Tracing the origin of continental HIMU-like intraplate volcanism using magnesium isotope systematics

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Abstract

The Mg isotopic compositions of 11 peridotite xenoliths and 21 HIMU-like intraplate basaltic lavas from New Zealand's South Island and Antipodes Islands, respectively, provide new evidence bearing on the origin of continental intraplate alkaline basalts. Olivine and orthopyroxene in the peridotites display very limited Mg isotopic variations (-0.28 to -0.17 ‰ for olivine; -0.24 to -0.14 ‰ for orthopyroxene). The clinopyroxenes have Mg isotopic compositions (-0.37 to -0.10 ‰) slightly lighter than coexisting olivine and orthopyroxene. Calculated whole rock $\delta^{26}\text{Mg}$ values of the peridotites vary from -0.27 to -0.16 ‰, falling within the normal mantle range (-0.25 $\pm$ 0.07 ‰). By contrast, the $\delta^{26}\text{Mg}$ values of continental intraplate lavas vary from -0.47 to -0.06 ‰ and negatively correlate with (Gd/Yb)_N ratios. The difference in Mg isotopic systematics between peridotites and intraplate basalts reveals that the former cannot be the sole magma source to the latter, and a lithology with a lighter Mg isotopic composition such as carbonated eclogite must also have contributed. Thus, the intraplate lavas with HIMU-like source signatures likely represent the product of mixing of melts derived from peridotite and recycled carbonated eclogite. Magnesium isotopes may therefore be a useful tracer of mantle sources to intraplate volcanism.

1. INTRODUCTION

Intraplate volcanism provides invaluable information on the chemical and physical evolution of the mantle (*e.g.*, Farmer, 2003; Dasgupta et al., 2010; Lee and Grand, 2012; McGee et al., 2015). Small-volume alkaline volcanoes that puncture Earth's continents have garnered significant interest due to their origin from a heterogeneous mantle that may have been modified directly or indirectly by melt extraction and a subducted component (Farmer, 2003; Pilet et al., 2008; Timm et al., 2010; Rooney et al., 2014).

High time-integrated $^{238}\text{U}/^{204}\text{Pb}$ signatures (HIMU) in oceanic intraplate basalts have been typically attributed to the incorporation of recycled ancient oceanic crust (*e.g.*, Stracke et al., 2003; Nebel et al., 2013; Castillo, 2015). Intraplate volcanoes erupted through the southwest Pacific continent of New Zealand, known as Zealandia, have HIMU-like isotopic affinities (*e.g.*, $^{206}\text{Pb}/^{204}\text{Pb}$ up to 20.67; Hoernle et al., 2006; Panter et al., 2006; Sprung et al., 2007; Timm et al., 2009, 2010; McCoy-West et al., 2010). The HIMU-like isotopic composition to these continental intraplate volcanoes has been ascribed to mantle melting in the presence of altered oceanic crust in the form of carbonated eclogite in the asthenosphere (Hoernle et al., 2006; Timm et al., 2009, 2010) or melting of amphibole veins in the lithosphere (Panter et al., 2006; Sprung et al., 2007). However, recent investigation of anhydrous Zealandia lithospheric mantle peridotites shows them to also have HIMU-like isotopic signatures (*e.g.*, $^{206}\text{Pb}/^{204}\text{Pb}$ up to 21.41; Scott et al., 2014a, b). Therefore, the source

region and melted components to these continental intraplate basalts with HIMU-like isotopic signatures remain unclear, and this is a debate that is repeated around the world (*e.g.*, Farmer, 2003; Pilet et al., 2008; Xu et al., 2012).

Magnesium (Mg), with three isotopes of ^{24}Mg , ^{25}Mg and ^{26}Mg , is a major element in the silicate Earth and can potentially offer unique insight into the sources of intraplate magma. Peridotitic mantle has a homogenous Mg isotopic composition ($\delta^{26}\text{Mg} = -0.25 \pm 0.07 \text{ ‰}$) because only very limited Mg isotope fractionation ($< 0.07 \text{ ‰}$) occurs during high-temperature igneous differentiation (Teng et al., 2007, 2010a; Handler et al., 2009; Yang et al., 2009; Bourdon et al., 2010; Huang et al., 2011; Pogge von Strandmann et al., 2011; Lai et al., 2015). By contrast, Mg isotopes strongly fractionate (up to 7 ‰) during Earth's surficial processes such as chemical weathering and seafloor alteration (Tipper et al., 2006; Pogge von Strandmann et al., 2008; Li et al., 2010; Teng et al., 2010b; Huang et al., 2012; Opfergelt et al., 2012; Liu et al., 2014). As a result, subduction should deliver altered oceanic crust containing isotopically distinct Mg to the mantle. Of the lithologies formed by metamorphosing subducted oceanic crust, carbonated eclogite has unusually low $\delta^{26}\text{Mg}$ values compared to normal peridotitic mantle, making it a prominent form of Mg isotopic heterogeneity in the upper mantle (Wang et al., 2014a). Thus, Mg isotopes may be a diagnostic tool for evaluating any contribution of carbonated eclogites to the mantle source of intraplate alkaline magmatism.

