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**Peridotite and pyroxenite xenoliths from the Muskox kimberlite, northern Slave craton,
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Abstract

We present petrography, mineralogy and thermobarometry for 53 mantle-derived xenoliths from the Muskox kimberlite pipe in the northern Slave craton. The xenolith suite includes 23% coarse peridotite, 9% porphyroclastic peridotite, 60% websterite and 8% orthopyroxenite. Samples primarily comprise forsteritic olivine (Fo 89-94), enstatite (En 89-94), Cr-diopside, Cr-pyrope garnet and chromite spinel. Coarse peridotites, porphyroclastic peridotites, and pyroxenites equilibrated at 650-1220 °C and 23-63 kbar, 1200-1350 °C and 57-70 kbar, and 1030-1230 °C and 50-63 kbar, respectively. The Muskox xenoliths differ from the neighboring and contemporaneous Jericho kimberlite by their higher levels of depletion, the presence of a shallow zone of metasomatism in the spinel stability field, a higher proportion of pyroxenites at the base of the mantle column, higher Cr₂O₃ in all pyroxenite minerals, and weaker deformation in the Muskox mantle. We interpret these contrasts as representing small scale heterogeneities in the bulk composition of the mantle, as well as the local effects of interaction between metasomatizing fluid and mantle wall rocks. We suggest that asthenosphere-derived pre-kimberlitic melts and fluids percolated less effectively through the less permeable Muskox mantle resulting in lower degrees of hydrous weakening, strain and fertilization of the peridotitic mantle. Fluids tended to concentrate and pool in the deep mantle causing partial melting and formation of abundant pyroxenites.

Key Words

Cratonic peridotite, cratonic pyroxenite, mantle metasomatism, thermobarometry, kimberlite xenoliths.

Introduction

Kimberlite - derived samples of the subcontinental lithospheric mantle (SCLM) are commonly used for insights on the composition and structure of the ambient SCLM (Nixon and Boyd 1973; Gurney and Harte 1980; Boyd et al. 1997a). Kimberlite-associated processes, however, may metasomatize the mantle in the vicinity of melt extraction and transport (Artemieva 2009; Kopylova et al. 2009; Doyle et al. 2004). Thus, differences found in the mantle sample in two adjacent contemporaneous kimberlites may reflect spatial compositional heterogeneity of the mantle, like those reported by Griffin et al. 1999a; Grutter et al. 1999; Carbno and Canil 2002; Davis et al. 2003; Hoffman 2003, or the varying effects of percolation of pre-kimberlitic fluids and kimberlite formation and ascent. To isolate the heterogeneity of the ambient mantle from the kimberlite-imposed mantle wall-rock metasomatism, we compared peridotites and pyroxenite xenoliths of the Muskox and Jericho kimberlites. The Muskox pipe was emplaced contemporaneously and within 15 km of the Jericho kimberlite that contains a well-studied suite of mantle xenoliths (Kopylova et al. 1999). Below we present petrography, mineralogy and thermobarometry for 53 Muskox peridotite and pyroxenite xenoliths and use their contrasting petrology to highlight interaction between pre-kimberlitic fluids and the ambient mantle.

Geological Setting

The Muskox and Jericho kimberlites belong to a mid-Jurassic cluster (172 ± 2 Ma, Heaman et al. 1997) of pipes located ~400 km NE of Yellowknife near the northern end of Contwoyto Lake (Fig. 1), emplaced in the Archean granite-granodiorite Contwoyto batholith (2589 ± 5 Ma, Van Breemen et al. 1987) of the Slave craton (Hoffman 1989). The Muskox kimberlite is made

up of volcanoclastic kimberlite, accounting for 60% of the pipe infill, and 40% coherent kimberlite. The Jericho kimberlite is located ~15 km NE of Musko and is a multiphase intrusion in 3 separate pipes (Kopylova and Hayman 2008). Jericho xenoliths include coarse peridotite (mainly low-temperature suite), porphyroclastic peridotite (mainly high-temperature suite), eclogite, megacrystalline pyroxenite, ilmenite-garnet wehrlite and clinopyroxenite (Kopylova et al. 1999). The xenoliths show the higher depletion of the shallow mantle, a layer of “fertile peridotite at 160-200 km (Kopylova and Russell 2000), the cold cratonic geotherm and a metasomatised, deformed and thermally disturbed mantle at depth below 160 km related to the lithosphere-asthenosphere boundary (Kopylova et al. 1999).

Petrography

70% of xenoliths in this study are recovered from the coherent kimberlite facies with the remaining 30% from the volcanoclastic facies. The xenoliths (2-8 cm in size) comprise 15% coarse spinel peridotite, 4% coarse spinel-garnet peridotite, 4% coarse garnet peridotite, 9% porphyroclastic peridotite (classification of Harte 1977), 60% websterite and 8% orthopyroxenite.

Coarse peridotite

Xenoliths are mainly harzburgites and span the spinel, spinel-garnet and garnet facies. Grain size correlates to sample facies, increasing from 0.5-2.5 mm in spinel facies samples to 1-10 mm in garnet peridotites. Olivine and orthopyroxene are subhedral and equant (Fig. 2A). Clinopyroxene forms smaller anhedral grains between olivine and orthopyroxene grains (Fig. 3F). Garnet is subhedral or rarely forms wispy, anhedral films on primary minerals (Fig. 3D). Subhedral garnets have variable thickness rims of small euhedral phlogopite (~200 µm) and

spinel (~10 μm) or are partially consumed by a fine-grained phlogopite- spinel aggregate. Both anhedral and subhedral garnet grains in spinel-garnet facies samples can have central inclusions of spinel (Fig. 3D, 3E). Spinel grains are irregularly shaped, anhedral, translucent brown and may be unevenly recrystallized into a cryptocrystalline aggregate of black spinel (Fig. 3A). Yellow-green amphibole is present in some spinel and spinel-garnet peridotites (5-10%) and can have prominent orthopyroxene rims (Fig. 3B). Phlogopite is rare, with one exceptional sample containing 10-15% phlogopite in parallel veins replacing orthopyroxene and olivine (Fig. 3C). Samples show minor serpentinization and deformation of olivine expressed as undulatory extinction and subgrain development, these features increase with the depth facies. Some coarse peridotites show recrystallization of olivine into finer grains and development of lattice-preferred orientation associated with the presence of amphibole and phlogopite. Samples contain veins and patches of a cryptocrystalline aggregate of serpentine, phlogopite and carbonate (Fig 2E). Carbonate is often present as infiltrating veins and as small patches of small equigranular rounded-hexagonal grains.

Coarse spinel peridotite contains 30-40% olivine (0.5-2.5 mm), 30-40% orthopyroxene (0.5-2.5 mm), $\leq 10\%$ spinel (200-600 μm) and 0-10% amphibole (< 2 mm), and minor amounts of clinopyroxene (< 2 mm), phlogopite and cryptocrystalline serpentine. Coarse spinel-garnet peridotite contains 30-40% olivine (40 μm -4 mm), 30-40% orthopyroxene (800 μm -3.5 mm), 5% garnet (600 μm -2 mm), spinel (100-800 μm) and clinopyroxene (1-3 mm) and minor amphibole and secondary serpentine. Coarse garnet peridotite is made up of 50-70% olivine (1 mm-10 mm) and lesser (10-20%) orthopyroxene (< 2 mm) with 10% garnet (1 mm-4 mm) and 5% clinopyroxene (1 mm).

