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Diapirism of depleted peridotite – a model for the origin of hot spots *

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A model is proposed for the origin of hot spots that depends on the existence of major-element heterogeneities in the mantle. Generation of basaltic crust at spreading centers produces a layer of residual peridotite ~20–25 km thick directly beneath the crust which is depleted in Fe/Mg, TiO_2, CaO, Al_2O_3, Na_2O and K_2O, and which has a slightly lower density than undepleted peridotite beneath it. Upon recycling of this depleted peridotite back into the deep mantle at subduction zones, it becomes gravitationally unstable, and tends to rise as diapirs through undepleted peridotite. For a density contrast of 0.05 g cm^{-3}, a diapir 60 km in diameter would rise at roughly 8 cm y^{-1}, and could transport enough heat to the base of the lithosphere to cause melting and volcanism at the surface. Hot spots are thus viewed as a passive consequence of mantle convection and fractionation at spreading centers rather than a plate-driving force.

It is suggested that depleted diapirs exist with varying amounts of depletion, diameters, upward velocities and source volumes. Such variations could explain the occurrence of hot spots with widely varying lifetimes and rates of lava production. For highly depleted diapirs with very low Fe/Mg, the diapir would act as a heat source and the asthenosphere and lower lithosphere drifting across the diapir would serve as the source region of magmas erupted at the surface. For mildly depleted diapirs with Fe/Mg only slightly less than in normal undepleted mantle, the diapir could provide not only the source of heat but also most or all of the source material for the erupted magmas. The model is consistent with isotopic data that require two separate and ancient source regions for mid-ocean ridge and oceanic island basalts. The source for mid-ocean ridge basalts is considered to be material upwelling at spreading centers from the deep mantle. This material forms the oceanic lithosphere. Oceanic island basalts are considered to be derived from varying mixtures of sublithospheric and lower lithospheric material and the rising diapir itself.

1. Introduction

In this paper, a model is proposed for the origin of hot spots or melting anomalies such as Hawaii and Iceland, whose origin was first considered in a modern context by Wilson (1963). A brief explanation of the model will be given first, and this will be followed by more detailed considerations of specific aspects. It is emphasized at the outset that the ideas presented here are speculative and the calculations supporting the model are rough. Before proceeding, several terms require definition. "Lithosphere" will be used in a petrologic sense, to refer to crustal and upper mantle material at subsolidus temperatures extending to a depth of ~85 km beneath the ocean basins. "Asthenosphere" will be used for sub-oceanic mantle material between ~85 and 220 km depth, and this material will be considered to be partially melted (Presnall, 1980). "Depleted peridotite" will refer to peridotite that has lost some or all of its major-ele-

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ment basaltic constituents. "Undepleted peridotite" will refer to peridotite that retains essentially all of its major element basaltic constituents. Oceanic basalts are generally considered to be derived from mantle material that was depleted in incompatible elements early in the Earth's history and hence has altered values of key ratios such as Rb/Sr, Sm/Nd and Lu/Hf relative to bulk Earth (chondritic) values. This depletion has been linked to the formation of enriched continental material (DePaolo and Wasserburg, 1976, 1979a; Jacobsen and Wasserburg, 1979). Because this type of depletion does not necessarily imply more than minor depletion in major elements, we will use the terms "depleted" and "undepleted" peridotite to refer only to major elements. For example, mantle peridotite that is the source of basaltic magmas at spreading centers will be referred to as undepleted even though it is reduced, for example, in Rb/Sr relative to the bulk Earth.

A fundamental requirement of the model is that major-element heterogeneities exist in the mantle. It is suggested that these heterogeneities are the result of the production of basaltic crust at mid-ocean spreading centers. On the assumption that basalts generated at spreading centers are derived by partial melting of peridotite mantle, the production of oceanic crust would leave behind a residue of peridotite or harzburgite with lower Fe/Mg, TiO_2, CaO, Al_2O_3, Na_2O and K_2O, and a lower density than undepleted peridotite. Thus, depleted peridotite would rest in stable gravitational equilibrium on top of undepleted peridotite. Presnall (1980) has argued that the thickness of the depleted layer in the oceanic lithosphere is $\sim$ 20–25 km, in rough agreement with earlier estimates (Kay et al., 1970; Green and Liebermann, 1976; Oxburgh and Parmentier, 1977). After recycling of this depleted peridotite back into the deep mantle at subduction zones, its lower density would cause it to become gravitationally unstable and rise as diapirs through undepleted mantle. Upon encountering the partially melted asthenosphere, which will be assumed to correspond to the zone of decoupling of the lithosphere from the deeper mantle, the tops of these diapirs would be incorporated into the asthenosphere and be bent over or sheared off by the overlying drifting lithosphere