Here, we present Mg isotope data for a set of well-characterized samples from New Zealand, including 11 spinel-facies peridotite xenoliths and 21 intraplate basaltic lavas, with the goal of examining the role of carbonated eclogite in continental intraplate magmatism that has a HIMU-like source signature. Our results show that the Mg isotopic compositions of olivine, orthopyroxene, and clinopyroxene minerals in the metasomatised peridotites display limited variations, with calculated whole rock $\delta^{26}\text{Mg}$ values (-0.27 to -0.16 ‰) falling in the normal mantle range. By contrast, $\delta^{26}\text{Mg}$ values of the Antipodes Volcano lavas vary significantly from -0.47 to -0.06 ‰, and negatively correlate with the (Gd/Yb)_N ratios, which reflects a mixture of both eclogite- and peridotite-derived melts.

2. GEOLOGIC BACKGROUND

Zealandia is a largely undersea continent formed around New Zealand in the southwest Pacific (Fig. 1). The peridotitic xenoliths entrained in the alkaline volcanoes are representative of the lithospheric mantle of Zealandia. These peridotites, including dunite, harzburgite, lherzolite, and wehrlite, were equilibrated within the spinel stability field with temperatures varying from 850 to 1000 °C (Scott et al., 2014a,b; McCoy-West et al., 2015). Geochemical studies of the peridotite xenoliths found that the lithospheric mantle of Zealandia have experienced varying degree of ancient to recent melt depletion followed by metasomatism (McCoy-West et al., 2013, 2015; Scott et al., 2014a,b; Liu et al., 2015). The very high spinel Cr# [>60 ; Cr# = $100 \times \text{Cr}/(\text{Cr} + \text{Al})$] are indicative of near complete removal of clinopyroxene during

melting, whereas LREE-enrichment, high Th/U and low Ti/Eu in the secondary clinopyroxene grains indicate subsequent interaction of peridotite with carbonatitic or CO₂-rich melt (Scott et al., 2014a, b). The regionally extensive metasomatism has imparted a HIMU-like Sr, Nd and Pb isotopic signature on the Zealandia peridotitic mantle (Scott et al., 2014b). Our study utilizes peridotite xenoliths from Kattothyrst and Burke River locations in the South Island because (1) the large geographical distribution of the xenolith locations indicates the HIMU-like isotopic features are widespread (Fig. 1); and (2) they contain the strongest ²⁰⁶Pb/²⁰⁴Pb and ⁸⁷Sr/⁸⁶Sr HIMU-like isotope ratios within the Zealandia dataset (Fig. 2; Scott et al., 2014a, b)

Intraplate volcanoes have been active and widespread on Zealandia since Cretaceous times. The resulting volcanic rocks are predominantly alkaline with HIMU-like isotopic signatures and have OIB-like major and trace element compositions (*e.g.*, Baker et al., 1994; Finn et al., 2005; Hoernle et al., 2006; Panter et al., 2006; Sprung et al., 2007; Timm et al., 2009, 2010; McCoy-West et al., 2010; McGee et al., 2013, 2015; Scott et al., 2013). The Antipodes Volcano on the eastern edge of the Campbell Plateau is composed of alkaline lavas and pyroclastic rocks, including tuff cone and ring deposits (Fig. 1; Cullen, 1969; Gamble et al., 1986; Scott et al., 2013). The lavas were erupted < 0.5 Ma and have the uniformly highest ²⁰⁶Pb/²⁰⁴Pb (20.5 to 20.6) and lowest ⁸⁷Sr/⁸⁶Sr (~0.7029) ratios among all the intraplate volcanoes across Zealandia (Fig. 2; Panter et al., 2006; Timm et al., 2010). Chemical analysis indicates little or no crustal contamination and steep heavy REE

profiles $[(\text{Gd}/\text{Yb})_{\text{N}} = 2.65 \text{ to } 5.20]$; normalized to chondrite value] suggest a derivation from a source that contains residual amphibole or garnet (Panter et al., 2006; Timm et al., 2010; Scott et al., 2013).

