

MANTLE PLUMES: DYNAMICS AND GEOCHEMICAL SIGNATURE

PLUMAS MANTÉLICAS: DINÁMICA Y SIGNATURA GEOQUÍMICA

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Abstract

Mantle plumes constitute a secondary type of convection. Their total heat flow represents 10% of the mantle heat loss, corresponding approximately to 8% of the global Earth's heat budget. Mantle plumes originate at the D'' core-mantle boundary as a result of the thermal destabilization triggered by the core heat. They ascent towards the base of the lithosphere with a nearly spherical-shaped head followed by an elongated tail connected to the D'' source area. During this upwelling process, plume heads liberate some of their heat into the neighbouring mantle zones, which they also partly assimilate.

Plumes are constituted by differentially enriched zones, as evidenced by the chemical composition of basalts in continental flood provinces, oceanic plateaux, and oceanic islands, which are supposed to be the result of plume activity. The geochemical and isotopic data of the last 15 years show that the more enriched domain is not uniform, and that at least three contrasted components can be identified: HIMU, EM1, and EM2. The HIMU component is enriched in high field strength elements and it is strongly radiogenic in Pb. The EM1 component is enriched in both large ion lithophile elements and light rare earth elements and it displays slightly elevated Sr isotopic relationships being also depleted in radiogenic Nd and Pb. Finally, the EM2 component is enriched in large ion lithophile elements, it is highly radiogenic in Sr, and its Nd and Pb isotopic values are intermediate between the two other components. The most widely accepted hypothesis suggests that the HIMU component derived from altered and dehydrated oceanic crust (~3%), while the EM1 and EM2 components represent a mixture between the HIMU component and a small percentage of previously dehydrated 1500-2000 Ma-old pelagic and terrigenous sediments. The less enriched domain is constituted by a mantle characterized by low $^{87}\text{Sr}/^{86}\text{Sr}$ and high Nd, Pb and He isotopic ratios. This component, called FOZO (Focus Zone), PHEM (Primitive Helium Mantle) and C (Common Component), is thought to be derived from the lower mantle.

Key words: Mantle plumes, continental flood basalts, oceanic plateaux basalts, oceanic island basalts, HIMU, EM1, EM2, lower mantle.

Resumen

Las plumas constituyen un modo de convección secundario. Su flujo de calor total representa el 10% del calor que emerge a través del manto y el 8% del que globalmente pierde la Tierra. Se originan en el límite manto-núcleo, por la desestabilización que genera el calor que desprende el núcleo en la capa D''. Ascenden hacia la base de la litosfera adoptando la forma de una cabeza más o menos esférica, seguida de un tallo que conecta con su zona-fuente. Durante su ascenso la cabeza transmite parte de su exceso térmico al material que atraviesa, por lo que asimila una parte del mismo.

La composición química de los basaltos continentales y oceánicos de plateaux y la de las lavas de islas oceánicas, que se supone que se generan por la fusión de plumas, sugieren que estas están compuestas por dominios diferentemente enriquecidos. Las recopilaciones geoquímicas e isotópicas llevadas a cabo en los últimos 15 años indican que el dominio más enriquecido no es uniforme, sino que está compuesto al menos por tres componentes a los que se les denomina HIMU, EM1 y EM2. El componente HIMU está enriquecido en elementos de elevado potencial iónico y es altamente radiogénico en Pb. El EM1 está enriquecido en elementos traza de elevado radio iónico así como en tierras raras ligeras y presenta moderadamente elevadas relaciones isotópicas de Sr, pero está empobrecido en Nd y Pb radiogénicos. Finalmente, el EM2 está enriquecido en elementos traza de elevado radio iónico, es altamente radiogénico en Sr y sus valores isotópicos de Nd y Pb son intermedios entre los de los dos componentes anteriores. La hipótesis más ampliamente aceptada es que el componente HIMU representa corteza oceánica alterada y deshidratada (~3%), mientras que los componentes EM1 y EM2 se explican por la mezcla de componente HIMU y un pequeño porcentaje (~3%) de sedimentos pelágicos y terrígenos antiguos (1500-2000 m.a.), previamente deshidratados (~6%). El dominio menos enriquecido está constituido por un manto relativamente

empobrecido en Sr radiogénico y con altas relaciones isotópicas de Nd, Pb y He, al que se le ha denominado FOZO (Focus Zone), PHEM (Primitive Helium Mantle) y C (Componente Común), y se supone que representa manto inferior.

