result of recycling subducted crust into the depleted mantle reservoir, whereas the rate for the alumina-depleted komatiite (ADK)-source parallels the average expected evolution curve for unbuffered mantle (note parallel arrows) suggesting a scenario of no crustal recycling in the Early Archean. The latter would allow the existence of highly depleted domains which become remixed back into the depleted mantle MORB-source during the Late Archean and Early Proterozoic. Index (in decreasing age): MM modern MORB, GI Gorgona Island, CM Cretaceous MORB, FF Fennell Formation (Slide Mountain terrane), UO Urals ophiolite, KR Kings River ophiolite, BOI Bay of Islands ophiolite, KG Køli Group, SA Saudi Arabian ophiolites, MA Matchless amphibolite, RM Rocky Mountain greenstones, CS Cape Smith, NW Norseman–Wiluna, AN Abitibi Newton township, AR Abitibi Rainy Lake, AM Abitibi Munro township, BE/DM Belingwe/Diemals-Marda, PB Pilbara, BA Barberton, data sources are given in Smith and Lambert (1995) and Smith and Ludden (1989).

lin and Allègre, 1988) has been interpreted as direct evidence for such a process, otherwise known as the marble-cake mantle model (Allègre et al., 1980; Allègre and Turcotte, 1986). The marble-cake model provides a ready explanation for the heterogeneity in the MORB source which is thought to comprise streaks (Fitton and James, 1986) or plums (Zindler et al., 1984) of enriched material embedded within depleted material residual from melt generation. However, the requirement to isolate subducted crust into a plume-source has switched the emphasis from marble-cake remixing to buffering the MORB-source with plume material (Stein and Hofmann, 1994; Kerr et al., 1995; Phipps Morgan et al., 1995; Kellogg and

King, 1997). Yet if plume material was the buffer, Nd and Hf isotopic ratios should be correlated in MORB as they are in OIB. The contaminant buffering the MORB source cannot therefore be the same age and composition as that producing the range of isotopic variation in OIB.

A further variation permitted by the plume model is buffering of the MORB-source with lower mantle material (Fig. 2a) (Sleep, 1984; Galer and O'Nions, 1985; Sun and McDonough, 1989). But neither mixing with primitive mantle (BE) nor a FOZO lower mantle composition between DM and HIMU as postulated by Hart et al. (1992) produces any better fit to the Nd–Hf isotopic data. Preferential incorporation of Hf relative to Lu into high pressure phases to give a lower mantle with a predominantly low εHf signature (Kato et al., 1988) might account for the low εHf–high εNd section of the MORB array. However, a problem arises in that if plumes originate in, or traverse the lower mantle, and since entrainment of material by plume heads is thought to be inevitable (Griffiths and Campbell, 1990; Hart et al., 1992), the signature of the lower mantle component should also be observed in the OIB array. Buffering from the lower mantle also fails to account for the high εHf part of the MORB array, leaving the plume model with a final option of appealing to decoupling between Sm–Nd and Lu–Hf isotopic systems as originally proposed by Patchett and Tatsumoto (1980).

2.8. Confirmation of plumes by seismic tomography?

Currently, the best hope of resolving whether plumes exist or not should lie with seismic tomography. A broad correlation exists between hotspots and seismically slow (hot) regions of the shallow mantle (Anderson et al., 1992), but may merely reflect tendencies for intraplate volcanism to be located close to continental margins and/or continents to aggregate over cold mantle (Anderson, 1998a,b). Correlations between low velocity regions at the base of the mantle with the surface expression of selected hotspots have been used in support of the plume model even though more than 50% of the mantle at this depth is seismically slow (e.g., Cadek et al., 1995), but the correlations are as imperfect as those made with the geoid. Not all examples of intraplate volcanism correlate with seismically slow mantle at potential plume source depths, while other regions such as the mantle beneath the Bering Strait have low seismic velocities but no hotspots (Anderson et al., 1992; Ray and Anderson, 1994; Smith, 1998).

Another approach has been to try to image individual plumes. The first plume claimed to have been detected was under the Kodiak–Bowie chain (Nataf and VanDecar, 1993) on the basis of a low velocity anomaly present only in the lower mantle, and offset from the surface expression of the supposed hotspot. Likewise, Helmberger et al. (1998) claimed the existence of the Iceland plume on the basis of the island overlying a broad low velocity region at the base of the lower mantle, while Wolfe et al. (1997) reported the Iceland plume from a low-velocity region in the upper 400 km of the mantle. Higher-resolution studies (Bijwaard et al., 1998; Bijwaard and Spakman, 1999) have now claimed to have imaged a whole plume conduit under Iceland, even though the anomaly shows a major deflection between depths of 700–1400 km. Considering upwelling associated with ridges may extend into the lower mantle and has no particular requirement to be vertical (Su et al., 1992, 1994), claims for plumes on the basis of two-dimensional mantle sections in the absence of consideration of any other explanation cannot be considered justified. The same applies to interpretations of low-velocity anomalies under continental lithosphere as plume heads still underlying the Massif-Central (Granet et al., 1995), Parana (VanDecar et al., 1995), and Deccan (Kennett and Widiyantoro, 1999) basalt provinces even up to 130 Myr after the volcanism, rather than less spectacular interpretations of asthenosphere upwelling into lithospheric thinspots. The interpretation of the Massif Central anomaly by Granet et al. (1995) also demonstrates the need to consider the anomalies in a regional context, since it may represent nothing more than fine structure on a large-scale low-velocity feature noted by Hoernle et al. (1995) under Europe. With dimensions of 4000 $\times$ 2500 km and 500 km deep, the latter is too large for a plume head but would be quite consistent with a hotcell induced by thermal insulation.

The focus on interpreting low-velocity features only in terms of the plume model is further illustrated by studies on Hawaii, which should have the most readily resolvable conduit as it is situated away from ridge systems and is supposedly the strongest plume. Russell et al. (1998) and Ji and Nataf (1998) searched for low-velocity anomalies in the lower mantle beneath the hotspot, but found no low-velocity anomaly which correlated with the surface expression of volcanism. The lower mantle just above the core mantle boundary in the region comprises a series of band-like slow velocity anomalies alternating with less-slow regions. Consequently, both Russell et al. (1998) and Ji and Nataf (1998) invoked deflection of plume upwelling following the ideas of Steinberger and O'Connell (1998), but while Russell et al. (1998) suggested the conduit to lie to the southeast of Hawaii, Ji and Nataf (1998) claimed a double conduit to the northwest of Hawaii. In effect, the proofs of the plume model given in the tomographic studies amount to choosing the most suitable anomaly to fit the preconceived idea. When data is evaluated only in terms of the plume model, the number of hotspots can be arbitrarily adjusted, proofs can be made from fragmentary evidence, and lack of suitable correlations cannot be interpreted against the plume model, none of the tomographic studies can be considered to have been a valid test of whether plumes exist or not.

2.9. Plumes in the Precambrian

Intraplate basalts in the terrane accretion record can only be recognised as far back in time as the beginning of the Middle Proterozoic with the oldest known examples found in the Circum–Ungava belt of Canada (Gaonac'h et al., 1992; Blichert-Toft et al., 1996). However, the high eruption temperatures and depth of origin suggested from interpretations of komatiites as products of dry melting makes them candidates for hotspot volcanism. Melting studies on Kilbourne Hole spinel lherzolite KLB-1 have been used to determine depths of origin for alumina-depleted komatiites (ADK) of 9–11 GPa in the Early Archean, and 5–7 GPa for alumina-undepleted komatiites (AUK) in the Late Archean (Takahashi, 1986, 1990; Herzberg, 1995; Zhang and Herzberg, 1996). Application of the same logic that plume sources are 200 to 300°C hotter than the MORB source leads to two parallel and gradually decreasing geotherms for the MORB and plume-source through time (Fig. 9) (Richter, 1988). The apparent lack of plume-derived volcanism in the Archean can be then redressed by re-interpreting greenstone belts, once thought to represent ocean ridges or marginal basin sequences (Arndt, 1983; Nisbet and Fowler, 1983; de Wit et al., 1987), as oceanic plateaus produced by plumes (Campbell et al., 1989; McDonough and Ireland, 1993; Abbott, 1996; Kent et al., 1996) (Fig. 2b). As greenstone belts and gneiss terranes are the building blocks of the Archean crust, the principal form of crustal growth has to change from the accretion of arc material in the Phanerozoic (e.g., Samson et al., 1989) to the amalgamation of oceanic plateaus in the Archean (Condie, 1998). The plume

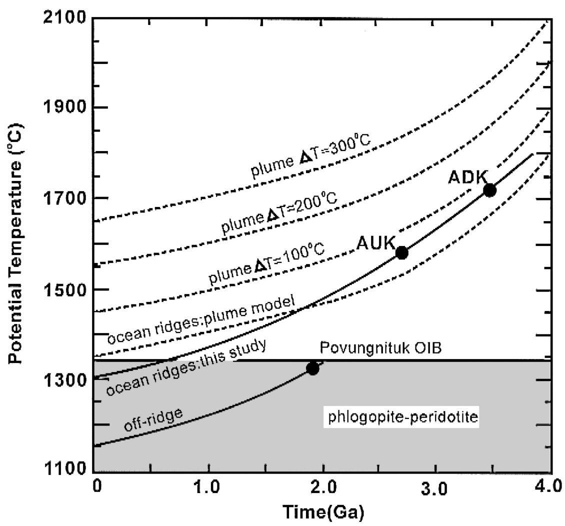


Fig. 9. Mantle thermal evolution through time. In the hotspot model, ridges and plumes define parallel decreases in potential mantle temperature while maintaining a temperature difference of $\Delta T = 100$ to 300°C. Alternatively, insulation from a voluminous early crust in conjunction with less efficient cooling of the mantle under slab melting regimes is suggested to lead to a thermal feedback scenario of high mantle temperatures in the Archean without having to invoke plumes. Combining the potential mantle temperatures of Takahashi (1990) for alumina-depleted (ADK) and alumina-undepleted (AUK) komatiite generation along ocean ridges with the estimate of McKenzie and Bickle (1988) of 1280°C for modern MORB generation, leads to a more rapid decline in temperatures reflecting increasingly efficient slab recycling and cooling of the mantle. The appearance of ocean island volcanism around 2.0 Ga corresponds to the time at which the average (off ridge) oceanic geotherm would permit the survival of volatile-bearing minerals, particularly phlogopite, in the shallow asthenosphere.

model also appears to reconcile crustal evolution curves (e.g., McLennan and Taylor, 1982) which depict a phase of major crustal growth in the Late Archean (Fig. 10) with the commonly used mantle evolution curves of DePaolo (1981) and Goldstein et al. (1984) which depict gradual evolution of the mantle (Fig. 8). If the continental crust is the geochemical complement of the depleted reservoir, the Late Archean pulse should be mirrored in the mantle evolution curves. However, if crustal growth takes place by an intermediate step such as plumes, a direct relationship is not expected since crustal growth becomes a function of maturation of plume sources and the depleted mantle can be considered the residue from plumes.

