

Nucleogenic neon in high $^3\text{He}/^4\text{He}$ lavas from the Manus back-arc basin: a new perspective on He–Ne decoupling

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Abstract

We report new neon isotope data obtained for well-characterised basaltic glasses from the Manus back-arc basin where helium studies have identified a mantle plume component (mean $^3\text{He}/^4\text{He} \sim 12 R_A$). In three-isotope neon space, seven of the Manus samples lie along a trajectory between air and an endmember more nucleogenic than MORB i.e., compared to typical MORB, samples have a higher $^{21}\text{Ne}/^{22}\text{Ne}$ ratio for a given $^{20}\text{Ne}/^{22}\text{Ne}$ ratio. Thus the slope of the Manus Basin line is less than that of the MORB line [Sarda et al., *Earth Planet. Sci. Lett.* 91 (1988) 73–88]. This is the first observation of lavas with high $^3\text{He}/^4\text{He}$ ratios having nucleogenic neon isotope systematics, indicating a unique decoupling of He from Ne. We evaluate five possible explanations for the observed trend. We discount: (1) crustal contamination, (2) devolatilisation of the subducting Solomon Sea plate and (3) addition of neon from an ancient recycled slab component – based upon mass balance considerations of the availability of nucleogenic Ne. Two possibilities remain – both of which must produce an elevated He/Ne ratio in the Manus Basin source region to account for the nucleogenic neon: (4) a previous degassing event which would leave a Ne-depleted residual reservoir, or (5) a deep mantle source heterogeneity preserving a unique signature inherited from Earth's accretion. We find that isolation times as short as 10 Ma for a previously degassed source are sufficient to grow in the nucleogenic Ne without significantly altering the plume-like $^3\text{He}/^4\text{He}$ ratios. Alternatively, solubility-controlled outgassing/ingassing of a magma ocean in contact with a proto-atmosphere may have produced the requisite high He/Ne ratio, although an open-system style of equilibration is necessary. At present, insufficient evidence is available to discriminate between these alternatives. © 2001 Elsevier Science B.V. All rights reserved.

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1. Introduction

Neon and helium isotopes are highly sensitive indicators of mantle plume involvement in petrogenesis. The current paradigm is that plumes tap the relatively undegassed lower mantle, which is thought to have preserved a high proportion of primitive noble gases [4,5]. In the case of helium

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and neon, these primitive gases may have had solar-like isotopic compositions [5]: however, radiogenic ^4He and nucleogenic ^{21}Ne in-growth over time will have altered the initial ratios. Potentially complicating factors, such as magma mixing, crustal assimilation, and air contamination, can further modify isotopic ratios of a magma body on its path towards the Earth's surface. In most cases, these modifications can be identified since the different sources (e.g., upper mantle, lower mantle, crust and air) have diagnostic helium and neon isotopic ratios.

In the case of helium isotope systematics, ^3He is overwhelmingly primordial in origin, whereas ^4He is the product of U and Th decay. Ocean island basalts (OIB) erupting in plume-related localities such as Hawaii (e.g. [6]) and Iceland [7] have high $^3\text{He}/^4\text{He}$ ratios (up to $37 R_A$, where R_A is the ratio in air $= 1.4 \times 10^{-6}$). In contrast, most mid-ocean ridge basalts have $^3\text{He}/^4\text{He}$ ratios equal to $8 \pm 1 R_A$ [8,9]. This constancy suggests that the upper mantle source for MORB is homogeneous in composition with respect to helium. Crustal $^3\text{He}/^4\text{He}$ ratios are significantly lower than the air value due to the dominance of radiogenic helium.

The three isotopes of neon (^{20}Ne , ^{21}Ne and ^{22}Ne) can be used to resolve atmospheric, primordial and nucleogenic contributions to mantle-derived magmas. ^{20}Ne and ^{22}Ne are predominantly primordial – trapped during accretion of Earth, whereas ^{21}Ne is produced by (α, n) reactions on ^{18}O [10]. The consequence of producing nucleogenic ^{21}Ne is to increase the $^{21}\text{Ne}/^{22}\text{Ne}$ ratio over time. In plots of $^{20}\text{Ne}/^{22}\text{Ne}$ versus $^{21}\text{Ne}/^{22}\text{Ne}$ (three-isotope neon space), both MORB and OIB follow linear arrays from the atmospheric value ($^{20}\text{Ne}/^{22}\text{Ne} = 9.80$ and $^{21}\text{Ne}/^{22}\text{Ne} = 0.029$ – since a certain degree of atmospheric contamination is inevitable in terrestrial samples) to more nucleogenic $^{21}\text{Ne}/^{22}\text{Ne}$ ratios. On such plots, OIB magmas tend to follow a trajectory with a slope steeper than MORB samples (i.e. they show lower $^{21}\text{Ne}/^{22}\text{Ne}$ values for a given $^{20}\text{Ne}/^{22}\text{Ne}$ ratio) – this observation has been explained by assuming that the lowermost mantle is less degassed than the source of MORB (i.e., lower mantle contains more primordial ^{22}Ne) so that the isotopic

shift induced by addition of nucleogenic ^{21}Ne is less pronounced. Significantly, because production of ^{21}Ne is coupled to that of ^4He ($^{21}\text{Ne}/^4\text{He} = 4.5 \times 10^{-8}$; [11]), an increase in $^{21}\text{Ne}/^{22}\text{Ne}$ should be accompanied by a decrease in $^3\text{He}/^4\text{He}$. Therefore, since the He–Ne isotope systematics are coupled, it is possible to predict the $^3\text{He}/^4\text{He}$ through knowledge of the Ne isotope systematics (see [5,12] for details).

Recent work on Icelandic [13–15] (and East Pacific Rise [16]) lavas has challenged the view that He and Ne isotope systematics are necessarily coupled in all mantle-derived materials. The picture that emerges from the Icelandic work is one where the Ne isotope systematics seem to show virtually no addition of nucleogenic ^{21}Ne (i.e., the neon is purely solar in composition), yet $^3\text{He}/^4\text{He}$ ratios are significantly lower than solar values indicating addition of radiogenic helium. Thus, in three-isotope neon space the gradient of the Icelandic trajectory is much steeper than expected from the helium isotope systematics. In this work, we present another example where the He and Ne isotope systematics are apparently decoupled; however, for the Manus Basin, the decoupling is in the opposite sense. The $^3\text{He}/^4\text{He}$ ratios are consistent with mantle plume involvement, yet the neon isotopes show significant addition of nucleogenic ^{21}Ne . The resulting gradient in three-isotope Ne space is significantly lower than expected. Such an observation is unique in noble gas studies of mantle plumes. In this contribution, we consider the ramifications of this result for both the development of the Manus Basin magma source region and its degassing history through time.