3. ANALYTICAL METHODS

Peridotite chips were crushed to $\sim 0.3 \text{ mm}$, and 1 to 3 mg of mineral grains (olivine, clinopyroxene and orthopyroxene) were handpicked under a binocular microscope. About 1 to 5 mg mineral grains and rock powders of the Antipodes Volcano basalts were digested using a mixture acid of Optima-grade HF-HNO₃-HCl. After complete dissolution, dried residues were taken up in 1N HNO₃ for ion column chemistry. Chemical separation and purification of Mg were achieved by cation exchange chromatography with Bio-Rad AG50W-X8 resin at University of Washington using established methods (Teng et al. 2007; 2010a; 2015a; Yang et al., 2009; Li et al., 2010). The same column procedure was performed twice in order to purify the Mg. The pure Mg solutions were then dried down, and re-dissolved in 3% HNO₃ in preparation for mass spectrometry.

The Mg isotopic compositions were analyzed by the sample-standard bracketing method using a *Nu* Plasma MC-ICPMS at low resolution mode (Teng and Yang, 2014). Each batch of sample analysis contains at least one well-characterized standard. Ratio measurements of sample solutions were analyzed for >4 times within a session.

The long-term precision based on the repeated analyses of olivine and seawater samples is better than ± 0.07 ‰ (2SD) for the $^{26}\text{Mg}/^{24}\text{Mg}$ ratio (Teng et al., 2015a, b). Magnesium isotopic results are reported in the conventional δ notation in per mil relative to DSM-3, $\delta^{26}\text{Mg} = [(^{26}\text{Mg}/^{24}\text{Mg})_{\text{sample}}/(^{26}\text{Mg}/^{24}\text{Mg})_{\text{DSM-3}} - 1] \times 1000$. Two in-house standards, San Carlos olivine and seawater, yield average $\delta^{26}\text{Mg}$ values of -0.27 ± 0.08 ‰ (2SD; $n = 5$) and -0.81 ± 0.05 ‰ (2SD; $n = 6$), respectively, which are in agreement with previously reported values (*e.g.*, Foster et al., 2010; Li et al., 2010; Teng et al., 2010a; Ling et al., 2011)

4. RESULTS

Olivine separates from the 11 peridotite xenoliths have similar Mg isotopic compositions, with $\delta^{26}\text{Mg}$ ranging from -0.28 to -0.17 ‰ (Fig. 3). Orthopyroxene and clinopyroxene display $\delta^{26}\text{Mg}$ variations of -0.24 to -0.14 ‰ and -0.37 to -0.10 ‰ respectively (Fig. 3). The bulk rock $\delta^{26}\text{Mg}$ values calculated from the mineral data range from -0.27 to -0.19 ‰. By contrast, Mg isotopic compositions of the 21 intraplate lavas span in a relatively wider range, with $\delta^{26}\text{Mg}$ varying from -0.47 to -0.06 ‰, most of which being isotopically lighter than those of the peridotite xenoliths (Fig. 3).

5. DISCUSSION

In this section, the Mg isotopic systematics of the peridotites samples is firstly

evaluated to examine the overprint of carbonatitic metasomatism. Then, the role of recycled carbonated eclogite in the source of continental intraplate volcanism is discussed based on the Mg isotopic systematics of Antipodes Volcano lavas.

5.1. Magnesium isotopic composition of the Zealandia peridotite

The lithospheric mantle of Zealandia has a complex history involving ancient melt depletion and subsequent carbonatitic metasomatism (McCoy-West et al., 2013, 2015; Scott et al., 2014a,b; Liu et al., 2015). Melt extraction should not have influenced the Mg isotopic systematics of peridotites because partial melting causes very limited Mg isotope fractionation between melts and residues (≤ 0.07 ‰; Teng et al., 2010a). Carbonatitic metasomatism potentially imparts a low- $\delta^{26}\text{Mg}$ signature to peridotites (Yang et al., 2012; Xiao et al., 2013; Huang et al., 2015; Hu et al., 2015) since recycled sedimentary carbonates and marbles have extremely light Mg isotopic compositions compared with the normal peridotitic mantle (*e.g.*, Saenger and Wang, 2014; Wang et al., 2014a). The overprint of carbonatitic metasomatism is evident in newly formed peridotite minerals, with clinopyroxene likely having formed by the reaction of carbonatite + orthopyroxene $\rightarrow$ clinopyroxene + olivine + CO_2 in the present peridotites (Brey and Green, 1975; Wallace and Green, 1988). These clinopyroxenes have high LREE contents and variable but low Ti/Eu ratios, which is consistent with the fingerprint left behind by carbonatitic metasomatism (Scott et al., 2014a, b).

