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Modification of an ancient subcontinental lithospheric mantle by continental subduction: insight from the Maowu garnet peridotites in the Dabie UHP belt, eastern China

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Abstract

Orogenic mantle-derived peridotites commonly originate from the subcontinental lithospheric mantle (SCLM) and thus provide a key target to investigate the modification of the SCLM by a subducting slab. The Maowu ultramafic rocks from the Dabie ultrahigh-pressure (UHP) metamorphic belt have formerly been debated as representing cumulates or mantle-derived peridotites. Detailed petrological and geochemical data presented in this study provide new constraints on the origin and formation of the peridotites involving melt depletion in the ancient SCLM and deep crustal metasomatism. The Maowu garnet dunites have refractory bulk compositions characterized by high Mg# (91.9–92.0) and Ni (2537–2892 ppm) values and low Al₂O₃ (0.26–0.76 wt.%), CaO (0.05–0.32 wt.%), TiO₂ (<0.03 wt.%), Pd/Ir (0.40–0.46) and ¹⁸⁷Os/¹⁸⁸Os (minimum 0.11461) values. The Paleoproterozoic model ages (T_{RD} = 2.1 Ga, T_{MA} = 2.3 Ga) of the most refractory dunites represent minimum estimates for the age of the initial melt extraction. The extremely depleted nature, high olivine Fo (92.7–93.9), high Cr# (82–87) of spinel, and Re–Os isotopic data suggest that the Maowu garnet dunites are the residues of ~40% partial melting and represent a Paleoproterozoic fragment of the SCLM beneath the southeastern margin of the North China craton. Many garnet orthopyroxenite veins crosscutting the Maowu dunites preserve abundant metasomatic textures and show variable enrichment in incompatible elements. Mineral and whole-rock chemistry indicate that these veins represent metasomatic products between the wall dunites and silica-rich hydrous melts under UHP conditions. The veins show large variations in platinum-group

element (PGE) signatures and Re–Os isotopes. The garnet-poor orthopyroxenite veins are characterized by low Al_2O_3 (<2 wt.%) and S (<31 ppm) contents and have PGE patterns and $^{187}\text{Os}/^{188}\text{Os}$ ratios similar to the wall dunites, whereas the garnet-rich orthopyroxenite veins have high Al_2O_3 (>6 wt.%) and S (99–306 ppm) contents and show melt-like PGE patterns and high $^{187}\text{Os}/^{188}\text{Os}$ ratios (up to 0.36910). These features, combined with the occurrence of interstitial sulfides in the garnet-rich orthopyroxenite veins, suggest that crust-derived sulfur-saturated silicate melts may have significantly modified the PGE signature and destroyed the Re–Os systematics the SCLM. However, when the crust-derived silicate melts became sulfur-depleted, such melts would not significantly modify the PGE patterns, radiogenic Os-isotope compositions or the Re-depletion model ages of the SCLM. Consequently, deep crust-mantle interactions in continental subduction zones could induce high degrees of Os isotopic heterogeneity in the SCLM wedge.

Keywords: garnet dunites, SCLM, crustal metasomatism, PGE, Re–Os isotopes

1. Introduction

Orogenic peridotites are commonly observed within the crustal rocks in continental subduction zones (Brueckner and Medaris, 2000; Menzies et al., 2001). These rocks usually show compositional layers or veins from refractory harzburgite or dunite to fertile lherzolite or garnet pyroxenite (Jahn et al., 2003; Ackerman et al., 2009). Several layered peridotites and pyroxenites have commonly been described as cumulates from mafic magmas in the crustal level that form lower crustal igneous intrusions (Chavagnac and Jahn, 1996; Zhang et al., 2000; Zheng et al., 2008) and have been termed 'crustal' type garnet peridotites (Brueckner and Medaris, 2000). In contrast to the cumulate model, some layered peridotites and pyroxenites were formed by the metasomatic reaction of crust- or mantle-derived melts/fluids with the mantle wedge (Rampone and Morten, 2001; Malaspina et al., 2006, 2009; Ackerman et al., 2009); these orogenic ultramafic rocks are commonly referred to as mantle type garnet peridotites (Brueckner and Medaris, 2000). Based on whole-rock chemistry, Reverdatto et al. (2008) provided geochemical criteria to distinguish these two types of peridotites even after the complex history they have been subjected to since their original emplacement. However, recent studies revealed that some mantle type peridotites with extensive crustal metasomatism exhibit geochemical characteristics of both crustal and mantle type peridotites (Vrijmoed et al., 2006, 2013), making discrimination between these two types difficult.

Increasingly, petrochemical studies have demonstrated that the mantle type peridotites originate primarily from the subcontinental lithospheric mantle (SCLM)

and contain abundant metasomatic phases such as garnet, pyroxene, amphibole, phlogopite, apatite, monazite, dolomite, zircon and sulfide (Yang and Jahn, 2000; Zhang et al., 2000, 2007; Yang, 2003; Malaspina et al., 2006, 2009; Ye et al., 2009; Chen et al., 2013b). The metasomatic agents are interpreted as fluids or melts derived from the subducting slabs (Scambelluri et al., 2006, 2014; Malaspina et al., 2006, 2009; Zheng, 2012). Therefore, mantle type orogenic peridotites are also excellent research targets for exploring the mechanical and chemical effects of element transfer into the mantle wedge at various depths (Bruckner, 2004; Scambelluri et al., 2006, 2008).

The Maowu mafic–ultramafic complex hosted by coesite-bearing paragneisses is a typical orogenic peridotite body in the Dabie UHP belt, eastern China (Fig. 1). The complex consists mainly of layered garnet orthopyroxenite, garnet clinopyroxenite and garnet websterite, in addition to minor peridotite and eclogite. This mafic–ultramafic complex has been interpreted as a crustal type peridotite subducted to great depth in the Triassic (Okay, 1994; Zhang et al., 1998; Jahn et al., 2003). However, the poikilitic orthopyroxene grains in the Maowu pyroxenites have high nickel content (> 2000 ppm) and are enriched in light rare earth elements (LREEs) and large ion lithophile elements (LILEs), suggesting that they were formed by interactions between olivine and a silica-rich hydrous melt (Malaspina et al., 2006, 2009). A detailed petrological study of the Maowu garnet orthopyroxenites indicates peak metamorphic conditions of ~ 800 °C at 5.3–6.3 GPa (Chen et al., 2013a). To investigate the origin of this body, most previous studies are focused on the garnet

pyroxenites (Okay, 1994; Fan et al., 1996; Zhang et al., 1998; Liou and Zhang, 1998; Jahn et al., 2003; Malaspina et al., 2006, 2009; Chen et al., 2013a, 2013b). However, the petrological and geochemical features of the Maowu peridotites are not well investigated because of their rare occurrence in the outcrop.

In this paper we present a study of garnet dunites with garnet orthopyroxenite veins in the Maowu mafic–ultramafic complex. This paper includes detailed mineral textural features, major and trace element compositions of bulk rocks and minerals, Re–Os isotopes and platinum-group element (PGE) data to investigate the origin, age and metasomatism of the Maowu garnet dunites. The new data indicate that the dunites (1) represent mantle residues after ~40% partial melting, (2) were derived from an ancient SCLM wedge beneath the NCC, and (3) experienced crustal metasomatism by silica-rich hydrous melts during continental subduction that disturbed the Re–Os systematics of the SCLM.

2. Geological setting

The Dabie UHP metamorphic terrane is located in the southwestern part of the Dabie–Sulu Orogen in eastern China (Fig. 1). It was formed by subduction/collision of the Yangtze Craton beneath the Sino-Korean Craton in the Triassic (e.g., Li et al., 2000; Zheng et al., 2006). The Dabie–Sulu UHP terrane is composed primarily of two sets of lithologies. One set is intercalated marble, eclogite, garnet–biotite schist, jadeite–quartzite and dark biotite–epidote-rich paragneisses, which are considered to be the metamorphic sedimentary cover of the northern margin of the Yangtze Craton

(Rolfo et al., 2004). The other set is tonalitic–trondhjemitic–granitic orthogneisses with rare eclogite blocks. Coesite is common and diamond is rarely reported from fresh UHP rocks, such as eclogite, jadeite–quartzite, garnet-bearing marble, and garnet–biotite schist (Xu et al., 1992, 2003, 2005; Hirajama and Nakamura, 2003; Rolfo et al., 2004), thus implying that the whole UHP terrane had ever subducted to a depth of more than 80 km as a coherent slab (Ye et al., 2000). Most of the UHP metamorphic rocks in the Dabieshan–Sulu UHP terrane recorded peak P–T conditions up to 4–6 GPa and 700–850 °C (Carswell et al., 1997; Rolfo et al., 2004; Shi et al., 2006). The ages of the UHP metamorphism are measured at 230–240 Ma (e.g., Zheng et al., 2004, 2006; Liu et al., 2006).

Several garnet peridotite bodies occur as meter- to kilometer-sized blocks within the amphibolite-facies paragneisses or granitic gneisses in the Dabie–Sulu UHP terrane (Fig. 1) (Okay, 1994; Zhang et al., 1998, 2000). Both crustal and mantle type peridotites have been identified in the Dabie–Sulu UHP terrane (Zhang et al., 2000). The Maowu mafic–ultramafic body occurs in granitic gneiss and biotite gneiss in the Dabie UHP terrane (Jahn et al., 2003). It is a lens (50 m × 250 m) composed of millimeter- to centimeter-thick compositional layers, which are dominated by garnet orthopyroxenite, garnet clinopyroxenite and garnet websterite, with minor garnet dunite and eclogite. No lherzolite has been discovered in the outcrop. The boundaries between various rock types are generally sharp but not planar. Phlogopite-rich horizons or veins develop adjacent to the orthopyroxenite layers. Many previous studies suggested that the protolith of the Maowu mafic–ultramafic body was a

cumulate complex formed by fractional crystallization of a basaltic magma intruded into the crust of the Yangtze Craton, which experienced deep subduction and UHP metamorphism in the Triassic (Okay, 1994; Fan et al., 1996; Zhang et al., 1998, 1999; Liou and Zhang, 1998; Jahn et al., 2003). On the other hand, several relict olivine and low-Ni orthopyroxene are overgrown by poikilitic orthopyroxene in the Maowu garnet orthopyroxenites, suggesting that the garnet orthopyroxenites were formed by metasomatic interactions between precursor refractory mantle peridotites and slab-derived hydrous melts (Malaspina et al., 2006, 2009; Chen et al., 2013a). Detailed textural and mineralogical evidence and thermodynamic modeling demonstrate that the Maowu garnet orthopyroxenites underwent six stages of metamorphism, corresponding to corner-flow process in the shallow–wet mantle wedge, recycling of the rocks deep into the upper mantle by downward flow, and the uplift process by the exhumation of continental country rocks (Chen et al., 2013a).

