Lithos xxx (2010) xxx–xxx

Contents lists available at ScienceDirect

Lithos

journal homepage: www.elsevier.com/locate/lithos

Subducted oceanic crust as diamond hosts revealed by garnets of mantle xenoliths from Nyurbinskaya, Siberia

Amy J.V. Riches [a,*], Yang Liu [a], James M.D. Day [b], Zdislav V. Spetsius [c], Lawrence A. Taylor [a]

[a] Planetary Geosciences Institute, Department of Earth & Planetary Sciences, University of Tennessee, Knoxville, TN 37996, USA
[b] Department of Geology, University of Maryland, College Park, MD 20742, USA
[c] Institute of Diamond Industry, ALROSA Co. Ltd., Mirny, Yakutia, Russia

ARTICLE INFO

Article history:
Received 10 May 2010
Accepted 9 September 2010
Available online xxxx

Keywords:
Yakutia
Eclogite
Craton
Diamond
Lithosphere formation

ABSTRACT

The ~380 Ma Nyurbinskaya kimberlite pipe, Yakutia, Siberia, sampled a highly-diamondiferous and unusual mantle xenolith population dominated by eclogites. New in-situ major- and trace-element data for garnets previously analyzed for oxygen isotope compositions show that Group A eclogitic garnets have Mg# >68 and are LREE-depleted. Group B and Group C eclogitic garnets cover a range of Mg#, and are each divided into two types based on their trace-element characteristics. Type B1 and C1 eclogitic garnets are dominant and are LREE-depleted. Less common Type B2 and C2 garnets generally have Mg# >60 and convex-upward REE-profiles. Harzburgitic garnets are a minor component of the Nyurbinskaya xenolith suite, and have high Mg# (~84), high Cr contents (~11 wt.% Cr_2O_3), and sinusoidal REE-patterns. Group A, Type B1, and C1 eclogitic garnets define a broad negative correlation between Mg# and Yb abundances consistent with a shallow origin as basaltic and gabbroic portions of oceanic crust. Harzburgitic, Type B2, and C2 eclogitic garnets have trace-element characteristics indicative of interaction with a C–O–H–N–S-rich fluid in lithospheric environments. These results provide clear evidence for the presence of subducted crustal materials in the Siberian mantle lithosphere and support models of craton formation by subduction zone stacking.

1. Introduction

Despite the relative rarity of eclogitic and pyroxenitic mantle xenoliths sampled by kimberlites, they offer key constraints on the formation of thickened Archean cratonic lithosphere, as well as the extent of modification resulting from interaction with melt or fluids in mantle environments and during kimberlite transport (e.g., Pearson et al., 2003; Carlson et al., 2005). Studies of eclogites and pyroxenites are not only important for refining models of global crust–mantle evolution, but also provide constraints on the origin of highly-diamondiferous mantle as eclogites and pyroxenites represent significant diamond reservoirs in some portions of Archean lithosphere (e.g., Orapa, Magondi Mobile Belt, Shee and Gurney, 1979; Roberts Victor, Kaapvaal Craton, MacGregor and Manton, 1986; Carswell et al., 1981; Lac de Gras, Slave Craton, Aulbach et al., 2007; Nyurbinskaya, Siberian Craton, Spetsius et al., 2008). Two general scenarios have been proposed to explain the compositional diversity of ancient cratonic mantle, namely; 1) subduction zone stacking (Helmstaedt and Gurney, 1995; Pearson and Witting, 2008), and 2) melt intrusion and ponding at the base of the lithospheric mantle (e.g., Griffin et al., 1999a; Griffin and O'Reilly, 2007). Several early studies suggested that eclogite and pyroxenite xenoliths may represent the products of high-pressure crystal fractionation (e.g., O'Hara, 1969; MacGregor and Carter, 1970; Garlick et al., 1971; O'Hara et al., 1975). However, mineralogical and reconstructed bulk-rock characteristics (major-, minor-, and trace-element systematics, and isotopic compositions) of many eclogites may not be accounted for by high-pressure crystal fractionation (e.g., Shervais et al., 1988; Taylor and Neal, 1989; Jacob, 2004; Schmickler et al., 2004). In this context, eclogites and pyroxenites may ultimately be derived from crustal precursors (e.g., Taylor and Neal, 1989; Pearson et al., 1993; Spetsius, 2004).

Compared to other Yakutian kimberlites (e.g., Udachnaya kimberlite, Daldyn kimberlite field, Pokhilenko et al., 1991; Boyd et al., 1997), the ~380 Ma Nyurbinskaya kimberlite (Fig. 1) matrix has lower LREE, Nb, Ta, U and Th contents (Agashev et al., 2001, 2004; Lapin et al., 2007). Further, the $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ compositions of the Nyurbinskaya kimberlite at the time of eruption are transitional between Group I and Group II kimberlites of the African cratons (e.g., Agashev et al., 2001; Kornilova et al., 2001). This distinctive kimberlite sampled an unusual and highly-diamondiferous mantle attested to by a xenolith population that is dominated by eclogites with only minor amounts of peridotite. Eclogite-rich xenolith populations have been identified in only one other Yakutian kimberlite (Zagadochnaya, 30 km Southeast of Udachnaya), but Zagadochnaya xenoliths are

* Corresponding author. Tel.: +1 865 974 6024; fax: +1 865 974 2368.
E-mail address: ariches@utk.edu (A.J.V. Riches).

0024-4937/$ – see front matter © 2010 Elsevier B.V. All rights reserved.
doi:10.1016/j.lithos.2010.09.006

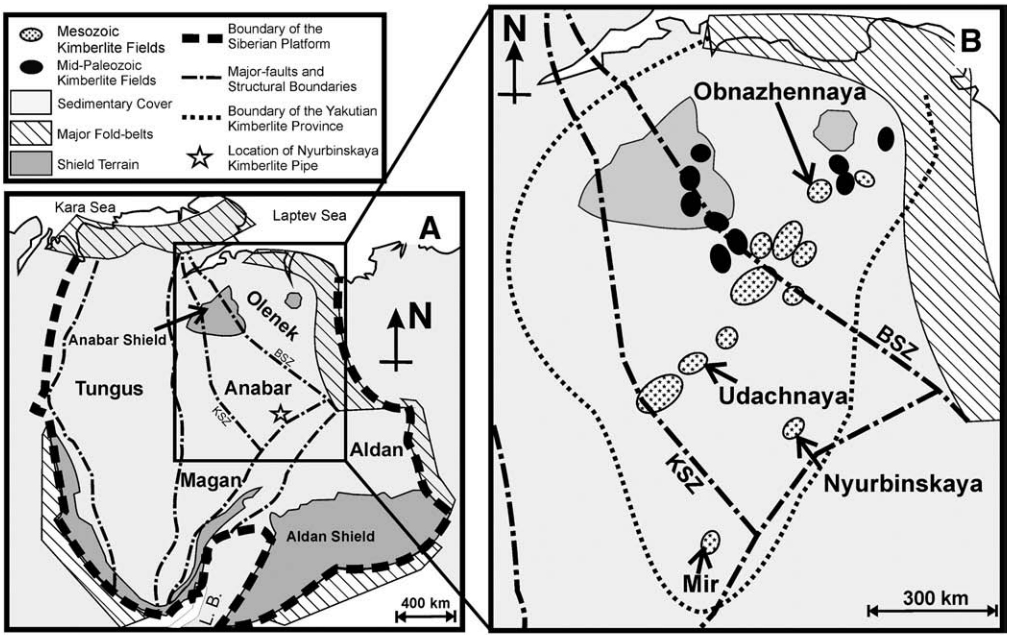

Fig. 1. Schematic diagram of prominent geological structures and crustal provinces of the Siberian Platform (A), and (B) location of Yakutian kimberlite fields with the site of Obazhennaya, Udachnaya, Mir and Nyurbinskaya kimberlite pipes (after Dawson, 1980 and Rosen et al., 1994). KSZ = Kotuykan Shear Zone. BSZ = Billyakh Shear Zone. L.B. = Lake Baikal.

devoid of diamonds, and grospydites (grossular ± clinopyroxene ± kyanite), for which it is the type-locality, are abundant (Bobrievich et al., 1960; Sobolev et al., 1968; Nimis et al., 2009). Importantly, the Nyurbinskaya xenolith suite has an unusual distribution of oxygen isotope compositions extending to the highest known $\delta^{18}O$ value (+9.65‰; Spetsius et al., 2008) observed in diamondiferous xenoliths worldwide. In contrast to xenolith suites from other localities across the Siberian craton, and even across the globe, garnets of Nyurbinskaya mantle xenoliths have no $\delta^{18}O$ values below the mantle range (5.3 ± 0.6‰, $2\sigma_{stdev}$, where garnet–zircon fractionation is <0.1‰; Valley et al., 1998 and Valley, 2003) and define two populations, the largest of which is centered at 6.6‰ (Fig. 2).

It is widely accepted that O-isotopes fractionate little (≤1‰) during high-temperature melting, metasomatism, and metamorphism in the upper mantle (Eiler, 2001). Larger fractionations of oxygen isotopes require processes involving fluids at or near Earth's surface (Clayton et al., 1975), which are considered analogous to low- and high-temperature alteration experienced by oceanic crust (Gregory and Taylor, 1981; Alt et al., 1986). The robust nature of oxygen isotopes during prograde metamorphism is demonstrated by the preservation of values consistent with low-temperature alteration in massif eclogites and eclogite-facies rocks with a complex structural history (e.g., Getty and Selverstone, 1994; Pulitz et al., 2000). Thus, the large positive $\delta^{18}O$ values of Nyurbinskaya garnets strongly supports the involvement of subducted oceanic crust which has been altered at low-temperatures (<350 °C) in the mantle lithosphere underlying this region of the Siberian craton (Spetsius et al., 2008).

An earlier study of eclogitic xenoliths from Nyurbinskaya performed by our group (Spetsius et al., 2008) did not identify whether these mafic lithologies represent; 1) subducted crustal material of low-pressure origin, or 2) high-pressure cumulates derived from a heterogeneous asthenosphere containing a portion of recycled crust. This study explores these two different models of eclogite origin using new in-situ major- and trace-element data for garnets of known

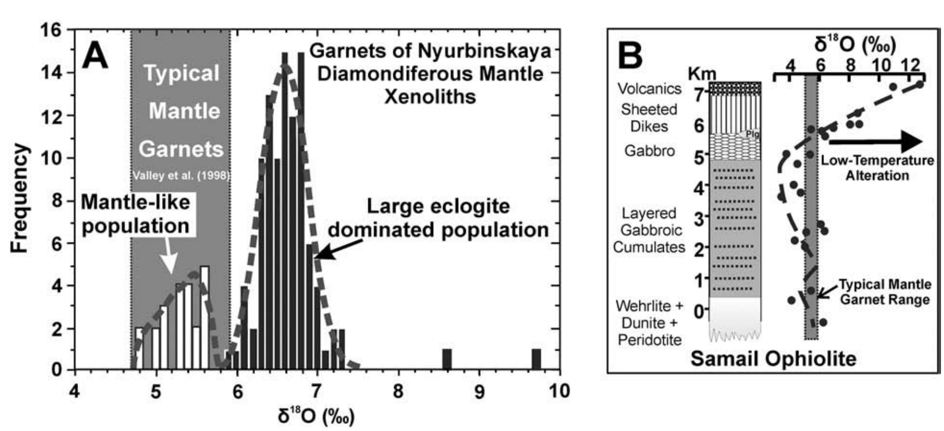

Fig. 2. A: Oxygen isotope compositions of garnets from mantle xenoliths of the Nyurbinskaya kimberlite pipe with corresponding probability density curves (after Spetsius et al., 2008), compared to the oxygen isotope composition of garnets of "typical" upper mantle (assuming garnet–zircon fraction is <0.1‰; Valley et al., 1998; Valley, 2003). The population with mantle-like $\delta^{18}O$ values (eclogite + peridotite) accounts for 18% of the analyzed xenoliths and ranges from 4.7 to 5.8‰ displaying a negatively skewed distribution. The large population dominated by eclogite xenoliths accounts for 82% of the population and ranges from 5.9 to 9.7‰ with a steep-sided bell-shaped curve that approaches a normal distribution (kurtosis = 0.188; calculated by including all values above 5.9‰). B: Oxygen isotope compositions observed in a suite of samples from the Samail ophiolite displaying high $\delta^{18}O$ values in the upper portion of oceanic crust altered at low-temperatures (after Gregory and Taylor, 1981; plg = plagiogranite).