While superficially appealing, the logic used to construct the Archean plume model encounters many difficulties. Not only does the crustal growth mechanism change, but also the properties of plumes themselves. In the Archean plume model of Campbell et al. (1989), basalts are produced by the cooler plume head as it interacts with asthenosphere while komatiites are produced by the hotter plume tail, whereas picritic basalts are produced during the early stages of modern plume activity (e.g., Storey et al., 1991). The mantle isotopic evolution curves of DePaolo (1981) and Goldstein et al. (1984) may also be oversimplified since they were based on early Nd data obtained when it was common practice to derive Sm–Nd isochrons from a range of rock types whose cogenetic relationships were often dubious. Mixing rock types from komatiites to felsic volcanic rocks has the effect of rotating an isochron to both greater slope and hence age, and to lower εNd (Smith and Ludden, 1989). While the ages were re-evaluated (e.g., Claoué-Long et al., 1984), the mantle evolution

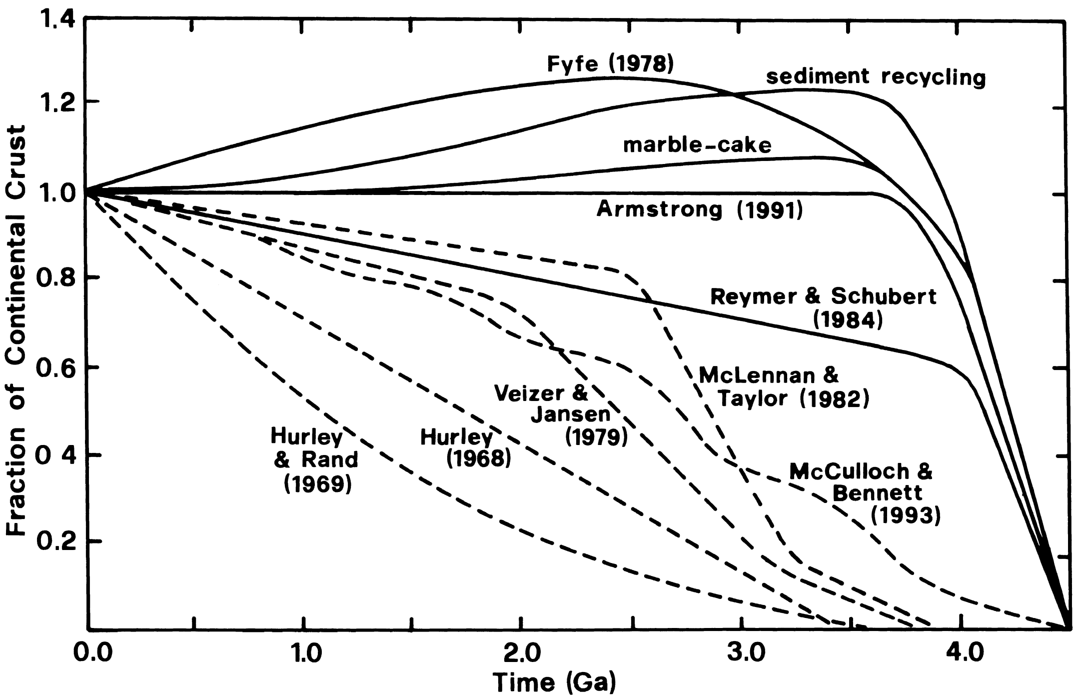


Fig. 10. Models for the evolution of the Earth's crust can be divided into those depicting rapid or gradual formation of the continents. The latter category (dashed lines) is compatible with the mantle plume concept and depictions of a gradually evolving depleted mantle reservoir. Models depicting rapid formation of a large early crust (solid lines) include those involving steady-state recycling (Armstrong, 1968, 1991) and those where recycling becomes more significant through time such that the volume of continental crust in the Early Archean is larger than in existence today (Fyfe, 1978; this study). Curves labelled sediment recycling and marble-cake correspond to the mantle buffering scenarios in Fig. 7b and c, respectively. Data from Hurley (1968), Hurley and Rand (1969), Fyfe (1978), Veizer and Jansen (1979), McLennan and Taylor (1982), Reymer and Schubert (1984), Armstrong (1991) and McCulloch and Bennett (1993).

curves were not. Recalculation of εNd values for mantle-derived volcanic rocks to best U–Pb or Pb–Pb ages results in far greater isotopic variation in the Precambrian mantle, with komatiites often having higher εNd than tholeiites (Smith and Ludden, 1989; Anderson, 1994b). The isotopic relationship of intraplate to asthenosphere-derived melts therefore also has to change through time, since modern OIB have lower εNd than MORB (Smith and Ludden, 1989; Anderson, 1994b). High εNd values and the requirement to delay crustal production until the Late Archean mean the plumes must have been of the recycled crustal type. But if the oceanic crust was basaltic, the recycling mechanism has to change to avoid the improbability of deriving a komatiite from a predominantly basaltic parent, such that peridotite could not be stripped from the slab as in models for modern plumes. However, any Archean plume model based on crustal recycling of basaltic material (e.g., Chase and Patchett, 1988; Condie, 1998) is potentially defeated by the paradox of Bickle (1978) whereby higher temperatures should result in greater ridge lengths along with thicker, younger and less readily subductable basaltic crust. A komatiitic crust would avoid the paradox (Nisbet, 1987), but the plume model would then become self-contradicting in duplicating an origin for komatiite generation for which it was itself invoked to avoid.

2.10. Toward a synthesis?

The stranglehold on geodynamic interpretations achieved by the plume model in the 1990s belies the complexity and internal contradictions within the model. Although plume theory now controls plate tectonics, crust–mantle evolution, and even the Earth's magnetic field (Fig. 11a) from suggestions that the release of material from the D″ layer controls convection in the outer core (Vogt, 1975; Larson and Olson, 1991), even the most basic problems such as the depth of plume-source layer(s) remain unresolved. Upper mantle plume sources have been argued to be less suitable in accounting for $^3He/^4He$ signatures (e.g., White and McKenzie, 1995) and cannot be integrated with the superplumes concept where it has been calculated that the diameter of massive plume heads would be greater than the depth of the upper mantle (Coffin and Eldholm, 1993). A plume-source layer at the core–mantle boundary has been the most favoured option, despite uncertainties as to whether geochemical signatures are appropriate for equilibration with high pressure assemblages (Ringwood, 1990) and whether plumes from such depth could penetrate a layered mantle regime (Nakakuki et al., 1994). The most recent trend has been toward models incorporating alternating phases of upper and lower mantle-derived plume activity. These MOMO models (for mantle overturn, major orogeny; Stein and Hofmann, 1994; Hofmann, 1997) follow from considerations of densities (Irifune and Ringwood, 1993) supported by tomographic evidence (Bijwaard et al., 1998; van der Hilst et al., 1998) that slabs may encounter difficulty penetrating into the lower mantle. Subducted oceanic crust is then thought to accumulate at the base of the upper mantle, akin to the megalith concept of Ringwood (1982), before foundering into the lower mantle and causing mantle overturn.

While the MOMO models avoid some of the difficulties of restricting plumes to upper- or lower mantle sources, they do little to address problems such as the number of hotspots, whether hotspots are fixed, or why plume features should reverse in the Precambrian, nor do they provide any explanation for residual volatile-bearing minerals in the sources of OIB. The inadequacy of a partial explanation for modern intraplate volcanism and duplication of processes also remains. Invoking ''fossil plumes'' (e.g., Halliday et al., 1990; Stein and Hofmann, 1992) or contamination of the mantle with plume material to distribute the appropriate source material throughout the shallow mantle does not avoid the fundamental problem that non-plume mechanisms are still required for the tapping of such sources. The startling aspect to the progression of thinking is that if non-plume mechanisms must exist, then why have they not been evaluated parallel with, and to the same extent as the plume model? Rather than looking for ways to perpetuate the plume model by combining different variations, the arguments against them should have been taken to indicate that all are fundamentally wrong. The question which should have been posed all along is not what additional variations would be required to make the plume model fit, but what would the Earth look like if intraplate volcan-

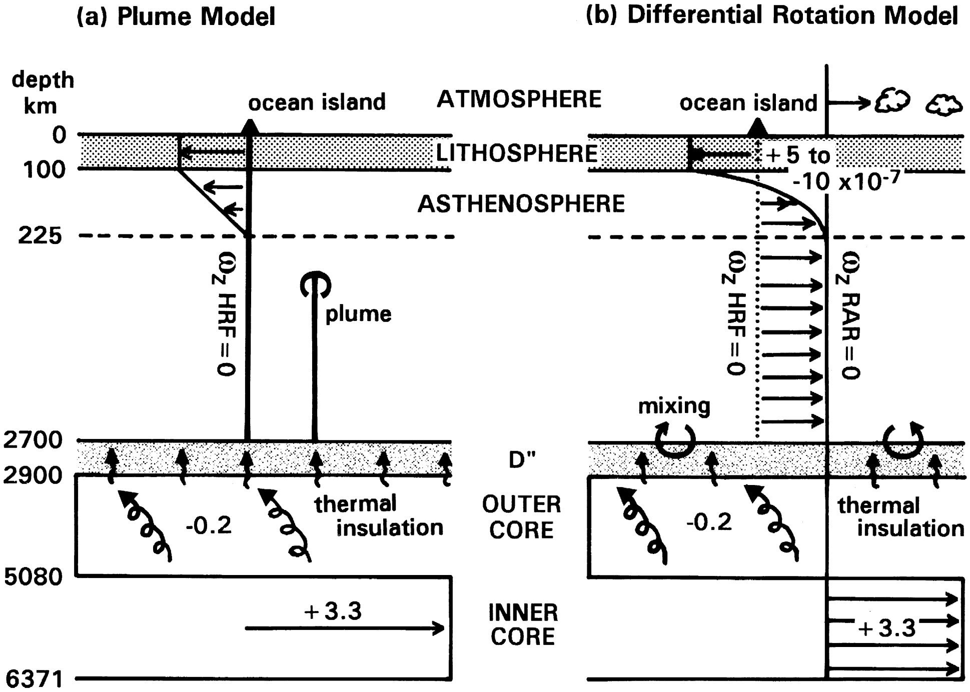


Fig. 11. Causes of the correlation between magnetic reversal frequency and tectonic events according to: (a) The plume model (D″ source layer variation only). The pattern of convection in the outer core generating the magnetic field is controlled by the insulating effect of the D″ layer which in turn is controlled by the release of plumes (b) Differential rotation model. Subducted lithosphere may insulate the core prior to remixing into the mantle, but the link with surface tectonic events is provided by the balance of angular momentum between the lithosphere (counterflow regime of Fig. 4e illustrated), mantle and core. If the eastward motion of the inner core and mantle is constant, changes in angular momentum related to lithospheric plate configuration can only be balanced by changes in the rate of westward drift of the convection pattern in the outer core. Numbers indicate the vertical components of motion in degrees per year relative to the hotspot frame (HRF) or relative to the rotation (Z) axis reference (RAR) as for Fig. 12. The location of sections relative to the Earth's structure is shown in Fig. 2a,c.

ism and mantle geodynamics were explained by the non-plume processes?

3. In retrospect: a path that might have been followed

3.1. Wetspots not hotspots

The starting point for the construction of any counter-model has to be an acceptance of the evidence for amphibole and phlogopite in the source of OIB which has been presented for as long a time as time plume models have assumed a thermal anomaly (e.g., Tuthill, 1969; Green, 1971; Flower et al., 1975; Clague and Frey, 1982; Bonatti, 1990; Francis, 1995; Francis and Ludden, 1995). The requirement for these volatile-bearing minerals indicates that intraplate volcanism results predominantly from compositional, not thermal anomalies. The low melting point of such minerals would make them susceptible to shear melting to generate intraplate tracks, while entrainment into ridge upwelling or local asthenospheric convection cells would lead to rapid generation of large quantities of melt thereby explaining the generation of large igneous provinces (Smith and Lewis, 1999a). Such minerals are also eminently suitable from the geochemical perspective as the parent/daughter ratios of amphibole would make it a candidate to carry the EM1 isotopic signature, while phlogopite would be a candidate to carry EM2. But invoking volatile-bearing peridotite assemblages leads to two problems, how to account for the origin of the material and how to explain its distribution as amphibole and phlogopite are conventionally associated with the sources of arc volcanism.