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Palabras clave: Plumás mantélicas, basaltos continentales de plateau, basaltos oceánicos de plateau, basaltos de islas oceánicas, HIMU, EM1, EM2, manto inferior.

Introduction

Observations of Earth's surface phenomena and the theoretical inferences that can be deduced from the application of the Physical principles both suggest that most of the geological phenomena are a direct consequence of mantle convection processes (e.g., Griffiths and Turner, 1998).

The progressive escape of the internal mantle heat towards the surface gave rise to a relatively cold thermal layer in its upper sector. The conduction of the core heat towards the base of the mantle triggered a weak hot thermal layer in the core-mantle boundary.

As shown by fluid dynamics, convection begins when a fluid thermal layer becomes unstable accompanied by either its downwelling or upwelling. In the Earth, the downwelling of the lithosphere into the mantle in the subduction zones results in a passive upwelling circulation towards the oceanic ridges. The cycle that begins in these spreading zones with the formation of the lithosphere, its progressive cooling during its drifting towards the zones of subduction, the subduction of these slabs and their downwelling till they reach the base of the convection zone (possibly the D'' layer), culminates with the passive ascent of the mantle towards the oceanic ridges. This complete cycle is thought to be responsible for the loss of most of the mantle heat (~85%; Davies, 1999).

In addition to the convection related to the subducting plates, the thermal layer of the base of the mantle can also become unstable thus ascending towards the surface. The same that occurs in the convection model, the upwelling processes are also counterbalanced by a passive return flow. This other mode of secondary convection involves a total heat flow of about 3.5 TW (Davies, 1999), representing the 10% of the mantle-derived heat, corresponding as a whole to 8% of the total heat loss of the Earth. As this heat is mostly transported by mantle plumes towards the base of the lithosphere, except in some anomalous cases such as Iceland, only a small proportion of it is liberated to the surface (e.g., Davies, 1999). This implies that plumes contribute to decrease the temperature of the core as they transfer the heat from the core to the mantle, but not outside of it (Stacey and Loper, 1984).

Numerical models show that the behaviour of the two thermal layers is independent, implying that they interact very slightly, i.e., plumes ascent through the mantle without interfering with the convective flow of the plates. This explains that there is no correlation between the location of plumes and the plate distribution. However, the correlation between the location of plumes, the Geoid anomalies, and the lower propagation rates of the seismic waves in the lower mantle indicate that plumes are mostly generated far from the lithospheric subduction zones (Stefanick and Jurdy, 1984; Richards and Engebretson, 1992).

In the first section of the paper we will describe the structure and dynamics of the plume, using the experimental and theoretical aspects. We will now discuss the chemical composition inferred from the geochemical signature of continental flood basalts, oceanic plateaux, and oceanic islands.

Structure, dynamics, and zone of generation

The core is at a higher temperature than the mantle as a result of the progressive heat loss by convection that characterized the later since the early stages of the Earth's evolution. Within this scenario, the core also loses heat and gives rise to a 100-200 km-thick thermal zone in the lower mantle (Davies, 1990) whose mean temperature is probably 300-400° C higher than the overlying lowermost mantle (Griffiths and Campbell, 1991). Therefore, the viscosity of this thermal zone is probably much lower than the rest of the mantle, triggering the destabilization and upwelling of some sectors of this layer.

Experiments by Whitehead and Luther (1975) evidenced that the ascent of a fluid with a lower viscosity than the surrounding one gives rise in its apical zone to a spherical "head" whose diameter is much wider than the column or "tail" that follows it and connects it to the deeper source. These results have been extrapolated to suggest that the ascent of the core-mantle boundary material might happen in a much similar way.