The Mg isotopic systematics of the peridotite minerals also supports carbonatitic metasomatism as many clinopyroxenes are isotopically lighter than the coexisting olivine and orthopyroxene (Fig. 4a). This trend is opposite to the one observed in equilibrated natural samples and that predicted from first principle calculations, which suggest that the enrichment of ^{26}Mg in peridotitic minerals follows a sequence of clinopyroxene > orthopyroxene > olivine (Handler et al., 2009; Yang et al., 2009; Huang et al., 2011, 2013; Liu et al., 2011; Schauble, 2011; Xiao et al., 2013; Lai et al., 2015). In addition, the inter-mineral isotope fractionation between clinopyroxene and olivine ($\Delta^{26}\text{Mg}_{\text{Cpx-Ol}}$) is positively correlated with $\delta^{26}\text{Mg}_{\text{Cpx}}$ (Fig. 4b), further suggesting that the variation of $\Delta^{26}\text{Mg}_{\text{Cpx-Ol}}$ is caused by the decrease of clinopyroxene $\delta^{26}\text{Mg}$ value. Nonetheless, because of the large Mg budget of peridotitic minerals, the degree of the influence of carbonatitic metasomatism on Mg isotopic systematics of peridotite minerals is small as evidenced by the limited Mg isotopic variation in clinopyroxenes ($\delta^{26}\text{Mg} = -0.37$ to -0.10 ‰).

The whole-rock $\delta^{26}\text{Mg}$ values for the peridotites, calculated based on mineral compositions, range from -0.27 to -0.16 ‰ with an average value of -0.22 ± 0.03 ‰ (Fig. 3). This value agrees very well with global peridotite xenoliths ($\delta^{26}\text{Mg} = -0.25 \pm 0.04$ ‰; Teng et al., 2010a). Studies have shown that no systematic $\delta^{26}\text{Mg}$ difference exists between spinel- and garnet-facies peridotites (Teng et al., 2010a; Pogge von Strandmann et al., 2011). Therefore, although no garnet-facies peridotites have been

found across Zealandia, they would be expected to have a similar bulk Mg isotopic composition to the spinel-facies peridotites.

5.2. Origin of magnesium isotopic variations in Antipodes Volcano lavas

Magnesium isotopic compositions of the 21 Antipodes Volcano basalts ($\delta^{26}\text{Mg} = -0.47$ to -0.06 ‰) extend well beyond the range of the Zealandia peridotites and typical oceanic basalts (Fig. 3). The little (<1 %) or no crustal contamination of the lavas (Panter et al., 2006; Timm et al., 2010; Scott et al., 2013), suggests that the input of crustal materials, if any, cannot have caused the isotopic variation. Fractional crystallization of olivine and clinopyroxene during magmatic differentiation causes Mg isotope fractionation of no more than ± 0.07 ‰ (Teng et al., 2007, 2010a) and thus cannot account for the Mg isotopic variation either. Therefore, the Mg isotopic compositions of basalts must reflect processes that occurred in the mantle source region. Below, we examine the possible sources for the Antipodes Volcano lavas and discuss the mechanism of generating the Mg isotopic variations.

The high $(\text{Na}_2\text{O} + \text{K}_2\text{O})/\text{TiO}_2$ ratio (up to 7.13) of some Antipodes Volcano lavas point toward the source containing variably carbonated peridotites (e.g., Dasgupta et al., 2007), despite that a large portion of lavas equally has low $(\text{Na}_2\text{O} + \text{K}_2\text{O})/\text{TiO}_2 < 1$ (Scott et al., 2013). Though having been extensively metasomatised by carbonatite or CO_2 -rich silicate melts (e.g., Scott et al., 2014a, b), Zealandia peridotites maintain a mantle-like $\delta^{26}\text{Mg}$ value (-0.27 to -0.16 ‰), higher than those of most Antipodes