Please cite this article as: Riches, A.J.V., et al., Subducted oceanic crust as diamond hosts revealed by garnets of mantle xenoliths from Nyurbinskaya, Siberia, Lithos (2010), doi:10.1016/j.lithos.2010.09.006

oxygen isotope compositions from the Nyurbinskaya kimberlite pipe, Yakutia, Siberia. We show that eclogitic garnets with high $\delta^{18}O$ have major- and trace-element compositions consistent with a low-pressure origin, and this provides some of the strongest evidence in support of models advocating subduction zone stacking during the generation of Archean cratonic lithosphere in the Siberian region.

2. Methods

In this study, a total of 89 garnets from 47 individual xenoliths were examined for their major- and trace-element compositions (Supplementary Tables A1 and B1). Due to significant alteration only fresh garnets were consistently recovered from these xenoliths, and these materials are no longer available for study as they have been crushed and used for other purposes (i.e., diamond extraction). For these reasons clinopyroxene analyses were not performed and xenoliths were classified on the basis of their garnet compositions (*c.f.*, Spetsius et al., 2008). In the majority of xenoliths (59%), between 2 to 5 high-purity garnet fragments were characterized, and these mineral separates correspond directly to sample aliquants on which oxygen isotope compositions were determined previously by Spetsius et al. (2008). This approach contrasts with major-element compositions reported by Spetsius et al. (2008), which were determined on polished sections of the same xenoliths. Major-element compositions were measured using a CAMECA SX-100 electron microprobe (EMP) at the University of Tennessee. Garnet compositions were determined using a 10 μm beam size, a 20 nA beam current, and an accelerating potential of 15 kV. Counting times for all elements were generally 20–30 s. with the exception of P (60 s.), Ni (60 s.), and Cr (40 s.). Standard PAP corrections were applied to all analyses. Precision and accuracy were monitored with natural and synthetic standards at intervals during each analytical session, and drift was within counting error. Detection limits (3σ above background) are typically <0.03 wt.% for SiO_2, TiO_2, Al_2O_3, MgO, CaO, Na_2O, and <0.05–0.1 wt.% for FeO, MnO, Cr_2O_3, NiO, and P_2O_5.

Trace-element analyses were performed on the same garnet fragments measured for their major-element compositions using a New Wave Research UP213 (213 nm) Nd:YAG laser-ablation system, coupled to a ThermoFinnigan Element 2 ICP-MS, at the University of Maryland. A laser repetition rate of 7 Hz and a photon fluence of 1.8 to 2.5 J/cm^2 were employed for all analyses, and each grain was analyzed with a 40 to 80 μm spot. Data were acquired across a mass range of ^{45}Sc to ^{238}U, and samples were ablated in a pure He atmosphere then mixed with argon before entering the plasma (*c.f.*, Günther and Heinrich, 1999). The NIST 610 glass standard was used for calibration purposes, and the BCR-2g glass standard was analyzed as an unknown to assess accuracy and reproducibility (Supplementary Table A2). Minor- and trace-element reproducibility of the standard was better than 3% (RSD) for all elements analyzed. Elemental abundances of BCR-2g were generally within error ($1\sigma_{stdev}$) of the accepted value, and reproducibility of the reference material was generally better than 5% (RSD). LA-ICP-MS analyses were normalized to the CaO value determined by EMP analysis. Detection limits across the ^{45}Sc to ^{238}U mass range were generally within the 10 to 200 ppb range and vary in response to the mass of sample ablated.

3. Results

3.1. Garnet major-element compositions

Garnets analyzed here have no observable exsolution textures. Major-element concentrations of garnets of Nyurbinskaya xenoliths cover a wide compositional range (Table 1, Fig. 3) similar to those determined on polished sections from the same xenolith suite (Spetsius et al., 2008). However, in detail, the major-element compositions of garnets from a limited number of xenoliths (notably N8, N14, N16, N20, N22, and N72) differ from those reported by Spetsius et al. (2008). These differences probably reflect heterogeneity between polished sections and grain separates from individual xenoliths, emphasizing the need for compositional data on the same population of garnets measured for oxygen isotope compositions. Several schemes exist for the classification of garnets on the basis of major-element concentrations (e.g., Sobolev et al., 1973; McCandless and Gurney, 1989; Taylor and Neal, 1989; Schulze, 2003; Grütter et al., 2004); each of these suggests lithological affinities and, in some cases, provides an indication of whether diamond may be present in the host xenolith. Here we adopt the classification scheme of Taylor and Neal (1989), which is based on chemical divisions suggested by Coleman et al. (1965).

Except for two garnets from one harzburgitic xenolith (N97), garnet grains ($n = 87$) from all other xenoliths ($n = 46$) within this suite are eclogitic (Supplementary Tables A1 and B1), and can be subdivided into three groups (A, B and C) based on the molar proportions of pyrope, almandine, and grossular (Fig. 3). Many individual xenoliths contain garnets of near-uniform major-element compositions, and only a limited number of xenoliths contain mixed populations of Group A, B, and/or C garnets (Supplementary Table A1). The majority (~78%) of eclogitic xenoliths contain Group B garnets. From studies of garnets in eclogitic and pyroxenitic xenoliths of the Slave Craton, Aulbach et al. (2007) suggested that distinct trends could be identified in CaO–Na_2O space. However, relatively large uncertainties on Na_2O abundances determined by EMP analysis (~0.1 wt.% Na_2O, 2σSt Dev, internal precision) make it difficult to identify systematic trends between these elements in the Nyurbinskaya data set (Fig. 3b).

Harzburgitic garnets studied here are derived from a single xenolith (N97) with a garnet $\delta^{18}O$ value of 4.98‰; the larger population of harzburgitic xenoliths studied by Spetsius et al. (2008) contain garnets of near-uniform oxygen isotope compositions (+4.98 to +5.07‰, where internal precision is typically 0.1‰; Valley et al., 1995), all of which are within the range of "typical" mantle garnet. In contrast, eclogitic garnets cover a wide range of $\delta^{18}O$ values and 75% of Group A, ~80% of Group B, and ~17% of Group C xenoliths have garnet $\delta^{18}O$ values above the mantle range. Further, a small number of xenoliths ($n = 3$) contain mixed garnet populations, and yield average garnet oxygen isotope compositions within and above the garnet mantle range. Major-element compositions of Nyurbinskaya garnets do not co-vary with oxygen isotopes and define a scattered distribution (e.g., Figs. 3a and 4a), which is consistent with the suggestion that major-element compositions are decoupled from oxygen isotope values (Spetsius et al., 2008).

3.2. Trace-element characteristics

Garnets from the 47 studied xenoliths display a wide range of trace-element compositions (Fig. 5, Table 1 and Supplementary Fig. A1 to A4). CI-chondrite normalized rare-earth-element (REE) patterns of eclogitic garnets generally show light-REE (LREE) depletion relative to heavy-REE (HREE), whereas abundances of HREE and middle-REE (MREE) are variable (Supplementary Fig. A2). All primitive mantle normalized trace-element patterns of these garnets display a saw-tooth pattern between Ba and Pr with Th_{PM}, La_{PM}, and Sr_{PM} depleted relative to adjacent elements, as broadly expected for trace-element partitioning into garnets (e.g., Green, 1994). The CI-chondrite normalized Zr/Hf values of all garnets are ~chondritic to supra-chondritic, and Ti anomalies ($[Ti/Ti^*]_{PM}$, where $Ti^* = [(Eu + Gd)/2]_{PM}$, and PM denotes normalization to primitive mantle) are commonly negative (Fig. 6). For the majority of garnets, Rb, Ba, Nb, La, Ta, Pb, and Th are below the limits of detection (Appendix B).

3.2.1. Group A eclogitic garnets

Group A eclogitic garnets have been identified in four xenoliths, and have HREE abundances ($Lu_N = 15.0$ to 32.5, where N denotes normalization to CI-chondrite) that are intermediate among eclogitic

Table 1
Major-, minor-, and trace-element abundances in representative garnets of Nyurbinskaya mantle xenoliths.

