

**Oceanic and continental lithospheric mantle in the 1.95 Ga Jormua Ophiolite Complex,
Finland: implications for mantle and crustal evolution**

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ABSTRACT

The ca. 1.95 Ga Jormua Ophiolite Complex (JOC), Finland, is a rare Paleoproterozoic ophiolite that preserves a record of diverse upper mantle materials and melting processes. Meter-scale grid sampling of four JOC outcrops, as well as non-grid samples, permits evaluation of meter- to kilometer-scale mantle heterogeneity within the JOC. Significant heterogeneity is observed between the four grids, and also among a number of the non-grid samples examined. Variations in the concentrations of fluid-mobile elements are particularly large among different samples and locations. New whole-rock major, lithophile trace, and highly siderophile element data (HSE: Os, Ir, Ru, Pt, Pd, Re), including ^{187}Re - ^{187}Os isotopic data, for serpentinized harzburgites indicate the presence of two distinct compositional types and probable modes of origin within the JOC. This is consistent with prior findings. Type 1 is similar to modern refractory abyssal-type mantle. Type 2 is more highly refractory than Type 1, and most likely represents samples from sub-continental lithospheric mantle (SCLM). Type 1 mantle is moderately heterogeneous with respect to major and trace element and Os isotopic compositions at both the meter and kilometer scales. By contrast, Type 2 mantle is considerably more homogeneous than Type 1 grids at the meter scale, but is more heterogeneous at the kilometer scale. The median initial γOs value for Type 1 mantle, calculated for 1.95 Ga, is ~ -2.0 (where γOs is the % deviation in $^{187}\text{Os}/^{188}\text{Os}$ relative to a chondritic reference calculated for a specified time). This isotopic composition is consistent with a moderate, long-term decrease in Re/Os relative to the estimate for Primitive Mantle, prior to JOC formation. The similarity in this γOs value to the value for the modern abyssal mantle, as well as the initial values for several Phanerozoic ophiolites suggests that the upper mantle achieved a Re/Os ratio similar to the chondritic reference by ~ 2 Ga, then evolved along a subparallel trajectory to the chondritic reference since then. For this to occur, only

limited Re could have been permanently removed from the upper mantle since at least the time the JOC formed. A localized secondary metasomatic event at ~ 2 Ga, concurrent with the estimated obduction age for the JOC and subsequent Svecofennian Orogeny, affected the HSE systematics of some Type 1 samples. By contrast, late Archean Os T_{RD} model ages for Type 2 rocks indicate a depletion event superimposed upon the long-term Re depletion of the abyssal mantle. This event was established no later than ~ 2.6 Ga and may have occurred during a period of significant, well-documented crustal production in the Karelia craton at ~ 2.7 Ga.

INTRODUCTION

Documenting changes in the geochemical composition of the upper mantle over time and the relationship of these changes to crustal formation is key to understanding the origin and evolution of important Earth processes, most notably plate tectonics. Chemical and isotopic characterization of the mantle portions of ophiolites provide one way to study changes in the upper mantle through time. Ophiolites are defined as crust-mantle assemblages of oceanic lithosphere, preserved in continental crust. Ophiolite sections include oceanic sediments, pillow basalts, sheeted dikes, gabbros, and mantle peridotites (e.g., Anonymous, 1972). It is normally assumed that the mantle and crustal lithospheric materials incorporated in ophiolites were formed either directly or indirectly from the convecting upper mantle within several tens of Myr before the obduction processes that led to their incorporation and preservation in stable continental crust (Dewey & Casey, 2011).

Ophiolites form along convergent margins and are often assigned a minimum obduction age using the ages of cross-cutting igneous rocks and/or detrital zircons in associated metasedimentary rocks (e.g., Peltonen et al., 1998; 2003). The majority of known ophiolites are

Phanerozoic. Ophiolites in Proterozoic and Archean terranes are rare, underscoring their importance in preserving evidence for the development and progression of plate tectonic processes, including melt depletion of the convecting upper mantle, creation of oceanic lithosphere and eventual subduction of oceanic lithosphere (e.g., Kontinen, 1987; Helmstaedt & Scott, 1992; Scott et al., 1992; Stern et al., 1995; Peltonen et al., 1996, 1998, 2003, 2004; Şengör & Natal'in, 2004; Furnes et al., 2007; Kusky & Li, 2010; Kumar et al., 2010; Garde & Hollis, 2010; Dilek & Furnes, 2011; 2014; Furnes et al., 2014; Stepanova et al., 2014).

Ophiolite assemblages are typically altered as a result of processes associated with their obduction into stable continental settings. In particular, the ultramafic portions of ophiolite sequences commonly undergo variable degrees of serpentinization before, during and after obduction, although they may still retain some of their primary geochemical signatures (e.g., Deschamps et al., 2013; Malvoisin, 2015). The most noteworthy modifications concerns fluid-mobile element abundances (Day & Brown, 2021). Given the alteration history of such rocks, it is important to utilize geochemical tools capable of seeing through the effects of secondary modifications. Excluding gold, the highly siderophile elements (HSE: Re, Os, Ir, Ru, Rh, Pt, Pd, Au) have been shown to be resistant to certain secondary processes, such as serpentinization of mantle assemblages, while at the same time recording the timing and extent of primary mantle melting, and mantle metasomatic events (e.g., Snow et al., 2000; Pearson et al., 2003; Rudnick and Walker, 2009; Luguet and Pearson, 2019). It has also been shown that extraction of basaltic melts from the oceanic mantle has acted as a major control on the Re-Os isotopic evolution and relative abundances of HSE in the residual convecting upper mantle (e.g., Walker et al., 2002; Pearson et al., 2003; 2004). Consequently, Os isotopes and HSE abundances in the mantle portions of ophiolites can potentially be used to place constraints on the volumes of prior oceanic

crust removal and isolation from the upper mantle at various stages of Earth history (e.g., Walker et al., 2002; Walker, 2016).

Despite their importance, only a limited number of Precambrian ophiolites have been studied with respect to Os isotopes and the HSE. Kusky et al. (2007) reported Re-Os isotopic data for chromitites from the ~2.6 Ga Dongwanzi-Zunhua ophiolite, North China. All of the chromitites were characterized by very low Re/Os and initial $^{187}\text{Os}/^{188}\text{Os}$ consistent with chondritic reference growth models for the upper mantle. Walker et al. (1996) reported Re-Os isotopic data for chromite, gersdorffite and laurite separates from chromitite hosted in the mantle section of the ~1.97 Ga Outokumpu ophiolite. That study found initial Os isotopic compositions that were broadly similar to a chondritic reference. By contrast, Tsuru et al. (2000) reported Re-Os isotopic data for chromitites and whole-rock mantle samples from the *ca.* 1.95 Ga Jormua Ophiolite Complex (JOC), Finland, and found that some chromitites and associated ultramafic whole-rock samples were characterized by having substantially subchondritic $^{187}\text{Os}/^{188}\text{Os}$ at the time of ophiolite formation, consistent with melt depletion of the rocks well before the ophiolite formed. Tsuru et al. (2000) concluded that at least a portion of the mantle materials sampled by the JOC consisted of sub-continental lithospheric mantle (SCLM) that had undergone melt depletion at ~3 Ga, preceding the obduction age by ~1 Gyr. Tsuru et al. (2000) also suggested that some portions of the JOC may represent a second, mid-ocean ridge basalt (MORB)-like subtype with generally chondritic initial $^{187}\text{Os}/^{188}\text{Os}$ for the time of ophiolite formation. Ancient MOR-like harzburgites preserved in the JOC, if confirmed, would represent a rare example of directly accessible ancient oceanic mantle material.

Here we re-investigate the JOC because of its importance for documenting the isotopic and chemical composition of the upper mantle during the Proterozoic. We report whole-rock

major, minor and lithophile trace element compositions, as well as HSE compositions and new Re-Os isotopic data for serpentinized peridotitic rocks. One major goal of this study is to constrain the melt depletion history of the portions of the mantle sampled by the JOC, prior to ophiolite formation, and re-examine the earlier conclusion of the presence of both MOR-like mantle and SCLM. A second goal is to attempt to link whole-rock analyses of HSE abundances and Os isotopic compositions with lithophile major and trace element compositions, in order to improve understanding of the distribution and behavior of these elements in the upper mantle during the Proterozoic.

GEOLOGY

The Jormua Ophiolite Complex is located in the central part of the N-S-trending Kainuu belt, approximately 70 km from the southwestern margin of the Archean Karelia craton (**Figure 1**). In addition, a chain of small ultramafic lenses of probable ophiolitic origin runs southward to the Outokumpu area (south-southeast of the JOC) in the North Karelia belt (Peltonen, 2005; **Figure 1**), one of the Paleoproterozoic supracrustal belts in eastern and northern Finland (Laajoki, 2005). On both sides of the Kainuu belt, the Archean basement consists of gneissic and plutonic tonalite-trondhjemite-granodiorite series rocks, and minor greenstone relics. Svecofennian (1.86–1.80 Ga) granodiorite and granite plutons cut the Archean rocks and are particularly prevalent in the area between the Kainuu belt and the Svecofennian arc complex. Broadly coeval with emplacement of the JOC and granitoid intrusions, the Kainuu belt was also deformed and metamorphosed during the 1.8–2.0 Ga Svecofennian Orogeny, with associated metamorphic conditions reaching upper greenschist facies grade in the north, and amphibolite facies grade in the south.

The JOC represents the most complete Paleoproterozoic ophiolitic sequence recognized, covering an area of $\sim 90 \text{ km}^2$ (**Figure 2**). In contrast to other Precambrian ophiolitic complexes where crustal rocks dominate, e.g., the Payson Ophiolite, Arizona (Dann, 1997), approximately half of the lithology of the JOC comprises rocks that were originally mantle peridotites (Peltonen et al., 1996; 1998). Other rock types present in the JOC include gabbros, sheeted dikes, pillow lavas, and minor pelagic sediments and plagiogranites (leucotonalite-trondhjemite). The mantle rocks are cut by several generations of pyroxenitic and hornblenditic dikes (Peltonen et al., 1998). A few small chromitite pods bearing primary chromite have been documented, while disseminated chromites throughout the ophiolite are abundant, but mostly altered (Tsuru et al., 2000). In these cases the original mineralogy is extensively replaced by secondary minerals, although some chromites preserve primary cores. The ultramafic rocks mainly occur as serpentinites, which have locally been further altered to talc-carbonate rocks and, in extreme cases, to carbonate or quartz rocks (Peltonen et al., 1996; S  ntti et al. 2006). A lower crustal ultramafic-mafic cumulate series typical of Phanerozoic ophiolites (e.g., dunite transition zone, or Moho transition zone; Boudier & Nicolas, 1995) is lacking, as sheeted dikes and gabbroic bodies interleave with mantle rocks, sometimes separated by lens-like ultramafic septa. All intrusive mafic rocks and pillow lavas appear to be co-magmatic, having been derived from the same kind of enriched-mid ocean ridge basalt (E-MORB)-like parental magma (Peltonen et al., 1996). The timing of the mafic magmatism in the JOC is constrained by U-Pb zircon ages of 1952 ± 2 and 1950 ± 3 Ma obtained for a high-level gabbroic pod and trondhjemitic differentiate, respectively (Huhma et al., 2018).

During its emplacement and the later Svecofennian-age orogeny, the JOC was dismembered into four main tectonic blocks; the Hannusranta, Lehmivaara, Antinm  ki, and

Kannas blocks (**Figure 2**). Of the four blocks, the Antinmäki block (also known as the Eastern block; Peltonen et al., 1996; 1998) is the largest and best studied. It is the source of the majority of samples reported in this study, including data for three grids (see below). The block consists of serpentinized mantle peridotites and mantle-derived intrusive and extrusive rocks including gabbros, Fe-gabbros, sheeted dikes and pillow basalts. A gabbroic dike that cross-cuts mantle peridotite in this block yielded a U-Pb zircon age of 1953 ± 2 Ma (Peltonen et al., 1998); such dikes are interpreted to represent feeder channels for gabbroic intrusions in the upper part of the ophiolitic sequence.

The Lehmivaara (or Central) block also preserves serpentinized peridotite and dikes with ocean island basalt-like (OIB) geochemical affinities, as well as sheeted dike complexes, all of which were later cross-cut by E-MORB-like basalt dikes (Peltonen et al., 2003). Six serpentinized peridotite samples from one grid were analyzed from this block.

The Hannusranta (or Western) block is composed almost totally of serpentinized lherzolitic mantle rocks, with a paucity of E-MORB-like dikes and sheeted dikes (e.g., Peltonen et al., 1996). A characteristic feature of the Hannusranta block is the presence of coarse-grained clinopyroxene±amphibole and amphibole±garnet dikes and veins crystallized from an alkali basalt magma at a depth of 30-50 km, which host Archean (xenocrystic) zircons (Peltonen et al., 1998, 2003). No grid sampling was undertaken in the Hannusranta Block, however, two individual harzburgitic samples from this block were analyzed. The Kannas block (also known as the Northern block) was not sampled for this study, and so is not discussed further.

SAMPLES

Samples were collected either from drill cores or hand samples from bedrock exposures (**Figure 3**; see **Table S1** for metadata). The majority of samples were collected during a field campaign in 2015 (JU15 series), and the rest are whole-rock samples collected for and analyzed by Tsuru et al. (2000) (samples with “ATK” designation).

While some samples reported here were collected as individual specimens from various locations, samples from three 1.5 m × 1.5 m grids (Grid 1/JU15-16, Grid 2/JU15-18, and Grid 3/JU15-21), with nine subsamples from each grid, were collected from the Antinmäki block. Six subsamples were collected from Grid 4 (JU15-25 to JU15-30) in the Lehmivaara block (**Figure 2**). All of the ultramafic rock sampled have undergone significant alteration, including serpentinization before peak regional metamorphism, which produced the present antigorite-dominated mineral assemblage. Consequently, hereafter we refer to the various petrologic groups by their presumed protoliths, based mainly on major element chemical compositions with particular emphasis on Mg#. The majority of the mantle rocks originated as harzburgites and dunites, and are the primary focus of this study. Comparison of geochemical variations on the ~0.5 m scale, versus variations in samples collected more than 1 km apart (**Figure 2**) is used to provide a metric for length scales of chemical and isotopic heterogeneity in the Paleoproterozoic mantle.

In total, we report whole-rock data for 53 samples from the JOC, with 41 collected from the Antinmäki block, 7 from the Lehmivaara block, and 7 from the Hannusranta block (**Figure 2**). Of these, 33 samples comprise the three complete and one partial grids, to assess meter-scale mantle heterogeneity. The 33 grid samples are the main focus of this study; ten non-grid harzburgite samples were analyzed to provide additional regional context (**Table S1**; **S3**).

Several clinopyroxenite, hornblendite and plagiogranite samples were also collected and analyzed. Data for these rocks are provided in the **Electronic Supplement**.

ANALYTICAL METHODS

Sample Preparation

Samples were trimmed with a rock saw and polished with various grit sandpaper to remove tool marks. Pieces were then crushed in a ceramic jaw crusher and finely powdered in a ceramic *Shatterbox*. Aliquots from the powders were used for all major, lithophile trace, and HSE analyses, as well as Re-Os isotopic measurements.

