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Mantle xenocrysts from the Arkhangelskaya kimberlite (Lomonosov mine, NW Russia): Constraints on the composition and thermal state of the diamondiferous lithospheric mantle

M. Lehtonen [a,*], H. O'Brien [a], P. Peltonen [a], I. Kukkonen [a], V. Ustinov [b], V. Verzhak [c]

[a] Geological Survey of Finland, P.O. Box 96, FI-02151, Espoo, Finland
[b] ALROSA Co Ltd., 128-A Nevsky pr., 193036 St. Petersburg, Russia
[c] ALROSA Co Ltd., 4/7 Kuznechihinskyi pr., 163045 Arkhangelsk, Russia

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ABSTRACT

The Arkhangelskaya kimberlite pipe belongs to the Zolotitsa kimberlite field in the Arkhangelsk region, NW Russia. It is the first pipe of the Lomonosov diamond mine to be put into production, with 2 million tons of ore already extracted. In this study major and trace element compositions of garnet, clinopyroxene (Cpx), Mg-ilmenite and chromite xenocrysts from the Arkhangelskaya pipe have been used to infer information about the compositional variability of the mantle underlying the Zolotitsa field. Single-grain thermobarometry of peridotitic Cpx xenocrysts yields a cool cratonic geotherm that follows a ca. 36 mW/m^2 conductive model. Equilibration temperatures of garnet and chromite grains based on Ni- and Zn-thermometry, respectively, indicate a sampling interval of ca. 70–230 km of the lithospheric mantle when projected onto the Cpx-derived geotherm. The major element chemistry of Mg-ilmenite xenocrysts suggests that almost optimal redox conditions for diamond preservation prevailed in the mantle during the time of emplacement of the host kimberlite magmas. Garnet major and trace element compositions combined with the Cpx-geotherm indicate that the peridotitic diamond window extends from 130 to 210 km under Zolotitsa and that the deeper parts of the lithosphere have been affected by metasomatic events. Arkhangelskaya seems to have sampled the bulk of its diamonds from the deepest portion of the diamond stability field, between 190 and 210 km. In comparison, the neighbouring Lomonosova and Pionerskaya pipes are known to have collected their diamonds from 130–160 km. The comparable grade of the three pipes suggests that diamondiferous material is generously distributed within the diamond stability field. The remarkable difference evidenced by garnet composition and thermobarometry between Arkhangelskaya and the two other Zolotitsa pipes probably derives from differences in rheology and eruption rates of the rising kimberlite magmas.

1. Introduction

The kimberlites of the Zimnyi Bereg (Winter Coast) area in Arkhangelsk, NW Russia (Fig. 1), belong to the Late Devonian, 360–380 Ma, Arkhangelsk Alkaline Igneous Province (Mahotkin et al., 2000; Pervov et al., 2005 and references therein). In addition to kimberlites, this province comprises a wide variety of ultramafic rock-types, including alkaline picrites and olivine lamproites (Mahotkin et al., 2000; Garanin 2004; Downes et al., 2005). Based on petrographical and geographical characteristics the kimberlites can be divided into two groups (e.g. Beard et al., 2000; Mahotkin et al., 2000; Sablukova and Sablukov, 2008; Ustinov et al., 2008): mica-poor Eastern Group (Kepino-Pachuga and Verkhotina-Soyana fields) and micaceous Western Group (Zolotitsa and Mela fields) that resemble South African Group I and II kimberlites, respectively, in terms of mineralogy and chemical composition. Both groups host an economic diamond deposit: the Lomonosov deposit consisting of 6 individual pipes on a 9.5 km N–S chain belongs to the western Zolotitsa field whereas the Grib pipe (Rubanova et al., this issue) is located in the eastern Verkhotina-Soyana field. Studies of inclusions in diamonds and carbon isotope work have shown that that majority of the diamonds in the Arkhangelsk kimberlites are of harzburgitic paragenesis (Sobolev et al., 1973, 1997; Galimov et al., 1994). The crystals are mostly rhombododecahedrons and octahedrons, the latter form being more common among the smaller grain sizes, <1.0 mm (Garanin et al., 1998a).

Kimberlites of both groups contain megacrysts (Mahotkin et al., 2000; Kostrovitsky et al., 2004) and other lithosphere-derived materials, including xenoliths of peridotite (Sablukova, 1995; Bobrov et al., 2003; Malkovets et al., 2003a; Sablukova et al., 2003) and eclogite (Bobrov et al., 2003; Malkovets et al., 2003b) and abundant mantle xenocrysts (Sobolev et al., 1992; Sablukov et al., 1995; Garanin

* Corresponding author. Tel.: +358 20 550 2183; fax: +358 20 550 12.
E-mail address: marja.lehtonen@gtk.fi (M. Lehtonen).

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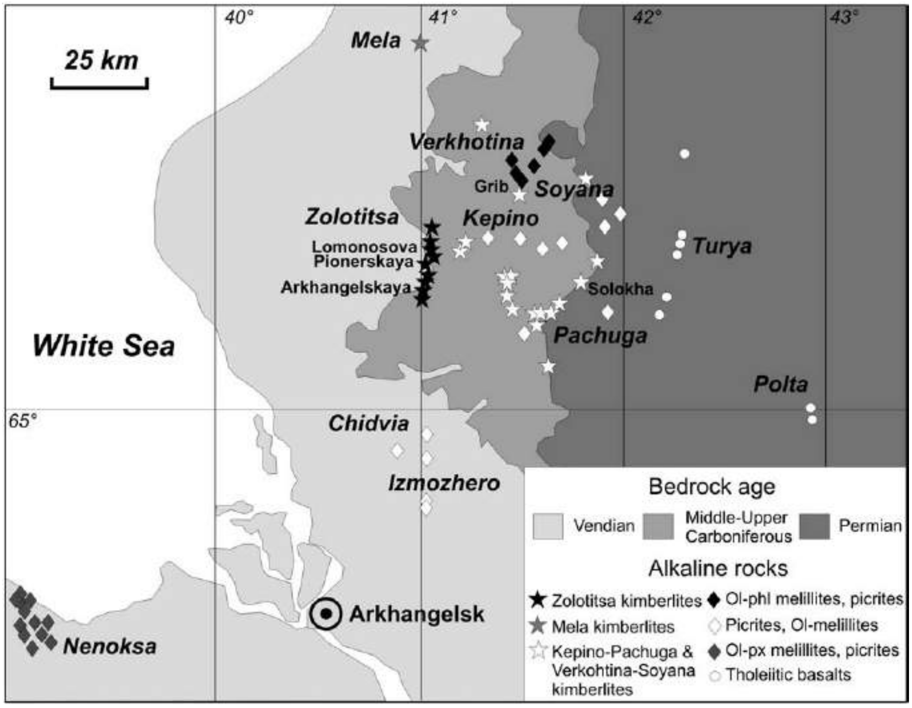

Fig. 1. Simplified geological map of the Arkhangelsk alkaline igneous province. Redrawn after Krotkov et al. (2001).

et al., 1998b). Crustal xenoliths of granulite (Markwick and Downes, 2000) and eclogite (Malkovets et al., 2003b; Bobrov et al., 2005) facies rocks have been also recovered from the kimberlites.

