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# Discrimination of diamond resource and non-resource domains in the Victor North pyroclastic kimberlite, Canada ☆

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### Abstract

The Jurassic Victor kimberlite complex comprises several steep-sided pipes, including the Victor North pyroclastic kimberlite (VNPK) pipe which encompasses one of the main diamond resources in the mine plan. The VNPK pipe also contains some low-grade domains that appear to be non-economic. Distinguishing between the high- and low-grade units within this pipe is complicated by the gradational nature of the boundary.

In this paper, detailed petrography and mineral chemistry analyses are used to establish reliable discriminators for the high- and low-grade kimberlite units. The high- and low-grade units are both spinel-bearing carbonate group-I kimberlites; dolomite is more abundant than calcite in all units. Based on textural relationships and composition, it is inferred that the carbonate is primary. Oxygen fugacity estimates on groundmass intergrowths of rutile and ilmenite indicate similar oxidation states for the magmas that produced the high- and low-grade kimberlite deposits. However, chromite core compositions within zoned groundmass spinels are markedly different between the high- and low-grade kimberlite units. In addition, compositions of groundmass ilmenite and dolomite show the same pattern as core compositions of chromite: the groundmass minerals in the low-grade unit have lower magnesiumnumbers than corresponding phases in the high-grade unit. The chemical compositions of groundmass chromite and ilmenite suggest crystallisation with olivine, and thus record the oxygen fugacity associated with the early crystallisation history of the kimberlite magma.

These observations suggest that the high- and low-grade units were formed from two distinct magma pulses that had a different origin in the asthenosphere and/or experienced different crystallisation histories. The similar oxygen fugacity estimates for the high- and low-grade kimberlite magmas indicate that differential resorption of diamond was not responsible for the grade difference. Therefore, it is most likely that the individual magma pulses entrained contrasting amounts of diamond during their ascent towards the surface, causing the observed grade differences in the VNPK. © 2008 Elsevier B.V. All rights reserved.

Keywords: kimberlite; diamond deposits; petrography; pyroclastics; fugacity; spinel group

## 1. Introduction

Kimberlite magma originates as a sub-cratonic, asthenospherederived melt. During its ascent towards the surface, the magma commonly incorporates abundant mantle xenocrysts, lesser mantle xenoliths and possible trace amounts of diamond. In order to improve our understanding of kimberlite formation, it is important to study the processes that produce kimberlite rocks, including melt generation, magma ascent, emplacement and eruption. This study of the Victor kimberlite focuses on understanding the deeper mantle processes involved in melt generation, sampling of xenocrysts ( $\pm$ diamond), and the evolution of magma during transport. Here, we show that deep mantle processes have a direct influence on the petrological and economic characteristics of diamondiferous kimberlite deposits.

The Victor kimberlite is located in the James Bay Lowlands, Northern Ontario, Canada and is scheduled for open pit mining by De Beers Canada in 2008. Victor is the largest kimberlite in the Middle-Late Jurassic Attawapiskat kimberlite province (Kong et al., 1999), and forms a complex of several pipes. The steep-sided

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Fig. 1. Macrodiamond content distribution (2001–2002 bulk sample programme) defining high-, moderate- and low-grade zones of the Victor North pyroclastic kimberlite pipe. Characteristic juvenile pyroclasts for each grade zone are shown schematically. Note the higher abundance of olivine phenocrysts in the low-grade juvenile pyroclasts and mixed pyroclasts (i.e. derived from both high- and low-grade kimberlite magma pulses) in the moderate-grade zone. Vertical axis is 300 m long. Vertical lines indicate the location of large diameter bulk sampling drill holes. The locations of drill cores V-00-117c and V-00-125c are also shown.

 $(\sim 70^{\circ})$  pipes occur in an Ordovician to Silurian sedimentary succession, unconformably overlying Precambrian granitoid basement. About 10–30 m of unconsolidated Quaternary overburden is currently present on top of the kimberlite.

The primary focus of this paper is the Victor North pyroclastic kimberlite (VNPK) pipe (Webb et al., 2004). The VNPK pipe comprises one of the main resources in the mine plan, but also includes some low-grade domains that, at the present time, appear to be a non-resource (Fig. 1). Mineral resource is used here to refer to "a concentration or occurrence of material of intrinsic economic interest in or on the Earth's

crust in such form and quantity that there are reasonable prospects for eventual economic extraction". Grade refers to "the relative quantity or the percentage of ore-mineral content in an orebody" (Glossary of Geology, 2005). One of the ongoing issues in terms of understanding the geology, delineating the ore and controlling grade, is distinguishing between the resource and non-resource domains of the VNPK pipe. The main rock type in the VNPK pipe is a massive, moderately sorted, loosely packed, crystal lapilli-tuff. The rock is dominated by discrete or free olivine crystals (crystals without selvages of crystallised melt) and minor juvenile pyroclasts (crystals rimmed by selvages of crystallised melt). The kimberlite contains the two different generations of olivine characteristic of most kimberlites: olivine macrocrysts (anhedral, commonly 0.5-10 mm in size, mantle-derived grains) and primary olivine phenocrysts (euhedral and <0.5 mm in size).

A close correlation between kimberlite petrography and macrodiamond content in the VNPK pipe was established by Webb et al. (2004). The macrodiamond content distribution (Fig. 1) in the VNPK pipe is consistent with an emplacement model of nested craters resulting from two discrete eruptions (Webb et al., 2004). First, a magma with a low diamond content and high abundance of olivine phenocrysts ( $\sim 28\%$ ) erupted explosively, producing juvenile pyroclasts with a high olivine phenocryst abundance (Figs. 1 and 2A). A second explosive eruption of kimberlite magma, having a high diamond content and lower abundance of olivine phenocrysts ( $\sim 8\%$ ), from the same pipe formed a nested crater. The juvenile pyroclasts of this second kimberlite pulse have a distinctly lower proportion of olivine phenocrysts (Figs. 1 and 2B). Intermediate diamond contents are found at the interface between the high- and low-grade deposits. The mixed interface is also characterised by the presence of both (e.g. olivine phenocryst-rich vs. olivine phenocryst-poor) juvenile pyroclast types (Fig. 1). An average sample grade of 23 carats per hundred tonnes has been established for the Victor kimberlite. With an approximate value of 419 US\$/ct, Victor represents one of the highest value/carat primary diamond deposits in the world (grade and value figures courtesy of De Beers Canada Inc., 2007).

