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Plume-ridge interaction revisited: Evidence for melt mixing from He, Ne and Ar isotope and abundance systematics

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ABSTRACT

Submarine volcanic glass He, Ne and Ar data from the Pacific-Antarctic-Rise at its intersection with the Foundation Seamount Chain have been obtained to further elucidate one of the main outstanding problems in plume-ridge interaction studies: the physical state in which mantle flow and mixing occur in those settings. Mixing of plume mantle and mid-ocean ridge basalt source material may either be in the form of two solids, of solid-melt or of two melt phases only. The positive correlation of $^4\text{He}/^{40}\text{Ar}^*$ with $^{206}\text{Pb}/^{204}\text{Pb}$ as well as the negative correlation of $^4\text{He}/^{40}\text{Ar}^*$ with $^{40}\text{Ar}^*$ and ^4He concentrations, combined with a lack of correlation between $^4\text{He}/^{40}\text{Ar}^*$ and ridge depth or MgO concentration, indicate that the enriched mantle plume material has been extensively degassed prior to mixing with the mid-ocean ridge source material. Degassing of the plume source and mid-ocean ridge source material prior to mixing is also suggested by the He versus Pb isotope systematics. However, the occurrence of degassing processes requires the existence of melts, implying that mixing between the plume and the mid-ocean ridge material occurs in the physical form of melts.

Keywords: plume-ridge interaction, noble gases, mixing processes, partial melting, degassing

1. INTRODUCTION

Hotspots and mid-ocean ridges (MORs) represent the main features above ascending material of the Earth's mantle where mass and thermal fluxing between the Earth's interior and the surface take place. In areas where hotspots are located close to MORs the mantle plume flow creating the hotspot at the surface interacts with the ridge system as indicated by along-axis bathymetric and gravitational anomalies as well as geochemical and isotopic anomalies of mid-ocean ridge basalts (MORBs) (e.g. Ito et al., 2003). Today 21 of the about 50 active hot spots have been identified to interact with MORs, inducing such anomalies along 15 to 20% of the length of the global MOR system. Plume-ridge interaction processes have been the subject of

geochemical and geophysical studies because they provide information on mantle flow and melting dynamics. An important question in this context is the physical form of mantle flow and mixing, which may either be as two solids, as solid-melt or as two melt phases only (e.g. Ito et al., 2003). Geophysical models of plume-ridge interaction indicate that solid plume material would simply push the depleted mantle material away as it expands along-axis with little or no mixing occurring, resulting in a sudden change of chemical and isotopic composition at the edge of plume influence instead of the observed gradients (Feighner and Richards, 1995; Ito et al., 1997; Braun and Sohn, 2003; Ito et al., 2003). Similar predictions arise from numerical models simulating the entrainment of ambient mantle material into plumes rising vertically through the mantle (Farnetani and Richards, 1995; Farnetani et al., 2002). Richards and Griffith (1989), on the other hand, have shown experimentally that in fast-spreading environments a strong deflection of a plume by horizontal flow of ambient mantle could occur, resulting in significant entrainment and stirring of upper mantle into the plume.

Gradients of Sr, Nd and Pb isotopes as well as gradients of incompatible trace element ratios and concentrations along ridge segments situated close to hotspots such as Iceland, Kerguelen, Easter or Galápagos are interpreted as being due to inflow of deep mantle plume material into adjacent ridge segments and mixing of enriched plume mantle with depleted asthenospheric material (Schilling et al., 1982; Graham et al., 1999; Schilling et al., 1999; Haase, 2002). Plume material flowing along the axis is progressively diluted by the MORB material. However, geochemical modeling of melt formation and mixing based on these trace element and isotope signatures seems to yield less consistent results and mixing of solids, of solids and melts as well as of melts only has been proposed to explain the isotopic and trace element patterns observed in plume-ridge settings (e.g. Poreda et al., 1993; Haase, 2002). An alternative model for plume-ridge interaction processes has been proposed by Phipps-Morgan (1999), Niu et al. (1999) and Niu and Hekinian (2004). They suggested that heterogeneous plume material consisting of enriched veins in a depleted matrix is sucked into the ridge, displacing the MORB source material. In this model, geochemical gradients are not produced by mixing and along-axis dilution of enriched plume melts with melts from the depleted mantle, but by increasing depletion during continuous melting of the enriched veins during along-axis flow.

Noble gases are, owing to their large and indicative isotope variations between the various terrestrial reservoirs, excellent tracers for mantle sources and mixing processes. In addition, caused by the coupling of He, Ne and Ar through distinct production ratios, noble gas relative abundance ratios are a tracer for fractionation processes occurring during magma formation, ascent, eruption and surface residence time. In contrast to the non-volatile trace elements, the behavior of which can be understood from consideration of crystal-melt equilibrium, noble gases can also be partitioned into a vapor phase, and thus they are also a useful tracer for degassing processes. All this makes the noble gases a unique tracer for plume-ridge processes

helping to solve the problem of the physical state in which mantle flow and mixing occurs in plume-ridge settings. Thus, we present new noble gas abundance and isotope data for a suite of volcanic glasses from the Pacific-Antarctic Rise (PAR) between 36 and 40°S (Fig. 1) to further clarify the mixing processes occurring at plume-ridge settings.