2. Samples and analytical procedures

Samples selected for this study have all been analysed previously for their $^3\text{He}/^4\text{He}$ ratios ([1,2]). They are fresh tholeiitic lavas ($< 130\,000$ yr old), which are well-characterised for their major and trace element chemistry and radiogenic isotope systematics [17]. They are distinguished into the following seven types: MORB-1, MORB-smt, MORB-2, E-MORB, Arc basalt,

back-arc basin basalt (BABB), and Extreme BABB (XBABB) (see discussion in [1,2]). Sample locations and eruption depths are given in Table 1.

Prior to analysis, a binocular microscope was used to select glass samples free of large vesicles, phenocrysts and surface alteration. In all cases, an effort was made to choose a small number of large grains to minimise the surface area to volume ratio. Prior to loading, the glasses were ultrasonically cleaned in a 50:50 acetone–methanol mixture and weighed to ± 0.05 mg.

Neon concentration and isotopic measurements (together with helium abundances) were carried out on a modified VG5400 mass spectrometer [18] using stepped heating as the extraction method. Samples were loaded onto a sample sled and pre-heated under UHV (to 150°C) for several days prior to analysis. Volatiles were extracted in four temperature steps: 600°C, 800°C, 1200°C and 1400°C, each for 30 min duration. Volatiles released during the 600°C steps were not collected since we assume that they are dominated by atmospheric contamination. Similarly, the fusion step (1400°C) was discarded as it contains residual volatiles – generally at blank levels. We report gas yields at the 800°C and 1200°C release steps (Table 1) and use the total concentrations in the following discussion. All results take into account corrections for procedural blanks and contributions to ^{20}Ne and ^{22}Ne due to doubly charged ^{40}Ar and CO_2 respectively. Best-fit regressions of $+/\text{++}$ ratios as a function of H_2 and CO_2 partial pressures were used to estimate the contributions of the doubly charged species (see procedure in [19]). Blanks for ^{20}Ne and ^4He throughout these experiments were in the range of $8 \pm 1 \times 10^{-13}$ cm^3STP and $2 \pm 1 \times 10^{-9}$ cm^3STP , respectively which typically represented $\sim 5\%$ of sample yields.

3. Results

A total of 19 glasses were analysed as part of this study. Neon isotopic compositions together with abundances of neon and helium are reported in Table 1.

3.1. Neon isotopes and abundances

Neon concentrations for the various petrogenic rock-types vary over approximately two orders of magnitude as shown in Fig. 1. In this plot, we illustrate the deviation of the neon isotopic composition of a given sample from the air value using the $\delta^{20}\text{Ne}$ and $\delta^{21}\text{Ne}$ notation (see details in [20]). We note that the arc-type, BABB and XBABB samples all have air-like Ne isotope compositions, regardless of concentration. However, the MORB lavas (particularly MORB-1) show a trend whereby the lowest concentration samples have the highest $\delta^{20}\text{Ne}$ and $\delta^{21}\text{Ne}$ values i.e., they are the least contaminated by air Ne. The standard three-isotope neon plot (Fig. 2) provides further illustration that the majority of the Manus Basin samples lie within error of the air value.

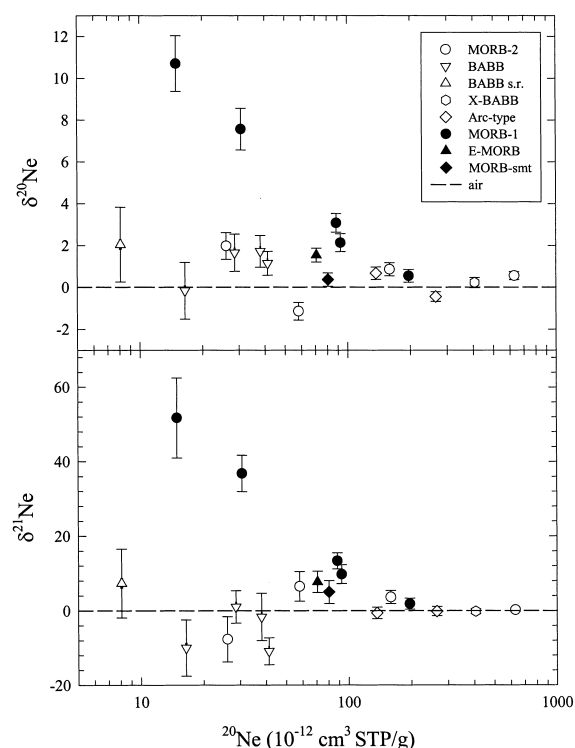


Fig. 1. Plot of the neon isotopic enrichment (in %) of Manus Basin samples relative to air as a function of total ^{20}Ne concentration. Neon isotope ratios are plotted in the delta notation where $\delta^i\text{Ne} = \{(^i\text{Ne}/^{22}\text{Ne}_{\text{sample}}/{}^i\text{Ne}/^{22}\text{Ne}_{\text{air}}) - 1\} \times 100$ ($^i\text{Ne} = ^{20}\text{Ne}$ or ^{21}Ne). The dashed line represents air ($\delta^{20}\text{Ne}$ and $\delta^{21}\text{Ne} = 0$). Errors are given at the 1σ level.