Later, Griffiths and Campbell (1990) showed with another laboratory experiment, that when a hot fluid with low viscosity is injected into another one of similar composition and higher viscosity, the head of the first one increases during its ascent as a result of two conjugate factors:

a) the higher ascent velocity of the tail resulting from its higher temperature, and b) the assimilation of the fluid that is being penetrated. When the tail material reaches the plume head it flows radially around it, inducing the quick diffusion of the heat. This heat gives rise to a thin thermal sheet in the periphery of the head and it is partially transferred to its inner portions thus rising the temperature of the assimilated fluid. Additionally, during plume ascent the tail and head transfer part of their excess heat to the fluid they are penetrating. This triggers a lower density thermal layer around the head that ascends and is partially assimilated by it. The circulation that occurs in the inner portions of the plume head gives rise to a helicoidal-type structure as a result of the ascent of the new hot fluid in its central sector through the tail and of the relative downwelling around the equator. According to the experimental results, we may infer that the head of the mantle plume is constituted by a mixture of plume-source material and mantle elements assimilated during the ascent. Additionally, the assimilation of the mantle material by the head of the plume, triggers its cooling and thus its mean temperature is lower than the tail. However, the most apical sector of the head is not cooled significantly as we have already said that this zone is the site of the continued flow of hot tail material.

Finally, the numerical models developed by Davies (1999) with 1:1, 1:30, and 1:200 plume-fluid viscosity rates showed that the size of the plume head is similar in the three cases. However, the tail is increasingly thinner with lower viscosities. This implies that the lowest viscosity fluid needs a thinner conduct than the higher viscosity one (assuming equal ascent velocities).

Regarding the zone of generation of mantle plumes, the most widely accepted hypothesis involves the core-mantle boundary (D'' layer). However, some authors also propound that these upwellings might also come from the boundary between the upper and lower mantle (670 km discontinuity). The first model implies that the core has a higher temperature than the lower mantle and that it gives rise to the required thermal flow for the ascent of plumes towards the base of the lithosphere. The second hypothesis suggests that the mantle convection should occasionally happen in two layers, separated by the upper/lower mantle boundary. Campbell and

Griffiths (1992) propose three important arguments favouring the core-mantle model: a) the existence of plumes with head diameters up to 1000 km, as required for the generation of the huge volumes of lavas characteristic of continental and oceanic plateau-basalts; b) the presumed fixed position of hotspots with respect to each other; and, c) the Earth's thermal budget requires at least 70% of its heat to be generated within the lower mantle. If there is a thermal upper-lower mantle boundary, this heat must be transmitted through the upper mantle to the surface in ascending plumes, but this is inconsistent with the low (< 10%) estimated heat flow for all known plumes.

Even if we do not want to go into great details concerning this controversial issue, we should mention that the plume composition is expected to be different depending on the depth of its source, i.e., the 670 km discontinuity or the core-mantle boundary (e.g., Fitton *et al.*, 1997). The head of a plume coming from the 670 km discontinuity will be characterized in its peripheral zone by overheated asthenospheric mantle and fragments of the enriched component (if it has been previously accumulated in this layer), and lower mantle in its axial zone. Therefore, both zones will show contrasted chemical and isotopic signatures, with an outer region with a composition similar to the MORB source, and an inner sector with a less depleted mantle. By contrast, the head of a plume originating at the core-mantle boundary will be essentially constituted by material of this thermal layer and the lower mantle, with a small envelop of asthenospheric mantle. Finally, the head of a plume coming from the D'' layer that is temporally ponded in the 670 km discontinuity (to trigger later one or several secondary plumes) will have a similar composition to the first case, with additional hot material from the D'' layer.

Geochemical signature

If we consider the mechanism of ascent of plumes and we accept the hypothesis that they were probably generated at the core-mantle boundary, the plume head must be constituted by a mixture of the thermal layer D'' and the lower mantle. Additionally, the tail which only exceptionally will assimilate the surrounding mantle, should be constituted by material from the core-mantle boundary. This means that the

volcanic rocks coming from the plume head will provide information about the geochemical signature of the lower mantle, whereas the lavas generated by the melting of the tail better identify the composition of the D'' layer, specially when the melting rates are high.