Models for modern arc volcanism have proposed that fluids or melt invading the mantle wedge crystallise hydrous assemblages which may be dragged to depths of 150 km above the subducting slab before dehydrating and releasing fluids to shallower levels (e.g., Wyllie and Sekine, 1982; Tatsumi, 1989). Dehydration of deeper assemblages is also postulated to occur on termination of subduction activity, imparting a geochemical layering with phlogopite surviving at greater depths whereas volatiles invading shallower levels of the mantle wedge may precipitate amphibole. But while arc and intraplate melts share a common requirement for equilibration with a harzburgitic rather than lherzolitic source as in the case of MORB genesis (Francis, 1995), differences in high field strength element (HFSE) content have appeared to preclude a simple relationship (Sun and McDonough, 1989). Retention of HFSE in phases such as rutile in the subducting slab has been emphasised (e.g., Ryerson and Watson, 1987), but experimental evidence has indicated that these elements are not exclusively retained in the subducting slab. If the amount of Nb lost to slab-derived fluids is only slightly less than the amount of Ce (1.0% vs. 2.3% of the slab budget; Brenan et al., 1995a,b), prolonged episodes of subduction could potentially allow significant refluxing of HFSE into the mantle wedge. The HFSE depletion in arc magmas could therefore result from partitioning of these elements into phases such as olivine and orthopyroxene in the mantle wedge which take little further part in subsequent melt generation (Kelemen et al., 1992). This type of metasomatic model has received support from analyses of minerals in lherzolite xenoliths from regions where subduction has formerly taken place (Ionov et al., 1995; Sun and Kerrich, 1995), and allows a relationship between residues from arc volcanism and sources for intraplate volcanism.

Alternatively, the slab dehydration regimes of modern subduction zones are not likely to be representative of subduction processes through time. With higher mantle temperatures in the Archean, the geothermal gradient intersects the breakdown curve for amphibole before the water-saturated basalt solidus, which in conjunction with the subduction of younger and possibly thicker crust, would result in melting before the stability limit of amphibole was reached (Martin, 1986, 1993; Peacock et al., 1994). Depending on temperatures and the amount of komatiitic material in the slab, such dehydration-melting processes could generate melt fractions up to 70% (Martin, 1986). While HFSE are incompatible in aqueous fluids, they are more strongly partitioned into silicate melts giving the remnants of ancient collision zones in the continental mantle an advantage over modern mantle wedges as later sources of intraplate volcanism. Signatures from ancient metasomatic events, of course, could only be preserved in a long-lived reservoir such as the continental mantle. It therefore becomes necessary to constrain whether wetspots are produced in situ throughout the shallow mantle and intraplate volcanism is controlled by processes tapping the enriched domains, or whether volatile-bearing minerals are periodically introduced into the asthenosphere such that the presence of wetspots is a necessary pre-condition for the generation of intraplate melts.

3.2. Perisphere or continental mantle as a source for intraplate volcanism?

There are essentially two models which have dealt with a shallow origin for the sources of intraplate volcanism on a global scale. The concept of an enriched ''perisphere'' layer residing between the lithosphere and MORB-source has been proposed by Anderson (1995; 1996; 1998a). The perisphere includes the upper part of the asthenosphere and the thermal boundary layer of the continental mantle. The layer undergoes continuous enrichment from subduction processes, but is essentially static and hence encounters difficulty in generating long-lived volcanism as along the Hawaiian chain (Hofmann, 1997). Other difficulties exist with regard to isotopic signatures, as intra-oceanic arcs are typically characterised by EM2 which is not a major signature in intraplate volcanism. The EM1 and HIMU signatures which predominate in intraplate volcanism are clearly ancient and cannot be related to relatively recent metasomatic events. Invoking sediment (e.g., Tu et al., 1991; Farley, 1995) as a perisphere component defers isotopic ageing to before the sediment enters the subduction zone, such that the EM1 and EM2 signatures can be linked to contamination with pelagic and turbidite compositions as in plume models. However, as isotopic evolution would produce a range of $^{176}Hf/^{177}Hf$, any sediment-contaminated mantle has to be less than 200 Myr old; otherwise the model faces the same problem as the plume model with regard to correlated Nd–Hf isotopic compositions outlined in Section 2.7. High $^{3}He/^{4}He$ signatures can potentially be explained by invoking cosmogenic components subducted with the sediment (Anderson, 1996). But while the sediment contamination mechanism has been proposed as a means for generating the DUPAL domain in Asia, similar domains in the Arctic and the globe encircling belt at 20°S either lie perpendicular to regions of current subduction or in areas where sediment subduction has not recently taken place. The sediment model also fails to explain the overlap in isotopic compositions between continental types of intraplate volcanism such as carbonatites (which are almost certainly not derived from subducted detrital sediments) and OIB (Bell and Blenkinsop, 1989).

The alternative to the perisphere model is that the sources of intraplate volcanism are derived solely from continental mantle (Smith, 1993). Structurally, this reservoir can be envisaged as comprising a cold buoyant mechanical boundary layer (lithosphere) overlying a more transient thermal boundary layer which is transitional to the asthenosphere in thermal profile and viscosity (Menzies, 1990). Cratonic parts of the continental mantle contain refractory peridotites, the depletion of volatiles shown by such material likely being a necessary part of the cratonisation process (Pollack, 1986). However, as continents usually collide by subduction, the cratonic roots will be surrounded by suture zones. The thermal boundary layer may thus be largely characterised by EM2 isotopic signatures, while the mechanical boundary layer displays EM1 signatures. Small melt fractions may also migrate into the continental mantle from the asthenosphere (McKenzie, 1989), potentially as far as the mechanical boundary layer (England, 1993). Such melts would inherit a depleted isotopic signature from the asthenosphere, but if crystallising as pyroxenite, parent/daughter ratios imparted on partial melting would cause evolution to HIMU signatures over timescales of a few hundred million years (Meijer et al., 1990; Smith, 1992, 1998). The isolation of the mechanical boundary layer would also make it a suitable reservoir to carry high $^{3}He/^{4}He$ signatures, as the observation that the highest $^{3}He/^{4}He$ characterise samples with lowest He abundance, strongly suggests such signatures result from shielding from sources of ^{4}He, not the presence of excess ^{3}He (Anderson, 1998c). Regions of continental mantle affected by subduction would likely be volumetrically greater than regions which have undergone shearing and infiltration by melts, hence the predominance of EM over HIMU compositions in intraplate volcanism (Fig. 6).

The advantage of the continental mantle is that it is a reservoir where isotopic signatures may age, and where signatures produced under different tectonic regimes throughout Earth history may be retained. It is also a dynamic reservoir which evolves as part of the supercontinent cycle. Thermal boundary layer material is susceptible to erosion on continental rifting and mechanical boundary layer material to delamination on continental collision (e.g., Gallagher and Hawkesworth, 1992; Black and Liégeois, 1993; England, 1993), though losses may potentially be compensated if cold asthenosphere attaches as continents move to rest over mantle downwellings. The present continental area is approximately 172×10^{6} km^{2} (Forsyth and Uyeda, 1975). To generate the volume of intraplate volcanism since the Cretaceous, even assuming a large 10% melt fraction, the amount of continental mantle required would only correspond to an average thinning of 16 km. Even allowing for losses being largely restricted to under Phanerozoic belts or small cratons, the amounts are still a factor of three less than most estimates of the thickness of the thermal boundary layer, and only 10% of the total thickness of the continental mantle estimated by Polet and Anderson (1995). The key to invoking such material as the source for intraplate volcanism lies in explaining its distribution throughout the asthenosphere. The perisphere model has its source components in place, so does not encounter the same problem. The perisphere model is also compatible with plate motions in the hotspot reference frame. But if the model is intended to replace plumes there is no requirement to retain the hotspot frame. When the effects of this reference frame are removed, it will be argued there not only arises a mechanism for the lateral distribution of eroded continental mantle throughout the asthenosphere, but also mechanisms for tapping such material as a result of Earth rotation.

3.3. Westward plate lag and eastward mantle flow

Before the hotspot frame was proposed, the Antarctic plate was used as a reference because it is nearly surrounded by ridges and so can be regarded as stationary while its position centred about the Earth's rotation axis allows plate motions to be directly related to the x, y, z components of a triaxial Earth (Lewis and Smith, 1998). Initial modelling of plate interactions using this plate as a fixed reference indicated a net westward drift of plates at an average rate of 5 cm yr^{-1} (LePichon, 1968; Knopoff and Leeds, 1972), somewhat reminiscent of the ''pollflucht'' of Wegener (1920). Unfortunately, many of the early explanations for a net westward plate drift focused on causes external to the Earth such as tidal lag (Bostrom, 1971; Moore, 1973). Such mechanisms were subsequently shown to be inadequate by Jordan (1974), who not only refuted the mechanism but also the observation of westward plate movement as being incompatible with plate motions in the hotspot frame. The possibility that it was the plume model that was wrong was never considered, even though this was what was indicated by the equivalence of plate motions relative to hotspots or a fixed Antarctic plate demonstrated by Lliboutry (1974) (Fig. 12).

An internal explanation for a westward plate drift can be found when the Earth is envisaged as a series

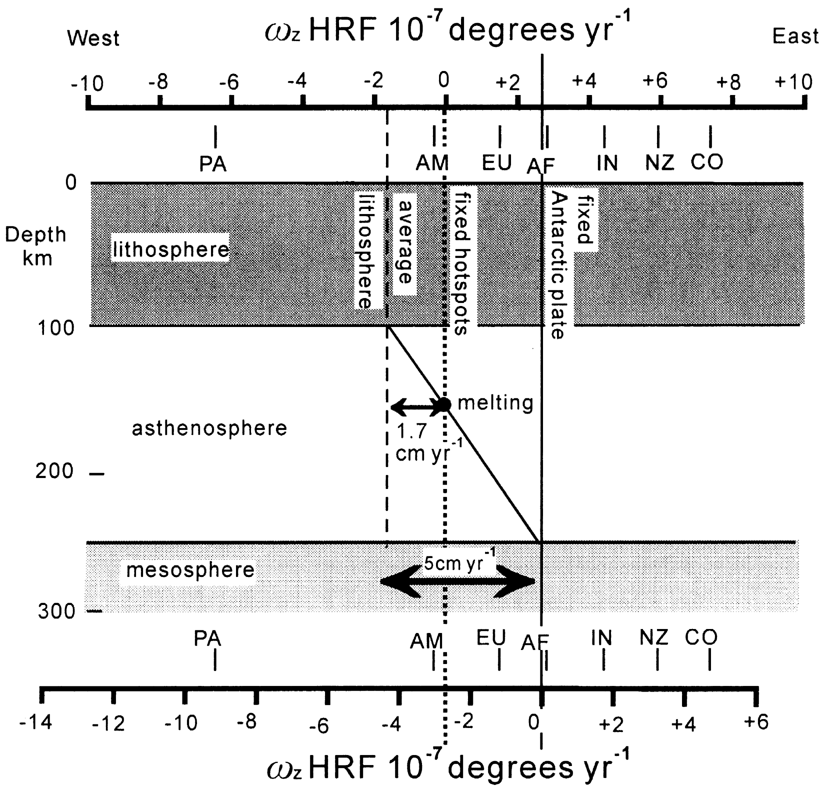


Fig. 12. Vertical components of plate motion vectors relative to the Earth's rotation axis in the hotspot reference frame (ω_Z HRF) and relative to a fixed Antarctic plate (ω_Z RAR; rotation axis reference) (after Lliboutry, 1974; Smith and Lewis, 1999b). Average lithospheric differential rotations of 1.7 cm yr^{-1} in the hotspot reference frame (Ricard et al., 1991) compared to 5 cm yr^{-1} relative to a fixed Antarctic plate (LePichon, 1968) can be explained by the sources of intraplate volcanism lying at shallow depths in the asthenosphere. Flow profiles under individual plates are expected to be modified by boundary forces as shown in Fig. 4.