Porphyroclastic peridotite

Porphyroclastic harzburgites with rare dunites are composed mostly of olivine porphyroclasts (60-70%) with olivine neoblasts (5-20%) and lesser orthopyroxene (0-10%), garnet (5%) and clinopyroxene (0-5%). Olivine porphyroclasts are large (8-20 mm), subhedral to anhedral grains (Fig. 2D) with undulatory extinction, deformation lamellae, subgrains and inclusions of orthopyroxene and garnet. Olivine is 10-100% replaced by serpentine with fine-grained magnetite and carbonate. Olivine neoblasts (10-150 μm) are polygonal isometric or tabular (Fig. 3G), rarely fully envelope olivine porphyroclasts and are 0-100% replaced by serpentine or a cryptocrystalline aggregate of serpentine + phlogopite + magnetite $\pm$ carbonate $\pm$ spinel. Orthopyroxene are large (2-4 mm) euhedral rectangles, rounded inclusions in olivine porphyroclasts (1 mm), or anhedral grains (<1 mm) between olivine porphyroclasts. Orthopyroxene is serpentinized and show variable undulatory extinction. Garnet forms large (1-4 mm), euhedral-subhedral grains in triple junctions of olivine porphyroclasts and orthopyroxene (Fig. 2D) or smaller inclusions in olivine porphyroclasts. Garnet often contains small dark inclusions interpreted to be partial melt; garnet rims are variably replaced by euhedral phlogopite (200 μm -0.5 mm) and spinel (10-80 μm). Garnet inclusions in olivine have a thin serpentine rim or a thick halo of olivine neoblasts. Clinopyroxene (0.2-1 mm) occurs as small anhedral-subhedral grains between larger olivine porphyroclasts, orthopyroxene and garnet.

Pyroxenite

Pyroxenitic rock types include websterites and rare orthopyroxenites, divided based on the clinopyroxene modes and mineral chemistry.

Websterites show allotriomorphic textures resulting from intergrown anhedral pyroxenes with curvilinear shapes typical of simultaneous magmatic crystallization. Their texture contrasts

the long-equilibrated metamorphic textures of coarse peridotites. Websterites are composed of orthopyroxene (20-70%) and clinopyroxene (5-40%) with olivine (0-40%), garnet (5-20%) and rarely spinel (2-5%). Rock types are dominantly pure websterites with rare olivine websterites. Some websterites are characterized by large (0.5 mm - 4 cm) euhedral-subhedral orthopyroxene hosting monogranular networks of anhedral wormlike clinopyroxene and garnet (100 μ m-1 cm) developing along certain crystallographic directions of host orthopyroxene (Fig. 2E). Clinopyroxene contains numerous dark rounded inclusions along grain boundaries and fractures, interpreted to be melt inclusions. Occasional subhedral garnet grains are partially melted and partly replaced by tabular pleochroic phlogopite (<200 μ m) and cubic spinel (10-40 μ m) or cryptocrystalline aggregates of these minerals. Olivine is rare, found as irregular grains with curvilinear grain boundaries. Websterites contain 5-10% patches and veins of a dark brown to opaque black cryptocrystalline aggregate probably composed of serpentine, magnetite, phlogopite, carbonate and spinel (Fig. 2E). The aggregate could be zoned, with magnetite or carbonate in the center and zones of serpentine and then phlogopite in the periphery. We interpret the cryptocrystalline patches as altered garnet grains.

Orthopyroxenite is always deformed, composed of orthopyroxene porphyroclasts (Fig. 2F) (90%) and neoblasts (5%) with minor olivine (<1%), replaced by serpentine (5-20%) phlogopite <1%) and secondary spinel (<1%). Porphyroclasts (200 μ m-1 cm) are 10-60% serpentinized, show undulatory extinction and development of subgrains and are surrounded by neoblasts (Fig. 3H). Orthopyroxene neoblasts (10-20 μ m) are sometimes replaced by serpentine, phlogopite and spinel. Anhedral olivine (<200 μ m) is present in interstices of orthopyroxene porphyroclasts.

Analytical Methods

Mineral compositions were analyzed using a fully automated Cameca SX-50 electron microprobe at the Earth and Ocean Science Dept. at the University of British Columbia. Samples were analyzed in wavelength dispersion mode at accelerating voltage of 15 mV and beam current of 20 mA. On-peak counting times were 10 s for major and 20 s for minor elements. Raw data were treated with the 'PAP' $\phi(\rho Z)$ on-line correction program. Individual phases in samples were analyzed as 15-20 points in cores and rims over 3-5 grains. Homogenous analyses were averaged, zoned mineral analyses were averaged separately as cores and rims (Table S1). Precision (2σ , relative %) and minimum detection limits (absolute wt. %) as follows: SiO₂ (0.7, 0.03); Na₂O (3, 0.02); MgO (1, 0.05), Al₂O₃ (2, 0.08); CaO (1.5, 0.02); TiO₂ (22, 0.03); Cr₂O₃ (9, 0.04); MnO (27, 0.03); FeO (3, 0.04); NiO (20, 0.03).

Mineral Chemistry

Olivine

Olivine Mg-number (Mg/(Mg+Fe) cpm) is highest (92-94) in coarse spinel and spinel-garnet peridotite; values are lower (89-91) for coarse garnet and porphyroclastic peridotite. Pyroxenite shows a wide distribution of Mg# (88-93; Fig. 4C) with higher values for orthopyroxenites.

Orthopyroxene

Orthopyroxene is enstatite, with more restricted Mg# than those in olivine, in the range 91-94 in peridotites and 90-92 in pyroxenites (Fig. 4B, D). Mg# in peridotitic orthopyroxene decreases with depth, from ~92-93 wt. % in coarse spinel and spinel-garnet peridotite, to 91-92 in coarse garnet and porphyroclastic peridotites. Orthopyroxene in orthopyroxenites ranges to

higher Mg# (92) than in websterites. In all depth ranges and rock types, Mg# is higher in orthopyroxene than in coexisting olivine. Zoning can be found in orthopyroxenes in all rock types except coarse garnet peridotite, showing a rimward decrease in Al_2O_3 (2 wt. %), Cr_2O_3 (0.5 wt. %) and CaO (2 wt. %) and an increase in MgO (3 wt. %). Exceptions to these ranges exist for metasomatized samples, where values can be variably raised or lowered in tandem or separately.

Al_2O_3 content of orthopyroxene was used to determine the depth facies of peridotites where garnet or spinel could not be found, based on lower Al_2O_3 content of orthopyroxenes in equilibrium with garnet (Boyd et. al. 1997). The Al_2O_3 content threshold varies from 0.6-0.9 wt. % in Udachnaya (Boyd et. al. 1997) to 0.6-0.8 wt. % in Jericho (Kopylova et al. 1999; McCammon and Kopylova 2004) and 0.5-0.6 in the Gahcho Kue xenoliths (Kopylova et. al. 2004). Al_2O_3 in orthopyroxene in all Muscox peridotites varies from 0.35 wt. % to 3.0 wt. %. We assigned the garnet facies to Muscox peridotites that contained orthopyroxenes cores with less than 0.9 wt. % Al_2O_3 .

Clinopyroxene

Clinopyroxene in all rock types is Cr-diopside (1.0-3.0 wt. % Cr_2O_3), with the highest values of Cr_2O_3 in websterites and the lowest in coarse spinel peridotites (Fig. 5B). Mg# values are highest in coarse peridotites (90-96) and lowest in pyroxenites (89-91) (Fig. 5B). Chromium is correlated with Na_2O . Al_2O_3 is higher in peridotitic clinopyroxene (2.0-5.0 wt. %) and lower and more restricted in pyroxenites (~2.0 wt. %). Zoning can be found in clinopyroxene in all rock types except for coarse garnet peridotite, expressed as rimward decreases in Al_2O_3 (2.5 wt. %), Cr_2O_3 (0.25 wt. %) and Na_2O (0.6 wt. %) and increases in MgO (1 wt. %), CaO (3 wt. %) and FeO (0.9 wt. %).