Volcano lavas (Fig. 3). Therefore, solely partial melting of carbonated peridotites may not fully explain the Mg isotopic systematics of Antipodes Volcano lavas. It has been proposed that incorporation of hydrous minerals (phlogopite and/or amphibole) that occurred as veins during mantle melting may account for some specific geochemical features found in intraplate continental basalts in New Zealand (*e.g.*, negative K anomalies; Panter et al., 2006). Phlogopites of metasomatic origin have been reported to have $\delta^{26}\text{Mg}$ values (-0.11 to -0.05 ‰) that are slightly higher than the normal mantle value (Liu et al., 2011). Although no Mg isotope data has been reported for amphiboles from metasomatised lithospheric mantle, amphiboles from crustal rocks do have mantle-like $\delta^{26}\text{Mg}$ values (Liu et al., 2010) and the similar Mg coordination geometry (CN = 6) between amphibole and phlogopite further predicts the absence of inter-mineral isotope fractionation. Thus, partial melting of hydrous minerals (*e.g.*, hornblende) alone or their participation in peridotitic melting cannot contribute to the light Mg isotopic compositions of the basalts.

Recycled sediments have highly heterogeneous $\delta^{26}\text{Mg}$ values compared to the mantle, with carbonated meta-sediments having light Mg isotopic compositions (*e.g.*, Wang et al., 2015a). The incorporation of recycled sediments or sediment-derived melts in peridotitic melting could generate basaltic melts with light Mg isotopic compositions but a large amount is required because of the low-MgO nature. However, the low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (~ 0.7029) of the Antipodes Volcano basalts are inconsistent with a significant contribution from sediments or their derivatives (Panter et al., 2006;

Timm et al., 2010). In addition, an increasing contribution of recycled sediments would dampen any HIMU signatures, whereas the Antipodes Volcano lavas have the highest $^{206}\text{Pb}/^{204}\text{Pb}$ ratio (20.5 to 20.6) across the Zealandia intraplate lavas (Scott et al., 2013).

Eclogitic components in mantle typically have very heterogeneous Mg isotopic compositions (Wang et al., 2012, 2015b). For example, cratonic eclogite xenoliths from South and West Africa have $\delta^{26}\text{Mg}$ values ranging from -1.38 to +0.05 ‰, reflecting the Mg isotopic heterogeneity of their protoliths, altered oceanic crusts (Wang et al., 2012, 2015b). Exhumed carbonated eclogites sampled in Dabie-Sulu orogenic belt also have unusually light Mg isotopic compositions ($\delta^{26}\text{Mg} = -1.92$ to -0.65 ‰), which are resulted from isotopic exchange between eclogites and surrounding carbonates during subduction (Wang et al., 2014a). Therefore, partial melting of cratonic eclogites or recycled carbonated eclogites in the mantle may have the potential to generate melts with variable and low $\delta^{26}\text{Mg}$ values. The radiogenic isotopic systematics of Antipodes Volcano basalts, that is high $^{206}\text{Pb}/^{204}\text{Pb}$ and low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (Panter et al., 2006; Timm et al., 2010), are also in accordance with eclogitic melting. However, it was suggested that melting of eclogite boudins dispersed within the mantle alone may not be able to produce the voluminous intraplate volcanoes widespread on Zealandia, especially those large composite shield volcanoes (*e.g.*, Hoernle et al., 2006). The low $(\text{Gd}/\text{Yb})_{\text{N}}$ (down to 2.65), low $\text{CaO}/\text{Al}_2\text{O}_3$ (down to 0.26) and high $(\text{Na}_2\text{O} + \text{K}_2\text{O})/\text{TiO}_2$ ratios (up to 7.13) in some

Antipodes Volcano basalts support the predominance of peridotite as their source region (Dasgupta et al., 2007; Scott et al., 2013), which is also true for other intraplate basalts on Zealandia (*e.g.*, McGee et al., 2013).

