^{186}Os - ^{187}Os and highly siderophile element abundance systematics of the mantle revealed by abyssal peridotites and Os-rich alloys

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Abstract

Abyssal peridotites are oceanic mantle fragments that were recently processed through ridges and represent residues of both modern and ancient melting. To constrain the nature and timing of melt depletion processes, and the composition of the mantle, we report high-precision Os isotope data for abyssal peridotites from three ocean basins, as well as for Os-rich alloys, primarily from Mesozoic ophiolites. These data are complemented by whole-rock highly siderophile element (HSE: Os, Ir, Ru, Pt, Pd, Re), trace- and major-element abundances for the abyssal peridotites, which are from the Southwest Indian (SWIR), Central Indian (CIR), Mid-Atlantic (MAR) and Gakkel Ridges. The results reveal a limited role for melt refertilization or secondary alteration processes in modifying abyssal peridotite HSE compositions. The abyssal peridotites examined have experienced variable melt depletion (2% to >16%), which occurred >0.5 Ga ago for some samples. Abyssal peridotites typically exhibit low Pd/Ir and, combined with high-degrees of estimated total melt extraction, imply that they were relatively refractory residues prior to incorporation into their present ridge setting. Recent partial melting processes and mid-ocean ridge basalt (MORB) generation therefore played a limited role in the chemical evolution of their precursor mantle domains. The results confirm that many abyssal peridotites are not simple residues of recent MORB source melting, having a more complex and long-lived depletion history.

Peridotites from the Gakkel Ridge, SWIR, CIR and MAR indicate that the depleted MORB mantle has $^{186}\text{Os}/^{188}\text{Os}$ of 0.1198356 ± 21 (2SD). The Phanerozoic Os-rich alloys yield an average $^{186}\text{Os}/^{188}\text{Os}$ within uncertainty of abyssal peridotites (0.1198361 ± 20). Melt depletion trends defined between Os isotopes and melt extraction indices (e.g., Al_2O_3) allow an estimate of the primitive mantle (PM) composition, using only abyssal peridotites. This yields $^{187}\text{Os}/^{188}\text{Os}$ (0.1292 ± 25), and $^{186}\text{Os}/^{188}\text{Os}$ of 0.1198388 ± 29 , both of which are within uncertainty of previous primitive mantle estimates. The $^{186}\text{Os}/^{188}\text{Os}$ composition of the PM is less radiogenic than for some plume-related lavas, with the latter requiring sources with high long-term time-integrated Pt/Os. Estimates of primitive mantle HSE concentrations using abyssal peridotites define chondritic Pd/Ir, which differs from previous supra-chondritic estimates for Pd/Ir based on peridotites from a range of tectonic settings. By contrast, estimates of PM yield non-chondritic

Ru/Ir. The cause of enhanced Ru in the mantle remains enigmatic, but may reflect variable partitioning behaviour of Ru at high pressure and temperature.

1. Introduction

The highly siderophile element (HSE: Re, Os, Ir, Ru, Pt, Rh, Pd, Au) composition of the mantle is a key constraint for understanding terrestrial accretion and differentiation. Core formation is predicted to have efficiently stripped the HSE from the mantle due to the high metal-silicate partition coefficients ($>10^4$) typical of these elements at relatively low pressures and temperatures (e.g., [Brenan et al., 2016](#)). This prediction is not matched by mantle peridotite compositions that commonly have HSE abundances only ~150 times less abundant than in chondrite meteorites, and in chondritic relative proportions ([Morgan, 1986](#); [Morgan et al., 2001](#); [Becker et al., 2006](#); [Fischer-Gödde et al., 2011](#)). Competing models to explain this discrepancy include late accretion of chondrite-like impactors to Earth (e.g., [Kimura et al., 1974](#); [Chou, 1978](#)), lower metal-silicate partition coefficients at higher pressures and temperatures ([Murthy, 1991](#)), and outer core addition ([Snow & Schmidt, 1998](#)).

The mantle has been sampled from oceanic and continental settings, with a strong bias towards continental lithospheric mantle and massif peridotites. These peridotites all come from the upper mantle and we refer to estimates of the upper mantle based on the HSE as primitive mantle (PM). Some prior studies (e.g., [Meisel et al., 2001](#); [Becker et al., 2006](#); [Fischer-Gödde et al., 2011](#)) have also referred to this as primitive upper mantle (PUM). The PM composition in this instance is broadly analogous to the bulk silicate Earth (BSE) since the immense majority of the HSE reside in the mantle relative to the crust. To obtain estimates of the PM composition, studies have used peridotite suites to extrapolate to compositions by accounting for melt-depletion ([Morgan, 1986](#); [Meisel et al., 2001](#)). In the most recent iterations, [Becker et al. \(2006\)](#) and [Fischer-Gödde et al. \(2011\)](#) proposed a PM composition with Pd/Ir and Ru/Ir ratios significantly higher than in bulk chondrite meteorites. Similar ‘non-chondritic’ HSE patterns have also been observed in mantle peridotites from different settings ([Pattou et al., 1996](#); [Snow & Schmidt, 1998](#); [Rehkamper et al., 1999](#)), as well as in the mantle sources of some lavas ([Peters et al., 2016](#)). However, the reliability of using massif peridotites compositions has been called into question, due to refertilization processes that these samples may experience (e.g., [Marchesi et al., 2014](#); [Lorand & Luguet, 2016](#)).

The oceanic mantle is the largest of the accessible terrestrial reservoirs. Despite the volume and extent of this reservoir, it is poorly studied compared with continental mantle peridotites, largely due to accessibility and sampling issues. The most common samples from the oceanic mantle are abyssal peridotites, which are widely interpreted as the depleted mantle residues of mid-ocean ridge magmatic processes (see [Warren, 2016](#) for a review). Abyssal peridotites are typically sampled at oceanic transform faults or at ridge segments, where mantle has been tectonically exhumed. Unfortunately, abyssal peridotites are not pristine samples of the mantle, as most have been modified by secondary alteration processes (e.g., [Snow & Dick, 1995](#); [Snow & Reisberg, 1995](#); [Malvoisin, 2015](#)). Additionally, during upwelling and partial melting beneath the ridge, abyssal peridotites may be infiltrated by associated melts ([Niu, 2004](#); [Warren, 2016](#)). Nonetheless, abyssal peridotites can be used to help to unravel the composition of the PM, as they are not subject to complex melt refertilization from exotic melts and fluids, as may occur in the continental lithospheric mantle ([Becker & Dale, 2016](#)).

Here we report new high-precision $^{186}\text{Os}/^{188}\text{Os}$ and $^{187}\text{Os}/^{188}\text{Os}$ data for abyssal peridotites from three ocean basins (Indian, Atlantic, Arctic), as well as Mesozoic to Proterozoic-aged Os-rich alloys from ophiolites. We compare these data with $^{186}\text{Os}/^{188}\text{Os}$ data obtained for the mantle from prior studies ([Walker et al., 1997](#); [2005](#); [Brandon et al., 2000](#); [2006](#)), using identical normalizations and reporting uncertainties as 2 standard deviations (2SD), unless otherwise stated. For convenience, we report uncertainties given on the last digits of the reported value (i.e., 0.1198388 ± 0.0000029 is abbreviated to 0.1198388 ± 29). These data are coupled with whole-rock HSE, trace-element and major-element data for abyssal peridotites, allowing a refined assessment of Earth's mantle composition from the view point of the oceanic mantle. In addition to providing extrapolations of HSE data to indices of melt depletion (e.g., [Morgan, 1986](#)), the Pt-Os and Re-Os isotope systems provide robust constraints on ratios of these elements in the mantle due to the long-lived ^{187}Re - ^{188}Os ($\lambda = 1.6668 \times 10^{-11} \text{ y}^{-1}$) and ^{190}Pt - ^{186}Os ($\lambda = 1.54 \times 10^{-12} \text{ y}^{-1}$) decay schemes ([Smoliar et al., 1996](#); [Walker et al., 1997](#); [Selby et al., 2007](#)). We use these results to investigate melt depletion processes in the mantle and to estimate the PM composition for the highly siderophile elements.

2. Samples and methods

2.1 Samples

Abyssal peridotites (**Figure 1**) were analysed from the ultraslow (<20 mm/yr full rate) spreading Oblique Segment at 9-16°E on the Southwest Indian Ridge (SWIR), the slow-spreading (46 mm/yr) central Indian Ridge (CIR) at the Marie Celeste transform fault, the ultraslow spreading (11-14 mm/yr) Gakkel Ridge from 5°W to 85°E, and from the Mid-Atlantic Ridge (MAR) south of the Kane transform fault (MARK area). Full details of sample locations and sample preparation protocols are in the *Supplementary Information*.

2.2 Major and trace element abundance analyses

Major element compositions were measured by X-ray fluorescence (XRF) at Franklin and Marshall College using a PW 2404 *PANalytical* XRF vacuum spectrometer following the procedures outlined in [Boyd & Mertzman \(1987\)](#). Major element analyses by XRF involved standard lithium tetraborate fusion techniques using 3.6:0.4 g LiBO₄:sample powder. Ferrous iron concentrations were determined by titration with potassium dichromate. Precision and accuracy are estimated using repeat analyses of standards, with long-term reproducibility (in wt.% and 2σ absolute standard deviation, $n = 13$) of ± 0.13 for SiO₂, ± 0.01 for TiO₂, ± 0.09 for Al₂O₃, ± 0.63 for FeO, ± 0.47 for Fe₂O₃, ± 0.10 for Fe₂O₃^T, ± 0.01 for MnO, ± 0.04 for MgO, ± 0.07 for CaO, ± 0.03 for Na₂O, ± 0.01 for K₂O, and $\pm <0.01$ for P₂O₅. Accuracy for the average of 13 runs of BHVO-2 relative to USGS values is better than 0.2% for SiO₂ and TiO₂, <1% for Al₂O₃, MgO, Fe₂O₃^T, CaO, Na₂O, P₂O₅, and <3% for K₂O (*Table S1*).

Trace-element abundances were determined at the *Scripps Isotope Geochemistry Laboratory (SIGL)*, Scripps Institution of Oceanography using methods described previously in [Day et al. \(2014\)](#). One hundred milligrams of powder was precisely weighed and digested in a 1:4 mixture of Teflon-distilled HNO₃:HF for >72 Hrs at 150°C on a hotplate. Rock standards (BHVO-2, BIR-1, BCR-2, AGV-2, GP13, DTS-2b) and total procedural blanks were prepared along with the samples. After drying down and sequential HNO₃ dry-down steps to break-down fluorides, clear sample solutions were diluted by a factor of 5000 in 2% HNO₃ and doped with a 1 ppb In solution to monitor instrumental drift. Solutions were measured using a *Thermo Scientific* iCAPq c quadrupole inductively coupled plasma mass spectrometer at the *SIGL* in

standard mode. Reproducibility of the reference materials was generally better than 5% (RSD) for basaltic and peridotite standards, and element abundances were generally within error of recommended values (*Table S1*).