Significantly, mantle xenoliths in the Lomonosov pipes are extremely altered in contrast to those at Grib which are more abundant, fresher and useful for extracting petrological information. For this reason, most of the xenolith P–T estimates of the Zimnyi Bereg area come from Grib, with peridotite xenoliths yielding 31–72 kbar and 600–1200 °C, corresponding to a 37–38 mW/m^2 conductive model geotherm of Pollack and Chapman (1977; Fig. 4) (Malkovets et al., 2003a; Sablukova et al.; 2003). Clinopyroxene-garnet pairs have yielded a very similar geotherm, filling many discontinuities seen in the xenolith data (Kostrovitsky et al., 2004; Fig. 4). The P–T conditions for mantle eclogites from Grib have been determined to be 35–39 kbar and 950–1050 °C (Bobrov et al., 2003). The P–T parameters for the rare pristine mantle xenoliths from Zolotitsa range between 17–53 kbar and 800–1200 °C (Sablukova, 1995). Based on single-grain Cr-pyrope thermobarometry (Ryan et al., 1996) the geotherm follows a 37 mW/m^2 conductive model (Sablukov et al., 1995).

The focus of this work is on the Arkhangelskaya kimberlite pipe that belongs to the Lomonosov diamond mine, being the first pipe put into production. The aim is to obtain additional information on the stratigraphy, compositional variability and evolutionary history of the lithospheric mantle underlying Zolotitsa by studying mantle xenocryst minerals and using the methodology described by Griffin et al. (1999, 2002). This study is closely linked to the modelling of a 1000-km mantle transect across the Karelian and Kuola-Kuloi cratons based on kimberlites in Finland and in NW Russia (Peltonen et al., 2008).

2. Samples

Xenocryst samples from Arkhangelskaya were hand picked from a heavy mineral concentrate made available for this study by ALROSA Co Ltd. Hundreds of garnet, clinopyroxene (Cpx), chromite and ilmenite xenocrysts were recovered and mounted for electron microprobe and LA-ICP-MS analyses. Garnet, chromite and ilmenite grains in this study are unsorted and represent the entire range of populations within the kimberlites. For the Cpx grains, an attempt was made to choose only peridotitic varieties, based on their bright green colour.

3. Analytical techniques

Pyrope garnet, chromite and ilmenite major element compositions were determined by a Cameca Camebax SX100 ele ctron microprobe at GTK E-beam Laboratory, applying an acceleration voltage of 25 kV, probe current of 48 nA and beam diameter of 1 µm. The parameters for eclogitic garnet were 15 kV, 20 nA and 15 µm, and for clinopyroxene 15 kV, 30 nA and 5 µm, respectively. Selected garnet xenocrysts were analyzed for trace elements by laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) at the University of Frankfurt, using a New Wave Research LUV213™ petrographic ultraviolet Nd-YAG laser coupled with a Finnigan MAT ELEMENT2™. The laser was run at a pulse frequency of 10 Hz and a pulse energy of 0.5 mJ with a 150 µm spot size for the measurement of garnet. Two in-house natural garnet samples (PN1 and PN2) were used as external standards, and calcium was used as an internal standard for the measurements.

The garnets were classified based on their major element compositions into harzburgitic (G10), lherzolitic (G9), wehrlitic (G12), high-Ti peridotitic (G11), low-Cr megacrystal (G1), eclogitic (G3, G4) and pyroxenitic (G4, G5) varieties according to the Grütter et al. (2004) revision of the scheme by Dawson and Stephens (1975). For equilibration temperatures of garnets the Ni thermometer (Griffin et al., 1989) was applied using the calibration of Ryan et al. (1996). The chromites were classified into Di (inclusion in diamond-type) and Cr–Ti (phenocrystic) varieties according to Fipke et al. (1995) and Sobolev (1977) and their equilibration temperatures were calculated using the Zn thermometer described also in Ryan et al. (1996). Equilibration pressures and temperatures of the peridotitic clinopyroxene were calculated using the single-grain Cpx thermobarometer of Nimis and Taylor (2000).

4. Results

The analytical database containing all major and trace element analyses obtained for this study is available as an Open file report at the Geological Survey of Finland website (Lehtonen et al., 2008a,b).

