Platinum-group element contents of Karelian kimberlites:
implications for the PGE budget of the sub-continental lithospheric mantle

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Abstract. We present high-precision isotope dilution data for Os, Ir, Ru, Pt, Pd and Re in Group I and Group II kimberlites from the Karelian craton, as well as 2 samples of the Premier Group I kimberlite pipe from the Kaapvaal craton. The samples have, on average, 1.38 ppb Pt and 1.33 ppb Pd, with Pt/Pd around unity. These PGE levels are markedly lower, by as much as 80%, than those reported previously for kimberlites from South Africa, Brazil and India, but overlap with PGE results reported recently from Canadian kimberlites. Primitive-mantle-normalised chalcophile element patterns are relatively flat from Os to Pt, but Cu, Ni and, somewhat less so, Au are enriched relative to the PGE (e.g., Cu/Pd >25.000). Pd/Ir ratios are 3.6 on average, lower than most other mantle melts. The PGE systematics can be largely explained by two components, (i) harzburgite / lherzolite detritus of the SCLM with relatively high IPGE (Os-Ir-Ru) / PPGE (Rh-Pt-Pd) ratios, and (ii) a melt component that has high PPGE/IPGE ratios. By using the concentrations of iridium in the
kimberlites as a proxy for the proportion of mantle detritus in the magma, we estimate that the analysed kimberlites contain 3-27% entrained and partially dissolved detritus from the sub-continental lithospheric mantle, consistent with previous estimates of kimberlites elsewhere (Tappe S. et al., 2016, *Chem. Geol.* [http://dx.doi.org/10.1016/j.chemgeo.2016.08.019]). The other major component in the samples is melt, modelled to contain an average of 0.85 ppb Pt and 1.09 ppb Pd. Assuming that Group II kimberlites are derived from relatively metasomatised SCLM, our data suggest that the metasomatised Karelian SCLM is relatively poor in Pt and Pd. If our data are representative of other Group II kimberlites elsewhere, this result could imply that the PGE enrichment in certain continental large igneous provinces, including Bushveld, is not derived from melting of metasomatised SCLM.

**Keywords.** Kimberlite, platinum-group elements, sub-continental lithospheric mantle, Finland, Karelian craton, Kaapvaal craton, Premier kimberlite pipe, Kaavi kimberlite

1. **INTRODUCTION**

Several recent workers have suggested that cratonic large igneous provinces (LIPs) are more prospective to host deposits of platinum-group elements (PGE) than off-cratonic LIPs because cratonic magmas can scavenge PGE during ascent through the sub-continental lithospheric mantle (SCLM) (Zhang et al., 2008; Griffin et al., 2013; Hughes et al., 2015). However, the global database on SCLM rocks, consisting largely of mantle xenoliths in kimberlite and basalt, indicates that the cratonic SCLM is mostly relatively Pt and Pd depleted compared to the primitive mantle (Irvine et al., 2003; Pearson et al. 2003, 2004; Wittig et al., 2010; Maier et al., 2012; Barnes et al., 2015). It could be argued that the xenoliths are not representative of the PGE content of the SCLM as the most metasomatised, potentially relatively PGE enriched, mantle components should be relatively more susceptible to entrainment, disaggregation and dissolution during mantle melting.
and magma ascent. In order to address the question whether the SCLM can be locally enriched in PGE we determined the concentrations of PGE in 9 samples of kimberlite from the Karelian craton and 2 samples from the Kaapvaal craton.

Kimberlites are volatile-rich, potassic, ultramafic igneous rocks that are the main hard-rock source of diamond. They are normally sub-divided into Group I and Group II. Group I has abundant large crystals of olivine, in a matrix of olivine, monticellite, perovskite, spinel, mica, calcite, and serpentine, whereas Group II typically has abundant phlogopite ± olivine in a matrix of phlogopite, K richterite, and other diagnostic minerals. Mitchell (1995) reinstated the name “Orangeites”, originally coined by Wagner (1928), for the Group II kimberlites. Mineralogically, the orangeites are akin to olivine lamproites (except for the lack of carbonate in the latter), and it is now generally agreed that orangeites are the southern African equivalents of lamproites (Mitchell, 2006), rather than being synonymous with Group II kimberlites in general.

Kimberlites are characterized by high abundances of not only the most incompatible trace elements (e.g., Rb, Ba, Th, Nb, LREE, Sr), but also of compatible elements such as Mg, Ni and Cr. This paradoxical duality is interpreted to reflect the mixing of incompatible element-rich kimberlite "melt" with MgO-, Ni- and Cr-rich lithospheric mantle rocks, the dissolved and undissolved remnants of which form the rock and mineral detritus in kimberlites known as xenoliths, macrocrysts and megacrysts. In view of the mostly relatively low Pt and Pd contents of subcontinental lithospheric mantle harzburgite and lherzolite sampled by nodules and massifs (average around 3-5 ppb Pt, <1-3 ppb Pd; Irvine et al., 2003; Pearson et al., 2004; Becker et al., 2006; Smith et al., 2009; Wittig et al., 2010; Maier et al., 2012), it is likely that, if the SCLM contained more PGE-rich portions, these were more fusible and may have formed from metasomatic melts or fluids. As such, they could be represented by the kimberlites, notably those of Group II for which SCLM derivation is least controversial (Mitchell, 2006). In contrast, if both kimberlites and SCLM derived xenoliths are PGE depleted compared to the sub-lithospheric mantle, this would lend
further credence to the model that the SCLM in general is relatively PGE poor.

2. SAMPLES

The Karelian Group II samples analysed here are from the 1.2 Ga Seitäperä intrusion and the Lentiira dike swarm located between the villages of Kuhmo and Lentiira in eastern Finland (O’Brien and Tyni, 1999; O’Brien et al., 2007), 350 km inwards from the craton margin (Fig. 1). The most distinctive feature of these potassic, ultramafic rocks is their phlogopite-rich nature (Fig. 2a). Phlogopite occurs rarely as macrocrysts, but it is abundant as phenocrysts and microphenocrysts with relatively Ti-rich compositions similar to those of microphenocrysts in lamproites. The more primitive potassic rocks may also contain large amounts of olivine macrocrysts and, in some cases, abundant xenocrysts and xenoliths of mantle peridotite. Additional groundmass minerals include K-rich richterite, Mn-rich ilmenite, Cr-rich spinel zoned to Ti-magnetite, apatite and perovskite in a calcite + serpentine matrix. More evolved versions of this rock type contain abundant olivine phenocrysts, low-Al clinopyroxene, and phlogopite that is zoned to low-Ti tetraferriphlogopite, similar to mica zoning trends in Group II kimberlites (Mitchell, 1995). Because the rocks contain primary carbonate, they are more akin to South African Group II kimberlites (orangeites) than olivine lamproites or ultramafic lamprophyres. Similarly to orangeites, the Seitäperä Group II kimberlites have enriched Nd isotopic signatures with εNd around -7, yet with distinct compositions of $^{87}\text{Sr}/^{86}\text{Sr}$ around 0.704-0.706 (Fig. 3; Kargin et al., 2013; O’Brien, 2015).