3. Petrography

The Maowu garnet dunites are generally fresh and consist of ~93 vol.% olivine, 2 vol.% orthopyroxene, and 2 vol.% garnet, with trace Cr-spinel and serpentine. No clinopyroxene has been observed. In hand specimen, the rock shows massive structure and the minerals show well-developed equilibrium granoblastic polygonal fabrics (Fig. 2a). Coarse-grained (2-3 mm) olivine is euhedral to subhedral and occasionally shows ‘triple junction’ texture (Fig. 2c, 2e). Coarse-grained orthopyroxene porphyroblasts (up to 1 cm) occur in the matrix olivine; they are typically anhedral to subhedral

grains and always contain chromite exsolution lamellae (Fig. 2b). The chromite lamellae are 10–500 μm long and 2–20 μm wide, and make up ~ 5 vol%, based on back-scattered electron images. Cr-spinel is identified as inclusions in olivine and orthopyroxene and rarely as euhedral intergranular matrix grains ranging from 0.2 to 0.5 mm in diameter. Rare garnet occurs as intergranular grains in the matrix olivine and usually contains rounded olivine (Fig. 2c).

Both in hand specimen and thin section, several garnet orthopyroxenite veins with variable thicknesses (1–20 mm) can be observed crosscutting the garnet dunites (Fig. 2a, 2d-e). These veins are composed mainly of medium-grained (0.5–1 mm) orthopyroxene, garnet, Cr-spinel, Ti-clinohumite, and olivine, with trace apatite and sulfide. Olivine commonly occurs as inclusions in orthopyroxene, garnet and Ti-clinohumite (Fig. 2e-g). Ti-clinohumite exhibits brown color and shows equilibrium texture with the matrix orthopyroxene; it currently contains abundant magnetite lamellae except along its margins (Fig. 2g). The rims of matrix garnet are partially replaced by later-stage kelyphites of amphibole (Fig. 2d). Based on the abundance of garnet and orthopyroxene, the veins can be subdivided into the following two groups: garnet-rich orthopyroxenite with >10 vol.% garnet and garnet-poor orthopyroxenite with <5 vol.% garnet. Sometimes the garnet dunites containing very tiny garnet-poor orthopyroxenite veins (1 mm thickness) cause dunite to tend locally towards harzburgite (Fig. 2e). In the garnet-poor orthopyroxenites, limited olivines are preserved in the matrix (Fig. 2f). In the garnet-rich orthopyroxenites, some sulfides (30–300 μm) can be observed interstitially along the

boundaries of orthopyroxene and garnet (Fig. 2h), which have been interpreted as secondary phases added from percolating melts (Luguet et al., 2003; Liu et al., 2009). However, no sulfides can be observed in the garnet-poor orthopyroxenites.

4. Analytical methods

4.1 Whole-rock major and trace elements

The garnet orthopyroxenite veins of the Maowu samples were carefully extracted from a 0.5–2 cm-thick slice of rock. Two harzburgite samples (MW38-2 and MW39-1) represent the garnet dunites containing tiny garnet-poor orthopyroxenite veins (approximately 1-2 mm thick). All samples were disaggregated between thick sheets with a rock hammer, and the small blocks were then cleaned in an ultrasonic cleaner with distilled water. After being placed in an oven at 60 °C for more than 24 hours, these blocks were ground in an agate mill to approximately 200 mesh for whole-rock analyses.

Whole-rock major element compositions were analyzed by X-ray fluorescence spectrometry (XRF) on fused glass disks at the Institute of Geology and Geophysics, Chinese Academy of Sciences (IGGCAS), with an analytical uncertainty ranging from 1% to 3% for elements present at >1 wt.% and approximately 10% for elements present at <1 wt.%.

Whole-rock trace element concentrations were analyzed on rock powders by Inductively Coupled Plasma Mass Spectrometry (ICP-MS) using a Finnigan Mat Element Spectrometer at the IGGCAS. Sample powders were digested in distilled

HF+HNO₃ in 15-ml, high-pressure Savillex Teflon bombs at 120 °C for seven days, evaporated to near dryness and then diluted to 50 ml using super-pure HNO₃ for analysis. A blank solution was prepared, and the total procedural blank was used as an internal standard to correct for matrix effects and instrument drift. Two reference standards (GSR-1 and GSR-3) were measured during the course of the analytical procedure. Precision for most trace elements was better than 5%.

4.2 Mineral major and trace elements

Major elements of rock-forming minerals were analyzed with a CAMECA SXFive microprobe analyzer at the IGGCAS. The analytical conditions were 15 kV accelerating voltage, 20 nA beam current, and 3 µm spot diameter for most minerals and 1 µm for the exsolution lamellae. The counting time for most minerals was 20 seconds on peak and 10 seconds on lower and upper background positions, respectively. The counting time for TiO₂, Al₂O₃, Cr₂O₃, MnO, and NiO in olivine and orthopyroxene was 40 seconds. The precision of all analyzed elements was better than 1.5%.

Mineral trace elements were analyzed *in-situ* on polished thin sections using a laser ablation inductively-coupled mass spectrometer (LA-ICP-MS) at the State Key Laboratory of Continental Dynamics, Northwest University. The LA-ICP-MS system consists of a GeoLas 200M laser-ablation system equipped with a 193 nm ArF excimer laser and a homogenizing imaging optical system, and an Agilent 7500a ICP-MS. The laser operated at a frequency of 8 Hz and an energy density of 14J/cm²

with a beam diameter of 60–120 μm according to grain dimensions. Counting times were approximately 20 s and 40 s for background and analysis, respectively. The GSE-1g and ^{29}Si contents were selected as the external and internal standards, respectively. The absolute content of SiO_2 was determined by in-situ EMPA (CAMECA SXFive). Three USGS glasses (NIST 610, BHVO-2g and BCR-2g) were used as monitoring standards. The external standard was analyzed after every 10 analyses. The trace element data were calculated using the GLITTER 4.0 Online Interactive Data Reduction for LA-ICP-MS program developed by GEMOC, Macquarie University. Additional analytical details and accuracy were reported in Gao et al. (2002).

4.3 Whole-rock Re–Os isotopes, PGE and S abundances

The Re–Os isotopes and PGE abundances were analyzed by the isotope dilution process at the IGGCAS, following the method of Chu et al. (2009). Whole-rock powders (~2.0 g) were digested and equilibrated with appropriate amounts of ^{187}Re – ^{190}Os mixed spike and ^{191}Ir – ^{99}Ru – ^{194}Pt – ^{105}Pd mixed spike, using inverse aqua regia (3 ml of purified concentrated HCl and 6 ml of purified concentrated HNO_3) in Carius tubes. Then the tubes were sealed and placed in an oven at 240 °C for 72 hours. Osmium was extracted from the inverse aqua regia into CCl_4 , then back-extracted into HBr , and finally microdistilled. Afterwards, Pt, Ir, Pd, Re and Ru were separated from the solution using the anion exchange resin AG-1 $\times 8$ (100–200 mesh in MQ).

Osmium isotopic compositions were analyzed by negative thermal ionization mass

spectrometry (N-TIMS) on a GV Isoprobe-T instrument in peak-jumping mode with a single electron multiplier, or in static mode on Faraday cups, depending on the amount of Os. The measured Os isotopic ratios were corrected for mass fractionation using $^{192}\text{Os}/^{188}\text{Os}=3.0827$. The Johnson–Matthey standard of UMD was used as an external standard with $^{187}\text{Os}/^{188}\text{Os}=0.11373\pm6$ for the electron multiplier and 0.11374 ± 10 for Faraday cups. The internal precisions of the $^{187}\text{Os}/^{188}\text{Os}$ ratios were better than $\pm0.2\%$ (2σ).

The concentrations of Pt, Ir, Pd, Re and Ru were measured on a Thermo-Electron Neptune MC–ICP–MS system using an electron multiplier in peak-jumping mode or Faraday cups in static mode, according to the measured signal intensity. Isotopic mass fractionations were corrected using Re, Ir, Ru, Pt and Pd standards (usually one per three sample analyses). WPR-1 (peridotite) was used as an external standard to evaluate the accuracy of this analytical method. The precision of the ICP–MS analyses was typically 0.1–0.3% (2σ). The total procedural blanks were approximately 5 pg for Re, 8 pg for Ir, 10 pg for Ru, 8 pg for Pt and 8 pg for Pd.

The total sulfur concentration was measured by an Elemental Infrared Analyser (HORIBA EMIA 220-V) at the University of Québec in Chicoutimi, Canada, following the method of Bédard et al. (2008). Five reference materials (AN-G, BE-N, BIR-1, DR-N, and MRG-1) were used as monitoring standards. The detection limit, as defined as 3σ on the blank, was 31 ppm.