of shells of differing viscosity rotating about a central axis (Fig. 1b). Taylor numbers for the mantle ($T \sim 10^{-17}$) are far less than the values around unity which would be expected if mantle convection was affected by Earth rotation. It is therefore important to emphasise that plate lag results from the asthenosphere acting as a zone of decoupling between the lithosphere and the deeper, convecting mantle. The mantle is thought to behave as a Newtonian fluid (e.g., Bott, 1984) such that transmission of stresses will depend on viscosity. The presence of a radial low-viscosity layer such as the asthenosphere will thus affect coupling between the surrounding more rigid shells (Smith and Lewis, 1999b). As the Earth rotates, a drag will be exerted on the asthenosphere by the mesosphere. Flow induced in the asthenosphere will in turn exert a drag on the base of the lithosphere, but as the viscosity of the asthenosphere is reduced by its thermal profile, the drag also is reduced. In conjunction with plate boundary forces, the result is a differential rotation characterised by what appears to be a net westward lag of plates (Fig. 1b) as mantle and lithosphere move about the Earth's rotation axis with different velocities. The effect will not be purely west–east as undulations in the flow lines are expected as the position of the rotation axis is expected to move in response to plate movements changing the mass distribution at the Earth's surface (polar wander). As the effect is also a plate lag relative to the deep mantle, it can alternatively be envisaged as eastward mantle flow. However, after introduction of the hotspot frame the concept of differential rotation was largely ignored until re-introduced by Doglioni (1990) as a cause of slab steepening beneath westward-dipping subduction zones to produce structural features of arcs. The re-evaluation of westward plate lag by Ricard et al. (1991) is particularly significant in demonstrating that a net westward drift of plates, albeit reduced to a global average of 1.7 cm yr^{-1}, still exists in the hotspot frame. The average polar wander rate plus the rotation differential in the hotspot frame equals the average 5 cm yr^{-1} plate lag determined from the rotation axis reference (Fig. 12). If hotspots were truly fixed and independent of mantle flow, there should be no net westward plate lag. That requirements for it still exist, and that the magnitude is reduced in the hotspot frame, suggests that (1) westward plate lag is a real effect and not an artefact of using the Antarctic plate as a reference and (2) hotspots are actually a series of shallow melting anomalies which are moving relative to both the lithosphere and mesosphere (Smith and Lewis, 1999b).

The outer core also constitutes a major radial reduction in viscosity within the Earth's structure and should also be susceptible to differential rotation effects. Westward drift of the non-dipole moment of the magnetic field at a rate of 0.20 yr^{-1} relative to the mantle has been known for some time from observations on the magnetic field (Jault et al., 1988). Recent studies (e.g., Su et al., 1997) have also shown the inner core to be rotating eastwards at a rate of 3.28° yr^{-1} relative to the mantle, in the opposite direction to the non-dipole moment. In the plume model, the motions of the various shells of the Earth appear chaotic with the inner and outer core moving in opposite directions about the rotation axis, the mantle rotating about an equatorial axis, and the lithospheric plates moving in no predominate direction over the surface (Figs. 1a and 11a). In the differential rotation model, the motion of the shells is more regular with the lithosphere and convection pattern in the outer core showing a net westward drift, while the mantle and inner core move eastwards relative to the rotation axis (Figs. 1b and 11b). Previous studies have argued for a balance of angular momentum between the core and mantle system and noted that torques exerted by atmospheric circulation can lead to short-term perturbations in the Earth's rotation (Hide, 1993; Jackson et al., 1993). Under such circumstances it is reasonable to expect changes in lithospheric plate configuration to also affect the balance of angular momentum. As the total angular momentum of the Earth must remain constant in the absence of any changes in external torques (Jackson et al., 1993), balancing angular momentum changes in the lithosphere by changes in core convection patterns would provide a link between magnetic reversals and plate tectonics.

3.4. Continental rifting by basal drag

Requirements for plate drag and boundary forces acting in conjunction can be readily accommodated in the differential rotation model as drag is continuously exerted on the base of the mesosphere and transmitted upwards to the plates according to an extent determined by thickness and viscosity of the asthenosphere (Smith and Lewis, 1999b). It is this drag force which is envisaged to replace the role of plumes in continental rifting. The mantle flow inherent in the mechanism also offers an explanation for an eastward migration of tectonic cycles described by Trurnit (1991). The limiting case for continental drag would be when the continental mantle is sufficiently thick to preclude the existence of any underlying asthenosphere. Momentum would then be transferred from the deep mantle directly to the lithosphere such that the continent would appear fixed relative to the mantle. With the exception of Antarctica which is located on a pole where the velocity differential is zero (hence use of this plate as a reference), the continent which most closely approaches this scenario is Africa, which has been depicted as stationary or as showing only a small lag relative to the rotation axis reference (Knopoff and Leeds, 1972; Lliboutry, 1974). The drag forces of 2×10^{19} to 10×10^{19} N estimated by Smith and Lewis (1999b) to act on this continent are in good agreement with the estimates of Chapple and Tullis (1977) and would be comparable to or exceed the combined effects of slab pull and ridge push. Stresses arising from drag on the base of a supercontinent 10 000 km in diameter would be around 20 MPa, though they would not be evenly distributed depending on positions of the continent relative to undulations in mantle flow pattern and to the equator. A tendency for plates in the northern hemisphere to rotate clockwise, and correspondingly plates in the southern hemisphere to rotate anti-clockwise similar to the Coriolis effect in surface fluids (Bostrom, 1986), may be attributed to the rotation differential being greatest at equatorial latitudes, producing differential hemispheric torques on plates which cross hemispheres. Such torques, in conjunction with membrane stresses arising from flexing and hence weakening of plates as they cross the Earth's equatorial bulge (Turcotte and Oxburgh, 1973), may contribute to the preferential breakup of plates in the equatorial region.

The other principal control on continental breakup is the lithospheric architecture itself since rifting invariably follows pre-existing sutures or lineaments which cut the lithosphere (e.g., Bailey, 1992; Smith, 1993). The sutures serve as lines of weakness from a mechanical standpoint (Vissers et al., 1995) and will be particularly prone to reactivation if they contain low melting point hydrous minerals remaining from subduction activity which brought the continental blocks together. Sutures and lineaments are also likely to mark topographic variations at the lithosphere–asthenosphere boundary, which are particularly critical as mantle flow around such features may induce small-scale convection which further erodes the thermal boundary layer of the continental mantle (Bailey, 1980; King and Anderson, 1995, 1998; Sheth, 1999b). Hence sutures orientated north–south and hence opposing eastward asthenospheric flow tend to be reactivated preferentially. Doglioni (1990) termed the latter first order rifts while lineaments perpendicular to mantle flow constitute second order rifts which tend to remain closed.

3.5. *Intraplate volcanism in opening ocean basins and other manifestations of convective melting*

The convection cells induced by topography at the continental mantle–asthenosphere boundary during rifting will be transverse rolls of the type envisaged by Richter (1973) with the axis of the cells parallel to that of the rift. Continental margin volcanic sequences which parallel the axis of rifting are readily interpreted in terms of decompression melting as continental mantle is eroded and cycled toward the rift axis (e.g., Mutter et al., 1988; Holbrook and Kelemen, 1993; Keen et al., 1994). The more localised occurrence of other intraplate volcanic features reflects lithospheric structures intersecting the axis of rifting (Smith, 1993), akin to observations of carbonatites and kimberlites being associated with, and often overlying intersection of lineaments cutting thick cratonic crust (Bailey, 1977; Mitchell, 1986; Woolley, 1989). Anderson (1996) pointed out that flood basalt provinces are generally located where younger crust is juxtaposed against Archean cratons. The third dimensional control is the extension of a lineament or suture into the zone where the continental lithosphere changes thickness (Smith, 1993). Volcanic lines progressing from the basalt province are governed by the suture, hence intraplate volcanic tracks in opening ocean basins invariably extrapolate into sutures/lineaments in the continental mantle which intersect the axis of rifting (LePichon, 1968; Marsh, 1973; Sykes, 1978). The convection cells expand as rifting progresses, cycling continental mantle from the suture toward the spreading axis, thereby producing a volcanic tracks which may be mistaken as plume tail effects, while erosion of continental mantle from sutures parallel to the rift axis may produce oceanic rises.

The controls of lithospheric structure are readily illustrated by intraplate volcanism in the Atlantic Ocean basin, which can be considered the simplest example of rifting as the ridge system still parallels the continental margins (Fig. 13). Opening of the North Atlantic followed a Caledonian suture between North America and Eurasia, while opening of the South Atlantic followed the Pan-African suture of the Adamastor Ocean. From north to south, volcanic activity on Jan Mayen can be interpreted as a southward propagating fracture emanating from the truncation of a segment of spreading axis by a major transform in conjunction with lithospheric weakening related to the presence of a nearby continental fragment (Hasse et al., 1996). Volcanism on Iceland extrapolates into a Proterozoic lineament cutting across Greenland and extending into the North Sea Rift system (Smith, 1993). The New England Seamounts mark a continuation of volcanism in the St. Lawrence graben system. The Canary Islands lie on a continuation of Hercynian structures in the Atlas Mountains and can be explained by a fracture propagating westwards into the ocean basin model (Anquita and Hernan, 1975). The Cape Verde and Bermuda Rises parallel north–south sutures in the Appalachians and along western Africa, respectively (Vogt, 1991; Smith, 1993). The Cameroon Line is associated with the Benue Trough, and the Walvis Ridge extrapolates into the Damara belt with the Parana–Etendeka flood basalt provinces overlying the intersection of the Pan African Adamastor and Damara sutures (Smith, 1993). The Rio Grande and Falkland plateaus follow Mesozoic fracture systems (LePichon, 1968).

The Indian Ocean is a more complex case in that the basin formed from the separation of three continental masses (Africa, India, and Australia–Antarctica), though lithospheric control of intraplate volcanism is similar to in the Atlantic basin. The Karoo flood basalts of southern Africa are located around and along the suture between the Kaapvaal and Zimbabwe cratons. The Ninety-east and Eighty-five-east ridges can be explained as leaking transform faults. Flood basalts of the Deccan Traps are associated with the intersection of the Narmada and Cambay rifts (Sheth, 1999a). The topographic structure of the former would have been perpendicular to the direction of motion, and is postulated to have induced convection beneath the continental block. Eroded continental mantle could later have been tapped by a southward propagating fracture to form the Chagos–Laccadive–Réunion ridge, with the track on the Indian plate following the Vishnu fracture zone and the track on the African plate following the Mauritius fracture zone (Sheth, 1999a). The volcanism also occurred as the continental block approached the equatorial region and hence may have experienced differential torques as it crossed hemispheres in addition to membrane stresses from crossing the equatorial bulge. The Emeishan flood basalts of South China were erupted under strikingly similar conditions to the Deccan, being associated with a southwest–northeast trending suture (Gilder et al., 1991) on a block which crossed the equatorial region at the end of the Permian-beginning of the Triassic.

While convection melting is particularly important during continental rifting, induced local convection is not necessarily limited to extensional settings, as can be illustrated by the Columbia River and Siberian Traps flood basalt provinces. The Columbia River province lies at the tip of a lithospheric thinspot created by Basin and Range extension, and is also adjacent to Archean crust of the Wyoming province. Generation of large volumes of melt in this province may be explained as a result of induced convection supplemented by back-arc upwelling, with later volcanism along Snake River Plain marking the reactivation of a Proterozoic suture by stresses induced from interactions between North America and oceanic plates to the west (e.g., Christiansen and McKee, 1978; Hart and Carlson, 1987; Smith, 1992). Correspondingly, the Siberian Traps erupted during a time of continental convergence can be explained by convection induced at the leading edge of the continental plate (Czamanske et al., 1998).