2. THE SAMPLE SET

The studied sample set comprises fresh, submarine basaltic glasses from the fast-spreading PAR (~10 cm/yr (Lonsdale, 1994)) between 36 and 40°S, at its intersection with the about 2500 km long Foundation Seamount Chain (FSC) (Fig. 1). Even though it has been shown in a previous publication (Haase et al., 2005) that some of the evolved magmas from this ridge segment have suffered contamination by probably altered oceanic crust, the selected samples are uncontaminated basaltic glasses (Table 1, supplement material), as indicated by their Sr isotopic compositions, which are not displaced to higher values such as those of the contaminated evolved lavas (Haase et al., 2005). Previous studies of the trace element and isotopic composition of samples from this ridge segment have established the existence of systematic variations in e.g. $^{87}\text{Sr}/^{86}\text{Sr}$, $^{206}\text{Pb}/^{204}\text{Pb}$, La/Sm and Ce/Yb ratios (Hekinian et al., 1997; Hekinian et al., 1999; Maia et al., 2000; Haase et al., 2005). The MORBs along this ridge section are variably enriched in incompatible elements and radiogenic isotopes, with lavas with the highest $^{87}\text{Sr}/^{86}\text{Sr}$, $^{206}\text{Pb}/^{204}\text{Pb}$ and La/Sm ratios occurring at the ridge/seamount intersection (Hekinian et al., 1997; Hekinian et al., 1999; Maia et al., 2000; Haase et al., 2005). These variations have been interpreted as evidence of mixing between incompatible element-enriched mantle plume material from the Foundation plume and MORB source material being relatively depleted in those elements (Maia et al., 2000; Haase et al., 2005). The spreading axis along this segment of the PAR is unusually shallow, rising up to 2100 m compared to a typical ridge depth of 3800 m for the Pacific ridge systems (Lonsdale, 1994). The bathymetric data of the PAR study region correspond well with gravity data indicating an increase in crustal thickness in this area (Maia et al., 2001).

3. SAMPLE PREPARATION AND ANALYTICAL PROCEDURES

We have analyzed nine ‘zero age’ glass samples from the PAR (Table 1, supplement material) and an additional about 3.5 Ma old sample (O'Connor et al., 2001) from the Foundation Seamount Chain (FSC) for their He, Ne and Ar isotopic ratios and abundances. The glass samples were crushed, sieved and hand-picked under a microscope. Glasses displaying evidence of alteration were discarded, as well as glass grains being smaller than 0.5 mm. The glass separates were ultrasonically cleaned using acetone before being wrapped in Al foil and placed into the sample carousel of the extraction furnace. To achieve a better separation of magmatic from atmospheric noble gas components, the glasses were degassed in three temperature steps (400°, 800° and 1500°C), using an all-metal, double-walled resistance-heated

furnace with a vacuum-tight Ta crucible and a Nb liner fitted into the Ta crucible without contact at the bottom. Following sample gas extraction, gas purification was achieved according to the procedure described by Niedermann et al. (1997). Analytical blanks were run routinely between sample measurements and depend on the history of the crucible and on extraction temperature. Typical values in 10^{-12} cm³ STP are: ⁴He 1-270, ²⁰Ne 0.5-2.8 and ⁴⁰Ar 480-4200, the maximum values corresponding to the highest extraction step. Noble gases were analyzed in a VG 5400 mass spectrometer having a mass resolution of ≥ 600 . With this resolution isobaric interferences such as ⁴⁰Ar⁺⁺ at $m/e=20$ or ¹²C¹⁶O²⁺⁺ at $m/e=22$ cannot be resolved, and were corrected for by using the method described in Niedermann et al. (1997). For more detailed information on the equipment and analytical procedures used during this study see Niedermann et al. (1997).

4. HE, NE AND AR SYSTEMATICS OF THE SAMPLE SUITE

The investigated PAR glasses have a range of ³He/⁴He ratios (Fig. 2) from values typical for MORB ($8 \pm 1 R_A$, (e.g. Graham, 2002)) to relatively high values of $10.86 \pm 0.11 R_A$. The Ne isotopic ratios lie between the Loihi-Kilauea (Honda et al., 1991) and the MORB trends (Sarda et al., 1988) in a Ne three-isotope diagram (Fig. 3). High ³He/⁴He and primitive Ne isotopic ratios (low ²¹Ne/²²Ne relative to MORB) are usually interpreted to indicate the presence of plume material derived from a relatively undegassed deep mantle source (Graham, 2002). The highest ³He/⁴He and the most primitive Ne isotopic ratios occur, similar to e.g. ²⁰⁶Pb/²⁰⁴Pb or Th/U ratios, at the ridge/seamount chain intersection (Fig. 2), approaching values normal for MORB towards the south. However, in contrast to the about 380 km long geochemical anomaly typical for e.g. ²⁰⁶Pb/²⁰⁴Pb (Fig. 2a) or Th/U ratios (Fig. 2e) the noble gases form, with a length of roughly 150 km, relatively narrow, spike-like anomalies. Similar spike-like anomalies have been observed e.g. along the E-Rift of the Easter Microplate (Poreda et al., 1993) and along the Kolbeinsey Ridge (Schilling et al., 1999). To test whether these narrow noble gas isotopic anomalies are controlled by source composition (a primitive source with high ³He/⁴He but low He concentration mixing with a depleted source with low ³He/⁴He but high He concentration) or magmatic processes, the relative elemental abundances of He and Ar in conjunction with He and Pb isotope systematics are used.

