

Newly obtained data on He isotopes in mantle xenoliths in basalts from the Vitim Plateau, Eastern Siberia, Russia

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The paper reports recently obtained data on the He isotopic composition of mantle xenoliths in basalts from the Vitim Plateau. Most of the samples, which were examined by the crushing technique, have atmosphere-normalized $^3\text{He}/^4\text{He}$ ratios indiscernible from those of the a source of the MORB type. The only exception is a single sample whose ratio is significantly higher than the value typical of MORB. The Sr-Nd systematics of all of the xenoliths and their host basalt are close to those of basalts at oceanic islands (OIB). The isotopic composition of the mantle xenoliths is proved to be formed by the mixing of two sources: mantle plume and MORB. **KEYWORDS:** *Vitim plateau; mantle xenoliths; basalts; He-Ar and Sr-Nd isotopes.*

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Introduction

After the discovery of solar ^3He on the Earth in 1969 [Clarke *et al.*, 1969; Mamyrin *et al.*, 1969], high $^3\text{He}/^4\text{He}$ (R) ratios, much higher than the atmospheric value of $R_a = 1.39 \times 10^{-6}$, have been repeatedly obtained and confirmed in numerous publications [Kyser and Rison, 1982; Mamyrin and Tolstikhin, 1984; and others].

The most stable of the terrestrial reservoirs is the source of the MORB type, which is related to mid-oceanic ridge basalts and has a virtually unchanging $^3\text{He}/^4\text{He}$ ratio ($R_a = 8 \pm 1$). Higher isotopic ratios ($> 20R_a$) were found in oceanic mantle plumes, such as those at Hawaii, Samoa, and Iceland [Dixon *et al.*, 2000; Kaneoka *et al.*, 1980; Poreda and Farley, 1992], where weakly degassed primitive mantle material ascends, is melted, and thus basalts of the OIB type are derived. Judging from their isotopic and geochemical signatures, these basalts were derived by the partial melting of lower mantle material [Rison and Craig, 1983].

Data on the geochemistry and isotopic composition of plateau basalts, such as flood basalts in the Deccan Plateau, eastern Siberia, Greenland, and South Africa, suggest a contribution of a source of the OIB type at continents, and data on the deep geophysics confirm anomalous characteristics of the mantle beneath such areas.

Starting with the publication by Tolstikhin *et al.* [1972], who presented He isotopic analyses of xenoliths from Kamchatka, the Stanovoi Highland in central Siberia, and Spitsbergen, such studies were focused mostly on xenoliths in basalts from oceanic islands [Kaneoka *et al.*, 1980; Kyser and Rison, 1982; Porcelli *et al.*, 1987; Poreda and Farley, 1992].

Until recently the most thoroughly examined ultramafic xenoliths were those from Hawaii. The $^3\text{He}/^4\text{He}$ ratio of these xenoliths turned out to be lower than in their host basalts because the xenoliths themselves are of residual nature (dunites and harzburgites) [Kaneoka *et al.*, 1980; Rison and Craig, 1983; and others]. High $^3\text{He}/^4\text{He}$ ratios, one order of magnitude higher than the atmospheric ones, were determined in xenoliths in Samoan basalts.

Studies of the He isotopic systematics of xenoliths in continental basalts were launched not long ago [Drubetskoi and Grachev, 1987; Dunai and Baur, 1995; Matsumoto *et al.*, 1998; Porcelli *et al.*, 1987; Shijine *et al.*, 2003; Staudacheer and Allegre, 1991] and the continental mantle was proved to usually have R/R_a ratios equal to or lower than those of MORB.

At the same time, isotopic geochemical studies of mantle xenoliths from continental basalts have demonstrated that the mantle beneath continents is highly heterogeneous [Grachev, 1989; Ionov *et al.*, 2005; Kogarko *et al.*, 1987; and others].