Garnet

Cr-pyrope plots within the 'G9' field of the Cr_2O_3 -CaO diagram (Fig. 6). Cr_2O_3 is highest for porphyroclastic peridotites (4.0-12.0 wt. %) and lowest for coarse spinel-garnet peridotites (1.5-4.0 wt. %). Websteritic garnet is the most calcic with 4.5-8 wt. % CaO, compared to 4.5-6.5 wt. % CaO in peridotitic garnet. Garnet compositions follow two trends, the 'lherzolitic' (Sobolev et al., 1973) along the G9/G10 boundary, and a trend of lower Cr_2O_3 enrichment with increasing CaO (Fig. 6) for a subset of websterites (Fig. 6). Mg# values are the highest and variable most in porphyroclastic peridotites (76-82), lower in pyroxenites (76-79) and intermediate in coarse peridotites (77-81) (Fig. 5C). Garnets are often zoned, with rimward increases in Al_2O_3 (<2 wt. %), FeO (1 wt. %) and MgO (1 wt. %) and decreases in Cr_2O_3 (2 wt. %). The zoning is most pronounced in websterites.

Spinel

Spinel is spinel-chromite containing 0-0.05 wt. % TiO_2 , 30-55 wt. % Cr_2O_3 , 15-33 wt. % Al_2O_3 and 11-16 wt. % MgO. It occurs only in coarse peridotites and demonstrates the common negative correlation of MgO and Cr_2O_3 . The low Cr_2O_3 compositions of the spinel place it below the diamond inclusion field (Gurney and Zweistra 1995), with a single outlying analysis with only 13 wt. % Cr_2O_3 .

Amphibole

Amphibole is pargasite, with 18-20 wt. % MgO, 10-13 wt. % CaO, < 3 wt. % Na_2O , 3 wt. % FeO and 1 wt. % Cr_2O_3 . Grains are homogenous and demonstrate higher Al_2O_3 in coarse spinel peridotites (13 wt. %) than in garnet peridotites (10 wt. %).

Thermobarometry

Compositions of homogeneous grains were preferentially used for thermobarometry; in samples that showed core-rim heterogeneities, rim compositions were employed. Because the Muskox mantle xenolith suite is diverse mineralogically, we devised different approaches to deal with rocks that lack different minerals.

For samples containing orthopyroxene, clinopyroxene and garnet, combined Brey and Kohler (1990) two-pyroxene temperature (BK T) and orthopyroxene-garnet pressure (BK P) can be computed (Table 1). This thermometer is calibrated in natural lherzolite for the conditions 900-1400 °C and 10-60 kbar, with an accuracy of ± 16 °C and ± 2.2 kbar (Brey et al. 1990). This thermobarometric combination was used because it is recommended for natural lherzolite compositions at P-T conditions (Smith 1999) comparable to that expected for the shallow mantle beneath the Muskox kimberlite, and has been widely applied in similar contexts. This BK thermobarometry allows us to make comparison to the Jericho geotherm which has also been computed with the same method (Kopylova et al. 1999). The BK T/BK P points for Muskox xenoliths match the Jericho geotherm very well (Fig. 7A). For further thermobarometry we therefore assumed that these spatially proximal and roughly contemporaneous pipes share the same thermal regime.

For samples containing two pyroxenes, but lacking garnet (Table 2) it was difficult to estimate the equilibrium pressure. We computed Brey and Kohler (1990) two-pyroxene temperatures (BK T) at two assumed pressures and projected the resulting univariant P-T line onto the Jericho geotherm (Figs. 7, 8). The geothermal intercept is the most likely pressure and temperature of the sample equilibrium (Table 2).

For samples lacking clinopyroxene, it is not possible to compute BK T two-pyroxene temperatures and other thermometers consistent with BK T should be employed. We have tested the match of the Brey and Kohler (1990) Ca-in-Opx thermometer against BK T (Fig. 7B, Table 1). The BK Ca-in-Opx values are 40-80 °C cooler than those reported by the BK T method for pyroxenites, and deviate ~30 °C above and below BK T for two spinel-garnet peridotites (Fig. 7B). The match is better than between BK T and O'Neill and Wood (1979) olivine-garnet temperatures that deviate up to 200 °C (Smith 1999; Kopylova and Caro 2004). Therefore, we estimated BK Ca-in-Opx temperatures for 2 assumed pressures for each clinopyroxene-free sample (Table 3). The pressure and temperature of the intersection of this line with the Jericho geotherm (Fig. 8, 9) constraints the equilibrium conditions for the samples.

For samples without orthopyroxene, but containing coexisting clinopyroxene and garnet, we have investigated the use of the Nakamura (2009) clinopyroxene-garnet thermometer (NK T). Nakamura (2009) thermometer (NK T) is based upon Fe-Mg exchange between coexisting garnet and clinopyroxene for mafic to ultramafic bulk compositions at 800-1820 °C and 15-75 kbar, with an accuracy of ± 74 °C. NK T has been plotted against BK T to check consistency (Fig. 4C and Table 1). The two thermometers are shown to be in good agreement, with NK T reported as 0-80 °C cooler than BK T for the same samples, with the exceptions of two deviating samples (Fig. 4C). We thus computed the intersection of the univariant P-T line of Nakamura (2009) temperatures with the Jericho geotherm (Table 3) to estimate the depth of origin for orthopyroxene-free samples.

Thermobarometric estimates computed by the above methods for peridotites and pyroxenites are compiled on Fig. 8 and 9, where samples with independently derived pressures and temperatures are plotted as single symbols and univariant P-T lines are shown intersecting

the Jericho geotherm. Finding the intersection points was straightforward for the steady-state, shallow geotherm that can be approximated as a curve or a short linear segment, but in the thermally-disturbed asthenosphere (Kopylova et al. 1999) this projection is not possible and a range of feasible P-T conditions is reported in Tables 2 and 3 instead.

Depth positions of xenoliths from Fig. 8 and Fig. 9 constrain the Muskox mantle cross-section (Fig. 10A) in comparison with the Jericho mantle (Fig. 10B). The Muskox mantle shows a consistent change with depth from coarse spinel peridotite (75 to 135 km) to coarse spinel-garnet peridotite (150 to 165 km) and to porphyroclastic peridotites (185 - 230 km). There is a sampling gap at 150 - 165 km, which occupies the transition to coarse garnet peridotite. Pyroxenites occur from 165 km to 215 km, coexisting with both coarse garnet peridotite and porphyroclastic peridotite. The pyroxenites are present at greater depth than eclogites at Muskox, 120-170 km (Fig. 10A and Kopylova et al. 2015).

Thermobarometry combined with mineral chemistry enables conclusions on the presence of low-T and high-T suites of peridotites commonly distinguished in the cratonic mantle (Harte and Hawkesworth 1989; Pearson et al. 2003). Most coarse peridotites and two porphyroclastic peridotites with magnesian olivines and orthopyroxenes (with $Mg\# > 92$ on Fig 4) can be classified as low-T. The rest of porphyroclastic peridotites (2 samples) and two coarse peridotites that plot together with pyroxenites on Fig. 5 can be considered high-T. The scarcity of high-T samples makes it impossible to comment on the P-T conditions of their formation.