	GROUP A		GROUP B						GROUP C			HARZBURGITIC	
Xenolith	N47†	N104	N65	N72	N72	N77†	N77†*	N84*	N8*†	N31	N121	N97	N97
Details	Grain 4	Grain 3	Grain 3	Grain 2	Grain 3	Grain 4	Grain 6	Grain 3	Grain 4	Grain 2	Grain 4	Grain 1	Grain 3
$\delta^{18}O$ (‰)	6.03	8.59	6.75	6.52	6.52	5.39	5.39	4.78	6.35	5.43	5.25	4.98	4.98
SiO_2 (wt.%)	42.3	41.4	40.2	40.4	40.4	40.5	41.5	41.1	40.3	40.5	40.5	41.0	40.9
TiO_2	0.37	0.36	0.27	0.25	0.24	0.25	0.49	0.34	0.38	0.20	0.42	0.17	0.12
Al_2O_3	23.4	23.1	22.6	22.6	22.7	22.9	22.9	23.0	22.9	23.0	22.8	15.4	15.2
FeO	10.0	14.1	19.7	19.9	19.7	19.0	10.0	13.3	10.5	9.00	11.1	6.80	6.85
MnO	0.33	0.36	0.39	0.35	0.35	0.37	0.23	0.21	0.15	0.17	0.19	0.39	0.41
CaO	3.16	3.62	4.54	3.74	5.77	4.75	8.63	6.68	17.2	17.3	15.0	5.27	5.44
MgO	20.6	17.3	12.9	13.4	11.8	13.4	16.4	15.7	9.06	9.94	10.1	19.8	19.6
Na_2O	0.09	0.12	0.11	0.11	0.11	0.12	0.10	0.11	0.17	0.10	0.17	0.05	0.04
P_2O_5	0.04	<0.03	0.04	0.06	0.06	0.05	0.08	0.11	<0.03	0.13	0.18	0.07	0.06
Cr_2O_3	<0.03	0.04	0.03	0.02	0.05	0.03	0.18	0.06	<0.04	0.07	<0.01	11.0	11.2
Total	100.3	100.4	100.7	100.8	101.2	101.3	100.5	100.7	100.9	100.4	100.4	100.0	99.8
Mg#	78.8	68.7	54.6	54.8	51.6	56.4	75.4	69.0	61.4	67.6	61.8	84.2	84.1
Ca/(Ca + Mg) (mol.)	0.10	0.12	0.20	0.17	0.26	0.21	0.27	0.23	0.16	0.56	0.52	0.16	0.17
Pyrope (mol.%)	71.8	61.7	47.0	48.8	43.3	48.3	57.8	56.0	33.1	36.1	37.0	71.7	71.1
Almandine	19.6	28.3	40.3	40.7	40.7	38.6	19.9	26.5	21.5	18.3	22.9	13.8	13.9
Spessartine	0.65	0.72	0.81	0.74	0.80	0.76	0.45	0.42	0.32	0.34	0.40	0.81	0.85
Grossular	7.94	9.28	11.9	9.82	15.2	12.4	21.9	17.1	45.1	45.2	39.7	13.7	14.2
Sc (ppm)	38.6	38.3	45.3	66.2	32.3	47.0	30.0	19.6	9.52	157	11.5	152	150
Ti	2270	1968	76.7	2144	796	1249	2712	1836	2134	3900	2324	845	848
V	92.3	58.6	51.2	101	49.4	72.2	144	66.8	48.6	262	104	369	377
Cr	268	227	203	283	168	246	1150	412	80.2	659	55	–	–
Co	58.3	55.7	48.1	88.2	34.7	57.4	57.3	81.9	56.0	197	57.1	38.7	37.7
Ni	33.0	15.9	<15.3	19.3	12.1	25.3	77.6	102	64.0	107	52.3	44.8	42.2
Rb	<0.28	<0.25	<0.30	<0.34	<0.15	<0.29	<0.30	<0.33	<0.24	<0.89	<0.29	0.86	1.19
Sr	0.13	0.17	0.17	0.40	0.28	0.39	0.41	0.51	0.94	0.99	0.63	1.38	0.80
Y	9.8	22.0	15.6	56.1	21.0	32.0	6.93	6.86	5.86	75.0	5.65	6.50	6.48
Zr	57.8	12.8	2.77	13.2	5.88	12.0	21.8	17.4	24.4	36.6	20.7	34.5	35.4
Nb	0.14	0.05	0.03	<0.02	<0.02	<0.01	0.06	0.03	0.12	<0.06	0.09	0.20	0.19
Ba	<0.23	<0.18	<0.11	<0.21	<0.07	<0.15	<0.16	<0.23	<0.16	<0.62	<0.12	<0.16	<0.21
La	<0.02	<0.02	<0.02	<0.03	0.01	0.01	0.02	<0.01	0.02	<0.10	<0.01	0.08	0.02
Ce	0.06	0.04	0.10	0.14	0.10	0.14	0.09	0.14	0.37	0.16	0.13	0.77	0.45
Pr	0.02	0.02	0.04	0.06	0.04	0.05	0.03	0.07	0.22	0.08	0.08	0.28	0.22
Nd	0.39	0.28	0.33	0.76	0.44	0.70	0.47	0.95	2.90	1.07	1.34	3.89	3.58
Sm	0.49	0.36	0.24	0.92	0.36	0.56	0.63	0.89	2.49	1.75	1.52	1.68	1.79
Eu	0.29	0.29	0.13	0.79	0.20	0.35	0.38	0.57	1.14	1.08	0.75	0.59	0.63
Gd	1.09	1.34	0.54	4.07	1.15	1.87	1.52	1.41	3.26	5.24	2.19	1.86	1.82
Tb	0.22	0.38	0.19	1.06	0.36	0.58	0.29	0.22	0.43	1.33	0.32	0.29	0.29
Dy	1.61	3.20	2.17	8.82	3.13	4.86	1.72	1.40	1.89	12.3	1.46	1.54	1.55
Ho	0.40	0.84	0.68	2.20	0.79	1.24	0.30	0.27	0.25	3.1	0.22	0.25	0.25
Er	1.49	2.81	2.51	6.87	2.66	3.88	0.61	0.70	0.40	10.2	0.42	0.59	0.56
Tm	0.27	0.47	0.42	1.07	0.44	0.63	0.07	0.10	0.04	1.66	0.05	0.08	0.08
Yb	2.45	3.83	3.77	8.28	3.32	4.96	0.44	0.69	0.24	13.3	0.26	0.57	0.54
Lu	0.39	0.61	0.57	1.29	0.50	0.77	0.07	0.09	0.03	2.10	0.03	0.10	0.10
Hf	0.61	0.28	0.07	0.17	0.10	0.17	0.60	0.38	0.59	0.73	0.52	0.64	0.75
Ta	0.02	<0.01	<0.01	<0.01	<0.01	<0.01	0.01	<0.02	0.02	<0.05	0.01	0.03	0.02
Pb	<0.07	<0.08	<0.08	<0.17	<0.05	<0.10	<0.09	<0.13	<0.12	<0.49	<0.10	<0.08	<0.05
Th	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	0.00	<0.01	<0.01	<0.01	0.02	<0.01

Major-element compositions were determined by EMP and are reported as wt.%. Minor- and trace-element abundances (ppm) were determined by LA-ICP-MS. Garnets are classified on the basis of major-element compositions (after Taylor and Neal, 1989). Xenoliths containing mixed garnets are marked by a **†**. Eclogitic garnets marked with an asterisk (*) have Type 2 trace-element characteristics (details are given in the text). During analyses by LA-ICP-MS detection limits (<X) show modest variations between analytical points in response to the mass ablated in-run. Oxygen isotope compositions were determined on multiple grain fragments in each xenolith and were reported previously by Spetsius et al. (2008). A complete set of garnet compositions for Nyurbinskaya xenoliths studied here is included in the supplementary material (Appendix B).

garnets. These garnets generally display a positive slope within the HREEs ($[Dy/Yb]_N = 0.4$ to 0.6), and commonly define LREE-depleted profiles ($[La/Sm]_N \approx 0.01$, and $[Sm/Yb]_N = 0.10$ to 0.23). Group A eclogite garnets display both small negative and positive Eu-anomalies ($[Eu/Eu^*]_N = 0.95$ to 1.17, where $Eu^* = [(Sm + Gd)/2]_N$, and N denotes normalization to CI-chondrite). However, the magnitude of these Eu-anomalies is relatively small and generally within analytical uncertainties. These garnets cover a wide range of Ti, Y, V, and Cr concentrations that overlap the range observed in Group B and Group C eclogitic garnets (Supplementary material). Group A garnets cover a range of Zr/Ti, Cr/Ti, and Y/Ti values, but the V and Ti abundances in garnets with $\delta^{18}O$ values >5.9‰ define a positive linear correlation (Supplementary Fig. A3). All Group A eclogitic garnets define a broad negative correlation between Mg# and Yb content (Fig. 6a). The Ti anomalies of Group A garnets range from weakly negative to small positive values, (0.61–1.04, Fig. 6c), and cover a range of Zr/Hf values at low Sm/Hf (Fig. 7). Three garnets from a single xenolith (N47) display unusual REE-patterns, with a near-flat profile between Eu and Ho ($[Eu/Ho]_N = \sim 0.70$), and elevated HREE relative to MREE ($[Sm/Yb]_N = 0.22$–0.23). In addition, among all eclogitic garnets, the N47 garnets contain the highest Zr abundances (up to 58 ppm), and have the highest Zr/Hf and low Sm/Zr values (86 to 98 and ~0.01, respectively).

3.2.2. Group B eclogitic garnets

Group B eclogitic garnets of Nyurbinskaya define two distinct populations (B1 and B2) on the basis of their trace-element characteristics (Fig. 5 and Supplementary Fig. A2). Type B1 and B2 garnets differ

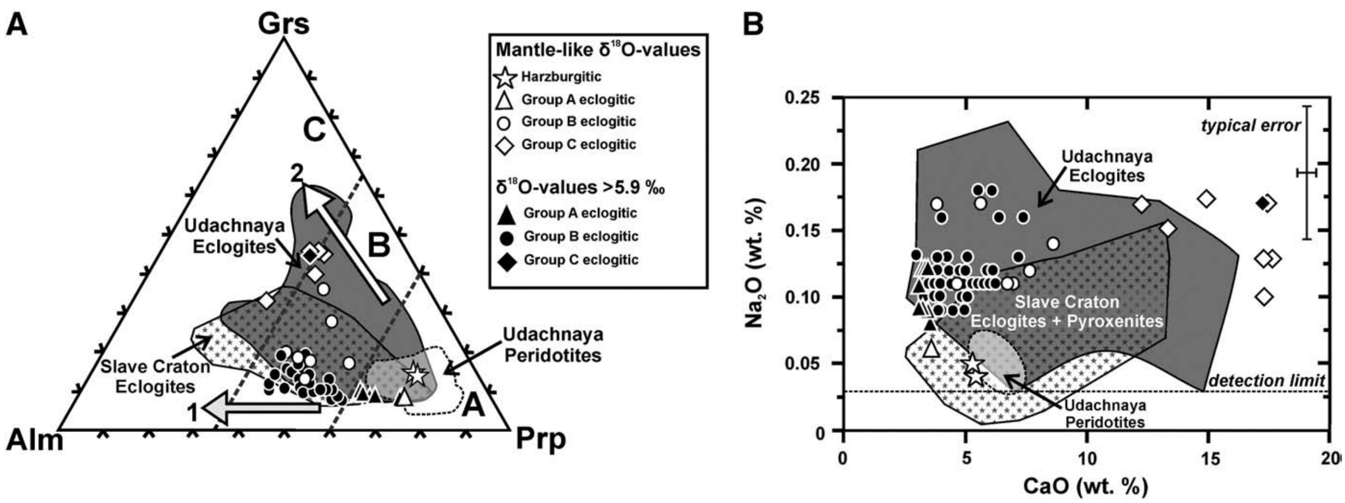

Fig. 3. Major-element compositions of Nyurbinskaya garnets, combined with oxygen isotope compositions reported by Spetsius et al. (2008). A: The A–B–C classification scheme of Taylor and Neal (1989) builds on that of Coleman et al. (1965). (Prp = pyrope, Alm = almandine, Grs = grossular). Arrow 1 indicates increasing almandine content, whereas arrow 2 indicates enrichment in grossular (after Snyder et al., 1997). B. Oxide content of garnet grains with typical error ($2\sigma_{stdev}$ derived from counting statistics) and detection limit of electron microprobe analyses. Data fields for xenoliths of the Udachnaya kimberlite, Siberia and Lac de Gras kimberlites, Central Slave Craton are derived from compositions reported by Jerde et al. (1993), Sobolev et al. (1994), Beard et al. (1996), Snyder et al. (1997), and Aulbach et al. (2007).

mostly in their HREE abundances and their MREE–HREE slope (Supplementary Fig. A2). The dominant population, Type B1, is the most common (69 grains of 38 xenoliths) and has trace-element patterns that are similar to those of Group A eclogitic garnets (Fig. 5). These garnets are LREE-depleted ($[La/Sm]_N = 0.01$ to 0.20, and $[Sm/Yb]_N = 0.03$ to 0.23), and have high HREE concentrations ($Lu_N = 18.7$ to 52.4) that generally display a shallow positive slope ($[Dy/Yb]_N = 0.31$ to 0.90). In contrast, rare Type B2 eclogitic garnets (2 grains, each from a different