Major Elements

Whole-rock major element data were obtained by X-ray fluorescence (XRF) analysis in two campaigns: one in 2015 (Haller, 2017) and a second in 2019. Data were collected using a *PANalytical Zetium* X-ray fluorescence vacuum spectrometer at Franklin and Marshall College, Lancaster, PA, following methods described in Boyd & Mertzman (1987) and Mertzman (2000). Loss on ignition (LOI) was determined by heating ~1 g sample powder at 950°C for approximately 90 minutes. The anhydrous sample powder was then mixed with lithium tetraborate, placed in a platinum crucible, and heated with a Meker burner until molten, whereupon the material was mixed for 10-12 minutes to ensure sample homogeneity. The molten sample was then quenched in a Pt casting dish. Repeated analyses of BHVO-2 yielded accuracies of <1.5% for most major elements (when comprising >0.5 wt.% of the anhydrous total). Minor elements (<0.5 wt.% of the anhydrous total) and the alkaline oxides had an accuracy of <3%. The data were standardized after methods outlined in Abbey (1983), Govindaraju (1994) and

Mertzman (2000). The typical reproducibilities ($\% 2\sigma$) of the analyses were 0.8% for SiO_2 , 1.3% for Al_2O_3 , 2.3% for $\text{Fe}_2\text{O}_3\text{T}$, 1.0% for MgO , and 0.6% for CaO , based on repeated analyses of BHVO-2.

Lithophile Trace Elements

Lithophile trace element data for whole-rock samples were collected in three campaigns at the Scripps Institution of Oceanography (SIO) using methods outlined in Day et al. (2014) and O'Driscoll et al. (2015, 2018). The earlier set (2016) includes the JU15-16 (Grid 1) and JU15-18 (Grid 2) grid samples, while all the other samples were analyzed in campaigns in 2019 and 2022. All samples were analyzed along with basalt and peridotite rock standards (BHVO-2, BCR-2, BIR-1a, HARZ-01), as well as total procedural blanks. For these analyses, ~100 mg of sample powder was digested in a 1:4 mixture of concentrated high purity HNO_3 :HF for >72 h at 150°C on a hotplate. After drying down and sequential additional dissolutions and dry downs with HNO_3 to break down fluorides, clear sample solutions were obtained. These solutions were diluted by a factor of 5000 in 2% HNO_3 and doped with a 1 ppb In solution to monitor instrumental drift. Sample and standard solutions were measured in standard mode on a *Thermo Scientific iCAP Qc* quadrupole inductively-coupled plasma mass spectrometer (ICP-MS) at SIO. The reproducibility (2σ) of laboratory standards was <5% for the majority of elements compared to recommended values, where available (**Table S2**). Reproducibility of Cs, and to a lesser extent Cr, Li, and Ba was lower than for the majority of trace elements reported. Per-element internal statistics for standards are also provided in **Table S2**, in order to assess analytical reproducibility of elements across a representative range of concentrations.

Highly Siderophile Element Abundances and Re-Os Isotopes

Highly siderophile element concentrations and Re-Os isotopic system ratios were determined at the University of Maryland (UMd). Sample powders were digested using the Carius tube method described in Shirey & Walker (1995). Approximately 1.5 g of sample powder was loaded into each tube with known quantities of ^{185}Re - ^{190}Os and HSE (^{191}Ir , ^{99}Ru , ^{194}Pt , ^{105}Pd) group spikes and ~4.5 mL of a 2:1 mixture of concentrated HNO_3 and concentrated HCl . Tubes were then sealed and heated for ~48 hours at ~260 °C. Replicates and a harzburgite standard (MUH-1) were included in the analyses. After Carius tube digestion, samples were transferred to centrifuge tubes pre-loaded with CCl_4 to extract Os from the sample solution. This extraction was repeated twice. Osmium was then back-extracted into concentrated HBr , dried down, and further purified using dichromate- HBr microdistillation. Sample Os separates were loaded onto Pt filaments along with a $\text{Ba}(\text{OH})_2$ activator solution. Osmium abundances and isotopic compositions were analyzed using a *ThermoFisher Triton* multicollector mass spectrometer, using an ion counter in peak-hopping mode. The remaining HSE were separated using a primary AG-1x8 100-200 mesh anion resin column. The Re-Ru and, where necessary, Ir-Pt fractions were purified using scaled down versions of the primary chemistry. Abundances of Re, Ir, Ru, Pt and Pd were determined by isotope dilution using either a *Nu Plasma* or *ThermoFisher Neptune Plus* multi-collector inductively-coupled plasma mass spectrometer. Total analytical blanks ($n = 9$; 2016 and 2020 campaigns) were 0.4-2.9 pg Os, ≤ 12 pg Ir, 6.8-69 pg Ru, 97-432 pg Pt, 0.3-32 pg Pd and 0.6-13 pg Re. The average percent internal uncertainty on individual $^{187}\text{Os}/^{188}\text{Os}$ harzburgite analyses is ≤ 0.1 % 2SE (Table S4; includes some data previously reported in Ginley, 2013; Haller, 2017). External reproducibility of $^{187}\text{Os}/^{188}\text{Os}$ of the UMCP Johnson-Matthey standard for the 2016

campaign was $\pm 0.2\%$ 2σ (average $0.1137 \pm 0.0002\ 2\sigma$; $n = 17$), and the 2019 campaign was $\pm 0.1\%$ 2σ (average $0.1138 \pm 0.0001\ 2\sigma$; $n = 7$).

RESULTS

Major and Lithophile Trace Element Abundances

Major and lithophile trace element compositions of 43 harzburgite and ten non-harzburgitic samples of various compositions are reported in **Tables 1** and **2**, respectively. Whole-rock major element compositions of harzburgites within each grid are characterized by only modest variations (**Figure 4A-F**). For example, within each grid, MgO typically varies by only ~ 2 wt.%. There is no evidence for systematic correlations between MgO and Al_2O_3 among the samples (**Figure 4A**), as would be expected if the abundances of both elements were dominantly controlled by variable extents of partial melting (e.g., Herzberg, 2004; O'Driscoll et al., 2012). With respect to most major and minor elements, compositional variations within each grid are less than the compositional variations between the grids, where MgO varies, for example, by more than 5 wt.% (e.g., **Figure 4**). The compositions of individual (non-grid) samples vary considerably and cover nearly the entire range defined by collective samples from the four grids.

When normalized to primitive mantle (PM; Sun & McDonough, 1989), Pb concentrations are elevated in all grids except for Grid 3, with the greatest enrichments found in Grids 1 and 2 (**Figure 5**). Of the high field strength elements (HFSE), Th, Zr, Hf, and Ti are depleted in the majority of the harzburgites, while Nb and Ta are PM-like to slightly enriched. By contrast, Nb/U is more variable than Nb/Yb, but broadly elevated relative to PM throughout the sample suite (**Figure 6**).

Primitive mantle-normalized rare earth element (REE) patterns are highly variable among samples from the four grids (**Figure 7**). Samples from Grid 1 define relatively flat, but highly depleted patterns, whereas samples from Grids 2, 3 and 4 have higher overall abundances of REE than Grid 1, and are light REE (LREE) enriched. Patterns for Grid 2 samples are characterized by small positive Eu anomalies, while samples from Grids 3 and 4 show strong negative Eu anomalies ($\text{Eu}/\text{Eu}^* = 0.07\text{--}0.53$; calculated as $\text{Eu}/\text{Eu}^* = \text{Eu}_N/\sqrt{\text{Sm}_N \cdot \text{Gd}_N}$ after Shields & Stille, 2001).

Highly Siderophile Element Abundances

Highly siderophile element compositions for JOC harzburgite samples vary substantially, although most fall within the range of compositions defined by modern abyssal peridotites (**Table 1; Figure 8A-D**). Most PM-normalized HSE patterns (PM values from Day et al., 2017) are characterized by variable extents of depletion in the incompatible HSE Pd and Re. The degree of HSE heterogeneity within a grid varies, with comparatively minor variability in Grids 3 and 4, and the greatest variability in Grids 1 and 2. The extent of heterogeneity among samples from a grid is best assessed through the standard error of the whole-grid median concentrations for each element (**Table 1**). Median values are discussed throughout this study due to their relative insensitivity to outlying data, and thus better representation of the typical harzburgite behaviors, in the small sample populations of the grid samples, compared to mean values. Both median and mean values for grid samples are provided in the data tables, along with their corresponding uncertainty estimates. Uncertainties of the median values are provided as standard errors (SE) of the median, which are, based on the asymptotic variance estimate of the median, assumed to be $\sim 1.253 \cdot (\text{SE}_{\text{mean}})$.

Grid 1 harzburgites are characterized by considerable variability in Ir, Ru, and Pd compared to other grids (**Fig. 8A**). This grid has the strongest overall depletion (normalized to PM, hereafter denoted by subscript “N”) in $\text{Re}_\text{N}/\text{Ir}_\text{N}$ of the samples analyzed, and $\text{Pd}_\text{N}/\text{Ir}_\text{N}$ varies from PM-like to strongly depleted (**Table 1**). While median Ir, Ru, and Pt concentrations (**Table 1**) in Grid 1 samples are similar to PM, individual samples within this grid show greater variation compared to the other grids. Most samples also record depleted $\text{Os}_\text{N}/\text{Ir}_\text{N}$ relative to PM, with the lowest median ratio of all grids.

Most Grid 2 samples are characterized by relatively flat, PM-like abundances of Os through Pt, but highly variable depletions in Pd (**Figure 8B**). Grid 2 patterns are more uniform than Grid 1, although Pd and Pt are variably depleted. Sample JU15-18 C2 is an outlier with pronounced Ru, Pt, and Pd depletions. Most samples from this grid have $\text{Os}_\text{N}/\text{Ir}_\text{N} < 1$, although they are less strongly depleted than Grid 1. Unlike the other grids, most Grid 2 samples have slightly elevated $\text{Re}_\text{N}/\text{Ir}_\text{N} \geq 1$, for all but one sample.

Grid 3 features relatively flat, PM-like HSE patterns in Os, Ir, and Ru, although several samples are characterized by modest depletions in Ir (**Figure 8C**). Since Ir partitions similarly to Os and to a lesser extent Ru, small depletions likely reflect involvement of phases that selectively incorporated Ir as well as Pt during depletion of some portions of JOC mantle (e.g. Pearson et al., 2004). By contrast, most samples from this grid are characterized by moderate depletions in Pt. Platinum can behave compatibly or incompatibly, depending on melting conditions, and for example, may be strongly fractionated in cratonic settings (Pearson et al., 2004). Palladium and especially Re are depleted in samples from this grid.

Grid 4 sample HSE patterns are characterized by remarkably uniform Os, Ir and Ru concentrations and very flat PM-normalized patterns (**Figure 8D**). Most Grid 4 patterns also

have consistently depleted Pt and Pd concentrations relative to PM, with the exception of PM-like Pt and Pd in JU15-29. Most samples are characterized by strong depletions in Re. Samples JU15-27 and JU15-30, however, have more PM-like Re concentrations.

The non-grid harzburgite samples record localized minor enrichments or depletions in HSE relative to PM, with similar overall magnitudes of Re and Pd depletion relative to Ir (**Figure 8E**). Most of these samples have HSE patterns and a range of variation more closely resembling Grids 1 and 2, than Grids 3 and 4, most notably lacking the strong Pt depletion of Grids 3 and 4.

Rhenium-Osmium Isotopic Systematics

Initial $^{187}\text{Os}/^{188}\text{Os}$ is calculated for the presumed time of obduction at 1.95 Ga, (Peltonen et al., 1996), the assumed closure age of the Re-Os isotope system within the ophiolite, and helps to assess the potential for ancient melt depletion events predating the obduction age (**Table 2**). Isotopic data for these rocks are also reported using the γOs_t notation, the percent deviation of the calculated $^{187}\text{Os}/^{188}\text{Os}$ of a sample relative to a chondritic reference at the time of interest (see Equation 1). The chondritic reference isotopic composition for 1.95 Ga is 0.11374, and $^{187}\text{Re}/^{188}\text{Os}_{\text{chondrite}} = 0.40186$ using the parameters of Shirey and Walker (1998).

$$\gamma\text{Os}_t = 100 * \left(\frac{\frac{^{187}\text{Os}}{^{188}\text{Os}}_{\text{sample}(t)}}{\frac{^{187}\text{Os}}{^{188}\text{Os}}_{\text{chondrite}(t)}} - 1 \right) \quad (1)$$

Harzburgites analyzed in this study have γOs_t values ranging from -10.7 to +8.3, with most values clustering between -5 and +2 (**Figure 10A**). The median value for all 43 harzburgites analyzed is -2.6 ± 0.7 (uncertainty reported throughout as 1SE of the median). The majority of samples from Grids 1 and 3 have relatively uniform, subchondritic γOs_t values that

largely vary by only a few percent. Grid 1 has a median γOs_t of -2.1 ± 0.3 ($n = 9$), while Grid 3 has the lowest median of the grids analyzed of -4.1 ± 0.4 ($n = 9$). Variability (1SE of the median) of values for Grids 2 and 4 is considerably greater. Grid 2 has a median γOs_t of -0.2 ± 1.6 ($n = 9$) that is similar to the chondritic reference. The greater variability may reflect the substantially higher $^{187}\text{Re}/^{188}\text{Os}$ ratios for samples from this grid compared to the others ($\sim 10\times$). This results in greater age corrections than for the lower Re/Os samples. Grid 4 is characterized by a γOs_t of -2.5 ± 1.9 ($n = 6$). The two samples with the highest $^{187}\text{Re}/^{188}\text{Os}$ ratios have the most negative γOs_t values, suggestive of comparatively recent Re addition or Os loss. If they are removed from the median calculation, the remaining four samples are characterized by a γOs_t of -1.8 ± 1.0 ($n = 4$).

Model Os melt extraction ages can be calculated using three methods: Mantle Extraction Ages (T_{MA}), Re-depletion Age type 1 (T_{RD1}) and Re-depletion Age type 2 (T_{RD2}) (Walker et al., 1989; McCoy-West et al., 2013; Rudnick & Walker, 2009; Luguet and Pearson, 2019). Details of these calculations are provided in the **Electronic Supplement**. These types of Os model ages are based on the assumption that the mantle material examined was affected by a single, significant melting event that removed some or even all Re from an originally fertile mantle peridotite composition whose $^{187}\text{Os}/^{188}\text{Os}$ ratio had been previously evolving along an average chondritic growth trajectory. For both types of T_{RD} model age calculations it is assumed that Re was largely removed from the mantle sample examined during a single melting event. As true quantitative Re extraction is unlikely except for high degrees of partial melting (e.g., Handler et al., 1997), T_{RD} model ages are normally presumed to place *minimum* age constraints on the timing of melt depletion of the upper mantle. Importantly here, the model ages are calculated for both the present isotopic composition (T_{RD1}), and a projected initial isotopic composition using the

measured $^{187}\text{Re}/^{188}\text{Os}$ (T_{RD2}). T_{MA} model ages, by contrast, are based on the assumption that the Re/Os of a peridotite has been maintained since the time of melt depletion. This type of model age calculation can be more accurate when partial melting, and therefore Re removal, was relatively modest. Secondary Re alteration as a function of melt-rock or water-rock interaction, however, becomes increasingly influential (and probable) in older rocks, resulting in over- or under-estimation of melting ages using the T_{MA} , for which growth trajectories are projected backward in time from the present measured $^{187}\text{Re}/^{188}\text{Os}$ to the chondritic evolution model.

All Os model age calculations require an assumption regarding the evolution of $^{187}\text{Os}/^{188}\text{Os}$ in the upper mantle. Choice of an evolutionary path is problematic for older rocks because Os isotopic data for upper mantle-derived rocks during the Proterozoic and Archean are rare. For this study, we utilized the parameters specified in Shirey and Walker (1998) in the model age calculations. These evolutionary parameters are broadly consistent with the evolution of the carbonaceous chondrite average, which by happenstance, appears to be broadly similar to the overall depletion history of the upper mantle, relative to estimates for the PM, which is more similar to the more radiogenic endmember compositions of ordinary and enstatite chondrites (e.g., Meisel et al., 2001; Walker 2016).