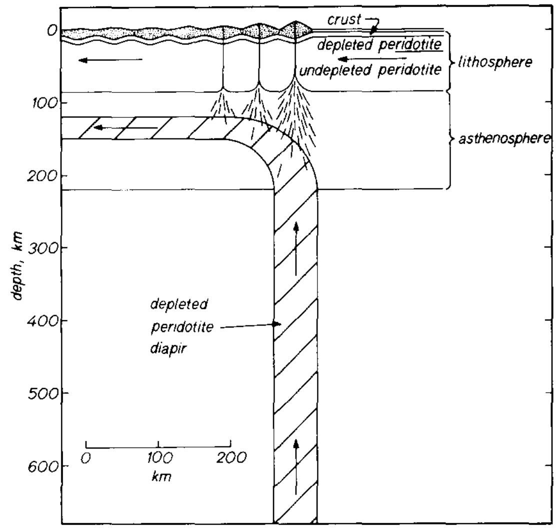

Fig. 1. Diagrammatic representation showing vertical section of oceanic lithosphere drifting across rising depleted diapir to produce a chain of islands. Section taken along the island chain. The diagram illustrates the case of a lithospheric plate velocity greater than the rate of rise of the diapir, resulting in thinning and extension of the diapir after it is bent over. Arrows indicate directions of motion. Elevation of volcanoes not to scale.

(Fig. 1). For a diapir rising at a rate of several centimeters per year, the vertical temperature distribution in the diapir at depths greater than $\sim$ 200 km would approach an adiabatic gradient, and the uppermost part of the diapir would generally be hotter than the surrounding mantle. Thus, the temperature, amount of melting of the asthenosphere, and probability of volcanism at the Earth's surface would be increased. As is discussed in more detail below, it is suggested that diapirs exist with varying amounts of depletion. Those that are only slightly depleted could provide most of the source material for the volcanic rocks, whereas those that are strongly depleted would provide very little of the source material. In the latter case, the asthenosphere and overlying lower lithosphere which moves across the diapir would be the source material, and the diapir would act primarily as a source of heat.

O'Hara (1975) mentioned diapirism of depleted peridotite as a possible explanation for Iceland, and the model proposed here has similarities to his. In this paper, we develop and expand the suggestion of O'Hara and apply the more generalized model to the origin of hot spots worldwide.

2. Density considerations

A critical assumption for our model is that the density of depleted mantle is less than undepleted mantle at any given depth. Various estimates are given in Table I for the density contrast between these materials. It is generally agreed that depleted peridotite is lower in Fe/Mg than undepleted peridotite. This chemical difference would tend to decrease the density of depleted peridotite because the densities of all the major mineral constituents (olivine, orthopyroxene, clinopyroxene, garnet and spinel) decrease with decreasing Fe/Mg. In addition, the most dense phase of peridotite, garnet or spinel, would be among the first to melt and would be preferentially removed during the extraction of basalt. The combined effect of these two factors explains the conclusion reached by all except Shaw and Jackson (1973) that depleted peridotite is slightly less dense than undepleted peridotite (Table I). The opposite conclusion of Shaw and Jackson (1973) was based on measured densities of nodules from Hawaii. They found 11 dunites, which they considered to represent depleted residual mantle, to have a density of 3.27–3.42 g cm^{-3}, whereas 5 garnet lherzolites and one garnet clinopyroxenite, which they considered to represent undepleted mantle material, were found to have densities of 3.14–3.34 g cm^{-3}. They noted that the average dunite was 0.07 g cm^{-3} more dense than the average garnet lherzolite, but the reality of this difference is questionable in view of the large overlap of the two ranges (0.07 g cm^{-3}) and uncertainties (acknowledged by Shaw and Jackson to be as large as 0.13 g cm^{-3}) in the measurements. Also, densities of the Hawaiian dunites are high because they are not depleted in Fe/Mg relative to the garnet lherzolites and pyroxenites. The dunites appear, in fact, to be crystal accumulates from a tholeiitic magma chamber rather than residual material from partial fusion (Sen and Presnall, 1980). In view of these problems and the general agreement of all other investigators, we believe that the density difference determined by Shaw and Jackson is in error and that depleted peridotite is less dense than undepleted peridotite. Because the gravitational anchor model of Shaw and Jackson for the origin of hot spots depends on a higher density for depleted peridotite, it appears that their model fails (Boyd and McCallister, 1976; Jordan, 1979).