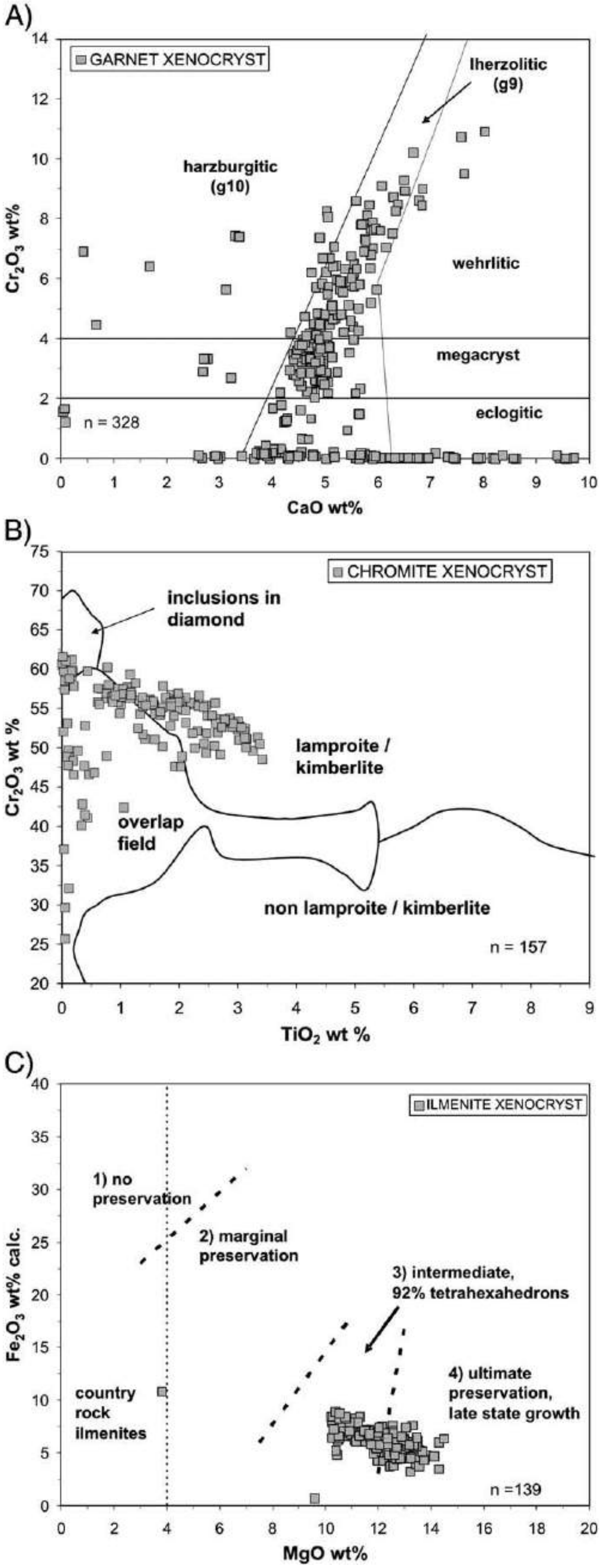

Fig. 2. Compositional classification diagrams for Arkhangelskaya xenocrysts. A) Cr_2O_3 vs. CaO for garnet. Crustal garnets are excluded from this diagram using the classification by Schulze (2003). The harzburgite, lherzolite and non-peridotite fields are redrawn after Gurney (1984) and the wehrlite field is separated according to Sobolev et al. (1973). B) Cr_2O_3 vs. TiO_2 diagram for chromites redrawn after Fipke et al. (1995). C) Fe_2O_3 vs. MgO diagram for ilmenites redrawn after Gurney and Zweistra (1995).

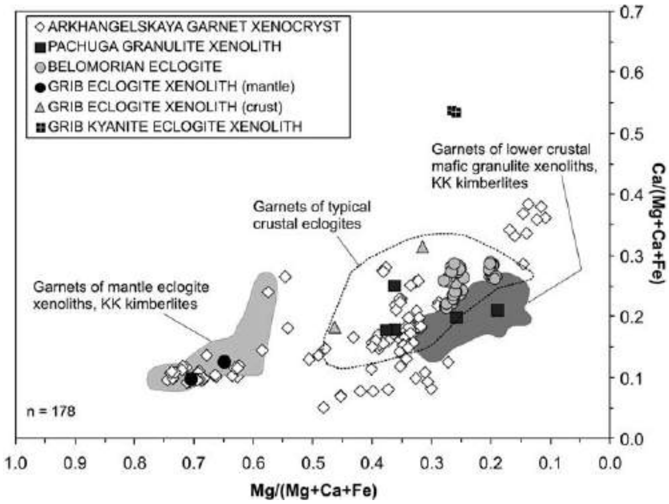

Fig. 3. Eclogitic garnet xenocrysts from Arkhangelskaya based on compositional classifications by Grütter et al. (2004) and Schulze (2003) plotted on a Mg/(Mg + Ca + Fe) vs. Ca/(Mg + Ca + Fe) variation diagram. For comparison, analytical data of garnets has been collected from the literature: lower crustal granulites from the Pachuga field (Arkhangelsk) by Markwick and Downes (2000), crustal and mantle eclogites from Grib by Malkovets et al. (2003b), kyanite eclogite from Grib by Bobrov et al. (2005), and eclogite xenoliths from the Belomorian mobile belt by Peltonen (unpublished data) and Volodichev et al. (2004). Compositional fields of garnets of mantle eclogites (Peltonen et al., 2002) and of mafic lower crustal granulites (Peltonen, unpubl. data) from Kaavi-Kuopio kimberlite field (KK) on the SW edge of the Karelian craton, are also shown for comparison. The field of typical crustal eclogites is based on crustal eclogite suites from Kaapvaal (Pearson et al., 1995), NE Greenland (Brueckner et al., 1998), Rongcheng eclogites, China (Nakamura and Banno, 1997), eclogitic gabbros from Zambia (John and Schenk, 2003) and Dora Maira Massif, Western Alps (Nakamura and Banno, 1997).

4.1. Major element geochemistry of mantle xenocrysts

4.1.1. Garnet

CaO and Cr_2O_3 contents of pyrope garnets from Arkhangelskaya kimberlite are shown in Fig. 2A (Sobolev et al., 1973; Gurney, 1984). It is evident that lherzolitic garnet (G9, G11) predominates over other garnet varieties. Wehrlitic garnets (G12) are literally absent, as could be expected based on the observation by Sobolev et al. (1992). Only a small percentage of the garnet xenocrysts are classified as subcalcic harzburgitic grains (G10). Cr-poor but Ti-rich pyropes of megacryst composition (G1) are common. The abundant grains at low levels of Cr fall into the "eclogitic" categories (G3, G4).

A closer look at the "eclogitic" xenocrysts reveals that many of them are probably of crustal origin (Fig. 3), similar in composition to garnets from crustal eclogites from Grib (Malkovets et al., 2003b) and lower crustal granulites from the Kepino-Pachuga field (Markwick and Downes, 2000; Fig. 1). The separate low-Mg, high-Ca xenocryst population is most likely also of crustal origin, apparently from even less mafic protoliths. None of the xenocrysts match with garnets from a kyanite-eclogite described from Grib by Bobrov et al. (2005). A subset of the Arkhangelskaya xenocrysts has similar compositions to garnets from Archean eclogites within the Belomorian mobile belt at the NE margin of the Karelian Craton (Peltonen, unpublished data; Volodichev et al., 2004). The remaining garnet xenocrysts with intermediate Mg/(Mg + Ca + Fe) could represent another type of lower crustal eclogite based on data collected from the literature

(Pearson et al., 1995; Nakamura and Banno, 1997; Brueckner et al., 1998; John and Schenk, 2003).

A mantle eclogite population with higher Mg/(Mg + Ca + Fe) can also be recognized and compares well with garnets from mantle eclogites from Grib (Malkovets et al., 2003b) and from Kaavi-Kuopio, located at the SW edge of the Karelian Craton (Peltonen et al., 2002). A diamond indicative Group I population with $Na_2O > 0.07$ wt.% (Sobolev and Lavrentìev, 1971; McCandless and Gurney, 1989) exists within the mantle eclogite garnets (not shown), as expected, given the presence of eclogitic paragenesis diamonds (ca. 10% of the diamond population) (Sobolev et al., 1973; Galimov et al., 1994) and the statement by Garanin et al. (1998b) of an abundance of Group I eclogitic garnets (ca. 9% of the garnet population) in the Lomonosov deposit.