Median T_{MA} model ages for the four grids range from 2.2 to 2.8 Ga, excluding the two samples with high Re/Os (JU15-27 and JU15-30) from the calculation of the Grid 4 median. T_{RD1} model ages for grids 1, 3 and 4 range from 1.8 to 2.5 Ga. The majority of samples comprising these grids have significantly subchondritic $^{187}\text{Re}/^{188}\text{Os}$ (median <0.06 ; **Table 1**), consequently, most samples have similar T_{MA} and T_{RD1} model ages (**Figure 10C**). The median T_{RD1} model age calculation for Grid 2 gives a meaningless forward age of -0.5 Ga because the present-day $^{187}\text{Os}/^{188}\text{Os}$ ratios of most samples are higher than the chondritic reference. T_{RD2}

model ages for all four grids range from 1.9 to 2.7 Ga. Again, because of the low Re/Os ratio for most samples from Grids 1, 3 and 4, T_{RD2} model ages are similar to their respective T_{MA} and T_{RD2} model ages. The median T_{RD2} for Grid 2 is broadly similar to the T_{RD2} model age for Grid 1 (1.9 versus 2.2 Ga) because the calculation takes into account the comparatively high Re/Os ratios of the Grid 2 samples.

Non-grid samples are characterized by highly variable model T_{MA} , T_{RD1} and T_{RD2} ages that range from 1.0 to 3.7 Ga, although some give impossibly old (ATK96) or forward ages (JU15-1) because they have $^{187}\text{Re}/^{188}\text{Os}$ ratios that are similar to or substantially greater than the chondritic reference (**Table 1; Figure 10B**).

DISCUSSION

Geochemical Evidence for Two Mantle Types in the JOC

Despite pervasive secondary alteration, numerous previous studies have identified ophiolite harzburgites as potential representatives of mantle that has undergone variable extents of melt depletion (e.g., Bernstein et al., 1998), as well as retaining broad characteristics suggestive of tectonic provenance. Consequently, several ophiolite subtypes have been identified, including mid-ocean ridge (MOR), suprasubduction zone (SSZ), and continental margin (CM) types (e.g., Dilek and Furnes, 2011). Suprasubduction-type ophiolites are the most common, forming at convergent margins during trench rollback, and are interpreted to be associated with onset of major subduction-related geological events, such as ocean basin closure. Often, SSZ ophiolites preserve classical Penrose-type sequences, and sample highly melt-depleted, harzburgitic mantle. Volcanic sequences are typically MORB-like, arc-like, and/or boninitic (Dilek and Furnes, 2011, and references therein).

Mid-ocean ridge type ophiolites, like SSZ ophiolites, commonly display a Penrose structural sequence, but form in association with a MOR system, rather than solely a convergent margin. Geochemically, these ophiolites are associated with normal MORB (N-MORB) or E-MORB-like, with occasional examples of crustally contaminated MORB (C-MORB) present in the crustal portion of the ophiolite (Dilek and Furnes, 2011). Continental margin type ophiolites, which are associated with continental breakup and early stages of ocean basin opening, sample SCLM, which can lie in direct contact with overlying mafic igneous rocks, and may be cross-cut by dikes.

A major objective of this study is to reassess the tectonic provenance of different portions of the JOC, given that prior interpretations of JOC mantle lithologies have included E-MORB and ocean island basalt (OIB)-type source mantle (Peltonen et al., 1996), as well as SCLM (Tsuru et al., 2000; Peltonen and Kontinen, 2004). In the case of the confirmation of the presence of MOR-type mantle, compositional data can be used to improve constraints on the chemical and isotopic characteristics of abyssal mantle during the Paleoproterozoic. In the case of SCLM-type mantle, compositional and chronologic data can lead to an improved understanding of the nature and timing of continental lithospheric mantle growth, and possible relationship to the evolution of plate tectonics.

Chemical data alone cannot be used to distinguish MOR-type mantle from SCLM, as both types of mantle have undergone variable extents of melt depletion. Some samples of SCLM tend to show chemical evidence for more extreme melt depletion than is evidenced in abyssal mantle, and for certain samples, SCLM tends to be characterized by greater levels of metasomatic modification and/or re-melt enrichment than abyssal mantle (e.g., Zhang et al., 2008). Although chemical composition fields defined by the global databases for each overlap

(**Figure 4**), possible distinguishing characteristics are present in trace elements. As examples, serpentized abyssal peridotites tend to have substantially greater Pb enrichment than SCLM, perhaps related to the amount of seawater interacting with the two domains. Also, while SCLM is often depleted in middle and heavy rare earth elements (MREE and HREE) compared to light rare earth elements (LREE; e.g., Brenan et al., 1995; Li and Lee, 2006; Zhang et al., 2008; Kodolányi et al., 2012), abyssal mantle is often LREE-depleted (**Figure 5-6**), although metasomatized abyssal mantle may have LREE enrichment relative to MREE and HREE as a result of fluid-rock interaction (Vander Auwera and Andre, 1991). Further, for some locales, Os isotope model ages are a reliable indicator of the presence of SCLM, compared with abyssal mantle, given that SCLM is commonly characterized by substantially older Os model ages than abyssal mantle materials of the same presumed formation age. Hence, it was the Archean Os model ages for some JOC peridotites that led Tsuru et al. (2000) to conclude that major portions of the JOC represent SCLM that formed during the late Archean. As with prior studies, our new data support the interpretation of two types of harzburgite in the JOC. Henceforth, they are referred to as Type 1 and Type 2 mantle.

Type 1 Mantle (abyssal)

Type 1 mantle is characterized by comparatively low MgO (**Figure 4**), relatively unfractionated REE patterns, and HSE patterns consistent with moderate melt depletion, as evidenced by depletions in the incompatible HSE, Pd and Re, relative to PM (**Figure 8**). Type 1 as represented by Grids 1 and 2 is found mainly in the Antinmäki block, and represents most of the harzburgite samples in this study. The Al₂O₃ contents of Grids 1 and 2 are both strongly depleted, ranging between 1 to 2 wt.%. Grid 1 rocks have higher average MgO and lower average SiO₂ contents

than Grid 2 (**Figure 4A-C**), consistent with a more refractory residue. Otherwise, the two grids have similar major element compositions.

Most trace element concentrations in Grid 1 samples, including the overall depletions in REE (~10× depletion relative to PM) and the minor LREE depletion relative to HREE are similar to modern abyssal lithospheric mantle (**Figures 5A, 6A**), as are the large positive Pb anomalies. Strong depletions in Pd are present in most Type 1 rocks, as well as especially strong Re depletions for all samples from this grid (**Figure 8A**). Osmium, Ir, Ru and Pt concentrations all show considerable variations within the grid, indicating processes in addition to simple melt removal, such as metasomatic overprinting.

Lithophile trace element patterns for Grid 2 are, as with Grid 1, broadly consistent with abyssal mantle (**Figure 5B**). Unlike Grid 1, however, Grid 2 samples are characterized by minor LREE enrichment and only minor HREE depletion, relative to PM (**Fig. 6A**). Rare earth element patterns of samples from this grid are similar to one another, indicating that the volume of material sampled was modified by a process that affected the REE in a uniform manner. In contrast to Grid 1 samples, Grid 2 samples are characterized by negligible Re depletion compared to PM, but show a similar, broad range of Pd concentrations, which are moderately depleted relative to PM. These compositions suggest that Grid 2 mantle was initially melt-depleted, but that its Pd and Re budgets were subsequently modified. While Re can be fluid-mobile under oxidizing conditions (e.g., Xiong et al., 2006), co-occurrence of elevated Re (and Pd) with elevated LREE suggests that melt percolation probably controlled the re-enrichment of these elements. The retention of strong depletion in Al_2O_3 , however, indicates that the effects of the putative melt-rock interaction was largely limited to incompatible trace elements.

Importantly, melt percolation is also evidenced by the presence of pyroxenitic veins in the Grid 2 outcrop (**Figure 3**).

Additional information about the nature of the Type-1 mantle is revealed by HFSE systematics. As elements that are relatively insensitive to fractionation during partial melting and alteration, the relative abundances of Th, Nb, and Yb can be used to evaluate the enrichment of both source mantle and mantle-derived melts (Pearce, 2008). Many of the JOC harzburgites, including Grid 2 samples, generally plot along the Th-Yb-Nb mantle array, intermediate to modern N-MORB and E-MORB (**Figure 11**). Grid 1 samples, however, have Th abundances below the measurement detection limits (**Table 1**), and therefore, must plot well below the array. Given the highly incompatible nature of Th, its evident low abundances are likely associated with the process(es) that also led to the overall REE-depletion. Thorium mobility is unusual in metamorphic and subduction zone settings, but has been shown to occur along with U and REE loss (e.g., Ague, 2017), all of which are observed in Grid 1 samples.

Titanium and redox-sensitive V systematics (relative to Yb) provide constraints on the extent of melt depletion, depth of melting and oxidation state of a fertile MORB mantle assumed to have a Primitive Mantle starting composition (Pearce & Parkinson, 1993). Modeled compositional fields for enriched fertile and refractory MORB mantle compositions are shown in **Figure 12**. Grid 1 and 2 samples plot as distinct clusters on the V and Ti versus Yb plots, reflecting interpretations of differing metasomatic histories of the two sites. Grid 1 samples are more depleted in incompatible elements than Grid 2 samples, clustering near the refractory MORB mantle field, consistent with ~20% melt extraction, slightly in excess of QFM redox conditions in the spinel-to-spinel+garnet stability field. Grid 2 samples most closely resemble fertile MORB mantle, consistent with only 5-10% melt depletion, and slightly favoring QFM to

QFM-1 conditions that are more reduced than the other grids. The apparent minimal amount of melt extraction is unique to Grid 2, compared to the rest of the JOC. As noted earlier in this subsection, the protolith to Grid 2 samples appears to have been initially melt-depleted, and was likely subjected to metasomatism that resulted in variable re-enrichment of portions of the grid.

All Grid 1 samples have low $^{187}\text{Re}/^{188}\text{Os}$ ratios <0.06 , resulting in only minor age corrections for γOs_t . All samples record modestly negative initial γOs_t values with a median value of -2.0, similar to the median value for modern abyssal peridotites of -1.3 (Walker, 2016). Given the low $^{187}\text{Re}/^{188}\text{Os}$ ratios for these samples, T_{MA} , T_{RD1} and T_{RD2} model ages are similar, with median ages ranging only from 2.1 Ga (T_{RD1}) to 2.3 Ga (T_{MA} and T_{RD2}). Thus, model depletion ages are older than the age of the ophiolite by only ~ 0.1 to 0.3 Gyr.

The Re-Os characterization of Grid 2 samples is more problematic, given the evidence for the likely irregular, metasomatic addition of Re to the samples. The median initial γOs_t value for this suite of -0.2 is significantly higher than for Grid 1, with greater variability among individual samples (2σ of 1.6 and 7.5 for Grids 1 and 2, respectively). The respective median T_{MA} and T_{RD2} ages of 2.2 and 1.9 Ga for this grid, however, are similar to ages for Grid 1, and are consistent with melt depletion and subsequent metasomatic Re enrichment, soon after ophiolite formation. The median T_{RD2} age is a meaningless future age because the $^{187}\text{Re}/^{188}\text{Os}$ of most samples are greater than the chondritic reference.

In summary, Os model ages for Grids 1 and 2 indicate melt depletion histories only shortly before the time of ophiolite formation. The HFSE suggest significantly different degrees of partial melting are represented by Grids 1 ($\sim 20\%$) and 2 ($<10\%$). Grid 1 melting occurred under comparatively oxidizing conditions between QFM and QFM+1 while Grid 2 “melting” occurring under somewhat less oxidizing conditions between QFM and QFM-1, although this

composition is more likely a result of metasomatic re-enrichment of the Grid 2 protolith rather than minor melt extraction. Both grid suites appear to have been modified by secondary processes. In the case of Grid 1, both the low REE and Th abundances, and variability in incompatible HSE (Os, Ir and Ru) are evidence for likely metasomatic processes. In the case of Grid 2, LREE enrichment and variability in Pt concentrations are also indicative of metasomatic processes, albeit different from those that affected Grid 1. Overall, the chemical and especially Os isotopic characteristics of the Antinmäki block Grids 1 and 2 are most consistent with Paleoproterozoic abyssal lithospheric mantle.

Type 2 Mantle (SCLM)

Type 2 is represented by samples from Grids 3 and Grid 4. Grid 3 has both the highest MgO and Al₂O₃ (on an anhydrous basis) of the four grids, as well as the highest LOI and P₂O₅ contents (**Figure 4A, 4C**). Elevated volatiles and phosphate, while otherwise retaining major and trace element characteristics of harzburgite, may indicate secondary alteration involving carbonate-phosphate phases of the ultramafic protolith, such as has been described by Sääntti et al. (2006). Secondary alteration is also consistent with a broader pattern of elevated LOI at high MgO contents in peridotites (see **Figure 4D** inset). Grid 4 samples, in contrast to Grid 3, are intermediate in various major element compositions to Type 1 Grids 1 and 2.

Lithophile trace element concentrations of both Grids 3 and 4 are more enriched than would be expected for refractory mantle, particularly considering that LREE abundances of the two grids are slightly enriched compared to PM, when REE depletion is more typical in refractory compositions. This enrichment is distinct from LREE-enriched Grid 2 harzburgites (**Figures 5-6**). The sinusoidal REE patterns of Grids 3 and 4 feature convex, slightly enriched

LREE (relative to PM), and concave, depleted MREE-HREE. The Type 2 REE patterns also feature well-developed negative Eu/Eu* anomalies that indicate that Type 2 mantle depletion occurred under more oxidizing conditions not present in Type 1 harzburgites.

Type 2 mantle is distinct from Type 1 in some, but not all, HFSE behavior (**Figures 11, 12**). In the Th-Yb-Nb plot (**Figure 11**), Grid 3 plots within the MORB-OIB mantle array, while Grid 4 samples plot below the mantle array. Unlike in the Type 1 case, Grid 4 Th and Yb abundances are comparable to Grid 3. Grid 4 samples, however, are more Nb-rich than Grid 3 (**Figures 5D, 7**), driving the shift away from the mantle array in **Figure 11**. Grid 3 and 4 samples exhibit evidence of 5 to 15% melt depletion along a Ti/Yb trajectory between the spinel and garnet stability fields (**Figure 12A**). This is also consistent with minor retention of HREE relative to MREE evident in Type 2 REE patterns, and is in contrast to Grids 1 and 2, which follow a Ti/Yb depletion trajectory consistent with melting in the spinel stability field. In the V-Yb diagram, Grids 3 and 4 plot in a scattered array with depletion levels consistent with variable Ti-Yb depletion (**Figure 12B**). The V-Yb data suggest that Grid 4 is slightly more depleted than Grid 3, and that both grids are considerably more oxidized than Type 1 harzburgites, with samples plotting in excess of QFM+1, consistent with the well-developed negative Eu anomalies in REE patterns for both grids.

The generally uniform HSE patterns for Grids 3 and 4 are depleted in Pt (**Figure 8**) consistent with fractionation patterns observed in cratonic peridotites (Pearson et al., 2004), and thus SCLM protoliths for Type 2 mantle. Some Grid 3 samples, where the most depleted γ Os in this study was found, also show depletions in Ir relative to Os and Ru. The uniformity and Pt $\pm$ Ir depletion found in the HSE patterns of the Type 2 grids are potentially controlled by a minor phase not present in Type 1 mantle.