2.3 *Highly siderophile element abundances and $^{187}\text{Os}/^{188}\text{Os}$*

Osmium isotope and HSE abundance analyses were performed at the University of Maryland, College Park (UMd) and at the *SIGL* using identical approaches. Precisely weighed homogenised powders were digested in sealed borosilicate Carius tubes, or in quartz high-pressure asher vessels, with isotopically enriched multi-element spikes (^{99}Ru , ^{106}Pd , ^{185}Re , ^{190}Os , ^{191}Ir , ^{194}Pt), and 12 mL of a 1:2 mixture of multiply Teflon distilled HCl and HNO_3 purged of excess Os by repeated treatment and reaction with H_2O_2 . Samples prepared in Carius tubes were digested to a maximum temperature of 270°C in an oven for 72 hours. Samples prepared in high-pressure asher vessels were subjected to a maximum temperature of 320°C for 6 hours, using an *Anton Paar* high-pressure asher. Osmium was triply extracted from the acid using CCl_4 and then back-extracted into HBr (Cohen & Waters, 1996), prior to purification by micro-distillation (Birck et al., 1997). Rhenium and the other HSE were recovered and purified from the residual solutions using standard anion exchange separation techniques (Becker et al., 2006; Day et al., 2016a).

Isotopic compositions of Os were measured in negative-ion mode on a *ThermoScientific Triton* thermal ionisation mass spectrometer at the University of Maryland and at the *SIGL*. Rhenium, Pd, Pt, Ru and Ir were measured using an Cetac Aridus II desolvating nebuliser coupled to a *ThermoFisher* Element 2 ICP-MS or to a *ThermoScientific* iCAPq c ICP-MS. Offline corrections for Os involved an oxide correction, an iterative fractionation correction using $^{192}\text{Os}/^{188}\text{Os} = 3.08271$, a ^{190}Os spike subtraction, and finally, an Os blank subtraction. Precision for $^{187}\text{Os}/^{188}\text{Os}$, determined by repeated measurement of the UMCP Johnson-Matthey standard was better than $\pm 0.2\%$ (2 SD; 0.11371 ± 21 ; $n = 22$). Rhenium, Ir, Pt, Pd and Ru isotopic ratios for sample solutions were corrected for mass fractionation using the deviation of the standard average run on the day over the natural ratio for the element. External reproducibility on HSE analyses was better than 0.5% for 0.5 ppb solutions and all reported values are blank corrected. The total procedural blanks ($n = 5$) run with the samples for both Carius tubes and

high pressure asher digestions had $^{187}\text{Os}/^{188}\text{Os} = 0.189 \pm 0.133$, with quantities (in picograms) of 2.4 $\pm$ 2.2 [Re], 13.5 $\pm$ 32.5 [Pd], 3.2 $\pm$ 5.8 [Pt], 3.6 $\pm$ 6.2 [Ru], 2.1 $\pm$ 3.9 [Ir] and 0.3 $\pm$ 0.3 [Os]. These blanks resulted in negligible corrections to samples (<1%). Two peridotite standards, HARZ-01 and GP13 were run at regular intervals during the analytical campaign and concentrations, as well as precision, have previously been reported in Day et al. (2012 and 2016a) and the data are provided in Table S2.

2.4 High-precision Os isotopic analyses

High-precision Os isotopic measurements were performed on large mass aliquots (23-53 g) of abyssal peridotite sample powders using a NiS fire assay procedure (Figure S1) and were prepared in laboratory ware that had never been used with isotopically enriched HSE ‘spikes’. While the NiS fire assay procedure enables variable recovery of Os, it allows for the pre-concentration of Os from large-mass samples. The disadvantage of this technique is that Re and Pt cannot be easily measured, so that Re/Os and Pt/Os ratios cannot be simultaneously obtained on the same aliquots.

The unspiked samples were placed in new alumina crucibles and fused at 1100°C in a muffle furnace for 120 minutes. Sodium tetraborate was used as flux, with a sample:flux:Ni:S ratio of 10:10:1:0.5. After quenching, the resultant NiS bead was extracted and dissolved in 4M HCl and filtered through cellulose paper. The cellulose and trapped Os-sulfides were then placed in a Carius tube and digested in 12 mL of a 1:2 mixture of multiply Teflon distilled HCl and HNO₃ purged of Os by treatment and reaction with H₂O₂. These samples were digested to a maximum temperature of 240°C in an oven for 72 hours. For Os-rich alloys measured in this study, grains were partially to fully digested in a 1:2 mixture of multiply Teflon distilled HCl and purged HNO₃ inside sealed Carius tubes (c.f., Walker et al., 2005; Brandon et al., 2006). Osmium was triply extracted from the acid using CCl₄ and then back-extracted into HBr (Cohen & Waters, 1996), prior to purification three times using standard micro-distillation procedures (Birck et al., 1997). Triple micro-distillation resulted in highly purified Os cuts compared with samples only micro-distilled once. Total procedural blanks for the integrated NiS fire assay-Carius tube-triple micro-distillation procedure were 0.74 $\pm$ 0.13 pg per gram of fused sample, with $^{187}\text{Os}/^{188}\text{Os} = 0.1551 \pm 62$ (2 SD) ($n = 3$). Total Os recovery was between 60 and >95%.

Some of the abyssal peridotite samples did not form homogeneous black glasses, as is typical after fusion of basaltic samples, but were green in colour and semi-crystalline, likely due to the degree of serpentinization and high Cr contents of some of the samples, presumably contributing to low yields, in some instances.

High-precision Os isotope data were obtained by negative thermal ionization mass spectrometry (N-TIMS) using a *Thermo Electron* Triton at the University of Maryland with a high purity O₂ bleed at a constant source pressure of 1×10^{-7} mbar. Data were collected in static mode using Faraday collectors with signal intensities of >100 mV on mass 234 ($^{186}\text{Os}^{16}\text{O}_3^-$), generated for >360 ratios (17s integration time per ratio, 30 s baselines measured every 20 ratios), with the goal of obtaining an in-run precision of ± 25 ppm (2 Sigma Error), or better, for $^{186}\text{Os}/^{188}\text{Os}$ on 75 ng Os loads, or larger. For some samples, with limited available Os (< 50 ng), signals intensities fell below 100 mV on mass 234 after ~180 ratios, resulting in more limited data collection.

A plethora of potential analytical issues associated with $^{186}\text{Os}/^{188}\text{Os}$ measurements by N-TIMS have been highlighted in recent work (e.g., [Luguet et al., 2008](#); [Chatterjee & Lassiter, 2015](#)). These issues include correction for the isotopic composition of oxygen, potential polyatomic mass interferences and residual interferences and correlations, many of which have been addressed for the technique employed here in [Ireland et al. \(2011\)](#). In this work, we used a $^{17}\text{O}/^{16}\text{O}$ composition of 0.0003749 and $^{18}\text{O}/^{16}\text{O}$ of 0.0020439, which is similar to the O isotopic values used by [Luguet et al. \(2008\)](#) and [Ireland et al. \(2011\)](#) to correct their data. Following O corrections, an instrumental mass fractionation correction was applied using the exponential law and a $^{192}\text{Os}/^{188}\text{Os}$ ratio of 3.083.

Potential interferences involving Pt, W and Re oxides were monitored during all runs with masses 230 ($^{198}\text{Pt}^{16}\text{O}_2^-$; *Figure S2*), 231 ($^{183}\text{W}^{16}\text{O}_3^-$) and 233 ($^{185}\text{Re}^{16}\text{O}_3^-$), using Faraday detectors in one routine. In another routine, after each baseline measurement, 230, 231 and 232 were all measured for 16 second integrations using the secondary electron multiplier, prior to switching the Faraday detectors to obtain static Os isotope compositions. In both cases, potential interferences had no measurable effects on Os isotopic compositions. We found that a major

improvement to measurements comes from triple micro-distillation of Os. Multiple distillation of Os from Os-rich alloys led to a significant reduction in potential interferences measured by scanning the mass range from 30-250 using the secondary electron multiplier (*Figure S2*).

During the analytical campaign at the University of Maryland, 31 separate 75 ng Os loads of the Johnson Matthey Os internal standard (UMCP) yielded average $^{186}\text{Os}/^{188}\text{Os}$ of 0.1198458 ± 15 (2 SD) and $^{187}\text{Os}/^{188}\text{Os}$ of 0.1137852 ± 23 (2 σ SD). These values agree well with measurements reported by [Brandon et al. \(2000\)](#) ($^{186}\text{Os}/^{188}\text{Os} = 0.1198464 \pm 12$; $^{187}\text{Os}/^{188}\text{Os} = 0.1138067 \pm 21$) and [Brandon et al. \(2006\)](#) ($^{186}\text{Os}/^{188}\text{Os} = 0.1198470 \pm 16$; $^{187}\text{Os}/^{188}\text{Os} = 0.1137908 \pm 36$) for their abyssal peridotite and Os-rich alloy datasets. For this reason, all published high precision Os isotope data for abyssal peridotites and Os-rich alloys are normalized to the UMCP value reported in this study ($^{186}\text{Os}/^{188}\text{Os} = 0.1198458$; $^{187}\text{Os}/^{188}\text{Os} = 0.1137852$). The data from [Chatterjee & Lassiter \(2016\)](#) has been renormalized to the $^{186}\text{Os}/^{188}\text{Os}$ UMCP composition. However, as these authors did not report the measured $^{187}\text{Os}/^{188}\text{Os}$ value for the UMCP standard, their $^{187}\text{Os}/^{188}\text{Os}$ data have not been renormalized. The UMCP standard values that we obtained are within the range of both 30 s and 300 s baseline measurements done by [Chatterjee & Lassiter \(2015\)](#) on <40 ng Os load sizes using a within-run O isotopic composition (30 s baseline integrations $^{186}\text{Os}/^{188}\text{Os} = 0.1198466 \pm 34$; $^{187}\text{Os}/^{188}\text{Os} = 0.113793 \pm 7$; 300 s baseline integrations $^{186}\text{Os}/^{188}\text{Os} = 0.1198474 \pm 36$; $^{187}\text{Os}/^{188}\text{Os} = 0.113791 \pm 5$).

As an independent measure of analytical reproducibility and precision, a new solution standard, OSUM8, was prepared from an Os-rich alloy. After digestion of the alloy, the Os fraction was multiply micro-distilled, leading to a highly-purified solution. This standard has been previously reported in [Ireland et al. \(2011\)](#), who obtained $^{186}\text{Os}/^{188}\text{Os} = 0.1198369 \pm 20$ and $^{187}\text{Os}/^{188}\text{Os} = 0.1314618 \pm 20$. When normalized to our value of UMCP, the values reported in [Ireland et al. \(2011\)](#) are 0.1198384 and 0.1314648. During this study, for highly purified OSUM8 solutions, prepared in an identical fashion to the samples, we obtained $^{186}\text{Os}/^{188}\text{Os} = 0.1198390 \pm 5$ and $^{187}\text{Os}/^{188}\text{Os} = 0.1314615 \pm 23$ ($n = 30$). Two aliquots of the Allende CV3 chondrite were all measured using the NiS protocol used for abyssal peridotites and yielded values within uncertainty of previous results ($^{187}\text{Os}/^{188}\text{Os} = 0.1254616 \pm 31$; $^{186}\text{Os}/^{188}\text{Os} = 0.1198345 \pm 16$; *Supplementary Information*).