The Finnish Group I kimberlite samples come from the 600 Ma old pipes of the Kaavi-Kuopio kimberlite clusters that are located in a craton margin tectonic setting in eastern Finland. Samples of pipes 1,2,3,5,9,10, and 14 are hypabyssal kimberlites. They all have typical Group I mineralogies (i.e. abundant large rounded grains (macrocrysts) of olivine in a matrix of subhedral to euhedral olivine, monticellite, perovskite, spinel, mica, apatite, calcite and serpentine), major and trace element and isotopic compositions (O’Brien and Tyni, 1999). They display the classic suite of
lithospheric mantle-derived xenoliths and xenocrysts (Fig. 2b), including (i) variable amounts of olivine showing a trimodal size distribution with abundant 0.1-0.4 mm size grains mostly interpreted as phenocrysts, 1-5 mm sized rounded macrocrysts, and occasional cm sized megacrysts. (ii) Mg-ilmenite, (iii) pyrope garnet derived from a range of sources including high Cr, Ca-depleted harzburgite to Ca-saturated lherzolite and Ca-rich wehrlite, to Ti-rich megacryst-compositions, and orange garnets derived from mantle eclogite, (iv) clinopyroxene comprising lherzolitic, low-Cr megacrystic, and eclogitic subgroups (O’Brien, 2015), and (v) spinels from upper mantle spinel lherzolites and rare chromites plotting within the diamond inclusion field (O’Brien, 2015). The rocks have relatively depleted isotopic signatures (Kargin et al., 2014) plotting close to South African Group I kimberlites as well as kimberlites in Canada (Tappe et al., 2016) in εNd vs Sr, albeit at slightly lower εNd (around 0) (Fig. 3, O’Brien and Tyni, 1999).

In addition to the Finnish kimberlites, we also analysed two kimberlite samples from the Kaapvaal craton. They represent Group I hypabyssal tuffissitic kimberlite breccias exposed in drill core from the 1151 Ma (Wu et al., 2013) Premier pipe, located in the vicinity of Pretoria. The rocks contain abundant large (up to 1 cm in diameter) macrocrysts and clasts of pelitic and quartzitic country rocks set in a fine grained groundmass. The lithophile element composition of the rocks has previously been reported in Maier et al. (2005), and geological-mineralogical data on the Premier pipe can be found in Bartlett et al. (1994) and de Wit et al. (2016).

3. ANALYTICAL METHODS

All of our samples have been characterised for major and trace elements at the Geological Survey of Finland in Espoo. The concentration of the lithophile elements in the Finnish samples is reported in O’Brien and Tyni (1999) and that of the Premier samples in Maier et al. (2005). Preparation of the Finnish kimberlite samples involved crushing of approximately 500 g of sample material into small chips with a hammer and hand-picking approximately 200 g of chips least contaminated by crustal
xenoliths. No attempt was made to separate mantle xenoliths from kimberlite. The hand-picked material was powdered in a steel grinding vessel. Major elements and S were determined by XRF on fusion beads, $\text{H}_2\text{O}$ and $\text{CO}_2$ by LECO analyzer, Ba, Ni, Cr, Cu, Zn, Sr and Zr by XRF on pressed pellets and REE, Hf, Ta, Pb, Th, U, Nb, Y, Rb, Sc, V, and Co by ICP-MS. Analytical methods are reported in O’Brien and Tyini (1999) and the data are provided in Table 1.

The PGE (Os, Ir, Rh, Pt, Pd) and Au have previously been determined by Ni-fire assay Te-co-precipitation followed by ICP-MS at the Geological Survey of Finland. Details of the analytical method are outlined in Juvonen et al. (2002) and the data are listed in Electronic Appendix 1. The concentration of Pt and Pd in the kimberlites were found to be very low (mostly <2 ppb) and lower than those reported in the literature for kimberlites (McDonald et al., 1995). In order to obtain better precision the samples were additionally analysed by high pressure asher digestion followed by online separation and quantification by isotope-dilution (HPA-ID) using ICP-MS following the method of Meisel et al. (2003). Rock powder (2 g) was weighed in a 30 ml quartz HPA tube. Five milliliters of 15 mol l$^{-1}$ $\text{HNO}_3$ and 2 ml of 12 mol l$^{-1}$ $\text{HCl}$ (both environmental grade) were added and variable amounts of a PGE-mixture spiked solution (UQAC-S-1) were added. The quartz lids were closed with a quartz disk and a PTFE (polytetrafluoroethylene) strip prior to high-pressure digestion at 300°C for 4.5 hr. The confining pressure was adjusted to 100 bars at the beginning and increased to 130 bars during digestion. After digestion, the samples were transferred to a 30 ml PFA container and directly sparged into the ICP-MS by bubbling argon through the sample for osmium isotope ratio determination. The samples were dried at 70–80 °C and the residuum was diluted with 2-10 ml of 0.1 mol l$^{-1}$ HCl. After centrifugation, 1 ml of the solution was introduced into a 1-m-long cation exchanged resin column (previously washed with 6 mol l$^{-1}$ HCl and equilibrated at 0.1 mol l$^{-1}$ HCl). The column was directly attached to the ICP-MS for 'online' separation. A multiple PGE-Au and Re stock solution (SCP Science, Canada) was used to monitor the mass drift. The enriched
isotopes in the spike mixture are $^{99}$Ru, $^{108}$Pd, $^{185}$Re, $^{190}$Os, $^{191}$Ir and $^{198}$Pt while the isotopes used for quantification are $^{101}$Ru, $^{106}$Pd, $^{187}$Re, $^{188}$Os, $^{193}$Ir and $^{196}$Pt. Because they are monoisotopic, Rh and Au cannot be determined by isotope dilution technique. The signal detected from the acid blank (0.1 mol l$^{-1}$ HCl) injected in the column prior to each analysis was subtracted from the raw count to the detector. When counts are present in the blank, the blank/sample proportion is far less than 1%. Because the on-line separation leads to changes in the signal intensity with time, it is difficult to convert the blank signal into an equivalent concentration, thus limits of detection can only be estimated and are ranging between 0.003-0.086 ng g$^{-1}$ for Re and Pd, respectively (Savard et al., 2010). Indeed, as it is stated in Savard et al. (2010): “The MLD cannot readily be calculated precisely for HPA-ID because the result is obtained by a summation of the signal integrated over a period of time that can vary, and during which the signal intensity also varies”. As also stressed in Meisel et al. (2003), the technique requires only concentrated acids and highly diluted HCl, thus procedural blank levels are controlled by the quality of the acid. Meisel et al. (2003) presented detection limits calculated on the absolute amount of total procedural blanks using reagent grade acids and found total procedural blanks ranging between 4 pg and 408 pg for Os and Pt, respectively. For the present work, the environmental acid grade used, in combination with a newer generation of ICP-MS instrument (Agilent 7700X) resulted in absolute procedural blanks ranging between 0.4 pg and 9.3 pg for Re and Pt, respectively (see Table 2). These values are lower than the detection limit (Table 3). More details on the method are provided in Meisel et al. (2003), Savard et al. (2010), Meisel and Moser (2004a, b) and Savard et al. (2010).