Grid 3 has a low $^{187}\text{Re}/^{188}\text{Os}$ median of 0.028, which, as with Grid 1, results in only minor age corrections for calculating initial γOs_t values and model ages. Grid 3 has the most strongly subchondritic median initial γOs_t value of -4.1, compared to the other JOC grids. This median value is slightly higher than the average γOs_t value of -5.1 previously reported for JOC chromites by Tsuru et al. (2000). As with Grid 1, the low $^{187}\text{Re}/^{188}\text{Os}$ ratios for these samples result in similar T_{MA} , T_{RD1} and T_{RD2} model ages ranging from 2.5 Ga (T_{RD1}) to 2.7 Ga (T_{MA}). Thus, the median model depletion ages for this grid is ~0.4 Gyr older than the median ages determined for Grids 1 and 2 of the Type 1 harzburgites.

The median initial γOs_t value for Grid 4 of -2.5 is intermediate between the higher values of Grids 1 and 2, and the lower value for Grid 3, albeit with considerable isotopic heterogeneity. Two samples from this grid have $\text{Re}_\text{N}/\text{Pd}_\text{N} > 1$ and correspondingly high $^{187}\text{Re}/^{188}\text{Os}$, requiring greater age correction than for the other grid samples, and these two samples (JU15-27 and JU15-30) also have the lowest initial γOs_t values (-10.1 and -4.6, respectively). The scatter among initial γOs_t values within the small area sampled, coupled with the presence of pyroxenitic veins in the grid outcrop (see **Figure 3**) suggests that there may have been highly localized (sub-meter scale) Re mobility at the sample site, sufficiently long after the main Re depletion event that single-stage age correction may not fully account for ^{187}Os ingrowth. Nevertheless, median T_{MA} and T_{RD2} model ages are in good agreement at 2.5 Ga and similar to Grid 3.

Non-Grid Samples

The non-grid samples represent a mixture of Type 1, including both LREE-enriched and REE-depleted examples, and Type 2 harzburgite (see **Table 2** for list of samples by category),

distinguished mainly by trace element characteristics (REE patterns, Eu/Eu*, and to a lesser extent, the presence of a positive Pb anomaly in extended trace element plots, as seen in Figure 5A-B). Type 1 LREE-enriched are the most common non-grid harzburgite type (n=5). Extended trace element patterns have large positive Pb anomalies (~10x Ce), and nearly PM-like, slightly LREE-enriched REE patterns with minor positive to negative Eu/Eu* anomalies, closely resembling Grid 2 (**Figures 5-6**). Two of the non-grid harzburgites considered to be Type 1 are more similar to Grid 1 samples, with large positive Pb anomalies (**Figure 5**), and REE-depleted patterns. Collectively, all of the non-grid Type 1 samples have T_{RD2} ages ranging from 1.6-2.4 Ga, with a median value of 1.9 Ga, and γOs_t values ranging from -10.7 to +1.8, with a median of +0.5. These samples are overall more radiogenic than Grid 1, more closely resembling Grid 2.

Three of the non-grid harzburgites belong to Type 2 (resembling Grids 3 and 4). For these samples, Pb enrichment is more muted (resembling Grid 4) than the other non-grid harzburgites, and while REE abundances vary, pattern shapes resemble the sinusoidal Grid 3 and 4 patterns, with well-developed negative Eu/Eu*. Unlike with the grid samples, the non-grid Type 2 samples only show small Pt depletions (**Figure 13A**). Two of the samples (ATK59 and ATK589) have negative γOs_t (both yielding values of -2.7) that give identical T_{RD2} ages of 2.4 Ga, which closely resemble Re-Os systematics of Grid 4 (median γOs_t of -2.5, and T_{RD2} of 2.4 Ga). The third sample (JU15-1) has high $^{187}Re/^{188}Os$ and was duplicated. Like most other samples reported here, the three model ages disagree for both replicates, and overall the sample replicated poorly (T_{RD2} of 2.4 Ga and 4.6 Ga for the duplicates; the first of which is identical to ATK59 and ATK589). Based on the elevated Re concentration and more radiogenic Os, the duplicate analysis may have included a small amount of material that had been subjected to Re

contamination significantly later than obduction, and its calculated initial $^{187}\text{Os}/^{188}\text{Os}$ should be discounted.

In summary, incompatible lithophile trace elements are more fractionated from PM values in rocks from Grids 3 and 4 compared to the rocks of Grids 1 and 2, e.g., LREE enrichment and strong negative Eu anomalies. They also show evidence for having formed or been modified in a more oxidizing environment. Highly siderophile element patterns, however, are indistinguishable from modern abyssal peridotites. Despite this, both Grids 3 and 4 are characterized by T_{MA} and T_{RD2} model ages that are ~400-500 Myr older than ophiolite formation. We conclude that these grid locales in the Antinmäki and Lehmivaara blocks sample SCLM that likely formed, at latest, at the end of the Archean.

Mantle Heterogeneity in the JOC

Overall, geochemical variability in the JOC at the meter scale is similar to that at the kilometer scale within the Type 1 harzburgites (Grids 1 and 2). By contrast, Type 2 grids (Grids 3 and 4) have distinctive, more uniform HSE behavior at the meter scale, suggesting that their protolith mantle was more homogeneous than Type 1 mantle. Most of the variance in Type 2 harzburgites is outlier-controlled, indicative of highly localized heterogeneities. Type 1 mantle is more common than Type 2 mantle harzburgites in our sample set, although this sampling is not statistically representative of the mantle in the JOC. The new data, however, lend some insight to the distribution of abyssal and SCLM domains within the JOC (**Figure 2**).

Within the Antinmäki Block, which is the best-characterized JOC block, both Type 1 and 2 harzburgites are present. Type 2 mantle is found in the central-western, northeastern, and southeastern margins of the lherzolitic portion of the block; all are separated by at least 2 km.

Two of these sites are located within several tens to hundreds of meters of Type 1 harzburgites. In both of these examples, LREE-enriched Type 1 harzburgites (e.g., resembling Grid 2) occur near Type 2 domains. The REE-depleted Type 1 harzburgites (resembling Grid 1) described in this study are all from the southeast corner of the Antinmäki Block and sit further from known outcrops of the Type 2 domain. The transition between LREE-enriched Type 1 mantle and Type 2 mantle appears to be abrupt, occurring within tens to hundreds of meters.

The harzburgites from the Lehmivaara and Hannusranta Blocks are all Type 2 mantle, but represent only limited sampling of both blocks compared to the Antinmäki samples. Hence, additional characterization of rocks from these blocks will be necessary to more meaningfully assess the distributions of mantle type.

The juxtaposition of Type 1 and Type 2 mantle within the ophiolite may be explained as a function of the tectonic history of the Kainuu belt, which hosts the JOC and related ophiolitic and ultramafic complexes. Prior studies interpreted the JOC as a “passive margin” ophiolite (Kontinen, 1987; Peltonen et al., 1996; 1998), where lithologies would be expected to transition between those characteristic of the oceanic versus continental (or cratonic) lithosphere, hence accounting for the presence of both mantle types throughout the JOC. After being “detached” from the ocean basin during early stages of continental breakup, the JOC was likely dismembered during the initial thrusting onto the margin of the Karelian craton (Peltonen and Kontinen, 2004). The presence of Type 1 mantle in a Paleoproterozoic ophiolite provides a rare opportunity to examine the Os isotopic composition of a portion of the ancient DMM.

Osmium isotopic composition of the abyssal mantle during the Proterozoic

Tsuru et al. (2000) reported a γOs_t value of ~ -5.1 for what is defined here as Type 2 JOC mantle. This is $\sim 1\%$ lower than the estimate of ~ -4 reported here for Grid 3. The value reported by Tsuru et al. (2000) was largely based on data for chromitites and chromites separated from serpentinites (originally harzburgites). Tsuru et al. (2000) also analyzed whole-rock serpentinites from which the chromites were separated, averaging a value of γOs_t of ~ -4 that is virtually identical to the Grid 3 median value. The difference in isotopic compositions between chromite and whole rock is puzzling. Chromite was originally viewed as the ideal recorder of mantle Os isotopic composition in ophiolite peridotites because of the high abundance of Os and low Re/Os typical of such chromites (e.g., Walker et al., 2002). It has been subsequently been argued, however, that chromite present in mantle assemblages is more likely to concentrate Os that is *more* radiogenic than is representative of the mantle domain within which it forms (O'Driscoll et al., 2012; 2015). This is because the Os that is concentrated is most likely derived from intergrain sulfides, which have been shown to have higher Re/Os than the bulk rock in which they originally resided and thus, become more radiogenic (Alard et al., 2005). The offset between the previously reported chromitite compositions in Type 2 JOC and whole-rock analyses cannot be easily explained and highlights the likelihood that the average model depletion age for the bulk samples here may serve as only minimum ages, and that the creation of the SCLM may have predated the average model age by several 100 Myr.

Tsuru et al. (2000) proposed that MORB-like (abyssal) mantle was also present in the JOC. This observation is confirmed here in the presence of Type 1 rocks. Our new data show that at least part of the Type 1 mantle (REE-depleted; e.g., Grid 1) records the time-integrated removal of Re from the upper mantle, relative to the PM, that may have begun during early Earth history, perhaps related to one or more episodes of significant crustal growth (e.g., Pearson et al.,

2007). Examples of Type 1 LREE-enriched mantle (e.g., Grid 2; **Table 2**) shows evidence of later metasomatic overprinting and Re addition not typical of the REE-depleted group, which resulted in considerable scatter in the Os isotopic composition of these samples compared to Type 1 REE-depleted and Type 2 samples (**Figure 10**). Despite the scatter, the median γ_{Os_t} for all Type 1 LREE-enriched samples ($+0.2 \pm 1.3$) is indistinguishable from the chondritic reference, consistent with Re mobility concurrent with JOC obduction. Whether or not the Type 1 LREE-enriched samples are a faithful record of Os isotopic compositions in the DMM at 1.95 Ga, or partially a record of metasomatic processes, merits further study.

Considering only Type 1 samples with good agreement between T_{MA} , T_{RD1} , and T_{RD2} model Re depletion ages (**Figure 10**, e.g. within variance of 0.3 Ga, after Rudnick & Walker, 2009), Type 1 mantle has a median γ_{Os_t} of -2.1 ± 0.9 . Including all Type 1 samples shifts the median γ_{Os_t} only slightly to -2.0 ± 0.8 , suggesting the median estimate is robust to outliers. This median value is nearly identical to the Snow and Reisberg (1995) estimate for modern DMM of -1.9 ± 1.1 based on unaltered examples of modern abyssal peridotite. Subsequent estimates of γ_{Os} for the modern DMM, using much larger datasets, have minimally changed this estimate, with γ_{Os} values ranging only from -2.0 (Lassiter et al., 2014) to -1.3 (Walker, 2016). Assuming JOC Type 1 mantle is representative of DMM at 1.95 Ga, the JOC median suggests that the offset of the Os isotopic composition of early Proterozoic DMM relative to the chondritic reference is nearly identical to that for the modern DMM (**Figure 14**).

Secular and/or spatial changes in the Os isotopic composition of the DMM are apparent when considering a global and temporal distribution of harzburgite and chromite samples from ophiolite complexes that have been previously characterized for Re-Os isotopic systematics (**Figure 14; Table 3**). Median γ_{Os_t} values are determined from samples not significantly

affected by high Re abundance ($\text{Re}_\text{N}/\text{Pd}_\text{N} > 1$). The median Os isotopic composition of Type 1 JOC mantle ($\gamma_\text{Os} = -2.0$) resembles that of many modern ophiolites in addition to the Snow & Reisberg (1995) estimate ($\gamma_\text{Os} = -1.9$) for modern DMM. Given that negative γ_Os values in abyssal mantle, relative to the chondritic reference and PM, at the time of obduction is assumed to reflect prior Re removal (via melt extraction) from the source, at least part of the JOC Type 1 mantle must have been depleted in Re no later than the Archean-Proterozoic boundary (e.g., Grid 1).

In a simple model of progressive extraction of crust (Re) from abyssal mantle over time, abyssal mantle would be expected to evolve to Os isotopic compositions that increasingly grows to less radiogenic compositions than the evolution trajectory of the PM. Although the availability of Os isotopic data for samples of abyssal mantle through time are limited, most ophiolites appear to show some level of long-term Re depletion relative to PM (**Fig. 14**). The γ_Os_t value of the abyssal mantle portion of the JOC of -2.0 is noteworthy in that it is similar to the value for the modern DMM. If representative of the DMM, this suggests that the level of Re depletion in it was nearly complete by ~2.0 Ga, and that the Os isotopic composition of the DMM has evolved along a trajectory that is parallel to the chondritic reference of Shirey and Walker (1998). This in turn suggests the $^{187}\text{Re}/^{188}\text{Os}$ of the DMM has been nearly identical to the chondritic reference ratio of 0.402 (Shirey and Walker, 1998), indicating that significant Re extraction had occurred by the early Proterozoic in at least some portions of the upper mantle reservoir, providing some of the earliest whole-rock evidence of ancient melt extraction.

Unlike peridotites from most younger, similarly well-documented ophiolites, the majority of JOC harzburgites have Re-Os model ages that predate obduction. Model ages for three of the four grid samples further demonstrate good internal agreement. By contrast, only sparse

examples of similarly ancient Re depletion remains in the modern upper mantle, largely confined to single chromites, sulfides, and/or Os-Ir alloys (e.g., Harvey et al., 2006; Liu et al., 2022), although some rare examples of ancient melt events have been found in modern whole-rock abyssal peridotite samples (Liu et al., 2008). These isolated examples from Phanerozoic peridotites tend to reproduce poorly, which is often attributed to heterogeneous Os-Ir-rich phases hosted in mantle rocks. It is possible that the whole-rock “memory” of ancient crustal building was mostly erased from the convecting upper mantle as a function of crustal recycling and younger crustal building events between the early Proterozoic and modern mantle, as suggested by Meisel et al. (2001). Further, the budgets of isolated recycled crust may not be constant, or homogeneously distributed over the temporal and spatial extent of the upper mantle, evidenced by the apparent shift in median Os isotopic composition of the DMM towards higher γ_{Os} values, implying higher Re/Os, in late Proterozoic to early Cambrian ophiolites before returning to lower Re/Os and a γ_{Os} value more like modern DMM (Figure 14; e.g., Haller et al., 2021; Paquet et al., 2022). While the spatial extent and contributions of fertile and depleted DMM domains through time remains ambiguous, the new JOC data, combined with information from younger ophiolites, reveals potential secular changes in the Os isotopic composition of the DMM. Thus, the Re-Os isotopic record of Type 1 JOC mantle confirms findings of ancient Re depletion, and together with the record obtained from younger ophiolites, suggests that siderophile cycling in the convecting mantle may be more complex and dynamic than previously thought.

CONCLUSIONS

The 1.95 Ga Jormua Ophiolite Complex, one of the oldest known complete ophiolite sequences, preserves a heterogeneous mixture of serpentinized mantle domains, likely consisting of both

abyssal- and SCLM-type mantle and associated mantle-derived melts. The harzburgites examined herein have refractory major element, HSE and Re-Os systematics consistent with a major mantle melting event that affected the majority of the protolith mantle that occurred at a minimum age of ~ 2.6 - 2.5 Ga, placing the formation of the majority of the mantle material in the JOC at the Archean-Proterozoic transition. We further identify the presence of abyssal harzburgite that existed no later than the Archean-Proterozoic transition, making the JOC a host of rarely-preserved ancient oceanic mantle material in addition to previously documented Archean SCLM.