4.1.2. Chromite

Chromite is an abundant indicator mineral in the Arkhangelskaya pipe. In Fig. 2B chromite analyses are plotted in a Cr_2O_3–TiO_2 diagram with discriminant fields after Fipke et al. (1995). A majority of the chromites plot within the kimberlite phenocryst field, which is typical for chromites derived from the Lomonosov deposit kimberlites (Sobolev et al., 1992). A subset of chromites also have diamond indicative compositions with high contents of Cr ($Cr_2O_3 > 58$ wt.%) and Mg (MgO >9 wt.%) and low contents of Ti ($TiO_2 < 0.6$ wt.%) and $Fe^{3+}/(Cr + Al + Fe^{3+})$ (<0.07), underlining the significance of chromite–harzburgite as a potential diamond source rock in the mantle.

4.1.3. Ilmenite

Based on previous work (e.g. Sobolev et al., 1992; Sablukov et al., 1995) ilmenite is not a common mineral in the Lomonosov deposit kimberlites. However, in the Arkhangelskaya sample it was surprisingly abundant. In Fig. 2C microprobe analyses of ilmenite xenocrysts are plotted in a Fe_2O_3–MgO-diagram (Gurney and Zweistra, 1995) that is widely used for describing the redox conditions in the mantle at or close to the time of emplacement of the host kimberlitic magma. The results indicate very low oxygen fugacity, implying almost optimal conditions for diamond preservation.

4.1.4. Clinopyroxene

Clinopyroxene xenocrysts are Cr-diopsides that contain 0.57–3.84 wt.% Cr_2O_3 and 0.56–3.53 wt.% Na_2O. Their Mg/(Mg + Fe) ratios range from 0.91 to 0.96. According to the Al_2O_3–MgO-screen by Nimis and Taylor (2000) (not shown) over 80% of the Cpx-grains analyzed for this study (ca. 200 in total) are classified as a low-Al garnet-free peridotite type. These grains were excluded from the P–T calculations described in the following section.

4.2. Cpx thermobarometry and thermal state of the lithosphere

Single-Cpx thermobarometry on clinopyroxene xenocrysts originating from garnet–peridotites (Nimis and Taylor, 2000) plot along a cool cratonic geotherm (Fig. 4). It lies between the 36 mW/m^2 geotherms by Pollack and Chapman (1977) and Kukkonen and Peltonen (1999), the latter defined for the 600 Ma Karelian Craton using heat flow constraints and xenolith modes and petrophysical properties. The Cpx geotherm is essentially indistinguishable from the 37 mW/m^2 garnet geotherm proposed by Sablukov et al. (1995).

In this paper, the calculated xenocryst pressures are converted into depth values by assuming that the pressures are representative of lithostatic pressures and by using the following relationship:

$$Z = \frac{P}{9.81\rho_M} - \frac{\rho_C}{\rho_M}H + H$$

where Z is depth (m), H is thickness of crust (m), P is the pressure (Pa) and ρM and ρC are the mantle and crust densities, respectively.

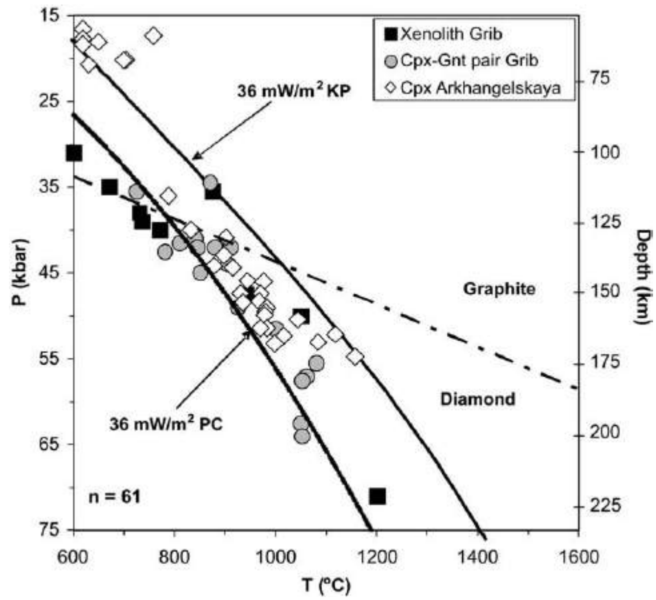

Fig. 4. P–T data from Arkhangelskaya clinopyroxene xenocrysts. Mantle xenoliths from Malkovets et al. (2003a,b) and Sablukova et al. (2003), garnet-cpx aggregates from Kostrovitsky et al. (2004). Reference geotherms from Pollack and Chapman (1977; PC) and Kukkonen and Peltonen (1999; KP). Diamond-graphite transition after Kennedy and Kennedy (1976).

For instance, applying 2800 kg m^{-3} for the crust and 3360 kg m^{-3} for the upper mantle density, a crustal thickness of 36 km (representative for the Arkhangelsk area), and a calculated Cpx pressure value of 4 GPa, the corresponding depth value is 127.4 km, which marks the depth where the Cpx-geotherm (Fig. 4) enters the diamond stability field (Kennedy and Kennedy, 1976).

4.3. Garnet and chromite thermometry

In Figs. 5 and 6 our Arkhangelskaya data are presented along with analyses made available for this study by Prof. W.L. Griffin. His database contains additional analyses from Arkhangelskaya as well as from other pipes in the region (Fig. 1). The database has been discussed in Sablukov et al. (1995) and Malkovets et al. (2007). Data from all of the pipes presented here are compared in detail in the Discussion (Section 5).

The T_{Ni} histogram for Arkhangelskaya garnet xenocrysts (Fig. 5) shows a bimodal distribution including a lower temperature peak at 650 to 1000 °C consisting mainly of lherzolitic (G9) and harzburgitic (G10) garnets, and a higher temperature peak at 1050 to 1450 °C containing low-Cr megacryst (G1) and high-Ti peridotitic (G11) varieties in addition to the previous two. With this number of analyses, a complete gap exists between the two garnet sampling peaks, and even if more data partially filled the gap, at the very least a garnet sampling minimum from 1000 °C to 1100 °C appears to be a real feature of this kimberlite. The origin of the megacrystal grains is still under debate (e.g. Hops et al., 1992) but the most Mg-rich varieties have been included on the diagram as they reach the same Mg (MgO>18 wt.%) and Cr contents ($Cr_2O_3 > 1.5$ wt.%) as G9 garnets, suggesting equilibration with similar composition olivine as the lherzolites.