Table 3 reports results obtained for the reference materials analyzed together with the kimberlite samples to monitor quality. Results for OKUM agree with certificate values within the confidence limits. For Re and Os there are no certificate values, therefore the values obtained by Savard et al. (2010) are listed for comparison. The results for Re are similar, however the Os value is considerably higher, at 1.2 ng/g vs 0.79 ng g$^{-1}$. It is possible that the difference in Os values is real,
and due to inhomogeneity in the sample, given the low concentration of Os and the low sample weight used in isotope dilution. A round robin has recently been conducted on PGEs, Re and Au in OKUM in order to obtain certification of OKUM following ISO guidelines (Kane and Potts, 1997), however the report is not yet available (Thomas Meisel, personal communication). From 8 participating laboratories who provided Os concentrations using various preparation and analytical techniques (fire assay, Carius tube, isotope dilution, acid digestion, fusion, Q-ICP-MS, MC-ICP-MS), the data range from ~0.4 to ~1.6 ng g\(^{-1}\) Os, with >20% of the provided results showing values >1 ng g\(^{-1}\) Os. It was suggested that the working value for Os in OKUM after “outlier rejection” should be 0.85 +/- 0.12 ng g\(^{-1}\) Os, thus RSD is ~14% suggesting Os distribution in OKUM is inhomogeneous (Thomas Meisel, personal communication). In the present work, two aliquots of OKUM were analyzed and the relative standard deviations for Ru, Pd and Pt are < 5%. However for Os, Ir and Re the relative standard deviations are higher, ranging between 15% and 21% RSD. The higher relative standard deviations for these elements may be the result of a combination of low sample weights, low concentrations and inhomogeneous distribution of the elements in the sample. Again, results from the round robin show that RSD for Ir, Ru, Pt are around 12-13%.

4. RESULTS

The results for PGE by HPA-ID-ICP-MS analyses cover a similar range to those obtained by Ni-fire assay (Fig. 4). Combined Pt + Pd contents of the analysed Group I and Group II kimberlites from Finland and South Africa are generally below 5 ppb (Table 2). On average, the samples contain 1.38 ppb Pt and 1.33 ppb Pd (Pt/Pd ~1). There is very little difference between Group I and Group II, except that average Au contents of Group II samples are about twice those in Group I (1.19 ppb vs 0.63 ppb, respectively). There is also little difference in PGE contents between the Premier samples and the other kimberlites, a surprising result in view of the fact that Premier intruded rocks of the Bushveld Complex, the world’s largest repository of PGE. Combined IPGE
contents in our kimberlite samples are between ~1 and ~4 ppb, with average values being 0.77 ppb Os, 0.51 ppb Ir, 1.22 ppb Ru, and 0.34 ppb Rh.

Most of the mantle normalised PGE patterns are moderately fractionated (Ir/Os 0.71*CI, Ir/Ru=0.64*CI, Ir/Rh=0.42*CI) (CI from McDonough and Sun, 1995), but Pd/Ir =2.18*CI and Pt/Ir =1.23*CI (Table 2). The Pd/Ir ratios are lower, and the metal patterns less fractionated, than in most other mantle derived magmas, with the exception of some komatiites and picrites (see compilations in Barnes et al., 1985; Maier and Barnes, 2004; Fiorentini et al. 2010; Day et al., 2013; Barnes et al., 2015). However, the metal patterns mostly show enrichment in Ni, Cu and Au relative to PGE on mantle normalized plots (Fig. 4a-c).

Our samples show markedly lower PGE contents than the kimberlites analysed by McDonald et al. (1995) who reported average concentrations of 1.45 ppb Os, 1.34 ppb Ir, 2.97 ppb Ru, 0.9 ppb Rh, 7.7 ppb Pt, 5.1 ppb Pd and 1.8 ppb Au in their Group I and Group II samples from South Africa and Brazil (Fig. 4e-f). Our PGE contents are also lower, by about 50%, than the results of Rao et al. (2014) for orangeites from the Bastar craton of India which showed average PGE contents of 3.5 ppb Ru, and ~3 ppb Pt and Pd each (Fig. 4g) However, our PGE levels are broadly similar to those of Tappe et al. (2016) who analysed Os, Ir, Ru, Pt, Pd and Re in two Canadian kimberlite clusters (Renard and Wemindji), using ID-N-TIMS for Os and ID-ICP-MS for the other PGE (Fig. 4h). It is thus possible that kimberlites have more variable PGE contents than reported by us. Alternatively, the relatively high PGE levels in the study of McDonald et al. (1995) could reflect the lower precision of the INAA method and the heterogeneity of the standard used; Tredoux and McDonald (1996) state coefficients of variation between 10 and 20% for the analysis of Pt and Pd in standard WITS-1 used by McDonald et al. (1995). Future work will undoubtedly clarify this issue.