The first studies of the He isotopic composition of mantle xenoliths from the Baikal rift zone have revealed broad variations in the $^3\text{He}/^4\text{He}$ ratios: from atmospheric values to those higher than in MORB [Drubetskoi and Grachev, 1987]. Inasmuch as the experiments were carried out by melting

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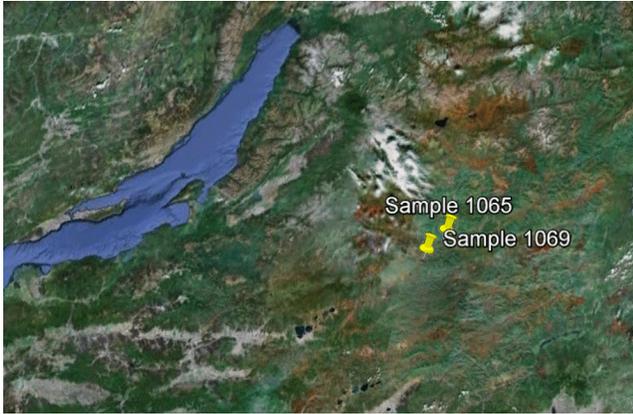


Figure 1. A Google-Earth map showing the location of the studied xenoliths samples.

techniques, the issue of the cosmogenic component has not been resolved, because the samples collected for the analysis were found on the surface, and hence, they could be exposed to cosmic radiation, and spallogenic He could be implanted in the crystal structures of minerals.

The effect of cosmic rays on the $^3\text{He}/^4\text{He}$ ratios of rocks was first quantified for mantle xenoliths in basalts from the Vitim Plateau in central Siberia and the Taryat depression in Mongolia, along with a sample of quartzite from Antarctica [Staudacher and Allegre, 1991]. The application of the crushing technique, which allowed these researchers to study in He isotopic composition of gas-liquid inclusions, and control with the application of the He signature in the Ne isotopic composition made it possible to elucidate the issue of the occurrence of the cosmogenic component.

This publication presents our newly obtained data on the He isotopic composition of mantle xenoliths in basalts from the Vitim Plateau that allowed us to conduct comparative analysis of earlier and recently obtained results.

Brief Description of the Samples

Helium isotopic systematics was studied in two samples. Sample 1069 was taken up in the Shlak-Gora opencast mine at a distance of 75 km from the Romanovka–Bagdarin road (Figure 1). In this quarry, cinder of a volcanic cone is mined. Since this mine could hardly be started to develop earlier than 300 years ago, the exposure time of the xenoliths could not be longer than this. The sample itself has an ellipsoidal shape close to 14 cm along its major axis and is enclosed into a tuff rind. It is important to mention that the samples analyzed by Staudacher and Allegre [1991] were taken in the same mine, and this allows us to compare the data of these researchers and ours.

Sample 1065 was collected from a flow of black aphanitic basalt on the left-hand side of the Bol'shoi Amalat River, 300 m downstream of the bridge across this river on the Romanovka–Bagdarin road (Figure 1).

The modal compositions of samples 1065 and 1069 place them into the spinel lherzolites and garnet-spinel facies, respectively. The chemical composition of the xenoliths and their host basalt are presented in Table 1.

Isotopic geochemical data on a representative collection of xenoliths from basalts from the Vitim Plateau were published in [Ashchepkov, 1991; Ionov et al. 2005].

As follows from Figure 2, the composition of the mantle beneath the Vitim Plateau is strongly heterogeneous, as is typical of the continental mantle as a whole [Grachev, 1989]. Although one of our xenoliths (sample 1069) has a composition closer to the primitive mantle in terms of $\text{MgO}-\text{Al}_2\text{O}_3$ proportion (Figure 2), both of the xenolith samples and the host basalt define a compact field of data point in a Sr–Nd diagram (Figure 3), with this field corresponding to a source of the OIB type. It is worth mentioning that this field also includes the composition of glass from the partial melting zone of the xenolith of sample 1069. The diagram also shows data from [Ionov et al., 2005], Sr–Nd characteristics of the samples, and their He isotopic systematics from [Staudacher and Allegre, 1991].

