Doklady Earth Sciences, Vol. 380, No. 7, 2001, pp. 858–862. Translated from Doklady Akademii Nauk, Vol. 380, No. 4, 2001, pp. 523–527.

GEOCHEMISTRY

Low-Density Anomalies in the Mantle: Ascending Plumes and/or Heated Fossil Lithospheric Plates?

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Presented by Academician F.A. Letnikov May 8, 2001

Received May 22, 2001

In the beginning of the 1970s, Morgan [1] introduced the concept of the mantle plume into the everyday terminology of geological research. In succeeding years, different aspects of this hypothesis were considered: the heat source and mechanisms of plume ascent [2–4]; isotope-geochemical features [3, 5–8]; manifestation in geophysical fields [3, 9–11]; and so on. In classical models, a plume is a low-density mantle material ascent to the Earth's surface from the core–mantle boundary. In general, the plumes are vertical, but it is considered that they can deflect from the vertical axis in a horizontal convective flow with increasing inclination and break into a series of separate "drops" [3]. Two characteristic structural parts of the plume, the head and conduit (Fig. 1a), were distinguished on the basis of mathematical calculations and laboratory experiments ([2] and review [3]).

The plumes are traced on the sections of deep seismic topography as subvertical or inclined zones of decreased velocities of seismic wave propagation (regions of deconsolidation), e.g., a strongly inclined low-density anomaly below Africa (Fig. 1b [10]). A clearly expressed inconsistence between the distinguished structure and the idealized model form of the plume engages our attention. This can be partially caused by the methodological peculiarities of deep seismic tomography. However, the discreteness and significantly greater horizontal sizes of this anomaly are not doubted. Since the appearance of this low-density structure in the mantle can be hardly explained from the positions of the classical plume model, it is necessary to consider other possibilities.

The lithospheric plates descending into the mantle along modern subduction zones manifest themselves as anomalies of the opposite type (Fig. 1c) [11]: the velocities are increased here due to a greater density of the cold subducted plates. A morphological similarity between these seemingly different low-density and high-density plates is, however, evident. Assuming that subduction continues during a long period of the Earth's geological history, we can suppose the existence of many ancient (buried) lithospheric plates in the mantle and that they were distinct from the surrounding medium in terms of composition and structure over long periods of time [3, 5]. During prolonged subduction, the plates can be broken and folded [3]. Consequently, their initial configuration would be distorted and complicated.

Let us consider a situation when low-density anomalies in the mantle can be formed due to the heating of subducted fossil plates. Unlike earlier works considering this possibility in the context of partial recycling of the plate material to the surface in the form of mantle plumes (see [2], references therein, and [3]), we suggest an approach with radiogenic heating *in situ*. This model satisfactorily explains anomalous isotope-geochemical characteristics of several Late Cenozoic basalts in the Southern Hemisphere.

Model of radiogenic heating of a buried plate. If we assume that a distributed energy source is localized within the body of a fossil lithospheric plate, the generated heat would be spent, on the one hand, on increasing the heat content (heating of the plate) and, on the other hand, on the heat emission into the surrounding medium (mantle) [12]. We can formally write the following equation:

$$Q = cV\rho(T_{m_k} - T_{m_0}) + \alpha S(T_m(\tau) - T_{c_0})\Delta\tau, \quad (1)$$

where c is the specific heat capacity of the plate rocks; V is the volume of the plate; T_{m_0}, $T_m(\tau)$, and T_{m_k} are the initial, current, and final temperatures of the body; S is the surface area of the body; α is the coefficient of heat emission; T_{c_0} is the temperature of the surrounding medium; and $\Delta\tau$ is the time interval considered here.

Let us assume as the initial value the moment when temperatures of the initially cold plate and the temperature of the surrounding medium are already equalized. To a first approximation, the heat emission can be calculated from the final temperature of the plate; then Eq. (1) will take the form:

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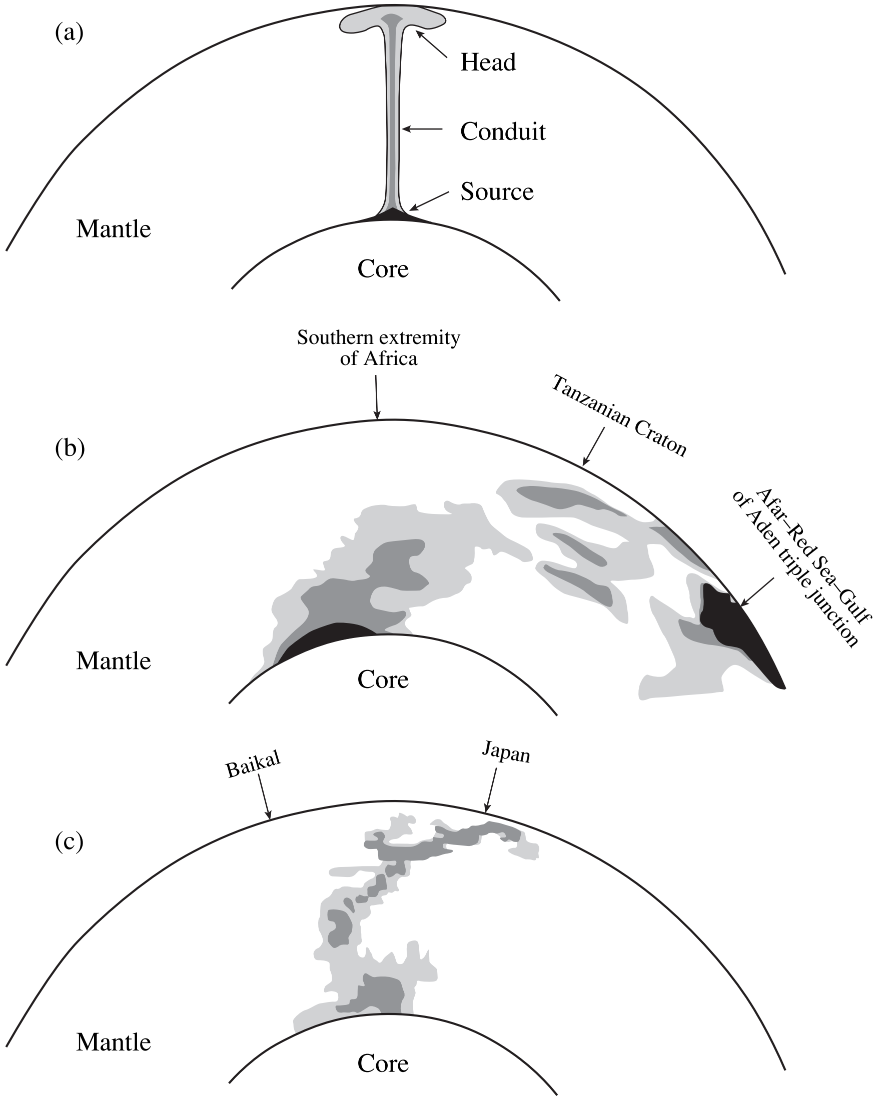

Fig. 1. Idealized form of the mantle plume (a), observed low-density anomaly below Africa [10] (b), and high-density anomaly (subducted plate) below the eastern boundary of Eurasia [11] (c). The shading is increased with decreasing density of the material in (a) and (b) and vice versa in (c).

$$Q = cV\rho\Delta T + \alpha S\Delta T\Delta\tau$$
$$\text{or} \quad Q = \Delta T(cV\rho + \alpha S\Delta\tau).$$

The coefficient of heat emission accounts for the thermal conductivity of the body (k_n) and the medium (k_c) and the thickness of the heat exchange layer from either side:

$$\alpha = \frac{k_n}{h_1} + \frac{k_c}{h_2}.$$

Based on parameters given in the table, it is possible to calculate the total heat needed to heat the plate to the supposed temperature difference: $Q \approx 1.346 \cdot 10^{25}$ J.

Since the fact of increased concentrations of radioactive elements in oceanic sediments compared to the mantle is well known, we can suggest that the distributed active heat source is the radioactive decay energy. Knowing the radioactive decay constant λ and the present-day content of radioactive isotopes in oceanic sediments (N) (table), it is easy to estimate their initial quantities (N_0):

$$N_0 = N/e^{-\lambda\tau}$$

and the amount of isotopes (ΔN) decayed within the time interval considered here (τ):

$$\Delta N = N\left(\frac{1}{e^{-\lambda\tau}} - 1\right).$$

This value is equal to $3.41 \cdot 10^{13}$ kg for ^{40}K, $6.09 \cdot 10^{10}$ kg for ^{238}U, $5.70 \cdot 10^9$ kg for ^{235}U, and $1.80 \cdot 10^{11}$ kg for ^{232}Th. We emphasize that our model considers the source as the sedimentary layer with a thickness

Physical parameters used in the calculations [3, 13, 14]