3. Results

3.1 Major and trace element abundances

Major element variations in the studied abyssal peridotites (*Table S1*) are broadly consistent with melt depletion trends expected for mantle residues (**Figure 2**). Mg-numbers ($\text{Mg}/[\text{Mg}+\text{Fe}]$) define a limited range of 0.87 to 0.91, typical for residual abyssal peridotites (e.g., [Niu, 1997](#)). They lie along melt depletion trajectories predicted for adiabatic decompression melting beneath ridges that equate to >10% melt extraction. Some of the Gakkel Ridge peridotites from this study and [Liu et al. \(2009\)](#) have high CaO (>4 wt. %), P_2O_5 (>0.04 wt.%) and LOI, which may be a characteristic of alteration. Our samples have high compatible element concentrations (1740 to 3600 $\mu\text{g g}^{-1}$ Cr, 1400 to 2500 $\mu\text{g g}^{-1}$ Ni) and generally low incompatible element concentrations, as expected for mantle residues of melting. However, they also have high Sr (0.4 to 1286 $\mu\text{g g}^{-1}$) and U (0.3 to 980 ng g^{-1}) concentrations, and similarly elevated Pb and K abundances (*Figure S3*). These signatures are commonly interpreted to reflect alteration during serpentinization of abyssal peridotites (e.g., [Niu, 2004](#); [Deschamps et al., 2011](#)).

Based on their rare earth element (REE) patterns (**Figure 3**), samples can be broken into three broad types. The first include samples with low REE abundances ($<1 \times \text{CI-chondrite}$) and which show pronounced light REE/moderate REE (LREE [e.g., Ce]/MREE [e.g., Sm]) depletion relative to the heavy REE (HREE [e.g., Yb]). MAR abyssal peridotites are all this type, and patterns like this are also present in the SWIR, CIR and Gakkel Ridge suites. The second type have higher absolute REE abundances (>1 to $<10 \times \text{CI-chondrite}$) and relatively flat, CI-chondrite normalized REE patterns. These peridotites occur in the Gakkel Ridge, CIR and SWIR suites. The final type has low absolute REE abundances ($<1 \times \text{CI-chondrite}$), but a broadly flat, CI-chondrite normalized pattern. This sample type only occurs at the Gakkel Ridge in refractory harzburgites containing less than 1 wt.% Al_2O_3 and negligible clinopyroxene (HLY102-40-24; HLY102-40-81).

3.2 Highly siderophile element (HSE) abundances

The HSE abundances of SWIR, CIR, Gakkel Ridge and MAR abyssal peridotites are similar to those reported previously for abyssal peridotites ([Snow & Schmidt, 1998](#); [Luguet et](#)

al., 2001; 2003; Becker et al., 2006; Liu et al., 2009) (**Table 1; Figure 4**). MAR samples have absolute and relative HSE abundances that are akin to PM estimates, with elevated Pd/Ir (1.8 ± 0.6 ; 2SD) and Ru/Ir (2.0 ± 0.4). SWIR and CIR samples have similar absolute and relative abundances of Os, Ir and Ru to PM estimates (Ru/Ir = 2.4 ± 0.4 and 2.1 ± 0.1 , respectively), but CIR samples exhibit relative depletions in Pd and Re (Pd/Ir = 0.7 ± 0.1), compared with PM. Some SWIR samples are depleted in Pt, Pd and Re relative to PM (Pd/Ir = 0.7 ± 0.3). The pyroxenite VAN7-96-16 has lower absolute HSE abundances than associated peridotites, with high Ru/Ir (3.1), Pd/Ir (2.8) and low Pt/Pd (0.8). Gakkel Ridge abyssal peridotites exhibit the largest range in absolute and relative HSE abundances (e.g., Pd/Ir = 0.8 ± 0.9 ; Pt/Ir = 3.5 ± 3.4 ; Ru/Ir = 3.9 ± 4.5), with some samples showing substantial Pd depletion and Re enrichment (**Figure 4**). The three samples with the most fractionated patterns (HLY0102-40-24 Pt/Ir = 12; HLY0102-92-24 Ru/Ir = 15; HLY0102-40-81) are refractory harzburgites with less than 1 wt.% Al_2O_3 . In general, the HSE are not correlated with major or trace elements, with the exception of SiO_2 and Os ($R^2 = 0.6$) and SiO_2 and Pd ($R^2 = 0.5$), which are reasonably well correlated.

3.3 Osmium isotope systematics

The average $^{187}\text{Os}/^{188}\text{Os}$ for all the peridotites in this study is 0.1275 ± 0.0153 (2 SD; $n = 25$), with a median value of 0.1261 (**Table 1**). For the SWIR and MAR, $^{187}\text{Os}/^{188}\text{Os}$ are broadly similar, with average ratios of 0.1263 ± 26 ($n = 5$) and 0.1257 ± 30 ($n = 6$), respectively. A single SWIR pyroxenite possesses more radiogenic $^{187}\text{Os}/^{188}\text{Os}$ of 0.1463. CIR peridotites have $^{187}\text{Os}/^{188}\text{Os}$ ratios (0.1230 ± 24 , $n = 3$) that are, on average, less radiogenic than SWIR and MAR samples. Casting these ratios as Re depletion ages ($T_{\text{RD}} = 1/1.67 \times 10^{-11} \times \ln\{[(0.127 - ^{187}\text{Os}/^{188}\text{Os}_{\text{sample}})/0.40186] + 1\}$; where T_{RD} ages represent minimum ages, due to the assumption of no ingrowth from ^{187}Re in samples since melt depletion), yields model ages of ~ 0.5 Ga for SWIR and MAR samples, and ~ 0.9 Ga for the CIR. Gakkel samples span a range in $^{187}\text{Os}/^{188}\text{Os}$, from 0.1230 to 0.1587. Removing the most radiogenic samples (HLY0102-92-24; PS59-206-10; PS59-238-73) yields average $^{187}\text{Os}/^{188}\text{Os}$ for the Gakkel Ridge of 0.1256 ± 48 ($n = 5$), similar to the SWIR and MAR. When split into individual ridges, some abyssal peridotite suites fall along a reference isochron relative to the chondritic $^{187}\text{Os}/^{188}\text{Os}$ value at ~ 1 Ga (**Figure 5**).

High-precision $^{186}\text{Os}/^{188}\text{Os}$ measurements of abyssal peridotites range from 0.1198336 to 0.1198391, with the limits of the range defined by samples from the Gakkel Ridge (**Table 2** and *Tables S3* and *S4*). Abyssal peridotites from the SWIR (0.1198344 to 0.1198389), CIR (0.1198349 to 0.1198363) and MAR (0.1198337 to 0.1198363) have values that fall within the Gakkel Ridge range. The samples do not define isochronous relationships in $^{190}\text{Pt}/^{188}\text{Os}$ - $^{186}\text{Os}/^{188}\text{Os}$ space, instead lying to the right and beneath the 1 to 2 Ga reference isochrons, assuming the primitive mantle composition of this study (**Figure 6**). Coupled $^{186}\text{Os}/^{188}\text{Os}$ - $^{187}\text{Os}/^{188}\text{Os}$ systematics for abyssal peridotites indicate a restricted range of $^{186}\text{Os}/^{188}\text{Os}$ for a comparatively large range in $^{187}\text{Os}/^{188}\text{Os}$ (**Figure 7**), allowing us to define an average $^{186}\text{Os}/^{188}\text{Os}$ for the depleted MORB mantle (DMM) of 0.1198356 ± 21 ($n = 19$).

Twenty-six Os-rich alloys were analyzed from Mesozoic to Proterozoic ophiolites, including re-measurement of 17 alloys reported previously in Walker et al. (1997; 2005) and Brandon et al. (2006) (**Table 2**). These analyses were conducted to provide points-of-comparison with the abyssal peridotites measured in the same analytical campaign, since Os-rich alloys have been used to estimate the $^{186}\text{Os}/^{188}\text{Os}$ mantle composition. After renormalization to $^{192}\text{Os}/^{188}\text{Os}$ of 3.083, and the standard value used in this study, we find excellent agreement with our new data compared with previous measurements (*Table S4*). All of the alloys have exceptionally low Re/Os and Pt/Os, such that no measureable ingrowth of ^{187}Os or ^{186}Os has occurred since their formation (e.g., Walker et al., 2005). The alloys span a range in $^{187}\text{Os}/^{188}\text{Os}$, from ~0.107 to 0.147 and a range in $^{186}\text{Os}/^{188}\text{Os}$, from 0.1198248 to 0.1198458, with most alloys clustering at $^{187}\text{Os}/^{188}\text{Os}$ and $^{186}\text{Os}/^{188}\text{Os}$ ratios of 0.1249 and 0.119836, respectively (**Figure 8**). Osmium-rich alloys ‘freeze in’ the Os isotopic composition inherited from their mantle sources at the time of their formation. Hence, they can reveal mantle compositions and heterogeneities complementary to those observed in bulk rock peridotites. Alloys that formed with high $^{187}\text{Os}/^{188}\text{Os}$ (>0.13) have come from precursor mantle sources with Pt/Os within the range of chondrites (**Figure 8**). Other alloys have developed from sources with chondritic Re/Os, but low Pt/Os, as evidenced by correspondingly unradiogenic $^{186}\text{Os}/^{188}\text{Os}$. Conversely, one alloy (4Os) has a ~3 Ga Re-Os depletion age, but radiogenic $^{186}\text{Os}/^{188}\text{Os}$ with a precursor source Pt/Re calculated to be >5000. Fifty-four discrete Phanerozoic alloys (this study; Brandon et al., 2006) yield an average $^{186}\text{Os}/^{188}\text{Os}$ of 0.1198361 ± 20 and $^{187}\text{Os}/^{188}\text{Os}$ of 0.1256 ± 23 (2 SE).

4. Discussion

Abyssal peridotites represent residues of recent MORB melt extraction from sources that may have experienced prior melt depletion and/or refertilization processes (e.g., Allegre & Luck, 1980; Martin, 1991; Niu, 1997; Brandon et al., 2000; Salters & Dick, 2002; Cipriani et al., 2004; Harvey et al., 2006; Liu et al., 2008; Stracke et al., 2011; Lassiter et al., 2014; Mallick et al., 2014; Warren, 2016). The new data presented for the SWIR, CIR, MAR and Gakkel Ridge peridotites indicate variable and significant melt extraction, with evidence for a range of prior melt depletion episodes inferred from Os isotopes. Our results also significantly expand the available $^{186}\text{Os}/^{188}\text{Os}$ database for mantle materials. The new ^{186}Os - ^{187}Os isotope data for Os-rich alloys points to heterogeneity preserved in these materials, inherited from their mantle sources. Nonetheless, both the abyssal peridotite and Phanerozoic alloy datasets overlap with respect to average $^{187}\text{Os}/^{188}\text{Os}$ (0.126) and $^{186}\text{Os}/^{188}\text{Os}$ (0.1198356 ± 21 and 0.1198361 ± 20 , respectively). Prior to considering the implications of these data for mantle evolution, we first consider processes that are known to modify abyssal peridotites and mantle material.