5. Results

5.1 Whole-rock chemistry

The major element compositions of the investigated samples are listed in Table 1. All samples have very low LOI (Loss on Ignition), reflecting very weak degrees of serpentinization. The Maowu garnet dunites have refractory compositions with extremely high MgO (46.70–48.15 wt.%) and low CaO (0.05–0.32 wt.%), Al₂O₃ (0.26–0.76 wt.%), Na₂O (<0.02 wt.%) and TiO₂ (<0.03 wt.%) contents. The low SiO₂ (43.12–43.85 wt.%) and high Mg# [atomic Mg/ (Mg+Fe)*100, 91.9–92.0], Ni (2537–2892 ppm), and Cr (2687–3211 ppm) values are similar to depleted mantle values. In contrast, the garnet orthopyroxenites have high SiO₂, Al₂O₃, CaO and TiO₂ but low MgO contents (Table1), because they contain additional orthopyroxene, Ti-clinohumite and garnet. The K₂O and Na₂O contents are very low, due to the scarcity of phlogopite and clinopyroxene. The abundances of compatible elements Cr, Co and Ni in the Maowu peridotites are higher than those of the Bixiling cumulates in Dabie and the Maobei cumulates in Sulu, and are comparable to those of the Dabie-Sulu mantle type peridotites (Table 1, Fig. 3b). Major-element compositional trends are summarized in selected Hacker diagrams in Fig. 4. Whole-rock MgO contents show negative correlations with Al₂O₃, CaO and several incompatible elements (Zr, Sr, Y, and REEs). Moreover, the Maowu peridotites and pyroxenites define a strong trend of sharp increasing Al₂O₃ content with slight increasing CaO content (Fig. 4b-c), which is different from those induced by mantle melt-rock reactions (or refertilization) or melt extraction.

All Maowu peridotites and pyroxenites show LREE-enriched, slightly fractionated

LREE patterns (Fig. 5a). The REE pattern of garnet orthopyroxenite shows continuously decreasing LREE from La to Eu and nearly flat (or slightly increasing) middle to heavy REE (MREE-HREE). The very low HREE abundances in the Maowu dunites (0.1-0.3 of those in chondrite) suggest a refractory origin and can be archived by incremental melting of a fertile peridotite (Ionov, 2004, 2010) or higher melting degrees for ‘open-system’ melting (Godard et al., 2008). LREE-rich patterns in refractory peridotites can be due to retention of melt in the residue, or more likely, to post-melting metasomatism (see below). The garnet orthopyroxenites have high REE abundances relative to the dunites, probably due to higher proportions of apatite and garnet. Compared to primitive mantle compositions (Fig. 5b), the Maowu peridotites and pyroxenites are large-ion lithophile element (LILE)-enriched, and show spikes at U, Ba, Pb and Th, indicating that they were metasomatized by crustal fluids or melts.

The S content in the garnet dunites and the garnet-poor orthopyroxenite veins is below the detection limit (<31 ppm); however, the garnet-rich orthopyroxenite veins have a high S content (99–306 ppm).

5.2 Mineral chemistry

The representative major and trace element compositions of minerals are listed in Tables 2–4. The Fe^{3+} content in garnet, orthopyroxene and spinel is calculated based on stoichiometry.

Garnet in the dunites is almost homogeneous and pyrope-rich, with 1.67–2.78 wt.%

CaO, 2.92–4.45 wt.% Cr_2O_3 and 75.4–77.3 Mg# (Table 2). Garnet in the garnet orthopyroxenite has lower Cr_2O_3 (0.36–1.22 wt.%) than that in the dunite. However, these two types of garnet have similar Mg# and pyrope components (Table 2) and show systematically LREE-depleted and HREE-enriched patterns (Fig. 6a). Moreover, the pyroxenite garnets have relatively higher Y and HREEs than the dunite garnets.

All olivine grains in the Maowu samples are characterized by relatively high Mg# (92.2–93.4) and low MnO (0.03–0.13 wt.%), Ca (29–84 ppm), Ti (2.6–14.9 ppm) and Al (4.3–25.2 ppm). However, their NiO content shows large variations among different samples (Fig. 7). In the dunites, the matrix olivine has a normal mantle NiO (0.33–0.44 wt.%) content. In the garnet-poor orthopyroxenites, the olivine inclusions have relatively higher NiO (0.72–0.89 wt.%) values. However, the olivine inclusions in the garnet-rich orthopyroxenites have the lowest NiO (0.17–0.25 wt.%) values.

Orthopyroxene porphyroblasts (Opx-P) in the garnet dunites contain abundant exsolution lamellae of chromite; they have high Mg# (92.6–93.1) and are weakly zoned with lower Al_2O_3 (0.04–0.07 wt.%) in the cores than the rims (Al_2O_3 = 0.11–0.17 wt.%). LA-ICP-MS analysis shows that the Opx-P has very low Ni (<700 ppm) and REE contents (some below the detection limits), with a LREE-depleted pattern (Fig. 6b). The orthopyroxene inclusions (Opx1) preserved in the garnet orthopyroxenites show the same composition as the Opx-P. The matrix orthopyroxene (Opx2) in the garnet orthopyroxenites shows approximately constant composition. It has higher NiO (0.17–0.25 wt.%) but lower Cr_2O_3 (<0.05 wt.%) than the Opx-P. Unlike the Opx-P, the Opx2 is characterized by LREE enrichment and HREE

depletion (Fig. 6b). It also has relatively higher Sr, Ba and Pb contents (Table 3), probably resulting from subduction-related metasomatism.

Cr-spinel grains occurring in the matrix and as inclusions in the garnet dunites have high Cr₂O₃ (51.13–53.02 wt.%), Cr# [atomic Cr/(Cr+Al)×100] (76–88), and low TiO₂ (<0.05 wt.%) and Mg# (27–33) (Fig. 8). The compositions are similar to those in orogenic residual dunites worldwide (Su et al., 2016a). The spinel grains occurring in the garnet orthopyroxenite veins have similar compositions to those in the matrix of garnet dunites. The chromite lamellae in the Opx-P have the highest Cr# (94–97).

5.3 PGE abundances and Re–Os isotopes

The results of the PGE abundances and Re–Os isotopes of the Maowu ultramafic complex, including 3 garnet peridotites, 2 garnet-poor orthopyroxenites, and 2 garnet-rich orthopyroxenites, are listed in Table 5. The Maowu peridotite samples have high concentrations of Ir-group platinum family elements (IPGEs) (Os=1.98–4.46 ppb, Ir=1.29–2.03 ppb, Ru=4.38–6.41 ppb) and low ratios of Pd/Ir (0.38–0.46). The patterns of IPGE are flat and similar to that of primitive upper mantle (PUM, Becker et al., 2006). However, the concentrations of Pd-group platinum family elements (PPGEs) show significant depletions [Ir/(Pt+Pd) = 0.43–0.76] relative to the PUM [Ir/(Pt+Pd) = 0.24] (Becker et al., 2006). Similar PGE patterns are also observed in the Lijiatun dunites which represent highly depleted mantle peridotites after >30% partial melting (Su et al., 2016b). Garnet orthopyroxenites show large variation in PGE abundances and patterns. The

garnet-poor orthopyroxenites have similar PGE abundances to the dunites, which are characterized by relatively high IPGE concentrations with depletion of PPGEs [$\text{Ir}/(\text{Pt}+\text{Pd}) = 0.96\text{--}1.52$] (Fig. 9). However, the garnet-rich orthopyroxenites contain relatively lower IPGEs and high PPGEs [$\text{Ir}/(\text{Pt}+\text{Pd}) = 0.05\text{--}0.14$].

The present-day $^{187}\text{Os}/^{188}\text{Os}$ ratios of the Maowu garnet peridotites are subchondritic (0.11461–0.11497) and yield Re-depletion ages (T_{RD}) ranging from 2.0 to 2.1 Ga. Except for the sample MW42-2, which has the lowest $^{187}\text{Re}/^{188}\text{Os}$ ratio at 0.03 and yields a 2.26 Ga model age (T_{MA}) relative to the PUM (Meisel et al., 2001), the other two samples yield future and thus meaningless T_{MA} model ages. The dunite sample MW38-1 has a very high $^{187}\text{Re}/^{188}\text{Os}$ but a similar $^{187}\text{Os}/^{188}\text{Os}$ to the other peridotites, implying recent Re addition. The garnet-poor orthopyroxenites have similar Re–Os isotopes to the garnet peridotites. The $^{187}\text{Re}/^{188}\text{Os}$ and $^{187}\text{Os}/^{188}\text{Os}$ ratios of the garnet-poor orthopyroxenites vary from 0.37 to 0.89 and from 0.11422 to 0.11607, respectively; their Re-depletion model ages range from 1.89 to 2.14 Ga. However, both garnet-rich orthopyroxenites (MW7-2 and MW29-6) have highly radiogenic Os-isotope compositions ($^{187}\text{Os}/^{188}\text{Os} = 0.18631\text{--}0.36910$) and high Re/Os ratios (1.34–1.60) (Table 5), indicating that significant Re addition disturbed the Re–Os system in the garnet-rich orthopyroxenite.

6. Discussion

6.1. Origin of the Maowu dunites

The origin of the Maowu mafic-ultramafic body has been debated for more than

twenty years. It has been considered to be as (1) crustal-derived peridotite (Okay, 1993; Fan et al., 1996; Zhang et al., 1998, 2000; Liou and Zhang, 1998; Jahn et al., 2003), or (2) mantle-derived peridotite (Liu et al., 1998; Malaspina et al., 2006, 2009; Chen et al., 2013a, 2013b). Cumulate origin for the Maowu mafic-ultramafic body was formerly proposed on the basis of the layer structure and bulk rock trace element compositions. However, many experimental and natural sample studies have demonstrated that mantle peridotites metasomatized by crust- or mantle-derived fluids or melts also showed complex compositional layers (e.g. Iizuka and Nakamura, 1995; Rampone and Morten, 2001; Vrijmoed et al., 2006, 2013; Ackerman et al., 2009). Such layers may be due to element fractionation during the metasomatic process (Iizuka and Nakamura, 1995; Chen et al., 2015). In this regard, it is difficult to identify these two types of orogenic peridotites in a field outcrop. In addition, to distinguish the two types of orogenic peridotites based on geochemical criteria, we should carefully take into account the effect of crustal metasomatism on the bulk rock composition, especially the incompatible elements. Several mantle type peridotites with extensive crustal metasomatism could also exhibit some geochemical characteristics of crustal type peridotites (Vrijmoed et al., 2013).

Extensive attention was paid for the Maowu garnet pyroxenites, hampering real insight into the origin of this complex. Refractory orogenic dunite consists primarily of olivine, with minor orthopyroxene and spinel/garnet and experienced the weakest metasomatism by crust-derived agents among orogenic peridotites. The mineral compositions (e.g., olivine) of orogenic dunite would not significantly change during

later processes related to subduction and exhumation because little element exchange would occur between the abundant olivine and the few other minerals (Chen et al., 2015; Su et al., 2016a). Therefore, refractory dunite is the best sample from which to trace the origins of orogenic peridotites.