Methods

Monomineralic fractions were separated from the xenolith samples with the use of heavy liquids and subsequent magnetic separation of minerals. If needed, the concentrates were additionally purified to 95–99% purity by hand-picking. He was extracted from minerals by means of melting [Kamensky et al., 1990] and by crushing the samples [Ikorskii and Kamensky, 1998] at the Laboratory of Geochronology

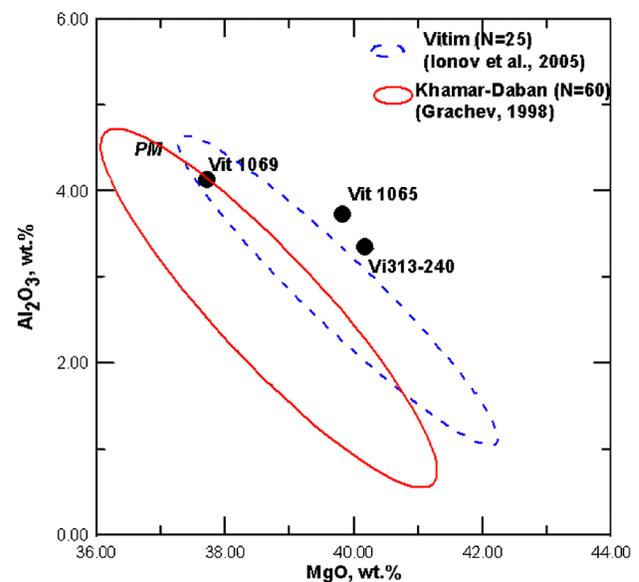


Figure 2. MgO– Al_2O_3 diagram for the studied xenoliths in comparison with the fields of the Vitim [Ionov et al., 2005] and Khamar-Daban xenoliths suites [Grachev, 1998].

Table 1. Chemistry (wt. %) of the studied ultramafic xenoliths, minerals and basalt

| Sample | 1065 | | | | 1069 | | | | 1069 |
|--------------------------------|-------|-------|-------|-------|-------|-------|-------|-------|--------|
| | Wr | Ol | Cpx | Opx | Wr | Ol | Cpx | Opx | basalt |
| SiO ₂ | 43.74 | 40.25 | 52.08 | 54.40 | 43.41 | 40.38 | 52.71 | 55.24 | 48.82 |
| TiO ₂ | 0.20 | – | 1.39 | 0.27 | 0.25 | – | 0.48 | – | 2.57 |
| Al ₂ O ₃ | 3.73 | – | 4.30 | 3.97 | 4.13 | – | 5.13 | – | 13.38 |
| FeO | 8.47 | 9.98 | 5.39 | 7.03 | 11.08 | 8.24 | 2.70 | 3.66 | 13.74 |
| MnO | 0.25 | 0.35 | 0.15 | – | 0.12 | 0.35 | 0.34 | 6.02 | 0.19 |
| MgO | 39.81 | 49.39 | 16.18 | 32.81 | 37.72 | 49.88 | 15.84 | 0.30 | 8.63 |
| CaO | 2.18 | 0.13 | 17.87 | 0.88 | 3.63 | 0.15 | 20.41 | 33.37 | 7.98 |
| Na ₂ O | 0.44 | – | 1.79 | 0.03 | 0.65 | – | 1.26 | 0.68 | 2.79 |
| K ₂ O | 0.17 | – | – | – | 0.04 | – | – | – | 1.56 |
| P ₂ O ₅ | 0.12 | – | – | – | 0.01 | – | – | – | 0.44 |
| Cr ₂ O ₃ | 0.29 | 0.87 | – | 0.53 | nd | – | 1.15 | 0.25 | 0.05 |
| NiO | 0.11 | – | – | – | nd | – | – | – | – |
| N | 1 | 10 | 6 | 3 | 1 | 6 | 6 | 6 | 1 |

and Geochemistry of Isotopes at the Geological Institute of the Kola Research Center, Russian Academy of Sciences. The crushing method made it possible to selectively extract gases from fluid inclusions and thus minimize the effect of radiogenic gases that were accumulated in the crystal structures of minerals [Kaneoka *et al.*, 1978]. In extracting the gases, 0.16–0.25 g of the material were placed, together with steel rolling crushers, into a glass ampoule, which was then evacuated and welded. The material was then crushed by the vibration of the ampoule. The He isotopic composition and concentrations were measured on a MI-1201 mass spec-

trometer (no. 22–78), the He sensitivity was 5×10^{-5} A/torr. The concentrations were calculated from the measured peak height accurate to 5% ($\pm 1\sigma$), the isotopic ratios were measured accurate to $\pm 20\%$ at ${}^3\text{He}/{}^4\text{He}$ ratios of $n \times 10^{-8}$ and $\pm 2\%$ at ${}^3\text{He}/{}^4\text{He}$ ratios of $n \times 10^{-6}$. Blanks were conducted after reloading, under the same conditions under which the samples were analyzed.