The geochemical characteristics of the basalts which are supposed to be derived from plumes (e.g., their concentration in compatible and incompatible elements and the Os-Sr-Nd-Pb isotopic ratios) suggest that all of them carry both enriched and less-enriched domains. The most accepted hypothesis about the enriched domain suggests that it corresponds to oceanic crust (Hofmann and White, 1982), whereas the less-enriched domain represents a material geochemically different from the MORB source, possibly lower mantle (e.g., Hauri *et al.*, 1994 and Campbell, 1998). On the other hand, the relatively high concentration in compatible elements in the most primitive lavas from plumes, require that the source is constituted by ~60-70% of less-enriched peridotitic material and ~30% of transformed oceanic crust (see for example Chauvel *et al.*, 1992 and Campbell, 1998).

Plume heads

If we accept the hypothesis proposed initially by Morgan (1981) and later on adopted by Richards *et al.* (1989) and Griffiths and Campbell (1990), which suggests that the basalts from oceanic and continental plateaux are the result of partial melting of plume heads, the geochemical composition of these basalts can provide information about this sector of mantle plumes. Unfortunately this is not an easy task as the plumes are composed of differently enriched domains and thus the melting degree plays an important role. In this respect, the picritic and komatiitic magmas better identify the source chemistry than the tholeiitic ones and the alkali basalts. On the other hand, the melts derived from the plume head can interact with liquids derived from the lithospheric mantle and/or assimilate crustal material during their ascent. For example, the geochemical modeling by Gibson *et al.* (1995) of the Paraná basalts indicates that all melts derived from the plume head carry high proportions (up to 50%) of liquids from the lithospheric mantle and, additionally, some of them have assimilated upper crust material.

Despite all this, if we remove from the continental plateaux basalts those rocks with obvious signs of crustal contamination and participation of lithospheric mantle, it can be concluded that magmas with a signature similar to the plume-related OIB have been reported in nearly all CFB large provinces, representing a small proportion of the total volume of magmas generated. For example, the Ambenali Formation basalts from the Deccan area, the earliest emissions from the Imnaha Formation in Columbia River and some other outcrops in the Eastern coast of Greenland, show low LILE/HFSE ratios (i.e. the source is enriched in HFSE) and low Sr and high Nd isotopic ratios (e.g., Hawkesworth *et al.*, 1990 and Saunders *et al.*, 1992).

Although the basalts and picrites from phanerozoic CFB provinces suggest the presence of an OIB-type mantle, the data from similar lavas in the Archean and Proterozoic indicate that during these times the depleted mantle prevailed (see Campbell, 1998). If this is correct, it would mean that the composition of the thermal core-mantle boundary has evolved through time.

The basalts, picrites and komatiites from oceanic plateaux (e.g., Ontong-Java and Caribbean), which do not traverse a thick and complex lithospheric mantle or continental crust, provide a more precise insight of the nature of plume-heads than their continental counterparts. In the plume model, the initial melts (which are presently located at the bottom of the plateau) proceed from the core-mantle thermal boundary, whereas the melts produced at later stages when the plume head expands under the lithosphere, derive from a mixture of the plume head material and the lower mantle. Finally, the melts extruded during the final stages arise from the entrained lower mantle (see Campbell, 1998). The available geochemical data for the shallower levels of these plateaux (see for example Neal *et al.*, 1997 and Kerr *et al.*, 1997, 2000) exhibit La/Lu ratios which are in general lower than those of OIB. In other words, their abundances in REE (and other incompatible elements), normalized to Primitive Mantle, is either flat or slightly enriched. The MgO-rich lavas (picrites and komatiites) are more heterogeneous than the basalts and they are usually more depleted in LREE than the MORB (e.g. Arndt *et al.*, 1997). These results suggest that the lavas from these plateaux were produced by high melting rates. Concerning their isotopic signatures, the Sr

ratios range from 0.7030 to 0.7044, the ϵ_{Nd} from +3 to +7.5, the $^{206}\text{Pb}/^{204}\text{Pb}$ between 18.2 and 19.2, $^{207}\text{Pb}/^{204}\text{Pb}$ from 15.48 to 15.59 and the $^{208}\text{Pb}/^{204}\text{Pb}$ ratios between 38.2 and 39.0 (Mahoney *et al.*, 1993 and Kerr *et al.*, 1996).