The new mineral and whole-rock compositions and the textural relationships presented above indicate that the Maowu dunites have an ancient SCLM origin and have experienced melt extraction and metasomatism that modified their primitive compositions.

6.1.1. Evidences from major and trace elements

Orogenic peridotites have a wide compositional range due to their complex formation processes. In the Dabie-Sulu orogenic belt, most of the orogenic peridotites belong to the mantle type, which commonly have higher Mg# values (85.9–93.8), Ni contents (1441–2926 ppm), and Mg/Si ratios (1.04–1.83). On the other hand, the crustal type peridotites (such as the Maobei and the Bixiling peridotites) have relatively lower Mg# values (<85), Ni content (760–1440 ppm), and Mg/Si ratios (0.85–1.51) (Fig. 3, see review in Chen et al., 2015). The Maowu dunites are highly refractory with high MgO, Mg# and Ni values and low FeO^{T} , Al_2O_3 , CaO and TiO_2 contents, corresponding to mantle type peridotites (Fig. 3). They also have low HREE contents and few subcalcic garnets, suggesting a refractory mantle residue.

The compositions of olivine and spinel are useful approaches to trace the dunite origin (Rehfeldt et al., 2007; Xu et al., 2010; Abily and Ceuleneer, 2013). The olivines

from the Maowu dunites have high Fo (92.2–93.4) and NiO (0.33–0.44 wt.%) values, similar to those in refractory mantle peridotites (Tatsumi, 1986) and in mantle peridotite xenoliths from the North China Craton (Zheng et al., 2006, 2008, 2014; Xu et al., 2008) but deviate from those in cumulative dunites (Su et al., 2016a) (Fig. 7). The extremely low Ca, Ti and Cr contents in olivine are also different from those in cumulative olivines (Foley et al., 2013). The compositions of Cr-spinel in the dunites plot within the field of residual dunites (Fig. 8b) and show distinctly higher Cr# and lower Mg# than those observed in abyssal peridotites. Although high Cr# spinel has also been found in cumulates of primitive melts formed from refractory mantle peridotites, it commonly shows large variation in Cr# because Cr is rapidly removed from a melt during its solidification (Barnes and Roeder, 2001). Moreover, the spinel TiO₂ contents in cumulative peridotites are typically higher (up to 0.8 wt.%) than those from the mantle type peridotites (Hattori et al., 2010; Arai et al., 2012; Rajesh et al., 2013). The consistently high Cr# values (76–88) and low TiO₂ contents (<0.05 wt.%) in the Cr-spinel from the Maowu dunites rule out the cumulative process and reflect a highly refractory nature (Pearce et al., 2000).

The Maowu dunites investigated in this study are very fresh samples, which are characterized by limited LOI (mostly <1.0 wt.%). Whole-rock chemistry shows that they are only slightly enriched in LREE and water-soluble elements (e.g. Ba, U, Pb). All these features indicate that the refractory dunites experienced slightly metasomatism after melt extraction. In this regard, whole-rock MgO, Al₂O₃ and FeO^T were not significantly modified and consequently can be used to constrain the melt

extraction process. MgO and Al₂O₃ are not affected by the melting depth (Walter, 2003; Herzberg, 2004), whereas FeO^T in the residues is sensitive to the depth (Walter, 2003). The Maowu dunites have experienced either ~40% batch melting at 3–4 GPa or >30% fractional melting from ~4 GPa to ~1 GPa (Fig. 10). Furthermore, the Maowu dunites have low FeO^T contents (5.99-7.47 wt.%, average 7.14 wt.%), corresponding to the unique characteristics of xenolith suites from Archean cratons (Griffin et al., 2009) and Archean peridotite massifs (Beyer et al., 2006). Such low-Fe dunites are clearly different from Phanerozoic mantle xenoliths, ophiolites, abyssal peridotites and orogenic replacive dunites (≥ 8.0 wt.%) formed by melt-rock reactions (Griffin et al., 2009; Su et al., 2016a).

6.1.2 Evidences from PGEs and Re–Os systematics

Using PGE systematics provides an effective approach to investigate the origin and melt extraction processes of the upper mantle. IPGEs (Os, Ir and Ru) behave compatibly during partial melting because these elements preferentially remain in residual phases, such as sulfide, oxides or alloys (Mitchell and Keays, 1981; Lorand et al., 2013). At high degrees of partial melting (>20%), sulfides are consumed by the melt, and PPGEs behave incompatibly and are released into the melt, whereas IPGEs still behave compatibly (Lorand et al., 1999; Becker et al., 2006). Therefore, mantle melts or cumulates commonly show low ratios of IPGE/PPGE, whereas the residual mantle should be enriched in IPGEs but have substantial and progressive depletions in Pt, Pd and Re, with high ratios of IPGE/PPGE (Pearson et al., 2004; Maier et al.,

2012). The two dunite samples (MW38-1 and MW42-2) show consistently high IPGE concentrations and Ir/(Pd+Pt) ratios (Figs. 8, 11), reflecting a residual origin after high degrees of partial melting. The extremely low S contents (<31 ppm) in the Maowu dunites indicate that the IPGEs may be retained in oxides or alloys and only limited refractory sulfide inclusions can be survived during such high degrees of partial melting.

Re–Os isotopes in residual peridotites can be used to constrain the timing of melt depletion. T_{RD} ages are accurate only if the peridotites had Re/Os ratios of zero following the initial melt extraction. In this case, if no Re was added subsequently, then both T_{RD} and T_{MA} ages would be the same. If a given peridotite's Re/Os ratio was non-zero, then the T_{RD} provides only a lower limit on the time of melting. Although the three peridotite samples have different Re/Os ratios (0.01–0.22), the T_{RD} ages are approximately equal (2.0–2.1 Ga), indicating that the Re-addition event was recent and did not significantly modify the radiogenic Os-isotope compositions. The dunite sample MW42-2 has the lowest Re/Os (0.01) and $^{187}\text{Re}/^{188}\text{Os}$ (0.03) ratios and yields the oldest T_{RD} (2.1 Ga). Therefore, the melt extraction event most likely occurred at ~2.1 Ga, and the Maowu peridotites represent a refractory fragment of the SCLM that formed in the Paleoproterozoic.

6.1.3 Mantle affinity

Mantle type orogenic peridotites may originate from the subducted oceanic or continental lithospheric mantle, the overlying SCLM wedge, or the asthenospheric

mantle (Brueckner, 1998; Zheng et al., 2009; Ye et al., 2009). Detailed petrological and geochemical studies indicate that most mantle type peridotites in the Dabie–Sulu orogen represent fragments of the SCLM beneath the southeastern margin of the North China Craton (NCC), which underwent metasomatism by fluids/melts derived from the subducting Yangtze continental crust (Zheng et al., 2005, 2009; Zhang et al., 2011; Zheng, 2012; Xie et al., 2013). However, Yuan et al. (2007) suggested that the Xugou peridotite originated from the SCLM of the Yangtze Craton. To date, five peridotite bodies in the Dabie-Sulu orogen have been dated by Re–Os isotopes, including Raobazhai, Xugou, Lijiatun, Ganyu (CCSD-PP3) and Maowu (Fig. 1). Somewhat surprisingly, these peridotite bodies show consistent T_{RD} ages, as follows: ~1.9 Ga for Raobazhai (Zheng et al., 2009; Jin et al., 2004), ~2.0 Ga for Xugou (Yuan et al., 2007), ~2.0 Ga for Lijiatun (Su et al., 2016b), ~2.0 Ga for Ganyu (Chen et al., 2006), and ~2.1 Ga for Maowu (this study). No Archean T_{RD} ages have been found in the Dabie-Sulu orogenic peridotites. In fact, Paleoproterozoic T_{RD} ages have been reported in the SCLMs both of the Yangtze and the North China cratons (e.g., Reisberg et al., 2005; Wu et al., 2006; Zheng et al., 2009; Liu et al., 2011). Therefore, the Re–Os isotopic data cannot be a good indicator to distinguish these two SCLMs.

Many lines of evidence support the derivation of the Maowu peridotites from the SCLM hanging wall of the NCC during continental subduction. The high IPGE/PPGE and Ir/(Pt+Pd) ratios in the Maowu dunites, as well as other Sulu orogenic peridotites (Chen et al., 2015), are consistent with the SCLM xenoliths in the NCC (Fig. 11). The refractory chemistry with olivine Fo_{92-93} in the Maowu dunites is similar to that of

relict ancient SCLM xenoliths in the NCC (Zheng et al., 2014; Su et al., 2016b) but differs from the northern margin of the Yangtze craton, which has a lower olivine Fo_{88-91} (Zhang et al., 2001; Xia et al., 2010; Liu et al., 2012). Furthermore, the Maowu peridotites are currently enclosed in continental supracrustal rocks (paragneisses) but are lacking associated lower crustal rocks. Thus, it is very difficult to use the SCLM of the Yangtze craton to interpret the mantle source of the Maowu peridotites. The NCC is very large and has been a stable craton since it was completely assembled (~1.8–1.9 Ga) (Zhao et al., 2001). The current Paleoproterozoic T_{RD} ages recorded both in the eastern SCLM of the NCC and the Dabie–Sulu orogenic mantle peridotites suggest that the NCC experienced large-scale re-working of its mantle roots (Zheng et al., 2005), and/or it was probably produced by amalgamation of different “blocks” which may have different ages. Therefore, we suggest that the Maowu peridotites were derived from the SCLM of the NCC and subsequently were scraped off by the low-density continental supracrustal rocks along a subduction channel and exhumed back to the Earth’s surface.

6.2 Orthopyroxenite vein formation by deep crustal metasomatism

Peridotites from different geological settings commonly have pyroxenite veins produced by silica-rich melt metasomatism (e.g., Suhr et al., 2003; Vrijmoed et al., 2013) or crystal precipitation from mafic melts migrating along conduits (e.g., Downes, 2007; Ackerman et al., 2009; Svojtka et al., 2016). The Maowu garnet orthopyroxenites have been well characterized by several petrological and

geochemical studies (Malaspina et al., 2006, 2009; Chen et al., 2013a, 2013b), which indicated that these orthopyroxenites were metasomatic products between refractory peridotite precursors and silica-rich hydrous melts under UHP conditions. The garnet orthopyroxenite veins presented in this study have similar metasomatic textures and mineral compositions to those investigated by previous studies, also reflecting this metasomatic process.