He and Ar Isotopic Composition

Table 2 and Figure 4, Figure 5 show the results of our analysis of the ${}^3\text{He}/{}^4\text{He}$ and ${}^{40}\text{Ar}/{}^{36}\text{Ar}$ ratios in olivine and clinopyroxene with the application of the melting and crushing techniques. For comparison, Figure 1 exhibits the earlier results [Staudacher and Allegre, 1991], which were obtained by the melting technique for the whole-rock composition of xenoliths and, in a single instance, of olivine and pyroxene, with the olivine also studied by both the melting and the crushing techniques (sample Vi313-240). The R/R_a ratios of the samples studied by crushing are almost indistinguishable (within the analytical error) and correspond to the MORB source. For example, olivine and orthopyroxene from sample 1069 were studied by the crushing technique and yielded R/R_a ratios of 7.5 and 7.3, respectively, and clinopyroxene from the same sample gave an R/R_a ratio of 7.9 (Table 2).

The R/R_a ratio of sample Vi313-240 (7.82) obtained using crushing [Staudacher and Allegre, 1991] is virtually indistinguishable from or results on the monomineralic separates (Figure 3). It is important to emphasize that the R/R_a ratios of all of these samples are practically independent of their ${}^4\text{He}$ concentrations (Table 2, Figure 3).

At the same time, clinopyroxene from sample 1069 yielded (with the use of the melting technique) an R/R_a ratio of 10.1. Utilizing this technique, Staudacher and Allegre [1991] obtained a value of $R/R_a = 12.6$ for sample Vi313-240. In both instances, the R/R_a ratios obtained by the melting and

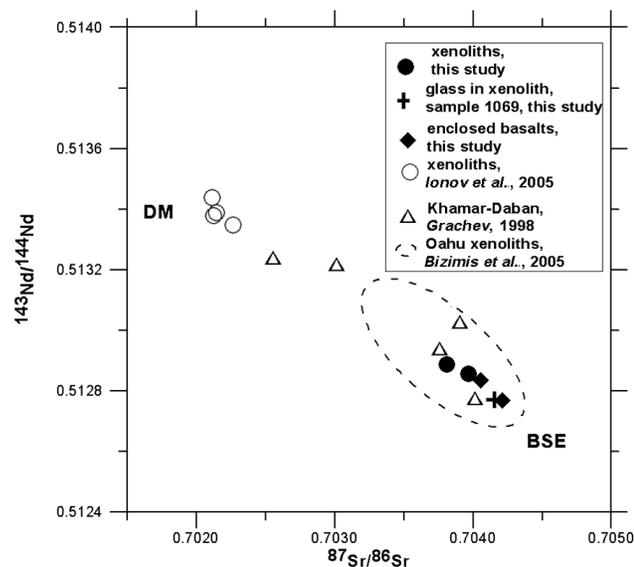


Figure 3. Sr-Nd diagram for the studied xenoliths in comparison with the fields of the Vitim [Ionov *et al.*, 2005], Khamar-Daban [Grachev, 1998] and Oahu Island xenoliths suites [Bizimis *et al.*, 2005]. DM – depleted mantle, BSE – chondritic uniform reservoir [Zindler, Hart, 1986].

Table 2. He-Ar isotopes (10^{-6} cm³/g) of ultramafic xenoliths and enclosed basalt

| No. | Sample | Rock, mineral | Weight, g | ⁴ He | ³ He/ ⁴ He (<i>R/R_a</i>) | ⁴⁰ Ar | ⁴⁰ Ar/ ³⁶ Ar | Method |
|-----|-------------|-----------------|-----------|-----------------|---|------------------|------------------------------------|----------|
| 1 | 1065 | Ol | 1.1844 | 0.0061 | 9.6 ± 5 | 0.70 | 317 ± 5 | melting |
| 2 | 1065 | Ol, leaching | 0.8530 | 0.0044 | 26.8 | | | melting |
| 3 | 1065 | Py | 0.9580 | 0.016 | 4.6 ± 2 | 0.68 | 359 ± 5 | melting |
| 4 | 1069 | Ol | 1.3673 | 0.204 | 10.1 ± 1 | 0.27 | 377 ± 5 | melting |
| 5 | 1069 | Ol | 1.8 | 0.034 | 7.5 | – | – | crushing |
| 6 | 1069 | Opy | 2.0 | 0.14 | 7.3 | – | – | crushing |
| 7 | 1069 | CPy | 1.6 | 0.31 | 7.9 | 0.12 | 325 ± 7 | crushing |
| 8 | 1069 basalt | | 0.3898 | 0.51 | 2.7 ± 0.2 | 3.5 | 312 ± 2 | melting |

crushing techniques for the same samples were 1.34 and 1.7.