As Ar is variably affected by atmospheric contamination, its measured abundance cannot be used directly. In the following discussion the calculated amounts of radiogenic Ar (Ar*) of each glass will be related to the measured ⁴He abundances, which are assumed to be completely radiogenic. Radiogenic Ar has been calculated using the equation: $^{40}\text{Ar}^* = [(^{40}\text{Ar}/^{36}\text{Ar})_{\text{measured}} - (^{40}\text{Ar}/^{36}\text{Ar})_{\text{atmosphere}}] \times ^{36}\text{Ar}_{\text{measured}}$. The investigated sample suite shows, similar to other MORB or OIB suites (e.g. Graham, 2002), large ranges in ⁴He/⁴⁰Ar*, exceeding the specific production ratios considerably (Fig. 4). In general, ⁴He/⁴⁰Ar* ratios correlate positively with ²⁰⁶Pb/²⁰⁴Pb ratios (Fig. 5) and negatively with ⁴⁰Ar* and ⁴He abundances (Fig. 4a and b). Similar

to the along-axis isotopic gradients (Fig. 2), along-axis gradients in $^4\text{He}/^{40}\text{Ar}^*$ can be observed, with the highest $^4\text{He}/^{40}\text{Ar}^*$ occurring again at the ridge/seamount chain intersection (Fig. 2d). The $^4\text{He}/^{40}\text{Ar}^*$ ratios show a similar range over a much smaller depth variation as MORBs from other plume-ridge settings (Fig. 6a), for example, from the South East Indian Ridge (SEIR) (Burnard et al., 2002; Burnard et al., 2004). The investigated samples show no trend of $^4\text{He}/^{40}\text{Ar}^*$ with MgO, a tracer for the extent of fractional crystallization (Fig. 6b). In contrast to the distinct along-axis geochemical gradients observed (Fig. 2) and the good correlations between $^4\text{He}/^{40}\text{Ar}^*$ and $^{206}\text{Pb}/^{204}\text{Pb}$, the He isotope ratio shows no simple correlation with e.g. $^{206}\text{Pb}/^{204}\text{Pb}$ (Fig. 7).

5. DISCUSSION

5.1. *Effects of shallow level magmatic processes on noble gases*

The behavior of noble gases during melt formation and evolution depends on their solubility in mantle minerals and their solubility in melts, which controls the noble gas content of the vapor phases produced during melt degassing. The solubility of noble gases in melts is a function of the atomic radius of the specific gas, depending mainly on pressure and melt composition and decreasing from He to Xe by a factor of ~ 50 (Jambon et al., 1986; Broadhurst et al., 1992; Carroll and Draper, 1994; Shibata et al., 1998). A consequence of this behavior is a relative enrichment of He compared to the heavier noble gases (Ar, Kr, Xe) in the melt in contrast to a relative depletion in the vapor phase. Thus, the concentration of noble gases and indirectly, their isotopic composition in melts is not only controlled by source composition and degree of melting but also by degassing during eruption and fractional crystallization processes (e.g. Marty and Zimmerman, 1999; Burnard et al., 2002; Burnard et al., 2004). When considering degassing processes it must be distinguished between those processes occurring during melt movement and stagnation in the shallow mantle or crust and those occurring during eruption. In this context, the amount of volatiles trapped in submarine basaltic glasses is known, independent of the degassing mechanism (e.g. equilibrium or disequilibrium degassing), to be partly controlled by the hydrostatic pressure during eruption exerted by the overlying seawater column – i.e. the eruption depth (Moore, 1979; Burnard et al., 2002; Burnard et al., 2004; Gonnerman and Manga, 2005; Gonnermann and Mukhopadhyay, 2007). Due to the fractionation of He from Ne or Ar during degassing (e.g. Marty and Zimmerman, 1999; Burnard et al., 2004), $^4\text{He}/^{40}\text{Ar}^*$ can be used as a tracer for degassing processes occurring during magmatic evolution. If the gas content of a sample suite is controlled by syneruptive degassing, $^4\text{He}/^{40}\text{Ar}^*$ should correlate negatively with water depth (Fig. 6a) in agreement with observations made along different spreading axes (e.g. Marty and Zimmerman, 1999; Burnard et al., 2004). However, in the PAR samples a large range in $^4\text{He}/^{40}\text{Ar}^*$ ($^4\text{He}/^{40}\text{Ar}^*$ ratios from 99 to 2960) is related to a depth difference of only ~ 50 m (2226 m - 2273 m), translating to a pressure difference of about 5 bar. This pressure difference is much too small to account for the observed ~ 1000 times enrichment of He relative to Ar

(initial $^4\text{He}/^{40}\text{Ar}^* \sim 3$), and it can be concluded that pressure variation due to variable water depth is not causing the observed gradient of $^4\text{He}/^{40}\text{Ar}^*$. This is in accordance with studies of Burnard et al. (2002, 2004), who have shown for a sample suite from the SEIR, comparable in ridge depth distribution to our samples (Fig. 6a), that large variations in the $^4\text{He}/^{40}\text{Ar}^*$ ratio over small depth differences are not due to degassing processes controlled by the confining pressure during eruption. Similarly, the observed variations in $^4\text{He}/^{40}\text{Ar}^*$ cannot be explained by pressure differences induced by the spatial position and depth distribution of melt lenses in the crust along the ridge. It has been shown that long lived, stable magma chamber systems with a limited depth range are developed along ridges with high melt supply (fast spreading ridges and plume influenced ridges) (e.g. Purdy et al., 1992; Rubin and Sinton, 2007). Thus a large variation in magma chamber depth is not expected for the studied PAR segment. In addition, fractional crystallization models indicate similar magma stagnation levels for the entire PAR segment presented in this study (Stroncik et al., 2002; Haase et al., 2005). In this context it should be noted that all studied samples are derived from the neo-volcanic zone and consequently no large age differences between those samples can be expected. Thus, even though later degassing during residence of melts in a magma chamber and syneruptive degassing cannot be ruled out, the observed $^4\text{He}/^{40}\text{Ar}^*$ trends are not controlled by pressure variations in the shallow crust or during eruption.