Table 1

Helium and neon abundances and isotopic ratios

Sample (type), location depth ^a , weight	³ He/ ⁴ He ^b (<i>R/R_A</i>)	<i>T</i> (°C)	⁴ He (10 ^{−9} cm ³ STP/g)	²⁰ Ne (10 ^{−12} cm ³ STP/g)	²⁰ Ne/ ²² Ne	²¹ Ne/ ²² Ne
32-5 (MORB-1) 3°25.2'S, 149°00.9'E 2300–2420 m, 282.68 mg	13.5 ± 0.2	800 1200 Total	3390 ± 30 263 ± 2 3650 ± 30	69.0 ± 0.2 23.6 ± 0.1 92.6 ± 0.2	9.95 ± 0.03 10.19 ± 0.06 10.01 ± 0.04	0.0311 ± 0.0008 0.034 ± 0.001 0.0318 ± 0.0007
33-3 (MORB-1) 3°31.7'S, 149°28.7'E 2090–2115 m, 287.60 mg	12.7 ± 0.3	800 1200 Total	5260 ± 50 669 ± 7 5930 ± 50	10.41 ± 0.07 4.54 ± 0.06 14.94 ± 0.09	10.8 ± 0.1 11.0 ± 0.2 10.9 ± 0.1	0.047 ± 0.004 0.037 ± 0.004 0.044 ± 0.003
36-2 (MORB-1) 3°26.8'S, 149°57.8'E 2155–2165 m, 308.67 mg	11.7 ± 0.2	800 1200 Total	8540 ± 70 77.5 ± 0.7 8620 ± 70	77.1 ± 0.2 11.21 ± 0.09 88.3 ± 0.2	10.10 ± 0.02 10.13 ± 0.06 10.10 ± 0.04	0.0329 ± 0.0006 0.032 ± 0.003 0.0329 ± 0.0006
38-3 (MORB-1) 3°19.9'S, 150°04.9'E 2200–2225 m, 298.81 mg	12.7 ± 0.2	800 1200 Total	11700 ± 100 28.1 ± 0.2 11700 ± 100	27.7 ± 0.1 3.06 ± 0.09 30.7 ± 0.2	10.58 ± 0.07 10.2 ± 0.1 10.5 ± 0.1	0.0402 ± 0.002 0.0349 ± 0.003 0.0397 ± 0.001
39-1 (MORB-1) 3°17.4'S, 150°07.7'E 2285–2370 m, 294.95 mg	12.7 ± 0.2	800 1200 Total	3560 ± 30 95.4 ± 0.8 3660 ± 30	186.8 ± 0.3 9.6 ± 0.1 196.4 ± 0.3	9.85 ± 0.02 9.97 ± 0.07 9.85 ± 0.03	0.0295 ± 0.0004 0.031 ± 0.003 0.0295 ± 0.0004
31-1 (E-MORB) 3°30.1'S, 149°15.5'E 2075–2245 m, 311.10 mg	10.3 ± 0.3	800 1200 Total	906 ± 9 63.7 ± 0.6 970 ± 9	52.37 ± 0.09 18.42 ± 0.06 70.8 ± 0.1	9.88 ± 0.03 10.16 ± 0.05 9.95 ± 0.03	0.0302 ± 0.0009 0.034 ± 0.002 0.0312 ± 0.0008
23-2 (MORB-smt.^c) 3°52.4'S, 149°58.0'E 1390–1865 m, 306.85 mg	15.1 ± 0.3	800 1200 Total	2220 ± 20 631 ± 6 2850 ± 20	67.4 ± 0.1 13.11 ± 0.06 80.5 ± 0.1	9.79 ± 0.02 10.06 ± 0.07 9.84 ± 0.03	0.030 ± 0.001 0.034 ± 0.002 0.0304 ± 0.0009
41-1 (MORB-2) 3°12.0'S, 150°12.5'E 2375–2400 m, 311.70 mg	0.8 ± 0.8	800 1200 Total	3.48 ± 0.05 2.05 ± 0.03 5.53 ± 0.05	626.9 ± 0.7 2.01 ± 0.06 628.9 ± 0.7	9.85 ± 0.01 10.1 ± 0.2 9.85 ± 0.02	0.0291 ± 0.0001 0.031 ± 0.007 0.0291 ± 0.0002
42-1 (MORB-2) 3°09.6'S, 150°17.1'E 2480–2490 m, 313.30 mg	11.4 ± 0.3	800 1200 Total	213 ± 2 7.86 ± 0.08 221 ± 2	10.18 ± 0.05 15.84 ± 0.06 26.02 ± 0.07	10.12 ± 0.08 9.91 ± 0.06 9.99 ± 0.06	0.029 ± 0.004 0.025 ± 0.002 0.027 ± 0.002
44-1 (MORB-2) 3°05.2'S, 150°23.7'E 2600–2630 m, 287.05 mg	12.1 ± 0.2	800 1200 Total	3570 ± 30 18.7 ± 0.2 3590 ± 30	135.2 ± 0.3 24.2 ± 0.1 159.3 ± 0.3	9.89 ± 0.02 9.85 ± 0.06 9.88 ± 0.03	0.0303 ± 0.0005 0.029 ± 0.001 0.0301 ± 0.0005
45-3 (MORB-2) 3°03.9'S, 150°27.3'E 2570–2670 m, 297.68 mg	12.7 ± 0.2	800 1200 Total	1190 ± 10 110.1 ± 0.9 1300 ± 10	52.7 ± 0.1 5.34 ± 0.09 58.0 ± 0.2	9.61 ± 0.02 10.5 ± 0.1 9.69 ± 0.04	0.030 ± 0.001 0.041 ± 0.005 0.031 ± 0.001
28-1 (BABB) 3°39.4'S, 149°40.4'E 2370–2440 m, 285.30 mg	12.0 ± 0.3	800 1200 Total	2010 ± 20 5.17 ± 0.05 2020 ± 20	28.1 ± 0.1 0.57 ± 0.09 28.6 ± 0.1	9.95 ± 0.05 10.7 ± 0.4 9.96 ± 0.9	0.029 ± 0.001 0.05 ± 0.01 0.029 ± 0.001

Table 1 (continued)

Sample (type), location depth ^a , weight	³ He/ ⁴ He ^b (<i>R/R_A</i>)	<i>T</i> (°C)	⁴ He (10 ^{−9} cm ³ STP/g)	²⁰ Ne (10 ^{−12} cm ³ STP/g)	²⁰ Ne/ ²² Ne	²¹ Ne/ ²² Ne
34-1 (BABB) 3°36.4'S, 149°43.9'E 2445–2510 m, 283.40 mg	12.1 ± 0.3	800 1200 Total	1130 ± 10 0.38 ± 0.02 1140 ± 10	37.1 ± 0.1 0.90 ± 0.09 38.0 ± 0.1	9.96 ± 0.05 10.4 ± 0.3 9.97 ± 0.08	0.028 ± 0.002 0.05 ± 0.02 0.029 ± 0.002
40-6 (BABB) 3°14.9'S, 150°07.7'E 2300–2310 m, 305.60 mg	11.5 ± 0.4	800 1200 Total	237 ± 2 4.09 ± 0.04 241 ± 2	40.8 ± 0.1 0.42 ± 0.09 41.2 ± 0.1	9.91 ± 0.03 10.5 ± 0.4 9.91 ± 0.06	0.026 ± 0.001 0.01 ± 0.01 0.026 ± 0.001
41-3 (BABB) 3°12.0'S, 150°12.5'E 2375–2400 m, 308.10 mg	11.6 ± 0.8	800 1200 Total	44.1 ± 0.4 0.87 ± 0.02 45.0 ± 0.4	15.9 ± 0.1 0.51 ± 0.09 16.5 ± 0.1	9.78 ± 0.06 10.0 ± 0.3 9.8 ± 0.1	0.027 ± 0.002 0.027 ± 0.02 0.027 ± 0.002
14-5 (arc-type) 3°42.9'S, 152°10.4'E 1755–1950 m, 318.10 mg	1.0 ± 0.2	800 1200 Total	1.90 ± 0.03 1.76 ± 0.03 3.67 ± 0.04	264.5 ± 0.3 0.4 ± 0.1 264.9 ± 0.3	9.76 ± 0.02 10.5 ± 0.4 9.76 ± 0.02	0.0289 ± 0.0004 0.042 ± 0.009 0.0290 ± 0.0004
17-1 (arc-type) 3°44.4'S, 151°38.8'E 1685–1860 m, 308.50 mg	0.6 ± 0.3	800 1200 Total	3.31 ± 0.05 2.89 ± 0.05 6.20 ± 0.07	136.1 ± 0.3 0.92 ± 0.05 137.0 ± 0.2	9.86 ± 0.2 10.7 ± 0.4 9.87 ± 0.03	0.0288 ± 0.0004 0.033 ± 0.007 0.0288 ± 0.0004
19-12 (BABB s.r.) 3°45.1'S, 151°09.5'E 2625–2635 m, 289.50 mg	3 ± 2	800 1200 Total	14.1 ± 0.2 2.43 ± 0.05 16.5 ± 0.2	7.03 ± 0.05 1.04 ± 0.07 8.07 ± 0.09	9.91 ± 0.09 10.6 ± 0.5 10.0 ± 0.2	0.031 ± 0.002 0.03 ± 0.01 0.031 ± 0.003
24-9 (XBABB) 3°32.7'S, 149°51.4'E 2155–2185 m, 299.45 mg	8.7 ± 0.4	800 1200 Total	9.5 ± 0.1 7.55 ± 0.07 17.1 ± 0.1	405.8 ± 0.6 0.5 ± 0.1 406.3 ± 0.6	9.82 ± 0.01 10.3 ± 0.3 9.82 ± 0.02	0.0290 ± 0.0002 0.03 ± 0.02 0.0290 ± 0.0002

