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Helium isotopes on the Pacific-Antarctic ridge (52.5°–41.5°S)

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1. Introduction

[1] The first isotopic data and concentrations of helium are reported for the Pacific-Antarctic ridge between 52.5°S and 41.5°S. The ⁴He/³He ratio is extremely homogeneous over more than 1200 km, with a mean ratio of 99,275 (R/Ra = 7.29) and a standard deviation of 2719 (0.19), which is the lowest dispersion observed for the global mid oceanic ridge system. Moreover, the Menard T.F. is a frontier between two mantles with slightly different helium isotopic ratios (96,595 ± 1520 and 100,347 ± 2330). No difference in the helium concentration between the two ridge segments defined by the Menard T.F. can be observed, as well as no significant difference in the U and Th contents suggesting that the difference in helium isotopic ratio is old (>500 My) and may represent a slight difference in degassing or/and trace element depletion history. Citation: Moreira, M. A., L. Dosso, and H. Ondréas (2008), Helium isotopes on the Pacific-Antarctic ridge (52.5°–41.5°S), Geophys. Res. Lett., 35, L10306, doi:10.1029/2008GL033286.

2. Sample Locations and Analytical Procedure

[4] PACANTARCTIC2 was a joint geophysical survey and geochemical sampling cruise to the Pacific-Antarctic Ridge between 41°15’S and 52°45’S. From December 17, 2004 to January 17, 2005, the French research vessel L’Atalante surveyed 1300 kilometers of the Pacific-Antarctic Ridge axis and made 3200 km of off-axis profiles to describe the structures located on either side of the Menard T.F. and west of the axis at 42°S. During the cruise, 43 dredges were performed among which 24 were on axis. We are focusing on these on-axis dredges on this paper [Dosso et al., 2005] (Figure 1). Details about ridge structure and sample chemistry are given by Klingelhoefer et al. [2006] and Hamelin et al. [2007].

[5] Pieces of glass, ~0.5 cm in size, were cleaned with distilled water, ethanol and acetone using an ultrasonic bath. Some samples were also cleaned with hydrogen peroxide in order to remove some Mn crust. Twelve glass samples with weights between 0.1 and 0.7 g were loaded in a glass sample tree connected to the Staudacher-type glassy High frequency furnace equipped with a Mo crucible. Samples were melted at ~1400°C in order to degas. Gas was purified with two titanium getters at 800°C during 10 minutes each. Furnaces were then stopped to decrease the Ti getter temperature to room temperature during 10 minutes. H₂ was purified with two SAES getters at room temperature. Rare gases were then trapped at 10 K on charcoal trap. Helium was desorbed at 25 K and then analyzed. Helium concentrations and the isotopic ratios were measured with our mass spectrometer ARESIBO II, equipped with a Faraday cup for ⁴He and an electron multiplier connected to an ion counting system for ³He. An aliquot of the helium was generally kept and analyzed immediately after the first helium measurement to confirm this analysis. In all cases, the duplicate gave the same isotopic ratio (within ±0.1 on the R/Ra ratio). Blanks were analyzed before each sample by heating the furnace at the same temperature as the samples. The ⁴He blank was ~10^{-8} ccSTP, which generally corresponds to less than 1% of the measured helium. For these analyzes, we have modified our old helium procedure.
for which $^3$He was measured at mass 3.016. $^3$He is now measured at mass 3.014, where the HD+ correction is negligible for low H2 signals. The H2 signal was always less than 15000 cps leading to a negligible contribution of HD+ at mass 3.014. This contribution is generally 3–5 cps at mass 3.016 and <0.1 at mass 3.014. Helium standards were repeated between samples to calibrate the mass discrimination and the sensitivity of the mass spectrometer. The helium standard is a gas collected in a spring source from Réunion Island (Indian Ocean), with a $^{4}$He/$^3$He ratio of 56,980 ± 450 (R/Ra = 12.68 ± 0.10). In order to propagate the uncertainties, the standard deviation of all the measured standards during the analysis period was considered. Moreover, samples previously analyzed by

Figure 1. Sample locations. Numbers inside circles are dredge numbers (e.g., DR5 and DR6). S1-3 and N1-5 are segment numbers as defined by Klingelhofer et al. [2006] based on the ridge morphology. Insert represents the studied area in the South Pacific. The figure is adapted from Dosso et al. [2005].
3. Results