Localized evidence of a second melting event ~ 2 Ga is present in a minor portion of the samples characterized in this study, consistent with prior age determinations for mantle dikes that cross-cut portions of the ophiolite, related to the obduction process and following Svecofennian orogeny and regional metamorphism. Trace element systematics indicate widespread, but highly variable, melt-rock and fluid-rock re-enrichment processes, which is also reflected as elevated Re in some of the harzburgite samples.

Grid sampling records HSE compositions more uniform at the meter scale than at the kilometer scale in three of the four grid sites within the JOC. The exception is Grid 1, which also records the least re-enrichment of the grid sites; most other locales exhibit highly localized (sub-meter scale) evidence of likely melt-rock interaction that re-enriched REE budgets and possibly Re. Overall, length-scale variations best resemble those found in the 161 Ma Coast Range Ophiolite in California, but contrast with the 497 Ma Leka and 492 Ma Shetland ophiolites, which have better developed meter-scale heterogeneities and evidence of less extensive melt depletion.

The presence of Type 1 (abyssal) mantle in the JOC provides an estimate of the DMM at ~ 1.95 Ga, showing that at least some portion of the Paleoproterozoic upper mantle was depleted

relative to a chondritic reference. Compared to modern DMM sampled by younger ophiolites, JOC abyssal mantle underwent a similar extent of depletion. By contrast, a cluster of ophiolites formed during and immediately following the Proterozoic-Cambrian transition are less melt-depleted, resembling a chondritic reference, suggesting that the osmium isotopic composition of the DMM is heterogeneous throughout its history or underwent long-term compositional shifts related to depletion and re-enrichment, or a combination of the two.

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DATA AVAILABILITY

The data reported in this article are available in this article and in its supplementary material.

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FIGURE CAPTIONS

Figure 1. Location of the Jormua Ophiolite complex in the Kainuu belt. The dashed line shows the inferred position of the S-W margin of the Archean Karelia craton. The Jormua Ophiolite Complex is labeled on the map as “JOC”, and the Outokumpu Ophiolite Complex is labeled as “OOC.”

Figure 2. Geological map of the Jormua area (modified after Kontinen, 1998) with sample locations. HaB = Hannusranta block, LeB = Lehmivaara block, KaB = Kannas block, AnB = Antinmäki block. Type 1a corresponds to Type 1 REE depleted harzburgite; Type 1b corresponds to Type 1 LREE-enriched harzburgite (see **Table 2**). See **Tables S1 and S3** for sample information marked with letters “a” through “b”.

Figure 3. Field photos showing the grid sample outcrops and sample sites. Locations of each grid are shown in **Figure 2**. Note the presence of ultramafic veins in the Grid 2 outcrop.

Figure 4. Major element (anhydrous) variations in the JOC harzburgites. Background data for oceanic lithosphere (closed green circles) and cratonic lithosphere (open black circles) are from a compilation provided by Jingao Liu (see **Electronic Supplement** for data sources). **A.** Al_2O_3 versus MgO . **B.** TiO_2 versus MgO . Most of the Grid 1 TiO_2 contents are below detection limits and are not plotted. **C.** P_2O_5 versus MgO . Grids 1 and most of 2 have P_2O_5 below detection limits and are not plotted. **D.** LOI versus MgO . High LOI in Grid 3 and some of the non-grid harzburgites is comparable to higher LOI at higher MgO in the background dataset (see inset).

Grid 3 is also notable because of its high average MgO, CaO, P₂O₅, and Al₂O₃, and low SiO₂ compared to other JOC harzburgites. **E.** MgO/SiO₂ versus Al₂O₃/SiO₂.

Figure 5. Primitive mantle-normalized trace element patterns, sorted by grid. Individual samples within each grid are plotted as thin purple lines, and grid median is shown as thicker black lines.

A. Grid 1 (Antinmäki). **B.** Grid 2 (Antinmäki). **C.** Grid 3 (Antinmäki). **D.** Grid 4 (Lehmivaara). **E.** Non-grid harzburgites.

Figure 6. PM-normalized rare earth element patterns, sorted by grid. Individual samples are shown as thin purple lines, and grid medians are given as thicker black lines. Median REE patterns of background oceanic lithosphere and cratonic lithosphere are also shown (from compilation provided by Jingao Liu). Thin black lines are representative E-DMM (enriched Depleted MORB Mantle), DMM (Depleted MORB Mantle), and D-DMM (depleted MORB Mantle) plotted for comparison, after Workman & Hart (2005). **A.** Grid 1 (Antinmäki). **B.** Grid 2 (Antinmäki). **C.** Grid 3 (Antinmäki). **D.** Grid 4 (Lehmivaara). **E.** Non-grid harzburgite samples.

Figure 7. Niobium/U versus Nb/Yb. Arrows indicate expected vectors for U-mobility compared to source enrichment.

Figure 8. PM-normalized HSE concentrations, sorted by grid. Median and 2SE (median) are also shown for all JOC harzburgites to compare extent of heterogeneity within a grid to that characterized for the entire ophiolite (proximity kilometer scale variability). **A.** Grid 1 (Antinmäki). **B.** Grid 2 (Antinmäki). **C.** Grid 3 (Antinmäki). **D.** Grid 4 (Lehmivaara). **E.** Non-

grid harzburgite samples. The 2SE of the median is calculated as $1.253 \cdot (2\sigma_{\text{mean}}/\sqrt{N})$, where σ_{mean} is the standard deviation of the mean, and N is the number of samples in the population.

Figure 9. A. PM-normalized Re/Pd versus Re, showing a correlation between increasing Re abundance and superchondritic Re/Pd, indicative of secondary Re re-enrichment. In some cases, notably with Grid 2, this re-enrichment correlates with poor agreement between T_{MA} , T_{RD1} , and T_{RD2} . Most of the rest of the samples lack evidence of Re modification and all three model ages tend to fall within ~ 0.3 Gyr (see **Table 1**). **B.** PM-normalized Re/Ir versus Pd/Ir, showing that, with the exception of most of Grid 2 and several other outlying samples, JOC harzburgites are typically depleted in both elements compared to PM. This also confirms that elevated Re/Pd in panel. **A** is controlled by elevated Re, and Pd is not unusually depleted in any high Re/Pd samples. Further, no systematic variation between Re and Pd is apparent.

Figure 10. A. $^{187}\text{Re}/^{188}\text{Os}$ versus γ_{Os} of JOC harzburgites. Elevated $^{187}\text{Re}/^{188}\text{Os}$ ($> \sim 0.2$ in JOC harzburgites) in some samples implies that the ^{187}Re - ^{187}Os system was disrupted at some point after the lithospheric mantle was generated, but trajectories (black arrows) suggest that samples in different locations within the ophiolite may have been affected differently. Most of the Grid 2 samples, a Grid 4 sample, and one of the non-grid harzburgite samples (JU15-1) have elevated $^{187}\text{Re}/^{188}\text{Os}$. The latter two samples have excess Re and so overcorrected initial $^{187}\text{Os}/^{188}\text{Os}$ relative to the rest of the harzburgites, whereas Grid 2 $^{187}\text{Os}/^{188}\text{Os}$ are indistinguishable from rest of the data. **B.** T_{RD2} versus T_{MA} , illustrating which samples have good agreement between the two model approaches. Samples plotting the furthest from the 1:1 correlation line are the same samples shown to have Re re-addition in **Figure 9**. One non-grid sample (ATK96) gave a T_{MA} of

Eu/Eu*. Although a weaker correlation than seen in **B**, γOs_t and Eu/Eu* differ between Type 1 and Type 2 harzburgites. The Type 1 outliers are likelier to skew to higher γOs_t while Type 2 outliers are likelier to skew to lower γOs_t .

Figure 14. γOs_t versus obduction age of a global ophiolite compilation. Harzburgites (dark blue circles) and chromites (light grey circles) are shown; see **Table 3** for details. Each ophiolite is shown as a median γOs_t , with error bars representing the standard error of the median. Models of the evolution of the abyssal mantle reservoir (dashed grey line after Haller et al., 2021) and PM (solid grey line: after Meisel et al., 2001) are shown for reference. The JOC Type 1 γOs_t value plots within the compositional range of modern ophiolites and depleted by roughly 2% relative to the chondritic reference ($\gamma\text{Os}_t = 0$). Data sources for compilation: Snow et al., 2000; Büchl et al., 2002; Walker et al., 2002; Melcher and Meisel, 2004; Frei et al., 2006; Agranier et al., 2007; Kusky et al., 2007; Shi et al., 2007; Schulte et al., 2009; Uysal et al., 2012; O'Driscoll et al., 2012, 2015, 2018; Gonzáles-Jiménez et al., 2014; Gong et al., 2020; Secchiari et al., 2020; Snortum and Day, 2020; Xu et al., 2020; Haller et al., 2021.