Compared with the wall garnet dunites, the garnet orthopyroxenite veins contain more modal amounts of the matrix orthopyroxene (Opx2), with the addition of garnet, Ti-clinohumite, and apatite. Although the veins exhibit different proportions of garnet and orthopyroxene, they commonly contain abundant metasomatic textures represented by the overgrowth of secondary orthopyroxene (Opx2), garnet and Ti-clinohumite at the expense of olivine. Only limited matrix olivine can survive in the garnet-poor orthopyroxenite veins (Fig. 2d). The Opx2 usually has high Mg# and Ni content and is LREE-enriched, which is different from the orthopyroxene porphyroblasts in the dunites (Opx-P) and the orthopyroxene inclusions (Opx1) in the orthopyroxenites. All these features indicate that the Opx2 is produced by interactions between silica-rich melt and olivine (Malaspina et al., 2006). Ti-clinohumite is a very important carrier of HFSEs and is also enriched in H₂O and LILEs (Scambelluri et al., 2006). The Ti-clinohumite in the veins shows equilibrium texture with Opx2 and also contains former mantle olivine, implying that the percolating melt is hydrous and enriched in TiO₂. Furthermore, the relict olivine inclusions in the veins have values of Fo, MnO, Al₂O₃ and CaO similar to those in the dunites, strongly suggesting that the

protolith of the garnet orthopyroxenite veins is the host dunites. However, the olivine grains in different veins show large discrepancies in NiO contents. For example, the relict olivines in the garnet-poor orthopyroxenites (bulk $\text{Al}_2\text{O}_3 < 2$ wt.%) have a very high NiO content (0.72–0.89 wt.%), whereas those in the garnet-rich orthopyroxenites (bulk $\text{Al}_2\text{O}_3 > 6$ wt.%) have a relatively lower NiO content (0.17–0.25 wt.%). This indicates that the percolating melts for the formation of two types of veins had different compositions.

Nickel is a highly compatible element in peridotites and is mainly hosted in olivine and sulfide (Kelemen et al., 1998; Kubo, 2002). If olivine reacts with a silica-rich melt to produce secondary orthopyroxene, the newly formed orthopyroxene is expected to inherit some Ni (Malaspina et al., 2006). However, most of Ni would still be retained in the relict olivine, resulting in a significant increase of its Ni content. Such high-Ni olivine relicts (up to 7700 ppm Ni) were also observed by Malaspina et al. (2006). On the other hand, if olivine reacts with a sulfur-bearing silicate melt in an open system, some sulfides are expected to precipitate from the melt when sulfur becomes oversaturated, and primitive Ni may be incorporated into the secondary sulfides or into the residual melt. The garnet-rich orthopyroxenite veins have a high S content, as evidenced by the formation of interstitial sulfides (Fig. 2h), implying that the percolating melt is also enriched in S. The low Ni content in olivine is most likely due to removal of primitive Ni from olivine into the residual melt or the secondary sulfide (Fleet et al., 1996; Hattori et al., 2010). Moreover, the garnet-rich orthopyroxenite veins also exhibit very low bulk Ni contents. The Ni contents of the

garnet-rich orthopyroxenites could be reduced by the addition of large amounts of silicate melt through a dilution effect or by removing the primitive Ni. The former process is impossible because an unrealistically large volume of melt ($>3:1$ melt/rock ratios) is required to decrease the Ni content from ~ 3000 ppm in the dunites to 997 ppm in MW25-2, which has the highest Ni content among the garnet-rich orthopyroxenites. Such high melt/rock ratios would totally consume the protolith olivine and significantly reduce the Cr contents of the garnet-rich orthopyroxenites, inconsistent with the petrological observations and the high Cr contents of the garnet-rich orthopyroxenites (up to 1844 ppm). Therefore, the primitive Ni content of the garnet-rich orthopyroxenites was most likely lost due to open-system metasomatism by a sulfur-saturated silicate melt. However, the garnet-poor orthopyroxenite veins show extremely low S content, similar to the wall dunites, most likely implying that the percolating silicate melt for the formation of garnet-poor orthopyroxenites was depleted in S. This is also evidenced by the high Ni contents in both olivine relicts (0.72-0.84 wt.%) and whole rocks (2224-3043 ppm).

The presented dataset points to the conclusion that the orthopyroxenite-forming melts were enriched in SiO_2 , Al_2O_3 , TiO_2 , REEs, and incompatible elements, with low Ca/Al ratios. Such silicate melts were most likely sourced from the country gneisses of the peridotites (Malaspina et al., 2006). Sm-Nd isochron ages, zircon U-Pb and monazite Th-Pb age data obtained from the Maowu body were constrained at 220-240 Ma (Rowley et al., 1997; Ayers et al., 2002; Jahn et al., 2003; Liu et al., 2006), corresponding to the time of continental subduction/exhumation. In addition, oceanic

crust and abyssal peridotite have restricted $^{187}\text{Os}/^{188}\text{Os}$ ratios of 0.126~0.148 (Gannoun et al., 2007) and 0.122~0.127 (Snow and Reisberg, 1995), respectively. In this regard, metasomatism by oceanic crust-derived fluids/melts cannot induce such high $^{187}\text{Os}/^{188}\text{Os}$ values (up to 0.36910) in the garnet-rich orthopyroxenites (Fig. 12a). The Maowu ultramafic complex is hosted by gneisses representing subducted upper continental crust that also experienced UHP metamorphism (Ye et al., 2000). The subducted upper continental crust has abundant hydrous minerals such as mica, chlorite and epidote. Dehydration of these hydrous minerals at subarc depths would generate aqueous solutions and hydrous melts (Hermann et al., 2006; Zheng et al., 2011), which have been widely recognized in UHP metamorphic rocks from continental subduction zones (e.g. Zheng and Hermann, 2014). These two types of metasomatic agents may be completely miscible to form a supercritical fluid (Hermann et al., 2006; Zheng et al., 2011). Hydrous melts and supercritical fluids are more efficient media to transport elements (especially HREE and HFSE) from the subducting crust to the mantle wedge (Zheng et al., 2011; Spandler and Pirard, 2013). The enrichment of HREE and HFSE in the Maowu garnet orthopyroxenites most likely points to hydrous melt metasomatism (Malaspina et al., 2006).

Although the two types of veins have different whole-rock compositions, their garnets and orthopyroxenes are almost homogeneous and have similar compositions. The peak metamorphic conditions for the veins are calculated by the Grt–Opx thermometer (Harley, 1984) and the Grt–Opx barometers (Nickel and Green, 1985; Brey and Kohler, 1990) and yield at 790–840 °C and 5.0–6.1 GPa, which are

consistent with the previous THERMOCALC calculation results (Chen et al., 2013a). This indicates that the metasomatic processes responsible for the formation of two types of orthopyroxenite veins both occurred under similar UHP conditions. The Maowu garnet dunites originating from the SCLM wedge of the NCC have opportunity to contact with the subducted upper continental crust at the slab–mantle interface. The metasomatic product, garnet orthopyroxenite, recorded a prograde P–T path characterized by isothermal compression (Chen et al., 2013a), which is different from those in the UHP continental crustal rocks. In this regard, the Maowu garnet orthopyroxenite was still located at the lower margin of the mantle wedge and was not incorporated into the continental crust during subduction. Therefore, the deep crustal metasomatism likely took place at the slab–mantle interface in the continental subduction channel.

Compared with the wall dunite, the garnet-poor orthopyroxenite only shows higher SiO₂ content with very limited addition of Al₂O₃ (Fig. 4), implying that the percolating melt is only enriched in SiO₂. Such a melt is not the typical felsic melt derived from the subducted continental crust. Therefore, the large variation of whole-rock compositions of these two types of orthopyroxenites was probably due to element fractionation during silicate melt–rock reactions, which was also revealed by several experimental and natural sample studies (e.g. Iizuka and Nakamura, 1995; Vrijmoed et al., 2013). The garnet-rich orthopyroxenite veins most likely represent the former metasomatic product because of their higher CaO, Al₂O₃, FeO and HREE and lower MgO contents (Table 1, Fig. 4), whereas the garnet-poor orthopyroxenite veins

were the later metasomatic product between the residual silica-rich melt and the wall dunite.

6.3 Effects of crustal metasomatism on the PGE signature and Re–Os isotopes of the SCLM

The PGE and Re–Os isotopic compositions of mantle rocks provide critical information on episodes of partial melting or melt–rock percolation-reaction processes in the mantle (e.g., Buchl et al., 2002; Pearson et al., 2004; Aulbach et al., 2014; Ionov et al., 2015), which are closely related to multiple generations of sulfides with contrasting Os isotopic compositions (primary residual sulfide vs. metasomatically introduced sulfide) (Alard et al., 2002; Griffin et al., 2004; Marchesi et al., 2010). Effects of mantle metasomatism on the PGE and Re–Os isotopic compositions of peridotites have been well constrained. It has been widely demonstrated that mantle metasomatic processes involving infiltration of sulfur-saturated silicate melts can lead to precipitation of metal sulfides as interstitial phases (Griffin et al., 2004; Aulbach et al., 2004; Liu et al., 2009; Alard et al., 2011). Such metasomatic sulfides typically display suprachondritic PPGE/IPGE (Luget et al., 2001, 2015; Alard et al., 2005, 2011) and are able to generate whole-rock scale enrichment in S and PPGE, and shift over time the $^{187}\text{Os}/^{188}\text{Os}$ ratios toward more radiogenic compositions (e.g., van Acken et al., 2010; Luget et al., 2015). Nevertheless, how subducted continental crust influenced the PGE and Re–Os signatures of the overlying SCLM wedge is still enigmatic (Zheng et al., 2009).