As the trace element abundances and isotopic data for these lavas are not characteristic of OIB, Campbell and Griffiths (1992, 1993) suggested that the oceanic plateau basalts come from a hybrid source in which the lower mantle is well represented. This interpretation implies that the composition of the enriched portion of the source is diluted as a consequence of the high melting degrees and thus the depleted portion (i.e. the one assumed to be derived from the lower mantle) prevails. As we will see, the isotopic composition of this source is close to the FOZO component by Hart *et al.* (1992) which has also been interpreted as lower-mantle.

Plume tails

According to the plume structure described above, the material from the thermal layer D'' is well represented in the uppermost sector of the head and tail. As a consequence, its geochemical signature can be obtained from the first melts generated in a CFB province and/or the lavas from oceanic islands. For the reasons exposed above, the composition of the later is more discriminatory.

The compilations of Sr, Nd and Pb isotopic ratios in OIB by Tatsumoto *et al.* (1984), White (1985), Zindler and Hart (1986), Hart (1988) among other authors, have shown that there is some heterogeneity. This can only be explained if we accept that the enriched domain is not uniform and is made up of at least three components which, following Zindler and Hart (1986), are termed HIMU, EM1 and EM2. The basalts from Santa Helena, Cook and Tubuai arise from a mantle source highly radiogenic in Pb ($^{206}\text{Pb}/^{204}\text{Pb} > 19.5$) and hence their HIMU denomination (high m, $^{238}\text{U}/^{204}\text{Pb}$), but not Sr ($^{87}\text{Sr}/^{86}\text{Sr} < 0.7035$). These basalts plot in the $^{87}\text{Sr}/^{86}\text{Sr}$ - $^{143}\text{Nd}/^{144}\text{Nd}$ diagram below the field of the ridge basalts, because their Nd isotopic ratios (0.5128-0.5129) are lower than in MORB, but within depleted or non-radiogenic quadrant. The basalts from Gough, Tristan and Kerguelen are supposed to be generated from a EM1 mantle source, moderately enriched in radiogenic Sr ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7050$ - 0.7055) but depleted in

radiogenic Nd and Pb. Finally, the rocks from some islands of Samoa and Sociedad have $^{87}\text{Sr}/^{86}\text{Sr}$ ratios > 0.7065 and $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios which are intermediate between the two other components and are assumed to be generated from an EM2 mantle.

The ratios of trace elements with different geochemical behavior (e.g. REE, LILE and HFSE) of oceanic basalts can provide additional information about the characteristics of the mantle components from which they derive (Weaver, 1991). The abundances in incompatible trace elements of representative basalts from each one of the above described components, normalized to primitive mantle, indicates that the HIMU is enriched in HFSE relative to LREE and LILE, thus leading to positive anomalies in Nb and Ta. The component EM1 has high LILE/HFSE and LREE/HFSE ratios, and thus Ba is enriched relative to other LILE and lack positive anomalies in Nb and Ta.

The geochemical signature of EM2 is not as well defined as the other two components because the analyses of basalts supposed to be derived from this component are less abundant. In general, the LREE/HFSE ratios (e.g. La/Nb) as well as the LILE/HFSE ratios (e.g. and K/Nb) of this component are significantly higher than those of HIMU. Additionally, the Rb/Nb, Ba/Nb, Ba/Th and Ba/La ratios are lower than in EM1.