Discussion

Depletion of the Muskox peridotitic mantle

The peridotitic mantle beneath the Muskox kimberlite is generally more depleted than that below the Jericho kimberlite. Mg# in peridotitic olivine in Muskox samples has a higher mode (93) and ranges to higher values (94) compared to Jericho samples (Fig. 4A). Mg# in orthopyroxene reflects a similar pattern, with a mode at 93-94 in Muskox samples and a lower Mg# of 93 at Jericho (Fig. 4B). This relationship can be seen in Mg# depth profiles for olivine (Fig. 11A) and orthopyroxene (Fig. 11B), with Mg# generally showing higher values at all depths in the Muskox samples except at 130-150 km and >200 km. A histogram of Cr₂O₃ in garnet (Fig. 12B) shows this depletion as well, as most Muskox garnets lie at or above 5 wt. %, whereas most Jericho analyses lie at or below 5 wt. %. To our knowledge, the contrast in the garnet and olivine mineral chemistry of peridotites at the same depth found at Jericho-Muskox has not been reported before for other kimberlites from a single cluster. Similar detailed depth profiles of mineral compositions are not available for other kimberlites sampling the same mantle segment within a short time period. Contrasting values for Mg# in peridotitic olivine and CaO content in garnet have been shown for peridotitic xenoliths from different Ekati pipes (Menzies et al., 2004) and from the main and satellite pipes at Letseng (Lock and Dawson 2004), but these peridotites do not originate from the same depth. At Ekati and Letseng, the peridotites with contrasting mineral chemistry are related to the stratified mantle with varied lherzolite/harzburgite ratios and to sampling of different depth intervals (Lock and Dawson 2004).

Two depth zones of metasomatism

Several MuskoX peridotites stand out from the rest of the suite. These samples show textural evidence of reequilibration and recrystallization, crystallization of late metasomatic minerals and zoned or outlying mineral compositions. We interpret these signs as evidence for mantle metasomatism at two depth intervals, a shallow zone at 130-150 km, and a deep zone at >200 km, discussed below.

Shallow zone of metasomatism

The shallow zone of metasomatism (130-150 km) is characterized petrographically by samples showing any of the following features (Table 4): 1) the presence of primary euhedral amphibole texturally equilibrated with surrounding minerals (Fig. 3D); 2) replacement of orthopyroxene by secondary anhedral phlogopite via a network of parallel veins (Fig. 3C); 3) spinel rimmed by garnet (Fig. 3D, E); 4) recrystallization of primary spinel into finer-grained secondary spinel (Fig. 3A); and 5) development of the lattice-preferred orientation (LPO) of smaller olivine grains. These metasomatized samples are characterized by low-Cr spinel, an increase in TiO₂, Na₂O, Cr₂O₃ and Al₂O₃ in pyroxenes as compared to other MuskoX peridotites, and less magnesian olivine (Fig. 11A) and orthopyroxene (Fig. 11B).

Metasomatism in the amphibole stability field is developed on the Kaapvaal craton, as observed in xenoliths from pipes in the Kimberley area, but is absent under the Slave and extremely rare under the Siberian (Solov'eva et al. 1994) cratons, the other two long-mined and well-studied kimberlite provinces. The Kaapvaal peridotites are altered and recrystallized to phlogopite–K-richterite-bearing peridotites (terminology of Erlank et al. 1987), which are thought to be formed by interaction of MARID-related hydrous fluids with peridotitic wall rocks (Dawson and Smith 1977; Erlank et al. 1987; Waters et al. 1989; Konzett et al. 2000; Winterburn

et al. 1990). Based on the stability field of amphibole, these peridotites probably originate at a maximum of 80-90 km depth under water-undersaturated conditions (Wallace et al. 1991). There are diverging opinions on the origin of the agent of metasomatism. It has been proposed to result from subduction and orogenesis (Sen et al. 1994, Koornneef et al. 2009) or to kimberlite melts (Griffin et al. 1996). Influx of hydrous fluid into the mantle, inferred as the source of metasomatism on and off cratons (O'Reilly et al. 1991), should change the strength of the lithosphere and its rheological properties. So-called "hydrous weakening" of the lithosphere should lead to enhanced recrystallization of olivine (Mei and Kohlstedt 2000). Corresponding with this model, metasomatism in the shallow Muscox mantle is also associated with recrystallization of olivine into finer grains and development of LPO. Influx of the hot fluid may cause transient heating (e.g. Jamtveit and Yardley 1997), which would act to enhance strain (Karato 1993). The subsequent cooling would be evidenced by garnet coronas on spinel and low-Al and Cr rims of zoned orthopyroxene equilibrated in the garnet stability field.

Deep zone of metasomatism

A deep zone of metasomatism starts at 200 km depth; we cannot constrain the deeper termination of the zone as the in the Muscox kimberlite did not sample below 220 km. The zone is manifested in development of late garnet and clinopyroxene along olivine-orthopyroxene grain margins (Fig. 3F) and the occurrence of Al- and Cr-zoned garnet, less magnesian olivine and orthopyroxene and Ti-rich pyroxenes (Table 4).

The deep zone of metasomatism is seen in plots of depth vs. Mg# of both olivine (Fig. 11A) and orthopyroxene (Fig. 11B), with a trend decreasing from ~92-93 at shallower depths (100-150 km) to ~89 at 200 km depth. These rare high-T coarse garnet peridotite samples demonstrate lower Mg# in olivine and orthopyroxene than porphyroclastic peridotites and

contain clinopyroxene with higher Al_2O_3 , TiO_2 , NaO and FeO and orthopyroxene with higher TiO_2 . Such mineral chemistry is typical for the more prevalent websterites found at this depth (Figs. 7, 8, 9A).

The metasomatized sample with interstitial clinopyroxene and garnet films contains primary orthopyroxene with higher CaO, compared to the sample without late development of clinopyroxene and garnet. This suggests metasomatic addition of CaO rather than the origin of garnet and clinopyroxene as exsolution phases from a higher-T orthopyroxene in a closed system, and could relate to refertilizing metasomatism often reported in cratonic mantle (Simon et al. 2007; Miller et al. 2014).

Post-craton stabilization metasomatism at the base of the SCLM below 190 km is common under cratons, is characterized by an increase with depth in FeO, CaO, Al_2O_3 and TiO_2 in bulk composition and mineral chemistry, and a decrease in olivine Mg# and may represent infiltration of asthenosphere-derived mafic melts and fluids (O'Reilly et al. 2010 and references therein). The deep metasomatic zone in the Muskox peridotitic mantle may be a continuation of the “fertile layer” found in the Jericho peridotitic mantle (Kopylova and Russell 2000) at 160–200 km (marked by arrow on Fig. 11). The fertility of peridotites at the base of the Jericho SCLM may result from the intrusion of and reaction with younger pyroxenitic magmas and related fluids (Kopylova and Russell 2000). A lower proportion of these fertile coarse peridotites in the Muskox sample suite may suggest a lower degree of penetration and fertilization by these pyroxenitic magmas and fluids.

Deformation of the Muskox peridotitic mantle

Porphyroclastic or ‘sheared’ peridotites are found at the deepest levels of the Muskox sample suite, co-existing with coarse peridotites and pyroxenites (Fig. 10A). Boyd and Gurney

(1987) interpret sheared and high-T peridotites as representing the beginning of melting in the presence of volatiles at the lithosphere-asthenosphere boundary. Kennedy et al. (2001) suggest that deformed peridotites accommodate localized shearing along a partially coupled lithosphere-asthenosphere boundary.