5. DISCUSSION
5.1. Petrogenesis of platinum-group elements in kimberlites

The estimated concentrations of the PGE in the Earth’s primitive mantle (PM) range from 1-7 ppb for individual elements (Palme and O’Neill, 2003; Barnes and Maier, 1999, Puchtel et al., 2004, Becker et al., 2006; Fischer-Gödde et al., 2011; Aulbach et al., 2016). Metal patterns in SCLM xenoliths show broadly similar levels on mantle normalized plots for Ni and the IPGE (Os, Ir, and Ru) at around 1 times primitive mantle, but Pt, Pd, Au and Cu tend to be variably depleted relative to the IPGE and PM (Pearson et al., 2003, 2004). This pattern is generally interpreted to be the result of extraction of basalt and komatiite from PM, with Ni and the IPGE behaving compatibly during much of the melting history due to partitioning into chromite, olivine, sulfides and alloys, whereas Pt, Pd, Au and Cu are incompatible (Barnes et al., 1985; Peach et al. 1990; Fleet et al. 1991, 1999; Sattari et al. 2002; Lorand et al., 2013; Mungall and Brenan, 2014; Aulbach et al., 2016; Brenan et al. 2016). Current estimates of the average PGE content of the SCLM suggest about 50-70% less Pt and Pd than in PM (Irvine et al., 2003; Pearson et al., 2004; Maier et al., 2012), but there is a certain degree of regional variation, with some cratons being relatively depleted compared to the global average (Greenland, Wittig et al., 2010) whereas others are slightly enriched (see compilation of data for various cratons in Maier et al., 2012). Rhodium behaves in a transitional manner, as seen by its relatively slight depletion relative to IPGE in the xenolith data (Fischer-Gödde et al., 2011; Maier et al., 2012; Barnes et al., 2015).

Assuming an average concentration of 250 ppm S in the convecting mantle (approximately 0.06% sulfide; McDonough and Sun, 2005), and a S solubility in basalt - komatiite at sub-lithospheric depth (>100-150 km) on the order of 500-1000 ppm (Mavrogenes and O’Neill, 1999), the bulk of the mantle sulfide is normally considered to be dissolved during moderate to large degree mantle melting (> 15-25%; Barnes et al., 1985; Mungall and Brenan, 2014). At lower degrees of melting, sulfide would be incompletely dissolved and thus the bulk of the highly chalcophile Pt, Pd, Au and Cu may remain in the mantle. The Pt and Pd depletion observed in the
mantle normalized patterns of mantle xenoliths therefore implies that the bulk of the SCLM is the residue of relatively large degree partial melting. Ballhaus et al. (2006) suggested that mantle sulfides can be entrained in magmas that formed at small degrees of mantle melting, but based on the low PGE contents reported here this model does not appear to apply to the kimberlites discussed in the present paper.

As few mantle SCLM xenoliths contain less than 1 ppb Pd, the SCLM appears to have been refertilized with melt or fluid that contained some Pd (Lorand et al., 2008; 2013), and this probably also introduced some Cu, Pt and Au. The precise nature and PGE content of the refertilising agent remains debated and was likely diverse (Simon et al., 2007; Kamenetsky et al., 2013), but the paucity of mantle samples that contain significantly higher Pt and Pd than PM (Pearson et al., 2004; Maier et al., 2012; Lorand et al., 2013) suggests that the agent was relatively PGE poor (but see Hughes et al. 2014 and Marchesi et al. 2014 for examples of relatively PGE rich SCLM xenoliths from the North Atlantic craton and the Rhonda peridotite massif).

Compared to most other mantle derived magmas (Fiorentini et al., 2010; Barnes et al., 2015) the kimberlites analysed in the present study have relatively unfractionated PGE patterns (but with several samples showing positive Rh anomalies) that can be modeled as a mixture of PGE-poor melt with high PPGE/IPGE ratios, and SCLM detritus (Fig. 4). For example, a 20% component of average SCLM (Maier et al., 2012) would result in ~ 0.7 ppb Ir, 0.8 ppb Pt, and 0.4 ppb Pd, accounting for most of the Ir in the analysed kimberlites, but only a fraction of the Pt and Pd. We can broadly replicate the measured patterns of the kimberlites, including the unique positive Rh anomalies, with a mixture of 10-20% average Kaapvaal peridotite xenoliths and 80-90% melt represented by the “liquid” component (see section 5.4 and Electronic Appendix 2 for calculation method) of MARID xenoliths, rocks that are composed mainly of mica, amphibole, rutile, ilmenite, and diopside (Dawson and Smith, 1977). The MARIDs have been interpreted, by some authors, as mixtures between SCLM derived melts and mantle peridotite (Waters, 1987; Gregoire et al., 2002).
They contain, on average, 2.4 ppb Pt and Pd each, and have more fractionated PGE patterns than the harzburgite xenoliths, with Pd/Ir ~11 and Cu/Pd ~26000 (Maier et al., 2012). It should be noted that in addition to peridotite and MARID melt, the elevated CO$_2$ contents of kimberlite requires a carbonate component, possibly derived from melting of trace carbonatite in the SCLM.

The similarity in PGE patterns between Group I and Group II kimberlites seen in figure 4 is notable, being that most authors interpret the 2 types of kimberlites to be derived from different mantle sources; Group I kimberlites have Nd and Sr isotope ratios around CHUR, putatively reflecting an asthenospheric source (e.g., Tappe et al., 2012), whereas Group II kimberlites have strongly radiogenic Sr and unradiogenic Nd isotopes, arguably reflecting a lithospheric mantle dominated source (Smith, 1983; Tainton and McKenzie, 1994; Mitchell, 2006; O’Brien, 2015). One might expect that melts from such diverse mantle domains should have contrasting PGE patterns.

We suggest that the similarity in PGE patterns can be understood within the context of a model whereby the bulk of the PGEs in both types of kimberlite is derived from refractory cratonic mantle material. The metasomatised component of the SCLM was PGE poor, consistent with the low sulfide and PGE contents of MARIDs and other metasomatised harzburgite/lherzolite xenoliths (Pearson et al., 2004; Maier et al., 2012; Lorand et al., 2013). Only Group II magmas fused potassic metasomes that control the lithophile trace element isotope systems such as Rb-Sr, Sm-Nd, Lu-Hf, thereby causing the enriched isotope signatures of the Group II magmas. In both the Karelian and Kaapvaal craton the Group II kimberlites are older than the Group I kimberlites; Possibly, Group I kimberlites retained their relatively depleted isotopic signature during traverse through the SCLM because the SCLM had become relatively refractory, following melting events that may have included the one which resulted in the Group I kimberlites.

We considered whether the PGE could have been extracted from the kimberlites by sulfide or sulfate that segregated during magma ascent and contamination in the crust. The kimberlites have variable amounts of lower, middle and upper crustal material, but none of these rocks contain
significant sulfur. In any case, kimberlite magmas are generally believed to ascend relatively fast within the crust, as suggested by the abundance of dense phases such as spinel crystals and large peridotite fragments, and by the presence of diamonds which would have been amenable to burning if the magma ponded or decelerated. This suggests that the ubiquitous PGE depletion of the kimberlites was not caused by the segregation of small sulfide droplets from the magmas.