The question now arises as to how to explain the value of $R/R_a = 26.8$ for olivine from sample 1065 (melting technique) (Table 2). Even with a correction for cosmogenic He, the R/R_a ratio of 15.7 is much higher than the ratio of MORB (8 ± 1), as follows from the comparison of the values obtained by the crushing and melting techniques.

A correction for cosmogenic He can be introduced if data are available on two minerals from a single xenolith [Dunai and Baur, 1995]. If the R/R_a ratio is equal in these two minerals but they have different He concentrations, then cosmogenic He is absent. This is the case with sample 1069, whose R/R_a ratios are similar and the He concentrations are different (Table 2). Hence, an average R/R_a ratio of 7.6 can be taken up for sample 1069.

For sample 1065, whose olivine and pyroxene yield different R/R_a ratios, a correction for $R/R_{a\text{cosm}}$ can be derived

from the relation $(^3\text{He}_{\text{min 1}} - ^3\text{He}_{\text{min 2}})/(^4\text{He}_{\text{min 1}} - ^4\text{He}_{\text{min 2}})$ [Dunai and Baur, 1995] and is $R/R_{a\text{cosm}} = 9.5$. This value (beyond the analytical error) differs from the value for sample 1069 and is significantly higher than the typical MORB values ($R/R_a = 8 \pm 1$).

Another approach to evaluating the effect of cosmic irradiation of the samples makes use of studying the Ne isotopic composition, as was done in [Staudacher and Allegre, 1991]. Sample Vi313-240, which has the highest R/R_a ratio (see Figure 4), $^{20}\text{Ne}/^{22}\text{Ne} = 8.13$, and $^{21}\text{Ne}/^{22}\text{Ne} = 0.1880$. These values testify that the Ne is cosmogenic and does not contain either the solar or the MORB component [Staudacher and Allegre, 1991]. These researchers obtained other $^{20}\text{Ne}/^{22}\text{Ne}$ and $^{21}\text{Ne}/^{22}\text{Ne}$ values, close to the atmospheric ones, for other samples.

The $^{40}\text{Ar}/^{36}\text{Ar}$ ratio ranges from 400, i.e., a value close to the atmospheric one (295.5), to 10,000 (Figure 4) but does not reach the maximum value for MORB (28,000). The

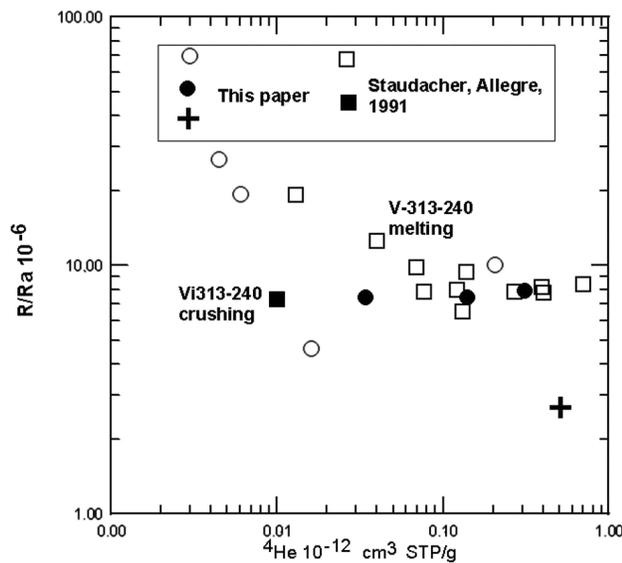


Figure 4. $^3\text{He}/^4\text{He}$ (R/R_a) – ^4He diagram for xenoliths in comparison with early published data [Staudacher, Allegre, 1991]. Filled circles and squares for gas released by melting procedure, open ones refer to crushing method data. Cross – enclosed basalt.