Concerning the signature of the less-enriched domain, the plot of the most primitive lavas of many oceanic islands in the DM-HIMU-EM1-EM2 tetrahedron converge towards a $^{87}\text{Sr}/^{86}\text{Sr}$ depleted (< 0.7025) composition with high $^{143}\text{Nd}/^{144}\text{Nd}$ (> 0.5131), $^{206}\text{Pb}/^{204}\text{Pb}$ (19.1-19.7) and $^3\text{He}/^4\text{He}$ ratios (> 8 Ra), which Hart *et al.* (1992) denominated FOZO (Focus Zone). If the components HIMU and EM are concentrated in the D'' layer, the FOZO reservoir would represent lower mantle entrained by the plumes during their ascent (See discussion in Hauri *et al.*, 1994). As we saw previously, the similarity of the geochemical signature of the FOZO and the source of the shallowest levels of oceanic plateaux suggests the presence of a common component in all plumes, which must constitute a significant volume of mantle distinct from the source of MORB.

Origin of the HIMU, EM1 and EM2 components

The characteristics of the HIMU component can be explained by a process of dehydration of altered oceanic crust (Weaver, 1991). As it is well known, the solubility of trace elements into a hydrous fluid is controlled by the ionic potential, Z/r , where Z is the charge of the ion in its usual oxidation state and r its ionic radius in Å. As a consequence, elements such as Cs, Rb, K, Ba and Sr, with low ionic potential are highly soluble, whereas Nb, Ta and other elements with high ionic potential are insoluble and stay as residual. If we assume that the MORB are hydrothermally altered during their drifting towards subduction zones with addition of LILE elements, about 3% of dehydration of that oceanic crust produces a residuum with an abundance in trace elements similar to that of HIMU.

In this sense, the high isotopic ratios of Pb which are characteristic of this component can be explained if we consider that during the dehydration there is an enrichment in U and Th relative to Pb in the residue. This is due to the higher solubility of Pb^{2+} (ionic potential = 1.68) relative to U^{6+} , U^{4+} and Th^{4+} (with ionic potentials of 8.22, 4.49 and 4.26, respectively). This enrichment increases the U/Pb and Th/Pb ratios of the dehydrated oceanic crust, whereas the Th/U ratio remains constant or changes only very slightly.

The high isotopic ratios of $^{187}Os/^{186}Os$ of the basalts from Santa Helena, Rurutu, Tubuai and Mangaia (1.1159-1.2474) (Hauri and Hart, 1993) confirm that the source of the HIMU basalts require the participation of recycled oceanic crust which is characterized by high Re/Os and $^{187}Os/^{186}Os$ ratios. Furthermore, the interaction of the oceanic crust with seawater can increase its Re abundance as well as their initial $^{187}Os/^{186}Os$ ratios so that its proportion in the source of those basalts is within the above indicated values. According to the calculations by Widon (1997), if we assume that the peridotitic component has values of $^{187}Os/^{186}Os$ of 1.1050,

to obtain the isotopic signature of the HIMU basalts the participation of oceanic crust must be between 15 and 30%.

Despite all the geochemical transformations described above, this crust remains of basaltic composition and hence retains a density different to the surrounding mantle. If we accept that, as a consequence of its relatively high abundances of Fe, Si and Ca, the density of that crust is higher than the surrounding mantle and it can detach from the ultrabasic refractory lithosphere which is subducting along, to finally accumulate at the core-mantle boundary (D" layer) constituting an irregular layer of "degenerated crust" (Hofmann and White, 1982). The isolation of this material in this layer during periods of time between 1.6 to 1.8 Ga, as suggested by the pre-sensible isochrons of recent OIB in the diagrams $^{87}Sr/^{86}Sr$ - $^{87}Rb/^{86}Sr$ and $^{207}Pb/^{204}Pb$ - $^{206}Pb/^{204}Pb$, allows the development of the isotopic characteristics (particularly Pb) of the HIMU component.