Prior to this study, the high- and low-grade kimberlite deposits were differentiated texturally in thin section using an optical microscope (Webb et al., 2004). This study finds that



Fig. 2. Photomicrographs of characteristic juvenile pyroclasts in the low-grade (A) and high-grade (B) kimberlite units. Low-grade juvenile pyroclasts have a higher abundance of olivine phenocrysts. OM: olivine macrocryst, OP: olivine phenocryst (altered in low-grade juvenile pyroclast), D: dolomite lath.

electron microprobe analyses of groundmass minerals can distinguish the low- and high-grade kimberlite units. In addition, it contributes to an improved understanding of the petrology of the Victor North kimberlite pipe, as well as the genesis and early evolution of the kimberlite magma that formed the Victor diamond deposit.

# 2. Analytical methods

Detailed core logging and sampling of two representative vertical drill cores (V-00-117c, V-00-125c, see Fig. 1 for location) in the VNPK pipe were undertaken. The selected drill cores both intersect all three of the contrasting grade zones, from bottom to top: low, moderate and high. The diamond grades in these two drill holes are inferred from bulk sampling of two reverse circulation large diameter (61 cm) drill holes (each located  $\sim 10$  m from the respective core holes) and detailed petrography on juvenile pyroclasts in samples from the aforementioned core holes (Webb et al., 2004; this study).

Representative polished core samples and thin sections of drill core V-00-117c and V-00-125c were studied petrographically. The goal of this study was to search for differences between the high- and low-grade kimberlite units in addition to those identified by Webb et al. (2004). The petrographic work focused on juvenile pyroclasts, comprising olivine grains rimmed by selvages of crystallised melt, which are viewed here as representative of the kimberlite magmas.

Detailed optical transmitted and reflective light microscope, scanning electron microscope (SEM) and electron microprobe (EMP) analyses were performed at the Department of Earth and Ocean Sciences, UBC. The SEM and EMP analyses were performed on four samples from drill core V-00-117c, two from the high-grade unit and two from the low-grade unit. A total number of 25 juvenile pyroclasts were analysed. Based on the presence of two petrographically distinct kimberlite units which are internally massive and show minor internal variation in componentry and petrography (Webb et al., 2004; this study), it is suggested that these samples form a representative subset for further analytical work. All groundmass minerals were studied under the SEM (Philips XL30) to determine the presence and nature of zoning. Subsequent quantitative EMP analyses (Tables 1 and 2) were performed using the fully automated CAMECA SX-50 instrument, operating in the wavelength-dispersion mode. Data reduction was done using the 'PAP'  $\phi(\rho Z)$  method (Pouchou and Pichoir, 1985). All oxides (spinel-group minerals, ilmenite and rutile) were analysed with an excitation voltage of 15 kV, beam current of 20 nA, beam diameter of 2 µm, peak counting time of 20 seconds (s) and background counting time of 10 s. Analyses of vanadium were corrected for titanium interference. The proportion of ferric  $(Fe^{3+})$  vs. ferrous  $(Fe^{2+})$  iron was calculated using stoichiometric and charge balance equations (Droop, 1987). Micas were analysed at 15 kV, 10 nA, with a beam diameter of 5-10 µm, peak counting time of 20 s (40 s for F, Cl) and background counting time of 10 s (20 s for F, Cl). Carbonates were analysed at an accelerating voltage of 15 kV, beam current of 10 nA, with a beam diameter of  $5-9 \mu m$ , a uniform peak counting time of 20 s and background counting time of 10 s. Apatite was analysed at an accelerating voltage of 15 kV, beam current of 10 nA, with a beam diameter of  $2-5 \mu m$ , and peak counting time of 10 s (14 s for La, Ce; 20 s for Cl; 50 s for F) and background counting time of 5 s (7 s for La, Ce; 10 s for Cl, 25 s for F). Loss of fluorine in apatite due to electron beam induced diffusion was tested in four larger  $(\geq 30 \,\mu\text{m})$  crystals, using methodology outlined in Stormer et al. (1993) and Coulson et al. (2001). Five analyses were conducted at

Table 1

Rep	resentative	electron	microprob	e (EMP	) analv	ses of	groundmass s	spinel.	ilmenite a	nd rutile	in high-	and lo	ow-grade	kimberlite	unit	iuvenile	pyroclas	ts
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Depth <sup>a</sup>	125.69	125.69	070.25	125.69	211.24	156.90	125.69	125.69	156.90	156.90	125.69	156.90
Grade	High	High	High	High	Low	Low	High	High	Low	Low	High	Low
Mineral <sup>b</sup>	Chr centre	Chr edge	Mag-Usp	Mag	Chr centre	Chr edge	Ilm	Ilm (+Rt)	Ilm	Ilm (+Rt)	Rt (+Ilm)	Rt (+Ilm)
MgO	10.28	11.69	11.26	2.56	10.24	10.87	19.55	18.01	17.11	16.80	1.63	3.22
$Al_2O_3$	1.97	7.34	4.22	0.21	2.57	7.43	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
SiO <sub>2</sub>	n.d.	0.06	n.d.	0.44	n.d.	0.07	0.11	0.04	0.07	n.d.	0.05	0.07
CaO	0.14	0.11	0.30	0.36	0.30	0.31	0.25	0.14	0.20	0.28	0.19	0.40
TiO <sub>2</sub>	5.29	2.24	12.07	0.54	5.59	4.18	55.22	55.09	53.90	54.96	91.39	86.20
$V_2O_5$	n.d.	0.17	n.d.	n.d.	n.d.	0.07	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
$Cr_2O_3$	50.69	54.45	0.79	0.71	47.27	47.57	2.14	3.88	3.53	3.87	3.22	3.76
MnO	n.d.	n.d.	0.39	0.29	0.26	0.27	0.40	0.31	0.41	0.36	n.d.	0.27
Fe <sub>2</sub> O <sub>3</sub> <sup>c</sup>	9.78	6.60	45.32	67.44	11.32	8.55	8.07	5.42	6.50	5.64		
FeO	20.45	17.44	24.55	27.45	20.42	19.23	14.27	16.92	17.35	18.82	1.43	4.65
NiO	0.21	n.d.	0.24	n.d.	0.13	0.17	n.d.	n.d.	0.10	n.d.	n.d.	n.d.
Nb <sub>2</sub> O <sub>5</sub>	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.13	n.d.	n.d.	n.d.	1.61	1.35
Ta <sub>2</sub> O <sub>5</sub>	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.08	n.d.
Total wt.%	98.81	100.10	99.14	100.00	98.10	98.72	100.14	99.81	99.17	100.73	99.60	99.92
$Mg/Mg + Fe^{t}$	0.385	0.471	0.235	0.049	0.374	0.419	0.618	0.596	0.568	0.556		
Cr/Cr+Al	0.945	0.833	0.111	0.692	0.925	0.811						