Within the suites of sheared peridotites from MuskoX and Jericho, strain may be higher at shallower depths. This is suggested by a negative correlation between the degree of shearing and the estimated depth of the samples. MuskoX porphyroclastic peridotites with less olivine neoblasts and less strain report minimum Cr-in-garnet pressure estimates (Grutter et al. 2006) of around ~45 kb, while more deformed samples with 20% neoblasts yield minimum pressures of around 30 kb. The MuskoX peridotite with 5% olivine neoblasts is sourced from $P=50$ kb, it contains the more chromian (10 wt.% Cr_2O_3) garnet. This pattern fits the Jericho sample suite. There peridotites showing extensive recrystallization of olivine and garnet (porphyroclastic disrupted peridotites) derive from depths of 170 km (Table 3 in Kopylova et al. 1999) and have only about 3 wt.% Cr_2O_3 in garnet, while less sheared samples (porphyroclastic non-disrupted peridotites) with no deformation of garnet or pyroxenes derive from depths of 165-195 km (Table 3 in Kopylova et al. 1999) and have the most chromian garnet (11 wt.% Cr_2O_3). An analogous relationship is seen in peridotites from pipes in the Kimberley area on the Kaapvaal craton (Boyd et al. 1987). We relate the higher strain at shallower depths to the higher deviatoric stress in the colder shallow mantle (Chu et al. 2012). This is corroborated by the smaller grain size of MuskoX coarse spinel peridotite compared to coarse garnet peridotites (Fig. 1), as grain size will decrease with increasing stress (Karato and Wu 1993). Another factor in the stronger deformation of the shallower MuskoX peridotite could be its dunitic mineralogy, as olivine is known to deform more readily than orthopyroxene (Harte 1977 and references therein).

Muskox peridotites with the most sheared textures contain tabular neoblasts (Fig. 3G), in contrast to other porphyroclastic peridotites with isometric neoblasts derived from 170-190 km. The tabular shape of neoblasts suggests fluid-assisted grain boundary migration (Drury and Van Roermund 1989) as the mechanism for static annealing. This fluid-assisted grain boundary migration must be restricted to shallower depth due to localized fluid penetration. The latter may have played a role in strain localization as the presence of water is known to decrease the effective viscosity of olivine (Mei and Kohlsted 2000).

Sheared peridotites from Jericho show more deformation than sheared peridotites from Muskox. In Jericho peridotites, stronger pyroxenes and garnet are deformed in mosaic, fluidal and disrupted textures (Kopylova et al. 1999). By contrast, Muskox peridotites show only moderately strained porphyroclastic textures with no more than 20% olivine recrystallization and non-deformed garnet and pyroxenes. There is a positive correlation in the degree of shearing and clinopyroxene modes between the Muskox and Jericho sample suites. Muskox deformed peridotites contain a maximum of 20% neoblasts and $\leq 5\%$ clinopyroxene, whereas the Jericho deformed peridotites (70-90% neoblasts) can have up to 10% clinopyroxene. This general correlation suggests that infiltration of metasomatizing melts or fluids that introduced CaO and triggered clinopyroxene formation (Simon et al. 2007; Miller et al. 2014) is possibly linked to deformation. The relationship between metasomatizing fluids and deformation of peridotitic cratonic mantle has been observed widely (Smith and Boyd 1987; Drury and van Roermund 1989; O'Reilly and Griffin 2010; Harte and Hawkesworth 1989). The presence of hydrous fluids has been shown experimentally to lower the effective viscosity of the mantle and enhance strain in both the diffusion and dislocation creep regimes (Skemer et al. 2013, Wang 2010). Refertilization events that formed late clinopyroxene and garnet may have had a larger impact in

the Jericho mantle than in the mantle below Muskox. The additional evidence for this is the presence of the ‘fertile’ peridotite layer at Jericho (Kopylova and Russell 2000), which is largely absent at Muskox.

Characteristics of the Muskox pyroxenitic mantle

Two types of pyroxenites occur in the Muskox mantle, orthopyroxenites and websterites. Both show unequilibrated textures (Figs. 1E, F) and therefore must be younger than the texturally equilibrated coarse peridotitic mantle with metamorphic textures (Goetze 1975; Kopylova et al. 1999). Textural disequilibrium in websterites is manifested in complex curvilinear grain boundaries (Fig. 1E) and widespread exsolution lamellae, while orthopyroxenites have sheared textures unequilibrated with respect to a wide range of grain sizes (Fig. 1F). A higher proportion of minerals in pyroxenites show chemical zoning than those in peridotites. Zoning is expressed in 1) A rimward decrease in CaO and increase in MgO and Na₂O in garnet; 2) a rimward decrease in Al₂O₃, Cr₂O₃ and CaO in orthopyroxene; and 3) a rimward increase in Al₂O₃, Cr₂O₃, CaO and decrease in MgO and Na₂O in clinopyroxene. The preserved zoning suggests that the pyroxenites formed within a short period of time (on the order of million years, Ivanic et al. 2012) before kimberlite sampling. The younger textures of the Muskox pyroxenites contrast with equilibrated, granoblastic textures of Siberian pyroxenites (Solovjeva et al. 1987) and Proterozoic (1.84 ± 0.14 Ga; Aulbach 2009) pyroxenites from the central Slave. Mineral compositions of the two pyroxenite groups are more Cr-rich than the respective minerals in the surrounding peridotitic mantle (Fig. 5). Commonly, pyroxenitic minerals are less chromian and magnesian than peridotitic minerals, typical for the general relationship between mafic and ultramafic rocks. At Muskox, less magnesian pyroxenitic minerals are more chromian (Fig. 5), which cannot be explained by a simple distinct degree of melting compared to

worldwide pyroxenites. High chromian pyroxenitic minerals, such as those found in the Muskox pyroxenites, have not been reported elsewhere.

Comparison with Jericho pyroxenites (Kopylova et al. 1999) may give us insights on formation of the Muskox pyroxenites. The following traits are common to pyroxenites from both pipes: 1) the presence of young, magmatic-textured websterites; 2) the occurrence of pyroxenites at restricted depth approaching the Jericho lithosphere-asthenosphere boundary, where they coexist with both coarse and sheared peridotites; 3) the thermal equilibration of pyroxenites along the ambient cratonic geotherm.

The following traits distinguish Muskox pyroxenites from Jericho pyroxenites: 1) the presence of two distinct types of pyroxenite in the Muskox sample suite; 2) more chromian compositions of all minerals; 3) the absence of rock types transitional to peridotites with respect to modal mineralogy and mineral chemistry (Fig. 5); 4) the wider range of formation pressures and temperatures (Figs. 9, 10); and 5) a higher proportion of these rock types in the Muskox xenolith suite.

