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Helium signature of the subcontinental lithospheric mantle

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Abstract

Peridotites and basalts from continental areas over the world show a helium isotopic ratio that is relatively homogeneous and more radiogenic than the MORB source. The mean ratio is $118\,000 \pm 15\,500$ ($R/Ra = 6.1 \pm 0.9$), compared to the MORB source ratio of $90\,000 \pm 10\,000$ ($R/Ra = 8 \pm 1$). This rather constant ratio indicates that the effect is worldwide and can be directly bound to subcontinental mantle geodynamics. A closed system model would imply low uranium content in the subcontinental mantle (< 2 ppb), disagreeing with the measured U concentrations in continental mantle peridotites (4–40 ppb). In this study, we evaluate a model where the continental lithospheric mantle is in steady state for helium. Helium residence time in the subcontinental mantle can therefore be estimated and gives values around 100 Ma, compatible with plate tectonic time scales and melt fluxes through the lithosphere. However, this model predicts ^3He fluxes between 110 and 1000 mol/yr depending on different U contents and depth of the lithosphere. This appears much higher than the flux observed in continental areas. © 2002 Elsevier Science B.V. All rights reserved.

Keywords: mantle; noble gases; helium; steady-state processes; models; xenoliths

1. Introduction

Noble gas systematics in oceanic basalts have given constraints on mantle evolution and the origin of the atmosphere [1]. One major conclusion was obtained from helium isotope systematics in mantle-derived material. The helium isotopic ratio is homogeneous in mid-ocean ridge basalts (MORB) with a mean ratio of $90\,000 \pm 10\,000$ ($R/Ra = 8 \pm 1$ where R is the $^3\text{He}/^4\text{He}$ ratio and Ra the atmospheric ratio) [2]. In contrast, the

$^4\text{He}/^3\text{He}$ ratios of ocean island basalts (OIB) varies from 16 800 in Iceland ($R/Ra = 42.9$) [3] to more than 160 000 ($R/Ra = 4$) in Sao Miguel island (Azores) [4]. This dispersion reflects different components in the source of plumes. The low $^4\text{He}/^3\text{He}$ ratios observed in Hawaii, Iceland or Galapagos reflect low $(\text{U}+\text{Th})/^3\text{He}$ ratios in the source [5–7]. OIB sources contain more uranium than the MORB source, indicating that the low $^4\text{He}/^3\text{He}$ ratio plume sources contain more ^3He than the MORB source. This can be interpreted as reflecting different degrees of degassing in the mantle where the MORB source is well degassed whereas the source of most mantle plumes (probably the lower mantle) contains little degassed material [1,8]. Moreover, a ‘low ^3He ’ hotspot cat-

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egory has been defined based on the helium isotopic ratio. These OIBs have high $^4\text{He}/^3\text{He}$ ratios (i.e. higher than the mean MORB ratio). Sao Miguel, Tristan da Cunha, Gough, St. Helena, Canaries, Cameroon line, Rurutu and Tubuai belong to this category [4,9–13]. Some of these OIB, such as St. Helena, Tubuai or Rurutu, have HIMU characteristics and the helium isotopic ratio may be due to the presence of old recycled oceanic crust [11]. Nevertheless, the origin of the low ^3He hotspot category is still debated [14,15]. Some authors argue for a mantle helium signature (recycling) whereas others propose that the high $^4\text{He}/^3\text{He}$ ratios reflect shallow level contamination [16].

The subcontinental lithospheric mantle (SCLM) is different from the oceanic mantle in its mineralogy and geochemistry. It is an isolated reservoir from the convective mantle and it has developed its own signature in term of major and trace elements and isotopic signature, distinct from the MORB source. Osmium isotopes give model ages higher than 1 Ga for the isolation from the convection. The mantle lithosphere is less dense than the convective mantle and therefore is stabilized under the continents permitting such old ages [17]. Trace element and isotopic studies on samples from the subcontinental mantle have shown that the subcontinental mantle has similarities with the depleted oceanic mantle but is re-enriched by fluids either from the asthenosphere or from slab dehydration added during past subductions. We examine in this study a new model for the origin of the homogeneity and greater radiogenicity than the MORB source of the helium isotopic ratio of the SCLM.