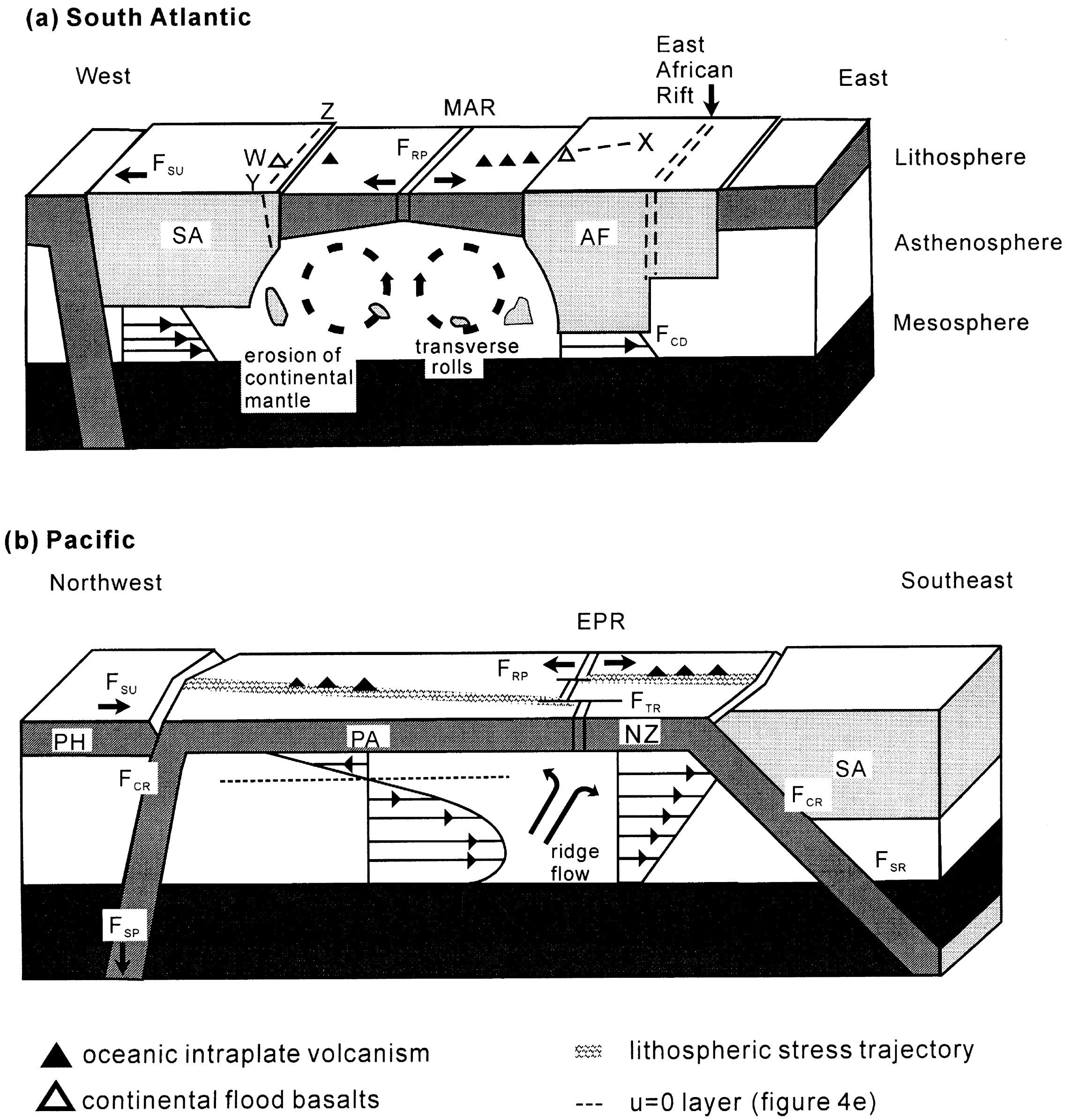


Fig. 13. Forces acting on plates in the differential rotation model illustrated by sections through: (a) the present South Atlantic basin. Rifting occurs parallel to pre-existing sutures (e.g., dashed line Y–Z; suture of Adamastor Ocean). Intraplate volcanism is found where second-order sutures (e.g., dashed line W–X; Damara belt) intersect the axis of rifting, indicating an origin related to erosion and cycling of continental mantle toward the ridge axis by local convection taking the form of transverse rolls. (b) The modern Pacific basin parallel to direction of asthenospheric flow. In the eastern part of the basin, flow profiles induced by mesospheric drag and boundary forces act in the same direction, giving rise to Couette flow under the Nazca plate (Fig. 4c). In contrast, conflicting patterns of flow induced by mesospheric drag and plate boundary forces set up a counterflow regime under the Pacific plate. Melt produced from shearing of volatile-bearing peridotite collects in a stationary layer at shallow level ($u = 0$; see Fig. 4e). Release of melt from this layer is governed by stress trajectories in the overriding plate, giving rise to ocean island volcanism. The pattern of upwelling under the East Pacific Rise corresponds to the sum of asthenospheric flow patterns under the Pacific and Nazca plates. Plate moving forces (following Forsyth and Uyeda, 1975) F_{CD} continental drag, F_{CR} continental resistance, F_{RP} ridge push, F_{SP} slab pull, F_{SR} slab resistance, F_{SU} trench suction, F_{TR} transform resistance; Plates: AF Africa, NZ Nazca, PA Pacific. PH Philippine, SA South America, Ocean ridges: EPR East Pacific Rise, MAR Mid Atlantic Ridge.

3.6. Stress fields and intraplate volcanism in long-lived ocean basins

Asthenospheric domains contaminated with eroded or delaminated continental mantle which were not immediately tapped to generate intraplate volcanism, would become displaced eastwards relative to the continent over time. The low melting point of volatile-bearing minerals would allow survival only in off-ridge environments (Fig. 3). The asthenospheric domains could potentially be tapped as a result of localised convection. Transverse rolls of the type invoked under the Atlantic basin would not be stable with the faster velocities of oceanic plates and would be transformed into longitudinal rolls with axes perpendicular to the ridge system (Richter, 1973), producing volcanic lines as in the ''hotline'' model of Bonatti and Harrison (1976). Alternatively, Shaw (1973) using Hawaii as an example, proposed that intraplate volcanism could reflect focusing of melts generated by shear melting, with the apparent episodicity of eruptions reflecting thermal buoyancy and collection of melts in shallow source regions. In the differential rotation model, shear stress profiles through the asthenosphere reflect the relative interactions of drag induced from the mesosphere and lithosphere (Fig. 4). When boundary forces act in the opposite direction to eastward mantle flow, the result is a counterflow regime characterised by the presence of a stationary layer at the crossover between opposing flow regimes. In the model depicted in Fig. 4e where lithosphere and asthenosphere are 100 km thick, the stationary layer would lie at a depth of around 128 km. Phlogopite, but not amphibole would be stable at depths greater than 100 km (Fig. 3). The stationary layer would rise to shallower depth with decreasing mesospheric velocity (higher latitude), higher plate velocity, or decreasing asthenospheric thickness, such that under oceanic plates less than 100 km thick, it could potentially lie within the stability field of amphibole. The stationary layer is also characterised by high shear stress, and thus with similar features to the horizontal stress trajectory of Shaw (1973), would be susceptible to the same thermal feedback mechanism for melting.

Release of melt from a horizontal stress trajectory was envisaged by Shaw (1973) and Jackson and Shaw (1975) to be governed by vertical lithospheric stress trajectories. When the outlines of Pacific plates are depicted as polygons, intraplate volcanic features formed since at least the Early Cretaceous can be seen to have followed a limited set of six orientations (N87°E, N54°E, N21°E, N11°W, N40°W, N67°W; Fig. 14). Plate boundaries along the margins of the basin also follow the same set of six directions, which would appear to define a classic fracture pattern since the means of each set of directions lie 30° apart. Eocene to Recent intraplate volcanism is concentrated largely into three zones spaced 1500 to 2000 km apart (Fig. 14). In the west, these zones coincide with major changes in orientation along the margin of the Pacific plate, while in the east the zones correlate with plate boundaries within the ocean basin or major fracture zones along the East Pacific Rise. Thus from south to north, the Louisville, Cook–Austral–Marqueses, and Hawaiian chains are parallel to the Eltanin–Heezen fracture zone system, the Nazca–Pacific plate boundary, and the Quebrada fracture zone, respectively. As the rigidity of plates allows transmission of stresses great distances from their boundaries (Sykes and Sbar, 1973; Solomon and Sleep, 1974; Zobak, 1992), the zones of volcanism appear to follow stress guides set up at the convergent margin as a result of tension applied by unequal boundary forces acting on different segments of the subducting slab, in agreement with the earlier stress field models.

If melts were generated by shearing within the stationary asthenospheric layer, or collected in such a layer, before being channelled by the lithospheric stress trajectories (note the asthenosphere immediately under the lithosphere in Fig. 4e behaves as lithosphere), the result would be analogous to the presence of a series of fixed, deep-seated melting anomalies under the plate. Cessation of volcanism in the absence of changes in the stress field can be explained by exhaustion of low-melting point minerals, though conversely, as the source material is also moving laterally, tapping of an extensive streak of eroded material would have the potential to give rise to very long-lived volcanism. Further zones of intraplate volcanism are possibly represented by the Christmas Ridge–Moonless Mountains and Kodiak–Bowie chains. The former may mark a poorly developed stress trajectory related to plate edge effects which are not coincident with ridge transforms, hence the minor amount of volcanism associated with these intraplate features. Conversely, the proximity of the Kodiak–Bowie chains to the Juan de Fuca ridge and their isotopic similarity to seamounts (Hegner and Tatsumoto, 1989) suggests an origin related to axial ridge processes.

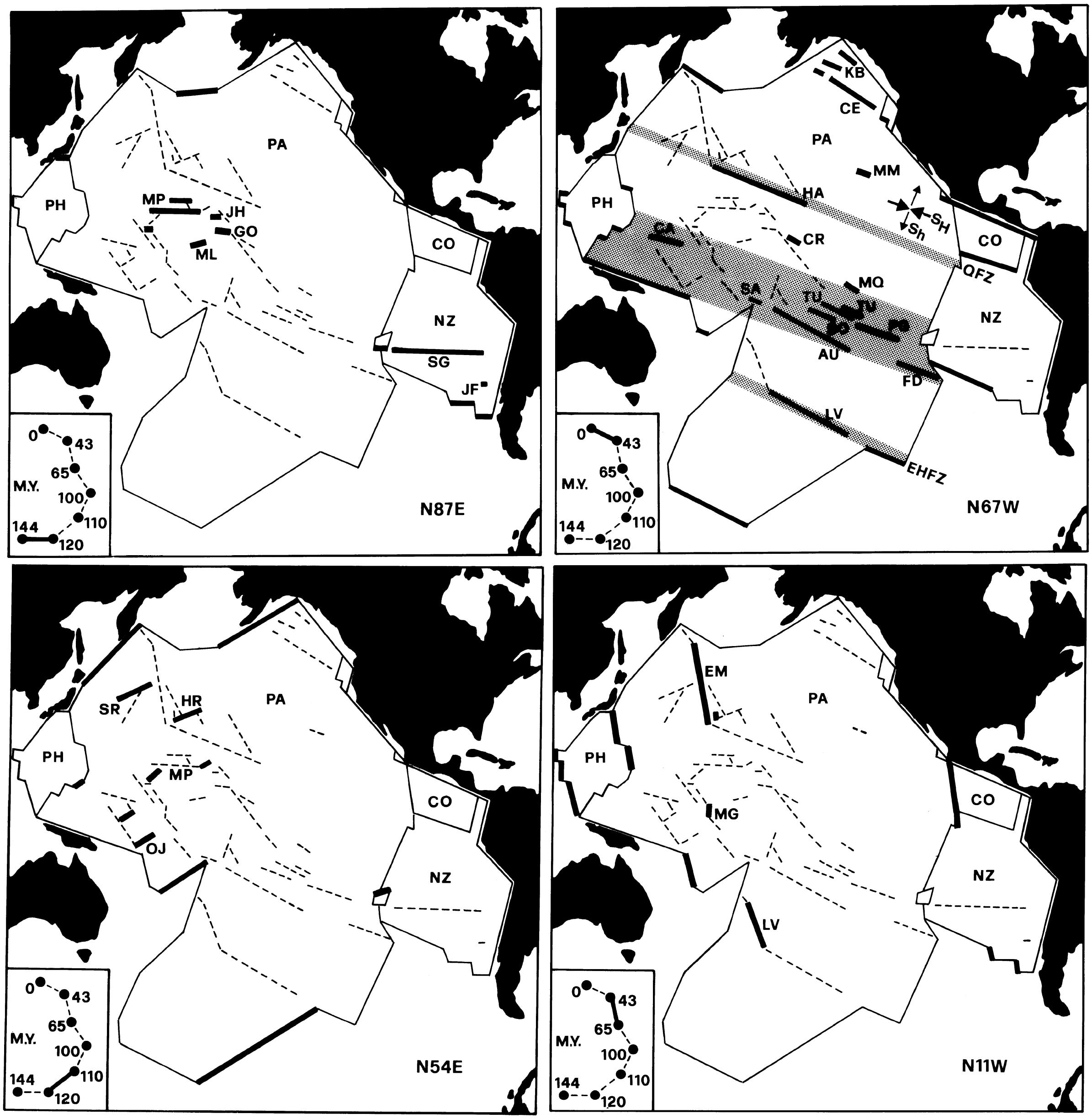
If boundary forces act in the same direction as eastward mantle flow, as in the example of the Nazca plate, velocities in the asthenosphere will be enhanced. Flow will resemble the Couette profile depicted for plate movement in the hotspot reference frame, but shear stresses and hence the potential for generating intraplate volcanism are reduced (Fig. 4). However, intraplate chains may be produced from incipient ''plate tearing'' effects related to the subduction geometry of different segments of the subducting slab (Anderson, 1998a). The most prominent intraplate feature on the Nazca plate, the Sala y Gomez chain, may be explained by this mechanism as may volcanism in the Galapagos Islands (Fig. 13b).