TABLE I

Density differences between undepleted and depleted peridotite

Reference	$\rho_{\text{undepleted}} - \rho_{\text{depleted}}$ (g cm^{-3})
Shaw and Jackson (1973)	−0.07
Carter (1970)	≤0.08
O'Hara (1975)	0.08
Boyd and McCallister (1976)	0.09
Green and Liebermann (1976)	0.02–0.05
Oxburgh and Parmentier (1977)	0.06
Jordan (1979)	≤0.06

The density differences summarized in Table I refer to spinel or garnet peridotite phase assemblages stable at depths less than ~330 km (Ringwood, 1975). The difference in density between spinel lherzolite and garnet lherzolite is only 0.015 g cm^{-3} (Oxburgh and Parmentier, 1977), so depleted peridotite would rise with about the same velocity through either material. At greater depths, the magnitude of the density contrast is uncertain, but its sign is unlikely to change because mineral structures enriched in Fe will generally be more dense than the same structures enriched in Mg. This is true, for example, in the case of olivine in the spinel structure (Ringwood and Major, 1970). Thus, at all depths in the mantle, it is probable that a diapir with the bulk composition of depleted peridotite will tend to rise through material with the bulk composition of undepleted peridotite.

3. Ascent velocity of diapirs

In order to determine the capacity of a depleted diapir to transmit heat upward by convection to the base of the lithosphere, it is necessary to estimate the ascent velocity of the diapir. In calculating the ascent velocity, the choice of the viscosity is important. In most calculations, mantle viscosities derived from glacial rebound studies are used. Anderson (1981a, 1981b) has pointed out that these viscosities are weighted toward values appropriate to the mantle beneath continental shields and that viscosities beneath spreading ridges could be several orders of magnitude lower because of increased temperatures and preferred orientation of crystals. He also argued that large deviatoric stresses and anomalously high temperatures adjacent to hot diapirs would further reduce the viscosity. Anderson then substituted these anomalously low viscosities into Stokes' law and obtained very rapid ascent velocities. However, Marsh and Morris (1981) pointed out that Stokes' law can be used to approximate the drag on a diapir only if the surrounding material has a constant viscosity for a distance away from the diapir of at least 10 times the diapir radius. They noted that for the case visualized by Anderson, Stokes' law can give a poor approximation of the ascent velocity because the thickness of the low viscosity material around the diapir is usually much less than the diapir radius (Marsh, 1978, 1981).

Morris (1980), as outlined by Marsh and Morris (1981), analyzed the problem of diapir ascent velocity in a medium of variable viscosity by considering two extreme cases. Both assume a spherical diapir surrounded by a sheath of material with a constant low viscosity η_1, which is in turn surrounded by material whose viscosity η_2 is undisturbed by the diapir. First, for constant η_2/η_1 and a sufficiently large radius of the diapir, the pressure required to transport material from in front of the rising diapir through the low-viscosity sheath greatly exceeds that required to deform the surrounding material of higher viscosity η_2. In this case, the order of magnitude of the ascent velocity can be obtained from Stokes' law using the undisturbed viscosity η_2. Second, if all parameters except viscosity are held constant and η_2/η_1 is made sufficiently large, material in front of the diapir passes around it through the low-viscosity sheath. Because the sheath is thin, Stokes' law cannot be used in this case, as discussed above. Marsh and Morris (1981) pointed out that the terminal velocity of a diapir will be governed by whichever of these two mechanisms yields the faster velocity. Furthermore, they pointed out that increased melting as the diapir approaches the surface will sharply increase the viscosity contrast η_2/η_1. Thus, the second mechanism will dominate near the surface but the first will dominate at depth.

For our calculations, we assume that the ascent velocity of a diapir for most of its journey can be approximated by the first mechanism. Thus, only the undisturbed viscosity far away from the diapir need be considered. Because viscosities based on glacial rebound studies are the most firmly based, these will be used in the calculations while remembering that the calculations for diapirs near or on spreading ridges may require revision because of the possibility of reduced viscosities in these regions. At least for Hawaii and other similar hot spots far from spreading ridges, glacial rebound viscosities will be assumed to be appropriate.

At strain rates appropriate for a rising diapir, a non-Newtonian fluid must be assumed, and for this case, we use an equation derived by B.D. Marsh (personal communication, 1982) and kindly made available to us. Based on the analysis of Acharya et al. (1976), Marsh expressed the total drag on a sphere in a non-Newtonian (power-law) fluid as

$$D = 24\pi m V^n a^{2-n}/2^{n+1}$$

where V = ascent velocity (cm s^{-1}), a = radius of the sphere, and m = apparent viscosity (c.g.s.) of a power law fluid of exponent n (that is where shear stress τ, strain rate $\dot{e}$ and m are related by $\tau = m\dot{e}^n$). By equating the total drag to the buoyancy of the sphere and solving for V, Marsh obtained the equation

$$V = \left[\frac{2^{n+1}\,\Delta\rho g a^{n+1}}{18m}\right]^{1/n} \tag{1}$$

where $\Delta\rho$ = density contrast between the diapir and the surroundings (g cm^{-3}) and g = acceleration due to gravity (cm s^{-2}). Note that for a Newtonian fluid ($n = 1$), this equation reduces to Stokes' law, where $m = \mu$ (viscosity). Equation (1) is for a spherical diapir but is also valid for approximating the ascent velocity of a cylindrical diapir (Marsh, 1981).