Results of Zn-thermometry (Ryan et al., 1996) of Arkhangelskaya chromites are presented in Fig. 6. The data reveal more or less continuous sampling of chromite from 600 °C to 1100 °C, with an obvious concentration at 850–1000 °C, which is due to the phenocrystic Cr–Ti chromites shown only for illustrative purposes. Even if they are excluded, the remaining chromite data fill the garnet data gap, suggesting that the chromites and garnets are derived from

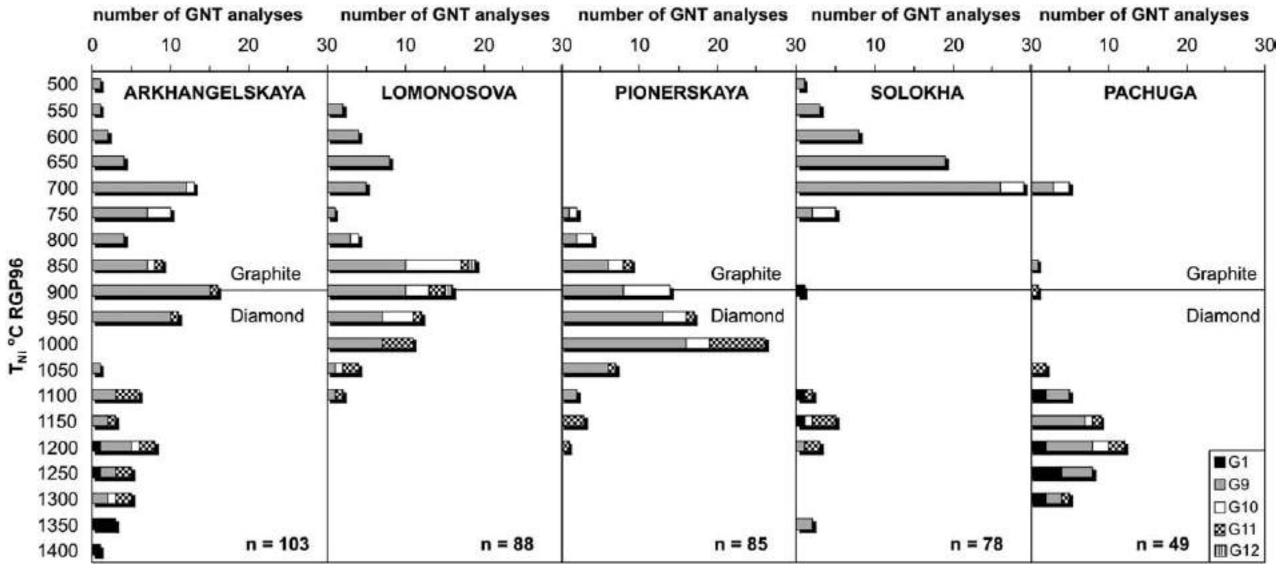

Fig. 5. Distribution of T_{Ni} for the Arkhangelskaya pyrope garnet xenocrysts. Lomonosova, Pionerskaya, Solokha and Pachuga data from W.L. Griffin database, discussed in Sablukov et al. (1995) and Malkovets et al. (2007). 48 of the total 103 analyses of Arkhangelskaya are also from this database. Diamond-graphite transition drawn based on the Cpx-geotherm in Fig. 4. Garnet classification according to Grütter et al. (2004). The absence of G1 garnet in other pipes than Arkhangelskaya is probably due to selecting grains based on color.

slightly different horizons within the lithospheric mantle. The range of Zn-temperatures of chromites fits very well with the lower temperature population of garnets but the most evident difference is the absence of chromites at deeper levels that would match with the higher temperature garnet population.

4.4. Trace and rare-earth element content of garnets

In Fig. 7 the TiO_2 contents and Mg# (Mg/(Mg + Fe)) of Arkhangelskaya garnets are plotted against depth by extrapolating the T_{Ni} temperatures to the Cpx-derived geotherm. The results are again presented along with the data from other kimberlite pipes in the Arkhangelsk region, see Discussion (Section 5).

The variations in the garnet TiO_2 contents reflect the degree of depletion or enrichment within the mantle (e.g. Griffin and Ryan, 1995), whereas Mg# of garnet reflects bulk composition but is also strongly controlled by the temperature dependent Mg–Fe exchange with olivine, orthopyroxene, clinopyroxene and Mg-chromite (Stachel et al., 2003). In addition, Ca has a significant crystal chemical effect on the Mg–Fe partitioning between garnet and olivine (O'Neill and Wood, 1979; O'Neill, 1980; Stachel et al., 2003). The temperature boundary at c. 800 °C in the Arkhangelskaya garnet data corresponds to a break in TiO_2 contents at approximately 110 km in depth, which separates low Ti content (TiO_2 < 0.2 wt.%) garnets with Mg# between 0.80 and 0.84 from a deeper horizon exhibiting a wider range in both parameters. In the temperature horizon 800 °C–1000 °C, corresponding to 110–140 km, TiO_2 contents of garnets vary between 0.05 to 0.5 wt.% and Mg# between 0.76–0.84. The garnet sampling gap between 1000 °C and 1100 °C corresponds to a depth of 140–160 km below which the lithospheric mantle continues to at least 230 km. The G1 and G11 grains in this lower layer exhibit very high contents of TiO_2 (up to 1.25 wt.%), interpreted to be a melt-metasomatic signature. The lower mantle layer

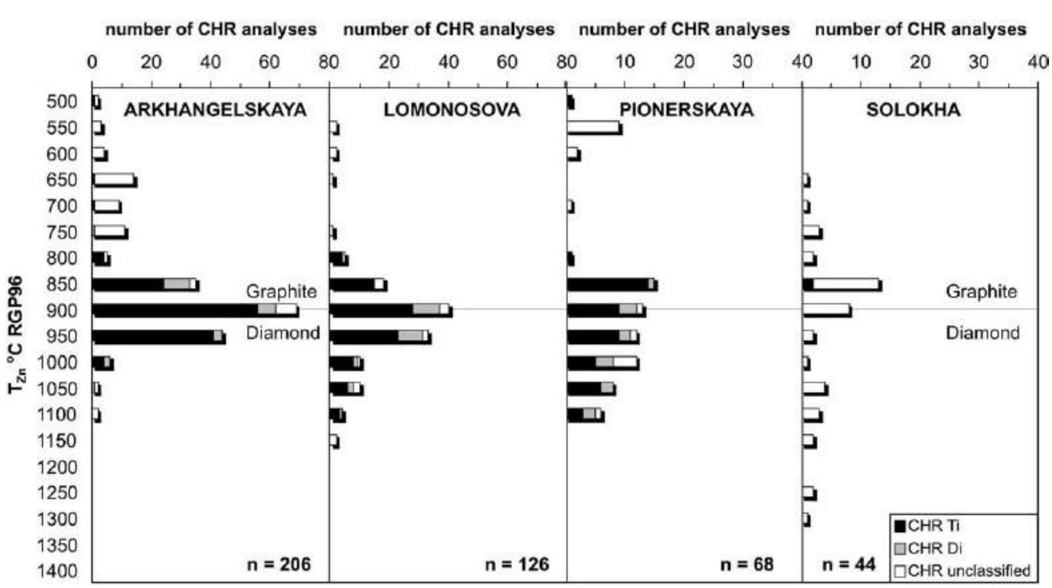

Fig. 6. Distribution of T_{Zn} for the Arkhangelskaya chromite xenocrysts. Lomonosova, Pionerskaya, Solokha and Pachuga data from W.L. Griffin database (Sablukov et al., 1995; Malkovets et al., 2007). 49 of the total 206 analyses of Arkhangelskaya are also from this database. Chromite classification after Fipke et al. (1995).