All errors are given to 1σ except for the ³He/⁴He ratios, which are reported at the 2σ level.

^a Start and finish depth of dredge.

^b From [1,2].

^c Sample 23-2 was dredged from a seamount.

However, we note that seven samples (five MORB-1, one E-MORB and one MORB-smt) are displaced from air, and lie on a linear array (dashed line in Fig. 2) passing through the atmospheric value. The correlation line defined by the seven samples has a gradient less than that of the MORB correlation line, i.e. samples show higher ²¹Ne/²²Ne ratios relative to MORB [3] for a given ²⁰Ne/²²Ne ratio. This observation is in stark contrast to other high ³He/⁴He samples, e.g. from Iceland, Loihi, Samoa, the Kola Peninsula and Reunion [6,13–15,21–23], that consistently fall on correlation lines with slopes steeper than the MORB Ne-trend [5]. We believe that this is the

first observation of oceanic lavas having high ³He/⁴He ratios and such nucleogenic neon isotope ratios.

3.2. Helium isotopes and abundances

As noted previously [1,2], the ³He/⁴He ratios of the Manus Basin samples are generally higher than the MORB-value of 8 ± 1 *R_A*. The MORB-type samples range from 10.3 to 15.1 *R_A* (with the exception of one sample, 41-1, which is air-like), and the BABB values range from 11.5 to 12.1 *R_A* (apart from one sample from the Southern Rift at 3 *R_A*). One XBABB sample has a ³He/⁴He ratio

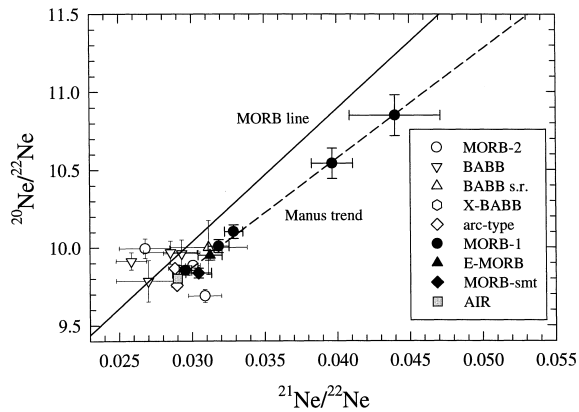


Fig. 2. Traditional three-isotope neon plot showing seven Manus Basin samples (filled symbols) trending towards an endmember composition more nucleogenic than typical MORB. The Manus trend (dotted line) is represented by the equation $(^{20}\text{Ne}/^{22}\text{Ne}) = 70.352(^{21}\text{Ne}/^{22}\text{Ne}) + 7.756$ ($r^2 = 0.994$). This line was calculated from an error-weighted regression through the seven data points and the air value. Also shown for comparison is the MORB trend [3] $(^{20}\text{Ne}/^{22}\text{Ne} = 85.06(^{21}\text{Ne}/^{22}\text{Ne}) + 7.47)$. Errors are reported at the 1σ level.

of $8.7 R_A$ and the arc-type samples have air-like $^3\text{He}/^4\text{He}$ values.

The helium abundances (Table 1) are highly variable ranging from 4 to 11 700 ncm^3 STP/g. In the majority of cases, most of the helium was released at the 800°C step, which we take to represent the vesicle-sited gas. The distribution of He between vesicles and glass along with its relationship to CO_2 will be considered elsewhere (Shaw et al., manuscript in preparation).

4. Discussion

A fundamental tenet of the solar neon hypothesis [5] is that the He and Ne isotope systematics are coupled. Therefore, for a given Ne isotopic composition, a corresponding $^3\text{He}/^4\text{He}$ ratio can be calculated (see methodology in [5]). Using the correlation line of samples in this work (Fig. 2) and assuming coupled He–Ne systematics (where $^3\text{He}/^{22}\text{Ne} = 7.7$ as in [12]), we would predict the Manus Basin samples to have a $^3\text{He}/^4\text{He}$ ratio of $4.6 R_A$. The average $^3\text{He}/^4\text{He}$ ratio of these seven samples, however, is $12.7 R_A$ suggesting

an apparent decoupling between the He and Ne isotope systematics. An alternative way of viewing this apparent decoupling is to calculate the difference in $^{21}\text{Ne}/^{22}\text{Ne}$ between the Manus sample trend and the value predicted assuming the solar neon hypothesis for a $^3\text{He}/^4\text{He}$ ratio of $12.7 R_A$ (Fig. 3 shows this schematically). Using this approach, we calculate a shift in the $^{21}\text{Ne}/^{22}\text{Ne}$ ratio ($\Delta^{21}\text{Ne}/^{22}\text{Ne}$) of 0.035 assuming a solar $^{20}\text{Ne}/^{22}\text{Ne}$ ratio of 13.8. If the solar neon component in Earth originated from a trapped meteoritic component ($^{20}\text{Ne}/^{22}\text{Ne} = 12.5$) rather than a pristine solar component ($^{20}\text{Ne}/^{22}\text{Ne} = 13.8$) then the calculated value of the shift would be lower (see [24,25] for a detailed discussion of Earth's mantle neon component); however, the trajectories in three-isotope neon space would be the same. In either case, it appears that the Manus Basin samples are characterised by anomalously nucleogenic neon.