4.1 Sea-floor alteration

Most abyssal peridotites have experienced serpentinization (<400 °C) and/or sea-floor weathering (~0°C) under a range of redox conditions (e.g., Snow & Dick, 1995; Bach et al., 2004; Paulick et al., 2006; Klein et al., 2013; Malvoisin, 2015). Samples in this study exhibit evidence for secondary alteration, manifested in the bulk rock data as loss on ignition (LOI) values up to 16 wt.%, which reflects the presence of serpentine and other alteration minerals observed petrographically. Only sample HLY102-40-81, from the Gakkel Ridge, is relatively fresh, with <5% serpentine (D'Errico et al., 2016) and 1% LOI. Elevated concentrations of fluid mobile elements, including U, K, Sr, P and Na, occur in many of the samples (Figure S3), consistent with modification by interaction with seawater (e.g., Niu, 2004; Harvey et al., 2006). However, most major and trace elements appear unaffected by alteration processes. For example, the peridotites do not deviate significantly from the 'terrestrial array' in $\text{Al}_2\text{O}_3/\text{SiO}_2$ - MgO/SiO_2 space, implying no significant addition of Si (cf. Malvoisin, 2015). This observation implies that positive correlations between SiO_2 and both Os and Pd reflect primary signatures in the samples.

Previous studies have suggested that $^{187}\text{Os}/^{188}\text{Os}$ (e.g., [Snow & Reisberg, 1995](#); [Standish et al., 2002](#)), Re and Pd (e.g., [Luguet et al., 2003](#); [Harvey et al., 2006](#)) can be modified by hydrothermal alteration and sea-floor weathering. Comparison of LOI and other indicators of alteration (e.g., U) with HSE contents and Os isotopic compositions yields no correlations (*Figure S4*), although the highest Re contents are found in MAR and Gakkel samples with high LOI. Absolute HSE abundances show greater dispersion at higher LOI, but this reflects a sampling bias, due to most peridotites in this study having LOI of ~12 wt. %, with few data for fresh peridotites. Overall, we conclude that the bulk-rock abyssal peridotite compositions predominantly reflect their high temperature mantle petrogenesis. This is consistent with preservation of mantle peridotite Os isotope and HSE abundances in steatized talc-magnesite samples (LOI = ~20 wt.%) associated with the Shetland Ophiolite Complex ([Day et al., 2017](#)).

4.2 Melt refertilization

Melt refertilization can affect abyssal peridotites, especially at slow and ultraslow spreading ridges ([Niu, 2004](#); [Seyler et al., 2004](#); [Warren et al., 2009](#)), leading to potential for precipitation of metasomatic sulfides rich in Pd and other HSE ([Luguet et al., 2003](#)). Clear evidence for melt refertilization in the sample suite occurs in two forms. First, enrichments in the LREE and other highly incompatible elements occur in some of the samples from the SWIR, MAR and Gakkel Ridges (**Figure 3**), a feature that has previously been attributed to refertilization processes (e.g., [Niu, 2004](#); [Warren, 2016](#)). A second independent line of evidence is that pyroxenite veins occur in one of the SWIR dredges and have been interpreted to result from recent melt-rock interaction related to ridge migration over the Bouvet hotspot at 15 to 25 Ma ([Warren et al., 2009](#)). The pyroxenite Van7-96-16 has significantly lower HSE contents than associated peridotites, with low contents of Re (0.011 ng g⁻¹). Due to the low Re abundance of Van7-96-16, the age-corrected $^{187}\text{Os}/^{188}\text{Os}$ at $\sim 20 \pm 5$ Ma is suprachondritic (0.1461) and, along with the $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7034) and $^{143}\text{Nd}/^{144}\text{Nd}$ (0.5128) compositions for clinopyroxene from this sample ([Warren et al., 2009](#)), is consistent with a recycled signature for infiltrating melts (e.g., [Le Roux et al., 1992](#); [Mallick et al., 2015](#)).

Evidence for melt interaction in the peridotites, based on their HSE concentrations and Os isotopic compositions, is limited. In the case of the Bouvet hotspot samples, the HSE

composition of the most extensively reacted peridotite sample (Van7-96-35: $^{187}\text{Os}/^{188}\text{Os} = 0.128$; Pd = 1.1 ng g⁻¹) does not deviate from depleted SWIR peridotites (Van7-85-27/Van7-85-30: $^{187}\text{Os}/^{188}\text{Os} = 0.125\text{--}0.127$; Pd = 2–2.1 ng g⁻¹), suggesting that interaction with hotspot melts did not significantly affect peridotite HSE compositions. Basaltic melts typically have more radiogenic $^{187}\text{Os}/^{188}\text{Os}$, but substantially lower Os contents than peridotites, meaning that only high degrees of melt-rock reaction will lead to modification of peridotite HSE compositions. For example, alkali basaltic melt (Os = ~0.067 ng g⁻¹; $^{187}\text{Os}/^{188}\text{Os} = \sim 0.146$; Day, 2013) and MORB (Os = ~0.010 ng g⁻¹; $^{187}\text{Os}/^{188}\text{Os} = \sim 0.133$; Gannoun et al., 2016) compositions have limited (<1%) effects on $^{187}\text{Os}/^{188}\text{Os}$ in abyssal peridotites such as Van7-85-27 (~3.5 ng g⁻¹ Os; $^{187}\text{Os}/^{188}\text{Os} = \sim 0.125$), until high melt-rock ratios (>10). Even a hypothetical melt with the composition of Van7-96-16 would only result in a 0.6% change in the ^{187}Os isotopic composition at 30% melt addition.

Effects on $^{186}\text{Os}/^{188}\text{Os}$ are even more limited than for $^{187}\text{Os}/^{188}\text{Os}$. Assuming a melt composition of 0.32 ng g⁻¹ Os (c.f., Van7-96-16) with the most radiogenic $^{186}\text{Os}/^{188}\text{Os}$ measured for oceanic island picrite to date (0.119848; Ireland et al., 2011) would yield a 4 ppm change in the Os isotopic composition of a typical abyssal peridotite at 30% melt addition. Sulfide melt addition has been advocated as a modifier of HSE compositions in abyssal peridotites (e.g., Rehkamper et al., 1999; Luguet et al., 2001; Alard et al., 2005) and as potential carriers of radiogenic $^{186}\text{Os}/^{188}\text{Os}$ (e.g., Luguet et al., 2008). Contents of Cu, which can be used as a proxy for Cu-Ni-rich sulfide melt addition, do not show a positive correlation with HSE content (Table S1). This suggests that sulfide melt addition is not a significant driver in HSE, $^{187}\text{Os}/^{188}\text{Os}$, or $^{186}\text{Os}/^{188}\text{Os}$ variations in the studied abyssal peridotites.

4.3 Partial melting

Models of non-modal fractional melt-loss for the abyssal peridotites based on REE abundances (Figure 3, e.g., Warren, 2016) and Cr-spinel compositions (e.g., Hellebrand et al., 2001) are broadly consistent. Considering only the MREE/HREE (rather than LREE-enriched signatures for some of the peridotites), the REE modelling calculations suggest 10–14% melt extraction for MAR, 2–9% for SWIR, 4–>16% for Gakkell Ridge, and 2–12% melt depletion for

CIR peridotites. These estimates are consistent with previous fractional melting estimates for Gakkel Ridge lherzolites (4-6%) and harzburgites (6->16%) (D'Errico et al., 2016).

Experiments on a range of sulfide compositions indicate that sulfide is partially to totally molten at upper mantle conditions (Zhang & Hirschmann, 2016). Melt extraction leads to the formation of Cu-Ni-rich sulfide melt enriched in Pt, Pd and Re, leaving a residual monosulfide solid solution (MSS) that concentrates Os, Ir and Ru (e.g., Alard et al., 2000; Luguet et al., 2003; Ballhaus et al., 2006). However, no primary MSS sulfide has been found in abyssal peridotites to date, with all observed sulfides being secondary, metasomatic sulfides (e.g., Luguet et al., 2003; Liu et al., 2009; Harvey et al., 2016), leading to the alternative possibility that Os, Ir, and Ru may be hosted in refractory metal nuggets (e.g., Luguet & Reisberg, 2016). Several of our samples are depleted in Re, Pd $\pm$ Pt, relative to Os, Ir and Ru, qualitatively reflecting extraction of sulfide melt, along with silicate melt.

To model the evolution of bulk rock HSE contents during partial melting, we assumed control of sulfides on the HSE during melting (**Figure 9**). This model is similar to that used previously for MAR and Gakkel Ridge abyssal peridotites (Luguet et al., 2003; Liu et al., 2009). We assume that sulfides are consumed by dissolution into the silicate melt and that they are exhausted when the partial extent of silicate melting reaches 15% (Luguet et al., 2003), consistent with a DMM with ~150 ppm S. The evolution of HSE concentrations in the sulfides is thus governed by the sulfide melt to silicate melt partition coefficients. We use two starting compositions: one with PM Pt/Ir and Pd/Ir values, and the other with CI-chondrite Pd/Ir. For consistency with Liu et al. (2009), we use the same partition coefficients as them ($D_{Pt} = 100,000$; 10,000; 1000; $D_{Pd} = 17,000$; 5000; 1000). Much higher partitioning between sulfide melt and silicate melt have recently been proposed by Mungall & Brenan (2014; Table 2), with D_{Ir} of $\sim 4.6 \pm 6.9 \times 10^5$, D_{Pt} of $\sim 8.5 \pm 11 \times 10^5$, and D_{Ir} of $\sim 1.9 \pm 1.6 \times 10^5$, and we also show models using these partition coefficients in the model (**Figure 9**).

Samples are consistent with fractionation of Pt from Ir at relatively high degrees of melt exhaustion (>10 %), consistent with exhaustion of sulfide from the residue. Conversely, many Gakkel samples, and all of the SWIR and CIR samples, have Pd/Ir ratios too low to be

accounted for by either high D_{Pd} (rock/melt partition coefficient) from a source with primitive mantle Pd/Ir, or melt depletion of a source with CI-chondrite-like Pd/Ir. In either case, the models require that Pd is more incompatible ($D_{\text{Pd}} < 500$) than the other studied HSE during partial melting. In general, abyssal peridotites with low Pd/Ir also have low Re (e.g., **Figure 4**) consistent with exhaustion of both Re and Pd by partial melting, likely through different fractionation processes. In the case of Pd, sulfide stability is likely to be a controlling factor, whereas for Re, melting of silicates may be more significant. No clear correlations are observed between indicators of melt extraction, such as Al_2O_3 and Re/Ir, Pt/Ir, Ru/Ir or Os/Ir, but there is a correspondence of lower Pd/Ir with low Al_2O_3 , and a stronger correlation between Al_2O_3 and $^{187}\text{Os}/^{188}\text{Os}$ (**Figure 10**). The variations in HSE inter-element ratios, but correspondence of $^{187}\text{Os}/^{188}\text{Os}$ with Al_2O_3 , appear to record melt depletion both as a function of relatively recent sulfide melt removal at the ridge, as well as more ancient melt depletion that has acted on all four of the studied ridges. Similar observations have previously been made for ~ 6 Ma peridotites from the Taitao Ophiolite, Chile (Schulte et al., 2009).