Helium concentration and isotopic ratios are given in Table 2 for PACANTARCTIC 2 samples (PAC2). \( ^4 \text{He} \) concentrations vary between 1.9 \( \times 10^{-8} \) (DR06-6g) and 2.1 \( \times 10^{-7} \) ccSTP/g (DR33-3g). Except for samples DR06-6g and DR27-1g, the helium concentrations are typical of MORB concentrations [Honda and Patterson, 1999; Moreira and Sarda, 2000; Sarda and Moreira, 2002]. \( ^4 \text{He}/^3 \text{He} \) ratios vary from 95,071 to 185,268 \( \pm 52,255 \). Except for the latter very high, and un-precise ratio, the highest helium isotopic ratio is observed for sample DR27-1g (118,450). These two radiogenic ratios correspond to the poorest samples in helium. The \( ^4 \text{He}/^3 \text{He} \) variation with latitude is represented in Figure 2 (with the exception of samples DR06-6g and DR27-1g). The helium isotopic ratios are similar to the mean MORB ratio, although slightly more radiogenic. The mean PAC2 ratio is 99,725 \( \pm 2720 \) (without samples DR06-6g and DR27-1g for which we will discuss below the origin of their ratios), compared to the mean MORB ratio of 90,000 [Allègre et al., 1995]. Good reproducibility within a single dredge is also observed as illustrated in samples DR33-2g and DR33-3g that show \( ^4 \text{He}/^3 \text{He} \) ratios of 99,524 \( \pm 1645 \) and 98,843 \( \pm 1487 \) respectively.

4. Discussion

4.1. Radiogenic Samples DR6-6 and DR27-1

[7] Samples DR6-6g and DR27-1g present low helium concentration for MORB samples (1.9 \( \times 10^{-8} \) and 1.9 \( \times 10^{-7} \) ccSTP/g respectively) as well as radiogenic helium isotopic ratio (185,268 \( \pm 52255 \) and 118,450 \( \pm 7767 \)) compared to other MORB from the area. The Th and U concentrations are 0.95 and 0.30 ppm for DR6-6g and 1.41 and 0.43 ppm for DR27-1g respectively. These concentrations are relatively high for MORB samples and also higher than found in MORB from the area [Hamelin et al., 2007]. These latter facts are related to the observation that these two samples are more evolved than the other samples from the area Indeed, these two samples present higher alkalis and SiO\(_2\) contents and lower MgO than the other [Hamelin et al., 2007].

[8] Considering the \( ^4 \text{He} \) production equation we can derive an “eruption age” if we assume that the radiogenic helium is post eruptive. Samples were melted and therefore the helium is the sum of mantle and radiogenic helium present in the sample. With an initial \( ^4 \text{He}/^3 \text{He} \) ratio similar to the mean ratio of the other MORB from this area, (99,730), and using the measured U and Th content, we get 140,000 and 330,000 years for samples DR6-6g and DR27-1g respectively, suggesting that the post-degassing
radiogenic ingrowth is responsible for the radiogenic isotopic ratios. Note that these ages are minimum since helium diffusion can occur. In the following discussion, we will then ignore these two samples.

4.2. Mean Isotopic Ratio: The “Normal” MORB Ratio?

The homogeneity of the helium isotopic ratio over more than 1200 km of mid oceanic ridge is clear. This part of the global ridge system is the most homogeneous for the helium isotopic ratio [Alle`gre et al., 1995; Georgen et al., 2003]. No oceanic island chain and topographic anomaly are present in the area. The closest hotspot that can be attributed to a mantle plume is the Foundation seamount chain, which crosses the Pacific Antarctic ridge at ∼37.5°S [Maia et al., 2005]. We can therefore assume that the mean ratio determined on the PAC2 samples represent the “normal” helium isotopic ratio for the pacific MORB source mantle.

4.3. Difference Between South and North of Menard Transform Zone

Figure 2 shows that the Menard T.F. is a natural boundary of the mantle convection. A difference in the \(^{4}\)He/\(^{3}\)He ratio can be observed. The mean isotopic ratios are 96,595 ± 1520 (1s) and 100,347 ± 2330 (1s) for the south and for the north zones of Menard T.F., respectively (samples DR6-6g and DR27-1g are not considered). In order to evaluate the possibility that the difference in the mean helium isotopic ratio between the two ridge segments reflects different mantle histories, including degassing or depletion during melting, we have developed a simple mantle evolution model that considers helium degassing and radiogenic production of \(^{4}\)He. The equations that govern the evolution of \(^{3}\)He, \(^{4}\)He and \(^{4}\)He/\(^{3}\)He are the following:

\[
d^{3}\text{He} \frac{d}{dt} = +\phi^{3}\text{He}
\]

\[
d^{4}\text{He} \frac{d}{dt} = +\phi^{4}\text{He} - 8\lambda_{8}^{238}\text{U}e^{\lambda_{8}^{238}\text{U}t} - 7\lambda_{5}^{235}\text{U}e^{\lambda_{5}^{235}\text{U}t} - 6\lambda_{2}^{232}\text{Th}e^{\lambda_{2}^{232}\text{Th}t}
\]

\[
R(\tau) = R_{A} - 8\lambda_{8}^{238}\text{U}_{A}\left[e^{(\lambda_{8}^{238}\text{U} - \phi)\tau} - 1\right] + 7\lambda_{5}^{235}\text{U}_{A}\left[e^{(\lambda_{5}^{235}\text{U} - \phi)\tau} - 1\right] + 6\lambda_{2}^{232}\text{Th}_{A}\left[e^{(\lambda_{2}^{232}\text{Th} - \phi)\tau} - 1\right]
\]

where \(R(\tau)\) is the isotopic ratio, \(R_{A}\) is the present-day isotopic ratio and \(\Phi\) is a constant (degassing constant in \(\text{yr}^{-1}\)). These equations are given as a function of age (\(\tau\)). The letter A stands for present.