5.2. Implications for the genesis of Bushveld Complex magmas

The Bushveld Complex is the largest layered intrusion on Earth. The extensive granitic, rhyolitic and mafic-ultramafic rocks outcropping across > 100,000 km$^2$ of southern Africa are the result of one of the world’s largest igneous events, with > 1 M km$^3$ of magma (Rajesh et al., 2013). The Complex has been interpreted as having crystallised from melts derived, at least in part, from the SCLM, based on Os isotopes and high Pt/Pd ratios of 1.6 that mirror, amongst known mantle reservoirs, only the SCLM (Maier and Barnes, 2004; Richardson and Shirey 2008, Barnes et al., 2010). However, if our kimberlite data are representative of not just the Karelian craton, but also the Kaapvaal craton, as suggested by the similarity in PGE patterns between the Karelian and Premier kimberlite samples, this interpretation seems difficult to reconcile with the low PPGE contents and the unfractionated PGE patterns in our samples, implying low PGE contents of the metasomatic component of the SCLM, as discussed above; Bushveld melts have not only much more fractionated PGE patterns than kimberlites (Fig. 5), but amongst the highest PGE contents of global magmas, and they also lack the positive Au anomalies observed in Group II kimberlites (Fig. 5). Furthermore, MELTS modelling indicates that partial melting of fertile SCLM at 30 kbar yields magma that has significantly lower SiO$_2$ and CaO, and far higher incompatible lithophile trace element contents than unevolved Bushveld B1 magma (Maier et al., 2016). We thus conclude that, if the PGE contents of the two analysed Premier kimberlites are representative of other Kaapvaal kimberlites, the PGE-rich Bushveld magmas are unlikely to be derived mainly from the
metasomatised portion of the SCLM. Alternatively, it is possible that the PGE contents of our
Premier kimberlite samples are not representative of Kaapvaal kimberlites, as potentially suggested
by the data of McDonald et al. (1995). However, as pointed out earlier, we feel that the INAA
method used by these authors lacks the precision required to reliably determine the low abundances
of PGE in kimberlites. Future studies will doubtlessly shed more light on this issue.

A further possibility is that the SCLM xenoliths so far analysed from the Kaapvaal craton
are non-representative. Marchesi et al. (2014) and Hughes et al. (2014) have reported elevated PGE
contents (and high Pt/Pd > 1) in the Rhonda SCLM massif and in xenoliths from the North Atlantic
craton. One could thus argue that the Kaapvaal SCLM too contains domains that are more enriched
in PGE than those samples by the xenoliths analysed so far, and that Bushveld magmas selectively
scavenged PGE from the ~ 150 km thick refractory SCLM during ascent. Additional work on the
Kaapvaal mantle rocks is required to resolve this issue.

5.3. Origin of elevated Au contents in the Group II kimberlites

The positive Au anomalies and elevated Au contents of the Group II samples relative to Group I
kimberlites could suggest that the metasomatised component of the SCLM is relatively enriched in
Au. This observation is potentially consistent with studies suggesting that the Au in orogenic Au
deposits is derived from alkaline magmas sourced from the lithospheric mantle (Hronsky et al.,
2012). However, due to the small number of Group II samples analysed by us this model remains
highly speculative and needs to be tested by analysis of further samples. Notably, a relative Au
enrichment was also observed in South African orangeites relative to South African Group I
kimberlites by McDonald et al. (1991).

5.4. Estimating the proportion of xenoliths, xenocrysts and primary kimberlite melt

Kimberlites contain abundant xenoliths and xenocrysts in various states of dissolution. This
is, for example, reflected in trajectories from kimberlite towards the lithospheric mantle data array in binary Ir vs Ru and, to a lesser degree, Ir vs MgO diagrams (Fig. 6). Previous authors have attempted to quantify the relative proportion of peridotite detritus and kimberlite liquid by using a variety of modelling approaches, including subtraction of olivine or peridotite from bulk kimberlite (Fraser and Hawkesworth, 1992; Le Roex et al., 2003; Kjarsgaard et al., 2009), mixing models of Os isotope ratios (Pearson et al., 2003; Tappe et al., 2016), Nd-Hf isotopes (Tappe et al., 2013), calculating melt composition from the composition of non-xenocrystic olivine (Arndt et al., 2010), analysis of quenched kimberlite melt (Price et al., 2000; Kopylova et al., 2007), and melt inclusions in olivine (Kamenetsky et al., 2013). With the exception of the melt inclusion work that yielded highly carbonaceous melt estimates, most of the proposed melt compositions show relatively good consistency (e.g., 20-30% SiO₂ and MgO, 1300-2400 ppm Cr, 800-1400 ppm Ni).

A potential shortcoming of the olivine-peridotite subtraction method is that kimberlites have MgO, Cr and Ni contents that differ by a factor of less than 2 from the composition of lithospheric mantle peridotites, rendering these elements relatively insensitive tracers of melt–peridotite mixing (Smith et al., 1985; Shee, 1986; Taylor et al., 1994; Berg and Carlson, 1998; Price et al., 2000, Le Roux et al. 2003, Harris et al., 2004, Becker and Le Roex, 2006).

In the present paper, we estimate the proportion of detrital material in kimberlite by using mass balance based on Ir contents. The reasoning behind this method is that Ir abundances in low-degree mantle melts and lithospheric peridotites show very different concentrations, and that Ir is mostly relatively immobile during mantle metasomatism and near surface alteration (but see Ackerman et al., 2009, and Lorand et al., 2013, for examples of Ir depleted metasomatized mantle rocks). Iridium was also chosen for more practical reasons: due to low reagent blanks and a lack of interfering ions it can be accurately determined not only by ID but also by conventional analytical techniques such as nickel sulfide fire assay followed by ICP-MS finish.
The following equations apply:

\[
Ci(pkm) = \frac{CIr(s) - f(a) \times Ci(a)}{1 - f(a)}
\]

(1)

\[
f(a) = \frac{CIR(s) - CIR(pkm)}{CIR(a) - CIR(pkm)}
\]

(2)

where \(Ci\) and \(CIR\) refer to the concentrations of an element of interest and iridium in the sample (s), the lithospheric mantle assimilant (a), and the primary kimberlite magma (pkm). \(f(a)\) refers to the volume fraction of the assimilant in the sample. To solve the equation for unknown elements one needs to know the composition of the assimilant and the Ir content of the primary kimberlite melt.