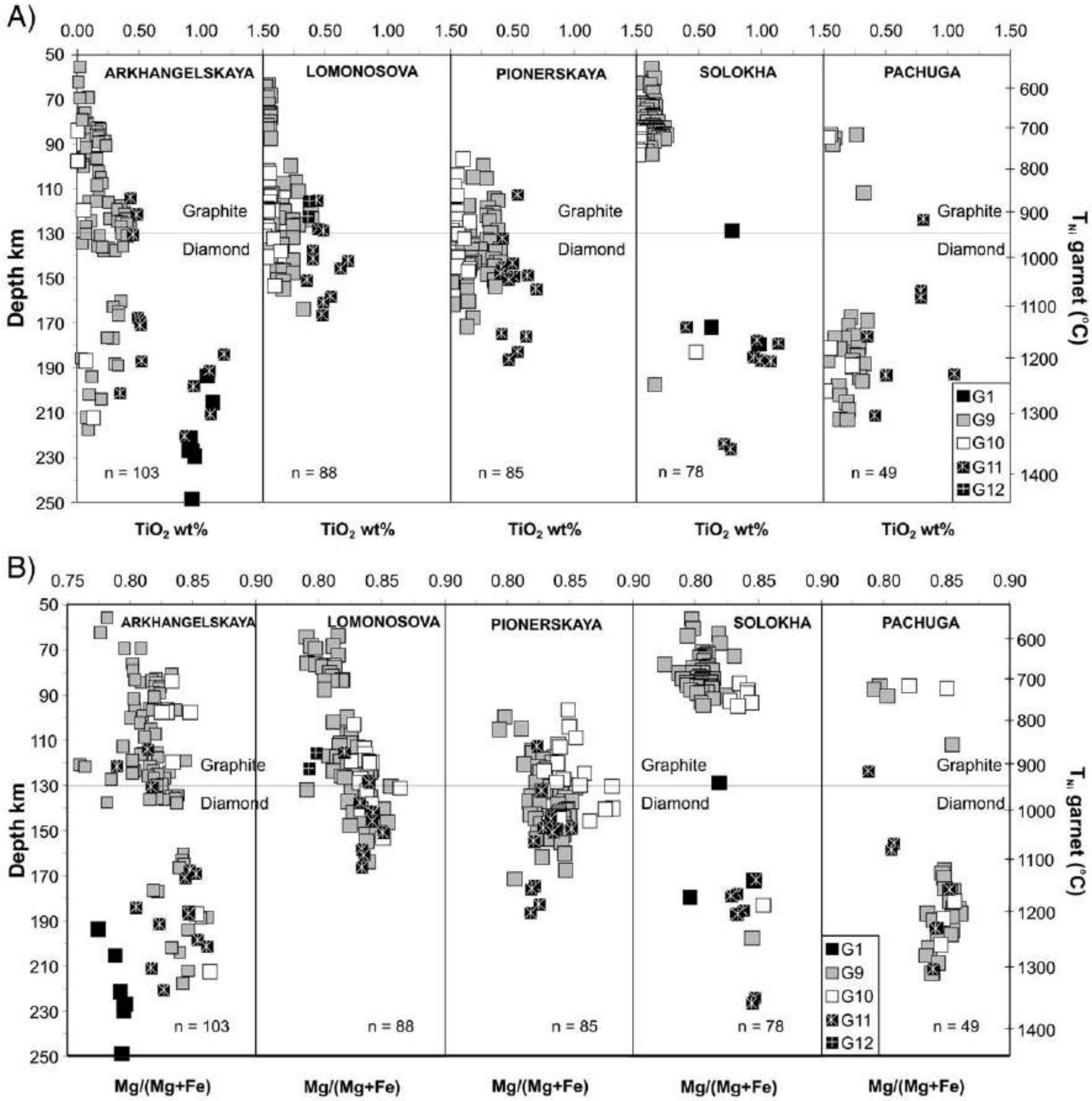

Fig. 7. Distribution of TiO_2 (A) and Mg# (B) vs. depth for the Arkhangelskaya xenocryst pyropes. Lomonosova, Pionerskaya, Solokha and Pachuga data from W.L. Griffin database (Sablukov et al., 1995; Malkovets et al., 2007). 48 analyses of Arkhangelskaya are also from this database.

appears to have slightly higher overall garnet Mg# and exhibits the widest compositional range when measured by garnet Mg#, ranging from 0.77 of the megacrysts up to 0.86 of some peridotitic varieties. The peridotitic diamond window stretches from the top of the diamond stability field at 130 km based on the Cpx geotherm (Fig. 4) to the base of G10 garnet bearing mantle at about 210 km.

The C1-chondrite (McDonough and Sun, 1995) normalized REE_N profiles (Fig. 8) of G9 and G10 garnet xenocrysts originating from the shallowest mantle layer, <110 km (<800 °C) are mostly so called S-type (sinusoidal), diagnostic of P-type garnet inclusions in diamond (e.g. Stachel et al., 2004). The S-type profile is characterized by enrichment in Sm_N over Dy_N (and Ho_N and Er_N) which mark a minimum in the profile and then a positive slope to Yb and Lu, giving rise to an S-shaped pattern. The positive slope from La_N to Nd_N is due to a strong increase in compatibility in the garnet structure with decreasing ionic radius of the LREE from La to Nd (e.g. Stachel et al., 1998). A subset of the G9 garnets are classified as N-type, showing stronger LREE depletion than the S-type profiles relative to MREE and HREE with steady enrichment from Sm_N to Yb_N, typical of Ca-saturated mantle garnets (e.g. Shimizu, 1975). In the middle mantle layer, 110–140 km (800–1000 °C), the S-type REE_N profile is also clearly dominant among G10 and G9 garnets with a few G9 and G11 garnets originating from this horizon belonging to the N-type. The deepest mantle horizon, >160 km (>1100 °C), is characterized by N-type G9, G11 and G1 garnets but still contains significant S-type G9 and G10 garnets.