Fresh peridotites with orthopyroxenite veins from Maowu provide an opportunity to evaluate the effects of crustal metasomatism on the PGE and Re–Os isotopic compositions of the SCLM. The two garnet-poor orthopyroxenite samples (MW5-1 and MW32-2) have PGE patterns similar to the host dunites, with flat Os to Ru and depletion of Pt and Pd (Fig. 9). These features, combined with the extremely low sulfur (<31 ppm), subchondritic Pd/Ir and $^{187}\text{Os}/^{188}\text{Os}$ ratios and metasomatic textures may be best explained by the addition of sulfur-depleted silica-rich melt to the melt-depleted residual dunites. This is also evidenced by the similar PGE and Re–Os signatures of the harzburgite sample MW38-2 which is a mixture of dunite and garnet-poor orthopyroxenite. In other words, such a sulfur-depleted silica-rich melt, mostly derived from deep subducted continental crust, would not significantly modify the PGE patterns, radiogenic Os-isotope compositions or the T_{RD} ages of the SCLM, even though the crustal metasomatism was sufficiently extensive to form orthopyroxenite. However, the two garnet-rich orthopyroxenite samples (MW7-2 and MW29-6) exhibit melt-like PGE patterns, characterized by low IPGE abundances and suprachondritic Pd/Ir ratios (Fig. 9). The occurrences of interstitial sulfides, combined with the high S contents and highly suprachondritic $^{187}\text{Os}/^{188}\text{Os}$ ratios, directly point to the addition of metasomatic sulfides precipitated from a sulfur-saturated silicate melt (Alard et al., 2000, 2011; Lorand et al., 2013). The garnet-rich orthopyroxenite sample MW7-2 has the highest S content (306 ppm) and also shows the highest $^{187}\text{Os}/^{188}\text{Os}$, strengthening this point. Such high radiogenic Os isotopes give unrealistic T_{RD} ages (Table 5). The highly suprachondritic $^{187}\text{Os}/^{188}\text{Os}$ ratios (up to

0.36910) and low bulk Ca/Al ratios (Figs. 12-13) in these two samples strongly suggest that the sulfur-saturated silica-rich melt was most likely derived from the country rock gneisses of the peridotites.

The upper continental crust (UCC) has extremely low PGE abundances but very high Re/Os and Pd/Ir ratios, with an average S abundance of 621 ppm and $^{187}\text{Os}/^{188}\text{Os}$ of 1.05 ± 0.23 (Peucker-Ehrenbrink and Jahn, 2001; Rudnick and Gao, 2003). The sulfur can be released into melts or fluids via melting or dissolution processes. However, the temperatures of subducted UCC recorded by the exhumed UHP rocks are most likely below 800 °C, which is lower than the melting temperature of sulfides to form a sulfide melt (e.g. Mavrogenes et al., 2001; Wykes and Mavrogenes, 2005). Therefore, dissolution is probably the primary approach for the release of sulfur from the subducted crust into the mantle (Klimm et al., 2012). The sulfur-bearing silicate hydrous melts released from the UCC would have higher Re/Os and Pd/Ir ratios than the residual crust. Such silicate melts have highly radiogenic Os isotopic compositions and would significantly affect the SCLM wedge when they become sulfur-oversaturated to precipitate the secondary sulfides, which is known to strongly depend on the changes of oxidation state and silicate melt composition during metasomatism (e.g. Jugo, 2009; Wallace and Edmonds, 2011; Webster and Botcharnikov, 2011; Klimm et al., 2012; Lesne et al., 2015). This extensive crustal metasomatism by sulfur-saturated silicate melts under UHP conditions may modify the SCLM to form fertile garnet pyroxenites with PPGE-enriched patterns and highly radiogenic Os isotopic compositions. However, when the crust-derived silicate melt

became sulfur-depleted, such a melt would not significantly modify the PGE patterns, radiogenic Os-isotope compositions or Re-depletion ages of the SCLM. Therefore, crust-mantle interactions in deep continental subduction zones could induce high degrees of Os isotopic heterogeneity in the SCLM wedge. In the meantime, the use of Os model ages of orogenic peridotites to constrain the ages of SCLM should be approached with great caution. However, an understanding of petrography, coupled with integrated studies of mineral and whole-rock compositions, particularly including PGEs and Re–Os dating of sulfides, can be helpful to reveal the significance of Re–Os isotopic ages and complex evolution of the mantle.

7. Conclusions

Garnet peridotites in the Maowu mafic–ultramafic complex from the Dabie UHP belt provide insights into origin and metasomatism of an ancient SCLM. The Maowu garnet dunites represent a Paleoproterozoic fragment of the SCLM beneath the North China craton; they underwent ~40% melt extraction before ~2.1 Ga. During subduction of the Yangtze continental crust, the garnet dunites were incorporated into the continental subduction channel and underwent metasomatism by silica-rich hydrous melts, with local formation of garnet orthopyroxenite veins cutting the dunites. The silica-rich hydrous melts were most likely derived from the country gneisses. The PGE signature and Re–Os systematics of the SCLM wedge were variously modified by these crust-derived melts, due to the melt sulfur concentrations and the secondary sulfide precipitation. Subduction of continental crust is therefore an

important process responsible for the Os isotopic heterogeneity of the mantle.

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Figure caption

Fig. 1. Distribution of orogenic peridotites in the Dabie–Sulu orogen (modified after Zhang et al., 2009). Note that the T_{RD} ages for several refractory peridotites are also shown (see the text for more details).

Fig. 2. Microphotographs showing mineral textures in the Maowu dunites and garnet orthopyroxenite veins. (a) Hand specimen showing massive structure and garnet orthopyroxenite veins crosscutting the wall dunite. (b) Orthopyroxene porphyroblast (Opx-P) shows abundant exsolution lamellae of chromite. (c) Fine-grained garnet occurs as intergranular grains in the matrix olivine. (d) Garnet-rich orthopyroxenite vein crosscutting the wall dunite. Note that some relict olivine grains are still preserved in the orthopyroxene. (e) Garnet-poor orthopyroxenite vein crosscutting the wall dunite. (f) Olivine occurs both as inclusions in orthopyroxene and as matrix phase in the garnet-poor orthopyroxenite. (g) Relict orthopyroxene (Opx1) and olivine are included in the matrix orthopyroxene (Opx2) and Ti-clinohumite in the garnet-poor orthopyroxenite. (h) Interstitial pyrite grains occur along the boundaries of garnet and orthopyroxene in the garnet-rich orthopyroxenite. Mineral abbreviations are from Whitney and Evans (2010).

Fig. 3. Diagrams of whole-rock Mg# versus Mg/Si (a) and Ni (ppm) (b) for the Maowu peridotites. $Mg\# = 100 \times \text{molar Mg}/(\text{Mg} + \text{Fe})$. The grey shaded regions of the Dabie-Sulu mantle-derived peridotites and the Bixiling and Maobei cumulative peridotites are taken from Chen et al. (2015). Arcton, Proton and Tecton fields are

from Griffin et al. (1999).

Fig. 4. Whole-rock major element compositions. (a) and (b) MgO (wt.%) versus Al_2O_3 (wt.%) and CaO (wt.%), respectively. (c) Diagram of Al_2O_3 versus CaO for the Maowu peridotites and garnet orthopyroxenites. The composition of Primitive Mantle (PM) is after McDonough and Sun (1995). Oceanic depletion trend shows expected compositions (refractory) of residues after progressive melt extraction from fertile Primitive Mantle compositions (Boyd, 1989). Refertilization trend mimics the oceanic trend, but runs in the opposite direction. Age fields for Archean, Proterozoic and Phanerozoic are from O'Reilly et al. (2001).

Fig. 5. (a) Chondrite-normalized REE, and (b) Primitive mantle-normalized trace element compositions of the representative whole rock samples. Normalizing values are taken from McDonough and Sun (1995).

Fig. 6. Chondrite-normalized REE composition of the representative garnet (a) and orthopyroxene (b).

Fig. 7. Fo versus NiO (wt.%) of olivine. Data sources: three types of orogenic dunites (Su et al., 2016a); ancient SCLM (Zheng et al., 2001, 2006; Ying et al., 2006; Xu et al., 2010); Cenozoic SCLM (Zheng et al., 2006; Ying et al., 2006; Wang et al., 2012), abyssal peridotites (Sobolev et al., 2005). Orange star represents peridotite that has suffered 60% serial depletion (Straub et al., 2008).

Fig. 8. Plots of Mg# versus Cr# (a) and TiO_2 versus Cr# (b) for spinel. The grey fields of three types of orogenic dunites are taken from Su et al. (2016a). The melt extraction trend are from Pearce et al. (2000).

Fig. 9. Whole-rock PUM-normalized platinum group element patterns of the Maowu dunites and orthopyroxenites. Primitive upper mantle (PUM) is taken from Becker et al. (2006).

Fig. 10. Diagrams showing degrees of partial melting. (a) and (b) FeO^T variation diagrams versus MgO and Al_2O_3 , respectively. The batch melting curves of the residual peridotites follow the model of Walter (2003). (c) and (d) Plots of FeO^T and Al_2O_3 vs MgO , respectively. The fractional melting curves refer to the model of Herzberg (2004). The black stars in both models represent the compositions of fertile peridotite KR4003 (Walter, 1998). Orogenic residual dunites (white diamonds with gray frame) refer to the complete dataset of orogenic dunites formed by high degrees of partial melting of mantle peridotites (collected by Su et al., 2016a).

Fig. 11. Diagram of Ir vs. $\text{Ir}/(\text{Pt}+\text{Pd})$ for the Maowu dunites. The whole-rock PGE data sources for the Dabie-Sulu orogenic peridotites are as follows: Ganyu (Chen et al., 2006); Raobazhai (Zheng et al., 2008); Yangkou, Suoluoshu and Hujialing (Xie et al., 2013); Bixiling (Zheng et al., 2008; Liu Q et al., 2012). The whole-rock PGE data for the ancient mantle xenoliths from the North China Craton are from Gao et al. (2002), Zheng et al. (2005), Becker et al. (2006), Zhang et al. (2008), Liu et al. (2010, 2011, 2015). The field for the Laiwu cumulative dunite is after Wang et al. (2012). The fields of PGE data from the cumulative peridotites in the Kohistan and Talketna arc are after Hattori and Guillot (2007). Primitive mantle value is also shown with dark star (McDonough and Sun, 1995).