[11] The global \(^{3}\)He flux at ridges is 1000 mol/year [Farley et al., 1995], which corresponds to \(d^{3}\text{He}/dt\). Therefore, knowing the mass of the degassing mantle, and the mean concentration of \(^{3}\)He in the mantle, one can get the mean degassing rate of the mantle \(\Phi\). Assuming that the popping rock 2πD43 reflects undegassed magma [Moreira et al., 1998], and that the melting rate is \(\sim 10\%\) one can get \([^{3}\text{He}] \sim 10^{-10} \text{cc/g}\). Therefore, \(\Phi = 2.2 \times 10^{-10} \text{yr}^{-1}\) if the degassing mantle is the upper mantle (\(n = 10^{23} \text{g}\)). If one considers that the popping rock 2πD43 is an exceptionally gas rich sample, helium concentration in the mantle should be lower. Therefore, we also did the calculations with a \([^{3}\text{He}] = 5 \times 10^{-11} \text{cc/g}\). We assume the
degassing constant is the same for each ridge segment, including south and north of Menard T.F. as this constant reflects only the oceanic crust formation rate. Two isotopic ratios, at identical ages, to be different, have to have different U/3He ratio. In order to get separation of the two mantles since 500 My (Figure 3), the difference in the U/3He ratio should be 1.38 or 1.26 depending of the helium concentration. If the separation occurred 1 Ga ago, smaller differences are required. For the calculation, the Th/U ratio was taken as 2.66 and 2.31 for the south and north segments respectively. These values are the mean value of all the samples. The mean uranium content appears to be the same (0.11 and 0.10 ppm). Using 10% melting we can estimate the U content of the mantle as 10 ppb.

[12] The most radiogenic mantle was probably either more degassed or less depleted in U by a factor of 1.2–1.4, even if it is not clear from the data, especially for helium. Helium concentration in MORB is not directly proportional to the inverse of melting rate like for non-volatile elements, since degassing at the surface is susceptible to change the helium content by a few orders of magnitudes [Moreira and Allègre, 2002]. Neon and argon measurements are necessary to quantify the degassing process [Moreira and Sarda, 2000; Sarda and Moreira, 2002].

[13] As a matter of fact, other isotopic tracers show a difference between south and north of Menard T.F. [Hamelin et al., 2007]. For example, the \(^{87}\text{Sr}/^{86}\text{Sr}\) of the south segment is slightly less radiogenic (\(<0.7024\)) than the north segment (\(<0.7025\)), suggesting a difference in the \(^{87}\text{Rb}/^{86}\text{Sr}\) fractionation history.

4.4. Correlation Between Spreading Rate and \(^{4}\text{He}/^{3}\text{He}\) Standard Deviation

[14] Using the same methodology as described by Allègre et al. [1995] for the calculation of the standard deviation, e.g. <\(\alpha\)> can be calculated using:

\[
(n - 1) <\alpha>^2 = \sum \left[ \sigma_i^2 + (X_i - X_m)^2 \right]
\]

where \(\sigma\) is the individual uncertainty on the \(^{4}\text{He}/^{3}\text{He}\) ratio, \(X_i\) is the measured \(^{4}\text{He}/^{3}\text{He}\) and \(X_m\) is the mean ratio (weighted for uncertainties). For PAC2 samples, the mean pondered \(^{4}\text{He}/^{3}\text{He}\) ratio is 99.173 and <\(\alpha\)> = 3371. The mean spreading rate of the studied segment is 9cm/year, which is one of the fastest among all the mid oceanic ridges. Therefore, our new set of data is consistent with the correlation between <\(\alpha\)> and the inverse of the spreading rate observed by Allègre et al. [1995]. This correlation was interpreted as the result of a better mixing of the MORB source when the spreading rate is high and assumed to directly represent the vigor of the mantle convection.

5. Conclusions

[15] The Pacific-Antarctic Ridge between 41.5 and 52.5°S presents the most homogeneous helium isotopic ratio among the worldwide ridge system. This can be interpreted in two ways. The first one is that this part of the ridge is free from any mantle plume that can “influence” the asthenospheric mantle. The relatively high spreading rate reflects large melting, and therefore more homogeneous
mantle zone, than for slower mid oceanic ridges such as the South West Indian Ridge or North Atlantic. Moreover, the Menard fracture zone appears to separate two mantles with slightly different helium isotopic ratio, suggesting a different history for each of these two mantles, including degassing or incompatible element depletion (in particular U and Th) at least more than 500 My ago.

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References