The composition of the assimilant is estimated by averaging the compositions of mantle xenoliths that have been recovered from kimberlite pipes. With the exception of a few pipes (notably Jagersfontein), this proved to be remarkably homogenous for the Kaapvaal and Karelian cratons, at 3.67 ppb Ir for the Kaapvaal craton (n=83, stdev 1.62ppb) and 3.62 ppb Ir for the Karelian craton (n=18; stdev 2.19, Pearson et al., 2004, Becker et al., 2006; Maier et al., 2012). Alard et al. (2000) have proposed that in cratonic peridotites Ir is mainly bound in sulfide minerals found as inclusions in olivine, but Luguet et al. (2007) showed that in the Lherz massif laurite inclusions may occur in addition to Pt-Ir-Os alloys. Although the Ir content of kimberlite thus basically is a measure of the amount of sulfide and alloy contamination, it is likely that kimberlite melt has assimilated bulk mantle peridotite, rather than selectively Ir-rich mineral phases; The latter are protected from selectively dissolving into kimberlite magma by the enclosing olivine crystals (macrocystal olivine) which show only minor reaction with the melt. Because the Ir content of kimberlite melt is low (see below) and \(D_{Ir}(ol-melt)\) is only around 2 (e.g., Brenan et al., 2005), possible removal of some phenocrystic olivine has no significant effect on the iridium content of the melt. Furthermore, a case can be made against gravitational removal of macrocrystal olivine (\(\rho=3.3\).
g/cm$^3$), because kimberlites contain abundant picroilmenite, garnet, and diamond xenocrysts, all of which have higher densities than olivine. Although nugget effect is always a possibility, we are confident that the extensive degree of assimilation by the rigorously mixing kimberlite magma provides us with well homogenised kimberlite samples in which Ir can be applied as a proxy for the degree of mantle assimilation.

The Ir content of the kimberlite parental magma can be estimated from general mantle melting constraints and the Ir content of other alkaline magma types; Firstly, Ir is largely bound in mantle sulfides (Alard et al., 2000) and, to a lesser degree, in olivine (Brenan et al., 2005). Partition coefficients between sulfide and silicate melt are between $10^5$ and $10^7$ (Fonseca et al., 2009; Mungall and Brenan, 2014), and between olivine and silicate melt are ~2 (Brenan et al., 2005). As a result, batch melting calculations yield <0.05 ppb Ir in melts with F<5%. Secondly, the available literature data indicate that most alkaline basalts have very low iridium contents (Vogel and Keays, 1997; Rehkämper et al., 1999; Tatsumi et al., 2000; Crocket et al., 2002; Barnes et al., 2015). Concentrations range from 0.002 to 0.365 ppb with a median value of 0.06 ppb (n=33). From these estimates we conclude that it is unlikely that primary kimberlite melts contain significantly more than 0.05 ppb iridium.

Calculations using Equation (2) suggest that the volume percentage of the lithospheric mantle material in Finnish Group I and II kimberlites is between 3-22 wt.%. The two Premier samples analysed have 18-27% mantle detritus. These estimates overlap with those of Tappe et al. (2016) for Canadian kimberlites (2-30%).

Equation (1) cannot be readily applied to calculate primary major and lithophile trace element “melt” compositions for our samples. This is because for major and lithophile trace elements, potential uncertainties are much larger than in the case of PGE, as the compositional difference between the contaminant (the xenoliths) and the kimberlite melt is far smaller than for PGE, and accumulation and fractionation effects are potentially far more significant; For example, a small
amount of cumulus phlogopite in the kimberlites dramatically changes the K and Rb contents of the rock.

6. CONCLUSIONS

(i) The present study presents high-precision isotope dilution data for Os, Ir, Ru, Pt, and Pd in kimberlites. Platinum and Pd contents of the rocks are generally very low (averaging around 1.3-1.4 ppb Pt and Pd each, 2-3 ppb combined IPGE), in contrast with some previous studies that proposed up to 4-7 times higher Pt and Pd contents and 2-3 times higher IPGE contents. We propose that the differences are largely due to the inferior precision and accuracy of those previous studies conducted by using INAA. In contrast, previous studies using ICP-MS (Tappe et al., 2016) have obtained PGE levels broadly overlapping with our results.

(ii) The PGE patterns of the kimberlites can be interpreted to represent mixtures of PGE poor melt and variably dissolved SCLM xenoliths and xenocrysts. Using Ir as a proxy, we have estimated the proportion of mantle detritus in the kimberlites to be between 3 and 28%.

(iii) Our data from the Karelian craton and the Premier pipe in South Africa suggest that the metasomatised component of the SCLM is relatively PGE poor. If the 2 Premier samples are representative of the Kaapvaal craton, this would imply that the high PGE contents and Pt/Pd ratios in Bushveld magmas are not derived from melting of the metasomatised Kaapvaal SCLM, and that the prospectivity of LIPs is not controlled by the interaction of the magmas with cratonic SCLM, as proposed in some recent studies.

(iv) The Group II kimberlites have positive Au anomalies in mantle normalised diagrams suggesting that the metasomatic component of the SCLM is relatively enriched in Au compared to PGE, and that the metasomatised SCLM could have supplied noble metals for orogenic Au deposits (Hronsky et al., 2012; Griffin et al., 2013).
We thank Dany Savard (UQAC) for conducting the ID analyses. S Tappe, L Ackerman, an anonymous reviewer, and associate editor C Dale provided helpful reviews.

9. REFERENCES


Meisel T. and Moser J. (2004a) Reference materials for geochemical PGE analysis: new analytical data for Ru, Rh, Pd, Os, Ir, Pt and Re by isotope dilution ICP-MS in 11 geological reference


**Figure Captions**

Fig. 1: Simplified geological map, showing locality of Finnish kimberlites analysed. Figure modified after Peltonen and Brügmann, (2006).

Fig. 2 (a) Phlogopite-rich group II kimberlite from dyke 16, Lentiira. Pool of late stage liquid in centre of photomicrograph is surrounded by phlogopite, tetraferriphlogopite, diopside, K-richterite. On right side of section are olivine pseudomorphs rimmed by perovskite. Plane polarised light. Modified after O’Brien and Tyni, 1995. (b) Thin section of hypabyssal kimberlite from Pipe 9 with abundant macro- and phenocrysts of olivine, as well as a fresh garnet lherzolite xenolith, containing olivine, orthopyroxene, red to purple pyrope, gray Mg-ilmenite, and bright green Cr-diopside. Plane polarized light. (c) Backscattered electron image of a Pipe 1 sample from Kaavi, Finland. Rounded olivine (bright green) is mostly altered in this sample, and all cases has a necklace of monticellite (red). Oxides are chromites (euhedral white) or late MUM spinel (skeletal small crystals, white to light yellow). Matrix is composed of serpentine (dk green), calcite (brown), kinoshitalite mica blades (yellow-orange) and euhedral apatite crystals (dark orange).