Fig. 12. $^{187}\text{Os}/^{188}\text{Os}$ vs. $^{187}\text{Re}/^{188}\text{Os}$ plot for the Maowu peridotites and pyroxenites

(note logarithmic scale of x axis). The Re-Os isotopic compositions of other peridotites from the Dabie-Sulu orogen are also shown in this diagram. Data sources: Raobazhai (Zheng et al., 2008); Xugou (Yuan et al., 2007); Ganyu (Chen et al., 2006); Lijiatusun (Su et al., 2016b). The Primitive Upper Mantle (PUM) composition is after Meisel et al. (2001) and Re-Os composition of the upper continental crust (UCC) is from Peucker-Ehrenbrink and Jahn (2001). (b) is an enlarged field for the lower part of (a).

Fig. 13. Os vs. $^{187}\text{Os}/^{188}\text{Os}$ plot for the Maowu peridotites and pyroxenites. Data sources are the same as in Fig. 12.

Fig. 1

Figure 2

Fig. 2

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Fig. 3

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Figure 13

Fig. 13

Table 1-5

Table 1. Whole-rock major element compositions of the Maowu peridotites and orthopyroxenites.

Sample No.	MW38-1	MW42-2	MW42-1	MW43	MW38-2	MW39-1	MW5-1	MW5-2	MW25-1	MW29-1	MW29-7	MW32-2	MW38-3	MW7-2	MW25-2	MW29-2	MW29-6	MW13
Lithology	Dunite				Harzburgite ^a		Grt-poor Opxenite ^b							Grt-rich Opxenite				
By XRF (wt.%)																		
SiO ₂	43.55	43.85	43.50	43.12	46.35	44.31	49.04	56.34	56.68	56.86	56.08	56.86	55.23	47.31	52.79	48.73	48.24	52.81
TiO ₂	0.01	0.02	0.01	0.01	0.11	0.01	0.02	0.04	0.09	0.08	0.09	0.01	0.15	0.24	0.03	0.47	0.33	0.02
Al ₂ O ₃	0.43	0.76	0.28	0.26	1.50	0.46	0.80	0.80	1.12	0.74	1.34	1.05	0.99	14.49	6.32	12.47	12.30	6.37
FeO _T	7.47	7.38	7.43	7.42	5.99	7.16	5.97	5.30	6.27	5.04	5.81	4.71	5.38	9.10	8.69	9.40	9.69	7.91
MnO	0.10	0.11	0.10	0.09	0.08	0.07	0.06	0.07	0.08	0.05	0.08	0.05	0.08	0.23	0.18	0.21	0.21	0.18
MgO	47.44	46.70	47.74	48.15	42.67	46.92	43.26	35.60	34.85	36.10	35.72	36.46	37.23	25.73	29.79	25.18	25.98	31.00
CaO	0.12	0.32	0.05	0.06	0.20	0.21	0.11	0.32	0.17	0.32	0.19	0.19	0.19	1.49	1.14	2.28	2.04	0.77
Na ₂ O	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.67	0.01	0.19	0.01	0.01	0.01	0.27	0.02	0.16	0.07	0.01
K ₂ O	0.01	0.01	0.02	0.01	0.02	0.02	0.02	0.06	0.02	0.02	0.02	0.02	0.02	0.04	0.03	0.04	0.03	0.02
LOI	0.46	0.26	1.48	0.30	2.18	1.08	1.54	0.94	0.30	0.94	0.80	1.56	1.40	0.40	0.08	0.26	0.32	0.06
Total	99.61	99.42	100.62	99.44	99.11	100.25	100.84	100.15	99.59	100.36	100.14	100.93	100.69	99.31	99.07	99.19	99.21	99.15
Mg# ^c	91.9	91.9	92.0	92.0	92.7	92.1	92.8	92.3	90.8	92.7	91.6	93.2	92.5	83.4	85.9	82.7	82.7	87.5
Mg/Si	1.62	1.59	1.64	1.66	1.32	1.58	1.31	0.94	0.92	0.95	0.95	0.96	1.00	0.81	0.84	0.77	0.80	0.88
By ICP-MS (ppm)																		
Cr	3211	2719	2744	2687	2603	2776	2383	3043	2224	2756	2033	3713	2776	1662	1509	1227	1617	1844
Co	119	111	116	118	98	124	114	92	85	82	74	97	90	88	105	68	73	83
Ni	2802	2537	2862	2892	2521	3011	2951	2342	1761	2354	2041	2589	2337	765	997	587	618	931

^a Harzburgite samples represent the garnet dunites containing tiny garnet-poor orthopyroxenite veins.

^b Opxenite refers to orthopyroxenite.

^c Mg# = atomic Mg/(Mg + Fe) × 100

Table 2. Representative major element compositions of minerals.

Mineral	Garnet									
Texture	Grt-matrix in dunites				Grt-matrix in Grt-poor orthopyroxenites			Grt-matrix in Grt-rich orthopyroxenites		
Sample no.	MW38-1	MW42-1	MW42-2	MW43	MW5-1	MW5-1	MW32-2	MW7-2	MW25-2	MW25-2
SiO ₂	41.78	41.79	41.46	42.04	42.01	41.76	41.24	41.43	42.51	41.93
TiO ₂	0.01	0.03	0.02	0.02	0.07	0.05	0.02	0.05	0.14	0.05
Al ₂ O ₃	21.03	19.84	20.48	20.27	23.07	22.67	22.81	22.27	22.93	22.82
Cr ₂ O ₃	3.37	4.45	2.92	3.71	0.82	1.22	0.68	0.40	0.54	0.36
FeO	11.66	12.12	11.82	11.86	11.42	11.59	11.76	11.45	10.82	11.00
MnO	0.65	0.75	0.66	0.68	0.40	0.41	0.42	0.44	0.47	0.47
MgO	19.42	19.66	19.66	20.55	19.45	20.00	19.25	19.57	20.24	19.29
CaO	2.78	2.01	2.36	1.67	3.34	2.59	3.25	3.46	2.77	3.38
NiO	0.01	0.04	0.01	0.01	0.01	0.00	0.03	0.01	0.01	0.02
Total	100.71	100.68	99.39	100.81	100.58	100.28	99.46	99.09	100.43	99.30
Si	2.998	3.012	3.010	3.008	2.991	2.989	3.001	2.990	2.997	3.009
Ti	0.000	0.002	0.001	0.001	0.004	0.003	0.001	0.003	0.007	0.002
Al	1.779	1.685	1.752	1.709	1.936	1.904	1.903	1.894	1.928	1.939
Cr	0.191	0.254	0.167	0.210	0.046	0.068	0.039	0.023	0.031	0.021
Fe ²⁺	0.667	0.697	0.658	0.648	0.651	0.644	0.655	0.592	0.611	0.647
Fe ³⁺	0.032	0.033	0.060	0.062	0.029	0.046	0.054	0.099	0.034	0.017
Mn	0.040	0.046	0.040	0.041	0.024	0.025	0.025	0.027	0.028	0.028
Mg	2.078	2.113	2.127	2.192	2.064	2.124	2.069	2.105	2.152	2.074
Ca	0.214	0.155	0.184	0.128	0.255	0.198	0.251	0.267	0.212	0.261
Ni	0.001	0.002	0.001	0.001	0.000	0.001	0.002	0.001	0.000	0.001
Mg#	75.7	75.2	76.4	77.2	76.0	76.7	76.0	78.0	77.9	76.2
Uv	9.6	12.9	8.5	10.6	2.3	3.4	2.0	1.1	1.5	1.0

Adr	1.6	1.7	3.0	3.1	1.4	2.3	2.7	4.9	1.7	0.8
Alm	19.8	19.8	19.4	18.6	20.9	20.3	20.8	18.6	19.7	21.1
Prp	61.6	60.0	62.6	62.9	66.4	67.0	65.8	66.1	69.3	67.6
Grs	6.3	4.4	5.4	3.7	8.2	6.2	8.0	8.4	6.8	8.5
Sps	1.2	1.3	1.2	1.2	0.8	0.8	0.8	0.9	0.9	0.9
Mineral	Olivine									
Texture	Ol-matrix in dunites				Ol-inclusion in Grt-poor orthopyroxenites			Ol-inclusion in Grt-rich orthopyroxenites		
	Sample no.	MW38-1	MW42-1	MW43	MW43	MW5-1	MW5-1	MW32-2	MW7-2	MW7-2
SiO ₂	41.41	40.81	41.68	41.45	41.20	41.69	41.49	41.35	40.84	41.92
TiO ₂	0.04	0.03	0.00	0.01	0.03	0.02	0.01	0.03	0.01	0.00
Al ₂ O ₃	0.01	0.00	0.01	0.01	0.00	0.00	0.01	0.00	0.00	0.01
Cr ₂ O ₃	0.00	0.01	0.02	0.00	0.00	0.00	0.00	0.06	0.02	0.01
FeO	7.35	7.50	7.39	7.34	7.18	6.97	6.90	7.49	7.85	7.14
MnO	0.05	0.13	0.08	0.06	0.09	0.07	0.07	0.07	0.07	0.11
MgO	50.82	50.19	50.70	50.89	50.27	49.83	50.23	51.35	50.99	50.82
CaO	0.01	0.01	0.01	0.01	0.03	0.01	0.02	0.02	0.02	0.01
NiO	0.41	0.39	0.45	0.41	0.72	0.77	0.80	0.18	0.21	0.25
Total	100.10	99.08	100.33	100.18	99.51	99.35	99.53	100.56	100.00	100.27
Si	1.004	1.001	1.009	1.004	1.006	1.011	1.013	0.997	0.991	1.005
Ti	0.001	0.001	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.000
Al	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Cr	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.000
Fe	0.149	0.154	0.150	0.149	0.147	0.144	0.141	0.151	0.159	0.145
Mn	0.001	0.003	0.002	0.001	0.002	0.001	0.001	0.001	0.001	0.002
Mg	1.837	1.834	1.830	1.838	1.830	1.828	1.828	1.845	1.844	1.842
Ca	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.001	0.000	0.000