4.4 Pre-existing versus recent ridge-generated heterogeneities

Studies of abyssal peridotites and ophiolite peridotites have found that their $^{187}\text{Os}/^{188}\text{Os}$ systematics can be explained by recent on-ridge melt extraction superimposed on ancient melt depletion trends (e.g., Harvey et al., 2006; Liu et al., 2008; Schulte et al., 2009; O'Driscoll et al., 2012; Warren and Shirey, 2012). Pre-existing heterogeneity recorded in abyssal peridotites has been noted in several ^{187}Re - ^{187}Os studies, with Re depletion (T_{RD}) ages as ancient as 2 Ga (Harvey et al., 2006; Liu et al., 2008). Using a compilation of abyssal peridotites, Lassiter et al. (2014) proposed that the 'average' T_{RD} age for abyssal peridotites was 0.7 ± 0.6 Ga and the average time of separation from the mantle (T_{MA}) was 1.7 ± 0.9 Ga, with $T_{\text{MA}} (= 1/1.668 \times 10^{-11} \times \ln[(0.127 - ^{187}\text{Os}/^{188}\text{Os}_{\text{sample}})/(0.40186 - ^{187}\text{Re}/^{188}\text{Os}_{\text{sample}}) + 1])$. Similarly, studies of ophiolite peridotites implicate ancient melt depletion (Büchl et al., 2004; Schulte et al., 2009; O'Driscoll et al., 2012; 2015). The samples that we studied have a range of T_{RD} ages, up to 1.1 Ga, with an average of $\sim 0.6 \pm 0.6$ Ga and T_{MA} of 1 ± 1 Ga. A modified compilation of abyssal peridotite data (Supplementary Information) yields a modified T_{RD} age (0.8 ± 0.5 Ga) identical, within uncertainty, of previous estimates (Lassiter et al., 2014).

A key question is whether ancient melt depletion experienced by abyssal peridotites can be distinguished from recent melt depletion at the ridge. There are no correlations between $^{187}\text{Os}/^{188}\text{Os}$ or T_{RD} with degree of melt depletion (F) to indicate that pre-existing melt depletion can be decoupled from recent-ridge melt depletion in a straightforward manner. However, a preponderance of abyssal peridotites with ancient T_{RD} ages also have low Pd/Ir, suggesting that they preserve significant pre-existing melt depletion ($>10\%$; **Figure 11**). Given the juxtaposition of ancient and more recently melt depleted abyssal samples at ridge segments, this observation leads to the conclusion that fractions of the upwelling mantle beneath ridges are already extensively melt-exhausted. Such a scenario is similar to observations of ancient melt depletion in ophiolites, where field relationships reveal different length-scales of heterogeneity, including juxtaposition of peridotite with different Re depletion ages within centimeter to meter scales (O'Driscoll et al., 2012; 2015). However, T_{RD} ages for samples with different degrees of melt depletion (i.e., harzburgite versus lherzolite) may reflect either temporal or spatial variations in melt depletion and thus may not reflect true age differences.

Studies of Sr and Nd isotope variations in abyssal peridotites have recognized that they extend to significantly more depleted values than mid-ocean ridge basalts (Snow et al., 1994; Kempton & Stephens, 1997; Salters & Dick, 2002; Cipriani et al., 2004; Warren et al., 2009). In a study of SWIR peridotites, Warren et al. (2009) found that elevated $^{143}\text{Nd}/^{144}\text{Nd}$ ratios in some abyssal peridotite clinopyroxene fractions reflect ancient mantle heterogeneity and recent modification by melting, melt-rock reaction and melt crystallization. Studies of Nd-Hf isotope ratios in abyssal peridotites have also revealed that SWIR peridotites mostly overlap those of MORB, whereas long-term depletion has led to Gakkel Ridge samples extending the MORB and ocean island basalt Hf–Nd isotope array to strongly depleted values (Stracke et al., 2011; Mallick et al., 2015). The correlation of Hf isotopes with indices of depletion (e.g., Al_2O_3 , Yb, spinel Cr#) and Os isotope ratios led Stracke et al. (2011) to argue that Hf and Os isotope compositions preserve a record of ancient mantle depletion, whereas their Nd isotope signatures often do not.

A compilation of available whole-rock Os isotope data with clinopyroxene separate Nd and Hf isotope data on the same abyssal peridotites reveals no clear correlations between $^{187}\text{Os}/^{188}\text{Os}$ and $^{143}\text{Nd}/^{144}\text{Nd}$, but a strong negative correlation between $^{176}\text{Hf}/^{177}\text{Hf}$ and

$^{187}\text{Os}/^{188}\text{Os}$ in Gakkel Ridge peridotites (**Figure 12**). The negative correlation between Hf and Os isotopes in the Gakkel Ridge is consistent with both isotopes recording ancient mantle depletion, with high long-term time-integrated Lu/Hf ratios preserved in some strongly depleted Gakkel peridotites with T_{RD} ages of 0.5 Ga or greater. This observation is similar to the correlation observed by [Warren and Shirey \(2012\)](#) between $^{187}\text{Os}/^{188}\text{Os}$ and Pb isotopes in sulfides from Gakkel and SWIR peridotites. However, there is no corresponding correlation between $^{176}\text{Hf}/^{177}\text{Hf}$ and $^{187}\text{Os}/^{188}\text{Os}$ in SWIR peridotites. Instead, these peridotites preserve evidence for melt depletion in Os isotopes, but not Hf isotopes. This suggests that the robustness of Sr-Nd-Hf-Os-Pb isotopes to preservation of ancient melt depletion decreases in the order of $\text{Os} > \text{Hf} > \text{Nd} > (\text{Pb?}) \geq \text{Sr}$.

Melt-depletion in bulk rock peridotites, recorded as a decrease in Al_2O_3 , can correlate with $^{187}\text{Os}/^{188}\text{Os}$ (e.g., [Reisberg & Lorand, 1995](#)) (**Figure 10**). Comparison of whole-rock Al_2O_3 contents with $^{187}\text{Os}/^{188}\text{Os}$ and $^{186}\text{Os}/^{188}\text{Os}$ indicates a positive relationship, though with poor correlation, suggesting that both long-term Re/Os and Pt/Os ratios track melt-depletion (**Figure 13**). Samples with low Al_2O_3 typically have the lowest $^{187}\text{Os}/^{188}\text{Os}$ and $^{186}\text{Os}/^{188}\text{Os}$. We used a similar approach to [Meisel et al. \(2001\)](#), who assumed a PM composition with 4.5 wt. % Al_2O_3 ([McDonough and Sun, 1995](#)). We also used a mean depleted mantle composition of 4 wt. % Al_2O_3 ([Workman & Hart, 2005](#)) as a lower bound on our estimate. Using these Al_2O_3 values and the 95% confidence limit bounds, we calculated primitive mantle $^{187}\text{Os}/^{188}\text{Os}$ of 0.1292 ± 25 and $^{186}\text{Os}/^{188}\text{Os}$ of 0.1198388 ± 29 (**Figure 13**).

4.5 *Pt-Re-Os evolution and HSE abundances of the mantle*

Prior constraints on the coupled Pt-Re-Os evolution of the PM from peridotites comes from the work of [Brandon et al. \(2000\)](#). That study showed that the $^{186}\text{Os}/^{188}\text{Os}$ ratios of abyssal peridotites were consistent with those of Os-rich alloys and that the Pt/Os ratio of the bulk mantle has not deviated by more than $\pm 30\%$ from chondritic Pt/Os over Earth history. Several high-precision Os isotope studies have also been reported on Os-rich alloy grains ([Walker et al., 1997](#); [Brandon et al., 1998](#); [2006](#); [Meibom & Frei, 2002](#); [Meibom et al., 2004](#); [Walker et al., 2005](#)). Osmium-rich alloys are typically found in placers downstream from ultramafic massifs ([Cabri & Harris, 1975](#); [Stockman & Hlava, 1984](#)). Possible origins for these grains include

reduction and release of Os, Ir and Ru following serpentinization of peridotite (Cabri & Harris, 1975), metasomatic products of melts passing through mantle lithosphere (Brenker et al., 2003) and crystallization from sulfide melts as sulphur is lost to the silicate melt (Fonseca et al., 2011, 2012).

The average $^{186}\text{Os}/^{188}\text{Os}$ compositions obtained for the abyssal peridotites (0.1198356 ± 21) and Phanerozoic Os-rich alloys (0.1198361 ± 20) are similar. These values are higher than the average $^{186}\text{Os}/^{188}\text{Os}$ of carbonaceous chondrite Allende (0.119834) or low-Fe ordinary chondrites (0.1198344), but are less radiogenic than high-Fe ordinary chondrites (0.1198386), being most similar to the average of enstatite chondrites (0.1198356 ; Brandon et al., 2005; 2006). A straightforward interpretation of this relationship would be a mantle that has evolved with a Pt/Os ratio similar to enstatite chondrites. However, as noted above, these averages likely incorporate long-term time-integrated melt depletion occurring in the depleted MORB mantle. Alternatively, if the Al_2O_3 - $^{187}\text{Os}/^{188}\text{Os}$ regression estimate used to establish PM composition is used ($^{187}\text{Os}/^{188}\text{Os} = 0.1292 \pm 25$; $^{186}\text{Os}/^{188}\text{Os} = 0.1198388 \pm 29$), then the mantle is most similar to ordinary H chondrites or to equilibrated enstatite chondrites (0.1198396 ± 9 ; Brandon et al., 2005). Here, we interpret the average ratio of abyssal peridotites and Os-rich alloys to reflect the average evolution of the DMM, with a $^{190}\text{Pt}/^{186}\text{Os}$ ratio of 0.0018, whereas a higher $^{190}\text{Pt}/^{186}\text{Os}$ ratio (0.0022) is estimated for the PM. The difference between these two values is that DMM is derived from the abyssal peridotite data whereas PM is the extrapolation of this data to a higher Al_2O_3 content (4.25 wt.%).