n.d.: not detected (below minimum detection limit).

<sup>a</sup> All samples are from V-00-117c, number represents depth/sample number in meters.

<sup>b</sup> Mineral abbreviations after Kretz (1983), (+..) denotes intergrowth with mineral in brackets.

<sup>c</sup> Ferric (Fe<sup>3+</sup>) vs. ferrous (Fe<sup>2+</sup>) iron calculated using stoichiometric and charge balance equations (Droop, 1987).

Table 2 Representative electron microprobe analyses of groundmass carbonate, apatite and phlogopite in high- and low-grade kimberlite unit juvenile pyroclasts

Depth <sup>a</sup>	125.69	156.90	156.90	156.90	070.25	211.24	125.69	156.90	125.69	125.69	156.90	125.69
Grade	High	Low	Low	Low	High	Low	High	Low	High	High	Low	High
Mineral <sup>b</sup>	Cal – lath	Cal – lath	Cal - repl.ol.	Cal - vein	Dol – lath	Dol – lath	Dol – base	Dol – base	Ap	Ap <sup>c</sup>	Ap	Phl – g.m.
Na <sub>2</sub> O	_	_	_	_	_	_	_	_	0.23	0.15	0.62	0.20
MgO	0.13	0.84	1.89	1.48	20.30	20.33	20.66	20.66	0.55	0.35	0.73	25.71
$Al_2O_3$	_	_	_	_	_	_	_	_	_	_	_	11.21
SiO <sub>2</sub>	_	_	_	_	_	_	_	_	_	_	_	42.51
$P_2O_5$	_	_	_	_	_	_	_	_	40.49	41.28	39.97	-
K <sub>2</sub> O	_	_	_	_	_	_	_	_	_	_	_	8.89
CaO	56.91	55.31	54.05	54.66	32.56	30.60	31.87	31.25	54.66	54.85	54.59	0.40
TiO <sub>2</sub>	_	_	_	_	_	_	_	_	_	_	_	0.66
Cr <sub>2</sub> O <sub>3</sub>	_	_	_	_	_	_	_	_	_	_	_	0.30
MnO	n.d.	0.02	n.d.	n.d.	0.32	0.28	0.33	0.21	n.d.	n.d.	n.d.	n.d.
FeO	0.16	0.15	0.23	0.09	0.18	1.28	0.51	0.56	0.22	0.19	0.20	3.91
SrO	0.23	0.44	n.d.	n.d.	0.29	0.57	0.14	0.12	0.27	n.d.	0.14	_
BaO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	_	_	_	n.d.
La <sub>2</sub> O <sub>3</sub>	_	_	_	_	_	_	_	_	n.d.	n.d.	n.d.	-
Ce <sub>2</sub> O <sub>3</sub>	_	_	_	_	_	_	_	_	n.d.	n.d.	n.d.	_
F	_	_	_	_	_	_	_	_	1.14	1.09	1.12	0.10
Cl	_	_	_	_	_	_	_	_	0.14	0.11	0.26	0.14
H <sub>2</sub> O <sup>d</sup>									1.20	1.23	1.19	4.12
O=F									-0.48	-0.46	-0.47	-0.04
O=Cl									-0.03	-0.02	-0.06	-0.03
$CO_2^{e}$	45.01	44.61	44.65	44.57	48.15	47.42	48.15	47.61				
Total wt.%	102.44	101.37	100.82	100.80	101.80	100.48	101.66	100.41	98.39	98.77	98.29	98.08

n.d.: not detected (below minimum detection limit); "-": not analysed.

<sup>a</sup> All samples are from V-00-117c, number represents depth/sample number in meters.

<sup>b</sup> Mineral abbreviations after Kretz (1983). Repl.ol: replacing olivine, g.m.: groundmass.

<sup>c</sup> Fluorine calculated by linear extrapolation from five successive measurements on same spot (for methodology, see Stormer et al., 1993; Coulson et al., 2001).

<sup>d</sup> H<sub>2</sub>O in apatite and phlogopite determined by stoichiometry, assuming 1 (OH, F, Cl) for apatite and 2 (OH, F, Cl) for phlogopite.

<sup>e</sup> CO<sub>2</sub> in carbonate calculated by stoichiometry.

the same point within each of these crystals. The initial value for fluorine was obtained by linear extrapolation back to the initial concentration. In these four cases the first measurement overestimated the fluorine concentration by only 1-11%. This overestimation has not been corrected for in other apatite analyses.

## 3. Petrography and mineralogy

#### 3.1. Macroscopic features

The kimberlite intersected in drill core is generally massive and shows only slight variations in grain size. Less common are 1-2 m thick, graded beds of kimberlite lapilli-tuff. The kimberlite has low concentrations of country rock lithic clasts (average  $\sim 7$  vol.%). The only macroscopic observation that might serve to distinguish the high- and low-grade zones is a  $\sim 10$  m thick interval with abundant country rock lithic clasts  $(\sim 17 \text{ vol.}\%)$  that occurs on the outer margin of the high-grade zone. Otherwise, there are only minor macroscopic differences between the high- and low-grade kimberlite units, which attest to the indistinct gradational nature of this grade boundary. These features include (Webb et al., 2004; this study): (1) slightly higher abundance of country rock basement clasts (granitic and dolerite); (2) greater degree of alteration of olivines; and (3) more abundant Cr-diopside macrocrysts in the low-grade unit. None of these features allows for a confident macroscopic division between the high- and low-grade kimberlite units.