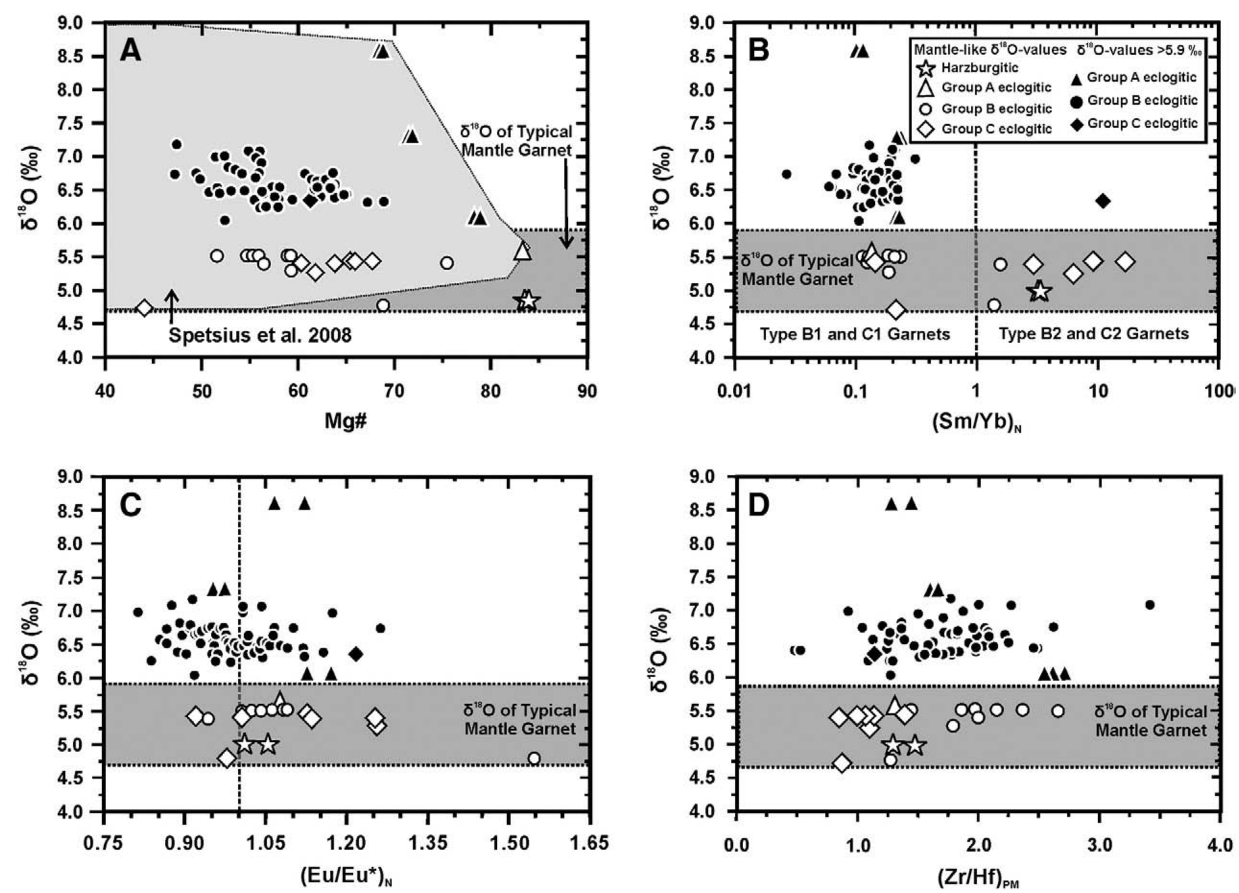

Fig. 4. Oxygen isotope compositions of Nyurbinskaya garnets (data from Spetsius et al., 2008) combined with A) Mg#, B) $(Sm/Yb)_N$, C) $(Eu/Eu^*)_N$ (where, $Eu^* = (Sm + Gd)/2$), and D) $(Zr/Hf)_{PM}$ values determined by EMP and LA-ICP-MS. Suffix N denotes normalization to CI-chondrite (McDonough and Sun, 1995), whereas suffix PM indicates normalization to primitive mantle (Sun and McDonough, 1989). The light gray field marks the range of compositions observed in a larger suite of garnets from Nyurbinskaya xenoliths (Spetsius et al., 2008), and the dark gray bar marks the range of "typical" mantle garnet $\delta^{18}O$ values (5.3 ± 0.6‰ $2\sigma_{stdev}$, Valley et al., 1998). The dashed line in C delineates the boundary between positive and negative Eu-anomalies that are discussed in the text.

Please cite this article as: Riches, A.J.V., et al., Subducted oceanic crust as diamond hosts revealed by garnets of mantle xenoliths from Nyurbinskaya, Siberia, Lithos (2010), doi:10.1016/j.lithos.2010.09.006

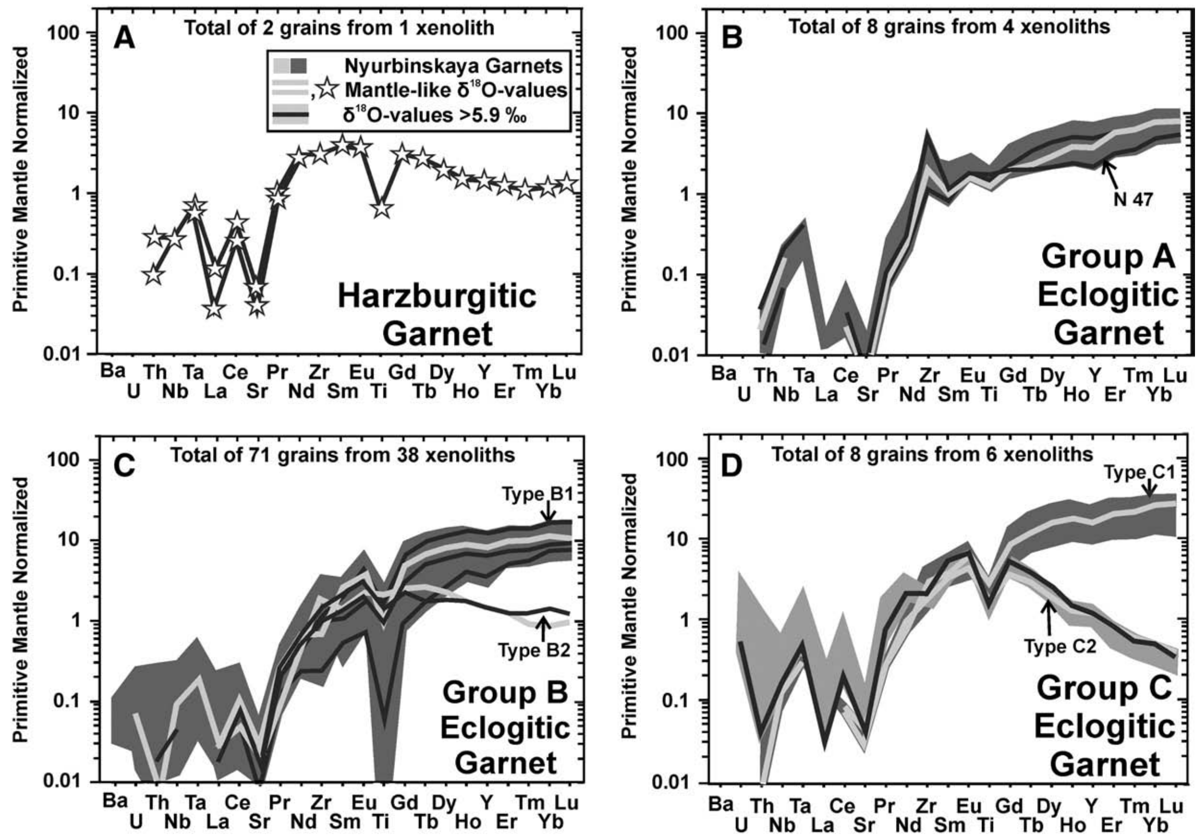

Fig. 5. Primitive mantle normalized trace-element abundances of Nyurbinskaya garnets determined by LA-ICP-MS. Representative grain compositions are shown as lines, and gray fields mark the data range of all Nyurbinskaya garnets studied here (Appendix B). Concentrations are normalized to the primitive mantle value of Sun and McDonough (1989).

xenolith) have convex-upward REE-patterns with a maximum CI-normalized value at Sm ($[La/Sm]_N \approx 0.02$, $[Sm/Yb]_N = 1.41$ to 1.56, and $[Dy/Yb]_N = 1.33$ to 2.56). The HREE abundances of Type B2 garnets ($Lu_N = 2.85$ to 3.66) are slightly lower than harzburgitic garnets ($Lu_N = 4.07$), and these eclogitic garnets display a near-flat slope between Tm and Yb ($[Tm/Yb]_N = 0.98$ to 1.00), which contrasts to the positive slope in harzburgitic garnets ($[Tm/Yb]_N \approx 0.80$). Compared to Type B1, B2 garnets with mantle-like $\delta^{18}O$ have the highest Mg# (69 to 75) and Ni abundances (77 to 102 ppm), but the lowest Y and Yb contents (~7 ppm and 0.44 to 0.69 ppm, respectively; Supplementary Table B1). Type B1 garnets generally have small positive and negative Eu-anomalies ($Eu/Eu^* = 0.85$ to 1.13; although many of these values are within analytical uncertainties), whereas Eu-anomalies of Type B2 garnets are positive and range from 1.13 to 1.55. Except for one garnet grain (Type B1) with a positive Ti-anomaly (1.14), all other Group B garnets have negative Ti-anomalies (0.01 to 0.90). Type B1 and B2 garnets show similar ranges of Sc, Ti, V, and Cr contents, with Zr/Ti values generally between 0.005 and 0.013, V/Ti values commonly range from 0.03 to 0.07, and Cr/Ti values are generally <1 (Supplementary Fig. A3). All Group B eclogitic garnets form a broad negative trend between Mg# and Yb contents (Fig. 6a), sub-parallel to the negative correlation observed in eclogitic xenoliths of the Slave Craton (Aulbach et al., 2007).

3.2.3. Group C eclogitic garnets

Group C garnets from 6 xenoliths also delineate two populations (C1 and C2) analogous to Type B1 and B2 groups (Fig. 5 and Supplementary Fig. A2). Garnets of Type C1 contain the highest Lu concentrations observed in the Nyurbinskaya xenolith suite ($Lu_N = 36.1$ to 103), whereas C2 garnets contain the lowest Lu abundances in all samples ($Lu_N = 0.81$ to 2.44). Type C1 garnets display no clear europium anomalies ($[Eu/Eu^*]_N = 0.92$ to 1.01), whereas $(Eu/Eu^*)_N$ values of C2 garnets are positive (1.13 to 1.25). All Group C garnets display negative Ti-anomalies (0.25 to 0.58), and Type C1 garnets have the highest Ti, Y and V contents observed in the Nyurbinskaya xenolith suite. Compared to Type C2, C1 garnets have the highest concentrations of Y (42 to 113 ppm), Cr (115 to 2626 ppm), and Sc (66 to 157 ppm). Group C garnets do not define a clear correlation between Mg# and Yb content, and generally scatter outside the elongate field defined by all other samples (Fig. 6a).

3.2.4. Harzburgitic garnets

Two harzburgitic garnet fragments from one xenolith (N97) have low HREE abundances ($Lu_N \approx 4.07$), and sinusoidal REE-profiles with a peak at Sm ($[La/Sm]_N \approx 0.02$, and $[Sm/Yb]_N \approx 3.21$ to 3.61). The sinusoidal shape of the harzburgitic REE-profile is supported by the shallow positive slope between Tm and Yb (Fig. 5a; $[Tm/Yb]_N \approx 0.8$), and no Eu-anomaly is present. In contrast to many eclogitic garnets that contain no detectable Rb, these harzburgitic garnets contain the highest Rb contents (0.9–1.2 ppm, Appendix B). In detail, Zr_{PM} and Ta_{PM} are elevated relative to Hf_{PM} and Nb_{PM} respectively (not plotted; $[Zr/Hf]_{PM} = 1.29$ to 1.48, $[Nb/Ta]_{PM} = 0.38$ to 0.45). Harzburgitic garnets have intermediate Ni and Sc contents (42.2 to 44.8 ppm and ~151 ppm, respectively) compared to eclogitic garnets. This type of garnet has low Ti and Y contents (~845 ppm, and ~6.48 ppm, respectively), but the highest Cr and V concentrations (~11.1 wt.% Cr_2O_3 and 369 to 377 ppm V) observed in the Nyurbinskaya sample suite (Supplementary Fig. A3). Further, these garnets have relatively low $(Ti/Ti^*)_{PM}$ values (~0.20) compared to eclogitic garnets (Fig. 6c). In addition, the Y and Zr contents of harzburgitic garnets (Supplementary Fig. A4) coincide with the field of low-temperature phlogopite metasomatism defined for peridotitic garnets by Griffin et al. (1999b).