Ni	0.008	0.008	0.009	0.008	0.014	0.015	0.016	0.003	0.004	0.005
Fo	92.5	92.3	92.4	92.5	92.6	92.7	92.8	92.4	92.1	92.7
Mineral	Orthopyroxene									
Texture	Opx-P ^a core	Opx-P ^a core	Opx-P ^a rim	Opx-P ^a rim	Opx1 incl. ^b in Grt orthopyroxenite			Opx2-matrix ^c in Grt orthopyroxenite		
Sample no.	MW43	MW43	MW43	MW42-1	MW5-1	MW5-1	MW32-2	MW5-1	MW7-2	MW25-2
SiO ₂	57.96	58.27	58.56	58.77	58.77	58.79	58.72	58.07	58.12	57.85
TiO ₂	0.01	0.01	0.00	0.00	0.00	0.02	0.02	0.01	0.00	0.08
Al ₂ O ₃	0.07	0.07	0.11	0.17	0.13	0.33	0.08	0.18	0.20	0.15
Cr ₂ O ₃	0.06	0.08	0.03	0.04	0.05	0.09	0.02	0.00	0.04	0.06
FeO	5.12	4.82	4.60	4.79	4.85	5.23	4.95	5.46	5.68	5.10
MnO	0.18	0.10	0.10	0.12	0.15	0.14	0.13	0.03	0.04	0.09
MgO	36.01	36.16	35.81	36.14	36.44	35.74	36.09	35.35	35.84	35.87
CaO	0.02	0.03	0.02	0.02	0.04	0.06	0.03	0.02	0.07	0.08
Na ₂ O	0.03	0.01	0.01	0.00	0.00	0.00	0.02	0.02	0.02	0.00
K ₂ O	0.00	0.00	0.00	0.01	0.01	0.03	0.00	0.00	0.02	0.00
NiO	0.05	0.08	0.07	0.06	0.05	0.07	0.04	0.25	0.19	0.17
Total	99.50	99.63	99.32	100.11	100.49	100.48	100.10	99.37	100.21	99.43
Si	1.993	2.000	2.010	2.008	2.000	2.007	2.007	2.006	1.989	1.992
Ti	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.000	0.002
Al	0.003	0.003	0.004	0.007	0.005	0.013	0.003	0.007	0.008	0.006
Cr	0.002	0.002	0.001	0.001	0.001	0.002	0.001	0.000	0.001	0.002
Fe ²⁺	0.136	0.138	0.133	0.137	0.138	0.149	0.141	0.158	0.147	0.142
Fe ³⁺	0.011	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.015	0.004
Mn	0.005	0.003	0.003	0.004	0.004	0.004	0.004	0.001	0.001	0.002
Mg	1.846	1.850	1.845	1.841	1.848	1.819	1.839	1.820	1.828	1.841
Ca	0.001	0.001	0.001	0.001	0.002	0.002	0.001	0.001	0.002	0.003
Na	0.002	0.001	0.001	0.000	0.000	0.000	0.002	0.001	0.001	0.000

K	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.001	0.000
Ni	0.001	0.002	0.002	0.002	0.001	0.002	0.001	0.007	0.005	0.005
Mg#	93.1	93.0	93.3	93.1	93.1	92.4	92.9	92.0	92.5	92.8

[illegible]

Na	0.000	0.000	0.001	0.000	0.000	0.000	0.002	0.001	0.000	0.002
K	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.000	0.000
Ni	0.004	0.004	0.001	0.003	0.004	0.002	0.001	0.006	0.006	0.006
Cr# ^d	88.1	85.6	78.4	84.4	85.2	86.8	89.4	94.3	95.1	95.4
Mg#	28.8	27.1	33.1	31.9	35.2	30.3	25.8	27.4	27.5	27.3

^a Opx-P, orthopyroxene porphyroblasts in dunites.

^b Opx1 incl., relict orthopyroxene inclusions in the matrix orthopyroxene.

^c Opx2-matrix, the matrix orthopyroxene.

^d Cr# = atomic Cr/(Cr+Al) × 100.

Table 3. Representative trace element compositions of garnet and orthopyroxene.

Sample no.	MW43	MW42-1	MW5-1	MW7-2	MW43	MW5-1	MW5-1
Texture	Grt in dunite		Grt in Grt-poor orthopyroxenite	Grt in Grt-rich orthopyroxenite	Opx-P	Opx1	Opx2
P	142	164	192	235	40	48	52
Ca	13180	13243	13468	14504	227	185	193
Sc	82	92	107	121	1.12	1.22	1.19
Ti	5.25	6.74	11.3	12.9	5.86	3.73	14.5
V	82	65	149	123	4.13	6.96	5.53
Cr	20794	20921	13501	7752	1227	1363	268
Mn	n.a.	n.a.	n.a.	n.a.	581	553	550
Co	55.8	54.7	116	123	55.5	53.3	54.8
Ni	8.75	12.6	16.0	9.79	685	654	1930
Sr	bdl	bdl	0.31	0.13	0.08	0.08	1.11
Y	19.7	18.8	29.7	33.4	0.03	0.04	0.06
Zr	3.99	5.26	6.30	8.22	0.01	bdl	0.02
Nb	n.a.	n.a.	n.a.	n.a.	0.01	0.006	0.03
Ba	n.a.	n.a.	n.a.	n.a.	0.14	0.30	5.29
La	0.02	0.01	0.02	0.01	0.003	0.004	0.02
Ce	0.05	0.06	0.03	0.03	0.01	0.01	0.03
Pr	0.02	0.02	0.02	0.02	0.002	0.002	0.004
Nd	0.10	0.10	0.21	0.18	0.02	0.02	0.02
Sm	0.52	0.45	0.99	0.84	0.007	0.007	0.008
Eu	0.44	0.50	0.57	0.43	0.002	bdl	0.002
Gd	2.23	2.80	2.52	2.47	bdl	bdl	0.01
Tb	0.56	0.59	0.55	0.53	bdl	bdl	0.002
Dy	3.41	2.95	3.66	3.75	0.009	0.008	0.01
Ho	0.62	0.71	0.71	0.77	0.002	0.002	0.002

Er	1.38	1.26	1.76	1.83	bdl	0.004	0.006
Tm	0.19	0.18	0.269	0.285	bdl	bdl	0.001
Yb	1.16	1.01	1.81	1.82	0.004	0.004	0.005
Lu	0.14	0.13	0.28	0.30	bdl	0.001	0.001
Pb	bdl	bdl	bdl	bdl	0.04	0.04	0.34
U	bdl	bdl	bdl	bdl	bdl	0.004	0.005

bdl, below detection limit.

n.a., not analysed.

Table 4. Representative trace element compositions of olivine.

Sample no.	MW38-1	MW42-1	MW43	MW43	MW5-1	MW5-1	MW25-1	MW7-2	MW29-6	MW29-6
Texture	Ol in dunite				Ol in Grt-poor orthopyroxenite			Ol in Grt-rich orthopyroxenite		
Li	2.84	1.31	2.34	2.90	3.93	3.84	3.55	3.03	2.11	4.26
B	2.19	2.22	2.19	2.27	2.84	2.63	3.38	3.37	3.40	5.01
Al	9.69	4.30	5.05	14.2	14.8	15.4	10.4	22.8	6.20	25.2
P	253	82	191	276	310	286	238	172	136	275
K	4.38	3.63	8.27	3.58	1.87	2.89	1.74	2.57	2.29	3.98
Ca	52.0	37.9	44.7	39.8	28.7	77.7	38.7	84.4	36.3	48.6
Sc	1.22	1.04	1.16	1.34	1.30	1.31	1.26	1.10	1.11	1.24
Ti	5.88	7.23	5.26	3.90	13.3	14.9	3.57	2.57	3.29	7.94
V	1.93	1.77	1.19	1.58	2.85	2.96	1.48	1.11	1.25	2.76
Cr	26.73	19.64	14.35	9.33	11.38	11.58	12.75	18.8	13.8	7.48
Mn	458	431	463	461	473	478	481	491	454	490
Co	161	161	160	161	165	164	169	167	166	163
Ni	3582	3325	3299	3179	6410	6424	5975	1722	1940	1438
Zn	41	47	41	41	42	41	49	48	48	59

Table 5. Re-Os isotopes, PGE and sulfur contents of the Maowu peridotites and orthopyroxenites.

Sample no.	MW38-1	MW42-2	MW38-2	MW5-1	MW32-2	MW7-2	MW29-6
Lithology	Dunite		Harzburgite	Grt-poor orthopyroxenite		Grt-rich orthopyroxenite	
Os (ppb)	3.29	4.46	1.98	1.87	0.60	0.05	0.24
Ir (ppb)	1.80	2.03	1.29	1.54	0.54	0.04	0.22
Ru (ppb)	5.14	6.41	4.38	3.24	1.71	0.10	0.36
Pt (ppb)	2.57	3.87	1.22	1.47	0.32	0.81	0.79
Pd (ppb)	0.83	0.81	0.48	0.14	0.04	0.13	0.78
Re (ppb)	0.74	0.03	0.13	0.15	0.11	0.08	0.33
$^{187}\text{Re}/^{188}\text{Os}$	1.081	0.032	0.315	0.374	0.888	7.920	6.491
$^{187}\text{Os}/^{188}\text{Os}$	0.11487	0.11461	0.11497	0.11607	0.11422	0.36910	0.18631
$2\sigma\%$	0.028	0.026	0.006	0.021	0.093	0.153	0.120
S (ppm)	bdl	bdl	bdl	bdl	bdl	306	99
T_{RD} (Ga) ^a	2.05	2.09	2.04	1.89	2.14	future	future
T_{MA} (Ga) ^a	-1.36	2.26	7.61	14.75	-2.02	1.89	0.56
Re/Os	0.22	0.01	0.07	0.08	0.18	1.60	1.34
Pd/Ir	0.46	0.40	0.38	0.09	0.07	3.07	3.55
Ir/(Pd+Pt)	0.53	0.43	0.76	0.96	1.52	0.05	0.14

^a They were calculated relative to the primitive upper mantle (Meisel et al., 2001) using the parameters: $\lambda_{\text{Re}}=1.666 \times 10^{-11}/\text{year}$, $^{187}\text{Re}/^{188}\text{Os}=0.423$, $^{187}\text{Os}/^{188}\text{Os}=0.1296$.

bdl, below detection limit.