Cratonic pyroxenites could be samples of locally pyroxene-enriched peridotitic mantle (Pearson et al. 1993), crystallized deep-seated mafic melts or their cumulates (e.g. Irving 1980) related in origin to a subducting slab (Foley et al. 2003), products of a reaction between peridotite and silicic melt (Rapp et al. 1999; Aulbach et al. 2002; 2009 Mallik and Dasgupta 2012), or metasomatic products of pre-kimberlitic melts with wallrock peridotites (Kopylova et al. 2009). Solving the origin of the pyroxenites would require additional studies of trace element, radiogenic and stable isotope compositions of the peridotitic and pyroxenitic phases, and cannot be dealt with in the framework of the present manuscript. Currently available data on the major element chemistry of Muskox mantle xenoliths enable only limited constraints on the origin of

the Muskox pyroxenites. They cannot be samples of the locally pyroxene-enriched peridotitic mantle, as their mineral chemistry differs from that of peridotites. The pyroxenites are unlikely to form in a reaction with peridotite, as evident from the lack of mineral compositions transitional between peridotite and pyroxenite (Fig. 5). Distinct mineral compositions of pyroxenites and their magmatic, young textures suggest they may be crystallized pockets of mafic melts. These melts were atypically Cr-rich. An Al- and Ca-poor variety of this Cr-rich melt crystallized into orthopyroxenites, with low garnet and clinopyroxene modes. The extremely high Cr₂O₃ content of orthopyroxenes in the orthopyroxenite can be partly explained by its relic high-pressure, high-temperature composition, which did not allow for exsolution of clinopyroxene component. The evidence for this is occurrence of clinopyroxene as lamellae in orthopyroxene and along grain boundaries and a negative correlation between the mode of clinopyroxene and the Cr₂O₃ content in orthopyroxene. The deep origin of the orthopyroxenites matches their high calculated equilibrium P-Ts (Fig. 9). These melts may have originated in response to the presence of fluids near the lithosphere-asthenosphere boundary. Timing of pyroxenite formation suggests that these fluids may have been proto-kimberlitic or shortly preceding kimberlite formation.

Concluding remarks

Xenoliths from the Muskox kimberlite emplaced simultaneously with the adjacent Jericho kimberlite provide the opportunity to examine the same mantle sampled by a different ascending magma batches and thus separate the ambient mantle characteristics from the traits imposed by the kimberlite formation and emplacement. Kimberlite sampling led to a more complete representation of the mantle column in the Jericho xenoliths, while the Muskox kimberlite did not sample in the depth range 150-165 km (Fig. 10) and did not entrain 4-phase

porphyroclastic peridotites that allow to check for a possible deviation from the ambient geotherm that marks the thermal lithosphere-asthenosphere boundary. Thus, the thickness of the lithosphere beneath MuskoX is unknown.

Besides sampling contrasts, the following distinctions are found when comparing more detailed features of sample suite from MuskoX against those from the Jericho kimberlites: 1) a slightly higher overall level of depletion in most depth ranges of the MuskoX peridotitic mantle; 2) metasomatism in the spinel facies shallow mantle at MuskoX; 3) the less deformed mantle expressed in peridotite textures resulting from lower strain and a smaller proportion of these rock types in the overall sample suite; 5) weaker refertilization of the peridotitic mantle; 6) a higher proportion of pyroxenites in the MuskoX sample suite and their wider depth distribution, and 6) the presence of 2 distinct varieties of pyroxenites with uniquely Cr-rich minerals.

The slightly higher depletion of the MuskoX mantle may represent the natural heterogeneity of the mantle, which typically shows a range of mineral and geochemical compositions (Walter 2003). All other features can be ascribed to the effect of kimberlite sampling, namely, to metasomatism with pre-kimberlitic fluids. The eclogitic component of the MuskoX and Jericho mantle is similar in bulk composition, mineralogy, origin and depth distribution (Kopylova et al. 2015) and cannot explain the distinct characters of the mantle suites from these two pipes. Metasomatic recrystallization of peridotite reacting with fluid to produce pyroxenes, garnet and megacrysts has been reported by Rawlinson and Dawson (1979), Boyd et al. (1984), Doyle et al. (2004) and Burgess and Harte (2004). This model of multi-stage recrystallization and metasomatism of peridotite wallrocks adjacent to percolating fluid was proposed for the origin of Jericho megacrysts and pyroxenites (Kopylova et al. 2009). This could be the same widespread type of metasomatism that introduced secondary clinopyroxene, garnet

and ilmenite ($\pm$ phlogopite and rutile) to cratonic xenoliths (e.g. Erlank et al. 1987, Gregoire et al. 2002; Simon et al. 2007, Rehfeldt et al. 2008). Our observations on mineral zoning in pyroxenites, textural disequilibrium of pyroxenites and phlogopite in peridotites agree with the conclusion that the metasomatism shortly preceded kimberlite formation (e.g. Dawson and Smith 1977; Hops et al. 1989; Konzett et al. 2000) and may be genetically related to pre-kimberlitic fluids (Konzett et al. 2000; Doyle et al. 2004; Moore and Belousova 2005; Kopylova et al. 2009).

Vagaries of fluid penetration may have led to metasomatism of the shallow mantle at Muskox, as well as the common concentration of fluid near the lithosphere-asthenosphere boundary causing formation of pyroxenites and sheared peridotites. Lower levels of strain and modal and cryptic metasomatism of peridotites suggest that these fluids have infiltrated and interacted less with the Muskox mantle than with the Jericho mantle, perhaps due to lower permeability of the mantle. Instead, the fluids have pooled and triggered local partial melting causing formation of abundant pyroxenites. Their distinct high-Cr character we ascribe to a distinct compositions of the Ti-, Al-, Fe-, Si- and Ca-rich Muskox metasomatizing fluids. The absence of ilmenite occasionally present in Jericho peridotites and pyroxenites (Kopylova et al. 1999) suggests the less Ti-rich character of the Muskox metasomatism. This, in turn, prevents Cr buffering by ilmenite (Kopylova et al. 2009) and ensures the chromian affinity of the pyroxenite minerals at Muskox. The variability in Cr_2O_3 contents of pyroxenitic minerals beneath various kimberlites parallels the range of Cr_2O_3 contents observed in high-Cr megacrysts and in thresholds between low-Cr and high-Cr megacrystal suites in individual kimberlites (Schulze 1987). The latter may imply the metasomatic agents variously endowed in Cr.

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Figure Captions

Fig. 1. Geologic map of the Jericho property showing kimberlite bodies after Couture and Michaud, 2004. Area shown is at the northern end of Contwoyto Lake, ~400 km NE of Yellowknife, NWT.

Fig. 2. Thin section photomicrographs. (A) Plane polarized light (PPL). Coarse spinel peridotite composed of equant olivine and orthopyroxene grains and smaller, light brown translucent spinel. There is minimal serpentinization of all phases along grain boundaries. (B) PPL. Coarse spinel-garnet peridotite composed of equant olivine and orthopyroxene grains, light pink garnet and dark opaque spinel grains. Spinel is often hosted as an inclusion inside larger garnet grains. (C) PPL. Coarse garnet peridotite composed of large equant olivine, orthopyroxene and garnet grains showing moderate serpentinization of all phases along grain boundaries and in fractures. (D) PPL. Porphyroclastic peridotite composed of large olivine porphyroclasts and garnet grains partially surrounded by networks of olivine neoblasts. Garnet is filled with small dark inclusions of crystallized partial melt. (E) Cross polarized light (XPL). Garnet websterite composed of a large, single orthopyroxene grains intergrown with single crystals of anhedral clinopyroxene. A large dark patch in the center of the photograph is a cryptocrystalline aggregate of serpentine, phlogopite, spinel and carbonate possibly replacing garnet. (F) (XPL) Orthopyroxenite composed of a large single orthopyroxene grain, recrystallized into subgrains and small neoblasts along boundaries of larger grains. Mineral abbreviations here and in Fig. 3 are OL – olivine; OP – orthopyroxene; CP – clinopyroxene; GA – garnet; SP – spinel; AMP – amphibole; PH – phlogopite. All scale bars are 4 mm.