To estimate the average HSE abundance of the primitive upper mantle, Becker et al. (2006) used lherzolites with >2 wt.% Al_2O_3 from a range of settings, including post-Archaeon continental lithospheric mantle, Archaean continental lithospheric mantle and Phanerozoic oceanic peridotites. These authors estimated Ir in the primitive upper mantle from the average value of their dataset, Os and Ru from the average values of Os/Ir and Ru/Ir, Pt and Pd from Pt/Ir and Pd/Ir extrapolated to 4.25% Al_2O_3 , and Re from $^{187}\text{Os}/^{188}\text{Os}$ extrapolated to 4.25% Al_2O_3 . We employ a similar method to estimate the HSE composition of the PM based on abyssal peridotites, but using two methods. First, we took the mean concentration of 34 abyssal peridotites with >2 wt. % Al_2O_3 from this work and published data (Table S5). This yields

primitive mantle values derived from abyssal peridotites of $3.6 \pm 1.0 \text{ ng g}^{-1}$ [Ir], $3.9 \pm 1.2 \text{ ng g}^{-1}$ [Os], $7.9 \pm 2.2 \text{ ng g}^{-1}$ [Ru], $7.6 \pm 0.5 \text{ ng g}^{-1}$ [Pt], $4.8 \pm 0.7 \text{ ng g}^{-1}$ [Pd] and $0.26 \pm 0.07 \text{ ng g}^{-1}$ [Re] (labelled as PM in **Figure 14**). Second, we regressed data to a PM Al_2O_3 value of 4 to 4.5 wt. %. We obtained similar values ($3.7 \pm 1.0 \text{ ng g}^{-1}$ [Ir], $3.7 \pm 1.0 \text{ ng g}^{-1}$ [Os], $7.4 \pm 1.0 \text{ ng g}^{-1}$ [Ru], $8.5 \pm 1.7 \text{ ng g}^{-1}$ [Pt], $5.3 \pm 1.1 \text{ ng g}^{-1}$ [Pd] and $0.25 \pm 0.04 \text{ ng g}^{-1}$ [Re]) to our first estimate, albeit using a larger database of abyssal peridotites ($n = 60$). These concentrations would be higher if volumetric changes in the rocks due to serpentinization act to dilute the HSE abundances (*Supplementary Information*).

The most precise estimates of Re and Pt contents in the PM come from the measured $^{187}\text{Os}/^{188}\text{Os}$ and $^{186}\text{Os}/^{188}\text{Os}$ values (**Figure 14**). A compilation of 209 abyssal peridotite $^{187}\text{Os}/^{188}\text{Os}$ measurements yields an average composition of 0.1266 ± 0.0075 . Using the Os concentrations in samples to generate a weighted mean yields less radiogenic $^{187}\text{Os}/^{188}\text{Os}$ of 0.1252, which is close to the average of samples with $>2 \text{ ng g}^{-1}$ Os (0.1247 ± 0.0075) and the recent estimate for convecting upper mantle of ~ 0.1245 (Lassiter et al., 2014). Taking these values, we obtain a long-term time-integrated $^{187}\text{Re}/^{188}\text{Os}$ ratio of 0.361 to 0.384 and Re concentrations of 0.29 to 0.31 ng g^{-1} . Similarly, using the average $^{186}\text{Os}/^{188}\text{Os}$ of the samples analysed here of 0.1198356 yields a $^{190}\text{Pt}/^{186}\text{Os}$ ratio (0.0018) that is lower than the $^{190}\text{Pt}/^{186}\text{Os}$ ratio (0.0022) inferred for a PM composition at 4 to 4.5 wt.% Al_2O_3 ($^{186}\text{Os}/^{188}\text{Os} = 0.1198388$). Assuming PM Os equals 3.9 ng g^{-1} , equates to a Pt concentration for the depleted MORB mantle of $\sim 7 \text{ ng g}^{-1}$ and for the PM of $\sim 8.5 \text{ ng g}^{-1}$. The new data for abyssal peridotites provide precise constraints on the Pt/Os (2.0 ± 0.2) and Re/Os (0.08 ± 0.01) evolution of the PM, indicating that it evolved with ratios of these elements within the range of chondrites (Re/Os = 0.06-0.08; Pt/Os = 1.7-1.9; Day et al., 2016b).

Our calculated PM composition based on oceanic peridotites is similar to the PM estimates provided by Becker et al. (2006) and Chatterjee and Lassiter (2016), with one notable exception. The estimate solely from abyssal peridotites does not have high Pd/Ir and the pronounced Pd enrichment noted in the Becker et al. (2006) PM composition. This result does not seem strongly affected by Pd melt-depletion in abyssal peridotites. Instead, both the Becker et al. (2006) and Chatterjee & Lassiter (2016) estimates of PM composition are partly reliant on

continental-derived peridotites, which have been shown to experience significant melt-refertilization, including enrichment in Pd (e.g., Aulbach et al., 2016; Luguet & Reisberg, 2016; Becker & Dale, 2016).

4.6 Implications for late accretion and $^{186}\text{Os}/^{188}\text{Os}$ isotope variations in the mantle

Metal-silicate segregation results in near-quantitative HSE depletion in the silicate portions of planetary bodies at low pressures (e.g., Brenan et al., 2016). Elevated and broadly chondritic-relative HSE abundances estimated for the mantle ($\sim 0.01 \times \text{CI-chondrite}$) are, therefore, too high to be explained by low-pressure metal-silicate partitioning alone (Kimura et al., 1974; Chou, 1978; Morgan, 1986). Several hypotheses have been proposed to account for this discrepancy, including late accretion of between 0.5-0.8% of planetary mass to the silicate Earth that followed the cessation of core growth (Chou, 1978; Day et al., 2016b), or high-pressure, high-temperature partitioning (e.g., Ringwood, 1977; Murthy, 1991).

The new results for abyssal peridotites and Os-rich alloys support the long-term evolution of the mantle with Re/Os and Pt/Os within the range of chondrites and place stringent constraints on models used to account for the HSE abundances present in the mantle (e.g., Walker et al., 1997). Our results are consistent with previous suggestions that the late accreting materials had Re/Os and Pt/Os ratios most similar to H ordinary chondrites or to equilibrated enstatite chondrites (Brandon et al., 2006). By contrast, it remains uncertain as to whether high-temperature, high-pressure equilibration experiments can reproduce the necessary inter-element HSE ratios consistent with upper mantle composition (e.g., Brenan et al., 2016), since the metal-silicate D values for all the HSE must be within a factor of two or closer to explain the measured abundances in terrestrial peridotites and Os-rich alloys. An important aspect of the estimates of HSE abundances in the PM from our study and that of Becker et al. (2006) is that both have high Ru/Ir, relative to carbonaceous, ordinary and enstatite chondrite meteorites. Whether this enrichment reflects a combination of metal-sulfide-silicate partitioning and late accretion (e.g., Peters et al., 2016), late accretionary impactors with high Ru/Ir ($\pm \text{Pd/Ir}$) (e.g., Morgan et al., 2001), or a partial melting and refertilization effect in the mantle, remains unresolved (e.g., Lorand & Luguet, 2016). One possible cause is fractionation by sulfide liquids at high temperatures and pressures, which can cause Ru/Ir fractionation (Laurenz et al., 2015).

Higher $^{186}\text{Os}/^{188}\text{Os}$ ratios (up to 0.1198463) have been measured in plume-related lavas than in the estimated PM value (0.1198388 ± 29) (e.g., Walker et al., 1997; Brandon et al., 1998; 1999; 2003; Puchtel et al., 2005; Ireland et al., 2011). While these enrichments were previously interpreted in terms of Re/Os and Pt/Os fractionation due to inner core crystallization followed by core-mantle exchange (e.g., Walker et al., 1995; Brandon et al., 1998), or from basal metal isolated in the mantle (Humayun, 2011), the young age of the inner core and lack of clear evidence for core-mantle or lower mantle interaction has led to some dispute of these hypotheses (e.g., Schersten et al., 2004; Lassiter, 2006; Willbold et al., 2011). More recently, models have focussed on HSE fractionation in the mantle through sulfide melting and metasomatism, or through the development of lithologies with high Pt/Os and/or Re/Os (e.g., Luguet et al., 2008; Day, 2013; Marchesi et al., 2014; O'Driscoll et al., 2015). The new data for abyssal peridotites and Os-rich alloys imply that alternative reservoirs, other than mantle peridotite may be required to explain high $^{186}\text{Os}/^{188}\text{Os}$ in some plume-related lavas. This requires that volumetrically significant reservoirs with high Os and Pt/Os must exist that are either not sampled as mantle peridotites or alloys, or that are only sampled during plume-related magmatism in some intraplate magmatic systems.

5. Conclusions

Abyssal peridotites from the Southwest Indian (SWIR), Central Indian (CIR), Mid-Atlantic (MAR) and Gakkel Ridges exhibit evidence for both recent melt extraction beneath the ridge, as well as ancient melt depletion. Abyssal peridotites with T_{RD} ages >0.5 Ga typically exhibit low Pd/Ir, and combined with high-degrees of estimated total melt extraction imply that they were relatively refractory residues prior to incorporation into the ridge. Consequently, they may have played little part in partial melting processes beneath their respective ridges. Of the radiogenic isotope systems (Sr, Nd, Pb, Hf, Os) that have been analyzed in abyssal peridotites, Os appears to provide the most robust record of melt depletion. SWIR, CIR, MAR and Gakkel Ridge peridotites yield an average $^{186}\text{Os}/^{188}\text{Os}$ for the oceanic upper mantle of 0.1198356 ± 21 . High-precision analyses of Os-rich alloys yield average $^{186}\text{Os}/^{188}\text{Os}$ within uncertainty of the abyssal peridotite value (0.1198361 ± 20). Evidence for melt depletion trends between Os isotopes and indices of melt extraction (e.g., Al_2O_3) allow an estimate of the hypothetical

primitive mantle (PM) composition, yielding $^{187}\text{Os}/^{188}\text{Os}$ (0.1292 ± 25) within uncertainty of previous estimates, and $^{186}\text{Os}/^{188}\text{Os}$ of 0.1198388 ± 29 . This value is still significantly less radiogenic than that measured in some plume-related lavas, requiring sources with long-term time-integrated Pt/Os higher than in the primitive or depleted MORB mantles. The $^{186}\text{Os}/^{188}\text{Os}$ value for the PM is similar to that previously suggested from Os-rich alloys and is most similar to the composition of H ordinary chondrites or equilibrated enstatite chondrites. Estimates of PM HSE concentrations using only abyssal peridotites have chondritic Pd/Ir, in contrast to estimates from peridotites from a wider range of tectonic settings. However, all peridotite estimates yield non-chondritic Ru/Ir. The cause of enhanced Ru in the mantle remains enigmatic, but may reflect variable partitioning behaviour of Ru during high pressure, high temperature partitioning. Our results confirm that abyssal peridotites are not simple residues of recent mid-ocean ridge basalt source melting. Instead, they have a more complex depletion history, reflecting partial melting processes that produce Re/Os and Pt/Os isotopic ratios within the range of chondrites, along with broadly chondritic abundances of the HSE.

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Figures and figure captions

Figure 1: Location map of sample sites used in this study. Abyssal peridotite samples come from the Arctic (Gakkel Ridge), Atlantic (mid-Atlantic ridge, Kane fracture zone [MAR (KANE), or MARK]), and Indian oceans (central Indian ridge [CIR] at the Marie Celeste fracture zone and south-west Indian ridge [SWIR]). Dotted lines show approximate locations of the mid-ocean rift system.

Figure 2: Variation of (a) Al_2O_3 , (b) TiO_2 , (c) FeO_T , (d) CaO versus MgO, (e) Al_2O_3 versus CaO, and (f) LOI versus P_2O_5 for abyssal peridotites measured in this study and published Gakkel Ridge data from Liu et al. (2009). Also shown are comparison data from the Leka ophiolite complex, Norway (O'Driscoll et al., 2015; unfilled diamonds), and models for partial melting from a primitive mantle (PM) composition from Niu (1997).