3.7. Mantle domains and the evolving pattern of intraplate volcanism in the paleo-Pacific basin

The history of intraplate volcanism in the Pacific basin can be used as an example to illustrate how the availability of suitable source material coupled with prevailing stress fields may control the occurrence of intraplate volcanism within long-lived ocean basins. Allowing for westward migration of the lithosphere, it can be demonstrated that mantle which now underlies the south-central Pacific was overlain by Gondwana in the mid Paleozoic (Fig. 15). Breakup of this continent involved detachment of several continental blocks including North China, South China, and Indochina which subsequently collided to form Asia (e.g., Coney, 1990). At least one of these blocks (North China) lost more than 120 km of lithospheric mantle between the Ordovician and Cenozoic (Menzies et al., 1993). By analogy with the model of Mahoney et al. (1989) for erosion of continental mantle from under India in the Cretaceous, it is envisaged that the lithospheric thinning created a South Pacific asthenospheric domain contaminated with Gondwanan thermal boundary layer material. The Mississippian ocean islands of the Cache Creek Group (Smith and Lambert, 1995), subsequently accreted to western North America, possibly represent the earliest preserved ocean islands derived from this domain. Paleontological evidence indicates the Cache Creek OIB originated close to eastern Gondwana (Monger, 1977), such that collision of the terrane with western North America around 225 Ma requires migration across the paleo-Pacific basin at an average rate of 15 cm yr^{-1}. A Late Paleozoic plate configuration with a large plate (Farallon) consumed at a convergent margin in the east of the basin, and ridge system in the western Pacific, would allow migration of the terranes across the Pacific basin. The high velocities would be consistent with a plate attached to an eastward-dipping slab where slab pull reinforces mantle drag (Smith and Lewis, 1999b). The paucity of accreted OIB in the Cordilleran orogen may reflect their subduction back into the mantle, but as the Farallon plate would be expected to be associated with a comparable asthenospheric flow profile to the modern Nazca plate, it is also possible that the asthenospheric regime was not conducive to the generation of intraplate volcanism and the eastern Pacific in the Late Paleozoic and Early Mesozoic had few ocean island chains.

A second phase of continental mantle loss, though involving mechanical boundary layer material, is postulated to have occurred by delamination during the formation of Asia in the Permo-Triassic. Formation of Asia would have occurred over mantle which now underlies Hawaii (Fig. 15b) creating a distinct North Pacific domain. Development of a convergent margin along southeast Asia in the Early Mesozoic, would have added a northwesterly slab pull component and altered the basin-wide stress field, favouring re-orientation of ridge systems and their propagation into the central paleo-Pacific. Volatile-bearing minerals would not survive entrainment into the upwelling associated with ocean ridge systems where their low melting point would be expected to lead to voluminous melting (Fig. 3). Intersection of ridge systems with enriched domains thus provides a mechanism for enhanced magma production to generate oceanic plateaus while at the same time causing depletion and fragmentation of the domains. Hence intersection of the Izanagi–Farallon ridge system with the southern Pacific domain gives rise to flood basalts of the Karmutsen Formation on the Insular terrane in the Permo-Triassic (Fig. 15b) while tapping of the southern domain to central Pacific by ridges produced in relation to major plate reorganisations offers a non-superplume explanation for the phase of major ocean plateau development in the mid Cretaceous. The Izanagi plate, in contrast to the Farallon, would have been associated with a counterflow regime, which in conjunction with the production of oceanic plateaus along western Pacific ridge systems, offers an alternative explanation to plume model of Kimura et al. (1994) for the generation of intraplate terranes accreted to Asia in the Mesozoic.

The counter-clockwise rotation of the stress field in the Cretaceous (Fig. 14) is attributed to the opening of the Atlantic and the westward motion of the



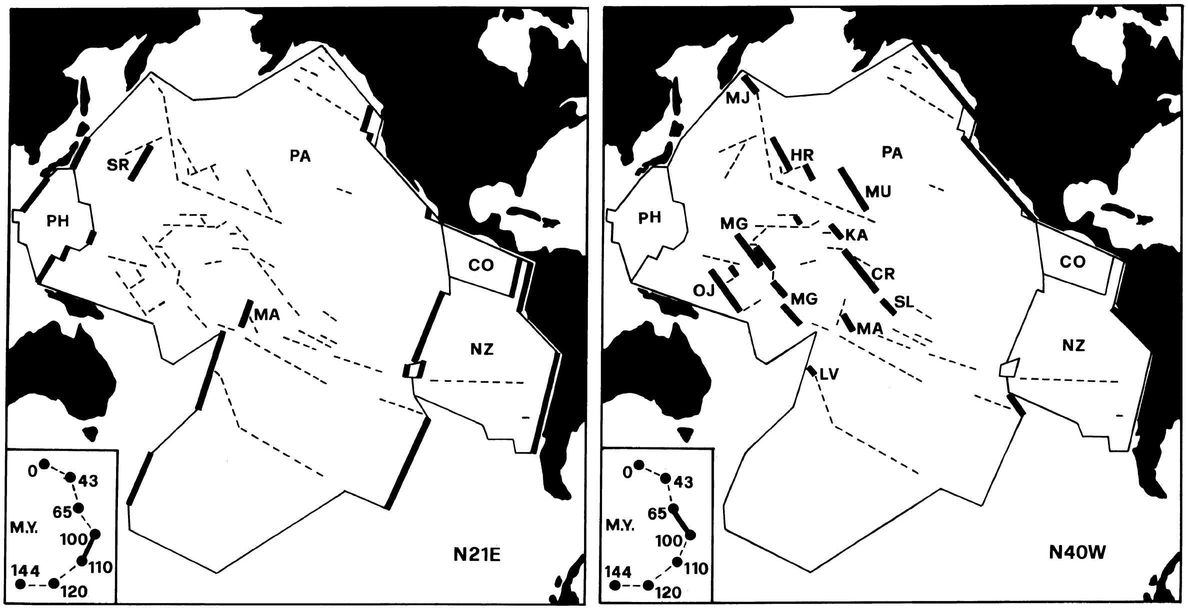


Fig. 14. Polygon model illustrating the six principal structural trends shown by plate boundaries and intraplate volcanic lines in the Pacific basin. Eocene–Recent intraplate volcanism on the Pacific plate (N67°W trend; top right) lies within three principal zones (shaded) representing stress trajectories where the average maximum horizontal shear stress (s_H) has been enhanced by plate interactions along the margins of the basin. Volcanism is interpreted to occur where the zones intersect eastward-migrating domains of asthenosphere contaminated with eroded/delaminated continental mantle (see also Figs. 13b and 15d). Most intraplate volcanic lines lie in the central zone where plate boundaries are most disrupted by the Earth's equatorial bulge. Changes in the orientation of the stress trajectories result from changing interactions along the margins of plates, producing stress wander curves as shown inset at bottom left for the Pacific plate. Plates: CO Cocos, NZ Nazca, PA Pacific, PH Philippine. Ocean islands and plateaus (numbers are ages (Ma) from Duncan and Clague (1985), Abrams et al. (1993), and references therein): AU Austral–Cook (0–29), CA Caroline (1–14), CE Cobb–Eickelberg (2–26), CR Christmas Ridge (33, 78–93), EM Emperor (43–65), FD Foundation (1–22), GO Goldsborough (128), HA Hawaii (0–43), HR Hess Rise (87–92, 115), JF Juan Fernandez (< 30), JH Johnston (128), KA Karin (83–93), KB Kodiak–Bowie (0–25), LV Louisville (0–65), MA Manahiki (110–123), MG Marshal–Gilbert (55–138), MJ Meiji (65–80), ML Magellan (100–135), MM Moonless Mountains (< 43), MP Mid Pacific (83–128), MQ Marqueses (1–9), MU Musicians (64–98), OJ Ontong Java (81–127), PG Pitcairn–Gambier (1–25), SA Samoa (5–14), SG Sala y Gomez ridge/Easter seamounts (0–26), SL South Line (69–73), SO Society (0–5), SR Shatsky Rise (98–120), TU Tuamotu (< 43). Fracture zones: EHFZ Eltanin–Heezen, QFZ Quebrada. Plate configuration based on earthquake epicentre data compiled by Tarr (1974) and Simkin et al. (1989). Mercator projection.

Americas. The brief clockwise rotation of the stress field between 65 and 43 Ma corresponds to a lengthening of convergent margins in the western Pacific as part of the "Eocene event" (Rona and Richardson, 1978) resulting from overriding of the Kula–Farallon ridge by North America and plate reorganisations in the Indian Ocean. Counter-clockwise rotation of the Pacific stress field resumed after the Eocene with the formation of new plate boundaries in the eastern Pacific. Modern South Pacific OIB have isotopic compositions characterised mainly by HIMU-EM1 components, though the occurrence of EM2 signatures in OIB from the southeastern Pacific basin (Juan Fernandez and San Felix) implies the domain may still compositionally mimic subduction events along the eastern margin of Gondwana in the Early Paleozoic. The mixing components for Hawaiian basalts are EM1 and DM (Fig. 6). The mixing trend passes through Bulk Earth (BE) and high $^{3}He/^{4}He$ ratios have been interpreted to reflect an abundance of the primordial isotope ^{3}He hence indicating a primitive source. However, BE is not a mixing component in Pb–Pb space. Similarities with the isotopic signatures of Cenozoic intraplate volcanism in Asia (Fig. 6) in conjunction with interpretations of high $^{3}He/^{4}He$ as a result of ^{4}He deficiency

(Anderson, 1998c) would be entirely consistent with an origin for Hawaiian basalts from delaminated Asian mechanical boundary layer continental mantle.