Fig. 3. Thin section photomicrographs. All images in plane polarized light. (A) A coarse spinel peridotite contains translucent brown spinel grains recrystallized into a dark opaque cryptocrystalline aggregate. (B) A coarse spinel peridotite contains large subhedral amphibole grains rimmed by orthopyroxene. A large, dark brown translucent spinel grain is in the lower right hand corner. (C) A coarse spinel peridotite contains phlogopite infiltrating the sample via a network of subparallel veins, forming large anhedral grains and resorbing orthopyroxene and olivine. (D) A coarse spinel-garnet peridotite contains light green, euhedral amphibole in textural equilibrium with olivine and orthopyroxene. Anhedral garnet-rimmed spinel is interstitial to other primary minerals. (E) A coarse spinel-garnet peridotite contains dark opaque spinel as an inclusion inside texturally equilibrated garnet. (F) A coarse garnet peridotite contains partially molten clinopyroxene with spongy margins as an anhedral interstitial grains between olivine and orthopyroxene. (G) A porphyroclastic peridotite contains large olivine porphyroclasts and garnet grains surrounded by small, isometric olivine neoblasts. Olivine in the center of the photograph has recrystallized into larger, tabular neoblasts. (H) An orthopyroxenite contains large orthopyroxene grains recrystallized into smaller subgrains and tiny neoblasts along boundaries of grains and subgrains. All scale bars are 1 mm; the bars in (A) and (H) are 200 μm .

Fig. 4. Histograms of Mg# (molar $\text{Mg}/(\text{Mg}+\text{Fe})$) for olivine (A) and orthopyroxene (B) in Muskox and Jericho peridotites and for olivine (C) and orthopyroxene (D) in Muskox and Jericho pyroxenites. Here and further Jericho data are from Kopylova et al. (1999).

Fig. 5. Compositions of Muskox minerals plotted as Cr_2O_3 wt. % vs. Mg# (molar $\text{Mg}/(\text{Mg}+\text{Fe})$) for (A) orthopyroxene, (B) clinopyroxene and (C) garnet. Muskox mineral compositions plotted

as symbols. Jericho mineral compositions plotted as fields, i.e. Jericho coarse peridotite (red line); Jericho porphyroclastic peridotite (blue line); Jericho pyroxenite (green line); Jericho low-Cr megacrysts (labelled); Jericho high-Cr megacrysts (labelled).

Fig. 6. Compositions of Muskox garnet plotted as Cr_2O_3 wt. % vs. CaO wt. %. Jericho garnet compositions (Kopylova et al. 1999) plotted as labelled fields. Fields for G9, G10 and eclogitic garnets are from Grutter et al. (2004).

Fig. 7. (A) Pressure-temperature estimates for Muskox peridotites and pyroxenites calculated by Brey and Kohler (1990) Al-in-orthopyroxene barometer (BK P) and Brey and Kohler (1990) two-pyroxene thermometer (BK T). For this figure and figures below, a solid line is the BK P/BK T Jericho geotherm (Kopylova et al. 1999). Here and further the solid line marked G/D is the graphite-diamond transition line (Kennedy and Kennedy, 1976). (B) Brey and Kohler (1990) two-pyroxene temperatures plotted against Brey and Kohler (1990) Ca-in-orthopyroxene temperatures computed for the same samples at the same BK pressure. Here and in (C), the solid line marks the 1:1 relationship between the temperatures. (C) Brey and Kohler (1990) two pyroxene temperatures plotted against Nakamura (2009) temperatures computed for the same samples at the same BK pressure.

Fig. 8. Equilibrium pressure-temperature estimates for Muskox peridotites (A) and pyroxenites (B) according to Brey and Kohler (1990). Solid line is the geotherm constrained for xenoliths from the Jericho kimberlite, calculated using the combined BK P/BK T (Kopylova et. al., 1999). Samples plotted as single points have both temperature and pressure calculated using the BK

P/BK T solution. Points joined by lines are calculated for two assumed pressures and shown to intersect the Jericho geotherm. Points joined by dashed lines use the BK Ca-in-Opx thermometer, points joined by solid lines use the BK T two-pyroxene thermometer. Pressure-temperature estimates for Jericho peridotites (Kopylova et al., 1999) are plotted as fields. Dashed univariant lines not connected to symbols are for 3 orthopyroxene-free porphyroclastic peridotites calculated with the Nakamura (2009) thermometer.

Fig. 9. Finer scale view of equilibrium pressure-temperature estimates for Muskox pyroxenites according to Brey and Kohler (1990). Solid line is the geotherm constrained for xenoliths from the Jericho kimberlite, calculated using the BK P/BK T (Kopylova et al., 1999). Samples plotted as single points have both temperature and pressure calculated using BK P/BK T solution. Points joined by lines are calculated for two assumed pressures and shown to intersect the Jericho geotherm. Points joined by dashed lines use BK Ca-in-Opx thermometer, points joined by solid lines use BK T two-pyroxene thermometer. Pressure-temperature estimates for Jericho pyroxenites (Kopylova et al., 1999) are plotted as a field.

Fig. 10. Depth distribution of rock types in the Muskox (A) and Jericho (B) mantle. Depths for Muskox samples are from Figs. 7, 8 and geothermal intercepts and BK P/BK T thermobarometric solution in Tables 1-3. Depths for Jericho samples are from BK P/BK T thermobarometric solutions (Kopylova et al., 1999). Depths of origin for Jericho and Muskox eclogites were computed using Nakamura (2009) intercepts with the Jericho geotherm and are based on 148 xenoliths reported in Kopylova et al., (accepted pending revisions).

Fig. 11. Depth profiles for Mg# (molar Mg/(Mg+Fe)) in Muskox and Jericho minerals. (A) olivine in peridotite; (B) orthopyroxene in peridotite; (C) olivine in pyroxenite; (D) orthopyroxene in pyroxenite. Depth estimates for Muskox samples are calculated as described in the text. Depth estimates and mineral compositions for Jericho samples are taken from Kopylova et al., (1999) and Kopylova and Russell (2000). Double arrow at 160-200 km is the localization of the “fertile” layer at Jericho manifest in the Ca- and Al-rich bulk composition of the mantle (Kopylova and Russell, 2000). The propagated error on Mg# of minerals from the microprobe analyses is $\pm .003$ Mg#, i.e. within the size of the symbol. The standard error of barometry is ± 6.6 km.

Fig. 12. Histograms of Cr₂O₃ (wt. %) for spinel (A) and garnet (B) in Muskox and Jericho peridotites.

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Table 1. Pressure and temperature estimates of Muskox samples containing two pyroxenes and garnet

Sample #	Rock Type	Combined		Ca-in-OPX T (°C) @ BK P	Nakamura T (°C) @ BK P
		BK P (kbar)	BK T (°C)		
MOX 3 33	Coarse Spinel Garnet Peridotite	42.1	860	824	862
MOX 7 62.38	Coarse Spinel Garnet Peridotite	32.1	739	771	670
MOX 1 45.5	Coarse Garnet Peridotite	61.2	1205	1132	1132
MOX 24 124.00A	Porphyroclastic Peridotite	56.7	1208	1169	1095
MOX 0 157.45	Websterite	58.8	1169	1103	1097
MOX 0 162.80A	Websterite	60.6	1206	1121	1188
MOX 0 162.80B	Websterite	59.6	1198	1122	1176
MOX 0 179.3	Websterite	60.0	1195	1130	1152
MOX 0 197	Websterite	59.6	1163	1085	1164
MOX 0 198.37	Websterite	57.1	1168	1104	1148
MOX 0 216.83	Websterite	61.1	1171	1098	1130
MOX 1 59	Websterite	58.6	1183	1113	1136
MOX 3 42.1	Websterite	59.7	1177	1117	1175
MOX 11 122.2	Websterite	57.9	1172	1114	1174
MOX 11 200.50A	Websterite	61.8	1163	1115	1129
MOX 24 43.35	Websterite	59.1	1183	1102	1195
MOX 24 206.73	Websterite	56.7	1174	1134	1147
MOX 24 240	Websterite	60.8	1193	1135	1086
MOX 31 230.74	Websterite	63.4	1174	1136	1168
MOX 31 281.2	Websterite	59.2	1204	1143	1163