Figure 1

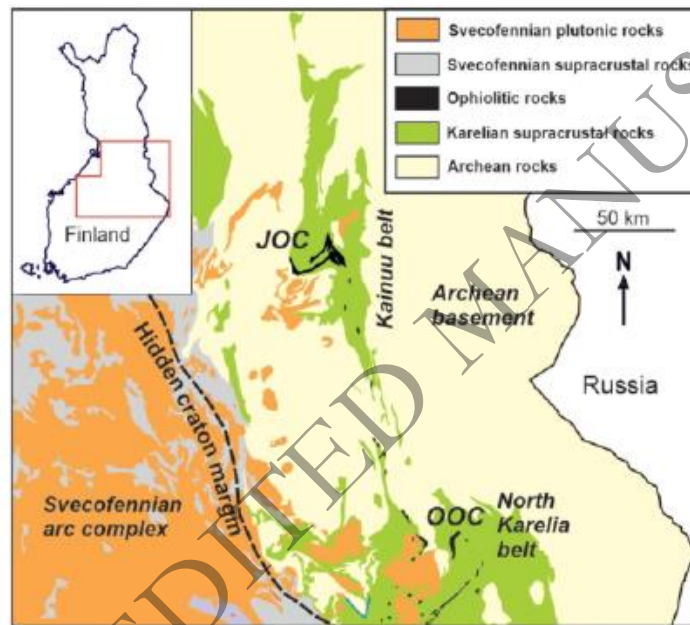


Figure 2

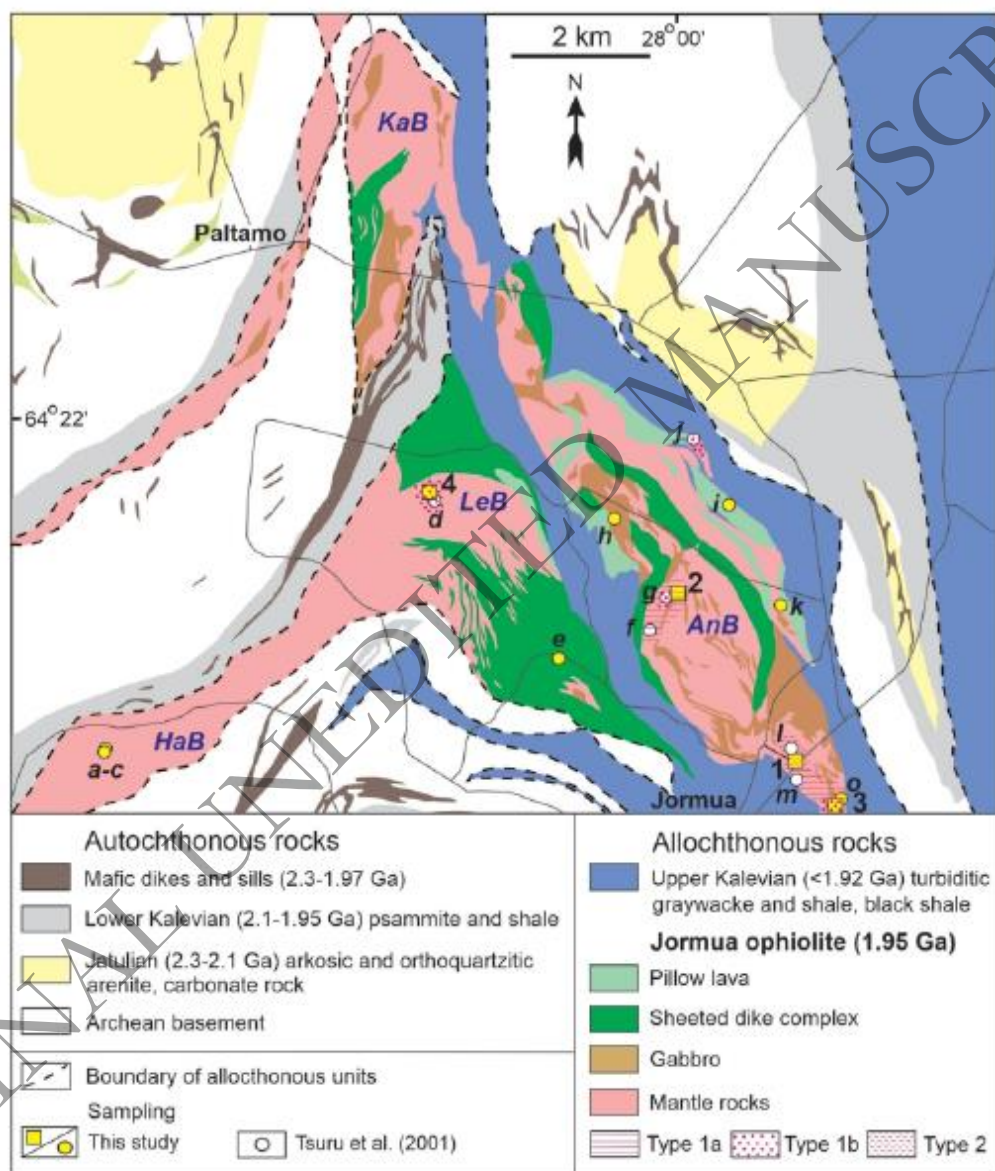


Figure 3

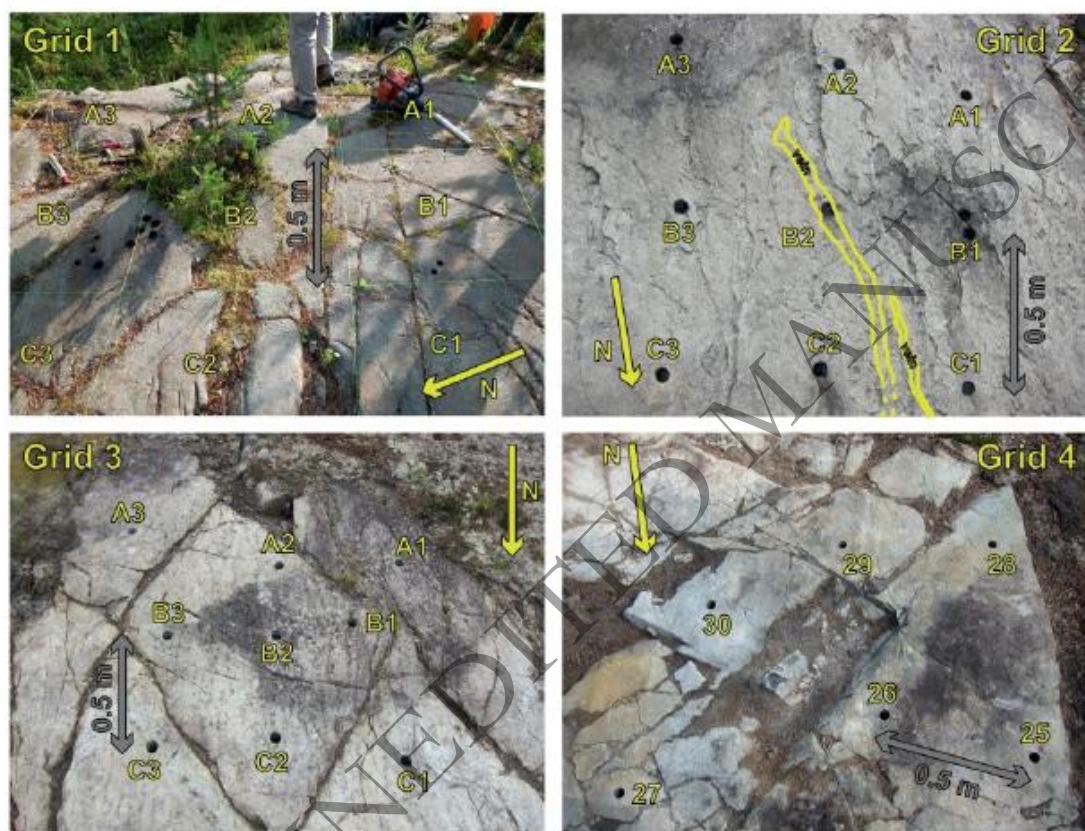


Figure 4

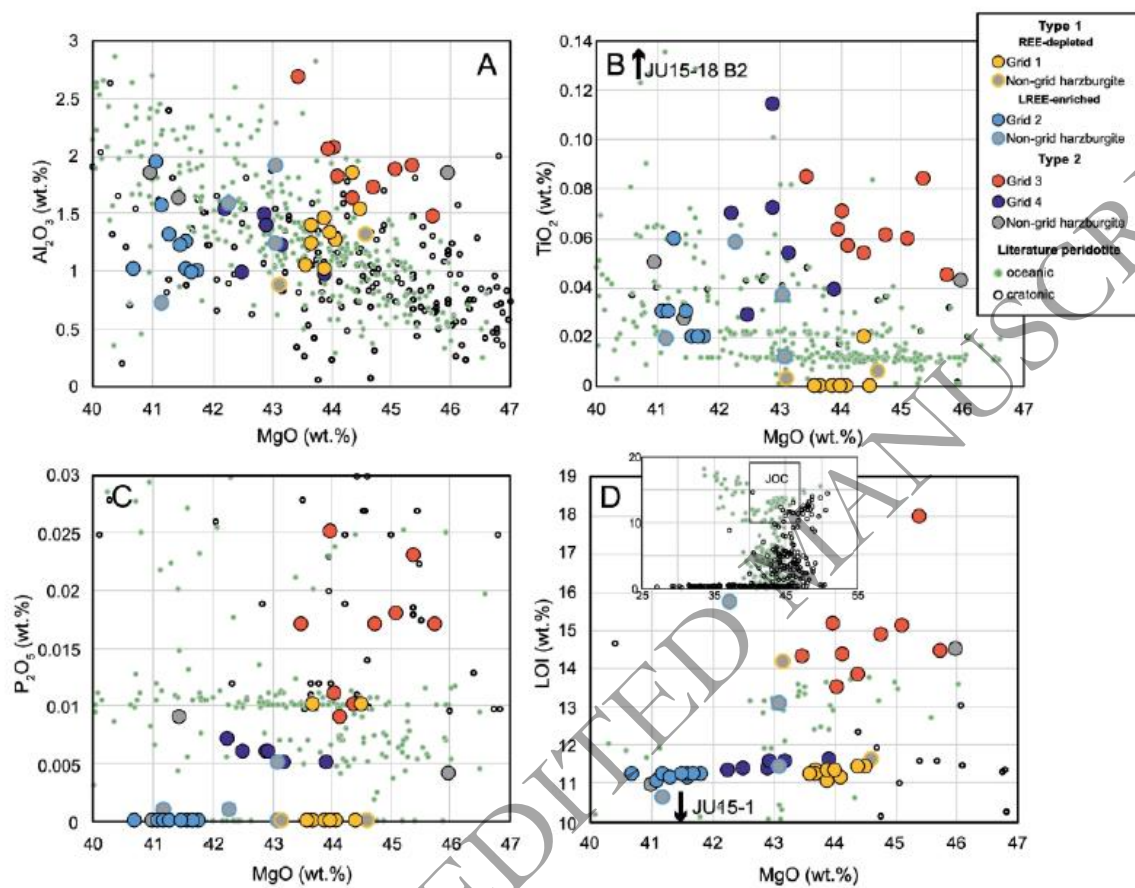


Figure 5

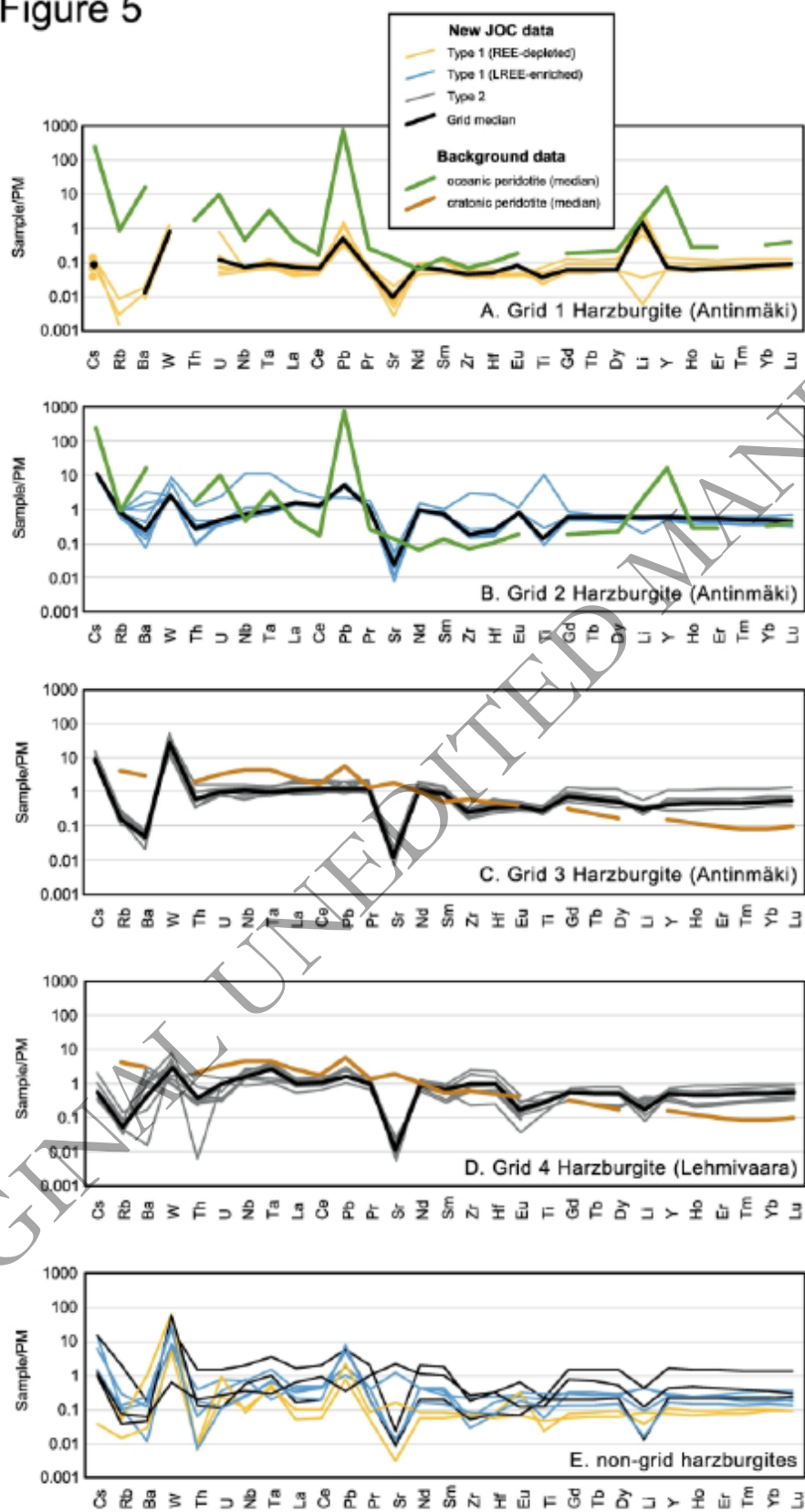


Figure 6

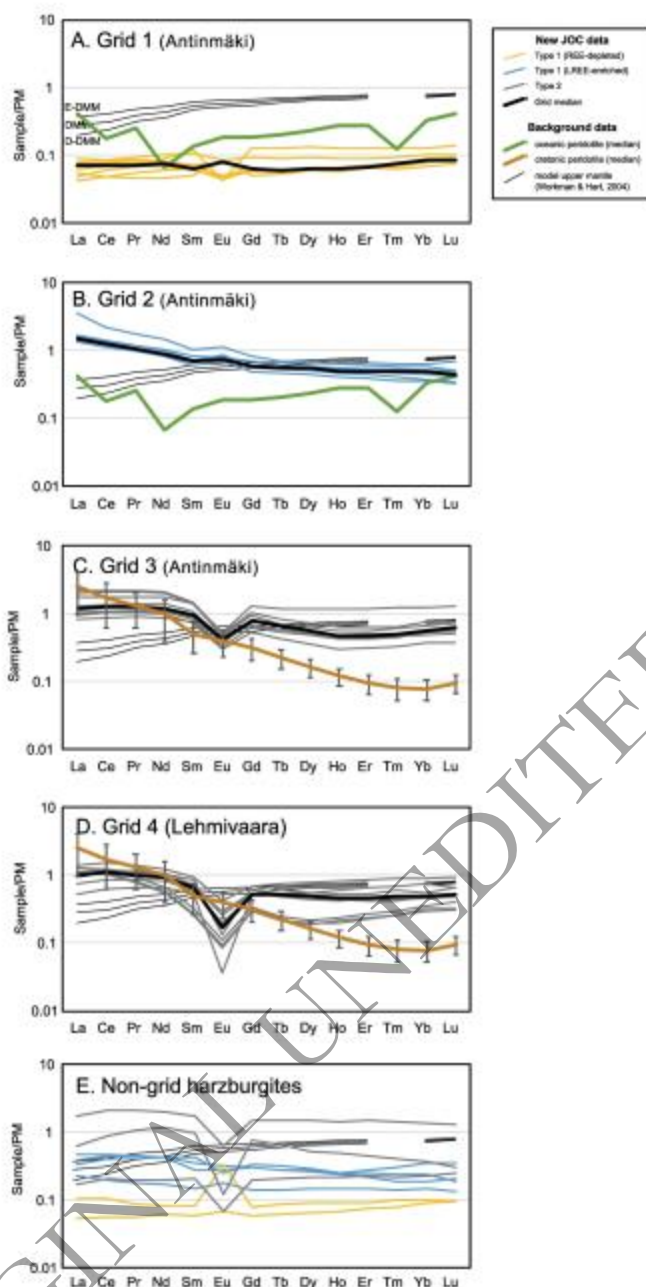
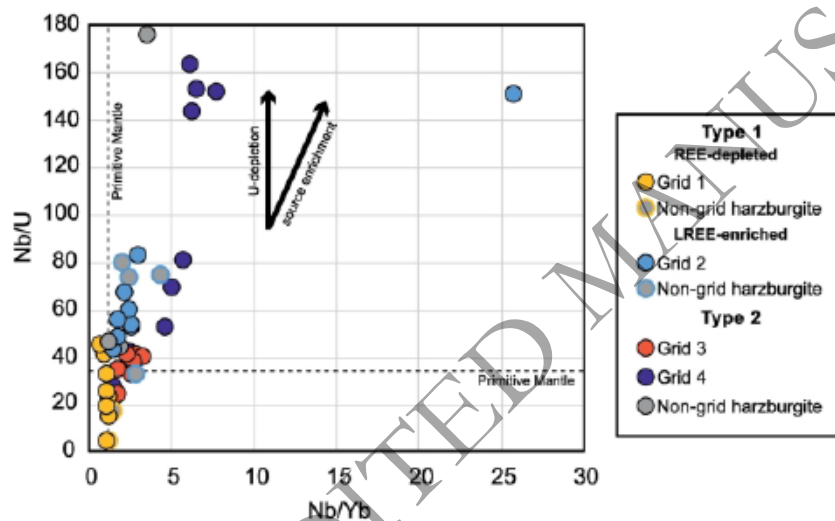