In the following discussion, we investigate five possible mechanisms by which this nucleogenic neon component could have been produced: (1) crustal contamination; (2) devolatilisation of the Solomon Sea plate; (3) the addition of slab-derived neon from ancient recycled oceanic litho-

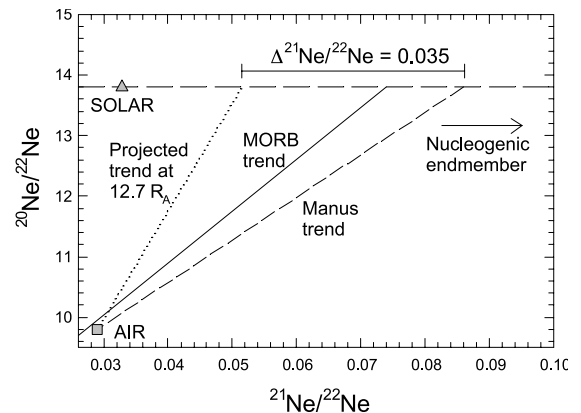


Fig. 3. Neon isotope plot showing the Manus trend (dashed line), which would extrapolate to a $^{21}\text{Ne}/^{22}\text{Ne}$ ratio of 0.0859 at a mantle source (solar) value of $^{20}\text{Ne}/^{22}\text{Ne} = 13.8$ [37]. In contrast, we would anticipate an extrapolated $^{21}\text{Ne}/^{22}\text{Ne}$ value of 0.0514 for a $^3\text{He}/^4\text{He}$ ratio of $12.7 R_A$ (the mean value of the seven Manus samples which define the Manus trend) assuming the solar Ne hypothesis [5]. Thus, the Manus Basin samples are more nucleogenic than expected by $\Delta^{21}\text{Ne}/^{22}\text{Ne} = 0.035$.

sphere; (4) sampling from a previously degassed source, or (5) a heterogeneous deep mantle plume component preserved since Earth's accretion.

4.1. Crustal contamination

The dominant reaction producing ^{21}Ne is the (α, n) reaction on ^{18}O . This reaction is particularly important in crustal lithologies due to the relatively high abundance of α -emitters (U- and Th-decay series nuclides). There are three possible mechanisms by which crustal contamination can occur: (1) in situ post-eruptive in-growth (2) magma ageing, and (3) wallrock contamination.

A potential means of adding nucleogenic neon to the Manus Basin lavas is in situ production since the time of eruption. Using a maximum age estimate of 130 000 yr [17], a U concentration of 0.04 ppm (based on average values in Manus lavas [17]), together with the ^4He production rate ($2.15 \times 10^{-7} \text{ cm}^3\text{STP/ppm U/Myr}$ [26]) and the $^{21}\text{Ne}/^4\text{He}$ production ratio (4.5×10^{-8} [11]), we calculate $5.03 \times 10^{-17} \text{ cm}^3\text{STP } ^{21}\text{Ne/g}$ can be produced. This amount is $< 5\%$ of the lowest concentration of excess nucleogenic ^{21}Ne in any of the 7 samples lying along the so-called Manus Ne trajectory (sample 23-2 has $0.11 \times 10^{-14} \text{ cm}^3\text{STP/g}$ – see below for details how this value is calculated). Therefore, due to the young age of these lavas, we can discount post-eruptive in-growth as a possible means of altering the $^{21}\text{Ne}/^{22}\text{Ne}$ ratio.

In the second case, radiogenic/nucleogenic gases may be added to magma prior to eruption via the process of magma ageing in crustal magma chambers [27]. We evaluate this possibility with reference to one of the seven samples lying along the so-called Manus Ne trajectory (sample 33-3 has a total amount of $^{21}\text{Ne} = 6.0 \times 10^{-14} \text{ cm}^3\text{STP/g}$). First, we calculate the proportion of air Ne in the sample assuming that its measured $^{20}\text{Ne}/^{22}\text{Ne}$ ratio (10.9) reflects mixing between air (9.8) and 'mantle+nucleogenic' Ne ($^{20}\text{Ne}/^{22}\text{Ne} = 13.8$). The fraction of air Ne in the sample = 0.73 which means approximately $1.66 \times 10^{-14} \text{ cm}^3\text{STP/g}$ (0.27 of the total) is comprised of mantle neon + ^{21}Ne excess. We can further resolve this non-air ^{21}Ne into mantle ^{21}Ne

($0.99 \times 10^{-14} \text{ cm}^3\text{STP/g}$) and excess ^{21}Ne ($0.67 \times 10^{-14} \text{ cm}^3\text{STP/g}$) assuming that the mantle component has $^{20}\text{Ne}/^{22}\text{Ne} = 13.8$ and a $^{21}\text{Ne}/^{22}\text{Ne} = 0.05146$ – the predicted value from the solar hypothesis for a measured $^3\text{He}/^4\text{He}$ ratio of $12.7 R_A$. Because of the constant $^{21}\text{Ne}/^4\text{He}$ production ratio of 4.5×10^{-8} , addition of this amount of ^{21}Ne would simultaneously add $0.15 \times 10^{-6} \text{ cm}^3\text{STP } ^4\text{He/g}$. A residence or ageing time of 17.2 Ma is required to produce this much ^4He (and ^{21}Ne) assuming a [U] of 0.04 ppm [17]. Calculated residence times vary between 21 Ma (samples 36-2 and 38-3) and 2.8 Ma (sample 23-2) for the other lavas which lie along the Manus trajectory (Fig. 2). These values are minimum estimates as magmatic degassing could have occurred simultaneously with production of nucleogenic ^{21}Ne over these calculated ages. These times are far in excess of postulated crustal residence ages of magma, which are generally on the order of $< 10^5$ yr [28,29]. Moreover, such long residence times are not consistent with the relatively young age of opening for the Manus spreading centre (estimated at 0.78 Ma [30]).

The third possibility involves addition of nucleogenic neon through crustal assimilation as magma ascends to the surface. This scenario also seems unlikely given that all samples lying along the 'nucleogenic MORB trajectory' (Fig. 2) were erupted along an active spreading centre or within an extensional zone. At such localities, the opportunity for crustal interaction appears minimal and, in any case, not greater than for mid-ocean spreading centres – where there is no evidence of such nucleogenic neon additions. Furthermore, even if crustal assimilation had occurred, the oceanic crust where these lavas are sited is young (< 1 Ma) and unlikely to have grown-in and stored sufficient ^4He and ^{21}Ne to modify source $^3\text{He}/^4\text{He}$ and $^{21}\text{Ne}/^{22}\text{Ne}$ ratios.

4.2. Devolatilisation of the Solomon Sea slab

Another possibility for the source of the nucleogenic neon is the down-going Solomon Sea plate. The source of the neon associated with subduction-related processes could be fluids or melts derived from either subducted sediments or the

underlying oceanic crust. In the case of sediments, noble gas data obtained on a variety of sediment types of different ages [31] indicate the overwhelming dominance of air-like isotope signatures. Therefore, even with prolonged ageing, an air-like neon signature would be expected to dominate over any neon produced in situ. For subducted sediments to be considered a viable possibility, the in situ-produced Ne component would have to be retained preferentially in the sediment matrix while the air-like component would have to be lost. One potential mechanism that could effect this separation would be loss of the air-like Ne via pore fluids which are expected to be expelled during sediment burial and diagenesis. The question then becomes whether the sediment matrix could still retain the nucleogenic Ne component. Diffusion experiments on interplanetary dust particles (IDP) in deep-sea sediments [32] suggest that neon is extremely mobile in sedimentary material at relatively moderate temperatures (200–500°C). If the diffusivity of Ne in sediments is similar to that in IDPs, it suggests that trapped (in situ-produced) Ne would be completely released from sediments at relatively shallow subduction depths, and probably able to escape to the forearc region. Previous studies have emphasised the difficulty of subducting helium [33]; the above considerations suggest that this may also be the case for neon.