The sinuosity and the degree of LREE depletion of a garnet REE_N profile can be measured using the ratios $(Nd/Ce)_N$ and $(Nd/Dy)_N$, as shown in Fig. 9. $(Nd/Ce)_N >>1$ suggests strong LREE depletion compared to MREE, whereas $(Nd/Dy)_N >>1$ refers to a high degree of sinuosity. Using these criteria, it can be observed, especially in the shallowest and middle mantle layers, that as the sinuosity of the REE_N patterns increase, the relative depletion in LREEs decrease.

5. Discussion

The mantle sampling pattern of Arkhangelskaya (Figs. 5–7) evidenced by single-grain thermobarometry differs clearly (especially

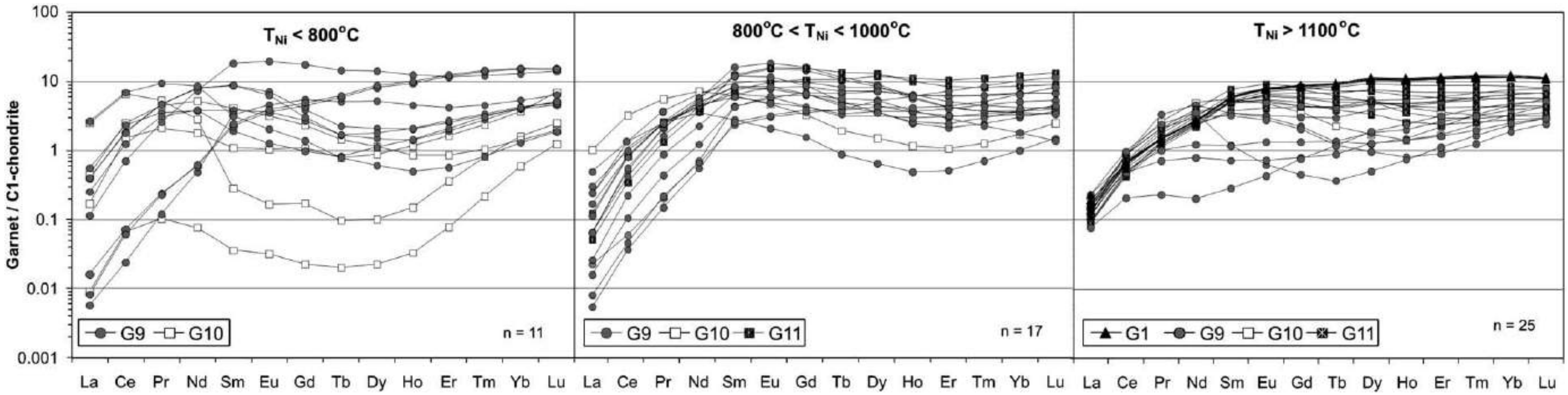

Fig. 8. C1-Chondrite normalized (McDonough and Sun, 1995) REE profiles of mantle-derived garnet xenocrysts from Arkhangelskaya. The samples are subdivided according to rock type (Grütter et al., 2004) and temperature horizon (Figs. 5 and 7).

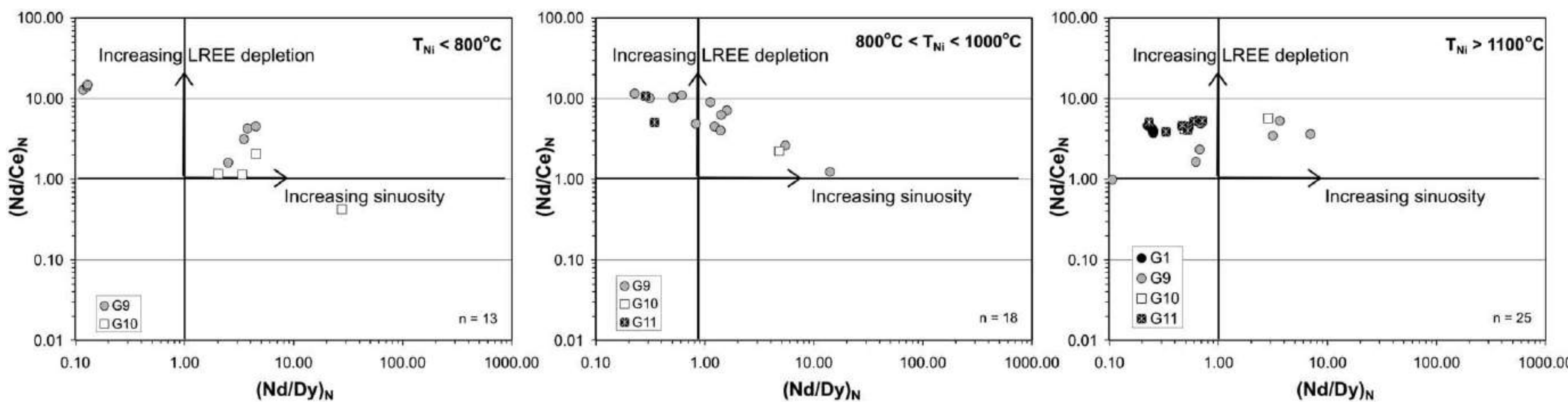

Fig. 9. $(Nd/Ce)_N$ vs. $(Nd/Dy)_N$ for garnet xenocrysts from Arkhangelskaya. (N = normalized to C1-chondrite composition after McDonough and Sun, 1995).

in regards to garnet) from that of Lomonosova and Pionerskaya which are other high-grade pipes of the Lomonosov deposit (Fig. 1). The difference is even more evident (both garnet and chromite) when compared to the low-grade kimberlites of Solokha and Pachuga from the Kepino-Pachuga field (Fig. 1). Lomonosova and Pionerskaya show continuous mantle sampling down to ca. 180–190 km with most of the grains originating from 110–160 km, which is also the G10 garnet and Di-type chromite bearing horizon. In comparison, Solokha has mostly sampled its garnets from a low-temperature (550–800 °C) graphite-bearing part of the mantle at 60–100 km, with another less prominent sampling peak at 160–210 km. The latter peak coincides with the main sampling horizon of Pachuga. Solokha chromites display a wide range of temperatures, 700–1350 °C (80–225 km), filling the discontinuity seen in the garnet thermometry. None of the Solokha chromites classifies as a diamond indicative Di-chromite, suggesting that even though the correct P-T horizon for diamond was sampled, it probably did not contain diamondiferous material. The sampling pattern of Arkhangelskaya is intermediate to the kimberlites above, with surprisingly few G10 garnet/Di-type chromite-bearing samples originating from the diamond stability field (130–210 km) considering the high-grade of the pipe. The chromite sampling pattern resembles closely that of Lomonosova and Pionerskaya but the bimodal distribution of garnet is more similar to Solokha, albeit less skewed towards low temperatures. The high temperature (1100–1350 °C) garnet population of Arkhangelskaya consisting of G9, G10 and G11 garnet varieties resembles that of Pachuga (Figs. 5 and 7).