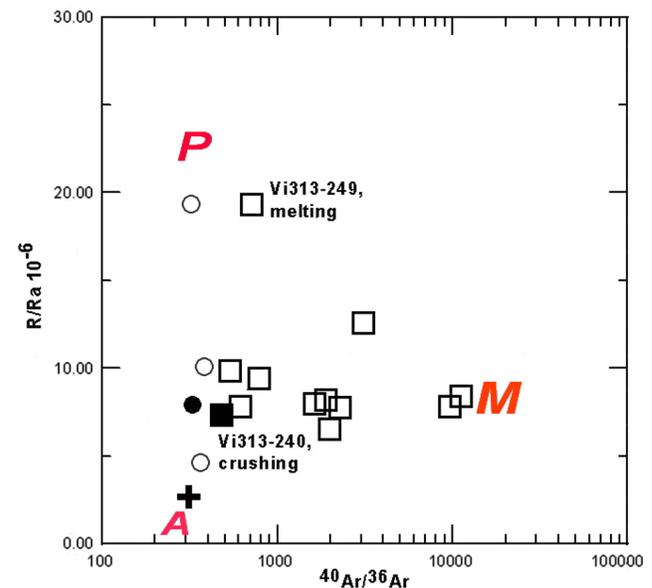


Figure 5. $^3\text{He}/^4\text{He}$ (R/R_a) – $^{40}\text{Ar}/^{36}\text{Ar}$ diagram for xenoliths. A – Atmosphere, M – MORB, P – Plume [Kaneoka, 1998]. The other signs are as in Figure 5.

Table 3. Sm-Nd and Rb-Sr isotopes of ultramafic xenoliths and basalts

| No. | Sample | Sm | Nd | Rb | normalise Sr | $^{147}\text{Sm}/^{144}\text{Nd}$ | $^{143}\text{Nd}/^{144}\text{Nd} \pm 2\sigma$ | $^{87}\text{Rb}/^{86}\text{Sr}$ | $^{87}\text{Sr}/^{86}\text{Sr} \pm 2\sigma$ |
|-----|--------------|-------|-------|-------|--------------|-----------------------------------|---|---------------------------------|---|
| 1 | 1065, Di | 0.927 | 2.597 | 0.749 | 56.44 | 0.21579 | 0.512887 ± 10 | 0.03839 | 0.703806 ± 24 |
| 2 | 1069, Di | 7.11 | 22.42 | 0.466 | 217.9 | 0.19156 | 0.512857 ± 6 | 0.00618 | 0.703963 ± 18 |
| 3 | 1069, glass | 6.483 | 29.43 | 20.13 | 572.2 | 0.13314 | 0.512770 ± 3 | 0.10171 | 0.704152 ± 17 |
| 4 | 1065, basalt | 9.67 | 48.49 | 28.96 | 1155 | 0.12056 | 0.512836 ± 6 | 0.0725 | 0.704054 ± 17 |
| 6 | 1069, basalt | 9.51 | 45.19 | 38.51 | 1169 | 0.12713 | 0.512768 ± 4 | 0.09525 | 0.704209 ± 14 |

Note: Sm, Nd, Rb and Sr concentrations in ppm (isotope dilution, precision is about 0.5%), error estimations on $^{147}\text{Sm}/^{144}\text{Nd}$ and $^{87}\text{Rb}/^{86}\text{Sr}$ are $\pm 0.3\%$ and 0.5% respectively. During the period of analytical work the weighted mean of 10 La Jolla Nd standart runs yielded 0.511852 ± 4 (2σ), using 0.241579 for $^{143}\text{Nd}/^{144}\text{Nd}$ to normalize; and NBS-987 standard yielded 0.710255 ± 15 (2σ), using 8.375210 for $^{87}\text{Sr}/^{86}\text{Sr}$ to normalize. Total procedure blanks for Nd and Sm are 0.08 and 0.03 ng respectively, and for Sr and Rb are 0.3 and 0.4 ng respectively. All isotopic analyses were carried out on the Finningan MAT-261 solid source machine under multicollector static mode. Chemical preparation of rock samples and elements separation were done using standart procedure similar to [Richard *et al.*, 1976] in the Precambrian Geology and Geochemistry Institute (St. Petersburg).