The subducted oceanic crust is also an improbable candidate for the nucleogenic Ne signal for the same reasons. First, any intrinsic volatiles would be expected to have been modified during seawater alteration to atmospheric-like values. Indeed, if the analogy of on-land ophiolite complexes is appropriate for oceanic crust, the zone of alteration may extend to depths of 3–5 km (e.g. [34]). Therefore, a significant fraction of the uppermost oceanic crust may have an air-like neon isotope composition. Second, we would have to invoke the scenario of preferential loss of the air-like Ne component, and the retention of the nucleogenic neon despite heating by subduction which would accentuate diffusional loss.

Perhaps the most persuasive argument against a slab origin for the nucleogenic neon in this study is that the most pronounced ^{21}Ne excess signal is

observed in the MORB Type-1 lavas (i.e., the four highest $^{21}\text{Ne}/^{22}\text{Ne}$ ratios along the Manus trend are found in MORB-1 lavas). Although some geochemical support for slab involvement is apparent in the Manus sample suite as a whole [17], the MORB-1 magmas are the magma-type least affected by subduction-related processes and it is these samples that possess the most nucleogenic neon.

4.3. *Ancient recycled slab*

The nucleogenic neon signal could result from the addition of slab-derived neon from ancient oceanic lithosphere, recycled via the deep mantle and entrained by the Manus plume en-route to the surface. A recycled altered oceanic crustal component has been suggested to explain anomalously low $\delta^{18}\text{O}$ values (5.3–5.8‰) found in the same samples [2]. In this scenario, the original noble gas signature (along with other volatiles) of the recycled slab would have been lost in the so-called ‘subduction barrier’ [31] but over time the slab would acquire a nucleogenic signature through in situ production.

To assess the plausibility of this suggestion, we have modelled the above scenario using an estimate of the ^{21}Ne concentration in 2-Ga-old subducted slab. We calculate that $1.3 \times 10^{-12} \text{ cm}^3\text{STP}$ $^{21}\text{Ne}/\text{g}$ would be produced in 2 Ga assuming a [U] value of 0.066 ppm for altered oceanic crust [35], a $^{21}\text{Ne}/^4\text{He}$ production ratio of 4.5×10^{-8} [11], a ^4He production rate of $2.15 \times 10^{-7} \text{ cm}^3\text{STP}/\text{ppm U}/\text{Myr}$ (which assumes a Th/U ratio of 3.3 and includes ^4He produced from Th decay in its production rate; see [26] for details) and 100% retention of produced ^{21}Ne . If the initial (mantle) plume ^{21}Ne concentration was $3.35 \times 10^{-12} \text{ cm}^3\text{STP}/\text{g}$ [36] with a $^{21}\text{Ne}/^{22}\text{Ne}$ ratio of 0.032 [37] then a slab/plume mixing ratio of ~ 4.4 is required to produce a resultant $^{21}\text{Ne}/^{22}\text{Ne}$ ratio of 0.0859. This slab/plume ratio could be as low as ~ 1 (i.e. a mixture of 50% slab and 50% plume components) if the altered oceanic crust had a U content of 0.3 ppm [38].

Both estimates of the slab/plume mixing ratio are exceptionally high. For comparison, Marcantonio et al. [39] used Os–Pb relationships of lavas

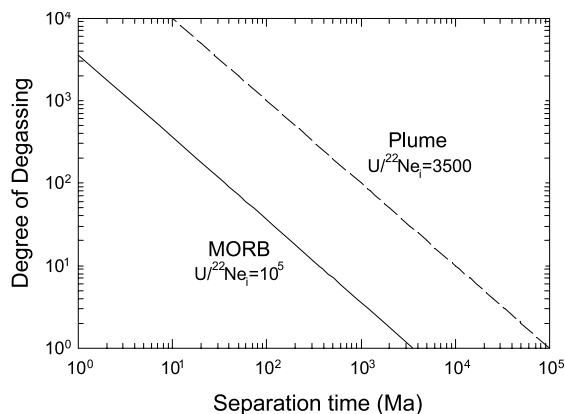


Fig. 4. Plot showing the relationship between degree of degassing and the required time since degassing (separation time) in order to produce a $\Delta^{21}\text{Ne}/^{22}\text{Ne} = 0.035$ for a MORB source (solid line) and plume source (dashed line). Initial source ($\text{U}/^{22}\text{Ne}$) ratios from Porcelli and Wasserburg [36].

erupted in the Canary Islands to constrain the contribution of ancient recycled crust to this (HIMU-type) mantle source to between 15 and 30%. In contrast, the Manus Basin lavas have a maximum $^{206}\text{Pb}/^{204}\text{Pb}$ ratio of 18.8 [17] which is considerably lower than that found at the Canaries. Therefore, if the recycled component is traced by radiogenic lead ($^{206}\text{Pb}/^{204}\text{Pb} > 20$), the Manus Basin data imply that the proportion of ancient slab must be considerably lower ($< 10\%$). Similarly, Macpherson et al. [2] used O-isotopes to limit the contribution of a deep hydrothermally altered oceanic crust component to $< 35\%$. Again this estimate is substantially less than the above values (50–85%) based on our Ne isotope results. More significantly, if the proportion of ancient slab material was as large as calculated above, the present sample suite would be expected to possess exceptionally strong slab-derived trace element (e.g., La/Nb) signals – these are conspicuously absent in the Manus Basin lavas.

4.4. A previously degassed plume source

During partial melting of a mantle source, noble gases are expected to act as incompatible trace elements partitioning into the melt phase as a function of their relative partition coefficients (D). Experimental data indicate that the partition

coefficients of He and Ne are similar ($D_{\text{He}} = 0.02$ – 0.7 , $D_{\text{Ne}} = 0.03$ – 0.6 [40] and references therein; and $D_{\text{He}} \approx D_{\text{Ne}}$, [41]). This implies that during the formation of melt, little or no fractionation between helium and neon is expected. However, this is not the case during formation of a free vapour phase. During melt–vapour partitioning (degassing), Ne will partition preferentially into the vapour phase, as its solubility (S) in a tholeiitic melt is less than that of helium ($S_{\text{He}}/S_{\text{Ne}} \sim 2$, [42]; $S_{\text{He}}/S_{\text{Ne}} \sim 10$, [43]). The consequence of a degassing event, therefore, is to increase the He/Ne ratio in the degassed melt compared with its starting value.