2. Helium data

In order to characterize the helium signature of the SCLM, data where helium was obtained only by crushing on olivines (phenocrysts and xenoliths) from different places were selected. There are two arguments for this choice. The crushing technique releases the gas from fluid inclusions, free of the cosmogenic and radiogenic helium produced in the matrix (^3He and ^4He respectively)

[18]. Moreover, for very low helium content ($< 10^{-9}$ cc STP/g), the helium in the matrix can be degassed by diffusion during crushing and can change the mantle isotopic signature [19]. Therefore, because olivine has no U and Th compared to the clinopyroxene (cpx) and orthopyroxene (opx), only this mineral was chosen for data selection.

Sample locations are from Europe (Massif Central, Eifel, Austria and Etna), the southwest of the USA (Sierra Nevada and Central basin), Antarctic, Australia and West Africa (Cameroon line). The data of Patterson and collaborators [20], which represent a subduction zone, are not included. Helium isotopic results from the East African and Ataq (South Yemen) xenoliths are not considered [21] because they were analyzed by melting. Continental flood basalts such as Deccan or Ethiopia traps are not considered in this study. Flood basalts are generally associated with extremely large volumes of lava erupted in a relatively small time period. For example the surface of the Deccan Traps is 1.0×10^6 km², compared to 7.5×10^4 km² for *all* the European volcanism. Flood basalts are considered to be the plume head arriving at the surface [22,23].

Fig. 1 shows the R/R_a ratio as a function of the ^4He concentration, and a histogram of the helium isotopic ratios for continental mafic material. The peridotites and the alkali basalts show a relatively homogeneous isotopic helium ratio as well as a large dispersion of the helium content. The mean $^4\text{He}/^3\text{He}$ ratio is $118\,000 \pm 15\,500$ ($R/R_a = 6.1 \pm 0.9$) for 114 samples. Moreover, if we also consider the published data obtained by melting and crushing on olivines, cpx and opx, the mean helium isotopic ratio becomes $109\,400 \pm 42\,200$ (or $R/R_a = 6.6 \pm 1.8$) for 343 data [13,21,24–32]. The larger dispersion is due to the cosmogenic and radiogenic component released from the matrix during melting that increases the natural variability [21,25,33]. The mean helium isotopic ratio of the crushed samples is 118 000. This ratio is relatively radiogenic compared to the MORB source ratio ($^4\text{He}/^3\text{He} = 90\,000 \pm 10\,000$ or $R/R_a = 8 \pm 1$ [2]). This observation was previously made by Dunai and Baur and Porcelli et al. [21,25,33]. The helium contents in olivines (xenoliths and

phenocrysts) are between 10^{-10} and 10^{-6} cc STP/g. There is no obvious difference between xenoliths and phenocrysts in helium content (Fig. 1).

3. Discussion

The origin of the continental volcanic provinces is still debated and especially the origin of European volcanic areas. Several authors argue for a mantle plume at the source of this volcanism [34–39]. Based on helium isotopes, a lower mantle origin can be excluded for European volcanism (France, Austria, Hungary and Germany). However, the existence of the ‘low ^3He ’ hotspot category (i.e. higher than MORB $^4\text{He}/^3\text{He}$ ratios) keeps a plume origin for the European volcanism possible, at least from the 670 km boundary. Fig. 2 shows the Pb isotope composition of the European volcanic areas (e.g. Massif Central, Eifel, Etna), compared to MORB and OIB from the Pacific and Atlantic. The European samples fall within the oceanic basalt field with a $^{206}\text{Pb}/^{204}\text{Pb}$ ratio close to 19.5 suggesting a plume origin for some authors. Fig. 2 shows that this volcanism should be considered homogeneous over a large province (>1600 km). A similar observation has already been pointed out by Hoernle et al. [36] and interpreted as reflecting the composition of a mantle upwelling and not the lithospheric mantle composition. If this component with a $^{206}\text{Pb}/^{204}\text{Pb}$ ratio of ~ 19.5 derives from the normal oceanic mantle, the model age of this isolation is between 2 and 2.5 Ga (based on the $^{207}\text{Pb}/^{204}\text{Pb}$ – $^{206}\text{Pb}/^{204}\text{Pb}$ diagram). Such a long time is possible in the lithospheric mantle and is compatible with

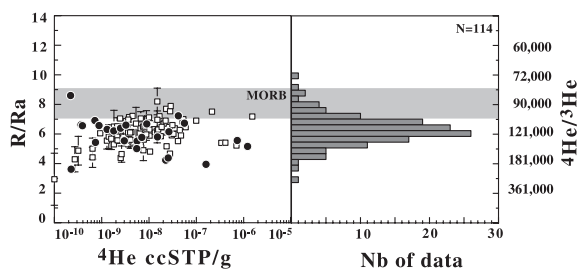


Fig. 1. Representation of the helium isotopic ratio as a function of the ^4He concentration and histogram of the helium isotopic ratios measured in phenocrysts (empty squares) and xenocrysts (black circles) from continental areas. References for data from literature are given in Table 1.