#### 3.2. Microscopic features

The juvenile pyroclasts in the high- and low-grade kimberlite units are remarkably similar in terms of their primary mineralogy. The only reliable microscale discriminator between these units is the abundance of olivine phenocrysts (Fig. 2). The low-grade kimberlite unit contains juvenile pyroclasts with a higher olivine phenocryst abundance (~28 vol.% vs. ~8 vol.% in the high-grade kimberlite unit; Webb et al., 2004). The groundmass of juvenile pyroclasts of both kimberlite units comprises, in order of abundance, euhedral calcite or dolomite laths, euhedral–subhedral spinel, minor small euhedral apatite and other less common opaque minerals (ilmenite, rutile; Figs. 2 and 3). In addition, rare laths of phlogopite were observed. Groundmass phases are set in a cryptocrystalline matrix of dolomite±serpentine (base). The juvenile pyroclasts are non-vesicular to poorly vesicular (typical abundance 0 to 4 vol.%) and have vesicle diameters of 50–150  $\mu$ m.

#### 3.3. Groundmass mineral composition

Olivine phenocrysts from the low-grade kimberlite unit are all altered and, thus, their original compositions cannot be measured. The chromites, however, are very resistant to alteration, and are pervasive in both the high- and low-grade kimberlite units. The chromite occurs as cores within grains of groundmass spinel (Fig. 3). The spinel is usually euhedral–subhedral,  $20-50 \ \mu m$  in diameter and strongly zoned from chromite core to magnetite rim



Fig. 3. Backscatter electron (BSE) images of groundmass phases within juvenile pyroclasts. (A) Typical low-grade kimberlite unit spinel with zoned chromite core (c: Al-rich centre, e: Al-poor edge) and magnetite (m) atoll-rim. (B–C) Typical high-grade kimberlite unit spinels with zoned chromite core and enveloping magnetite rim. Note irregular shape of chromite centre (c). (C–D) False colour BSE images of chromite, ilmenite (i) and rutile (r) intergrowths. Note that only the Al-poor chromite edge (e) is in direct contact with ilmenite. (E) Juvenile pyroclast selvage (S) around altered olivine macrocryst (OM) with abundant calcite laths (C) and one altered olivine phenocryst (OP). ICM: inter-clast matrix. (F) Dolomite laths (D) and apatite crystal (A) in the groundmass of a juvenile pyroclast.

(Fig. 4A; Table 1). The chromite cores in the VNPK are themselves zoned and this zoning trend is observed in almost all larger ( $\geq 20 \ \mu m$ ) cores. They are zoned from an early Al-rich chromite centre to an Al-poor and Cr-rich chromite edge (Figs. 3 and 4B; Table 1). The shape of the chromite centre is commonly anhedral (n=11/15), whereas the edge of the chromite usually has a more euhedral–subhedral outline.

There are clear differences between the spinels occurring in the high- and low-grade kimberlite units. The first difference is the development of magnetite rims (Fig. 3A–C). The low-grade kimberlite unit spinel grains (n=10) commonly have an obvious atoll-textured rim (80-90% of spinel grains, Fig. 3A). In addition, the magnetite in the atoll-rim is never in contact with the chromite core (100% of spinel grains, Fig. 3A). The highgrade kimberlite unit spinel grains (n=17) typically have no atoll-textured rim (71% of spinel grains, Fig. 3B, C). The magnetite rim is usually in contact with the chromite core (59% of spinel grains, Fig. 3C). An additional 23% of the spinel grains have a magnetite rim in contact with the chromite core on one side of the crystal, but no magnetite rim in contact with the chromite core on the other side (Fig. 3B). The remaining 18% of the high-grade unit spinel grains have no magnetite rim at all. The major difference between spinels from the high- and lowgrade kimberlite is the distinct contrast in the chromite core compositions (Fig. 4B). Both the centres and the edges of the chromite cores of spinel crystals in the low-grade kimberlite unit have lower Mg-numbers (Mg/Mg+Fe<sup>t</sup>) at a similar Crnumber (Cr/Cr+Al), relative to those in the high-grade unit. A slight difference in Ti-content of the chromite cores is also observed between the high- and low-grade units in the VNPK, although not as distinct as the difference in Cr# (Fig. 4B). The low-grade unit chromites generally have a higher Ti-content than those in the high-grade unit at similar Mg#.

Minor opaque phases in the groundmass include ilmenite and rutile. The range (average  $\pm 1$  SD) in mole fractions for the ilmenite analyses (n=33) are 0.57-0.65 geikielite (MgTiO<sub>3</sub>), 0.29-0.37 ilmenite (Fe<sup>2+</sup>TiO<sub>3</sub>), 0.04-0.06 haematite ( $Fe^{3+}_{2}O_{3}$ ), and <0.01 pyrophanite (MnTiO<sub>3</sub>); see Table 1. The groundmass ilmenites plot in the same field as groundmass ilmenites and ilmenite inclusions in olivine from the Wesselton kimberlite (Shee, 1984; Fig. 5A). Rutile analyses (Table 1; n=22) yield an average  $X^{Ti}$  (Ti p.f.u./ $\Sigma$ all cations) of  $0.88 \pm 0.04$  (1 SD). Analyses with  $X^{\hat{T}\hat{t}}$  below 0.82 were excluded. Oxide intergrowths (Fig. 3C, D) are commonly observed in the juvenile pyroclast groundmass. Touching Al-poor chromites and Mg-rich ilmenite and rutile, as well as chromites being partly surrounded by ilmenite and rutile indicate that ilmenite and rutile crystallised during or after crystallisation of the Al-poor chromite. The mineral composition of the groundmass ilmenites shows a similar difference between the high- and low-grade kimberlite units as the chromites (Fig. 5B): the low-grade kimberlite unit ilmenites have a slightly lower Mg# than their high-grade counterparts. The difference in Mg# is less pronounced than that for the chromite centres.