Another corollary of volatile loss is that the intrinsic Ne and He concentrations in the degassed melt would be reduced relative to U and Th: therefore, as a function of time and extent of degassing, $^3\text{He}/^4\text{He}$ ratios would decrease and $^{21}\text{Ne}/^{22}\text{Ne}$ ratios increase in the melt phase residual after degassing. Thus, the original neon isotope values could be altered more readily than those of helium due to the lower (residual) neon contents compared to helium. Such a scenario has been invoked to explain anomalously nucleogenic neon in samples from the North Chile Ridge [44] although, in that case, the $^3\text{He}/^4\text{He}$ ratios were less than MORB. Here, we explore whether this type of model can explain the nucleogenic neon measured in the Manus Basin samples.

The principal point to consider is whether the plume-degassing hypothesis can account for the shift in $^{21}\text{Ne}/^{22}\text{Ne}$ ($\Delta = 0.035$) in reasonable time periods and/or degrees of degassing without drastically altering the helium isotope systematics. In Fig. 4, we plot degree of degassing versus the time required following degassing (the separation time) to grow-in the amount of ^{21}Ne to produce the requisite $\Delta^{21}\text{Ne}/^{22}\text{Ne}$. We see that for a plume source ($\text{U}/^{22}\text{Ne} \sim 3500$ [36]) the grow-in time would be 1 Ga if degassing of initial Ne was 100-fold, or 100 Ma if degassing was 1000-fold. For comparison, the separation times for a MORB-type source ($\text{U}/^{22}\text{Ne} \sim 10^5$; [36]) would be 36 Ma and 3.6 Ma for 100- and 1000-fold degassing factors respectively (because Manus Basin $^3\text{He}/^4\text{He}$ ratios \gg MORB such a source is clearly inappropriate in the present case). Since

^{21}Ne in-growth is accompanied by ^4He production, the $^3\text{He}/^4\text{He}$ ratio of the source will decrease simultaneously with increasing $^{21}\text{Ne}/^{22}\text{Ne}$ although to a lesser extent due to greater retention of helium after the degassing event (helium is more soluble in basaltic melt than neon). Assuming a Rayleigh ‘open-system’ style of degassing and using the above degassing factors ($\times 100$ and $\times 1000$) with associated grow-in times for a plume source (1 Ga and 100 Ma respectively), an initial $^3\text{He}/^4\text{He}$ ratio of $30 R_A$ (with $5.6 \times 10^{-5} \text{ cm}^3\text{STP } ^4\text{He/g}$ [36]) will decrease to $26.6 R_A$ or $29.5 R_A$ respectively for a $S_{\text{He}}/S_{\text{Ne}} = 10$ [43], and to $16.7 R_A$ or $24.0 R_A$ respectively for a $S_{\text{He}}/S_{\text{Ne}} = 2$ [42]. Increased separation times and/or degassing factors will lower these $^3\text{He}/^4\text{He}$ ratios even further.

As illustrated above, production of nucleogenic neon (plus retention of $^3\text{He}/^4\text{He}$ ratios $> \text{MORB}$) by the plume-degassing mechanism depends upon two key factors: (1) the extent of degassing and (2) the separation time. There is evidence to suggest that helium loss during plume-degassing can be extensive. For example, Hilton et al. [45] estimated that the (plume) source region of Kilauea volcano ($15 R_A$) is degassed by a factor of 2000 compared to model-dependent values for a lower mantle source [36]. Thus, one of the key constraints of the plume-degassing hypothesis under consideration, i.e., the requirement of significant degrees of degassing, may be a feature common to the process of plume impingement with oceanic lithosphere. The second factor is the separation time, which can be relatively short provided the degassing factor is sufficiently high. For example, a conservative 100-fold decrease in plume He (corresponding to 10^4 decrease in Ne for a Rayleigh distillation with $S_{\text{He}}/S_{\text{Ne}} = 2$) would require a separation time of only 10 Ma in order to achieve a $\Delta^{21}\text{Ne}/^{22}\text{Ne} = 0.035$. From these considerations, therefore, we suggest that the degassing factor and/or separation times required to produce the nucleogenic neon in the Manus Basin lavas fall within reasonable limits.

The other major constraint is the requirement that previously degassed melt forms part of the source region. Macpherson and Hall [46] have

tracked the Manus Basin plume from the middle Eocene to the present-day, and it appears to have experienced a complex history of interaction with plate boundaries in the western Pacific. Although speculative, we suggest that plume material, which was extensively but incompletely degassed (therefore able to retain high $^3\text{He}/^4\text{He}$ ratios), could have been retained in the mantle in or around the upwelling plume for the relatively short time periods suggested above to produce the requisite $\Delta^{21}\text{Ne}/^{22}\text{Ne}$. Therefore, we conclude that plume-degassing (following melting but prior to eruption) could provide a viable means by which to effectively decouple He–Ne systematics with the superimposition of a nucleogenic neon signal onto a plume-like $^3\text{He}/^4\text{He}$ ratio.

4.5. Lower mantle heterogeneity

Finally, we consider the possibility that the mantle source supplying the Manus Basin has distinct He–Ne systematics compared to other plume sources. Recent work [13–15] on the Icelandic plume has identified cases of apparently decoupled He–Ne isotope systematics: solar-like Ne isotope compositions are accompanied by $^3\text{He}/^4\text{He}$ ratios ranging from $12\text{--}29 R_A$ – significantly lower than estimated solar $^3\text{He}/^4\text{He}$ ratios ($145 R_A$ [15]). This observation has been attributed to either a high concentration of solar Ne in the source, which has overwhelmed nucleogenic ^{21}Ne , or extremely low (U+Th) source contents. The Manus Basin glasses uniquely reveal a decoupling in the opposite sense: the $^3\text{He}/^4\text{He}$ ratios preserve evidence of a plume contribution whereas the $^{21}\text{Ne}/^{22}\text{Ne}$ systematics are more evolved, consistent with a source having lower solar Ne contents and/or high U and Th.

A simple test of whether the Manus Basin lavas sample a distinct source is to consider their $^3\text{He}/^{22}\text{Ne}$ ratio in comparison to other high- $^3\text{He}/^4\text{He}$ lavas. Following the approach of Moreira et al. [15], the $^3\text{He}/^{22}\text{Ne}$ ratio of the Manus Basin source can be calculated from the following equation:

$$\{(^4\text{He}/^3\text{He})_{\text{Manus}} - (^4\text{He}/^3\text{He})_{\text{solar}}\} =$$

$$(^4\text{He}/^{21}\text{Ne})^* \{ (^{21}\text{Ne}/^{22}\text{Ne})_{\text{Manus}} -$$

$$(^{21}\text{Ne}/^{22}\text{Ne})_{\text{solar}} \times (^{22}\text{Ne}/^3\text{He})$$

where $(^{21}\text{Ne}/^4\text{He})^* = 4.5 \times 10^{-8}$ [11], $^4\text{He}/^3\text{He}_{\text{solar}} = 5000$ [15], $^{21}\text{Ne}/^{22}\text{Ne}_{\text{solar}} = 0.032$ [37], $^3\text{He}/^4\text{He}_{\text{Manus}} = 12.7 R_A$ (this work) and $^{21}\text{Ne}/^{22}\text{Ne}_{\text{Manus}} = 0.0859$ (this work). In this way, we determine the $^3\text{He}/^{22}\text{Ne}$ ratio of the Manus Basin source to be 23.4. This value is much higher than estimates of 4 for Iceland, 9 for Loihi Seamount and 11 for MORB-mantle (see [15]). It implies that the plume source sampled by Manus Basin lavas is highly fractionated in its He–Ne elemental abundance compared to the Icelandic and Loihi plume sources. Clearly, the fact that its He/Ne ratio is higher than that of other mantle sources (including MORB mantle) is why its $^{21}\text{Ne}/^{22}\text{Ne}$ ratio is more susceptible to change than its $^3\text{He}/^4\text{He}$ ratio.