As noble gas solubility is also depending on melt composition, especially on the melt's CO_2 concentration, the degree of degassing can also be related to the extent of fractional crystallization (Marty and Zimmerman, 1999; Burnard et al., 2002). CO_2 is the first volatile phase to separate from the magma (e.g. Dixon et al., 1995; Dixon, 1997). E.g., it has been shown by Gerlach (1985) and Stolper and Holloway (1988) that a depth of about 40 km is necessary to dissolve the parental CO_2 content present in the tholeiitic magmas from Kilauea. The exsolution of CO_2 from the magma induces the degassing of the noble gases according to their melt solubilities (e.g. Marty and Zimmerman, 1999; Burnard et al., 2002). As CO_2 behaves like an incompatible element it gets successively enriched in the melt during the processes of fractional crystallisation. In response to this enrichment more and more CO_2 and, along with the CO_2 , more and more noble gases, are lost from the melt to a gas phase during the evolution of a magma (e.g. Marty and Zimmerman, 1999; Burnard et al., 2002). In such a case the $^4\text{He}/^{40}\text{Ar}^*$ ratio should correlate positively with the MgO concentration as a tracer for the extent of crystallization. However, no such correlation is observed for the investigated sample suite. Instead it is characterized by a large variation of $^4\text{He}/^{40}\text{Ar}^*$ at nearly constant MgO concentrations (7 to 8 wt.%) (Fig. 6b). In agreement with these similar MgO concentrations of the investigated lavas, variable residence times in crustal magma reservoirs can also be ruled out as possible reasons for the $^4\text{He}/^{40}\text{Ar}^*$ variations. Thus it can be concluded that none of the processes affecting the noble gases in the shallow crust can explain the observed along-ridge noble gas isotopic and abundance patterns.

5.2. *Plume-ridge interaction: melt mixing or progressive depletion?*

Because shallow level degassing processes cannot cause the observed trends in the PAR MORBs, especially the positive covariation observed between $^4\text{He}/^{40}\text{Ar}^*$ and $^{206}\text{Pb}/^{204}\text{Pb}$ (Fig. 5), we infer that these trends are most likely produced deeper in the mantle by interactions of melts from enriched plume material with melts from depleted upper mantle material. Incompatible element and Sr, Nd and Pb isotope data suggest that MORBs emplaced along plume-influenced ridges are generated either by variable degrees of melting and binary mixing of two separate sources (e.g. Douglass and Schilling, 2000; Haase, 2002; Murton et al., 2002; Niu and Hekinian, 2004; Kokfelt et al., 2005) or by progressive depletion or a progressive decrease of the enriched parts of a heterogeneous plume (“melting-induced mixing”) (Morgan, 1999; Niu et al., 1999; Niu and Hekinian, 2004). Progressive depletion of heterogeneous plume material as it flows along the PAR axis appears unlikely for the following reasons. (1) In such a scenario the extent of a geochemical anomaly is mainly controlled by melting of the enriched components and thus depends primarily, for a given source composition, on partition coefficients. Therefore the He and Ne isotopic anomaly should be similar to the extent of the anomalies displayed by e.g. Th/U (Fig. 2e) as they show relatively similar degrees of incompatibility taking published partition coefficients into account (e.g. Beattie, 1993; Heber et al., 2007). (2) The degree of degassing along the ridge should increase towards the periphery of plume influence and the most degassed samples should not be observed at the point of maximum plume influx. (3) The enriched portions of a heterogeneous plume are generally believed to consist of recycled material (e.g. Hofmann, 1997). In case of the FSC-PAR system the high $^{206}\text{Pb}/^{204}\text{Pb}$ ratios observed along the FSC (Maia et al., 2000) would require a recycled component with high time integrated $[\text{U}+\text{Th}]/\text{Pb}$ ratios resulting in low $^3\text{He}/^4\text{He}$ ratios, which is in contrast to the high $^3\text{He}/^4\text{He}$ ratios observed at the point of maximum plume influx. (4) In the heterogeneous plume model the depleted component is part of the plume, however the depleted endmember observed along the PAR resembles typical average East Pacific Rise MORB as postulated by White et al. (1987), making it, in our opinion, unlikely that the depleted endmember is part of the plume. Thus, we suggest that the $^4\text{He}/^{40}\text{Ar}^*$ systematics observed along the PAR are controlled by binary mixing processes between more degassed, enriched mantle material and less degassed, depleted mantle material.