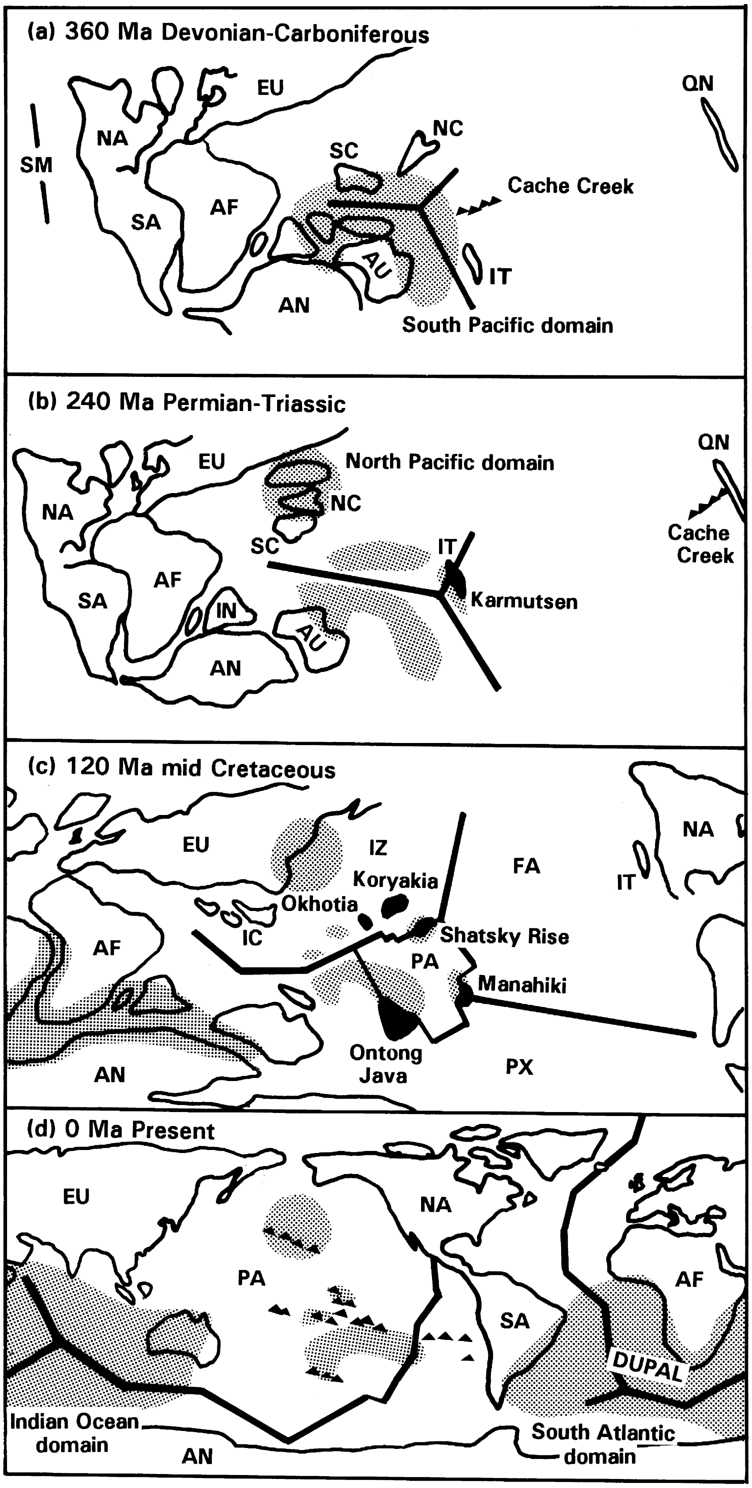
Further large-scale contamination of asthenosphere with Gondwanan continental mantle is envisaged to have occurred as the South Atlantic and Indian Oceans started to open (Fig. 15c), the distribution of such material now corresponding to the equatorial DUPAL belt of Hart (1984) (Fig. 15d). After Hawaii, the next highest $^{3}He/^{4}He$ signatures are found in Réunion lavas, where again the signatures can be ascribed to mechanical boundary layer material in consideration of evidence for extensive continental mantle loss accompanying the northward migration of India. Lastly, the differential rotation model predicts replacement of the South Pacific domain by Indian Ocean asthenosphere as appears to be happening from the discovery (Crawford et al., 1995; Hickey-Vargas et al., 1995) of Indian Ocean isotopic signatures in basalts from the western Pacific.

3.8. Marble-cake mantle as the fate of subducted crust

Attributing intraplate volcanism to the demise of once thicker, continental mantle roots, removes any requirement to isolate subducted crust into a plume source, thereby allowing re-assessment of the marble cake model for subducted crust as a buffer on the composition of the MORB-source. The most suitable isotopic system for testing this model is the Lu–Hf system by virtue of the fractionation of Lu from Hf in the sedimentary system. The necessary data (Patchett et al., 1984) for testing the model became available only two years after Ringwood (1982) had suggested that subducted oceanic crust would not attain sufficient buoyancy to generate plumes and would become remixed with the overlying mantle. However, the control on geodynamic thinking by the plume model was such that the alternative was never tested, and the field of chemical geodynamics progressed into the quagmire of paradoxes outlined in Section 2.7. Yet the supposed decoupling of Hf from Nd isotopes is exactly what would be expected from remixing crust into the mantle. Sm/Nd ratios are not severely fractionated during sedimentary processes, whereas concentration of Hf into zircon results in turbidite having very low Lu/Hf, while pelagic sediment lacks zircon and has high Lu/Hf. Both sediment types will therefore evolve toward low $^{143}Nd/^{144}Nd$, but while turbidite will maintain low $^{176}Hf/^{177}Hf$, pelagic sediment will evolve to high $^{176}Hf/^{177}Hf$ such that mantle contaminated with such material would attain a wide range in εHf but narrow range in εNd. The other barrier to testing alternative models has been a belief that trace element evidence precludes recycling crust into the depleted mantle (e.g., Hofmann et al., 1986; Sun and McDonough, 1989). It is therefore important to emphasise that in this model, depleted mantle is the product, not a starting component. In the mixing models presented in Fig. 7b,c crust is mixed with a very depleted mantle (VDM) reservoir which is more depleted than the depleted mantle in order to create the depleted mantle. For recent tectonic regimes, the VDM reservoir may be equated with a residue from melt extraction at ocean ridges and/or the peridotite portion of the subducting slab, for earlier times in Earth history it could also be equated with a residue from a melting event which generated continental crust from a primitive reservoir.

The calculations in Fig. 7b,c are the complement to the plume model of Patchett et al. (1984) shown in Fig. 7a, and to facilitate comparison the same sediment starting compositions have been used. How-

Fig. 15. Positions of mantle domains (shaded) produced by erosion/delamination of continental mantle into the asthenosphere during continental rifting/collision events relative to westward lag of the lithosphere through the Phanerozoic. Plate reconstructions are based on the paleogeographic maps of Coney (1990) and terrane accretion models of Debiche et al. (1987) and Zonenshain et al. (1987). The model demonstrates the suitability of differential rotation as a mechanism for the lateral introduction of geochemically enriched material into the oceanic asthenosphere to serve as the source of intraplate volcanism. Migration of the domains corresponds to an average rate of 4 cm yr^{-1} reflecting the reduction in the rotation differential with latitude (Fig. 1b) and the location of hydrous minerals in the flow profiles depicted in Fig. 4. Intersection of a ridge system is envisaged to cause depletion of a domain by formation of an oceanic plateau (solid pattern) such that the domains become fragmented over time. Abbreviations, continents and terranes: AF Africa, AN Antarctica, AU Australia, EU Eurasia, IC Indochina, IN India, NA North America, NC North China, SA South America, SC South China, IT Insular terrane; oceanic plates: FA Farallon, IZ Izanagi, PA Pacific, PX Phoenix.


(a) 360 Ma Devonian-Carboniferous
EU
QN
NA
SM
NC
SC
Cache Creek
AF
SA
AU
IT
AN
South Pacific domain
(b) 240 Ma Permian-Triassic
EU
North Pacific domain
QN
NA
NC
Cache Creek
SC
IT
AF
SA
IN
Karmutsen
AU
AN
(c) 120 Ma mid Cretaceous
NA
EU
IZ
FA
Koryakia
IT
Okhotia
AF
IC
Shatsky Rise
PA
Manahiki
AN
Ontong Java
PX
(d) 0 Ma Present
EU
NA
PA
AF
SA
DUPAL
Indian Ocean domain
AN
South Atlantic domain

ever, while the plume model extrapolates compositions into the future, the crustal recycling models are illustrated between 1 Ga and the present, the time period being intermediate between mixing estimates of 240–960 Myr (Kellogg and Turcotte, 1990) and 1280 Myr (Allègre and Turcotte, 1986). The models demonstrate that the high $^{176}Hf/^{177}Hf$-low $^{143}Nd/^{144}Nd$ part of the MORB array can be explained by mixtures of subducted basalt and pelagic sediment, whereas the low-$^{176}Hf/^{177}Hf$-high-$^{143}Nd/^{144}Nd$ part of the array corresponds to buffering by mixtures of subducted basalt and turbidite. The curvature of the mixing lines is controlled by the Nd/Hf ratios of the sediment endmembers, whereas mixing proportions are determined by the age of the sediment and the concentrations of Nd, Hf in the VDM component. Concentrations in the latter are constrained to be less than in the depleted mantle, which must also be replicated in the model. The fit to the shape of the array can be improved by increasing the number of mixing stages or by allowing heterogeneity to evolve in the VDM reservoir as for simplicity, the VDM component has been depicted as having a single composition. In the sediment-only recycling model, depleted mantle compositions can be produced with 0.05 to 0.15% pelagic sediment or 0.3 to 0.7% turbidite in the mantle. For the marble-cake model, the amounts of sediment correspond to 2–5% of the volume of the upper mantle when sediment comprises 1–7% of subducted crust (equivalent to 100–350 m of pelagic material or up to 900 m of turbidite on a basaltic crust 7 km thick). The amount of sediment corresponds to the flux of material subducted through the Phanerozoic. In these proportions, mixing produces not only the isotopic features of the MORB Nd–Hf isotopic array, but also arrests the mantle evolution rate to 2.2 εNd units G yr^{-1} as observed in Fig. 8.

3.9. Correlation of the geoid with plate configuration and mantle convection

Remixing subducted crust into the mantle to form the MORB source allows re-interpretation of the Earth's geoid anomalies in terms of mantle convection and lithospheric effects. The residual geoid anomalies in any event, constitute only a second order effect as they represent only a departure of the Earth's shape from an oblate spheroid. The first order effect is the departure from a perfect sphere manifest as polar flattening and the equatorial bulge. As the first order effect is caused by Earth rotation, it is logical to also examine possible effects on the second order geoid. Early studies (e.g., Chase, 1979) concluded there is no relationship between geoid and plates on an individual basis; however, relationships between plate groupings were not considered. Plotting the magnitude of angular momentum vs. percentage plate surface area for each plate yields an asymptotic division into continental and oceanic plates which yields a two megaplate configuration for the Earth (Fig. 5c) (Lewis and Smith, 1998). Megaplate M1 is mostly oceanic consisting of the Pacific, Philippine, Cocos, Nazca, Indian, and China plates, while megaplate M2 is mostly continental consisting of Europe, America, Africa and Antarctic plates. M1 and M2 comprise 46% and 54%, respectively, of the Earth's surface area. Alternatively, the China plate can be considered part of M1, separate from the Europe plate on account of rifting in the Lake Baikal region, to give a configuration where both M1 and M2 are closer to 50% in surface area. The two major geoid anomalies (closed highs; Fig. 5c) are restricted to the plates involved in each of the two megaplate distributions: the Pacific geoid anomaly lies within M1, whereas the African geoid high lies within the plates comprising M2. The Earth's two largest positive antipodal equatorial geoid anomalies are directly coincident with the poles of centre of gravity axis of the two megaplates and also coincide with the poles of one of the longitudinal principal moments of inertia axes. Likewise, poles of the other longitudinal principal inertial axis coincide with the maximum closure of the South American positive geoid anomaly and the minimum negative closure of the India geoid anomaly.