One possible process that can fractionate He from Ne is diffusion: however, this mechanism would require a helium and/or neon concentration gradient between the Manus Basin source region and ambient mantle. We suggest that for a significant portion of the history of the Manus plume, assuming a lower mantle derivation, it would have a helium concentration greater than that of ambient (upper) mantle [36]. In this case, a transient lowering of the $^3\text{He}/^{22}\text{Ne}$ ratio in the plume would occur by diffusive exchange with lower concentration ambient mantle due to the higher diffusivity of He [47]. Degassing of the plume would lead to lower He concentrations and thus eventually to a He–Ne fractionation in the opposite sense, i.e. a transient increase in the plume $^3\text{He}/^{22}\text{Ne}$ ratio. However, if the estimated separation time (~ 10 Ma) is correct, He would diffuse a maximum of only 1.8 m (for a diffusion coefficient (D_{He}) $\sim 10^{-4}$ cm²/s [26]) and thus a negligible volume of degassed plume would be affected. We conclude, therefore, that diffusion is not a viable mechanism by which to change the He/Ne ratio of the Manus magma source.

If diffusion fails to account for the high He/Ne ratios, the alternative is that the high He/Ne (Manus plume) ratio is inherently representative of its source region in the lowermost mantle. If

correct, then this feature would have to have been (1) a primary feature of the Earth accretion process, (2) resistant to > 4 Ga of convective mantle mixing, and (3) sampled uniquely by the Manus Basin plume. Should these conditions prevail, there are important implications for modelling the means by which Earth acquired its noble gases. Two models – both variants of the magma ocean hypothesis (e.g., [48]) have been suggested as capable of fractionating mantle He/Ne ratios. In the first case, noble gases acquired during accretion would have degassed from the magma ocean as a function of their relative solubilities. The greater solubility of He in silicate melts would result in preferential degassing of Ne. This hypothesis is supported by the fact that mantle-derived materials have an average $^3\text{He}/^{22}\text{Ne}$ (7.7) which is twice that of the solar ratio (3.8) [49], consistent with equilibrium degassing models assuming $S_{\text{He}}/S_{\text{Ne}} = 2$ [42]. In the Manus Basin case, the high He/Ne ratio ($^3\text{He}/^{22}\text{Ne} = 23.4$) of the source could be achieved only by both extensive and open-system style degassing. For example, to increase the $^3\text{He}/^{22}\text{Ne}$ ratio from the solar value to 23.4 in the (residual) magma ocean by Rayleigh fractionation would involve loss of 84% of the original helium (assuming $S_{\text{He}}/S_{\text{Ne}} = 2$ [42]).

The second possibility is that the Earth acquired its noble gases by ingassing of solar noble gases from a massive proto-atmosphere [50]. In this scenario, once the Earth had reached sufficient size it would become capable of capturing rare gases gravitationally from the solar nebula, thus forming a massive atmosphere of solar gases. The accretional energy along with a heat-trapping atmosphere would provide the heat necessary to induce melting of the growing Earth [51]. The solar rare gases from the atmosphere would have dissolved into a molten silicate Earth as a function of their solubilities [52].

Consequently, due to the greater solubility of He, ingassing would lead to an enhanced He/Ne ratio in the Earth's mantle. Again, we would have to invoke an open-system style of ingassing to produce such a high Manus Basin $^3\text{He}/^{22}\text{Ne}$ ratio. Honda and McDougall [49] discuss the relative merits of the two alternative models: at present, however, there are insufficient grounds for favour-

ing one model over the other. Therefore, at this stage, we suggest that the magma ocean hypothesis (either variant) could also be the potential cause of the high $^3\text{He}/^{22}\text{Ne}$ ratios found in the Manus Basin. Further studies on the Ne isotope systematics of plume-related materials may reveal if the presence of nucleogenic neon is a widespread phenomenon giving greater credence to this heterogeneous mantle hypothesis.

5. Summary

This study of the Manus Basin shows new evidence for a mantle source in which the helium and neon isotope systematics are decoupled. Significantly, the gradient of the Manus slope in three-isotope neon space is lower than that of the MORB line, a unique observation for high $^3\text{He}/^4\text{He}$ lavas. A mechanism is required to enrich the source in nucleogenic neon, without obscuring the plume signature manifested in the $^3\text{He}/^4\text{He}$ ratios ($> \text{MORB}$).

We have assessed a number of possible scenarios to explain these observations. Post-eruptive contamination (by in-growth of ^{21}Ne) cannot produce sufficient nucleogenic neon to affect measured $^{21}\text{Ne}/^{22}\text{Ne}$ values. Crustal contamination immediately prior to eruption was dismissed also due to the fact that the crust through which magma erupts is young and therefore lacking in a discernible nucleogenic neon signature. Likewise, magma residence times would have to be unrealistically large if the nucleogenic neon signal were to be grown-in through magma ageing. Although a slab component is present in some Manus Basin lavas, it is unlikely that slab-related noble gases would survive the subduction process to erupt at the Manus spreading centre. Addition of ancient recycled crust to a plume-like source was also evaluated but the proportion of ancient crust required would be too large to account for the observed geochemical traits of the Manus Basin lavas.

We conclude, therefore, that the source supplying the Manus Basin must have a highly fractionated He/Ne ratio (at least twice that of the MORB source) to account for the nucleogenic

neon. We present two scenarios to explain this feature. Either the Manus Basin mantle has undergone extensive degassing followed by a period of in-growth of ^{21}Ne , or the fractionated He–Ne systematics were inherited during the accretion process. In the first case, plume material may have undergone degassing through interaction with the oceanic lithosphere and become re-entrained in a region of plume upwelling.

Alternatively, He and Ne may be heterogeneously distributed in the source region of mantle plumes, presumably as a result of the accretion process. Additionally, this region of anomalous mantle must have survived ~ 4 Ga of mantle convection processes acting to homogenise the source. Gas partitioning between a magma ocean and proto-atmosphere has the potential to produce the requisite $^3\text{He}/^{22}\text{Ne}$ ratio but would involve an open-system degassing/ingassing mechanism. At present, there are insufficient grounds for distinguishing between the two possibilities.

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