Modeling of these mixing processes indicates that the $^{40}\text{Ar}^*/\text{Pb}$ ratio of the depleted endmember must be about 30 times higher than that of the enriched endmember. This is also supported by the negative correlations of $^{40}\text{Ar}^*$ and ^4He with $^4\text{He}/^{40}\text{Ar}^*$ (Fig. 4). The Pb isotopic systematics require relatively high $^4\text{He}/^{40}\text{Ar}^*$ for both endmembers in this mixing scenario (Fig. 5). These relatively high $^4\text{He}/^{40}\text{Ar}^*$ ratios cannot be source-intrinsic, as rocks being in secular equilibrium should have $^4\text{He}/^{40}\text{Ar}^*$ ratios within the theoretical production ratio of both elements (Fig. 4) (e.g. Graham, 2002). This production ratio has been estimated according to the method described by Graham (2002), using $[\text{U}+\text{Th}]/\text{K}$ typical for the mantle. Even if we assume that a source ratio must not necessarily reflect the production ratio, there is no

likely process to produce a mantle source with high initial $^4\text{He}/^{40}\text{Ar}^*$, because this would either require that He is more compatible than Ar or that [U+Th] is more compatible than K during melting. Thinking in terms of crustal recycling during subduction, higher proportions of U and Th would have to be recycled compared to K to generate a source with high $^4\text{He}/^{40}\text{Ar}^*$. However, studies of recycling processes during subduction indicate that the proportions of [U+Th] recycled are not larger than the proportions of K (Scambelluri et al., 1997; Philippot et al., 1998; Scambelluri et al., 2001). In addition such a source would result, in contrast to what is observed at the FSC, in radiogenic $^3\text{He}/^4\text{He}$. Generally, higher $^4\text{He}/^{40}\text{Ar}^*$ ratios are mainly produced by magmatic degassing during melt formation and evolution (e.g. Burnard, 1999; Moreira et al., 2001; Burnard et al., 2004). This indicates that the mixing between the plume endmember and the MORB source endmember must take place in the molten state, because only melts can have experienced degassing processes prior to mixing. Similar findings have been reported by Burnard et al. (2002), who have shown that the $^3\text{He}/^4\text{He}$ vs. K/Ti systematics observed along the SEIR at the Amsterdam-St. Paul Plateau must have been produced by degassing at mantle depth and subsequent mixing.

The melt mixing model is also supported by the He/Pb isotope systematics of the PAR magmas (Fig. 7). The linear positive covariations of $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ (Haase et al., 2005) indicate binary mixing between an enriched source and a depleted source which have similar Sr/Pb ratios. In contrast, the mixing lines for $^3\text{He}/^4\text{He}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ (Fig. 7) suggest mixing of two materials both with variable He/Pb ratios. The required variations can be explained by geochemical heterogeneities being either intrinsic to the source or inherited during melt formation and evolution. As heterogeneous mixing relations are only observed in He/Pb isotope space (Haase et al., 2005), geochemical heterogeneities intrinsic to the sources can be ruled out, because this would require that the source of the Foundation plume shows only a heterogeneous He composition. This is, in our opinion, unlikely as such heterogeneities can only be produced by processes having an impact on the entire composition of a source, such as e.g. melting. Thus the varying He/Pb ratios required to explain the He/Pb isotope systematics must have formed during melt formation and/or evolution. These processes can be modeled to constrain the required He/Pb ratios suggested by He/Pb systematics. The following endmembers have been used in these model calculations: (1) the depleted component is represented by the sample with the lowest $^{206}\text{Pb}/^{204}\text{Pb}$ and a MORB-like He isotopic ratio of $7.42 \pm 0.09 R_A$, (2) the enriched component is represented by sample 70DS2 from the FSC with a $^{206}\text{Pb}/^{204}\text{Pb}$ ratio of 20.2. As this glass sample is about 3.7 Ma old (O'Connor et al., 2001) and has Th and U concentrations of 1.73 and 0.55 ppm, respectively (Devey et al., 1997), age-corrected He isotope data of 22 ± 2 have to be used in the mixing model. Modeling of the mixing processes shows that $(\text{He}/\text{Pb})_{\text{MORB}}/(\text{He}/\text{Pb})_{\text{Plume}}$ source ratios varying between 0.6 and 10 are necessary to explain the observed He/Pb isotopic patterns. This requires that the He/Pb ratio of the plume-derived material must vary from values lower to values higher than that in the MORB material. A

lowering of the He/Pb ratio can be the result of magmatic degassing, whereas an increase in the He/Pb ratio can be achieved during low degrees of partial melting because the mineral/melt distribution coefficients of He (e.g. Heber et al., 2007) are lower than those of Pb (Beattie, 1993; Hauri et al., 1994) (Fig. 8). Thus He/Pb fractionation at the onset of melting in the deepest part of the melting column may produce the higher He/Pb ratios of melts derived from the enriched source. Taking all the different isotope and element systematics of the magmatic rocks from the studied PAR segment into account, the observed along-axis geochemical gradients can best be explained by mixing of variably degassed melts from the MORB and the Foundation plume source.

Even though the exact degrees of melting, degassing and mixing cannot be modeled because of the large uncertainties in the mineral/melt distribution coefficients of He, the combined noble gas isotope and element data suggest that mixing in the FSC-PAR system (and probably other plume-ridge settings) occurs between melts from the two sources. Thus our data support experimental and numerical geophysical models indicating that little or no solid state mixing occurs in the asthenosphere (see review by Ito et al., 2003).