There are several possible explanations for the M1, M2-geoid relationship. The Earth, as it attempts to reach rotational stability may redistribute masses about the poles of its three instantaneous inertia axes. The masses may be lithospheric, distributed via the process of plate tectonics. Alternatively, the correlation does not preclude, and would be consistent with the geoid anomalies reflecting a mantle mass distribution with an arrangement of large-scale mantle convection cells centred about the inertia axes

(Fig. 2c). The three-dimensional form of large-scale upwelling is likely to be characterised by a central upwelling column with downwelling around the flanks (Bercovici et al., 1989). Such structures have been loosely referred to as megaplumes (e.g., Cadek et al., 1995), but it is important to make a distinction. Not only is there a potential for confusion in dimensions with the superplume concept, geochemically the plume model implies a recycled or primitive mantle component, while the upwelling considered here is nothing more than a representation of large-scale mantle convection which differs from the surrounding mantle only in thermal profile. Several previous studies have suggested a bipolarity in such upwelling (e.g., Busse, 1983; Pavoni, 1991), and the location of such three-dimensional structures represents one possible explanation for the M1–M2 correlation. A modification to the model is that spreading centres would not be likely to be positioned above the upwelling, but around the margins of upwelling regions thereby accounting for the positions of the plates. In this respect, it is significant to note that M1 which is associated with the largest geoid anomaly is bounded by ocean ridges in the Indian and Pacific oceans and the rifting in the Lake Baikal area. The Mid Atlantic ridge lies within M2; however; it is possible that there is more than one mechanism at work. Anderson (1982b; 1994a; 1996) has emphasised the roles slab cooling and plate insulation in controlling the thermal structure of the mantle. The location of M1 corresponds to mantle which has not been cooled by subduction, while the M2 anomaly can be interpreted as the consequence of lithospheric insulation under Pangea inducing hotcell conditions in the mantle. Thermal anomalies created by such means would take several tens of millions of years to dissipate (Anderson, 1982b); the present position of the Mid Atlantic ridge bisecting M2 may be part of that process.

3.10. Large continents and very depleted mantle in the Archean

If the depleted mantle is buffered by subducted crust, the question arises as to how long such a process may have operated throughout Earth history as the present rate of Nd isotopic evolution of the MORB-source may only be traced to the Early Proterozoic (Fig. 8). Isotopic compositions as calculated by Smith and Ludden (1989) indicate a far greater relative range of variation in the Archean, with komatiites generally displaying higher εNd than tholeiites. Important isotopic differences also exist between komatiite types with the source region of ADK appearing to evolve at a faster rate than the source of AUK (Fig. 8). Concepts of high εNd values in the Archean mantle (Smith and Ludden, 1989; Bennett et al., 1993) were challenged by Lahaye et al. (1995) who suggested the variation in Nd isotopic composition reflected the susceptibility of the low trace element contents of komatiites to alteration. The basis of the isotopic arguments made by these authors was a similarity of εNd values calculated for clinopyroxene to the mantle evolution curves of DePaolo (1981) and Goldstein et al. (1984). However, it should be noted that the results of Lahaye et al. (1995) were adjusted relative to a very low $^{143}Nd/^{144}Nd$ value of 0.511822 for the La Jolla standard. Recalculation of the ratios to more commonly quoted ratio of 0.511855 for this standard would place the clinopyroxene separates above the standard mantle evolution lines and within the fields depicted for komatiites in Fig. 8. The high εNd values shown by komatiite samples are thus considered to be a feature inherited from their source.

High εNd values could potentially result from isotopic evolution in an early layered remnant from crystallisation of a magma ocean(s) in the Earth's mantle as discussed by Smith and Ludden (1989); however, high mantle temperatures in the Archean would imply more vigorous convection making the survival of any layered structure unlikely. The evolution rate of 4.4 εNd units per Gyr inferred for the ADK-source is double that of the Phanerozoic MORB-source and corresponds to the rate calculated for unbuffered mantle (note parallel lines in Fig. 8). The preferred interpretation is that the Nd data records the operation of a regime such as obduction tectonics (e.g., de Wit and Hart, 1993) or slab melting in subduction zones (e.g., Martin, 1986, 1993; McCulloch, 1993) which would limit the efficiency of recycling crust back into the mantle. Once formed, continental crust could therefore not be reduced in volume until the thermal regime of the mantle had declined, implying the existence of a large crust early in the Earth's history. The paucity of ancient model ages measured from the present crust can be attributed to intense impact cratering and accelerated tectonic cycles resulting from elevated thermal regimes (Warren, 1989) causing extensive reworking and mixing of the ancient crust with younger material. Upper limits to crustal volume can be calculated by assuming that the volume of crust required to buffer the depleted mantle in Fig. 7b,c resided at the Earth's surface before thermal gradients declined sufficiently to permit its recycling into the mantle. The volume of crust in the sediment-only recycling model of Fig. 7b corresponds to 20% of the volume of the present day continental crust, whereas the volume in the marble-cake model of Fig. 7c corresponds to 8% of the present day crust. The marble-cake scenario is preferred such that the volume of crust is considered to have been comparable to that which exists today, similar to in the models of Armstrong (1968; 1991). However, the model of Armstrong (1991) involves a steady state recycling regime with constant crustal volume and a depleted mantle reservoir which maintains high εNd throughout the Precambrian (Figs. 2d and 10), with the implication that the absence of komatiites after the Archean is a consequence of source isolation, or a change in thermal regime of the most depleted regions of the mantle. In the model advocated in this study, the volume of crust decreases towards the present because crustal recycling becomes more efficient. The occurrence of komatiites in Archean becomes a consequence of compositional variation in the mantle, with the very depleted mantle (unbuffered) reservoir only being capable of producing komatiite. The decline in apparent growth rates for Nd isotopic composition of the AUK-source (Fig. 8) suggests the very depleted reservoir was largely erased by the late Archean such that the isotopic composition of the mantle describes a sigmoidal rather than smooth curve.

The tectonic regime at convergent margins may have been not only a consequence of, but also a contributing factor to the thermal regime in the Archean mantle. Deep subduction of slabs is one of the principal mechanisms for cooling the modern mantle (Anderson, 1994a). In Archean tectonic regimes such as envisaged by McCulloch (1993) where slabs are fragmented at depths shallower than 300 km, the deep mantle is deprived of one of its two principal cooling mechanisms. Continental lithosphere may also insulate the mantle (Gurnis, 1988; Anderson, 1994a), such that the presence of a large continental crust (Fig. 2d) would assist in maintaining high temperatures in the mantle. The temperatures for komatiite genesis along ocean ridges estimated by Takahashi (1990) are only 80°C higher than the estimates of Richter (1988) for average Archean mantle. Temperature elevations of such magnitude well within the +200°C attainable under hotcell conditions alone in modern tectonic regimes (Anderson, 1982a, 1994a). More recent interpretations of komatiite textures have pointed toward a significant volatile content with the melts forming at temperatures as little as 100°C above present mantle temperatures (Parman et al., 1997; Stone et al., 1997). A low mantle temperature and large crust would also readily explain cool continental geotherms indicated by crustal assemblages, which are problematic in models where the continental crust is very small (Takahashi, 1990). The appearance of alkaline intraplate volcanism may also mark the decline in mantle thermal regime. Phlogopite-bearing assemblages can exist in the shallow mantle at temperatures up to 1350°C (Sato et al., 1997). The modern average oceanic geotherm away from ridges is approximately 125°C less than the ridge geotherm at a depth of 100 km. On the suggested thermal evolution curve in Fig. 9, the average mantle temperature would then exceed the stability field for phlogopite peridotite until around 2.0 Ga. Only after this time could hydrous peridotite assemblages survive to be tapped in stress field models, hence the appearance of OIB in the Circum–Ungava belt at 2.0 Ga.

4. Concluding remarks

The rise of the plume model can be attributed to the initial simplicity of the model and the timing of its introduction, following on the heels of the inception of plate tectonics which fomented a time of testing of many ideas. The number of variations found to be required on further evaluation, rather than being a cause for concern as under normal circumstances, may through the shear number of plume models, have given rise to a impression of the model as being the preferred option. The potential of hotspots as a reference frame further insured consideration of the plume model. However, the progression of the model, characterised by a focus on exploring new variations to the neglect of potentially fatal flaws, may have worked against the objectives of its proponents as demonstration of superiority over all possible alternatives is the only justification for a concept to achieve ruling theory status. What should not be forgotten is that when a concept lacks direct observation, it is inevitable that a certain amount of circularity in argument based on the assumption that the phenomenon exists will be used to define its features. Such logic, nor any amount of supposed correlation between the phenomenon and observations used in defining it, does not constitute a scientific proof. With regard to what is now known as ‘‘plume theory’’, it should be noted that constraints put on temperatures, depth of origin, size, composition and degree of melting of plumes have all been made under the assumption that plumes exist and consequently, do not constitute a test of, nor confirm the validity of the plume model. That many observations appear to be consistent with the plume model, merely reflects the logic used in the construction of plume model. Consideration of alternatives, of course, involves the risk that the alternative will provide a more elegant synthesis than the model being tested, and with the circularity of argument that has been used in construction of the plume model, a singular flaw has the potential to prove fatal to the entire concept. Had such alternatives been given the same consideration as the plume model was given when it was first introduced, a less complex picture of the Earth could potentially have been developed in the 1980s from the following concepts:

(1) Volatile-rich sources (‘‘wetspots’’): A major role for volatile-bearing minerals such as amphibole and phologopite in the sources of intraplate volcanism, thereby removing the need for large thermal anomalies in the generation of intraplate melts.

(2) Eastward mantle flow: Differential rotation of lithosphere and mantle resulting from the transmission of stress through the asthenosphere exerts a drag on the base of continental plates which leads to continental rifting, and provides a mechanism for the lateral introduction and tapping of source material for intraplate volcanism into the oceanic asthenosphere.

(3) Stress fields: Drag and plate boundary forces acting in opposing directions on oceanic plates, set up a counterflow regime in the asthenosphere which functions as the horizontal stress trajectory of Shaw (1973) and causes melting anomalies generated by shearing to appear stationary. A lithospheric field, equivalent to the vertical stress trajectory of Shaw (1973) is imposed by transmission of stress from convergent margins and is responsible for focusing melts into linear chains.

(4) Marble cake mantle: Re-mixing of subducted crust into the depleted mantle MORB-source as the fate of subducted oceanic crust in the Proterozoic and Phanerozoic.

Within this model, the roles of plate boundary and mantle drag forces acting on continental plates can be resolved. There is only one category of intraplate volcanism which, arising from a common cause (large-scale plate interactions) with a common source material (eroded continental mantle), becomes part of the plate tectonic evolution of the Earth. There are no MORB paradoxes as the simplest option of remixing subducted crust into the mantle can be followed. The Earth’s evolution becomes dominated by its cooling, with major changes in tectonic style and petrological association marking changes in the composition of recycled lithosphere, not the development of plume-sources. The plume model, in contrast, is beset by internal contradiction and superimposes only a partial explanation for intraplate volcanism onto plate tectonics. Faced with a comprehensive and simpler alternative, the partial nature of the explanation given for intraplate volcanism becomes the plume model’s most significant flaw. That the plume model grew to the complexity it currently exhibits with an almost total lack of criticism should make it remarkable among scientific endeavours of this century. Every option, except the possibility that it was fundamentally wrong all along, seems to have been explored. Wegener, Jeffreys and their contemporaries may have had the will but not the data to solve the debate over continental drift. During the rise of the plume model, it seems the data but not the will, existed to properly evaluate the problem at hand, resulting in a quarter century of attempts to combine plate tectonics with a concept that, in effect, controlled it. The ‘‘babies thrown out with the bathwater’’ were the alternative concepts and it is ironic that even within the plume model today, such concepts are still required to account for all the manifestations of intraplate volcanism that the ruling theory cannot explain.

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