osmium-derived ages. Moreover, some relatively high $^{206}\text{Pb}/^{204}\text{Pb}$ ratios (>19) exist in the subcontinental mantle as shown in orogenic peridotite massifs such as Ronda or Beni Bousera (Spain and Morocco) [40]. No plume is invoked for these ‘high’ $^{206}\text{Pb}/^{204}\text{Pb}$ ratios and therefore the volcanism should be considered a witness of the European–North African subcontinental mantle.

The xenoliths from Australia and Europe have a lithospheric origin and therefore helium in fluid inclusions probably represents lithospheric mantle composition [38,41–44]. One can argue that the helium in the peridotite inclusions may derive from the host magma itself by contamination. Because the melt also certainly derives from the lithospheric mantle, the helium isotopic signature in xenoliths has to reflect the isotopic signature of the SCLM.

Therefore we will consider this peculiar helium isotopic signature as reflecting the lithospheric mantle without invoking any mantle plumes, especially from the lower mantle where the helium

Table 1
References for the data used in Fig. 1

Authors	Location	Sample type
Marty et al. [31]	Italy	phenocrysts
Reid and Graham [28]	Southwestern USA	phenocrysts
Dodson et al. [29]	Western USA	phenocrysts
Matsumoto et al. [30]	Australia	xenoliths
Ackert et al. [32]	Antarctic	phenocrysts
Barfod et al. [13]	Cameroon	phenocrysts and xenoliths
Matsumoto et al. [27]	Australia	xenoliths
Gautheron and Moreira [24]	Europe	xenoliths

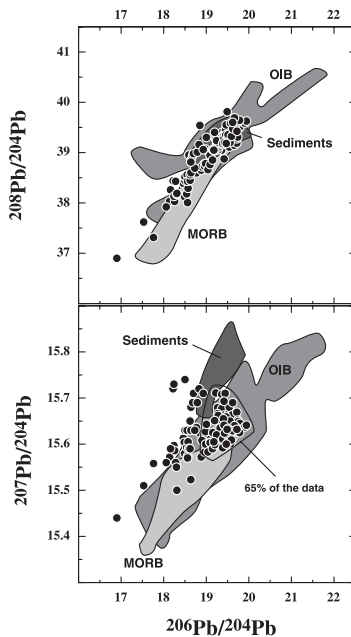


Fig. 2. $^{208}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ diagram for cpx from xenoliths, phenocrysts and whole rocks from the European volcanic provinces. Germany [34,60,61]; Carpathian–Pannonian [38]; southwest Poland [62]; Italy [63]. The fields for the Atlantic and Pacific MORB, Atlantic and Pacific OIB and the sediments are also reported.

isotope ratio $^4\text{He}/^3\text{He}$ is completely different ($< 30\,000$; $R/\text{Ra} > 20$) than the observed ratio of $118\,000$ ($R/\text{Ra} \sim 6.1$).

3.1. Previous models for the helium isotopic ratio in the subcontinental mantle

To explain the helium isotopic ratio of the subcontinental mantle, a mixing between a MORB-like material and a radiogenic component (e.g. sediments from subducted slab) [25] has been invoked, or a closed system evolution [28,29]. The mixing model would imply a much greater heterogeneity in the helium isotopic ratios with probably measurements of more radiogenic isotopic ratios because mixing is not efficient in the lithospheric mantle. A closed system evolution is also difficult to consider for two reasons. Different degrees of degassing and ages of isolation of the different subcontinental lithospheres are not in agreement with the constant helium isotopic ratio over the