Based on above discussions, both peridotite and eclogite were involved in the source region of Antipodes Volcano basalts. We thus proposed a hypothesis that the Antipodes Volcano basalts represent a mixture of melts derived from different mantle lithologies (*i.e.*, peridotite and eclogite). The negative correlation between $\delta^{26}\text{Mg}$ value and $(\text{Gd}/\text{Yb})_{\text{N}}$ supports our hypothesis (Fig. 5). Partial melting of either peridotites or eclogites alone cannot generate such negative correlation. Melts derived from the carbonated eclogites are characterized by the lowest $\delta^{26}\text{Mg}$ values and highest $(\text{Gd}/\text{Yb})_{\text{N}}$ ratios, while melts derived from low-degree melting of garnet-facies peridotites would have mantle-like Mg isotopic compositions and relatively lower $(\text{Gd}/\text{Yb})_{\text{N}}$ ratios. Melts from spinel-facies peridotites would also yield mantle-like $\delta^{26}\text{Mg}$ values, but the lowest $(\text{Gd}/\text{Yb})_{\text{N}}$ ratios because of the absence of residual garnet during partial melting. Sequential mixing between these melts would generate the general negative correlation between $\delta^{26}\text{Mg}$ and $(\text{Gd}/\text{Yb})_{\text{N}}$ (Fig. 5). The one exception to this trend (110212-1; Fig. 5) has high $(\text{Gd}/\text{Yb})_{\text{N}}$ ratio but mantle-like $\delta^{26}\text{Mg}$ value, perhaps caused by partial melting of either a garnet-facies peridotite containing a relatively high abundance of garnet (*e.g.*, > 10 %) or a normal eclogite with mantle-like $\delta^{26}\text{Mg}$ value. Nevertheless, the overall co-variation between $\delta^{26}\text{Mg}$ and $(\text{Gd}/\text{Yb})_{\text{N}}$ supports a lithologically-mixed source for most, if not all, Antipodes

Volcano basalts.

It is difficult to determine whether the carbonated eclogite + peridotite magma source comes from the asthenosphere, the lithosphere, or both. Eclogitic boudins extracted from peridotitic lithospheric mantle (*i.e.*, cratonic eclogite) have light Mg isotopic compositions (Wang et al., 2012, 2015b), and so corresponding melts are the satisfactory candidate for the low- $\delta^{26}\text{Mg}$, high-(Gd/Yb)_N end-member. Eclogitic pieces in the asthenospheric mantle may equally have light Mg isotopic compositions. Continuous pollution of the uppermost asthenosphere with recycled carbonated eclogites derived from subduction could have added light- $\delta^{26}\text{Mg}$ components to the Zealandia continental lithospheric mantle. Melting of the mantle would be initiated in these eclogites/pyroxenites that have a lower solidus temperature than peridotites (Hirschmann and Stolper, 1996), producing melts with low $\delta^{26}\text{Mg}$ and high (Gd/Yb)_N signatures. As melting progresses, dilution of these melts by those from the garnet- and spinel-facies peridotites would eventually delineate the trend by increasing $\delta^{26}\text{Mg}$ and decreasing (Gd/Yb)_N of the hybridized melts.

6. CONCLUSIONS

We present high-precision Mg isotopic data for peridotite xenoliths and HIMU-like continental intraplate basalts from New Zealand. The following conclusions are drawn:

(1) Peridotitic minerals display a limited variation of Mg isotopic compositions,

with $\delta^{26}\text{Mg}$ values varying from -0.28 to -0.17 ‰ in olivine, from -0.24 to -0.14 ‰ in orthopyroxene and from -0.37 to -0.1 ‰ in clinopyroxene.

(2) The clinopyroxenes are slightly lighter in Mg isotopes than coexisting olivine and orthopyroxene, consistent with their formation via carbonatitic metasomatism. The calculated whole rock $\delta^{26}\text{Mg}$ values for these metasomatised peridotites range from -0.27 to -0.19 ‰, with the average of -0.22 ± 0.03 ‰ (2SD), that is indistinguishable from the normal mantle value (-0.25 ± 0.07 ‰).

(3) The $\delta^{26}\text{Mg}$ values of HIMU-like Antipodes Volcano basalts are generally lower and more variable (-0.47 to -0.06 ‰) compared to the peridotite xenoliths, suggesting that the latter cannot be the sole mantle source for the basalts.

(4) The negative correlation between $\delta^{26}\text{Mg}$ value and $(\text{Gd}/\text{Yb})_{\text{N}}$ ratio indicates that the HIMU-like intraplate basalts represent the product of mixing of melts derived from peridotite and recycled carbonated eclogites.

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Figure captions:

Fig 1. (A) Schematic map of the Zealandia continent showing portions of continental crust below (light gray) and above (dark gray) sea level. The studied peridotite xenoliths and intraplate lavas, as represented by red stars, are from South Island and Antipodes Islands, respectively. (B) Geological map of the Antipodes Islands showing the rock types and sample localities (after Scott et al., 2013).