Fig. 3: Nd and Sr isotopic characteristics of Finnish kimberlites, compared to kimberlites and other alkaline rocks globally (modified after O’Brien, 2015). Data for Superior kimberlites are from
Fig. 4. PGE patterns of kimberlites, normalized to primitive mantle (normalization factors from Becker et al. (2005) and Day et al. 2016). Data is best modelled by mixing of 10-20% SCLM and 80-90% MARID. Note that MARID data are normalised to mantle detritus-free composition, as outlined in chapter 5.3.

Fig. 5: PGE patterns of Bushveld B1 sills (data from Barnes et al., 2010). Patterns of Premier kimberlites (ID and Ni-sulfide analyses) shown in grey. Normalization factors from Becker et al. (2005) and Day et al. (2016).

Fig. 6: Binary variation diagrams comparing composition of analysed kimberlites with Kaapvaal and Karelian mantle xenoliths. Also plotted are diamond-bearing (Renard) and barren kimberlites (Wemindji) from the Superior craton, Canada (Tappe et al., 2016). (A) Ru vs Ir, (B) MgO vs Ir. Most Karelian and Kaapvaal kimberlites plot near tielines between probable kimberlite melt and average Karelian SCLM (from Maier et al., 2012), suggesting that the main contaminant is garnet peridotite. In contrast, Superior kimberlites are more MgO rich, possibly suggesting more dunitic SCLM contaminants. Estimate of kimberlite melt is based on Price et al. (2000).

Table 1: Whole rock data for Karelian and Kaapvaal kimberlites

<table>
<thead>
<tr>
<th>Sample</th>
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<th>16-002</th>
<th>9729</th>
<th>9729</th>
<th>9729</th>
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<td>margin</td>
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<td>centre</td>
</tr>
<tr>
<td></td>
<td>Kaapvaal</td>
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<td>centre</td>
<td>centre</td>
<td>centre</td>
<td>centre</td>
<td>centre</td>
<td>centre</td>
<td>centre</td>
<td>centre</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

26.95  23.83  5621  5611  5596  5603  5599  5604
| Group | SiO₂ (wt%) | Zr (ppm) | Sr | Co | Cr | Ni | V | Cu | Sc | Rb | Y | Nb | Ba | La | Ce | Pr | Nd | Sm | Eu | Gd | Tb | Dy | Ho | Er | Tm | Yb | Lu | Hf | Ta |
|-------|------------|---------|----|----|----|----|----|----|----|----|----|----|----|----|----|----|----|----|----|----|----|----|----|----|----|----|----|----|
| pipe/dyke | 23.44 | 217 | 1183 | na | 750 | 471 | 243 | 38 | 28 | 52 | 20 | 217 | 968 | 240 | 445 | na | na | na | na | na | na | 217 | na | na | na |
| II | II | II | I | I | I | I | I | I | I | I | I | I | I | I | I | I | I | I | I | I | I | I | I | I | I | I | I |
| d 16 | d 16 | p 10 | p 14 | p 2 | p 5 | p 3 | p 9 | Premier | Premier |
| 42.95 | 36.07 | 35.28 | 34.96 | 32.08 | 35.99 | 29.69 | 39.43 | 39.81 | 37.72 |
| 1.99 | 2.41 | 2.39 | 2.84 | 2.34 | 1.31 | 2.29 | 1.95 | 1.96 | 2.00 |
| 3.14 | 4.01 | 2.30 | 5.02 | 3.48 | 5.43 | 2.74 | 4.88 | 3.62 | 2.11 | 2.46 |
| 0.28 | 0.14 | 0.15 | 0.19 | 0.19 | 0.26 | 0.17 | 0.35 | 0.15 | 0.14 | 0.15 |
| 15.02 | 21.68 | 29.47 | 24.19 | 29.96 | 22.58 | 28.39 | 20.89 | 26.95 | 27.31 | 27.06 |
| 22.48 | 5.44 | 4.87 | 8.30 | 3.78 | 10.09 | 7.27 | 12.59 | 6.19 | 7.16 | 8.24 |
| dl | 0.40 | dl | 0.21 | 0.14 | 0.10 | 0.14 | 0.11 | 1.05 | 0.06 | 0.01 |
| 0.76 | 5.73 | 2.31 | 1.57 | 2.27 | 1.95 | 0.31 | 0.63 | 2.44 | 1.04 | 1.07 |
| 0.87 | 0.12 | 0.39 | 0.58 | 0.34 | 0.60 | 0.27 | 0.69 | 0.34 | 0.33 | 0.42 |
| H₂O⁺ | na | na | na | na | 10.14 | 9.68 | 8.96 | 10.50 | 10.50 | 9.36 | na | na |
| LOI | na | na | na | na | na | na | na | na | 8.96 | 9.43 | na | na |
| C | na | na | na | 0.74 | 0.47 | 1.01 | 1.27 | 1.90 | 0.10 | na | na |
| Total | 75.95 | 93.64 | 87.76 | 98.20 | 99.60 | 97.19 | 97.83 | 95.81 | 100.90 | 98.24 | 98.19 |
Table 2: Absolute amount of total procedural blanks (in picograms) for HPA-ID procedure (n=12)

<table>
<thead>
<tr>
<th></th>
<th>Ru</th>
<th>Pd</th>
<th>Re</th>
<th>Os</th>
<th>Ir</th>
<th>Pt</th>
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<td>0.11</td>
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<tr>
<td>Lod (X + 3s)</td>
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<td>0.42</td>
<td>1.91</td>
<td>0.55</td>
<td>9.32</td>
</tr>
</tbody>
</table>

1 HPA+ID+ICP-MS
2 FA+ICP-MS
dl= value below detection limit; na=not analysed

Table 3: Comparison of Results Obtained for Reference Materials by Isotope Dilution Analyses at UQAC with Certificate Values