world. The second argument is the uranium content that is necessary. The closed system model implies a $\text{U}/^4\text{He}$ ratio near $1000\text{--}2000$ (U in ppm and ^4He in cc STP/g) [28,29]. Therefore, the U concentration for a ^4He content of 10^{-6} cc STP/g in the subcontinental mantle should be near $1\text{--}2$ ppb. This concentration is low compared to the uranium content of the MORB source ($5\text{--}6$ ppb) and does not agree with the enriched signature of the subcontinental mantle [45]. Much less uranium is necessary if the subcontinental has less than $1\ \mu\text{cc STP/g } ^4\text{He}$. There is no evidence for helium content in the subcontinental mantle higher than $1\ \mu\text{cc STP/g}$ (e.g. similar to MORB source) that could be compatible with the measured U contents (Fig. 1). In order to illustrate this problem, different closed system evolutions of the R/Ra ratio are reported as a function of ^3He contents and U content (ppb) in Fig. 3. A closure age of 1 Ga was chosen in this calculation [21,46,47]. An initial $^4\text{He}/^3\text{He}$ ratio of $74\,800$ or $R/\text{Ra} = 9.7$ was taken (in this model, the upper mantle has been evolving in closed system since 1 Ga). The Th/U ratio is taken equal to 3. Fig. 3 shows that the closed system model hardly explains the value of the subcontinental mantle, unless it has a very low uranium content (< 2 ppb) or a very high ^3He content (higher than the MORB source concentration).

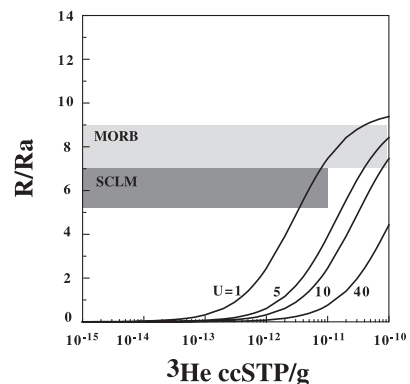


Fig. 3. Evolution in a closed system of the helium isotopic ratio as a function of the ^3He concentration in the SCLM. The age is taken as 1 Ga and four uranium contents (in ppb) were used for the different curves.

3.2. Steady-state model

As an alternative, we have explored another model. We assume that the subcontinental mantle is in steady state for He. The idea is simple and is similar to the model developed for the upper mantle by Porcelli and Wasserburg [48]. The idea that the SCLM is not in closed system was already suggested by Ballentine [49] who suggested that “the subcontinental mantle is not in closed system with respect to gas phases generated by asthenospheric melt emplacement”. Our model assumes that all the ^3He is coming only from the underlying asthenosphere using melts and/or fluids. The hypothesis of infiltration of helium in the SCLM has already been suggested [21,27–30]. In this model, the ^4He derives from two sources. One is the ^4He from the asthenosphere like the ^3He (with a $^4\text{He}/^3\text{He}$ ratio of 90 000; $R/\text{Ra} = 8$, comparable to the ratio of the convective mantle sampled by the MORB). The second origin for ^4He is the in situ decay of U and Th during the residence time. Fig. 4 illustrates the different reservoirs that have been used in this model and their isotopic compositions. Subduction may introduce noble gases into the system. However, the helium concentration in the atmosphere is so small that helium

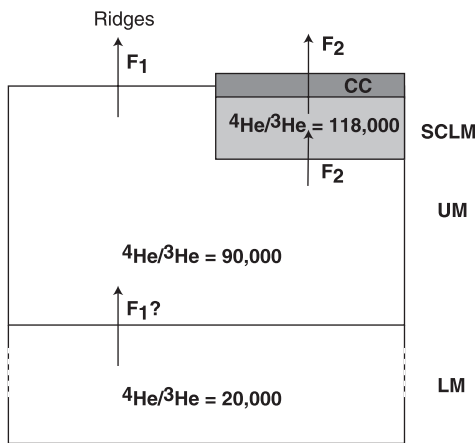


Fig. 4. Cartoon showing the steady-state model for helium in the subcontinental mantle. CC = continental crust; UM = upper mantle; LM = lower mantle. F_1 is the degassing flux at ridges and, if a steady-state model is assumed for the upper mantle, is the flux from the lower mantle. F_2 is the flux from the asthenosphere to the continental lithospheric mantle.

recycling in the mantle is negligible. Knowing the helium content, the uranium content and the $^4\text{He}/^3\text{He}$ ratio in the subcontinental mantle, it is possible to estimate a helium residence time and the ^3He flux. One can write the following equation:

$$P^* + \left(\frac{^4\text{He}}{^3\text{He}}\right)_{\text{MORB}} \cdot F - \left(\frac{^4\text{He}}{^3\text{He}}\right)_{\text{SCM}} \cdot F = 0 \quad (1)$$

where SCM is subcontinental mantle, P^* is the ^4He production (cc STP/yr) and F is the ^3He flux in cc STP/yr (entering and exiting). The residence time R_T of ^3He in the subcontinental mantle is defined as $R_T = [^3\text{He}]/F$ where $[^3\text{He}]$ is the total ^3He (cc STP) in the subcontinental mantle.