R/R_a and $^{40}\text{Ar}/^{36}\text{Ar}$ ratios are not correlated, as is also typical of MORB [Staudacher *et al.*, 1989].

The Sr-Nd isotopic systematics of xenolith samples 1065 and 1069 and the sample of the host basalt (sample 1069) define a compact swarm of data points in the compositional field of the Oahu, Hawaii lherzolites. This swarm is close to the BSE source in the Sr–Nd diagram (Figure 3). Comparing the data of Table 2 and Table 3, one can readily see that the xenoliths do not display any correlations between the He isotopic ratios and analogous Sr and Nd ratios, and these systems behave independently in the xenoliths.

Discussion

Our newly obtained data on the He isotopic composition of mantle xenoliths in basalts from the Vitim Plateau are generally consistent with the results reported in [Staudacher and Allegre, 1991], but the higher R/R_a ratio (higher than the typical MORB value) of one of the xenolith samples is significant and suggests a contribution from a source of the P (Plume) type. The occurrence of a mantle plume beneath the Vitim Plateau was independently inferred from isotopic geochemical data on the basalts [Johnson *et al.*, 2005] and from geophysical evidence [Zorin, 2003].

Certain issues remain unresolved concerning the effect of the cosmic irradiation of the samples when they resided for a long time on the surface. Is the implanted He somehow related to the composition of the minerals? Kurz [1986] believed that the production of spallogenic He can only insignificantly depend on the composition of minerals, but data derived in the course of our long-term studies of He isotopic composition of xenoliths [Grachev and Kamensky, 2005] are at variance with this conclusion. Moreover, it has long been known that mantle xenoliths can contain at least two olivine populations (porphyroblasts and neoblasts), which were produced in situ by solid-plastic deformations in the presence of volatiles before the xenolith was entrapped by melt [Mercier and Nikolas, 1975]. Such processes, which are associated

with partial melting and the development of glass zones in xenoliths, result in the loss of primary He by means of its diffusion. The diffusion rate of He through olivine at 1240°C is 1.8 mm/year [Hart, 1984], which is close to the size of olivine grains in xenoliths. Thus, the disturbance of equilibrium in the He isotopic system before the xenoliths were brought to the surface could be one of the factors that affected the He isotopic signature.

It is important to stress that there are still no reliable information on the R/R_a ratio of mantle xenoliths of the continental sublithospheric mantle, whereas ultramafic xenoliths in basalts at oceanic islands (Hawaii, Reunion, and Samoa) usually display $^3\text{He}/^4\text{He}$ ratios 20–30 times higher than the atmospheric ones [Poreda and Farley, 1992; Rison and Craig, 1983; and others]. At the same time, continental basalts, such as at the Deccan [Basu *et al.*, 1993] or Greenland [Graham *et al.*, 1998], have R/R_a ratios much higher than those of MORB (more than $30R/R_a$).

A probable reason for this is the fact that the continental lithospheric mantle, which is thicker than beneath oceans, has a more complicated evolutionary history. The U-Pb zircon dates of xenoliths from the Vitim Plateau group within the following are ranges: 1835–1955, 1562–1506, 264–295, and 164–183 Ma, with zircons of each of these age groups differing in composition and REE patterns [Saltykova, 2008].

It is also worth mentioning that mantle xenoliths usually exhibit evidence of their partial melting (triple junction boundaries of mineral grains). As is demonstrated in experiments on acid etching of olivine grains, this process results in the enrichment of the olivine in radiogenic He, because the etching zones are always enriched in U [Grachev and Komarov, 1994].

It is thus uncertain as to whether ultramafic xenoliths from the continental mantle have He isotopic signatures similar to those of xenoliths in lavas from oceanic islands.

Although strongly degassed (i.e., devoid of its primary ^3He , Table 3), the basalt hosting the mantle xenoliths displays Sr and Nd isotopic parameters (Table 3) close to those of the OIB and BSE sources. It should be stressed that the isotopic composition of glass the in the partial melting zone

in one of the xenoliths (sample 1069, Table 3, Figure 3) is also close to those of the OIB and BSE sources [Zindler and Hart, 1986], and this furnishes further arguments in support for a plume nature of volcanism at the Vitim Plateau [Johnson et al., 2005; Zorin et al., 2003].

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