Euhedral calcite or dolomite laths are common and occur within a cryptocrystalline base of dolomite and minor serpentine (Fig. 3E, F). The carbonate phases are stoichiometrically close to pure end-member compositions of calcite and dolomite (Table 2). In addition, all calcite laths have a moderate Sr content (Fig. 6A) with respect to calcite replacing olivine and calcite veins. There is a difference in composition of the groundmass dolomite laths between the high- and low-grade kimberlite units. The low-grade kimberlite unit dolomite laths have slightly lower Mg# (higher Fe content, see Fig. 6B), similar to observations of oxide groundmass minerals.



Fig. 4. Groundmass spinel compositions. (A) Chromite to magnetite zoning trend in spinels from the VNPK kimberlite, shown in a  $Mg/Mg+Fe^{2+}$  vs.  $Fe^{3+}/Cr+Al+Fe^{3+}$  plot. All chromites in (B) plot in the lower part of the graph. Typical zoning trends in kimberlites are shown by the black arrow, and the fields marked M86 (data from Mitchell, 1986) and S84 (Shee, 1984). The small field marked FC04 (Fedortchouk and Canil, 2004) represents chromite inclusions in olivine phenocrysts from the Lac de Gras area. (B)  $Mg/Mg+Fe^{t}$  vs. Cr/Cr+Al plot with Cr–Al zoning trend from chromite centres to edges of the cores. Fields marked FC04 represent chromite inclusions in olivine phenocrysts from the Torrie (lower left), Leslie, Grizzly and Aaron (upper right) kimberlites in the Lac de Gras area. Field marked S84 represents groundmass chromites (crosses) and a chromite included in olivine (star), each representing a different unit in the Wesselton kimberlite. Error bar indicates 2 SD analytical precision of EMP analysis.

Groundmass apatites are fluorine-bearing hydroxylapatites (Table 2), whereas the (OH, F, Cl) cation site in phlogopite is dominated by hydroxide with trace amounts of fluorine and chlorine (Table 2).

# 4. Discussion

#### 4.1. Early crystallisation sequence

The first minerals to crystallise from kimberlite magma are commonly olivine phenocrysts and chromites (e.g. Mitchell, 1986; Fedortchouk and Canil, 2004). Within the VNPK pipe, chromite in the groundmass of juvenile pyroclasts is found as cores within multiphase spinel crystals. The rim enveloping the chromite core is commonly composed of magnetite. The mineral zoning pattern from chromite to magnetite, and the presence of abundant atoll-textures on these spinels is typical of Group I kimberlites (Mitchell, 1986; Fig. 4A). The magnetite rims are commonly interpreted as products of late-stage crystallisation (Mitchell, 1986). Similar zoning trends and texturally comparable vermiform rims are found in chromites in basalts (Roeder et al., 2001, 2006). We have compared the chemical composition of the earlier formed chromite cores of groundmass spinel crystals in the VNPK with chromites from other kimberlites, including euhedral chromite inclusions in olivine phenocrysts and euhedral chromites in the groundmass (data from Shee, 1984; Fedortchouk and Canil, 2004). The chemical composition of the VNPK groundmass chromites is comparable to euhedral chromites included in olivines from the Lac de Gras kimberlites (Fedortchouk and Canil, 2004; similar range in Cr#, slightly lower Mg#), and very similar to both the euhedral chromite inclusion in olivine and groundmass chromites from the Wesselton kimberlite (Shee, 1984; Fig. 4B). During contemporaneous olivine and chromite crystallisation, some chromites are included in olivine phenocrysts during its growth, while others remain in the groundmass. The comparison of chromite compositions in kimberlite (Fig. 4B) suggests that the groundmass chromite grains in the VNPK crystallised at the same time as olivine and have not reequilibrated with the evolving melt. The chromite cores of groundmass spinel crystals in the VNPK show an atypical Cr-Al trend of Al-depletion from the centres towards the edges of the cores (Figs. 3A-C and 4B). This trend could be explained



Fig. 5. Groundmass ilmenite compositions. (A) Cr<sub>2</sub>O<sub>3</sub> vs. MgO wt.% plot comparing VNPK groundmass ilmenites with macrocryst, inclusion (in olivine phenocryst) and groundmass fields established by Shee (1984) for the Wesselton kimberlite. Ilmenite inclusions in olivine phenocrysts (diamonds) and ilmenite in the groundmass (crosses) from the Lac de Gras kimberlites (Fedortchouk and Canil, 2004) are shown for reference. The 18.0 wt.% MgO line represents the suggested upper limit for ilmenites crystallising together with olivine. (B) Mg-number histogram of groundmass ilmenite. 2 SD indicates precision.



Fig. 6. Groundmass carbonate compositions. (A) Sr (wt.%) vs. Ca (wt.%) plot for groundmass calcite laths (upper fields) and calcite in veins and olivine replacements (lower field). (B) Fe (wt.%) vs. Ca (wt.%) plot for groundmass dolomite that occurs in laths and in the interstitial base. DL: upper and lower detection limit; error bar indicates 2 SD precision.

by co-crystallisation of olivine and chromite (e.g. Cr–Al trend, Barnes and Roeder, 2001).

Common oxide intergrowths (Fig. 3C, D) give important information about the timing of crystallisation of the various oxide mineral phases. Touching Al-poor chromites and Mg-rich ilmenite and rutile, as well as chromites being partly surrounded by ilmenite and rutile, indicate that the latter two phases crystallised during or after crystallisation of the Al-poor chromite. A comparison of groundmass ilmenite compositions in the VNPK with groundmass ilmenite and ilmenite included in olivine phenocrysts in other kimberlites (Shee, 1984; Fedortchouk and Canil, 2004) was made. More than two thirds of the groundmass ilmenite analyses from the VNPK plot in the field for ilmenite inclusions in olivine phenocrysts (Fig. 5A) established by Shee (1984) for the Wesselton kimberlite. In addition, almost all of the analyses plot on the low MgO-side of the groundmass ilmenite field. Based on this comparison, it is suggested that all ilmenites with  $\leq 18.0$  wt.% MgO are the result of early crystallisation together with olivine, irrespective of their position in the groundmass or as inclusion in olivine.