Based on glacial rebound studies (Yokokura and Saito, 1978), we will assume that for the mantle $n = 1/3$ and $m = 10^{12}$ (c.g.s.). From Table I, we use a representative density contrast $\Delta\rho$ of 0.05 g cm^{-3}. From these parameters, we calculate that a diapir with a radius of 30 km would rise at a velocity of roughly 8 cm y^{-1}, a velocity quite adequate for convective transport of heat to be effective (Marsh, 1978). The choice of a diapir radius of 30 km applies to Hawaii, and will be discussed in the next section.

Epp (1978) has noted that hot spots in the Pacific Ocean begin at different times, have varying lifetimes, and have widely varying rates of production of volcanic material. Such a pattern would be expected for depleted diapirs that originate from source layers of varying volume and that rise through the mantle at different velocities because of different density contrasts and diameters. When the source layer for a given diapir becomes exhausted, the diapir would decrease in diameter and ascent velocity and finally become ineffective as a heat source, at which time volcanism would cease.

4. Heat and mass balance requirements

In the case of the Hawaiian Islands, it is possible to make a rough comparison between the amount of heat required to produce the observed volcanism and the amount of heat likely to be produced by a depleted diapir impinging on the base of the lithosphere. McDougall (1979) has shown that over the last 28 Ma, the center of volcanism has migrated at a uniform rate of ~9.4 cm y^{-1}. From this migration rate and the volumes of the volcanoes along the island chain calculated by Bargar and Jackson (1974), the average rate of production of lava from Kohala Volcano * to Midway Island is 0.024 km^3 y^{-1}. This calculation includes only material that stands above the sea floor, and to account for intrusion of magma into the crust and depression of the crust due to loading of erupted lavas, the total rate of production of lava will be assumed to be approximately a factor of 2.5 greater (Suyenaga, 1979), or ~0.06 km^3 y^{-1}. This rate is similar to the estimated steady-state lava production rate for Kilauea of 0.1 km^3 y^{-1} (Swanson, 1972). If the source of the lavas is predominantly in the partly melted asthenosphere, the heat supplied would have to increase the percentage of melt in the source region and raise the temperature a small amount, say 50°C (Mysen and Kushiro, 1977). As additional parameters, we take the density of basalt to be 2.77 g cm^{-3}, the heat of fusion of basalt to be 104 cal g^{-1}, the heat capacity and density of lherzolite to be 0.25 cal g^{-1}°C^{-1} and 3.35 g cm^{-3}, respectively, and the increased amount of fusion in the source region to be 20% (Green and Ringwood, 1967). From these parameters, the total amount of heat required is 3×10^{13} kcal y^{-1}, slightly more heat being required to cause melting than to raise the temperature of the source region (five times the volume of lava produced) by 50°C. If it is assumed that the diapir material incorporated into the asthenosphere supplies the required amount of heat by dropping in temperature by ~150°C, about 8×10^{14} g y^{-1} of diapir material would have to be supplied. From eq. 1, it can then be calculated that the diapir would be ~60 km in diameter and would rise at ~8 cm y^{-1} if the density contrast between the diapir and the surrounding mantle is 0.05 g cm^{-3}. This diameter is roughly half the width of the island of Hawaii measured at the ocean floor and therefore appears to be a reasonable size.

We now inquire if it is reasonable, from a mass balance viewpoint, to obtain most or all of the lavas from the asthenosphere and lower undepleted lithosphere without any contribution from the diapir. Again, we shall use Hawaii as an example. The volume of the five volcanoes on the island of Hawaii is 1.13×10^5 km^3 (Bargar and Jackson, 1974). If we multiply this volume by 2.5, as before, to account in an approximate way for depression of the crust and intrusion of magma beneath the surface, the total volume of lava produced is 2.8 ×

* McDougall's linear fit of age versus distance from Kilauea intersects the distance axis at about 90 km for zero age. This is roughly the distance of Kohala from Kilauea. Thus, Hualalai, Mauna Kea, Mauna Loa and Kilauea, all of which are younger than Kohala and would have small "negative ages" on McDougall's plot, are omitted from the calculation. All except Mauna Kea are still active and even Mauna Kea is probably not completely extinct.