Figure 7



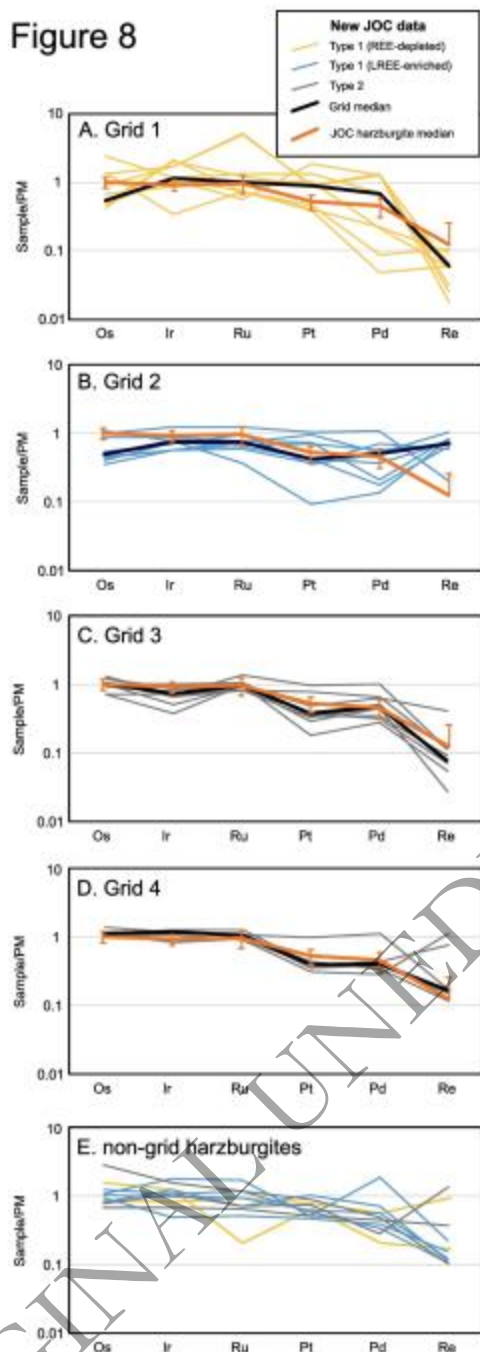


Figure 9

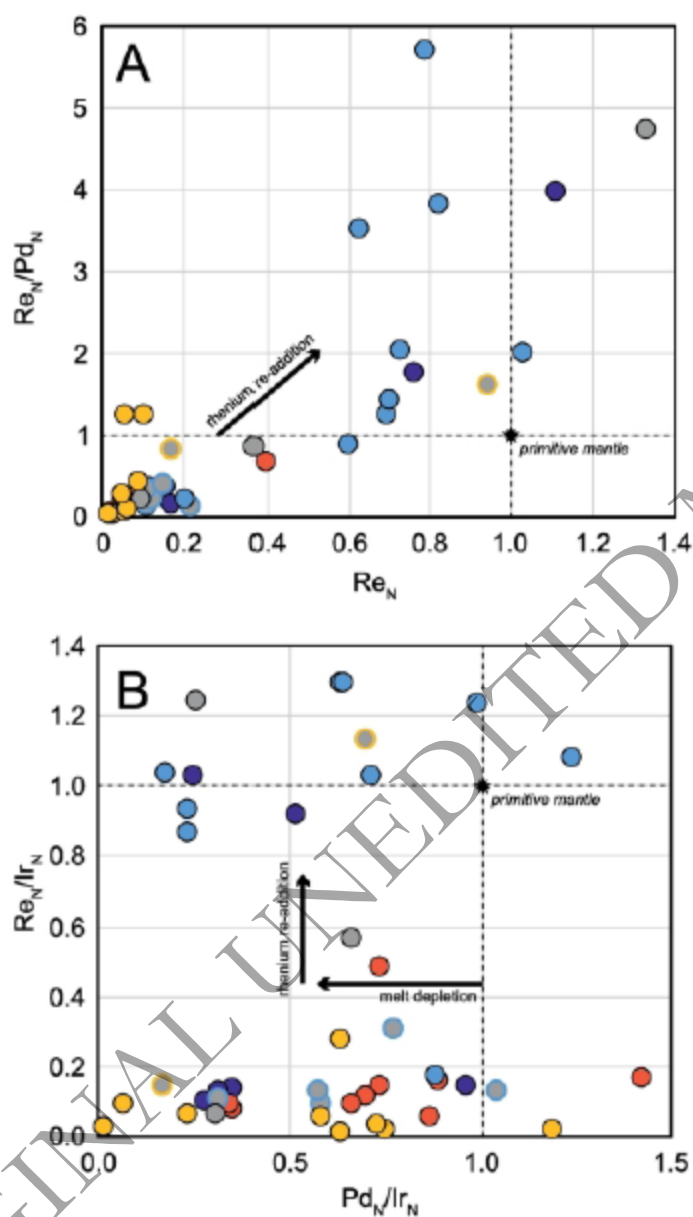


Figure 10

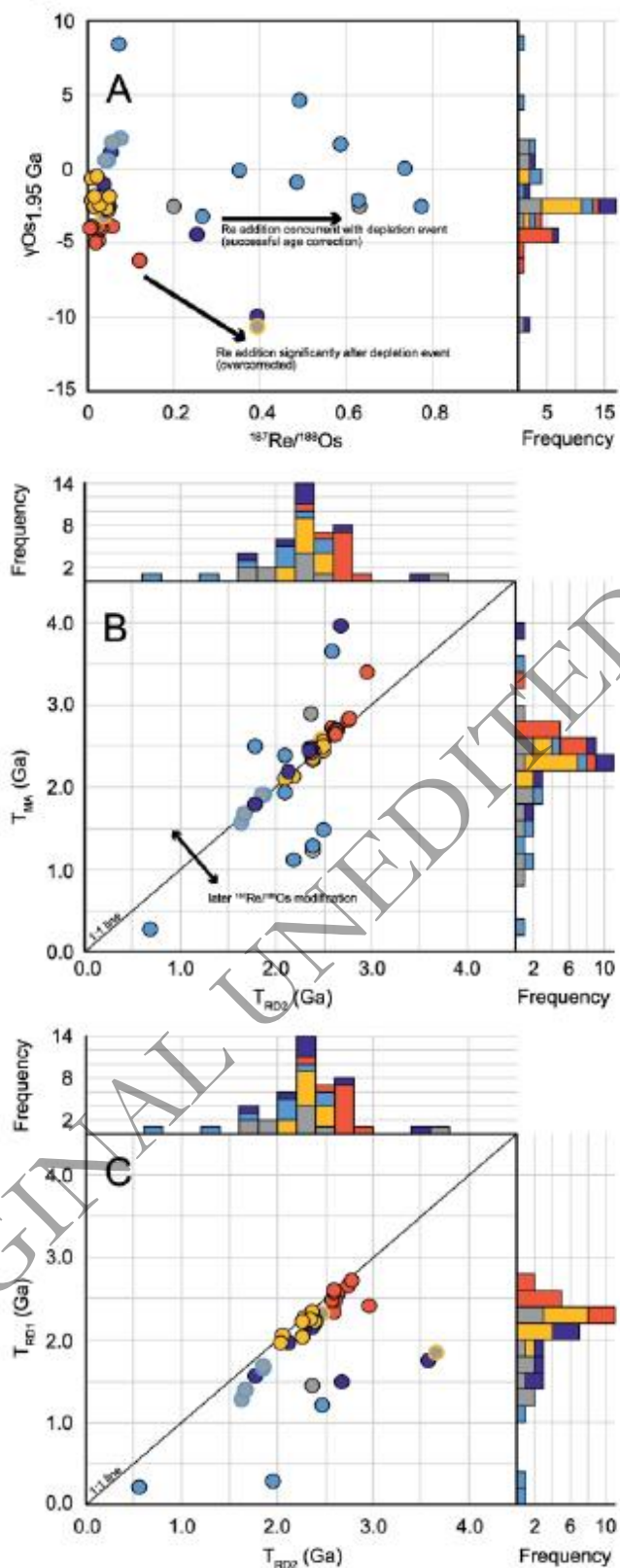
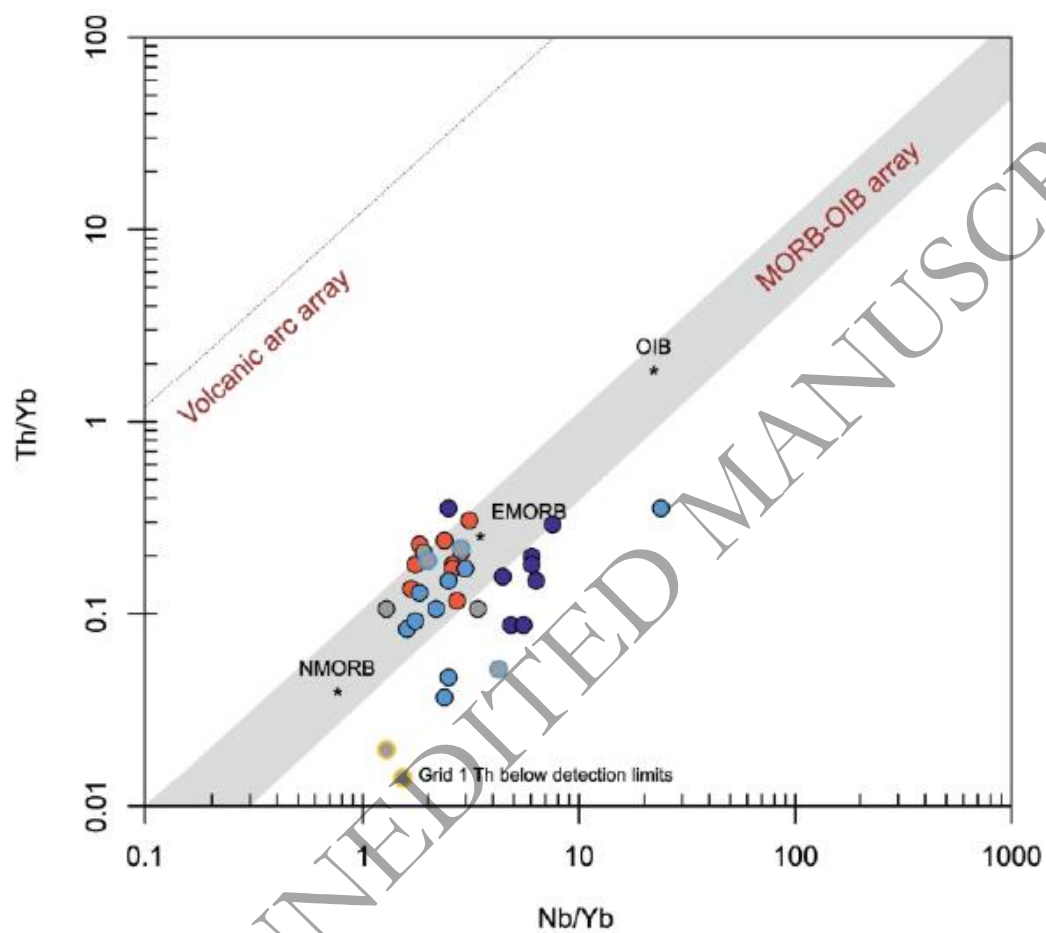


Figure 11



1328



Figure 13

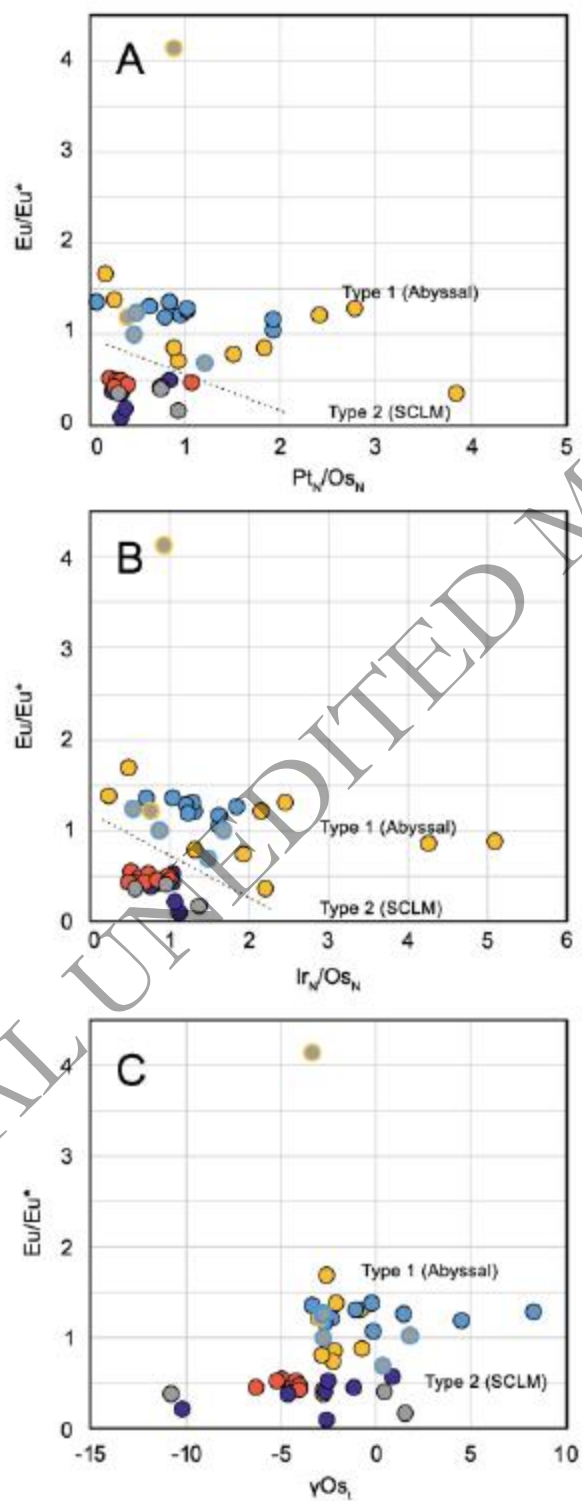


Figure 14

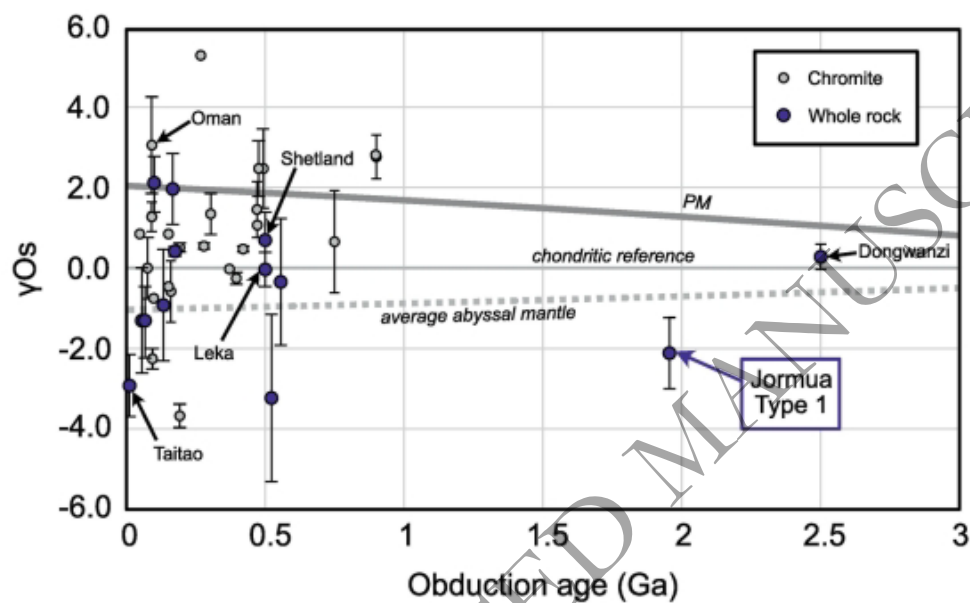


Table 1. Highly siderophile element data (ppb) and Re-Os systematic for IOC serpentinitised harzburgites.