For Tables 1-3: BK P - Brey and Kohler (1990) orthopyroxene-garnet barometer

BK T - Brey and Kohler (1990) two-pyroxene thermometer

BK Ca-in-Opx T- Brey and Kohler (1990) Calcium in orthopyroxene thermometer

Nakamura - Nakamura (2009) clinopyroxene-garnet thermometer

Geothermal Intercept - Intersection of calculated univariant line with geotherm constrained
for xenoliths from the Jericho kimberlite (Kopylova, 1999)

Table 2. Pressure and temperature estimates for Muskox samples containing two pyroxenes

Sample #	Rock Type	Combined		Combined		Geothermal Intercept	
		Assumed P ₁	BK T (°C) @ P ₁	Assumed P ₂ (kbar)	BK T (°C) @ P ₂	Pressure (kbar)	Temperature (°C)
MOX 3 69.25	Coarse Spinel Peridotite	25	809	40	831	35.6	824
MOX 24 31.10A	Coarse Spinel Peridotite	25	736	35	749	30.0	743
MOX 24 65.16	Coarse Spinel Peridotite	30	798	40	811	34.0	800
MOX 24 124.00B*	Coarse Spinel Peridotite	50	1130	60	1150	58.0	1145
MOX 25 161.52B	Coarse Spinel Peridotite	25	811	40	832	36.0	825
MOX 31 224.5 ⁺	Coarse Garnet Peridotite	55	1191	65	1217	50-62.8	1178-1213
MOX 3 74.42	Websterite	55	1194	65	1214	62.5	1211
MOX 3 78.85B	Websterite	55	1173	65	1194	61.0	1183
MOX 7 30	Websterite	50	1115	60	1135	57.1	1128
MOX 7 54	Websterite	55	1174	65	1194	60.9	1186
MOX 24 42.6	Websterite	60	1192	70	1214	61.4	1195
MOX 24 105.2	Websterite	60	1207	70	1228	62.9	1213
MOX 25 120.6A	Websterite	55	1162	65	1183	60.0	1170
MOX 25 161.52C	Websterite	60	1196	70	1217	62.0	1200
MOX 31 242.5	Websterite	55	1123	60	1133	56.7	1126
MOX 31 242.6	Websterite	60	1210	65	1221	63.5	1218
MOX 1 48.6	Orthopyroxenite	55	1192	65	1213	62.5	1208
MOX 24 31.1B	Orthopyroxenite	55	1192	65	1213	62.5	1208
MOX 28 269.6	Orthopyroxenite	55	1113	60	1123	55.9	1113

*Coarse Spinel Peridotite MOX24 124.0B is not plotted on Fig. 7, because orthopyroxene composition is unusually low-Al and was considered to be re-equilibrated by late processes.

⁺ High-T sample, geothermal intercept reported as possible range of intersections with disturbed Jericho geotherm. Here and Table 3.

Table 3. Pressure and temperature estimates for Muskox samples lacking (A) clinopyroxene or (B) orthopyroxene

A.		Combined		Combined		Geothermal Intercept	
Sample #	Rock Type	Assumed P ₁ (kbar)	BK Ca-in- Opx T @ P ₁	Assumed P ₂ (kbar)	BK Ca-in- Opx T @ P ₂	Pressure (kbar)	Temperature (°C)
MOX1 107.57	Coarse Spinel Peridotite	35	835	45	876	37	842
MOX7 97.68	Coarse Spinel Peridotite	35	861	45	902	39.5	877
MOX24 84.8	Coarse Spinel Peridotite	20	635	30	670	23	645
MOX11 162.1	Porphyroclastic Peridotite ⁺	55	1161	65	1209	50-61	1135-1190
MOX3 78.85A	Websterite**	45	1013	55	1058	50	1035
MOX25 120.60B	Orthopyroxenite	55	1207	65	1256	65	1256
B.		Combined		Combined		Geothermal Intercept	
Sample #	Rock Type	Assumed P ₁ (kbar)	Nakamura T (°C) @ P ₁	Assumed P ₂ (kbar)	Nakamura T (°C) @ P ₂	Pressure (kbar)	Temperature (°C)
DDH2 91.5	Porphyroclastic Peridotite ⁺	50	1228	70	1340	52-70	1250-1340
MOX25 124.8	Porphyroclastic Peridotite ⁺	50	1156	70	1264	50-62	1140-1200
MOX28 320.1	Porphyroclastic Peridotite ⁺	50	1114	70	1217	50-59	1110-1145
MOX0 158.8A	Websterite**	40	986	60	1080	50	1033
MOX0 233.5	Websterite**	40	1050	60	1152	58	1150
MOX11 287.67	Websterite**	40	998	60	1094	52	1050
MOX24 255	Websterite**	40	1099	70	1258	64	1230
MOX25 246.19	Websterite**	40	1063	60	1167	60	1167

**Clinopyroxene or orthopyroxene data not available due to extensive kimberlite alteration of xenolith.

Table 4. Evidence for metasomatism in Muskox peridotite xenoliths

Shallow Zone				
Sample	Rock Type	Depth (km)	Petrographic Features	Chemical Features
MOX7 62.38	Coarse Spinel-Garnet Peridotite	95	Amphibole; garnet-rimmed spinel	
MOX24 31.1A	Coarse Spinel Peridotite	99	Recrystallization of spinel	Low (36.09 wt. %) Cr ₂ O ₃ in spinel High (2.23 wt. %) Na ₂ O in clinopyroxene
MOX3 69.25	Coarse Spinel Peridotite	117	Amphibole rimmed by orthopyroxene	Low (35.28 wt. %) Cr ₂ O ₃ in spinel; High (2.11 wt. %) Na ₂ O in clinopyroxene; High (0.22 wt. %) TiO ₂ in clinopyroxene;
MOX25 161.5B	Coarse Spinel Peridotite	119	Phlogopite; recrystallization of olivine	High (1.23 wt. %) CaO in orthopyroxene
MOX7 97.68	Coarse Spinel-Garnet Peridotite	130		Low (13.76 wt. %) Cr ₂ O ₃ in spinel; Low Mg# in olivine (89.5) and orthopyroxene (89.8)
MOX3 33.00	Coarse Spinel-Garnet Peridotite	139	Garnet rimmed spinels	Core-rim depletion in Al ₂ O ₃ in orthopyroxene (0.5 wt.%)
MOX24 124.00B	Coarse Spinel Peridotite	153		High (1.23 wt. %) TiO ₂ in clinopyroxene
Deep Zone				
Sample	Rock Type	Depth (km)	Petrographic Features	Chemical Features
MOX1 45.5	Garnet Peridotite	202		High (0.11 wt. %) TiO ₂ in orthopyroxene; High (0.24 wt. %) TiO ₂ in clinopyroxene; Low (90) Mg# in olivine; Low (91) Mg# in orthopyroxene
MOX31 224.5	Garnet Peridotite	207	Interstitial clinopyroxene and garnet films	High (0.11 wt. %) TiO ₂ in orthopyroxene; High (0.23 wt. %) TiO ₂ in clinopyroxene; Low (89) Mg# in olivine; Low (91) Mg# in orthopyroxene