## 4.2. Geothermometry and oxygen barometry

Co-crystallised spinel–olivine and rutile–ilmenite were used to estimate the temperature and/or oxygen fugacity ( $fO_2$ ) of the

magma at the time of crystallisation. The oxygen fugacity, or the oxidation state, of kimberlite magma is an important parameter that controls diamond resorption (Fedortchouk et al., 2005; Kozai and Arima, 2005).

The following geothermometers and oxygen barometers were used: (1) Spinel–olivine geothermometer and oxygen barometer (Ballhaus et al., 1991): Based on groundmass chromite compositions from VNPK and a range in worldwide olivine phenocryst composition. Unknown variables:  $P, X^{Fo}$  (forsterite) in olivine; and (2) Rutile–ilmenite oxygen barometer (Zhao et al., 1999): Based on intergrown groundmass rutile and ilmenite in VNPK. Unknown variables: P, T.

#### 4.2.1. Spinel-olivine

Groundmass chromite compositions in this study suggest crystallisation together with olivine phenocrysts (see Section 4.1.). As no olivine phenocrysts were analysed in this study, a range in composition from worldwide kimberlite occurrences was used (Fo<sub>87</sub>-Fo<sub>93</sub>; Mitchell, 1995). The spinel-olivine Fe/ Mg exchange geothermometer (Ballhaus et al., 1991) and oxygen barometer based on the olivine-orthopyroxene-spinel equilibrium (Ballhaus et al., 1991) were used. The geothermometer and oxygen barometer have been applied successfully to numerous mantle rocks and mantle-derived melts. The oxygen barometer requires the presence of orthopyroxene, and for orthopyroxene-undersaturated rocks such as kimberlites the barometer gives maximum oxygen fugacity values (Ballhaus et al., 1991). A correction can be made if the silica activity of the rock is known. However, due to the absence of suitable silicabearing mineral phases in the groundmass that could limit the silica activity below the forsterite-enstatite buffer (Ballhaus et al., 1991; Barker, 2001; Fedortchouk and Canil, 2004), a correction was not made.

The temperature and oxygen fugacity were calculated using 46 analyses of chromites (23 from the high-grade kimberlite unit, 23 from low-grade), assuming a pressure of 10 kbar (1 GPa). Pressure has a fairly small effect on temperature and oxygen fugacity calculations (up to 1.5% decrease in temperature, up to 0.02 log unit decrease in  $fO_2^{NNO}$  from 10 kbar to 1 bar). The dominant influence on T and  $fO_2$  is the olivine composition  $(X^{Fo})$ , and the average result (n=46) varies from 919 °C, 2.0 log units above NNO for Fo<sub>93</sub> to 1209 °C, 0.8 log unit above NNO for Fo<sub>87</sub>, both at 10 kbar (Fig. 7A; where NNO is the nickel-nickel oxide buffer, after Ballhaus et al., 1991). This range in temperature correlates well with estimates from chromite inclusions in olivine by Fedortchouk and Canil (2004) using the same methodology. Results from the latter study are: temperatures of 1030-1170 °C ± 50 °C (mean 1080 °C), and oxygen fugacity values of 0.4-0.9 log unit above NNO (when uncorrected for silica activity). The oxygen fugacity results from this study show only slight differences between the highand low-grade kimberlite units (low-grade kimberlite unit is up to 0.3 log unit more oxidised).

## 4.2.2. Rutile-ilmenite

Rutile-ilmenite intergrowths (Fig. 3D) are common in the groundmass of juvenile pyroclasts of both kimberlite units, and



Fig. 7. Oxygen fugacity estimates for the VNPK. (A) Range in oxygen fugacity values relative to the NNO buffer at different pressures and temperatures (rutile–ilmenite; Zhao et al., 1999) and variable olivine phenocryst compositions (olivine–spinel; Ballhaus et al., 1991; at 10 kbar). Average values at different *T*, *P*, *X* conditions (heavy black lines)  $\pm 2$  SD are shown. Field of FC04 (Fedortchouk and Canil, 2004) shows the range in oxygen fugacity results for chromite, ilmenite and rutile inclusions in olivine for the Lac de Gras kimberlites (1080 °C, 10 kbar; not corrected for silica activity). (B) Oxygen fugacity values relative to the NNO buffer (1080 °C, 10 kbar) for the high- and low-grade kimberlite units, based on groundmass rutile–ilmenite intergrowths and the oxygen barometer of Zhao et al. (1999).

allow for estimation of the oxygen fugacity of the magma. The oxygen barometer of Zhao et al. (1999) was applied to touching mineral pairs and analyses with good totals (Tables 2 and 3). The barometer has been used successfully for rutile-ilmenite assemblages in mantle xenoliths incorporated in kimberlite (Zhao et al., 1999) and for rutile and ilmenite inclusions in olivine phenocrysts in kimberlites (Fedortchouk and Canil, 2004). The oxygen barometer (Zhao et al., 1999, Eq. 4) depends on the temperature, pressure, and the activities of FeTiO<sub>2</sub> in ilmenite, Fe<sub>2</sub>O<sub>3</sub> in ilmenite and TiO<sub>2</sub> in rutile. The oxygen barometer of Zhao et al. (1999) uses the solid solution models by Ghiorso (1990) and Ghiorso and Sack (1991) to calculate the activities for the ilmenite end members (an online version for calculating the activities of rhombohedral oxides is available at http://melts.ofm-research.org/CalcForms/index.html). In the oxygen barometer calculations, the activity of TiO<sub>2</sub> in rutile is equal to  $X^{Ti}$ . The rutile–ilmenite oxygen barometer is calibrated on ilmenites with a range in  $X^{ilm}$  of 0.42–0.67, significantly more iron-rich than ilmenites in kimberlites (our study, average X<sup>ilm</sup>=0.29-0.37; Fedortchouk and Canil (2004), X<sup>ilm</sup> ranges from 0.34-0.36). Despite the difference in composition, a comparison of the rutile-ilmenite (Zhao et al., 1999) and the olivine-spinel oxygen barometer (Ballhaus et al., 1991) by Fedortchouk and Canil (2004) yielded similar results, when the latter results are uncorrected for silica activity.