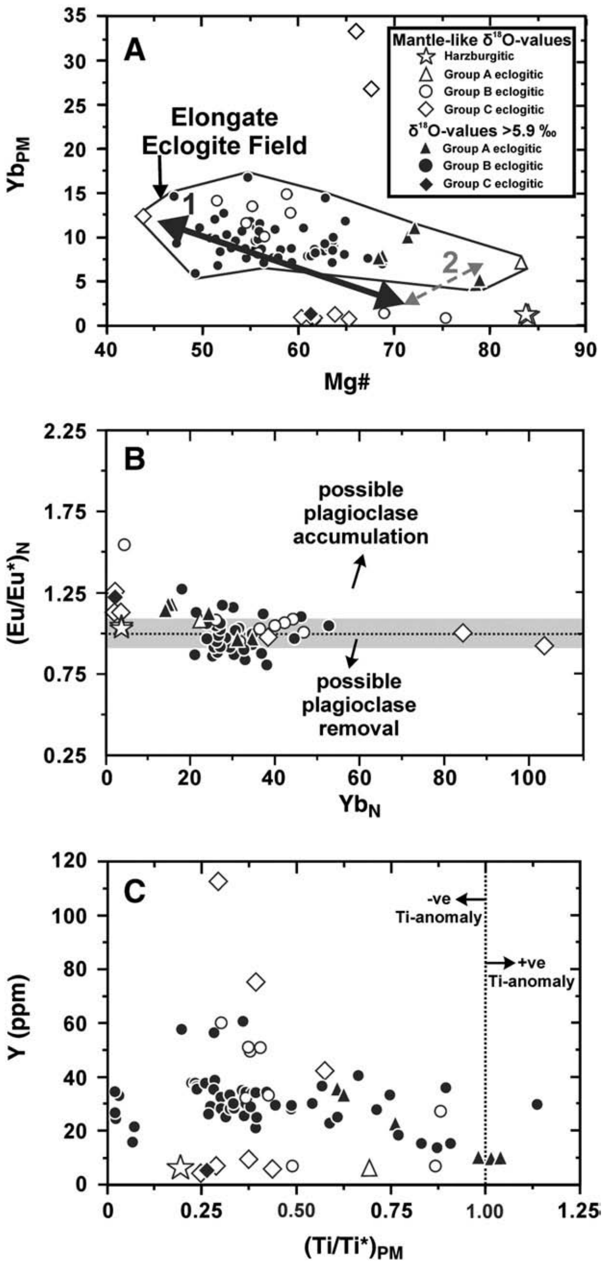

Fig. 6. Trace-element abundances and Mg# of Nyurbinskaya garnets measured by EMP and LA-ICP-MS analyses, and combined with oxygen isotope compositions reported by Spetsius et al. (2008). Subscripts N and PM denote normalization to the CI-chondrite value of McDonough and Sun (1995), and the estimate of primitive mantle given by Sun and McDonough (1989), respectively. Arrow 1 (A) corresponds to the negative correlation between Yb-abundances and Mg# of eclogitic garnets of Aulbach et al. (2007), and contrasts to the positive correlation (arrow 2) attributed to pyroxenitic garnets by these authors. The majority of Nyurbinskaya garnets have Mg# and Yb_{PM} values defining a broad field of data with a shallow negative slope (outlined in A). $Eu^* = [Sm + Gd]_N/2$, and $Ti^* = [Eu + Gd]_{PM}/2$. Typical uncertainties on Eu-anomalies determined by LA-ICP-MS ($2\sigma_{error}$) are shown by the gray bar that brackets a value of 1.

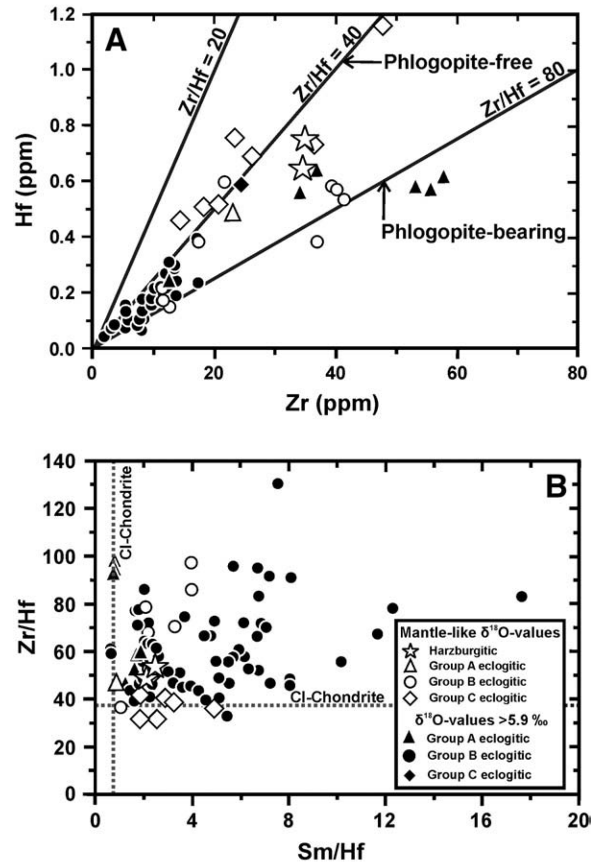

Fig. 7. Zirconium and hafnium concentrations determined by LA-ICP-MS (A) and ratios relative to Sm/Zr values (B) of Nyurbinskaya garnets combined with oxygen isotope compositions reported by Spetsius et al. (2008). Labels in A correspond to low and high Zr/Hf values associated with phlogopite-free and phlogopite-bearing eclogitic xenoliths, respectively (Jacob et al., 2009), and CI-chondrite values of B are from McDonough and Sun (1995).

3.3. Summary of results

Within the entire population of garnets from mantle xenoliths of Nyurbinskaya (47 xenoliths, 89 garnet fragments), minor- and trace-element abundances and inter-element ratios do not co-vary with $\delta^{18}O$ values and commonly display a scattered distribution indicating decoupling between oxygen isotope and trace-element compositions (Fig. 4b–d). Garnets with $\delta^{18}O > 5.9$‰ contain LREE-depleted patterns (with only one exception), whereas those grains with mantle-like $\delta^{18}O$ contain both the LREE-depleted and convex-upward REE-patterns (e.g., Fig. 4b). Although some correlations between $\delta^{18}O$, CaO, and FeO content, and trace-element ratios were reported in garnets from eight xenoliths of the Udachnaya kimberlite (Jacob et al., 1994; Jacob and Foley, 1999), these trends have not been observed in studies of larger xenolith populations (*c.f.*, Snyder et al., 1998), and are not evident in the large suite of garnets studied here. Generally, abundances of Hf, HREE, and elements of greater incompatibility do not co-vary with structural element (Fe–Mg–Ca) abundances. However, the abundance of Sm displays a broad positive correlation with Ca/(Ca + Mg) (Supplementary Fig. A1) suggesting that the concentration of this element generally increases as Ca-content increases in garnet. Additionally, Group B and C garnets contain sub-groups dominated by LREE-depleted profiles (Type B1 and C1), with minor amounts of garnets displaying convex-upward REE-profiles (Type B2 and C2). These major- and trace-element characteristics of garnets of Nyurbinskaya mantle xenoliths are broadly similar to those observed in previously studied peridotitic and eclogitic xenoliths from the Siberian Platform (e.g., Jerde et al., 1993; Sobolev et al., 1994; Snyder et al., 1997; Taylor et al., 2003a).

4. Discussion

4.1. Origin of highly-diamondiferous Nyurbinskaya xenoliths

A substantial portion of studied garnets fall in the classification for eclogites. As presented in Coleman et al. (1965) and Taylor and Neal (1989), garnets represent and resemble the major-element variations

Please cite this article as: Riches, A.J.V., et al., Subducted oceanic crust as diamond hosts revealed by garnets of mantle xenoliths from Nyurbinskaya, Siberia, Lithos (2010), doi:10.1016/j.lithos.2010.09.006

of their host xenoliths. Therefore, we discuss the origin of the host eclogites according to the groupings based on garnet compositions.

4.1.1. Origin of Group A eclogitic garnets

Group A eclogitic garnets of Nyurbinskaya have compositions that correspond to the broad negative correlation between Mg# and Yb-abundances evident in the vast majority of samples from Nyurbinskaya. High-pressure cumulates that fractionate garnet will define a positive slope between Mg# and Yb content, therefore the opposite trend observed in these eclogitic garnets suggests a low-pressure origin (after Taylor and Neal, 1989 and Aulbach et al., 2007). Moreover, the majority of Group A eclogitic garnets of Nyurbinskaya have $\delta^{18}O$ above the mantle range, and this may indicate the involvement of a component altered in near-surface environments during low-temperature (<350 °C) hydrothermal alteration. The small Eu-anomalies observed in the Group A eclogitic garnets of Nyurbinskaya are within analytical uncertainties, and the major- and trace-element compositions of these garnets resemble those of Group A eclogites recovered from the Mir kimberlite (Beard et al., 1996). In this case, Group A eclogites may not derived from gabbroic materials, but these magnesian eclogites may be derived from basaltic protoliths that have been subjected to low-temperature sea-water alteration prior to subduction and metamorphism (Beard et al., 1996).

Group A eclogites display some complexity and this is evidenced by garnets of N47 that show notable enrichment in Zr relative to Sm, and have Hf concentrations up to 0.61 ppm with Zr/Hf values up to 98. These characteristics are similar to garnets of phlogopite-bearing eclogites of the Kimberley diamond pipe, South Africa (Jacob et al., 2009), and Y and Zr abundances also place this sample within the field of low-temperature phlogopite metasomatism defined by Griffin et al. (1999b) (Supplementary Fig. 4A). For these reasons, the bulk-rock of xenolith N47 may contain accessory phases such as phlogopite and the conditions under which phlogopite may have formed in this xenolith are not well constrained.

The major- and trace-element systematics, and oxygen isotope compositions of garnets of Group A Nyurbinskaya eclogites contrast to those of several previous studies of magnesian (Group A) eclogitic xenoliths of Siberia, the Kaapvaal, and Western Africa cratons in which garnets have mantle-like $\delta^{18}O$ values, often have Cr_2O_3 contents >1 wt.%, and display extreme LREE-depletion (e.g., Taylor and Neal, 1989; Neal et al., 1990; Snyder et al., 1997; Barth et al., 2002). Additionally, eclogites with mantle-like oxygen isotope compositions are characterized by relatively unradiogenic $^{87}Sr/^{86}Sr$ (0.70298–0.70315), and depleted whole-rock Nd isotopic signatures at the time of kimberlite eruption ($\varepsilon Nd_{390\ Ma} = +4.3$ to $+6.7$; Snyder et al., 1997). These high-Cr Group A eclogites have not been interpreted as subducted materials, rather they are thought to represent high-pressure cumulates from melting of garnet-facies asthenosphere (Taylor and Neal, 1989; Neal et al., 1990). In addition, some Group A eclogitic garnets of the Udachnaya and Obnazhennaya kimberlites have $\delta^{18}O$ values >5.9‰, and have been attributed to two distinct protoliths; 1) subducted magnesian cumulates (Jacob et al., 1994), and 2) metasomatism of peridotitic mantle by carbonatitic and TTG melts (Taylor et al., 2003a). Together, this information suggests that Group A eclogites sampled around the globe, or even by a single kimberlite pipe (e.g., Nyurbinskaya), may not be accounted for by a single genetic model.

Group A eclogites are known to be transitional between eclogites of mantle origin (high-pressure cumulates) and those derived from subducted crustal materials (Taylor et al., 2003a). An increasing number of studies show that there are notable variations in the major-, minor-, and trace-element abundances, and oxygen isotope compositions of this group of eclogites when different sample suites are considered. These differences reflect not only the possibility of derivation from different mantle protoliths, but may also testify to distinct histories experienced by lithospheric blocks sampled by discrete kimberlite pipes.

4.1.2. Origin of Type B1 and C1 eclogitic garnets

Garnets of Type B1 and C1 eclogites cover a range of Mg#, and generally have low grossular contents but varying molar proportions of almandine (i.e., arrow 2 of Fig. 3a). The majority (85%) of LREE-depleted Type B1 and C1 eclogitic garnets define a broad negative array in Yb–Mg# space, and have $\delta^{18}O$ values above the mantle range These compositional characteristics, combined with the presence of omphacitic pyroxene and relatively high bulk SiO_2, CaO, Al_2O and Na_2O abundances (e.g., Beard et al., 1996), have been used to indicate that these groups of eclogite xenoliths are derived from gabbroic cumulates and basaltic portions of oceanic crust that have experienced plagioclase accumulation and removal at low-pressures prior to low-temperature alteration in near-surface environments (*c.f.*, Snyder et al., 1997 for a discussion). The absence of large Eu-anomalies in Nyurbinskaya eclogitc garnets is comparable to that observed in garnets of diamond-bearing eclogites from the Mir kimberlite, and it has been argued that this may reflect the influence of sea-water alteration and a small-degree of melt loss during subduction (Beard et al., 1996). Further, textural observations reported in previous studies of Group B and C eclogites (e.g., spongy-pyroxene rims and kelyphitic garnet boundaries recognized by Taylor and Neal, 1989; Jerde et al., 1993; Snyder et al., 1997) may support the notion that these xenoliths have experienced a small-degree of melt removal, and this has been linked to LREE-depletion observed in reconstructed bulk-rock patterns of fresh eclogite assemblages (e.g., Snyder et al., 1997). In addition, Jerde et al. (1993) suggested that Siberian eclogites derived from basaltic protoliths may sample both Type I and Type II basalts extruded in oceanic environments (BVSP, 1981), and Jacob and Foley (1999) argued that some eclogites with low HFSE abundances may be derived from island arc tholeiites.