Arkhangelskaya Cpx sampling (Fig. 4) runs uninterrupted from 100 km to 175 km and across the discontinuity at 140–160 km seen in the garnet data (Figs. 5 and 7). This is problematical, in that all of the Cpx data plotted in Fig. 4 should, based on the compositional criteria of Nimis and Taylor (2000), be derived from garnet-peridotites. Unfortunately, we do not have trace element data for these Cpx grains to determine whether they have depleted HREE contents, a robust test for coexistence with garnet. However, we have no reason to doubt the reliability of the Nimis and Taylor (2000) method in this case given the reasonable shape and continuous nature of the geotherm derived, and the fact that it coincides very well with previous P–T data from xenoliths. In other words there should be garnets from the entire Cpx temperature–pressure range, but that is not what we see. One possible way that this inconsistency might develop is if the estimates of depth for the high temperature (>1100 °C) garnets are not accurate in which case that they could be instead derived from shallower layers of the lithosphere. Such might be the case for recent heating of the lowermost lithosphere that would consequently no longer lie on the conductive geotherm. However, two important observations argue against this scenario: 1. The Cpx geotherm shows no sign of such a heating event because it has a realistic shape to high temperatures, especially when the Grib data are included. It would then be necessary to call upon some lag between a “speedy” garnet thermometer and a slower Cpx thermometer. 2. Even though the high temperature group of garnets includes many high-Ti G1s and G11s which are markers of melt metasomatism, there occurs within the same high-T group, a significant population of G9 and G10 garnets with depleted contents of Ti and sinusoidal REE_N profiles. These grains at least must represent pristine Archean mantle unmodified by more recent melt metasomatic events, and consequently should follow the conductive geotherm. The absence of garnet xenocrysts between 140 and 160 km remains, therefore, somewhat enigmatic. The most likely reason for the gap is that this horizon is relatively garnet-poor (but not garnet-free) and that the number of samples was simply not adequate to identify it.

One similarity between all three kimberlites is the correlation of garnet Mg# and depth (Fig. 7B), which can be due to either temperature effects on gt/ol Fe/Mg Kd (O'Neill and Wood, 1979; O'Neill, 1980) or an increase in the degree of partial melting with increasing depth. Since no correlation between garnet Cr# (Cr/(Cr+Al), not shown) and depth can be detected in any of the localities, the first alternative seems to be more realistic. In fact, the correlation of Mg# with increasing temperatures calculated from Ni concentration data lends confidence to the Ni thermometry methodology employed in this study. This is important when considering, e.g., the speculation about the depth of derivation of the high-T garnet population. The correlation also implies a stable and relatively closed system without strong disturbance due to influxes of additional metasomatic components.

The evidence from the five individual kimberlites demonstrates that there is considerable compositional variability in the mantle underlying Zimnyi Bereg area. Chromite thermometry from Lomonosova, Pionerskaya and Arkhangelskaya pipes bears evidence of a relatively uniform chromite-bearing layer underlying the Zolotitsa field, major differences being reflected in the presence/absence and the composition of garnet. Malkovets et al. (2007) suggested that garnet and diamond are secondary phases, created by the same metasomatic event, which included invasion of the highly refractory lithosphere by Si-bearing methane-rich fluids from asthenosphere-derived melts producing subcalcic garnet and diamond. As the melt invaded further into the lithosphere, it introduced Ca, Al and Fe into the system, refertilizing harburgites to lherzolites, and changing the garnet REE_N patterns from S- to N-type, which has been observed, for instance, in xenoliths from Lesotho (Simon et al., 2003). The distribution of garnet (and diamond) in the mantle would then be controlled by the presence of ancient structural conduits and, thus, strong lateral mineralogical variation on small scales, such as between the Zolotitsa and Kepino-Pachuga fields, would be expected. The REE_N distribution of Arkhangelskaya garnet xenocrysts indicates that the transformation from harzburgite to lherzolite has been most efficient at the deepest level of the lithospheric mantle, whereas in the shallowest and the middle lithospheric mantle layers, S-type REE_N patterns have remained better preserved. Interestingly the sinusoidal REE_N patterns seen in the lherzolitic garnets reported here are very similar to those in lherzolitic inclusions in diamonds reported by Stachel et al. (2004). In a number of respects, including an abundance of S-type G9 and G10 garnets similar in composition to inclusions in diamond and the existence of an “undisturbed” garnet Mg# with depth correlation, the Arkhangelskaya mantle section is quite similar to that at the Kuhmo-Lentiira area on the Central Karelian Craton (Lehtonen and O'Brien, in press).

6. Conclusions

Mantle xenocrysts in the Arkhangelskaya pipe demonstrate that the lithosphere underlying the Kola-Kuloi Craton is comparable in depth to the adjacent Karelian Craton (Peltonen et al., 1999; Lehtonen et al., 2004; Peltonen et al., 2008) and it has been affected by metasomatic events, especially near the asthenosphere–lithoshpere boundary. The distribution of G10 garnets in Arkhangelskaya shows that the peridotitic diamond window extends down to 210 km under Zolotitsa, even though it seems that the pipe itself has sampled its diamonds dominantly from between 190 and 210 km – whereas the neighbouring Lomonosova and Pionerskaya pipes have sampled their diamonds from 130–160 km. The comparable grade of these three pipes suggests that diamondiferous material is generously distributed within the mantle section. A subset of the Arkhangelskaya, Lomonosova and Pionerskaya chromites originating from 130–180 km have diamond-indicative compositions, underlining the importance of chromite bearing harzburgite as a diamond source rock in the Zolotitsa mantle. The sinusoidal REE_N profiles of G9 garnets in Arkhangelskaya could be related to diamondiferous lherzolites. Moreover, the major element chemistry of Mg-ilmenite xenocrysts in the pipe demonstrates that almost optimal redox conditions for diamond preservation prevailed in the mantle during the time of emplacement of the host kimberlite magmas.

The differences between the sampling patterns of Zolotitsa (Arkhangelskaya, Lomonosova, Pionerskaya) and Kepino-Pachuga (Solokha, Pachuga) kimberlites can probably best be explained by the model suggested by Malkovets et al. (2007). However, the remarkable difference evidenced by garnet composition and thermometry between the Arkhangelskaya and the other two Zolotitsa pipes, located just within a few kilometres distance from each other (Fig. 1), most likely derives from some other factors, such as differences in rheology and eruption rates of the rising kimberlite magmas – or even from inhomogeneities in the kimberlites themselves.

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