10^5 km^3. If this lava represents $\sim 20\%$ of its source region, a volume of mantle material of $\sim 1.4 \times 10^6$ km^3 is required. Because the diameter of the island of Hawaii measured at the ocean floor is ~ 140 km, we will use this same dimension for the width of the melted mantle. Then the amount of mantle material required could be accounted for by a volume of the dimensions 140 km perpendicular to the island chain, 140 km along the island chain, and 70 km deep. Because Leeman et al. (1977, 1980) have argued from rare earth element (REE) data that Hawaiian tholeiites cannot be derived from a spinel lherzolite mantle, the minimum depth to the top of the volume of mantle that is partially melted will be assumed to lie at ~ 65 km where the spinel lherzolite to garnet lherzolite transition is believed to occur (Hales et al., 1970; Asada and Shimamura, 1976; Presnall, 1980). Earthquakes beneath Hawaii have been related to movement of magma, and are observed to a minimum depth of ~ 60 km (Koyanagi and Endo, 1971). These data also suggest that the zone of melting lies at greater depths. Thus, the partially melted source region could extend from ~ 65 to 135 km depth, and it appears that the lower lithosphere and asthenosphere would be a volumetrically adequate source region for the lavas, with no contribution being required from the diapir itself.

In general, diapirs are considered to be intimately involved in the fusion process, and mixing of material between diapirs and the asthenosphere would be expected. Despite such mixing, the already depleted condition of diapirs with low Fe/Mg would prevent them from contributing any significant amount of magma. However, an important aspect of the depleted diapir model is the notion that diapirs exist with a complete range of depletion. If Fe/Mg in the diapir is high (but still slightly lower than the surrounding mantle), the diapir could represent a large proportion or even all of the source material, as in the case of Iceland discussed below.

5. Locations of hot spots and plate driving forces

Morgan (1971, 1972a, 1972b) suggested that hot spots are caused by rising thermal plumes and that upward and lateral convection from these plumes is the driving mechanism for plate tectonics (see also Wilson, 1973; Anderson, 1975). However, there appears to be no special relationship between the locations of hot spots and diverging plate boundaries. Some hot spots such as Iceland are located directly on spreading ridges, others such as St. Helena, Tristan da Cunha and Galapagos are near spreading ridges, and still others such as Hawaii and Macdonald seamount are far removed from ridges. Hawaii, in particular, is one of the most persistent and active hot spots, yet it is totally unrelated geographically to any sea floor spreading axis. These relationships suggest that hot spots do not drive plates. Instead, it is suggested that hot spots are a passive consequence of mantle convection and chemical fractionation at spreading centers.

It is reasonable to suppose that there will exist chemically depleted diapirs in the mantle that are either small in diameter or have only a small density contrast with normal mantle material or both. For such diapirs, the ascent velocity and amount of heat transported upward would be reduced, which would, under normal circumstances, prevent volcanism at the surface. However, if a thermally inefficient diapir happens to be located near a spreading ridge where regional temperatures are higher than normal, and the effective upward velocity of the diapir is increased both by the upwardly convecting mantle surrounding it and by reduced regional viscosities, enough additional heat may be supplied to cause volcanism above the diapir. It is suggested that the common occurrence of hot spots near spreading ridges may be explained in this way, and that the existence of a spreading center causes the formation of nearby hot spot volcanism rather than the converse.

In the case of Iceland, which is currently situated directly on the Mid-Atlantic Ridge, there would be no lithosphere and upper asthenosphere drifting across the diapir that could serve as the source region for the volcanism. The diapir itself would be the source for the erupted magmas and the large rate of lava production on Iceland could be explained by an ascent velocity for the diapir faster than that for material elsewhere beneath the Mid-Atlantic Ridge. Because the upward velocity varies as the fourth power of the diapir radius (eq. 1), only a modest increase in the diameter results in a significantly enhanced upward velocity for a diapir with only a very small amount of major-element depletion. For example, if the diapir is assumed to be 200 km in diameter and rising 1 cm y^{-1} faster than material rising elsewhere beneath the Mid-Atlantic Ridge, it can be calculated from eq. 1 that the difference in density between the diapir and the surrounding mantle would be ~ 0.005 g cm^{-3}, which amounts to a difference in *mg* number, 100 $Mg/(Mg + Fe^{2+})$, of only ~ 0.5 (Carter, 1970; Boyd and McCallister, 1976). Thus, the diapir would be only very slightly depleted and therefore capable of yielding abundant lavas. This end-member of our generalized model is similar to one of O'Hara's (1975) suggested explanations for Iceland. If Anderson's (1981a, 1981b) suggestion of lower mantle viscosities beneath ridges is correct, Iceland could be explained by a diapir with a smaller diameter or a smaller density contrast with the surrounding mantle or both. Thus, the parameters would change somewhat, but the overall concept would not.

6. Abundance of hot spots

The layer of depleted mantle produced at spreading ridges and recycled back into the deep mantle at subduction zones would be expected to create an extensive layer of depleted material in the deep mantle, and one might therefore expect to see a large number of hot spots closely spaced over the Earth. The number of hot spots is debatable but appears to be somewhere between 20 (Morgan, 1972a) and 122 (Burke and Wilson, 1976). If the maximum accounting of Burke and Wilson is used, there are only 12 on the entire Pacific Plate that have been active in the last 10 Ma.