Taken together, this evidence suggests that Group B1 and C1 eclogites are derived from subducted crustal materials altered in near-surface environments that sample both gabbroic and basaltic precursors widely considered to be produced in oceanic-spreading environments, with only minor proportions of subducted crust from arc-terranes that, for reasons of buoyancy, may be significantly more difficult to force to mantle depths (e.g., Cloos, 1993).

4.1.3. Origin of harzburgitic, Type B2, and C2 eclogitic garnets

Harzburgitic garnets of Nyurbinskaya are Mg and Cr-rich, and have mantle-like oxygen isotope compositions. Results of experimental work conducted within the pressure range of the garnet stability field and under conditions appropriate for diamond formation, indicate that sub-calcic Cr-rich garnets analogous to those of Nyurbinskaya harzburgitic xenoliths cannot be produced by high-pressure melting of fertile peridotite (Green et al., 1986; Brey et al., 1990; Canil and Wei, 1992), but may instead reflect metamorphism during subduction of spinel-facies melting residues with high-Cr/Al values (Bulatov et al., 1991; Stachel et al., 1998; Stachel and Harris, 2008). The sinusoidal REE-patterns observed in two harzburgitic garnet grains of a single Nyurbinskaya mantle xenolith are inconsistent with LREE-depleted patterns predicted by models of equilibrium mantle melting and may instead reflect post-melting modification (e.g., Stachel et al., 1998; Stachel and Harris, 2008). Sinusoidal REE-patterns are characteristic of harzburgitic lithologies observed in cratonic mantle across the globe (e.g., Pearson et al., 2003), and are typical of harzburgitic garnet inclusions in diamonds worldwide (see reviews by Taylor and Anand, 2004; Stachel et al., 2004; Stachel and Harris, 2008), suggesting that the trace-element characteristics of harzburgitic garnets from Nyurbinskaya do not result from interaction with the host kimberlite. The trace-element systematics of these garnets may reflect a two-stage process involving melt depletion that generates a LREE-depleted harzburgite (containing LREE-depleted garnet) and this may be recorded by the positive slope between Tm and Yb. Later, elements of greater incompatibility may be added to create sinusoidal REE-patterns in harzburgitic garnets (e.g., Taylor et al., 1996; Stachel et al., 1998), and this two-stage model may also apply to Group B and C garnets with Type 2 trace-element characteristics. Metasomatic agents considered responsible for the modification of incompatible element contents in mantle environments include; 1) Ti-poor carbonatitic melts (e.g. McDonough and Frey, 1989), 2) C–O–H–N–S-rich fluids (e.g., Stachel et al., 1998), and 3) migration of a silicate melt ± fractionation of that melt during melt percolation (e.g., Navon and Stolper, 1987; Bodinier et al., 1990).

To assess whether the trace-element compositions of harzburgitic, Type B2, and Type C2 garnets can be formed through interaction with kimberlite melt, we have utilized garnet-melt partition coefficients observed in kimberlite megacrysts (Fujimaki et al., 1984) and experimental work (Green et al., 1989; garnet–tholeiitic basalt partitioning at 25 kbar and 1100 °C). The results of this calculation show that equilibrium melt contains relatively high LREE abundances, significantly lower HREE abundances, and have relatively low Ti_{PM} compared to $MREE_{PM}$ (Fig. 8). Although the concentration of some elements (e.g., Ce, Sm, Eu, Gd, Dy, Ti, Hf, and Zr) overlaps the range observed in Yakutian kimberlites, other elements (e.g., Sr, Nb, Yb, Er, Y, Lu; Fig. 8 and Supplementary Table B1) are not accounted for by interaction with kimberlite melt, suggesting the involvement of other types of fluid or melt. Further, trace-element compositions of a calculated equilibrium melt of N97 harzburgitic garnets from Nyurbinskaya contrast to those calculated for harzburgitic silicate inclusions in diamonds of Akwatia, Ghana (Stachel and Harris, 1997), but display similar trace-element profiles to equilibrium melts of lherzolitic diamond of this locality.

The metasomatic agent responsible for modifying the trace-element characteristics of harzburgitic, Type B2 and C2 eclogitic garnets of Nyurbinskaya is LREE-enriched, HFSE-depleted (Ti, Zr, Hf) and it may be tempting to attribute these characteristics to carbonatite interaction. However, Ba and Sr contents of harzburgitic garnets are lower than those resulting from equilibrium with carbonatitic magmas (*c.f.,* Ionov, 1998, for an example of the trace-element compositions of carbonatite products). Additionally, garnet trace-element compositions requiring interaction with a LREE-enriched, HFSE-depleted agent have been identified in diamond inclusions recovered from kimberlites of Mir, Udachnaya, and Aikhal (*c.f.,* Taylor et al., 2003b and references therein), demonstrating the wide-spread nature of this metasomatic signature within the Siberian Craton. The physical and chemical constraints on the origin of this type of metasomatism have been discussed by several authors (e.g., Stachel et al., 1998; Taylor et al., 2003b; Taylor and Anand, 2004) and are not repeated here. These authors have concluded that although limited data are available on the solubility of major- and trace-elements in C–O–H–N–S bearing fluids, these may be likely metasomatic agents in diamond-bearing lithosphere and may even be linked to diamond formation during metasomatism (e.g., Liu et al., 2009).

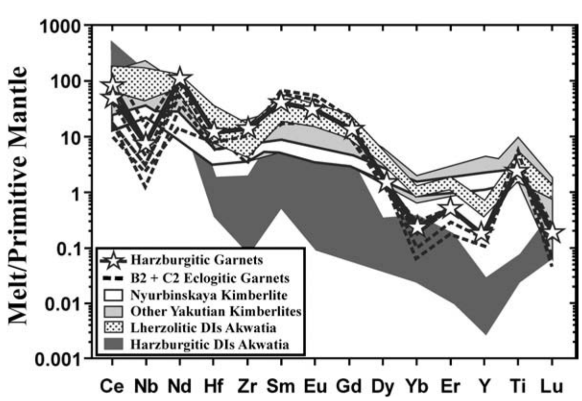

Fig. 8. Calculated trace-element composition of melt in equilibrium with harzburgitic, Type B2, and Type C2 eclogitic garnets of Nyurbinskaya (Supplementary Table B1). The trace-element composition of kimberlites from Nyurbinskaya and other kimberlite pipes of the Siberian Platform were determined by Lapin et al. (2007), and melts in equilibrium with peridotitic diamond inclusions (DIs) of Akwatia, Ghana were reported by Stachel and Harris (1997). All values are normalized to the primitive mantle estimate of McDonough and Sun (1995), and details of the calculation are given in the text.

4.2. Synopsis of the petrogenesis of Nyurbinskaya garnets

Combined oxygen isotope, major-, and trace-element characteristics of Nyurbinskaya garnets indicate the involvement of a significant portion of subducted crustal material in the mantle lithosphere underlying the Nyurbinskaya kimberlite. In particular, a negative slope between Mg# and Yb-contents, and oxygen isotope compositions above the mantle range suggest that Group A, Type B1, and C1 garnets are derived from the upper portions (basaltic and gabbroic) of oceanic crust that have been metamorphosed during subduction and incorporation into the lithosphere. Further, eclogites may experience a small-degree of melt loss, and in many cases have bulk-rock compositions that are not consistent with an origin as restites resulting from TTG melt production (*c.f.,* Snyder et al., 1997 and Taylor et al., 2003a for a discussion). In addition, incompatible trace-element characteristics of harzburgitic, Type B2, and C2 garnets provide evidence of metasomatic interaction, and indicate that C–O–H–N–S-rich fluids may have modified these materials in lithospheric environments. The diverse major- and trace-element characteristics of Nyurbinskaya garnets, and those of xenoliths across the Siberian craton, suggest that these mantle xenoliths sample a number of distinct protoliths of shallow origin, and indicates that the lithospheric mantle underlying the Anabar terrane may have a complex metasomatic history.

4.3. Evidence of subducted crust: implications for lithosphere formation

Our study provides strong evidence in support of subducted crustal materials in the mantle lithosphere underlying the Nyurbinskaya kimberlite pipe. In addition, our findings indicate that metasomatism by small-volume migrating CO–H–N–S-rich fluids may be a feature amongst these diamond-bearing xenoliths. The former study of Nyurbinskaya xenoliths by Spetsius et al. (2008) reported chlorotized phlogopite selveges surrounding diamonds, and suggested that their occurrence as chains or veinlets may indicate that diamonds formed during metasomatism by carbon-bearing fluids. Evidence of the C–O–H–N–S-rich metasomatic fluid(s) is widely recognized in cratonic lithosphere and diamond inclusions. Moreover, textural information, radiogenic (e.g., Rb–Sr, Sm–Nd) and carbon isotope compositions of global mantle xenolith suites commonly suggest that fluid and melt generated from crustal materials during subduction may be associated with metasomatism of the lithospheric mantle, hybridization of the mantle wedge, and diamond formation (e.g., Taylor et al., 2003b; Taylor and Anand, 2004; Spetsius and Taylor, 2008; Stachel and Harris, 2008). In addition, many cratonic mantle xenolith populations exhibit $\delta^{18}O$ values outside the mantle range (*c.f.,* reviews by Eiler, 2001; Jacob, 2004; Taylor and Anand, 2004), and it seems increasingly apparent that this isotope system may be a faithful indicator of the involvement of subducted crust despite metasomatism by melts and/or C–O–H–N–S-rich fluids capable of modifying incompatible trace-element characteristics.

Chronological information on mantle materials sampled by kimberlites compliments compositional constraints on the origin and chemical evolution of mantle lithosphere beneath cratons, and may provide a means for understanding the rates of global crustal generation. Although precise chronological constraints are not available for the Nyurbinskaya eclogites a number of previous studies have constrained the timing of major differentiation events

Please cite this article as: Riches, A.J.V., et al., Subducted oceanic crust as diamond hosts revealed by garnets of mantle xenoliths from Nyurbinskaya, Siberia, Lithos (2010), doi:10.1016/j.lithos.2010.09.006

experienced across the Siberian craton. Yakutian xenoliths and diamond inclusions have yielded osmium isotope compositions indicative of crustal generation and concomitant stabilization of peridotitic and eclogitic portions of lithosphere during the Archean (e.g., Pearson et al., 1995a,b, 1999), and this may be supported by Pb–Pb isotope systematics (e.g., Jacob and Foley, 1999). Further, the Yakutian kimberlite province is located within a region of thickened lithosphere (~200 km; Preistley and Debayle, 2003) that contains lithological and structural evidence of volcanic arc accretion. This basement sequence has a complex structural history involving significant prograde metamorphism (*c.f.,* Rosen et al., 1994, 2007). These features are common characteristics of cratonic regions around the globe (*c.f.,* Pearson and Witting, 2008 and references therein), and provide important geodynamic constraints that aid our understanding of the processes of craton formation. Moreover, signatures of subducted crust (major-, minor-, and trace-element, and stable isotope compositions) are widely recognized in these ancient terranes (e.g., Jacob and Foley, 1999; Barth et al., 2001, 2002; Aulbach et al., 2002; Pearson et al., 2003; Aulbach et al., 2007, 2009; Jacob et al., 2009; Liu et al., 2009; Spetsius et al., 2009), suggesting that subduction zone processes are fundamental components linked to the long-term survival of Earth's continental lithosphere. This widespread evidence of subducted protoliths supports models of thickened, persistent lithosphere formation by subduction zone stacking during ancient collision events (Helmstaedt and Gurney, 1995; Bleeker, 2003; Pearson and Witting, 2008), and these observations may have important implications for our understanding of Precambrian supercontinent assembly and models of crust mantle evolution.