The ilmenites used in the oxygen fugacity calculations have an average MgO content of 17.6 wt.% (range: 15.3–19.8; n=17), and 11 out of these 17 ilmenites have MgO  $\leq$  18.0 wt.%, suggesting co-crystallisation with olivine (see Section 4.1.). Based on these data it is proposed that all rutile–ilmenite intergrowths crystallised during or slightly after olivine crystallisation. The redox state calculations in this study were conducted for spinel–olivine crystallisation temperatures, as estimated for the Lac de Gras kimberlites, Canada (1080 °C at 10 kbar, Fedortchouk and Canil, 2004). These conditions are comparable to the range in temperatures derived from olivine–spinel geothermometry for the VNPK rocks (919–1209 °C, see Section 4.2.1.). To check the sensitivity of the oxygen barometer to pressure and temperature, oxygen fugacities were also calculated at intermediate pressure–temperature conditions (~800 °C, 1 kbar), and low pressure–temperature conditions prevalent during groundmass crystallisation at the surface (~600 °C, 1 bar; Kopylova et al., 2007).

Similar oxygen fugacities were obtained for the high- and low-grade kimberlite units (Fig. 7B), irrespective of P, T conditions (Table 3). At 1080 °C and 10 kbar the results are about 3.2-3.8 log units above the diamond/graphite-CO<sub>2</sub> (D/GCO) buffer of Frost and Wood (1997), and 0.9-0.3 log unit below the nickel-nickel oxide (NNO) buffer of Ballhaus et al. (1991). These results are  $\sim 1.3$  log units lower (i.e. less oxidised) than estimates by Fedortchouk and Canil (2004) for Lac de Gras kimberlites, when compared to the oxygen fugacities not corrected for silica activity (Fig. 7A). Similar oxygen fugacities in both the high- and low-grade kimberlite units suggest that diamond resorption was not responsible for the observed grade difference. In addition, the oxygen fugacity at the time of rutile and ilmenite crystallisation could, in principle, allow diamond resorption. In light of the above, it is interesting to look at experimental data of diamond dissolution. Experiments by Kozai and Arima (2005) suggest that the diamond oxidation rate decreases with increasing carbonate concentration in the melt. However, recent studies by Fedortchouk et al. (2007) imply that diamond oxidation requires the presence of a free fluid phase. The latter authors suggest that the decreasing oxidation rate at increasing carbonate concentration in the melt obtained in the Kozai and Arima (2005) study is caused by an increase of water solubility in the more carbonate-rich melt. This would decrease the amount of fluid in the system, and decrease the diamond oxidation rate. Thus, the carbonate-rich (possibly fluid phasepoor) kimberlite magma at Victor (see Section 4.3) might have prevented much of the diamond dissolution in both the highand low-grade kimberlite magmas.

# 4.3. Nature of carbonates

Late stage groundmass crystallisation included precipitation of various carbonate phases, and most likely, apatite. The calcite laths contain moderate amounts of Sr (Fig. 6A) relative to the concentrations found in the secondary calcite that forms veins or replaces olivine. The concentration of Sr found in calcite laths is consistent with magnatic crystallisation (e.g. Mitchell, 1986; Armstrong et al., 2004). The preservation of variable Mg numbers of groundmass dolomite laths between the two kimberlite units (Fig. 6B) presents additional evidence that these carbonates are

Table 3 Oxygen fugacity results from high- and low-grade kimberlite units, based on rutile–ilmenite intergrowths in juvenile pyroclast groundmass

Rutile			Ilmenite <sup>a</sup>							At 1080 °C, 10 kbar						At 600 °C, 1 bar				
	EMP # <sup>c</sup>	X <sup>Ti</sup>	EMP # <sup>c</sup>	X <sup>IIm</sup>	X <sup>Gei</sup>	$X^{Prn}$	X <sup>Hem</sup>	Fe <sup>3+</sup> /Fe <sup>t</sup>	Fe <sub>2</sub> O <sub>3</sub>	FeTiO <sub>3</sub>	log fO <sub>2</sub> [kbar] <sup>d</sup>	$\Delta log fO_2 NNO^e$	$\Delta log fO_2$ D/GCO <sup>e</sup>	Fe <sub>2</sub> O <sub>3</sub>	FeTiO <sub>3</sub>	$\log fO_2 [kbar]^d$	$\Delta log fO_2 \ NNO^e$	$\Delta log fO_2$ D/GCO <sup>e</sup>		
High-grade	1-125 69-20	0.937	1-12569-21	0.327	0.620	0.006	0.047	0.223	0.014	0.441	-8.26	-0.43	3.73	0.056	0.468	-17.49	0.68	2.43		
	1-12569-1	0.905	1-12569-3	0.279	0.673	0.007	0.041	0.227	0.011	0.389	-8.32	-0.48	3.67	0.045	0.408	-17.50	0.67	2.43		
	3-07025-23	0.916	3-07025-b_01	0.334	0.611	0.013	0.041	0.197	0.011	0.451	-8.55	-0.72	3.44	0.044	0.481	-17.78	0.39	2.14		
	3-07025-18	0.914	3-07025-20	0.334	0.616	0.006	0.044	0.208	0.012	0.450	-8.44	-0.60	3.55	0.050	0.478	-17.67	0.50	2.26		
	3-07025-17	0.920	3-07025-20	0.334	0.616	0.006	0.044	0.208	0.012	0.450	-8.43	-0.59	3.56	0.050	0.478	-17.66	0.51	2.27		
	3-07025-6	0.833	3-07025-7	0.284	0.659	0.009	0.048	0.251	0.014	0.393	-8.24	-0.40	3.75	0.058	0.413	-17.45	0.72	2.48		
	3-12569-3	0.912	3-12569-5	0.280	0.676	0.007	0.037	0.208	0.009	0.392	-8.47	-0.64	3.52	0.038	0.411	-17.65	0.53	2.28		
	3-12569-4	0.876	3-12569-5	0.280	0.676	0.007	0.037	0.208	0.009	0.392	-8.54	-0.71	3.45	0.038	0.411	-17.72	0.45	2.21		
	3-12569-13	0.848	3-12569-14	0.399	0.555	0.006	0.040	0.166	0.010	0.518	-8.97	-1.14	3.02	0.042	0.554	-18.21	-0.04	1.71		
Low-grade	3-15690-5	0.838	1-15690-7	0.305	0.642	0.008	0.044	0.225	0.012	0.418	-8.47	-0.63	3.53	0.050	0.442	-17.68	0.49	2.25		
	1-15690-8	0.828	1-15690-7	0.305	0.642	0.008	0.044	0.225	0.012	0.418	-8.49	-0.65	3.51	0.050	0.442	-17.70	0.47	2.23		
	1-15690-8	0.828	1-15690-9	0.338	0.613	0.008	0.040	0.191	0.011	0.456	-8.78	-0.95	3.21	0.043	0.486	-18.00	0.17	1.92		
	1-21124-3	0.861	1-21124-4	0.377	0.559	0.007	0.057	0.231	0.019	0.487	-8.31	-0.47	3.68	0.073	0.521	-17.58	0.59	2.34		
	1-21124-3	0.861	1-21124-5	0.361	0.572	0.007	0.060	0.251	0.021	0.470	-8.17	-0.33	3.82	0.080	0.502	-17.44	0.73	2.48		
	3-15690-7	0.824	1-15690-12	0.277	0.662	0.008	0.053	0.277	0.017	0.383	-8.07	-0.24	3.92	0.068	0.402	-17.28	0.89	2.65		
	3-15690-7	0.824	1-15690-13	0.305	0.639	0.008	0.048	0.239	0.014	0.416	-8.36	-0.52	3.63	0.058	0.440	-17.58	0.59	2.35		
	3-15690-9	0.820	3-15690-10	0.397	0.545	0.008	0.050	0.201	0.015	0.510	-8.67	-0.83	3.32	0.059	0.546	-17.93	0.24	1.99		
	3-15690-9	0.820	3-15690-11	0.364	0.580	0.007	0.049	0.212	0.015	0.478	-8.58	-0.74	3.41	0.058	0.511	-17.83	0.34	2.09		
	3-15690-13	0.832	3-15690-14	0.406	0.545	0.006	0.043	0.176	0.012	0.523	-8.91	-1.08	3.08	0.047	0.560	-18.16	0.01	1.76		
	3-15690-13	0.832	3-15690-15	0.385	0.553	0.007	0.055	0.221	0.018	0.496	-8.45	-0.62	3.54	0.069	0.530	-17.72	0.45	2.20		