A possible explanation for the rarity of hot spots is as follows. It has been argued earlier that the thickness of the depleted layer subducted into the deep mantle is ~ 20–25 km, and Marsh (1979) has shown that the diameter of a diapir is approximately the same as the thickness of the source layer if the viscosity of the diapir and surroundings are the same. For a diapir with a diameter of 20–25 km, eq. 1 indicates that the ascent velocity would be only ~ 0.01 cm y^{-1} for a density contrast of 0.05 g cm^{-3}. Such a diapir would take on the order of 10 times the age of the Earth to rise from the core–mantle boundary to the base of the lithosphere. Even if such a diapir started its ascent from a shallower depth, such that it could reach the base of the lithosphere in a reasonable amount of time, its ascent velocity would be so slow and its diameter so small that it would not be effective as a heat source. Thus, hot spots are expected to be uncommon. They would occur only when the depleted layer is locally thickened by some process after it is subducted into the mantle.

Now consider what happens to depleted peridotite at a subduction zone. Oxburgh and Parmentier (1977, 1978) advocated a stratified lithosphere nearly identical to that favored here, and they proposed that the layer of depleted peridotite rises diapirically, shortly after subduction. Despite the buoyancy of the depleted peridotite, diapirism of this material in the vicinity of subduction zones would not be expected to produce hot spots. As noted above, such diapirs, if they formed, would have diameters of ~ 20–25 km with upward velocities of only ~ 0.01 cm y^{-1}, assuming no temperature difference between the diapir and the surrounding mantle. A diapir starting from a subduction zone at 500 km depth would take roughly the age of the Earth to ascend to the surface and would therefore be totally ineffective as a convective heat source for a hot spot.

The rate of rise of diapirs near subduction zones would be suppressed by the cooler temperature of the depleted layer in comparison to material in the mantle wedge immediately above the subducting lithosphere. A density contrast of 0.05 g cm^{-3} would be neutralized by a temperature difference of about 400°C (Skinner, 1966). This is roughly half the maximum temperature contrast between subducting slab and surrounding mantle calculated by Schubert et al. (1975) for a slab penetration depth of 700 km (8 cm y^{-1} plate velocity, 45° slab inclination). Because the greatest temperature contrast at this depth is at ~ 30 km from the top of the slab, the average temperature of the depleted layer would be ~ 300–400°C less than the surrounding mantle. Thus, the buoyancy of the depleted layer would be almost completely neutralized by this effect (see also Oxburgh and Parmentier, 1977), and the ascent velocity of 0.01 cm y^{-1} calculated above should be considered a maximum value.

Because the descent velocity of a subducting lithospheric slab would be two to three orders of magnitude greater (assuming plate velocities of 1–10 cm y^{-1}) than the ascent velocity of diapirs originating from the embedded depleted peridotite layer, we argue that diapirs of depleted peridotite would usually not escape from the slab during its descent. The buoyancy of the depleted peridotite layer would, however, be expected to produce local thickening of this layer as it descends, leading to emplacement in the deep mantle of irregular, thickened lenses of depleted peridotite. These lenses would then serve as isolated source regions for diapirs rising to the surface to form hot spots. Localized thickening of the depleted peridotite layer is an important aspect of the model, for without it, the source layer would be too thin to produce diapirs of sufficient diameter to serve as effective heat sources for hot spots. Larger diapirs may also be expected to form sometimes from smaller diapirs that collide with each other and coalesce.

7. Dynamics of the Hofmann–White hot spot model

Thickness considerations similar to those just discussed bear on the model for the origin of hot spots proposed by Hofmann and White (1982). In their model, subducted oceanic crust is assumed to settle and accumulate at some location in the deep mantle such as the core–mantle boundary. After 1–2 Ga, the settled crust (thickened to ~ 100 km) becomes buoyant due to radioactive heating and rises as diapirs, which provide the heat and part of the source material for hot spots.

There are difficulties with the dynamics of the settling process due to the fact that the crust is only 5–7 km thick, thereby resulting in downward-moving diapirs with a maximum radius of ~ 3.5 km (Marsh, 1979). Even if a large density contrast of 0.2 g cm^{-3} between crustal material and normal mantle at the time of settling is assumed, eq. 1 indicates that downward-moving diapirs with a maximum radius of 3.5 km would sink at a rate of only ~ 0.1 cm y^{-1}. This velocity is 1–2 orders of magnitude lower than the convective flow velocity, assuming this velocity to be similar to rates of plate motion. Thus, the crustal layer would tend to remain entrained in the flow regime of the convection cell and would not settle out. Our model does not encounter this problem, because the buoyancy of the depleted peridotite layer causes it to thicken during its descent in the subducting slab, and the thickened layer is then able to produce large diapirs that rise at rates similar to convective flow velocities.