6. CONCLUSIONS

Our noble gas study (He, Ne and Ar) of MORBs from the FSC-PAR plume-ridge system established indirectly that the Foundation plume is derived from a relatively primitive mantle source which resembles those of the Galápagos (Kurz and Geist, 1999) or Easter mantle plumes (Poreda et al., 1993). We have also shown that the along-axis geochemical anomaly defined by trace elements and e.g. $^{206}\text{Pb}/^{204}\text{Pb}$ also occurs in the noble gas isotopes, but that the noble gas isotope gradient is narrower than that of, for example, the Pb isotopes. The $^4\text{He}/^{40}\text{Ar}^*$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ and the $^3\text{He}/^4\text{He}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ variations imply that the geochemical gradients observed along the PAR can only be explained by mixing of variably degassed melts formed by variable degrees of melting of two sources. We suggest that the Foundation plume rises beneath the western flank of the melting region of the PAR, supplying melts into the PAR where they mix with melts produced during shallower level melting of depleted mantle. The relatively high CO_2 concentrations and deep onset of melting of the enriched plume material lead to increased degassing of the rising plume magmas, thereby creating lower He/Pb and higher $^4\text{He}/^{40}\text{Ar}^*$ compared to the MORB melts. The data presented here in conjunction with similar $^4\text{He}/^{40}\text{Ar}^*$ and incompatible element systematics from other plume-influenced ridge segments (Burnard et al., 2002) indicate that mixing processes at plume-ridge settings generally occur in the form of melts.

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FIGURE CAPTIONS

- Fig. 1: Bathymetric map of the study area showing the position of the samples discussed in this contribution. Arrows indicate samples from which noble gas data have been obtained.
- Fig. 2: Along axis geochemical gradients of $^{206}\text{Pb}/^{204}\text{Pb}$ (a), $^3\text{He}/^4\text{He}$ (b), $(^{21}\text{Ne}/^{22}\text{Ne})_{\text{ext}}$ (c) and $^4\text{He}/^{40}\text{Ar}^*$ (d). $(^{21}\text{Ne}/^{22}\text{Ne})_{\text{ext}}$ is the $^{20}\text{Ne}/^{22}\text{Ne}$ ratio corrected for atmospheric contamination by extrapolation to the solar $^{20}\text{Ne}/^{22}\text{Ne}$ ratio of 13.8 (Graham, 2002). Note that only Ne data which are different from the atmospheric composition at the 2σ uncertainty level have been considered in this study. In addition, the sample with the high analytical uncertainties in $^3\text{He}/^4\text{He}$ will not be used in the discussion. This sample, being the most evolved sample analyzed (3.5 wt.% MgO, Fig. 6b), has suffered a high gas loss during melt crystallization. The Pb isotope data are taken from Haase et al. (2005). 2σ uncertainties are shown where larger than the symbol.
- Fig. 3: Neon three-isotope diagram. Ne data of all temperature steps being isotopically distinguishable from air at the 2σ confidence level are shown. Data lying between the Loihi-Kilauea line (Honda et al., 1991) and the MORB line (Sarda et al., 1988) indicate mixing between a more primitive and a less primitive, more radiogenic mantle source. 2σ uncertainties are shown.
- Fig. 4: $^4\text{He}/^{40}\text{Ar}^*$ vs. $^{40}\text{Ar}^*$ (a) and vs. ^4He (b). The negative trends observed between $^4\text{He}/^{40}\text{Ar}^*$ vs. $^{40}\text{Ar}^*$ and ^4He in this case not only indicate solubility-controlled degassing but also contain a mixing signal between degassed and less degassed melts. The $^4\text{He}/^{40}\text{Ar}^*$ production range is taken from Graham (2002). 2σ uncertainties are shown where larger than the symbol.
- Fig. 5: $^4\text{He}/^{40}\text{Ar}^*$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$. The observed positive covariation indicates mixing of a more degassed, enriched endmember with a less degassed, depleted one. As the $^4\text{He}/^{40}\text{Ar}^*$ ratios of both endmembers are higher than the production range (see Fig. 4) the differences in the $^4\text{He}/^{40}\text{Ar}^*$ ratios are not source-intrinsic but must have been obtained during melt formation and evolution. The mixing lines have been calculated using the samples with the highest and the lowest $^{206}\text{Pb}/^{204}\text{Pb}$ ratios found along-ridge and their associated $^4\text{He}/^{40}\text{Ar}^*$

ratios as endmembers for the enriched and the depleted source, respectively. The r_{mix} posted at the mixing lines stand for the $(^{40}\text{Ar}^*/\text{Pb})_{\text{MORB}}/(^{40}\text{Ar}^*/\text{Pb})_{\text{plume}}$ source ratios. R^2 of $r_{\text{mix}} = 30$ is 0.93.

Fig. 6: $^4\text{He}/^{40}\text{Ar}^*$ vs. depth (a) and vs. MgO concentration (b). No obvious correlation can be observed between $^4\text{He}/^{40}\text{Ar}^*$ vs. depth or MgO concentration, indicating that the variations observed in the $^4\text{He}/^{40}\text{Ar}^*$ ratio are neither controlled by confining pressure nor by the degree of fractional crystallization alone. For comparison samples from the SEIR are shown (circles and squares) (Burnard et al., 2004). The circles denote the sample suite being controlled by degassing mixing processes, whereas the squares show samples controlled by confining pressure during eruption (Burnard et al., 2004). 2σ uncertainties are given where larger than the symbol.