Fig 2. $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ diagram for Kattothyrst peridotites and Antipodes Volcano lavas. Data are from Panter et al., (2006), Timm et al., (2010), and Scott et al., (2014a). The fields of Zealandia peridotite xenoliths, Zealandia intraplate lavas, FOZO (focus zone) and HIMU are modified after Scott et al., (2014b).

Fig 3. Magnesium isotopic data for the peridotitic minerals and Antipodes Volcano lavas. The gray bar is the average whole-rock $\delta^{26}\text{Mg}$ value ($-0.22 \pm 0.03 \text{ ‰}$) of peridotite calculated from mineral compositions. The Mg isotopic compositions are from Table 1 and Table 2. Mineral chemical compositions and modal abundances are from Scott et al., (2014a).

Fig 4. Variations of olivine $\delta^{26}\text{Mg}$ value ($\delta^{26}\text{Mg}_{\text{Ol}}$) and inter-mineral fractionation between olivine and clinopyroxene ($\Delta^{26}\text{Mg}_{\text{Cpx-Ol}}$) as a function of clinopyroxene $\delta^{26}\text{Mg}$ value ($\delta^{26}\text{Mg}_{\text{Cpx}}$). The theoretically calculated equilibrium fractionation line at

temperature of 1000 °C is shown as dashed line. The solid line represents no inter-mineral fractionation between olivine and clinopyroxene. Data are from Table 1.

Fig 5. Variation of $\delta^{26}\text{Mg}$ vs $(\text{Gd/Yb})_{\text{N}}$ for the Antipodes Volcano lavas. The gray area represents the Mg isotopic composition of Zealandia peridotites. Model assumes 5-10% non-modal batch melting of spinel-facies peridotite with mode and melt mode of $\text{Ol}_{0.55} + \text{Opx}_{0.26} + \text{Cpx}_{0.15} + \text{Sp}_{0.04}$ and $\text{Ol}_{0.10} + \text{Opx}_{0.27} + \text{Cpx}_{0.50} + \text{Sp}_{0.13}$, respectively (Thirlwall et al., 1994; McGee et al., 2013); 5-10 % non-modal batch melting of garnet-facies peridotite with mode and melt mode of $\text{Ol}_{0.54} + \text{Opx}_{0.23} + \text{Cpx}_{0.17} + \text{Grt}_{0.06}$ and $\text{Ol}_{0.03} + \text{Opx}_{0.16} + \text{Cpx}_{0.88} + \text{Grt}_{0.09}$, respectively (Walter, 1998; McGee et al., 2013); 5-10% modal batch melting of eclogites with mode of $\text{Cpx}_{0.65} + \text{Grt}_{0.35}$ (Williams et al., 2009). The Mg isotopic compositions of clinopyroxene and garnet are -0.41 ‰ and -1.08 ‰, respectively (Wang et al., 2012). The Mg isotopic compositions of olivine, clinopyroxene, and orthopyroxene in spinel- and garnet-facies peridotites are taken from samples Kat-3 and Kat-14, which have the highest and lowest whole-rock $\delta^{26}\text{Mg}$, respectively. The $\delta^{26}\text{Mg}$ values of spinel and garnet are assumed +0.13 ‰ and -0.90 ‰, respectively, based on the first principle calculations (Schauble, 2011; Huang et al., 2013). The spinel- and garnet-facies peridotites are assumed to have a primary mantle-like $(\text{Gd/Yb})_{\text{N}}$ ratio (Sun and McDonough, 1989), and the eclogite is assumed to have a N-MORB-like $(\text{Gd/Yb})_{\text{N}}$ ratio (Sun and McDonough, 1989).

615 **Table 1:** Magnesium isotopic compositions of the minerals from peridotitic xenoliths