5. Summary

New major- and trace-element data of Nyurbinskaya garnets with oxygen isotope compositions dominated by values above the mantle range indicate that this eclogite-rich mantle xenolith population samples materials of low-pressure origin. Harzburgitic garnets have the highest Mg# and Cr contents, and sinusoidal REE-patterns that are most consistent with spinel-facies melting followed by interaction with C–O–H–N–S-rich fluids. Group A, Type B1, and C1 eclogitic garnets commonly have $\delta^{18}O > 5.9‰$, and cover a range of Mg# that defines a negative correlation with Yb-abundances that is consistent with a low-pressure origin. These eclogites correspond to basaltic and gabbroic portions of oceanic crust and provide clear evidence of the presence of subducted crustal materials in the Siberian lithosphere. These results support models of craton formation by subduction zone stacking. In addition, metasomatic fluids that may themselves originate from subducted crust potentially play an important role in the modification of mantle lithologies and may be linked to diamond formation.

Supplementary materials related to this article can be found online at doi:10.1016/j.lithos.2010.09.006.

Acknowledgements

We are grateful to Allan Patchen and Dr. Richard Ash for their assistance during EMP and LA-ICP-MS analyses, respectively. Constructive comments by S. Foley, an anonymous reviewer, and the editor-in-chief, A. Kerr, are greatly appreciated. This research was supported by the Planetary Geosciences Institute at the University of Tennessee.

References

Agashev, A.M., Watanabe, T., Bydaev, D.A., Pokhilenko, N.P., Fomin, A.S., Maehara, K., Maeda, J., 2001. Geochemistry of kimberlites from the Nakyn field, Siberia: evidence for unique source composition. Geology 29, 267–270.

Agashev, A.M., Pokhilenko, N.P., Tolstov, A.V., Polyanichko, V.V., Mal'kovets, V.G., Sobolev, N.V., 2004. New age data on kimberlites from the Yakutian Diamondiferous Province. Doklady Earth Sciences 399, 1142–1145.

Alt, J.C., Muehlenbachs, K., Honorez, J., 1986. An oxygen isotopic profile through the upper kilometer of oceanic crust, DSDP Hole 504B. Earth and Planetary Science Letters 80, 217–229.

Aulbach, S., Stachel, T., Viljoen, S., Brey, G., Harris, J., 2002. Eclogitic and websteritic diamond sources beneath the Limpopo Belt — is slab-melting the link? Contributions to Mineralogy and Petrology 143 (1), 56–70.

Aulbach, S., Pearson, N.J., O'Reilly, S.Y., Doyle, B.J., 2007. Origins of xenolithic eclogites and pyroxenites from the Central Slave Craton, Canada. Journal of Petrology 48, 1843–1873.

Aulbach, S., Creaser, R.A., Pearson, N.J., Simonetti, S.S., Heaman, L.M., Griffin, W.L., Stachel, T., 2009. Sulfide and whole rock Re–Os systematics of eclogite and pyroxenite xenoliths from the Slave Craton, Canada. Earth and Planetary Science Letters 283, 48–58.

Barth, M.G., Rudnick, R.L., Horn, I., McDonough, W.F., Spicuzza, M.J., Valley, J.W., Haggerty, S.E., 2001. Geochemistry of xenolithic eclogites from West Africa, part I: a link between low MgO eclogites and archean crust formation. Geochimica et Cosmochimica Acta 65 (9), 1499–1527.

Barth, M.G., Rudnick, R.L., Horn, I., McDonough, W.F., Spicuzza, M.J., Valley, J.W., Haggerty, S.E., 2002. Geochemistry of xenolithic eclogites from West Africa, part 2: origins of the high MgO eclogites. Geochimica et Cosmochimica Acta 66 (24), 4325–4345.

Beard, B.L., Fraracci, K.N., Clayton, R.A., Mayeda, T.K., Snyder, G.A., Sobolev, N.V., Taylor, L.A., 1996. Petrography and geochemistry of eclogites from the Mir kimberlite, Yakutia, Russia. Contributions to Mineralogy and Petrology 125 (4), 293–310.

Bleeker, W., 2003. The late Archean record: a puzzle in ca. 35 pieces. Lithos 71, 99–134.

Bobrievich, A.P., Smirnov, G.I., Sobolev, V.S., 1960. On the mineralogy of xenoliths of the grossular–pyroxene–disthene rock (grospydite) from kimberlites of Yakutia. Geologiya i Geophysika 3, 18–24.

Bodinier, J.-L., Vasseur, G., Vernières, J., Dupuy, C., Fabriès, J., 1990. Mechanisms of mantle metasomatism: geochemical evidence from the Lherz Orogenic Peridotite. Journal of Petrology 31 (3), 597–628.

Boyd, F.R., Pokhilenko, N.P., Pearson, D.G., Mertzman, S.A., Sobolev, N.V., Finger, L.W., 1997. Composition of the Siberian cratonic mantle: evidence from Udachnaya peridotite xenoliths. Contributions to Mineralogy and Petrology 128, 228–246.

Brey, G., Kőhler, T., Nickel, K.G., 1990. Geothermobarometry in four-phase lherzolites I: experimental results from 10 to 60 kbar. Journal of Petrology 31, 1313–1352.

Bulatov, V., Brey, G.P., Foley, S.F., 1991. Origin of low-Ca, high-Cr garnets by recrystallization of low-pressure harzburgites. 5th International Kimberlite Conference, Extended Abstracts, pp. 29–31.

BVSP, 1981. Basaltic Volcanism on the Terrestrial Planets. Pergamon, New York.

Canil, D., Wei, K.J., 1992. Constraints on the origin of mantle-derived low Ca garnets. Contributions to Mineralogy and Petrology 109, 421–430.

Carlson, R.W., Pearson, D.G., James, D.E., 2005. Physical, chemical, and chronological characteristics of continental mantle. Reviews of Geophysics 43, 1–24.

Carswell, D.A., Dawson, J.B., Gibb, F.G.F., 1981. Equilibration conditions of upper-mantle eclogites: implications for kyanite-bearing and diamondiferous varieties. Mineralogical Magazine 44, 79–89.

Clayton, R.N., Goldsmith, J.R., Karel, K.J., Mayeda, T.K., Newton, R.C., 1975. Limits on the effect of pressure on isotopic fractionation. Geochimica et Cosmochimica Acta 39, 1197–1201.

Cloos, M., 1993. Lithospheric buoyancy and collisional orogenesis: subduction of oceanic plateaus, continental margins, island arcs, spreading ridges, and seamounts. Geological Society of America Bulletin 105, 715–737.

Coleman, R.G., Lee, D.E., Beatty, L.B., Brannock, W.W., 1965. Eclogites and eclogites: their differences and similarities. Geological Society of America Bulletin 76, 483–508.

Dawson, J.B., 1980. Kimberlites and their Xenoliths. Springer, Berlin.

Eiler, J.M., 2001. Oxygen isotope variations of basaltic lavas and upper mantle rocks. Reviews in Mineralogy and Geochemistry 43, 319–364.

Fujimaki, H., Tasumoto, M., Aoki, K., 1984. Partition coefficients of Hf, Zr, and REE between phenocrysts and groundmasses. Journal of Geophysical Research 89, B662–B672 (supplement).

Garlick, G.D., MacGregor, I.D., Vogel, D.E., 1971. Oxygen isotope ratios in eclogites from kimberlites. Science 172, 1025–1027.

Getty, S., Selverstone, J., 1994. Stable isotopic and trace element evidence for restricted fluid migration in 2 GPa eclogites. Journal of Metamorphic Geology 12, 747–760.

Green, D.H., 1994. Experimental studies of trace-element partitioning applicable to igneous petrogenesis — Sedona 16 years later. Chemical Geology 117, 1–36.

Green, D.H., Falloon, T.J., Brey, G.P., Nickel, K.G., 1986. Peridotite melting to 6 Gpa and genesis of primary mantle-derived magmas. Proceedings of the Fourth International Kimberlite Conference, pp. 181–183.

Green, T.H., Sie, S.H., Ryan, C.G., Cousens, D.R., 1989. Proton microprobe determined partitioning of Nb, Ta, Zr, Sr and Y between garnet, clinopyroxene and basaltic magma at high pressure and temperature. Chemical Geology 74, 201–216.

Gregory, R.T., Taylor Jr., H.P., 1981. An oxygen isotope profile in a section of Cretaceous Oceanic Crust, Samail Ophiolite, Oman: evidence for $\delta^{18}O$ buffering of the oceans by deep (>5 km) seawater-hydrothermal circulation at mid-ocean ridges. Journal of Geophysical Research 86 (B4), 2737–2755.

Griffin, W.L., O'Reilly, S.Y., 2007. Cratonic lithospheric mantle: is anything subducted? Episodes 30, 43–53.

Griffin, W.L., Ryan, C.G., Kaminsky, F.V., O'Reilly, S.Y., Natapov, L.M., Win, T.T., Kinny, P. D., Ilupin, I.P., 1999a. The Siberian lithosphere traverse: mantle terrains and the assembly of the Siberian Craton. Tectonophysics 310, 1–35.

Griffin, W.L., Shee, S.R., Ryan, C.G., Win, T.T., Wyatt, B.A., 1999b. Harzburgite to lherzolite and back again: metasomatic processes in ultramafic xenoliths from the Wesselton kimberlites, Kimberly, South Africa. Contributions to Mineralogy and Petrology 134, 232–250.

Grütter, H.S., Gurney, J.J., Menzies, A.H., Winter, F., 2004. An updated classification scheme for mantle-derived garnet, for use by diamond explorers. Lithos 77, 841–857.

Günther, D., Heinrich, C.A., 1999. Comparison of the ablation behaviour of 266 nm Nd: YAG and 193 nm ArF excimer lasers for LA-ICP-MS analysis. Journal of Analytical Atomic Spectrometry 14, 1369–1374.

Helmstaedt, H.H., Gurney, J.J., 1995. Geotectonic controls of primary diamond deposits: implications for area selection. Journal of Geochemical Exploration 53, 125–144.

Ionov, D., 1998. Trace element composition of mantle-derived carbonates and coexisting phases in peridotite xenoliths from alkali basalts. Journal of Petrology 39, 1931–1941.

Jacob, D.E., 2004. Nature and origin of eclogite xenoliths in kimberlites. Lithos 77, 295–316.

Jacob, D.E., Foley, S.F., 1999. Evidemce for Archean ocean crust with low high field strength element signature from diamondiferous eclogite xenoliths. Lithos 48, 317–336.

Jacob, D.E., Jagoutz, E., Lowry, D., Mattey, D., Kudrjavtseva, G., 1994. Trace elements in diamondiferous eclogites from Siberia: remnants of Archaean oceanic crust. Geochimica et Cosmochimica Acta 58, 5191–5207.

Jacob, D.E., Viljoen, K.S., Grassineau, N.V., 2009. Eclogite xenoliths from Kimberley, South Africa — a case study of mantle metasomatism in eclogites. Lithos 112, 1002–1013.

Jerde, E.A., Taylor, L.A., Crozaz, G., Sobolev, N.V., Sobolev, V.N., 1993. Diamondiferous eclogits from Yakutia, Siberia: evidence for a diversity of protoliths. Contributions to Mineralogy and Petrology 114, 189–202.

Kornilova, V.P., Fomin, A.S., Zaitsev, A.I., 2001. New type of diamondiferous kimberlite rocks on the Siberian platform. Regional Geology and Metallogeny 13, 105–117.

Lapin, A.V., Tolstov, A.V., Vasilenko, V.B., 2007. Petrogeochemical characteristics of the kimberlites from the Middle Markha Region with application to the problem of the geochemical heterogeneity of kimberlites. Geochemistry International 45, 1197–1209.