Parameter	Fossil plate				Mantle		
Specific heat capacity, c, J/(kg K)	1250						
Volume, m^3	$1.5 \cdot 10^{16}$						
Density, ρ, kg/m^3	3000–4000						
Thermal conductivity, $k_{n(c)}$, W/(m K)	3				7.5		
Surface area of the plate, S, m^2	$6.04 \cdot 10^{12}$						
Thickness of the heat exchange layer, $h_{1(2)}$, m	500				200 000		
	Isotopes				Elements		
	^{238}U	^{235}U	^{232}U	^{40}K	U	Th	K
Radioactive decay constant, λ_0, 10^{-10} yr^{-1}	1.55125	9.8485	0.49475	5.543			
Specific heat of radioactive decay, J/(kg yr)	2970	17 987	852	883.6			
Content of isotopes in the natural mixture, %	99.2745	0.7196	100	0.0117			
Content of elements in deep-water sediments, wt %					$2.6 \cdot 10^{-6}$	$1.3 \cdot 10^{-6}$	2.5
Supposed temperature difference between the plate and surrounding mantle, 300 K							
Supposed duration of the process, 1.5 Ga							

of approximately 1 km, where the content of radioactive elements significantly exceeds their typical amounts for deep lithospheric layers and the corresponding concentrations in the mantle (assuming that the content of radioactive elements in the ancient subducted sediments were almost the same as in modern pelagic sediments [13]). The total radiogenic heat (Q_R) emitted, with account for the specific heat of decay [14], is equal to $4.59 \cdot 10^{25}$ J (the account for the radiogenic heat emitted within the oceanic basaltic crust will allow us to increase this value).

If we assume that the lithospheric plate loses up to 50% of its radioactive elements during subduction due to partial melting and degassing, the Q_R value will be equal to $2.29 \cdot 10^{25}$ J. It is obvious that the heating of the plate due to radiogenic heat is realistic, even if the other sources of energy (phase transitions, chemical reactions, friction energy, etc.) are not taken into account.

Geodynamic and geochemical implications of the suggested model. The most reliable argument for the uplift of mantle plumes from the core–mantle boundary is the high values of $^3He/^4He$ ($(^3He/^4He)_{meas}/(^3He/^4He)_{atm} = R/R_a > 35$). Isotope ^3He has a cosmogenic nature and can be contained in significant quantities in the relatively nondegasified lower mantle and the core (the PHEM component, Fig. 2 [6]), whereas isotope ^4He is mainly a product of the radioactive decay of U and Th, which are abundant in the lithosphere.

Maximal values of $^3He/^4He$ were recorded in the lavas of the Loihi Volcano in the Hawaii Archipelago. In the isotopic coordinates $^3He/^4He$–$^{87}Sr/^{86}Sr$ (Fig. 2), part of the oceanic islands (e.g., Iceland) occupies an intermediate position between compositions of the Loihi Volcano and MORB rocks that are identified with the DM component ($R/R_a = 8 \pm 1$). This agrees well with the classic plume model, where the PHEM component is mixed with the DM component. However, the MORB rocks form a trend directed to the recycled component (RC) with lower values of $^3He/^4He$ (R/R_a = 6–8) and higher values of $^{87}Sr/^{86}Sr$. Volcanic islands Tristan da Cunha and Gough are located at the end of this trend. The lavas of Samoa also stretch in this direction. The former bear signs of the enriched EM1 component, while the latter bear signs of the EM2 component. All these islands are in the regions of the DUPAL isotope anomaly of the Southern Hemisphere [5].

The appearance of mantle components with enriched isotopic characteristics is generally attributed to the recycling of the ancient lithospheric material (pelagic and terrigenous sediments and continental lithosphere with an age on the order of 1–2 Ga) [3, 5, 6]. In plume models, recycling is usually understood as burial of the lithospheric material into the lower mantle (sometimes into the boundary between the lower and upper mantles), its conservation for long periods of time, and further transportation to the upper levels by the mantle plumes. The Earth's core is assumed to be the main source of heat for the generation of plumes [2, 4].

It is clear that the realization of this recycling model is overcomplicated by additional conditions. The question arises as to whether the isotope-geochemical inhomogeneities are conserved during a period on the order of 10^9 yr in conditions of sharp temperature jumps between the core and lower mantle. In addition, the observed low values of $^3He/^4He$ in the RC remain unexplained, since the contribution of the lower mantle material inevitably leads to an increase in $^3He/^4He$ over the values characteristic of the mid-oceanic ridges.