Sample	Rock	Type	Os	Ir	Ru	Pt	Pd	Re	Os/Ir	Ru/Ir	Pt/Ir	Pd/Ir	Os/Ru	Ru/Pd	Pt/Pd	Os/Pt	Re/Pd	Re/Os	Ir/Os	Pt/Os	¹⁸⁷ Os/ ¹⁸⁸ Os	±2SE	±2SE	¹⁸⁷ Os/ ¹⁸⁸ Os	¹⁸⁷ Os/ ¹⁸⁸ Os	γ _{Os} ± 1σ	T _{DM} (Ga)	T _{DM} (Ga)	T _{DM} (Ga)	
Non-grid harzburgites																														
JUL15-23*	Antimaki	1, UREE-enriched	4130	6432	1328	3940	9370	0.067	0.6	2.1	0.6	1.3	0.6	0.9	0.3	1.0	0.1	0.1	1.7	0.5	0.1184	7.8E-05	0.07	0.1158	0.078	1.8	1.6	1.3	1.7	
AT173-1*	Antimaki	1, UREE-enriched	3254	4256	1428	7128	1344	0.04	0.7	1.3	0.8	0.7	0.6	0.8	0.4	0.1	0.1	1.5	1.2	0.1129	0.20	0.1162	0.090	3.0	1.9	1.4	1.9			
AT173-8*	Antimaki	1, UREE-enriched	4037	3244	7517	4738	3015	0.037	1.1	2.1	1.3	0.9	1.1	1.0	0.6	0.6	0.1	0.2	1.0	0.6	0.1157	0.20	0.1143	0.044	0.5	1.9	1.7	1.9		
AT173-9*	Antimaki	2, UREE-enriched	2.665	2.336	5.065	3.877	2.341	0.112	1.1	2.1	1.6	1.0	1.0	1.0	0.8	0.7	0.6	0.8	1.0	0.8	0.1174	0.20	0.1107	0.203	-0.7	2.9	1.4	2.4		
AT173-7*	Antimaki	1, UREE-enriched	4520	3277	9206	4256	1542	0.033	1.2	2.4	1.1	0.5	1.1	1.1	0.6	0.3	0.1	0.3	0.9	0.5	0.1115	0.20	0.1104	0.033	-0.9	2.5	1.3	2.4		
AT173-6*	Antimaki	1, UREE-depleted	6159	4422	1656	4976	1120	0.052	1.4	0.4	1.1	0.3	1.3	0.2	0.5	0.2	0.1	0.8	0.8	0.4	0.1113	0.20	0.1100	0.040	-9.3	2.6	1.3	2.5		
AT173-5*	Antimaki	1, UREE-depleted	3.457	3.020	7.451	6.037	3.105	0.284	1.1	2.5	2.0	1.0	1.1	1.1	1.0	0.7	0.1	1.6	0.9	0.9	0.1146	0.20	0.1016	0.395	-11	63	1.8	3.7		
AT173-8*	Antimaki	2, UREE-enriched	10.84	5.602	9.336	6.715	2.575	0.030	1.9	1.7	1.2	0.5	1.8	0.8	0.6	0.3	0.1	0.2	0.6	0.3	0.1111	0.20	0.1107	0.014	-2.7	2.4	2.3	2.4		
AT173-9*	Antimaki	1, UREE-enriched	3.623	1.512	4.021	3.507	1.071	0.046	2.0	2.1	2.0	1.1	1.8	1.0	0.9	0.8	0.3	0.4	0.5	0.5	0.1176	0.20	0.1126	0.061	1.6	1.7	1.4	1.7		
JUL15-1*	Hammurabi	2, UREE-enriched	3.041	3.885	5.520	5.52	1.484	0.400	0.8	1.4	1.4	0.4	0.7	0.6	0.7	0.3	1.2	4.7	1.4	0.9	0.1317	9.7E-05	0.07	0.1107	0.634	-2.7	1.2	-0.7	2.4	
JUL15-10a*	Hammurabi	2, UREE-enriched	1.899	-	-	-	-	0.611	-	-	-	-	-	-	-	-	-	-	-	-	0.1467	1.1E-04	0.07	0.0952	1.554	-16	1.0	-1.0	4.6	
mean non-grid			4546	3944	6877	5156	3128	0.120	1.2	1.8	1.4	0.8	1.1	0.8	0.7	0.5	0.4	0.9	1.0	0.7	0.1173		0.1106	0.201	-2.7	8.2	1.4	2.4		
2SD			4.912	2.783	6.496	2.681	5.006	0.312	0.9	1.3	0.9	0.8	0.9	0.6	0.4	0.5	0.9	2.8	0.8	0.6	0.0163		0.0099	0.669	8.7	38.6	2.4	1.4		
median non-grid			3.630	3.891	6.576	4.857	2.468	0.049	1.1	2.1	1.4	0.9	1.1	1.0	0.7	0.6	0.1	0.4	0.9	0.6	0.1150		0.1107	0.066	-2.7	2.2	1.6	2.4		
1SE (median)			0.973	0.551	1.287	0.531	0.992	0.062	0.2	0.3	0.2	0.2	0.2	0.1	0.1	0.1	0.2	0.6	0.2	0.1	0.0032		0.0020	0.133	1.7	7.7	6.5	0.3		
median non-grid Type 1																					0.1142		0.1142	0.060	0.5	1.9	1.7	1.9		
1SE (median)																					0.0024		0.0024	0.062	2.1	10.8	6.2	0.3		
median non-grid Type 2																					0.1107		0.1107	0.203	-2.7	2.4	1.4	2.4		
1SE (median)																					0.0032		0.0032	0.417	2.8	0.7	1.4	0.5		
Grid 1																														
JUL15-16 A1*	Antimaki	1, UREE-depleted	1.636	7.322	9.525	6.543	6.799	0.007	0.3	1.3	0.9	0.9	0.2	0.6	0.4	0.6	0.0	0.0	4.3	1.9	0.1119	0.20	0.1113	0.018	-2.1	1.3	1.2	2.4		
JUL15-16 A2*	Antimaki	1, UREE-depleted	3.350	5.994	40.34	6.200	6.642	0.009	0.6	6.7	1.0	1.1	0.5	3.1	0.5	0.8	0.0	0.0	1.9	1.0	0.1117	0.20	0.1112	0.012	-2.2	0.8	1.5	2.4		
JUL15-16 A3*	Antimaki	1, UREE-depleted	4.943	6.111	38.93	6.479	6.574	0.017	0.8	6.4	1.1	1.1	0.7	2.9	0.5	0.7	0.0	0.0	1.3	0.7	0.1112	0.20	0.1106	0.017	-2.7	1.4	2.3	2.5		
JUL15-16 A3*	Antimaki	1, UREE-depleted	1.880	3.778	4.371	13.360	6.620	0.005	0.5	1.2	3.7	1.8	0.5	0.7	1.8	1.2	0.0	0.0	2.2	3.9	0.1132	0.20	0.1128	0.012	-0.8	1.1	2.0	2.2		
JUL15-16 B1*	Antimaki	1, UREE-depleted	2.087	2.494	7.492	2.506	2.639	0.112	0.8	3.0	1.0	1.1	0.8	1.4	0.5	0.7	1.0	1.4	1.3	0.6	0.1127	0.20	0.1104	0.090	-3.0	1.2	1.6	2.5		
JUL15-16 B2*	Antimaki	1, UREE-depleted	2.076	4.117	10.29	9.723	3.561	0.019	0.5	2.5	2.4	0.9	0.5	1.1	1.1	0.6	0.1	0.1	2.1	2.4	0.1120	0.20	0.1105	0.045	-2.8	1.5	2.0	2.5		
JUL15-16 B3*	Antimaki	1, UREE-depleted	2.534	3.079	7.734	7.427	1.094	0.016	0.8	2.5	2.4	0.4	0.8	1.1	1.2	0.2	0.1	0.3	1.3	1.5	0.1119	0.20	0.1109	0.051	-2.5	2.4	2.2	2.4		
JUL15-16 C1*	Antimaki	1, UREE-depleted	9.004	4.063	5.464	3.357	0.446	0.031	2.2	1.3	0.8	0.3	2.0	0.6	0.4	0.1	0.1	1.1	0.5	0.2	0.1120	0.20	0.1114	0.016	-2.0	2.8	1.2	2.4		
JUL15-16 C2*	Antimaki	1, UREE-depleted	5.270	1.193	5.637	3.950	1.123	0.027	4.4	4.7	2.5	0.9	4.1	2.2	1.2	0.6	0.3	0.4	0.2	0.3	0.1138	0.20	0.1130	0.034	-0.7	2.1	1.9	2.1		
JUL15-16 C3*	Antimaki	1, UREE-depleted	1.590	7.472	5.537	2.749	0.242	0.017	0.2	0.7	0.4	0.0	0.2	0.3	0.2	0.0	0.0	1.2	5.1	0.9	0.1132	0.20	0.1115	0.052	-2.0	2.3	2.0	2.4		
mean Grid 1			3.346	4.560	16.52	6.958	3.334	0.017	1.1	2.6	1.8	0.8	1.0	1.2	0.9	0.5	0.1	0.4	2.2	1.6	0.1124		0.1114	0.023	-0.0	2.3	2.1	2.4		
2SD			4.931	4.037	21.99	7.422	5.541	0.017	2.8	3.8	2.2	1.1	2.5	1.7	1.0	0.7	0.2	1.0	3.2	2.5	0.0016		0.0013	0.032	1.8	0.3	0.2	0.3		
median Grid 1			2.076	4.072	7.72	6.543	3.512	0.017	0.5	2.5	2.4	0.9	0.5	1.1	1.1	0.6	0.1	0.1	2.1	1.5	0.1120		0.1113	0.024	-2.1	2.3	2.2	2.4		
1SE (median)			1.030	0.943	4.533	1.550	1.157	0.004	0.6	0.8	0.5	0.2	0.5	0.4	0.2	0.2	0.2	0.7	0.5	0.0003		0.0004	0.007	0.3	0.1	0.0	0.1			
Grid 2																														
JUL15-16 A1*	Antimaki	1, UREE-enriched	1.677	2.051	3.957	3.135	1.928	0.220	0.8	2.9	1.5	0.9	0.5	0.8	1.3	0.7	0.6	1.3	2.0	1.3	0.1130	0.20	0.1112	0.632	-2.3	1.3	-0.8	2.4		
JUL15-16 A2*	Antimaki	1, UREE-enriched	1.555	2.051	5.146	3.135	0.952	0.190	0.6	1.9	1.2	0.4	0.5	0.9	0.6	0.2	0.9	3.5	1.8	1.1	0.1149	0.20	0.1145	0.589	1.5	2.5	-1.2	1.8		
JUL15-16 A3*	Antimaki	1, UREE-enriched	1.373	2.057	5.312	3.168	3.007	0.210	0.7	2.6	2.5	1.5	0.6	1.2	1.2	1.0	1.2	1.2	1.6	2.0	0.1180	0.20	0.1136	0.739	-1.1	1.9	-1.7	2.1		
JUL15-16 B1*	Antimaki	1, UREE-enriched	2.087	2.494	7.492	2.506	2.639	0.112	0.8	3.0	1.0	1.1	0.8	1.4	0.5	0.7	1.0	1.4	1.3	0.6	0.1127	0.20	0.1104	0.090	-3.0	1.2	1.6	2.5		
JUL15-16 B2*	Antimaki	1, UREE-enriched	3.912	4.344	9.534	7.852	5.673	0.062	0.9	2.2	1.8	1.3	0.8	1.0	0.9	0.9	0.2	0.2	1.2	1.0	0.1126	0.20	0.1131	0.076	8.3	0.3	0.2	0.7		
JUL15-16 B3*	Antimaki	1, UREE-enriched	1.762	2.011	4.760	2.779	3.680	0.180	0.9	2.4	1.4	1.8	0.8	1.1	0.7	1.2	1.1	0.9	1.2	0.8	0.1152	0.20	0.1139	0.493	4.5	5.1	-1.2	1.3		
JUL15-16 C1*	Antimaki	1, UREE-enriched	3.337	3.202	6.185	3.521	1.142	0.247	1.0	1.9	1.7	0.4	1.0	0.9	0.8	0.2	0.9	3.8	1.0	0.9	0.1153	0.20	0.1135	0.356	-0.2	1.4	0.3	2.1		
JUL15-16 C2*	Antimaki	1, UREE-enriched	4.241	2.769	2.927	0.688	0.788	0.238	1.5	1.1	0.2	0.3	1.4	0.5	0.1	0.2	1.0	5.7	0.1	0.1	0.1119	0.20	0.1100	0.270	-8.3	1.6	1.2	2.6		
JUL15-16 C3*	Antimaki	1, UREE-enriched	1.928	2.893	5.881	7.214	2.734	0.310	0.7	2.0	2.5	0.9	0.6	0.9	1.2	0.6	1.3	2.0	1.6	1.9	0.1164	0.20	0.1108	0.776	-2.6	1.5	-1.4	2.5		
mean Grid 2			2.430	2.719	5.908	4.235	2.499	0.208	0.9	2.2	1.5	0.9	0.8	1.0	0.7	0.6	1.0	2.3	1.3	1.0	0.1146		0.1144	0.491	0.5	2.2	-0.5	2.0		
2SD			2.187	1.474	3.665	4.762	3.120	0.133	0.6	1.2	1.4	1.1	0.5	0.5	0.7	0.7	0.7	3.5	0.7	1.2	0.0127		0.0065	0.453	7.5	2.8	1.9	1.2		
median Grid 2			1.928	2.651																										

Table 2. Summary of distinguishing characteristics of the mantle domain types identified in JOC harzburgites.

	Subtype	Possible protolith domain	Samples	Major elements	REE pattern	Eu/Eu* anomaly	Ir/Os	Pt/Os
Type 1	REE-depleted (1a)	abyssal	Grid 1, ATK76, ATK96	lower MgO	planar, often slight LREE depletion relative to HREE	positive or negative	highly variable	frequently positive
	LREE-enriched (1b)	abyssal	Grid 2, JU15-23, ATK57(A-B), ATK75, ATK600	lower MgO	LREE-enriched planar to concave	often positive, some slightly negative	flat to slightly positive	positive to negative
Type 2	NA	SCLM	Grids 3, 4, ATK59, ATK589, JU15-1	higher MgO	sinusoidal with higher LREEs. Often La(N)<Ce(N)	strongly negative	flat to negative	frequently negative

Table 3. Median osmium isotopic compositions for a globally representative set of ophiolites.

Locale	Study	Age (Ga)	median γOs^*	1SE(median)	no. samples
Harzburgite/Peridotite					
Taitao Ophiolite, Chile	Schulte et al. (2009)	0.006	-2.9	0.8	18
New Caledonia, Oceania	Secchiari et al., (2020)	0.053	-1.3	1.3	4
Mugla Ophiolite, Turkey	Uysal et al. (2012)	0.065	-1.3	0.9	9
Troodos Ophiolite, Cyprus	Büchl et al. (2002)	0.09	2.1	0.7	3
Western YZO (Purang), Yarlung-Zangbo Ophiolite, Tibet	Xu et al. (2020)	0.13	-0.9	1.4	17
Coast Range, California, USA	Snortum and Day (2020)	0.161	2.0	0.9	3
Mamonia Complex, Cyprus	Batanova et al. (2008)	0.2	0.4	NA	1
Ligurian Ophiolites	Snow et al. (2000)	0.18	0.4	0.1	18
Shetland Ophiolite Complex, Scotland	O'Driscoll et al. (2012)	0.492	0.7	0.7	21
Leka Ophiolite, Norway	O'Driscoll et al. (2015); Haller et al. (2021)	0.497	0.0	0.4	41
Gangou Ophiolite, Tianshan ophiolites, NW China	Gong et al. (2020)	0.516	-3.2	2.1	4
Eastern Alps, Austria	Melcher and Meisel (2004)	0.55	-0.3	1.6	4
Jormua Ophiolite Complex	This study	1.95	-2.0	0.9	17
Dongwanzi Ophiolite (peridotites only)**	Kusky et al. (2007)	2.5	0.3	0.3	4
Chromites					
Tiebaghi, New Caledonia	Walker et al. (2002)	0.05	0.9	NA	1
St. Elena, Costa Rica	Walker et al. (2002)	0.08	0.1	0.8	2
Mayari-Baracoa Ophiolite	Frei et al. (2006)	0.09	-2.2	0.3	17
Troodos, Cyprus	Walker et al. (2002)	0.091	1.3	0.4	4
Semail, Oman	Walker et al. (2002)	0.096	3.1	1.2	5
Elstratova Ophiolite, Russia	Walker et al. (2002)	0.1	-0.7	NA	1
Bulquiza, Albania	Walker et al. (2002)	0.155	-0.4	NA	1
Kukes-Kalamash, Albania	Walker et al. (2002)	0.155	0.9	NA	1
Horokanai, Japan	Walker et al. (2002)	0.16	-0.6	0.7	4
Luobasa and Dongqiao, Tibet (Type I)	Shi et al. (2007)	0.196	0.5	0.1	16
Luobasa and Dongqiao, Tibet (Type II)	Shi et al. (2007)	0.196	-3.7	0.3	7
Dun Mtn., New Zealand	Walker et al. (2002)	0.27	5.4	NA	1
Yakuno, Japan	Walker et al. (2002)	0.28	0.6	0.1	3
La Cabaña ultramafic bodies, central Chile	González-Jiménez et al. (2014)	0.3	1.4	0.5	4
Kempirsai, Urals, Russia	Walker et al. (2002)	0.4	-0.2	0.2	3
Oeyama, Japan	Walker et al. (2002)	0.42	0.5	0.1	2
Ray Iz, Polar Urals, Russia	Walker et al. (2002)	0.47	1.5	0.7	3
Unst, Shetland Islands, Scotland	Walker et al. (2002)	0.473	1.1	NA	1
Leka, Norway	Walker et al. (2002)	0.497	2.5	1.0	3
Al' Ays, Saudi Arabia	Walker et al. (2002)	0.75	0.7	1.3	5
Morro Feio, Brazil	Walker et al. (2002)	0.9	2.9	NA	1
Crominia, Brazil	Walker et al. (2002)	0.9	2.9	0.5	2

* Datasets were filtered to exclude samples with high $187\text{Re}/188\text{Os}$ (>0.25 ; Tsuru et al., 2000). γOs was recalculated for each locale, using the same approach as for the JOC data in Table 1.

**The status of Dongwanzi as an Archean ophiolite complex has been the subject of debate (Kusky et al., 2001; 2002; 2007; Zhai et al., 2002; Zhao et al., 2007). We include it here for thoroughness.