More recently, Hanan and Graham (1996) have proposed that it is possible to generate the HIMU signature in about 300 Ma, and Hoernle (1998) has analyzed portions of the altered oceanic crust under the Canary Islands that well before their hypothetical future subduction (they stand on a passive margin) show the isotopic characteristics and concentrations in incompatible trace elements similar to the HIMU component. The latter author has also shown that the maximum age in which the average oceanic crust under Gran Canaria would produce the isotopic signature of the Santa Helena- and Mangaia-type HIMU should be of 770 Ma and 1070 Ma, respectively.

The isotopic signature of the EM1 and EM2 components is explained as a mixture of HIMU and a small percentage (~3%) of old pelagic and terrigenous sediments (1500-2000 Ma)¹ previously dehydrated (~6%), with high LILE/HFSE and LREE/HFSE ratios (Weaver, 1991). The isotopic composition of oxygen in fresh volcanic glasses of EM-type submarine

1. The Pb isotopic data require that the pelagic sediments of the EM1 component are contemporaneous with the subducted oceanic crust and different from modern sediments as their $^{206}Pb/^{204}Pb$ and $^{207}Pb/^{204}Pb$ ratios are not adequate (Weaver *et al.*, 1986). Finally, although the present terrigenous sediments have congruent Sr and Nd isotopic ratios, the old sediments that evolved to the present with moderately high Rb/Sr and low Sm/Nd ratios also meet the isotopic requirements, so it is also assumed that the sediments of the component EM2 are old and they subducted into the mantle together with the associated oceanic crust (Weaver, 1991).

volcanoes (e.g. Pitcairn in Polynesia) confirm the crustal participation in these mantle components (Woodhead *et al.*, 1993)

Conclusions

Plumes constitute a secondary type of convection. Their total heat flow represents 10% of the heat coming from the mantle, and 8% of the global Earth's heat budget.

They originate at the core-mantle boundary as a result of the destabilization induced by the heat transferred by the core to the D'' layer, and they ascend towards the base of the lithosphere with a nearly spherical-shaped head, followed by a tail rooted in the source region. During this upwelling process, the head transfers part of its heat to the surrounding materials, assimilating part of it.

They are composed of fragments of enriched domains included within a less enriched matrix. The abundance in compatible trace elements of the most primitive lavas generated by plume activity indicates that the less enriched domain (peridotitic) represent 60-70%, while the enriched fragments constitute 30% of the total.

The geochemical and isotopic data from plume-derived lavas show that the enriched domain is not uniform, and it is composed by at least three components: HIMU, EM1, and EM2. The HIMU component is enriched in high field strength elements and it is highly radiogenic in Pb. The EM1 component is enriched in both large ion lithophile elements and light rare earth elements and it displays slightly elevated Sr isotopic relationships being also depleted in radiogenic Nd and Pb. Finally, the EM2 component is enriched in large ion lithophile elements, it is highly radiogenic in Sr, and its Nd and Pb isotopic values are intermediate between the two other components.

The most widely accepted hypothesis suggests that these components correspond to the dehydrated oceanic crust (HIMU), with a small percentage (~3%) of previously dehydrated ancient pelagic and terrigenous sediments (EM1 and EM2). In order to obtain the characteristic isotopic signatures of these three components, the "degenerated" oceanic crust has to remain during a long period of time isolated from the convective mantle, probably in the core-mantle boundary. The high $^3\text{He}/^4\text{He}$ ratios of plume-

derived lavas (with respect to MORB) comes from the degasification of the core or the lower mantle assimilated during the ascent.

There are many arguments favouring the idea that the less enriched domain corresponds to the lower mantle. The high $^3\text{He}/^4\text{He}$ rate of the plume-derived lavas and the supposed high temperature of the plume with respect to the surrounding mantle clearly suggest a deep origin. Additionally, the most primitive lavas of many oceanic islands plot in the tridimensional space of the components DM-HIMU-EM1-EM2 oriented towards a region with geochemical and isotopic characteristics clearly different from the source-mantle of the mid-oceanic basalts [variously called FOZO, Hart *et al.*, (1992); PHEM, Farley *et al.* (1992); or, C, Hanan and Graham (1994)].

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