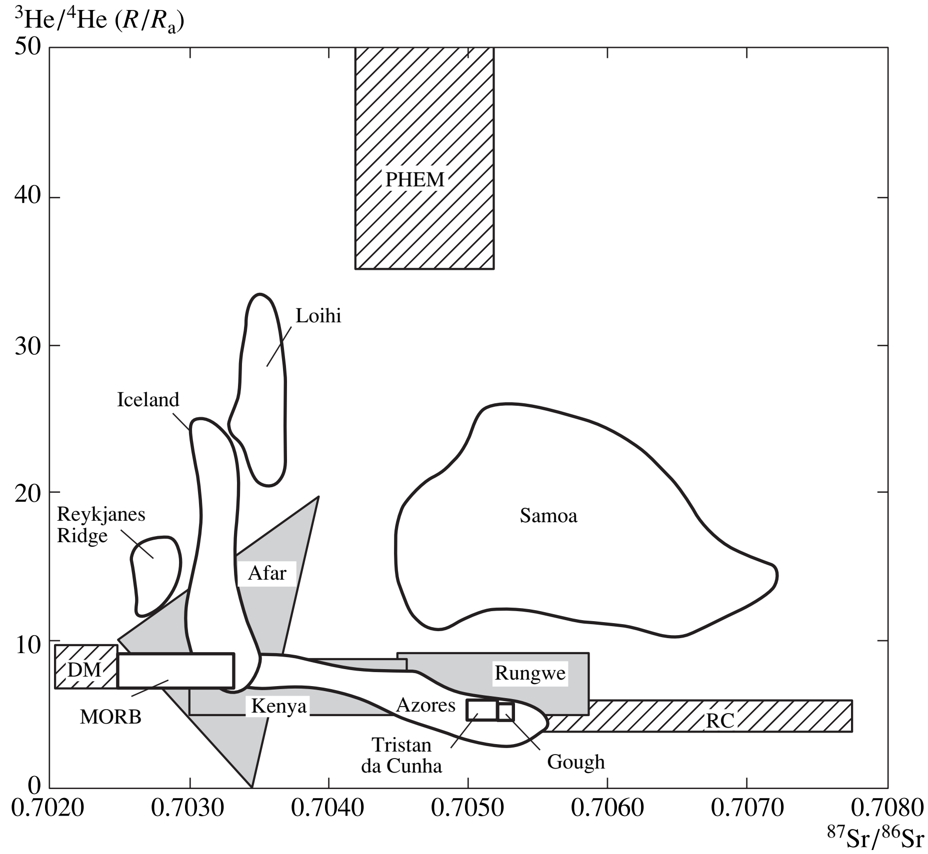

Fig. 2. Covariations of isotope ratios $^3He/^4He$ (R/R_a) and $^{87}Sr/^{86}Sr$ in oceanic basalts (modified after [6]). Figurative fields of the Late Cenozoic volcanic rocks of the eastern and northeastern Africa are shown for comparison (adopted from [7, 8, 15]). Final components: (PHEM) component of the nondegasified lower mantle, (DM) component of the depleted upper mantle, (RC) component of the recycled lithosphere. The latter component is heterogeneous and includes enriched mantle components EM1 + DUPAL and EM2 + DUPAL.

In our model, the lithospheric material retains its place within the buried plate. Degassing of the plate starts at the moment of significant heating, which finally leads to melting immediately above the plate or within the lower lithosphere above the degassing area. In such conditions, the appearance of enriched magmas with signs of melting of the recycled material and conserved low $^3He/^4He$ values is easily explained.

Intraplate oceanic islands can probably be divided into the following types: (1) islands formed as a result of the mantle plume activity (e.g., Hawaii); (2) islands formed over heated fossil plates (Tristan da Cunha, Gough); and (3) islands bearing signs of the interaction between mantle plumes and fossil plates (Samoa).

In order to make a comparison with oceanic basalts, the figurative fields of lavas of the Late Cenozoic rift systems in Africa are shown in Fig. 2 [7, 8, 15]. The location of the lavas of the Kenya rift and Western branch of the East African rift system (Rungwe field) on the mixing line between the DM and RC components is evident [8, 15]. In the Afar lavas, $^3He/^4He$ varies in a wide range (R/R_a varies from 0.009 to 16.9). The appearance of magmas with the lowest values of $^3He/^4He$ is explained by melting of the Pan-African basite core, and the appearance of magmas with the highest values of this parameter is explained by a combined source (mixture of the DM, RC, and PHEM components), where the lower mantle material does not exceed 5% [7]. Unfortunately, the isotope ratio $^{87}Sr/^{86}Sr$ is given in [7] only for the lava samples characterized by maximum values of the $^3He/^4He$ ratio in chosen regions, which did not allow us to fully discriminate the role of the recycled component.

Conclusion. A radiogenic source within a buried lithospheric plate appears powerful enough for its heating in situ. Thus, certain low-density anomalies in the mantle distinguished from the seismic tomography data can turn out to be ancient heated fossil plates but not mantle plumes. In the geochemical sense a component of such a plate would be expressed by low $^3He/^4He$ values (R/R_a = 6–8) combined with heterogeneous characteristics of the components of an enriched mantle (EM1 + DUPAL or EM2 + DUPAL). The low-density anomaly under Africa probably reflects the modern position of the ancient subduction zone postulated by

Hart [5] to explain the DUPAL anomaly of the basalts of the Southern Hemisphere.

ACKNOWLEDGMENTS

The authors thank Academician F.A. Letnikov for constructive remarks during revision of the article. This work was supported by the Russian Foundation for Basic Research (project nos. 99-05-65636, 00-05-64628, and 01-05-06058).

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