A further problem with the model of Hofmann and White is that they suggest the density decrease in the settled crustal material caused by heating to be only ~ 0.05 g cm^{-3}. Even if one makes the unrealistic assumption that the initial density contrast required to make the crust sink effectively is very small, a subsequent density decrease of 0.05 g cm^{-3} would probably only cause the settled crust to approach a condition of neutral buoyancy at best, and diapirs would not form.

8. Isotopic and rare earth element constraints on source regions

An important constraint on models for the evolution of the oceanic mantle is the requirement imposed by Pb, Sr, Nd and Hf isotope data that:

(1) most mid-ocean ridge basalts are derived from mantle material that has lower time-integrated U/Pb and Rb/Sr and higher Lu/Hf and Sm/Nd than most oceanic island basalts;

(2) the source regions for oceanic island and mid-ocean ridge basalts have been at least partially separated for a long period of time (Tatsumoto, 1966, 1978; Sun and Hanson, 1975; Carlson et al., 1978; O'Nions et al., 1979; Wasserburg and DePaolo, 1979; Patchett and Tatsumoto, 1980). The problem is complicated by the fact that the apparent age of separation of the mantle sources determined from the Rb–Sr and U–Pb systems is ~ 1–2 Ga whereas that for the Lu–Hf system is only ~ 0.65 Ga. Therefore, as pointed out by Patchett and Tatsumoto (1980), the array

of points on an "isochron plot" must be due not only to time but also to some time-integrated combination of processes such as fractionation and mixing.

Although the existence of different and ancient source regions for mid-ocean ridge and oceanic island basalts is well established, the isotopic data exert very little control on the spatial distribution of these source regions. The result has been a proliferation of models with important differences that probably cannot be resolved by the isotopic data alone. In one class of models, the source for mid-ocean ridge basalts is taken to be the asthenosphere, whereas the source for oceanic island basalts lies in the deep mantle (Schilling, 1973; Brooks et al., 1976). In another class of models, the source regions are inverted. Ridge basalts are considered to be derived from material rising from a deep region of the mantle, and oceanic island basalts are derived, at least in part, from a more shallow mantle source (Allègre and Bottinga, 1977; Tatsumoto, 1978; Anderson, 1981c). In still another model based mainly on Nd isotope data (Wasserburg and DePaolo, 1979), the mantle is divided into a lower portion below 600–1000 km depth that contains chondritic Sm/Nd and an intermediate and upper mantle that has Sm/Nd greater than the chronditic (bulk Earth) value. Ridge basalts are considered to originate from the intermediate and upper mantle whereas oceanic island basalts are derived from hot or partially melted diapirs ascending from the lower mantle that are mixed with the intermediate and upper mantle as they rise. Thus, ridge and oceanic island basalts would both arrive at the surface with $^{143}Nd/^{144}Nd$ greater than the chondritic value, but ridge basalts would have a greater $^{143}Nd/^{144}Nd$ than oceanic island basalts, as is observed.

We suggest that the entire oceanic lithosphere originates by upward convective transport from the deep mantle at spreading ridges and constitutes one of the mantle sources. This source has Pb, Sr, Nd and Hf isotope ratios like mid-ocean ridge basalts. The upper portion of the lithosphere becomes major-element depleted during the formation of oceanic crust while the lower portion at depths greater than ~ 30 km remains undepleted in major elements and is available to be tapped later at a hot spot (Presnall et al., 1979; Presnall, 1980). The other mantle source is assumed to lie in and beneath the asthenosphere, and consists of material with generally higher U/Pb and Rb/Sr and lower Lu/Hf and Sm/Nd than the lithosphere. Based on data from Kerguelen (Dosso and Murthy, 1980), it appears that this second source region, at least locally, contains even higher Rb/Sr and lower Sm/Nd than the bulk Earth values. Unless the depleted diapirs were contaminated during their ascent through the sub-lithospheric mantle, they would have an isotopic signature similar to mid-ocean ridge basalts. Large variations in $^{87}Sr/^{86}Sr$ occur along segments of the Mid-Atlantic Ridge not associated with any nearby hot spot volcanism (White et al., 1976) and heterogeneity of $^{143}Nd/^{144}Nd$ in ridge basalts has also been observed (Carlson et al., 1978). Thus, although the two mantle source regions are broadly different isotopically, at least one and probably both are somewhat heterogeneous.

A depleted diapir rising from within the mantle would produce magmas with an isotopic signature derived from some mixture of the lithospheric and sub-lithospheric source regions, as well as the diapir itself. For a diapir rising at a spreading ridge, as in Iceland, the isotopic signature would be derived almost entirely from the diapir. The isotopic data are consistent with this model. Values of $^{87}Sr/^{86}Sr$, $^{143}Nd/^{144}Nd$ and $^{176}Hf/^{177}Hf$ for Icelandic lavas are within the range of values for mid-ocean ridge basalts (White et al., 1976; O'Nions et al., 1977; Patchett and Tatsumoto, 1980) and the $^{207}Pb/^{204}Pb$ and $^{206}Pb/^{204}Pb$ values partially overlap those for mid-ocean ridge basalts (Tatsumoto, 1978). Other oceanic islands not lying directly on a spreading ridge have varied isotopic compositions in general, but display only small isotopic variations within a given island group. Furthermore, the isotopic data lie along linear arrays that could represent mixing lines between two isotopically heterogeneous but generally different source regions (Brooks et al., 1976; Tatsumoto, 1978; DePaolo and Wasserburg, 1979b; Dosso and Murthy, 1980).