Figure 3: CI-chondrite normalized plots of SWIR (filled circles), CIR (unfilled circles), Gakkel Ridge (filled triangles) and MAR (unfilled squares) abyssal peridotites measured in this study. Gray lines indicate 1% melt increments for a non-modal fractional melting model, where increasing melt depletion results in lower absolute abundances of the REE and increasing depletion in the LREE relative to HREE. Model parameters and normalization are provided in the *Supplementary Information*.

Figure 4: Chondrite-normalized highly siderophile element patterns for abyssal peridotites from (a) the South-West Indian Ridge (SWIR), (b) Central Indian Ridge (CIR), (c) Gakkal Ridge, and (d) Mid-Atlantic Ridge at the Kane fracture zone (MARK). New data shown as lines and symbols and published data shown as grey lines (*Table S5*). Refractory harzburgites (<1 wt.% Al_2O_3) from the Gakkal Ridge (HLY102-40-24, -40-81, -92-24) are identified and have the most fractionated HSE patterns of all of the suites of abyssal peridotite studied. Primitive mantle (PM) composition (thick green line) from [Becker et al. \(2006\)](#). Data are shown with most incompatible elements on the left (cf. [Day, 2013](#)).

Figure 5: $^{187}\text{Re}/^{188}\text{Os}$ - $^{187}\text{Os}/^{188}\text{Os}$ diagrams for abyssal peridotites from this study, showing 0.1, 1 and 2 Ga reference isochrons assuming chondritic Os isotope evolution.

Figure 6: $^{190}\text{Pt}/^{188}\text{Os}$ - $^{186}\text{Os}/^{188}\text{Os}$ diagram for abyssal peridotites from this study, showing 0.1, 1 and 2 Ga reference isochrons assuming a primitive mantle composition (this study) and chondritic Pt/Os.

Figure 7: Osmium isotope systematics of abyssal peridotites. The global average $^{186}\text{Os}/^{188}\text{Os}$ of abyssal peridotites (0.1198356 ± 21) is shown as a black stippled line, while the mean and 95% confidence limits for samples in this study are shown as grey lines. The average $^{186}\text{Os}/^{188}\text{Os}$ composition at a PM $^{187}\text{Os}/^{188}\text{Os}$ value of 0.1296 ± 0.0008 (Meisel et al., 2001) is 0.1198358 ± 14 , using 95% confidence limits. All samples in this study plot within uncertainty of this value. Also shown are data for Lena Trough abyssal peridotites (filled diamonds), Rio Grande Rift and Colorado Plateau (grey diamonds) and Salt Lake Crater (Hawaii; unfilled diamonds) peridotites from Chatterjee & Lassiter (2016). Error bars for the Chatterjee and Lassiter (2016) data are based on individual sample measurements, whereas the larger error bars for our data are based on replicate analyses using multiple, separate digestions of samples. Hence, error associated with the Chatterjee and Lassiter (2016) data likely under-estimate reproducibility.

Figure 8: Osmium isotope systematics of Os-rich alloys compared to abyssal peridotites (shaded field, with average denoted by stippled lines) and average compositions of chondrites (from Brandon et al., 2006; Allende from this study). Also shown is a chondritic evolution growth curve with time increments in Ga, from which evolution curves at 3.5 Ga with Pt/Re of 2.2 and 6000 are shown. For these curves, Solar System initial values at 4.567 Ga of 0.0952 ± 0.0002 for $^{187}\text{Os}/^{188}\text{Os}$ and 0.119823 ± 5 for $^{186}\text{Os}/^{188}\text{Os}$ were used (Day et al., 2016b).

Figure 9: Modeling of primitive mantle normalized (a) Pt/Ir and (b) Pd/Ir in abyssal peridotites during partial melting. Degree of partial melting (F , in percent) is derived from Cr-spinel compositions (Hellebrand et al., 2001). For models with associated partition coefficients for Pt of 100,000, 10,000, and 1000 and for Pd of 17,000, 5000 and 100, a constant sulfide-silicate melt partition coefficient was used for Ir ($D_{Ir} = 26,000$; Fleet et al., 1999) with partition coefficients marked along the model curves. The model curve denoted ‘M&B’ utilizes silicate melt-sulphide melt partitioning experiments from Mungall & Brenan (2014). A depleted mantle with 150 ppm S was selected for modeling, with a dissolved S content in melts of 1000 ppm. Under these conditions the magmatic sulfide in the residue are consumed by $F = 15\%$ (Luguet et al., 2003). Two starting compositions were used for the Pd/Ir ratio. One has primitive mantle Pd/Ir (solid line) versus Pd/Ir with CI-chondritic relative abundances (stippled lines for M&B and $D_{Pd} = 1000$). Primitive mantle normalization values from Becker et al. (2006).

Figure 10: Al_2O_3 variation diagrams for PM normalized (a) Os/Ir, (b) Ru/Ir, (c) Pt/Ir, (d) Pd/Ir, and (e) Re/Ir, and for (f) $^{187}\text{Os}/^{188}\text{Os}$ values measured by peak jumping of abyssal peridotites. PM compositions from Becker et al. (2006) and Meisel et al. (2001) shown in the figures and CI chondrite average values for HSE inter-element ratios are shown as horizontal lines in (a) to (e). Average $^{187}\text{Os}/^{188}\text{Os}$ values for carbonaceous (0.1262 ± 0.0005), enstatite (0.1280 ± 0.0008) and ordinary (0.1284 ± 0.0020) chondrites are shown as horizontal lines in (f), from the compilation of Day et al. (2016b). The stippled line in (f) shows the regression of data from this study, excluding PS59-206-10.

Figure 11: Time of Re depletion model ages (T_{RD}) versus Pd/Ir in abyssal peridotites. Primitive mantle Pd/Ir is from [Becker et al. \(2006\)](#). Literature sources of abyssal peridotites can be found in the *Supplementary Information*.

Figure 12: Osmium isotopic ($^{187}\text{Os}/^{188}\text{Os}$) composition versus $^{176}\text{Hf}/^{177}\text{Hf}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ for abyssal peridotites. Shown are the zero, 0.5 and 1 Ga T_{RD} isopleths. Osmium data are from this study and Liu et al. (2009). Hafnium and Nd data for the same samples are from Warren et al. (2009), and Stracke et al. (2011).

Figure 13: Relationship between Al_2O_3 and $^{186}\text{Os}/^{188}\text{Os}$ and $^{187}\text{Os}/^{188}\text{Os}$ in abyssal peridotites. R-squared values are from linear regression and 95% confidence limit regressions are shown. At the mean between depleted mantle Al_2O_3 compositions of ~4 wt. % to PM composition of ~4.5 wt. %, $^{187}\text{Os}/^{188}\text{Os}$ is predicted to be 0.1292 ± 0.0025 and $^{186}\text{Os}/^{188}\text{Os}$ is 0.1198388 ± 29 , from 95% confidence limit regressions. Also shown are recently published data for Lena Trough abyssal peridotites (filled diamonds) from [Chatterjee & Lassiter \(2016\)](#).

Figure 14: Estimates of the primitive mantle composition, here considered to be represented by abyssal peridotites (black line and labelled ‘PM*’), versus the primitive upper mantle (PUM – red line) estimate from [Becker et al. \(2006\)](#). Shaded region shows the 2 SD uncertainty on the calculations. Black dots show the Re and Os abundances of the upper mantle calculated from $^{187}\text{Os}/^{188}\text{Os}$ and $^{186}\text{Os}/^{188}\text{Os}$ systematics of abyssal peridotites.

Table 1: Highly siderophile element (ng g⁻¹) and Re-Os isotope compositions of abyssal peridotites and peridotite standards

Sample ID	Location	Lithology	Method	Al ₂ O ₃ (wt.%)	(Nd/Yb) _{ch}	LOI (wt.%)	Cu (μg g ⁻¹)	Os	Ir	Ru	Pt	Pd	Re	¹⁸⁷ Re/ ¹⁸⁶ Os	¹⁸⁷ Pu/ ¹⁸⁶ Os	¹⁸⁷ Os/ ¹⁸⁶ Os	2σ	T ₉₀
Van 7-85-27	SWIR	Harzburgite	CT	2.12	0.066	12.0	4.3	3.48	3.45	8.37	6.06	2.00	0.228	0.3162	0.1654	0.12472	0.00006	667
	SWIR	HPA				12.0	4.3	4.64	4.45	9.83	8.78	2.09	0.205	0.2132	0.1797	0.12497	0.00005	651
Van 7-85-30	SWIR	Harzburgite	CT	2.59	0.080	11.1	2.5	2.98	2.45	4.75	5.55	2.10	0.204	0.3305	0.1767	0.12680	0.00008	395
Van 7-85-49	SWIR	Lherzolite	CT	1.96	0.035	3.5	15.9	2.68	2.97	6.68	5.18	3.84	0.247	0.4430	0.1835	0.12701	0.00008	366
Van 7-96-16	SWIR	Pyroxenite	CT	10.5	0.313	4.8	4.2	0.32	0.39	1.19	0.86	1.07	0.011	0.1713	0.2581	0.14630	0.00013	-
Van 7-96-25	SWIR	Lherzolite	CT	2.67	1.51	13.8	19.3	2.15	1.95	5.92	3.31	1.32	0.100	0.2242	0.1457	0.12550	0.00009	578
Van 7-96-35	SWIR	Lherzolite	CT	1.67	2.46	16.2	28.0	1.97	1.92	5.27	4.07	1.08	0.164	0.4004	0.1961	0.12799	0.00008	228
KNOX11RR D-2-12	CIR	Harzburgite	CT	1.54	0.283	12.3	14.5	3.31	4.05	8.82	6.32	2.94	0.089	0.1300	0.1812	0.12468	0.00006	682
	CIR	HPA				12.3	14.5	3.67	4.05	8.62	6.25	2.06	0.086	0.1127	0.1614	0.12345	0.00005	864
KNOX11RR D-2-13	CIR	Harzburgite	CT	1.68	0.825	12.6	5.9	2.81	3.54	7.30	5.50	2.96	0.047	0.0815	0.1861	0.12175	0.00007	1100
	CIR	HPA				12.6	5.9	3.20	3.36	7.46	5.92	2.57	0.055	0.0826	0.1755	0.12283	0.00004	950
KNOX11RR D-2-18	CIR	Harzburgite	CT	1.55	1.497	12.4	6.8	2.27	3.10	5.87	6.36	1.89	0.031	0.0654	0.2662	0.12205	0.00005	1058
HLV0102-40-24	Gakkel	Harzburgite	CT	0.45	0.496	13.8	1.9	0.65	0.37	0.37	4.39	0.05	1.123	8.3508	0.6437	0.12698	0.00009	369
HLV0102-40-81	Gakkel	Harzburgite	CT	0.98	0.382	1.2	5.7	1.40	1.31	5.87	2.50	0.22	0.017	0.0588	0.1687	0.12321	0.00007	897
HLV0102-70-75	Gakkel	Lherzolite	CT	2.81	0.180	13.5	11.7	2.60	3.14	7.37	5.72	3.42	0.062	0.1152	0.2093	0.12695	0.00006	374
HLV0102-92-24	Gakkel	Harzburgite	CT	0.98	0.275	14.4	2.8	0.13	0.23	3.46	0.75	0.40	0.020	0.7398	0.5413	0.15669	0.00022	-
PS89-206-10	Gakkel	Lherzolite	CT	4.18	1.130	13.2	32.7	1.69	2.56	5.67	5.57	0.67	1.122	3.2040	0.3136	0.14456	0.00008	-
PS89-238-73	Gakkel	Lherzolite	CT	2.76	1.538	13.5	28.3	2.78	4.55	10.57	11.56	1.39	0.184	0.3190	0.3953	0.13140	0.00007	-
PS89-242-35	Gakkel	Lherzolite	CT	3.38	0.150	9.6	16.2	3.26	3.20	7.46	7.77	7.62	0.210	0.3098	0.2261	0.12808	0.00005	215
PS89-257-15	Gakkel	Lherzolite	CT	1.89	0.109	12.9	14.0	2.58	4.42	8.68	7.84	0.78	0.080	0.1488	0.2880	0.12291	0.00007	939
AP-2	MARK	Harzburgite	CT	1.64	0.017	13.1	14.4	6.61	6.51	12.53	9.46	9.35	0.613	0.4467	0.1358	0.12710	0.00010	353
AP-3 ^a	MARK	Harzburgite	HT CT	1.82		12.1		3.9	2.83	6.58	6.18	4.49	1.66			0.1246		
AP-4	MARK	Harzburgite	CT	1.21	0.012	13.5	17.1	3.34	3.95	6.16	5.17	3.32	0.202	0.2911	0.1466	0.12320	0.00010	899
AP-5	MARK	Harzburgite	CT	1.43	0.014	12.9	25.9	1.97	1.93	4.02	3.88	4.40	0.159	0.3881	0.1863	0.12610	0.00010	493
AP-6 ^a	MARK	HT CT	1.52		13.5			4.54	3.48	5.05	7.88	7.64	0.474			0.12626		
AP-7	MARK	Harzburgite	CT	1.45	0.013	13.6	20.2	3.61	3.28	7.78	4.18	7.19	0.213	0.2847	0.1101	0.12670	0.00010	409
HARZ-01		Harzburgite	CT					4.21	3.38	5.73	6.71	5.12	0.062			0.1254		
	2SD							1.30	1.36	2.19	2.69	1.60	0.023			0.0014		
GP13		Lherzolite	CT					3.37	3.38	6.90	6.78	5.50	0.286			0.12615		
	2SD							0.23	0.43	0.46	0.70	0.96	0.076			0.00062		
	n = 3		HPA					3.35	3.20	6.98	7.11	5.39	0.286			0.12669		
	2SD							0.10	0.13	0.25	1.42	0.58	0.001			0.00066		