We thus have:

$$F = \frac{P^*}{\left(\frac{^4\text{He}}{^3\text{He}}\right)_{\text{SCM}} - \left(\frac{^4\text{He}}{^3\text{He}}\right)_{\text{MORB}}} \quad (2)$$

Here $P^* = 2.8 \times 10^{-14} (4.35 + \text{Th}/\text{U}) \cdot U \cdot M$ in cc STP/yr. U is in ppm and M is the mass of the subcontinental mantle.

3.2.1. Parameters: U and He contents

The $(^4\text{He}/^3\text{He})_{\text{SCM}}$ ratio is taken as 118 000 and $(^4\text{He}/^3\text{He})_{\text{MORB}}$ is taken as 90 000. We use a subcontinental mantle mass M of 7.4×10^{25} g (depth: 150×10^3 m, surface: 149×10^{12} m² and density: 3.3). The Th/U ratio is taken as 3, but the effect of this parameter is small compared to the U content. We have chosen a mean thickness value of 150 km, which reflects the different ages and so thicknesses of the subcontinental lithosphere. In fact, for an old craton, the maximum thickness can be 250 km [17,47]. For the Proterozoic and Phanerozoic subcontinental lithosphere, the thickness is ~ 100 km (typical values for the European lithosphere [39,50]). It is therefore justified to take 150 km.

The very low helium concentration in some continental mantle xenoliths can be interpreted by low helium content in the mantle at a specific area or by helium loss during xenolith transport. The first case is unlikely because the helium isotopic ratios are the same everywhere, indicating

that xenoliths probably suffered helium loss. The helium concentrations in the Massif Central xenoliths (France) are the lowest and helium isotopic ratios are identical to German or Austrian samples. Moreover, the samples from the Massif Central present only rare inclusions as shown by microscopic observations [24]. This may explain the low helium content of these xenoliths. Fluid inclusions contained in xenolith minerals are often decipitated during decompression [41]. Inclusions are opened and may form secondary inclusions. Moreover, the helium contents obtained by crushing are similar to those obtained by melting. The data of Matsumoto and collaborators [27] show several examples of this similarity. For olivine sample 9709, the authors measured by melting a ${}^4\text{He}$ content of 3.9×10^{-8} cc STP/g and by crushing 4.2×10^{-8} cc STP/g. More examples are given for the comparison of helium in olivine performed by crushing and melting [24,25]. This suggests that mantle-derived rare gases are in inclusions. Inclusion losses are common and therefore the helium content of the subcontinental lithospheric mantle cannot be derived from measurements. In the following, different helium contents will be used for the model (from 10^{-7} to 10^{-6} cc STP/g ${}^4\text{He}$). These values are realistic compared to the measurements (Fig. 1).

Due to low concentrations (ppb level), it is difficult to measure uranium in peridotites and to estimate the real content of uranium in the subcontinental lithosphere. The subcontinental mantle is depleted, which should imply a low uranium content compared to the primitive mantle. However, as a result of enrichment of this mantle by fluids and/or melts, the uranium content is certainly higher than in the MORB source, which contains 5–6 ppb [51]. Matsumoto and collaborators [27] have measured the uranium and thorium contents on Australian xenoliths, which are amphibole and apatite metasomatized peridotites. These samples give probably the highest possible U and Th concentrations on mantle xenoliths. Based on their measurements on separate minerals, we can estimate a maximum of ~ 40 ppb U for the bulk peridotite. Eggins and collaborators [45] have measured the U and Th contents in two distinct peridotites from Australia (a fertile lher-

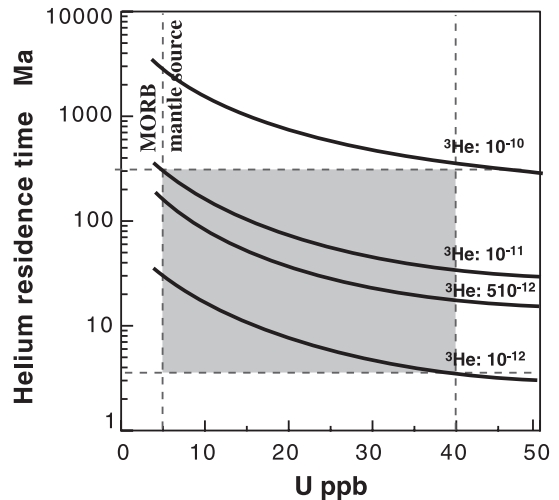


Fig. 5. Representation of the helium residence time in the subcontinental mantle as a function of the U concentration (in ppb) and the concentration of ${}^3\text{He}$ (in cc STP/g).

zolite and a light rare earth element-enriched harzburgite). They measured U contents for the bulk rock of 4 and 10 ppb respectively. The uranium content interval is between 4 and 40 ppb depending on the amount of metasomatism. Moreover, uranium and thorium may be located in grain boundaries that cannot be analyzed by conventional techniques. Nevertheless, in the model, we will use a 4–40 ppb interval with a mean value probably around 10 ppb.