Liu, Y., Taylor, L.A., Sarbadhikari, A.B., Valley, J.W., Ushikubo, T., Spicuzza, M.J., Kita, N., Ketcham, R.A., Carlson, W., Shatsky, V., Sobolev, N.V., 2009. Metasomatic origin of diamonds in the world's largest diamondiferous eclogite. Lithos 112, 1014–1024.

MacGregor, I.D., Carter, J.L., 1970. The chemistry of clinopyroxenes and garnets of eclogite and peridotite xenoliths from the Roberts Victor Mine, South Africa. Physics of the Earth and Planetary Interiors 3, 391–397.

MacGregor, I.D., Manton, W.I., 1986. Roberts Victor eclogites: ancient oceanic crust. Journal of Geophysical Research 91 (B14), 14061–14079.

McCandless, T.E., Gurney, J.J., 1989. Sodium in garnet and potassium in clinopyroxene: criteria for classifying mantle eclogites. In: Ross, J. (Ed.), Kimberlites and Related Rocks Vol. 2: Their Mantle/Crust Setting, Diamonds and Diamond Exploration: GSA Special Publication, N° 14, pp. 827–832.

McDonough, W.F., Frey, F.A., 1989. Rare earth elements in upper mantle rocks. In: Lipin, B.R., McKay, G.A. (Eds.), Reviews in Mineralogy and Geochemistry, 21, pp. 100–145.

McDonough, W.F., Sun, S.-S., 1995. The composition of the Earth. Chemical Geology 120, 223–254.

Navon, O., Stolper, E., 1987. Geochemical consequences of melt percolation: the upper mantle as a chromatographic column. Journal of Geology 95, 285–307.

Neal, C.R., Taylor, L.A., Davidson, J.P., Holden, P., Halliday, A.N., Nixon, P.H., Paces, J.B., Clayton, R.N., Mayeda, T.K., 1990. Eclogites with oceanic crustal and mantle signatures from the Bellsbank kimberlite, South Africa, part 2: Sr, Nd, and O isotope geochemistry. Earth and Planetary Science Letters 99, 362–379.

Nimis, P., Zanetti, A., Dencker, I., Sobolev, N.V., 2009. Major and trace element composition of chromian diopsodes from the Zagadochnaya kimberlite (Yakutia, Russia): metasomatic processes, thermobarometry and diamond potential. Lithos 112, 397–412.

O'Hara, M.J., 1969. The origin of eclogite and ariégite nodules in basalt. Geological Magazine 106, 322–330.

O'Hara, M.J., Saunders, M.J., Mercy, E.L.P., 1975. Garnet-peridotite, primary ultrabasic magma and eclogite; interpretation of upper mantle processes in kimberlite. Physics and Chemistry of the Earth 9, 571–604.

Pearson, D.G., Witting, N., 2008. Formation of Archaean continental lithosphere and its diamonds: the root of the problem. Journal of the Geological Society of London 165, 895–914.

Pearson, D.G., Davies, D.R., Nixon, P.H., 1993. Geochemical constraints on the petrogenesis of diamond fades pyroxenites from the Beni Bousera Peridotite Massif, North Morocco. Journal of Petrology 34, 125–172.

Pearson, D.G., Snyder, G.A., Shirey, S.B., Taylor, L.A., Carlson, R.W., Sobolev, N.V., 1995a. Archaean Re–Os age for Siberian eclogites and constraints on Archaean tectonics. Nature 374, 711–713.

Pearson, D.G., Shirey, S.B., Carlson, R.W., Boyd, F.R., Pokhilenko, N.P., Shimizu, N., 1995b. Re–Os, Sm–Nd, Rb–Sr isotope evidence for thick Archaean lithospheric mantle beneath the Siberian Craton modified by multi-stage metasomatism. Geochimica et Cosmochimica Acta 59 (5), 959–977.

Pearson, D.G., Shirey, S.B., Bulanova, G.P., Carlson, R.W., Milledge, H.J., 1999. Re–Os isotope measurements of single sulphide inclusions in a Siberian diamond and its nitrogen aggregation systematics. Geochimica et Cosmochimica Acta 63 (5), 703–711.

Pearson, D.G., Canil, D., Shirey, S.B., 2003. Mantle samples included in volcanic rocks; xenoliths and diamonds. In: Holland, H.D., Turekian, K.K. (Eds.), Treatise on Geochemistry, 2, pp. 171–275.

Pokhilenko, N.P., Pearson, D.G., Boyd, F.R., Sobolev, N.V., 1991. Mega-crystalline dunites: source of Siberian diamonds. Carnegie Institution of Washington Yearbook 90, 11–18.

Preistley, K., Debayle, E., 2003. Seismic evidence for a moderately thick lithosphere beneath the Siberian Platform. Geophysical Research Letters 30 (3), 1118.

Pulitz, B., Matthews, A., Valley, J.W., 2000. Oxygen and hydrogen isotope study of high-pressure metagabbros and metabasalts (Cyclades, Greece): implications for the subduction of oceanic crust. Contributions to Mineralogy and Petrology 138, 114–126.

Rosen, O.M., Condie, K.C., Natapov, L.M., Nozhkin, A.D., 1994. Archean and Early Proterozoic Evolution of the Siberian Craton: a preliminary assessment. In: Condie, K.C. (Ed.), Archean Crustal Evolution, pp. 411–460.

Rosen, O.M., Levsky, L.K., Zhuravlev, D.Z., Spetsius, Z.V., Rotman, A.Y., Zinchouk, N.N., Manakov, A.V., Serenko, V.P., 2007. The Anabar Collision System as an element of the Columbia Supercontinent: 600 Ma of compression (2.0–1.3 Ga). Doklady Earth Sciences 417A (9), 1355–1358.

Schmickler, B., Jacob, D.E., Foley, S.F., 2004. Eclogite xenoliths from the Kuruman kimberlites, South Africa: geochemical fingerprinting of deep subduction and cumulate processes. Lithos 75, 173–207.

Schulze, D.J., 2003. A classification scheme for mantle-derived garnets in kimberlite: a tool for investigating the mantle and exploring for diamonds. Lithos 71, 195–213.

Shee, S.R., Gurney, J.J., 1979. The mineralogy of xenoliths from Orapa, Botswana. In: Boyd, F.R., Meyer, H.O.A. (Eds.), The Mantle Sample: Inclusions in Kimberlites and Related Rocks, pp. 37–49.

Shervais, J.W., Taylor, L.A., Lugmair, G.W., Clayton, R.N., Mayeda, T.K., Korotev, R.L., 1988. Early Proterozoic oceanic crust and the evolution of subcontinental mantle: eclogites and related rocks from southern Africa. Geological Society of America Bulletin 100, 411–423.

Snyder, G.A., Taylor, L.A., Crozaz, G., Halliday, A.N., Beard, B.L., Sobolev, V.N., Sobolev, N.V., 1997. The origins of Yakutian eclogite xenoliths. Journal of Petrology 38 (1), 85–113.

Snyder, G.A., Taylor, L.A., Beard, B.L., Crozaz, G., Halliday, A.N., Sobolev, V.N., Sobolev, N. V., 1998. Reply to a comment by D. Jacob et al. on 'The Origins of Yakutian Eclogite Xenoliths'. Journal of Petrology 39 (8), 1535–1543.

Sobolev, N.V., Kuznetsova, I.K., Zyuzin, N.I., 1968. The pretrology of Grospydite Xenoliths from the Zagadochnaya Kimberlite Pipe in Yakutia. Journal of Petrology 9 (2), 253–280.

Sobolev, N.V., Lavrent'ev, Y.G., Pokhilenko, N.P., Usova, L.V., 1973. Cr-rich garnets from kimberlites of Yakutia and their paragenesis. Contributions to Mineralogy and Petrology 40, 39–52.

Sobolev, V.N., Taylor, L.A., Snyder, G.A., Sobolev, N.V., 1994. Diamondiferous eclogites from the Udachnaya kimberlites pipe, Yakutia. International Geology Review 36, 42–64.

Spetsius, Z.V., 2004. Petrology of highly aluminous xenoliths from kimberlites of Yakutia. Lithos 77, 525–538.

Spetsius, Z.V., Taylor, L.A., 2008. Diamonds of Siberia: Photographic Evidence for their Origin. Tranquility Base Press, Tennessee, United States.

Spetsius, Z.V., Taylor, L.A., Valley, J.W., Deangelis, M.T., Spicuzza, M., Ivanov, A.D., Banzerak, V.I., 2008. Diamondiferous xenoliths from crustal subduction: garnet oxygen isotopes from the Nyurbinskaya pipe, Yakutia. European Journal of Mineralogy 20, 375–385.

Spetsius, Z.V., Wiggers de Vries, D.F., Davies, G.R., 2009. Combined C isotope and geochemical evidence for a recycled origin for diamondiferous eclogite xenoliths from kimberlites of Yakutia. Lithos 112, 1032–1042.

Stachel, T., Harris, J.W., 1997. Diamond precipitation and mantle metasomatism — evidence from the trace element chemistry of silicate inclusions in diamonds from Akwatia, Ghana. Contributions to Mineralogy and Petrology 129 (2–3), 143–154.

Stachel, T., Harris, J.W., 2008. The origin of cratonic diamonds-constraints from mineral inclusions. Ore Geology Reviews 34, 5–32.

Stachel, T., Viljoen, K.S., Brey, G., Harris, J.W., 1998. Metasomatic processes in lherzolitic and harzburgitic domains of diamondiferous lithospheric mantle: REE in garnets from xenoliths and inclusions in diamonds. Earth and Planetary Science Letters 159, 1–12.

Stachel, T., Aulbach, S., Brey, G.P., Harris, J.W., Leost, I., Tappert, R., Vijoen, K.W., 2004. The trace-element composition of silicate inclusions in diamond; a review. Lithos 77, 1–19.

Sun, S.-S., McDonough, W.F., 1989. Chemical and isotopic systematics of oceanic basalts: implications for mantle composition and processes. In: Saunders, A.D., Norry, M.J. (Eds.), Magmatism in Ocean Basins: Geological Society Special Publication, (London), pp. 313–345.

Taylor, L.A., Anand, M., 2004. Diamonds: time capsules from the Siberian Mantle. Chemie der Erde 64, 1–74.

Taylor, L.A., Neal, C.R., 1989. Eclogites with oceanic crustal and mantle signatures from the Bellsbank kimberlite, South Africa, part 1: Mineralogy, Petrography, and whole rock chemistry. Journal of Geology 97, 551–567.

Taylor, L.A., Snyder, G.A., Crozaz, G., Sobolev, V.N., Yefimova, E.S., Sobolev, N.V., 1996. Eclogitic inclusions in diamonds: evidence of complex mantle processes over time. Earth and Planetary Science Letters 142 (3–4), 535–551.

Taylor, L.A., Snyder, G.A., Keller, R., Remley, D.A., Anand, M., Wiesli, R., Valley, J.W., Sobolev, N.V., 2003a. Petrogenesis of group A eclogites and websterites: evidence from the Obnazhennaya kimberlites, Yakutia. Contributions to Mineralogy and Petrology 145, 424–443.

Taylor, L.A., Anand, M., Promprated, P., Floss, C., Sobolev, N.V., 2003b. The significance of mineral inclusions in large diamonds from Yakutia, Russia. American Mineralogist 88, 912–920.

Valley, J.W., 2003. Oxygen isotopes in zircon. Reviews in Mineralogy and Geochemistry 53, 343–385.

Valley, J.W., Kitchen, N., Kohn, M.J., Niendorf, C.R., Spicuzza, M.J., 1995. UWG-2, a garnet standard for oxygen isotope ratios: strategies for high precision and accuracy with laser heating. Geochimica et Cosmochimica Acta 59 (24), 5223–5231.

Valley, J.W., Kinny, P.D., Schulze, D.J., Spicuzza, M.J., 1998. Zircon megacrysts from kimberlite: oxygen isotope variability among mantle melts. Contributions to Mineralogy and Petrology 133, 1–13.