a = Data from Becker et al. (2006). Abbreviations: SWIR = Southwest Indian Ridge; CIR = Central Indian Ridge; Gakkel = Gakkel Ridge; MARK = Mid-Atlantic Ridge south of the Kane Transform Fault; CT = Carius tube; HPA = High-pressure ashier; HT-CT = High-temperature (300 °C) CT. T₉₀ = Time of rhenium depletion age, calculated as: $1/\lambda \times \ln[(^{187}\text{Os}/^{186}\text{Os})_{\text{standard}} - ^{187}\text{Os}/^{186}\text{Os}_{\text{sample}}] / (^{187}\text{Re}/^{186}\text{Os}_{\text{standard}} + 1)$. Full standard data are provided in Table S2.

Table 2: High-precision Os isotope data for Abyssal Peridotites and Os-Ir-Ru alloys

Sample	Location	Lithology	n	$^{186}\text{Os}/^{188}\text{Os}$	2 σ	$^{186}\text{Os}/^{188}\text{Os}$	2 σ	$^{187}\text{Os}/^{188}\text{Os}$	2 σ
Abyssal Peridotites									
Van 7-85-27	SWIR, Oblique	Harzburgite	3	0.0013126	0.0000003	0.1198344	0.0000024	0.1254481	0.0000015
Van 7-85-30	SWIR Oblique	Harzburgite	2	0.0013242	0.0000046	0.1198363	0.0000004	0.1266180	0.0000007
Van 7-85-49	SWIR Oblique	Lherzolite	1	0.0013087	0.0000151	0.1198389	0.0000027	0.1269288	0.0000010
Van 7-96-25	SWIR Oblique	Lherzolite	1	0.0013539	0.0000078	0.1198376	0.0000036	0.1279704	0.0000060
Van 7-96-35	SWIR Oblique	Lherzolite	1	0.0013082	0.0000131	0.1198351	0.0000035	0.1276044	0.0000068
KNOX11RR D-2-12	CIR	Harzburgite	3	0.0013039	0.0000023	0.1198349	0.0000009	0.1245508	0.0000227
KNOX11RR D-2-13	CIR	Harzburgite	4	0.0013094	0.0000051	0.1198358	0.0000015	0.1217605	0.0000684
KNOX11RR D-2-18	CIR	Harzburgite	4	0.0013100	0.0000065	0.1198363	0.0000035	0.1221254	0.0000245
HLY0102-70-75	Gakkel Ridge, EVZ	Lherzolite	1	0.0012959	0.0000093	0.1198391	0.0000044	0.1272704	0.0000096
PS59-238-73	Gakkel Ridge	Lherzolite	2	0.0013124	0.0000010	0.1198336	0.0000021	0.1322697	0.0000077
PS59-242-35	Gakkel Ridge	Lherzolite	2	0.0013128	0.0000066	0.1198371	0.0000038	0.1282778	0.0000026
PS59-257-15	Gakkel Ridge, SMZ	Lherzolite	1	0.0013166	0.0000050	0.1198348	0.0000025	0.1235735	0.0000042
AP-1	MARK ^a	Harzburgite	3	0.0013143	0.0000038	0.1198357	0.0000034	0.1266505	0.0000079
AP-1	MARK ^a	Harzburgite	3	0.0013178	0.0000039	0.1198363	0.0000014	0.1275760	0.0000017
AP-3	MARK ^a	Harzburgite	3	0.0013209	0.0000030	0.1198341	0.0000049	0.1250446	0.0000029
AP-4	MARK ^a	Harzburgite	1	0.0013206	0.0000050	0.1198339	0.0000062	0.1226477	0.0000080
AP-5	MARK ^a	Harzburgite	3	0.0013122	0.0000048	0.1198344	0.0000036	0.1260936	0.0000053
AP-6	MARK ^a	Harzburgite	3	0.0013191	0.0000025	0.1198347	0.0000032	0.1271363	0.0000062
AP-7	MARK ^a	Harzburgite	4	0.0013292	0.0000081	0.1198337	0.0000041	0.1266874	0.0000399
Os-Ir-Ru Alloy Grains									
3 Savage River	W USA	Os-alloy	3	0.0013058	0.0000037	0.1198365	0.0000015	0.1199643	0.0000017
5 Bald Hills	W USA	Os-alloy	3	0.0013030	0.0000007	0.1198362	0.0000015	0.1200609	0.0000043
California #2	W USA	Os-alloy	3	0.0013059	0.0000056	0.1198374	0.0000026	0.1233518	0.0000042
4Os	W USA	Os-alloy	2	0.0013132	0.0000001	0.1198458	0.0000029	0.1069994	0.0000017
L1	W USA	Os-alloy	3	0.0013062	0.0000058	0.1198354	0.0000013	0.1222857	0.0000017
L1	W USA	Os-alloy	3	0.0013054	0.0000005	0.1198362	0.0000012	0.1227784	0.0000022
M6	W USA	Os-alloy	4	0.0013099	0.0000064	0.1198363	0.0000009	0.1228891	0.0000193
Os8	W USA	Os-alloy	2	0.0013046	0.0000012	0.1198365	0.0000020	0.1237957	0.0000021
#8 Oregon	W USA	Os-alloy	2	0.0013041	0.0000014	0.1198383	0.0000005	0.1238626	0.0000009
M1	W USA	Os-alloy	8	0.0013050	0.0000057	0.1198367	0.0000026	0.1243467	0.0000046
M2	W USA	Os-alloy	3	0.0013039	0.0000033	0.1198376	0.0000013	0.1245478	0.0000075
Os4a	W USA	Os-alloy	1	0.0013095	0.0000019	0.1198344	0.0000010	0.1261981	0.0000017
Os5	W USA	Os-alloy	2	0.0013104	0.0000016	0.1198293	0.0000004	0.1265035	0.0000028
Os3	W USA	Os-alloy	2	0.0013130	0.0000166	0.1198306	0.0000014	0.1266066	0.0000091
Os4	W USA	Os-alloy	3	0.0013040	0.0000025	0.1198295	0.0000025	0.1266509	0.0000035
Os1	W USA	Os-alloy	2	0.0013096	0.0000024	0.1198248	0.0000006	0.1276398	0.0000016
Os2	W USA	Os-alloy	2	0.0013075	0.0000006	0.1198290	0.0000070	0.1278899	0.0000036
8Os	W USA	Os-alloy	2	0.0013065	0.0000012	0.1198367	0.0000003	0.1301538	0.0000040
Os8	W USA	Os-alloy	2	0.0013068	0.0000003	0.1198363	0.0000008	0.1305158	0.0000031
L3	W USA	Os-alloy	4	0.0013046	0.0000050	0.1198384	0.0000024	0.1468688	0.0000097
W5	Unknown (BM)	Os-alloy	1	0.0012988	0.0000039	0.1198354	0.0000014	0.1242709	0.0000023
W5B	Unknown (BM)	Os-alloy	2	0.0013092	0.0000007	0.1198378	0.0000008	0.1242725	0.0000004
W5A	Unknown (BM)	Os-alloy	2	0.0013030	0.0000046	0.1198371	0.0000004	0.1242781	0.0000007
Finland	Finland	Os-alloy	4	0.0013063	0.0000066	0.1198344	0.0000018	0.1126372	0.0000084
9 Urals	Ural Mountains	Os-alloy	2	0.0013061	0.0000003	0.1198356	0.0000016	0.1198947	0.0000032
Yodda	Papua New Guinea	Os-alloy	2	0.0013034	0.0000026	0.1198385	0.0000001	0.1289971	0.0000061
Allende		CV3	2	0.0013062	0.0000035	0.1198345	0.0000016	0.1254616	0.0000031
OSUM8		Os-Alloy	30	0.0013052	0.0000031	0.1198390	0.0000005	0.1314615	0.0000023
UMCP		Standard	31	0.0013061	0.0000035	0.1198458	0.0000015	0.1137852	0.0000023

AP samples normalized to UMCP value measured in this study using measured $^{186}\text{Os}/^{188}\text{Os}$ (0.1198464 $\pm$ 12) and $^{187}\text{Os}/^{188}\text{Os}$ (0.1138067 $\pm$ 21). *from Brandon et al. (2000) - numbers in italics denote normalization relative to UMCP values measured in this study.

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