Fig. 7 $^4\text{He}/^3\text{He}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$. The geochemical heterogeneity of the different mixing endmembers required to reproduce the isotopic signatures is shown. The mixing lines have been calculated using the sample with the lowest $^{206}\text{Pb}/^{204}\text{Pb}$ and MORB-like He isotopic ratio as endmember for the depleted component and the sample with the most radiogenic $^{206}\text{Pb}/^{204}\text{Pb}$ and relatively primitive, age corrected He isotopic ratio (see text for further explanation) as endmember for the enriched one. The r_{mix} posted at the mixing lines stand for the $(\text{He}/\text{Pb})_{\text{MORB}}/(\text{He}/\text{Pb})_{\text{plume}}$ source ratios. 2σ uncertainties are shown where larger than the symbol.

Fig. 8: Modeled evolution of He (black curve) and Pb (gray curve) melt concentrations with the degree of melting (F). The melt composition has been calculated using the non-modal batch melting equation of Shaw (1970) and the mantle and melt modes of van Westrenen et al. (2001) for garnet peridotite. To exclusively demonstrate the influence of the partition coefficient on the evolution of He and Pb in the melt, the initial He and Pb source concentrations have been arbitrarily set to 0.15 ppm each. The partition coefficients have been taken from Heber et al. (2007) and Beattie (1993), the uncertainties of those partition coefficients have not been included in the model calculations.

Fig. 1

Fig. 2

Fig. 3

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Fig. 7

Fig. 8

Table 1: Concentration and isotope data of He, Ne and Ar as well as MgO concentrations of the investigated sample suite