3.2.2. Results and discussion

Helium residence time is reported in Fig. 5 as a function of the concentrations of U and ${}^3\text{He}$ in the subcontinental mantle. For example, with $U = 10$ ppb (and $\text{Th}/U = 3$) and ${}^4\text{He} = 1 \mu\text{cc STP/g}$ (${}^3\text{He} = 10^{-11}$ cc STP/g), the residence time is ~ 200 Myr. With a lower ${}^3\text{He}$ content of about 5×10^{-12} cc STP/g, the helium residence time decreases to ~ 60 Myr. The associated ${}^3\text{He}$ flux is 270 mol/yr when $U = 10$ ppb (or for a smaller lithospheric thickness, for example 100 km, the associated flux is 180 mol/yr when $U = 10$ ppb). For a MORB-like helium content (${}^4\text{He} = 10^{-5}$ cc STP/g), the helium residence time is similar to that proposed for the upper mantle (~ 1 Ga) compatible with a closed system model. The

closed system model can therefore be considered a special case of our model.

Using the surface of the continents, the flux estimated above (270 mol/yr) becomes 3.5 at/s/cm². This value is equivalent to the ³He flux at ridges estimated by Craig and collaborators [52] (4 at/s/cm²). This result implies that the flux is of the same order as the ridge flux. This indicates that the continental volcanism may be as efficient as the ridge volcanism for degassing in this model. Aeschbach-Hertig and collaborators [53,54] have shown that there is an actual important flux of mantle-derived helium in European volcanic provinces. For example, the ³He flux is 500 at/s/cm² in Lac Pavin (France), with a ⁴He/³He ratio of 110 000 ± 170. This degassing rate is local but gives an example of a strong degassing in a continental area during a pause in the volcanism (the Massif Central is sleeping). On a larger scale, the ³He flux is estimated between 0.1 and 0.4 at/s/cm² for the Pannonian basin (Hungary) [49], indicating that during crustal extension massive ³He degassing occurs. However, the observed flux is lower than the predicted flux. This gives limits to the model. We can suggest that the important degassing flux expected from the model can occur during the rifting and opening of an ocean as this melt flux is probably associated with the heat flux, which is at the source of rifting. Thicker lithosphere (<100 km) would lead to smaller ³He flux in the model. This may suggest that the model can be applied only to young lithosphere.

3.3. *Origin of the flux from the asthenosphere?*

Oceanic and continental lithospheres undergo metasomatism. This includes fluid and/or melt percolation in veins or a porous system with a chemical and modal change [55–58]. These CO₂+H₂O fluids are the metasomatic agents [58]. The migration of the noble gases in the shallow mantle is in majority controlled by CO₂ fluids [21,50]. Therefore, this metasomatism can explain the transfer of noble gases from the convective asthenosphere to the lithospheric mantle. Furthermore, Asmerom and collaborators [59] have proposed that melt from the lithospheric mantle derived from a very low melting rate (0.1%) and

may have an upwelling rate of less than 1 cm/yr. Assuming a vertical trajectory, such a melt may arrive at the surface (150 km) in 15 Myr, compatible with the estimated residence times if one considers that the trajectory is not exactly vertical.

4. Conclusions

Olivine phenocrysts and xenoliths in samples from the SCLM show a homogeneous and slightly radiogenic helium isotopic ratio (⁴He/³He = 118 000; *R/Ra* ~ 6.1) compared to the mean MORB value of 90 000 (*R/Ra* = 8). We have explored a model where the SCLM is in steady state for helium. This model assumes there is a flux of ³He from the convective mantle (associated with metasomatism with fluids and/or melts coming from the asthenosphere) and equal to the degassing flux. With this model, one can estimate a helium residence time of ~100 Myr, similar to the time scale of the plate tectonics. Another important conclusion derived from this model is that the ³He flux should be the same under the continents and at mid-ocean ridges (~3.5–4 at/s/cm²). This result seems not realistic considering the observed fluxes on continents. However, this important degassing can occur during rifting and opening of an ocean.

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[AH]

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