Sample	Mineral	$\delta^{26}\text{Mg}$	2SD	$\delta^{25}\text{Mg}$	2SD
Kat-2	Ol	-0.20	0.05	-0.07	0.05
	Cpx	-0.22	0.06	-0.07	0.04
	WR	-0.20	0.08	-0.07	0.06
Kat-3	Ol	-0.28	0.04	-0.12	0.05
	Opx	-0.22	0.08	-0.11	0.05
	Cpx	-0.18	0.05	-0.07	0.04
	WR	-0.27	0.11	-0.12	0.08
Kat-6	Ol	-0.24	0.04	-0.11	0.05
	Opx	-0.18	0.08	-0.11	0.05
	Cpx	-0.20	0.05	-0.08	0.04
	WR	-0.23	0.11	-0.11	0.08
Kat-8	Ol	-0.24	0.04	-0.12	0.05
	Opx	-0.14	0.08	-0.07	0.05
	Cpx	-0.21	0.05	-0.11	0.04
	WR	-0.23	0.11	-0.11	0.08
Kat-9	Ol	-0.19	0.04	-0.09	0.05
	Opx	-0.15	0.08	-0.09	0.05
	Cpx	-0.30	0.08	-0.06	0.07
	WR	-0.19	0.13	-0.09	0.10
Kat-13	Ol	-0.22	0.04	-0.12	0.05
	Opx	-0.18	0.08	-0.07	0.05
	Cpx	-0.24	0.05	-0.09	0.04
	WR	-0.22	0.11	-0.12	0.08
Kat-14	Ol	-0.17	0.05	-0.07	0.04
	Opx	-0.16	0.08	-0.09	0.05
	Cpx	-0.10	0.07	-0.06	0.05
	WR	-0.16	0.12	-0.07	0.08
Kat-15	Ol	-0.22	0.04	-0.10	0.03
	Opx	-0.19	0.08	-0.10	0.05
	Cpx	-0.28	0.05	-0.13	0.04
	WR	-0.22	0.11	-0.10	0.07
BUR-3	Ol	-0.27	0.05	-0.15	0.04
	Opx	-0.23	0.08	-0.13	0.05
	Cpx	-0.37	0.08	-0.15	0.07
	WR	-0.27	0.13	-0.15	0.10
BUR-14	Ol	-0.21	0.05	-0.11	0.04
	Opx	-0.24	0.08	-0.11	0.05
	Cpx	-0.31	0.05	-0.17	0.04
	WR	-0.21	0.11	-0.11	0.08
BUR-17	Ol	-0.25	0.07	-0.13	0.06
	Opx	-0.21	0.08	-0.11	0.05

	Cpx	-0.32	0.08	-0.14	0.07
	WR	-0.25	0.14	-0.13	0.11

616

617 Note:

618 Kat = Kattothyrst; BUR = Burke River

619 Ol = olivine; Opx = orthopyroxene; Cpx = clinopyroxene; WR = whole rock

620 The whole rock compositions are calculated based on the mineral compositions and
621 modal abundance that are taken from Scott et al., (2014a)

622 2SD equals two times the standard deviation of the population of n (n>20) repeat

623 measurements of the standards during an analytical session.

Table 2: Magnesium isotopic compositions of Antipodes Volcano lavas

Sample	$\delta^{26}\text{Mg}$	2SD	$\delta^{25}\text{Mg}$	2SD	(Gd/Yb) _N
110124-5	-0.28	0.05	-0.13	0.05	3.88
110126-2	-0.37	0.07	-0.18	0.06	3.98
110211-6	-0.24	0.05	-0.11	0.05	3.39
110201-4	-0.34	0.05	-0.17	0.05	3.36
110124-12	-0.35	0.05	-0.18	0.04	4.20
110125-9	-0.38	0.05	-0.22	0.05	4.29
110122-11	-0.27	0.05	-0.11	0.04	3.60
110128-11	-0.40	0.05	-0.20	0.04	3.84
110203-4A	-0.30	0.05	-0.16	0.05	3.68
110128-18	-0.33	0.07	-0.18	0.06	3.89
110204-1	-0.13	0.07	-0.07	0.06	3.19
110203-3	-0.37	0.07	-0.19	0.06	3.82
110123-10A	-0.34	0.07	-0.18	0.06	3.46
110204-4	-0.18	0.07	-0.11	0.06	2.77
110125-10	-0.29	0.07	-0.13	0.06	3.57
110125-6	-0.47	0.11	-0.20	0.09	4.28
110203-8	-0.21	0.07	-0.09	0.06	2.99
110212-1	-0.31	0.07	-0.15	0.06	5.20
A8	-0.06	0.08	-0.04	0.07	2.65
A105	-0.35	0.08	-0.19	0.07	3.75
A112	-0.31	0.08	-0.17	0.05	3.66

Note:
2SD equals two times the standard deviation of the population of n (n>20) repeat measurements of the standards during an analytical session.
(Gd/Yb)_N ratios are from Scott et al. (2013)

Figure 1 Wang et al.

Figure 2 Wang et al.

Figure 3 Wang et al.

Figure 4 Wang et al.

Figure 5 Wang et al.