An additional constraint is the REE data for oceanic island basalts. For example, in Hawaii, if the tholeiites are derived from at least 65 km depth, as discussed earlier, and if a conventional lherzolite mantle is assumed, phase equilibrium constraints require the amount of melting to be large, about 20% or more (for example, see Green and Ringwood, 1967; Presnall et al., 1978). Nd isotope data (O'Nions et al., 1977) for the tholeiites require that the time-integrated Nd/Sm ratio of the source be less than the chondritic value. However, because light REE are significantly enriched in the tholeiites, it is not possible to obtain these basalts by large amounts of melting from a source with Nd/Sm less than the chondritic value. If the constraint of large amounts of melting is retained, a metasomatic event is required to enrich the source in light REE. However, this event must have been recent in order to satisfy the Nd isotope data (Frey et al., 1978). The same argument applies for the alkalic basalts and nephelinites in Hawaii and for alkalic basalts at other oceanic hot spots, but the phase equilibrium data permit the amount of fusion to be somewhat less in these cases. An exception is Kerguelen. In this island group, metasomatism is not required because the Nd isotope data indicate that the volcanic rocks are derived from an ancient source enriched in light REE (Dosso and Murthy, 1980).

Direct evidence for mantle metasomatism is found in ultramafic nodules from many localities that contain amphibole, phlogopite, apatite and sphene as minor phases (Lloyd and Bailey, 1975; Frey and Green, 1974; Wilshire and Trask, 1971; Francis, 1976; Wilshire et al., 1980). Frequently these phases occur in veins cutting through lherzolite, which in many cases predate inclusion of the nodule in the host basalt. In addition to the mineralogical and structural evidence, Menzies and Murthy (1980) have emphasized that lherzolite and even harzburgite nodules have been found to contain high concentrations of light REE. They considered the harzburgites to be residues of partial melting that were later enriched in light REE and other trace elements by metasomatism. Some metasomatism of the asthenosphere and lower lithosphere may take place near ridges, as suggested by Frey and Green (1974). We suggest that it may also take place immediately prior to and during the life of an island volcano as a result of lateral and upward migration of small amounts of liquid within and beneath the asthenosphere surrounding the diapir.

9. Conclusions

Previous models for the origin of hot spots include those of Wilson (1963), Green (1971), McDougall (1971), Morgan (1972a, 1972b), Turcotte and Oxburgh (1973), Shaw and Jackson (1973), Anderson (1975) and Hofmann and White (1982). Anderson (1975) proposed a model based on the assumed preservation in the mantle of CaO-, Al_2O_3- and TiO_2-enriched regions dating from the original accretional history of the Earth. These regions would then produce "chemical plumes" that would rise diapirically and produce volcanism at hot spots. This model in its original form has now been implicitly abandoned by its creator (Anderson, 1981c). As discussed earlier, the gravitational anchor model of Shaw and Jackson (1973) appears to be ruled out by density considerations and the model of Hofmann and White (1982) is unlikely because of density and dynamic arguments. Models that rely on a propagating fracture (Green, 1971; McDougall, 1971; Turcotte and Oxburgh, 1973) would not explain the observation that hot spots move very little with respect to one another (Morgan, 1972b; Minster et al., 1974). This feature seems to require an origin that incorporates a heat source lying beneath the asthenosphere, a requirement that is met by the plume hypothesis of Morgan (1972a, 1972b) and by our model. An unexplained aspect of Morgan's model is the process by which localized regions of the deep mantle develop sufficient concentrations of radioactive heat-producing elements that they rise as hot plumes.

Our model involving diapirism of depleted peridotite satisfies heat, mass balance, and density requirements, and is consistent with constraints imposed by isotopic and trace element data. In addition, it explains: (1) the varying ages and lifetimes of hot spots; (2) the widely varying rates of lava production; and (3) the common occurrence of hot spots near spreading centers and their rarity near subduction zones, while still allowing major hot spots like Hawaii to exist far from any accreting plate boundary.

Our model differs from the plume hypothesis of Morgan in two major respects. First, the depleted diapirs are not anomalously hot except in their upper parts where the geotherm for the surrounding mantle falls below the nearly adiabatic temperature curve of the diapir. Second, hot spots are considered to be a passive consequence of mantle convection and fractionation at spreading centers rather than a driving force for plate tectonics.

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