<table>
<thead>
<tr>
<th>OKUM</th>
<th>LK-NIP</th>
<th>LLD</th>
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<tr>
<td></td>
<td>UQAC</td>
<td>Geolabs</td>
</tr>
<tr>
<td></td>
<td>This run n=2</td>
<td>Certificate</td>
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<tr>
<td>Ru ppb</td>
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</tr>
<tr>
<td>Element</td>
<td>Pd ppb</td>
<td>Re ppb</td>
</tr>
<tr>
<td>---------</td>
<td>--------</td>
<td>--------</td>
</tr>
<tr>
<td></td>
<td>11.29</td>
<td>0.57</td>
</tr>
<tr>
<td></td>
<td>0.07</td>
<td>0.12</td>
</tr>
<tr>
<td></td>
<td>11.70</td>
<td>*0.566</td>
</tr>
<tr>
<td></td>
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<td>0.04</td>
</tr>
<tr>
<td></td>
<td>17.03</td>
<td>0.79</td>
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<tr>
<td></td>
<td>17.96</td>
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<tr>
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<td>0.15</td>
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<tr>
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<td>0.003</td>
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</table>

OKUM = Abtibi komatiite; LK-NIP = Niping Diabase both supplied by Geolabs, Ontario; n.d. not determined; *for OKUM no certificate values are available for Os and Re so values from Savard et al. (2010) are presented. +For LK-NIP no standard deviations were given on the certificate so in addition to certificate values working values and standard deviations obtained at UQAC by Ni-FA-ICP-MS are presented for comparison; # Lower limit of detection for HPA-ID-ICP-MS analysis Savard et al. (2010)
Table 1: Whole rock data for Karelian and Kaapvaal kimberlites

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<td>p 14</td>
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Zr (ppm) 217 319 149 103 63.7 94.4 52.8 96.6 69.1 80
Sr 1183 467 681 776 457 890 477 889 484 441
Co na 80.8 85.7 62.8 78.2 59.9 80.5 52.7 67.8 88
Cr 750 1238 905 1146 1334 1171 1810 1011 1339 1385
Ni 471 1166 1419 840 1027 599 1266 487 986 1305
V 243 65.2 78.8 118 155 173 114 170 107 99
Cu 38 58 60 105 119 119 80 128 77 33
Sc 28 14.4 8.5 17 16.8 20.5 15.1 19.3 11.2 9.96
Rb 52 187 98 99.3 86.2 100 22.9 68.3 106 92
Y 20 10.7 13.1 13.4 9.15 13.8 7.99 14.81 10
Nb 217 176 168 176 177 202 98.8 219 126 95.1
Ba 968 1254 1702 1971 1254 1523 627 1344 1164 1029
La 240 150 188 140 100 141 65 142 71.9 43
Ce 445 278 324 228 171 237 109 239 119 79
Pr na 29.1 33 22.2 17.3 23.5 11.1 23.6 12.5 8.2
Nd na 89.6 105 69.9 53.4 76.1 36.3 77.9 38.5 31
Sm na 9.92 12.7 9 6.66 9.45 4.54 9.72 4.81 4.48
Eu na 2.29 2.83 2.03 1.6 2.16 1.17 2.18 1.04 1.24
Gd na 7.26 8.4 6.37 4.95 6.37 3.48 6.31 3.79 3.23
Tb na 0.79 0.94 0.75 0.54 0.74 0.42 0.79 0.45 0.37
Dy na 2.57 3.42 2.87 2.01 2.89 1.72 3.01 1.76 1.69
Ho na 0.43 0.5 0.46 0.34 0.47 0.3 0.48 0.3 0.26
Er na 0.86 0.97 1.15 0.72 1.15 0.72 1.3 0.77 0.63
Tm na 0.09 0.11 0.13 0.09 0.17 0.09 0.16 0.09 0.074
Yb na 0.47 0.69 0.8 0.58 0.92 0.58 0.9 0.67 0.43
Table 2: Absolute amount of total procedural blanks (in picograms) for HPA-ID procedure (n=12)

<table>
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<tr>
<th></th>
<th>Ru</th>
<th>Pd</th>
<th>Re</th>
<th>Os</th>
<th>Ir</th>
<th>Pt</th>
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<tr>
<td>Average</td>
<td>0.24</td>
<td>0.66</td>
<td>0.11</td>
<td>0.49</td>
<td>0.11</td>
<td>3.16</td>
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<tr>
<td>Stdev</td>
<td>0.3</td>
<td>0.51</td>
<td>0.1</td>
<td>0.47</td>
<td>0.15</td>
<td>2.05</td>
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<tr>
<td>Lod (X + 3s)</td>
<td>1.14</td>
<td>2.2</td>
<td>0.42</td>
<td>1.91</td>
<td>0.55</td>
<td>9.32</td>
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</table>
Table 3: Comparison of Results Obtained for Reference Materials by Isotope Dilution Analyses at UQAC Certificate Values

<table>
<thead>
<tr>
<th></th>
<th>OKUM</th>
<th>Geolabs</th>
<th>LK-NIP</th>
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<tr>
<td></td>
<td>UQAC</td>
<td>Geolabs</td>
<td>UQAC</td>
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<tr>
<td></td>
<td>This run</td>
<td>Certificate</td>
<td>working value</td>
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<td></td>
<td>n=2</td>
<td>stdev</td>
<td>+ value</td>
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<tr>
<td>Ru ppb</td>
<td>4.72</td>
<td>0.15</td>
<td>4.25</td>
</tr>
<tr>
<td>Pd ppb</td>
<td>11.29</td>
<td>0.07</td>
<td>11.70</td>
</tr>
<tr>
<td>Re ppb</td>
<td>0.57</td>
<td>0.12</td>
<td>*0.566</td>
</tr>
<tr>
<td>Os ppb</td>
<td>1.20</td>
<td>0.18</td>
<td>*0.79</td>
</tr>
<tr>
<td>Ir ppb</td>
<td>1.11</td>
<td>0.23</td>
<td>0.99</td>
</tr>
<tr>
<td>Pt ppb</td>
<td>11.33</td>
<td>0.42</td>
<td>11.00</td>
</tr>
</tbody>
</table>

OKUM = Abtibi komatiite; LK-NIP = Niping Diabase both supplied by Geolabs, Ontario; n.d. not determined; *for OKUM no certificate values are available for Os and Re so values from Savard et al. (2010) are presented. +For LK-NIP no standard deviations were given on the certificate so in addition to certificate value working values and standard deviations obtained at UQAC by Ni-FA-ICP-MS are presented for comparison; # Lower limit of detection for HPA-ID-ICP-MS analysis Savard et al. (2010)
Figure 2
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Figure 4

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Electronic Annex 1

Click here to download Electronic Annex: Electronic Appendix 1 ICP-MS data of PGE in kimberlites.xlsx
Electronic Annex 2

Click here to download Electronic Annex: Electronic Appendix 2 Modelling table.xls