Sample	T [°C]	⁴ He [10 ⁻⁸ cm ³ /g]	²⁰ Ne [10 ⁻¹² cm ³ /g]	⁴⁰ Ar [10 ⁻⁸ cm ³ /g]	³ He/ ⁴ He 10 ⁻⁶	²⁰ Ne/ ²² Ne	²¹ Ne/ ²² Ne	⁴⁰ Ar/ ³⁶ Ar	³⁸ Ar/ ³⁶ Ar	MgO [wt.%]
60DS1SO157 40.13°S 111.46°W	400	15.64 ± 0.78	585 ± 30	15.21 ± 0.76	11.00 ± 0.60	9.921 ± 0.079	0.02928 ± 0.00049	292.5 ± 1.1	0.18636 ± 0.00065	
	800	449 ± 16	1856 ± 94	5.28 ± 0.28	10.38 ± 0.16	9.965 ± 0.053	0.02930 ± 0.00035	616.5 ± 5.8	0.1870 ± 0.0018	
	1500	44.8 ± 1.6	48.7 ± 2.7	18.31 ± 0.96	9.72 ± 0.30	9.74 ± 0.12	0.0312 ± 0.0015	1231 ± 30	0.1884 ± 0.0016	
	Total	510 ± 16	2489 ± 98	38.8 ± 1.3	10.34 ± 0.15	9.950 ± 0.044	0.02933 ± 0.00029	514 ± 14	0.18683 ± 0.00059	
61DS1SO157 40.13°S 111.42°W	400	52.6 ± 1.9	816 ± 41	1.70 ± 0.11	10.72 ± 0.38	9.939 ± 0.055	0.02916 ± 0.00034	331.2 ± 3.8	0.1893 ± 0.0021	
	800	441 ± 16	2850 ± 150	5.38 ± 0.30	10.34 ± 0.20	9.947 ± 0.068	0.02923 ± 0.00049	1337 ± 41	0.1870 ± 0.0039	
	1500	29.7 ± 1.1	44.9 ± 2.5	21.3 ± 1.2	9.13 ± 0.49	9.56 ± 0.14	0.0312 ± 0.0014	4900 ± 1100	0.1873 ± 0.0049	
	Total	523 ± 16	3710 ± 150	28.4 ± 1.2	10.31 ± 0.18	9.940 ± 0.054	0.02924 ± 0.00038	2100 ± 410	0.1880 ± 0.0021	
02DS1SO157 37.66°S 110.87°W	400	23.91 ± 0.85	252 ± 13	11.81 ± 0.59	15.42 ± 0.54	9.697 ± 0.071	0.02898 ± 0.00076	295.0 ± 1.3	0.1877 ± 0.0015	
	800	206.8 ± 7.3	105.2 ± 5.5	5.45 ± 0.28	15.13 ± 0.30	9.81 ± 0.11	0.0296 ± 0.0012	298.2 ± 1.2	0.18754 ± 0.00090	
	1500	17.73 ± 0.89	11.8 ± 1.2	0.88 ± 0.25	14.06 ± 0.81	8.80 ± 0.67	0.0287 ± 0.0036	343 ± 16	0.1834 ± 0.0057	
	Total	248.4 ± 7.4	369 ± 14	18.14 ± 0.70	15.08 ± 0.26	9.698 ± 0.063	0.02916 ± 0.00064	298.0 ± 1.3	0.1875 ± 0.0010	
29DS1SO157 37.97°S 110.97°W	400	15.28 ± 0.54	68.1 ± 3.7	2.63 ± 0.14	12.36 ± 0.49	9.82 ± 0.13	0.0290 ± 0.0020	295.5 ± 1.9	0.1861 ± 0.0022	
	800	348 ± 12	107.9 ± 6.1	1.07 ± 0.13	12.13 ± 0.23	10.01 ± 0.19	0.0307 ± 0.0017	1480 ± 200	0.1865 ± 0.0052	
	1500	137.2 ± 4.9	16.9 ± 1.8	4.00 ± 0.87	11.31 ± 0.33	10.80 ± 0.32	0.0415 ± 0.0055	5900 ± +∞ / -4500	0.166 ± 0.094	
	Total	500 ± 13	192.9 ± 7.4	7.70 ± 0.89	11.92 ± 0.18	10.01 ± 0.12	0.0309 ± 0.0013	778 ± 31	0.1849 ± 0.0080	
106DS2SO157 36.90°S 110.63°W	400	0.089 ± 0.011	853 ± 43	0.951 ± 0.064	12 ± 10	10.011 ± 0.037	0.02926 ± 0.00047	305.1 ± 5.1	0.1878 ± 0.0035	
	800	0.210 ± 0.015	2350 ± 120	1.62 ± 0.16	12 ± 14	9.957 ± 0.030	0.02908 ± 0.00040	312.7 ± 4.2	0.1881 ± 0.0037	
	1500	0.0504 ± 0.0087	15.8 ± 1.0	3.50 ± 0.28	18 ± 16	9.60 ± 0.22	0.0291 ± 0.0031	305.2 ± 3.8	0.1879 ± 0.0018	
	Total	0.350 ± 0.020	3220 ± 130	6.06 ± 0.33	12.8 ± 8.9	9.970 ± 0.024	0.02913 ± 0.00032	307.2 ± 2.6	0.1879 ± 0.0015	
45DS1SO157 38.58°S 111.13°W	400	33.4 ± 1.2	14.9 ± 1.2	1.124 ± 0.073	11.18 ± 0.51	9.55 ± 0.50	0.0291 ± 0.0030	304.3 ± 7.6	0.1862 ± 0.0048	
	800	617 ± 22	34.7 ± 2.1	3.27 ± 0.19	10.80 ± 0.14	10.70 ± 0.10	0.0386 ± 0.0037	716 ± 25	0.1881 ± 0.0028	
	1500	98.8 ± 4.9	22.7 ± 1.7	5.38 ± 0.40	10.07 ± 0.22	10.13 ± 0.39	0.0354 ± 0.0041	1250 ± 170	0.1889 ± 0.0033	
	Total	749 ± 22	72.3 ± 2.9	9.77 ± 0.45	10.72 ± 0.12	10.26 ± 0.18	0.0354 ± 0.0023	776 ± 69	0.1878 ± 0.0021	
DR09/02AL 38.70°S 111.15°W	400	236.0 ± 8.3	10.24 ± 0.85	0.647 ± 0.054	11.00 ± 0.24	9.64 ± 0.34	0.0306 ± 0.0035	308.6 ± 4.4	0.1886 ± 0.0020	
	800	566 ± 20	47.0 ± 2.6	8.35 ± 0.54	10.52 ± 0.10	10.96 ± 0.23	0.0417 ± 0.0019	1048 ± 29	0.1879 ± 0.0027	
	1500	0.135 ± 0.012	10.1 ± 1.0	6.86 ± 0.36	12 ± 16	9.47 ± 0.43	0.0299 ± 0.0040	424.1 ± 5.6	0.1880 ± 0.0019	
	Total	803 ± 22	67.4 ± 2.9	15.85 ± 0.66	10.66 ± 0.10	10.49 ± 0.18	0.0379 ± 0.0015	605 ± 16	0.1880 ± 0.0014	
17DS1SO157 37.67°S 110.88°W	≤400	0.0485 ± 0.0049	<1.2	<0.067	15.5 ± 6.1	> 9.4	>0.021	>290	<0.20	
	800	192 ± 12	140.1 ± 7.5	5.41 ± 0.30	15.18 ± 0.24	10.02 ± 0.10	0.0300 ± 0.0010	299.4 ± 2.1	0.1878 ± 0.0016	
	1500	17.0 ± 1.0	6.1 ± 1.8	0.28 ± 0.66	14.16 ± 0.32	11.57 ± 0.91	0.0359 ± 0.0076	610.0 ± + ∞ / -270	0.186 ± 0.032	
	Total	209 ± 12	146.4 ± 7.8	5.69 ± 0.72	15.10 ± 0.22	10.08 ± 0.10	0.0302 ± 0.0010	307 ± 52	0.1878 ± 0.0018	
24DS1SO157 37.90°S 110.95°W	400	0.1083 ± 0.0056	0.1 ± 1.2	0.004 ± 0.030	11.8 ± 2.7	9 ± 15	0.04 ± 0.19	1460.9 ± + ∞ / -1200	0.3666 ± 6.3118	
	800	337 ± 24	71.7 ± 4.1	0.82 ± 0.10	11.06 ± 0.23	10.035 ± 0.068	0.0308 ± 0.0016	307.0 ± 4.1	0.1883 ± 0.0029	
	1500	75.3 ± 3.8	14.6 ± 1.5	1.47 ± 0.19	10.29 ± 0.18	11.42 ± 0.24	0.0412 ± 0.0064	888 ± 181	0.1885 ± 0.0048	
	Total	413 ± 24	86.4 ± 4.6	2.30 ± 0.22	10.92 ± 0.19	10.244 ± 0.076	0.0324 ± 0.0017	530 ± 91	0.1885 ± 0.0064	
Atmospheric values:					1.39	9.8	0.0290	295.5	0.1880	