<sup>a</sup>Mineral abbreviations after Kretz (1983). Gei: geikielite, Prn: pyrophanite.

<sup>b</sup>Calculated using Ghiorso (1990) and Ghiorso and Sack (1991). An online version for calculating the activities of rhombohedral oxides is available at http://melts.ofm-research.org/CalcForms/index.html.

<sup>c</sup>All samples are from V-00-117c, first digit represents EMP day #, next five digits represent depth/sample number (12569 is 125.69 m), last digits are EMP spot #.

<sup>d</sup>Calculated using oxygen barometer by Zhao et al. (1999).

<sup>e</sup>Buffer equations after Ballhaus et al. (1991) for NNO; Frost and Wood (1997) for D/GCO.

primary. There is no compositional evidence (e.g. moderate Sr content) that the dolomite base is a result of direct crystallisation from the magma. However, in all cases where olivine is fresh it is considered unlikely that the dolomite has been altered while adjacent carbonate laths and olivine have been preserved. The abundant carbonate in the juvenile pyroclasts at Victor indicates that the late-stage melt was dominantly carbonatic in composition. The fact that co-crystallisation of carbonate and apatite is commonly observed in carbonaties (Le Bas, 1981; Deans and Roberts, 1984; Stoppa et al., 2000; Rosatelli et al., 2003) provides further evidence for a primary origin of these carbonates.

#### 4.4. Magma genesis, evolution and entrainment of diamonds

There are two important observations that constrain the origins of the high- and low-grade kimberlite units in the VNPK pipe. The first is related to the chemistry of various groundmass minerals in the juvenile pyroclasts. Despite the fact that not all mineral phases in the high- vs. low-grade kimberlite units show statistically different populations, there is a consistent shift of all minerals in the high-grade kimberlite unit towards more Mgrich compositions (Figs. 4-6). A difference in Mg# of chromite between different phases of kimberlite and different kimberlite localities is very common (e.g. Shee, 1984; Mitchell, 1986; Fedortchouk and Canil, 2004; see Fig. 4B). Therefore, it seems likely that the high- and low-grade kimberlite units in the VNPK are derived from two distinct kimberlite magma pulses with a different genesis and/or evolution. These results are in good agreement with the study by Webb et al. (2004), which showed that the high- and low-grade kimberlite units are derived from separate magma pulses, based on different olivine phenocryst abundances in the high- and low-grade kimberlite juvenile pyroclasts (Figs. 1 and 2).

The second observation relates to the diamond content and oxygen fugacity estimates for the high- and low-grade kimberlite units. Similar oxygen fugacity estimates for the high- ( $\Delta \log fO_2^{NNO} - 0.6 \pm 0.4$  (2 SD) for Rt-Ilm at 1080 °C, 10 kbar) and low-grade kimberlite magma pulse ( $\Delta \log fO_2^{NNO} - 0.6 \pm 0.5$ ) were obtained. This negates the possibility that a difference in oxygen fugacity between the two magmas caused preferential resorption of diamond in the earlier, diamond-poor, kimberlite. Thus, it is most likely that the difference in diamond content between the two magmas is caused by differential mantle sampling. The later kimberlite magma, forming the present-day high-grade deposit, picked up more diamonds during its rise through the mantle.

Based on the above considerations, we are proposing four processes that could explain the consistent difference in Mg# between the various mineral phases that crystallised in the highvs. low-grade kimberlite magma pulses: (1) Slight differences in composition of the mantle protoliths where the kimberlite melts are formed, resulting in a different bulk chemistry of the two kimberlite magmas. (2) A difference in the source degree of melting, also resulting in a different bulk chemistry of the magmas. (3) Variable amount of olivine phenocryst crystallisation in the magmas during ascent. Olivine crystallisation will deplete the remaining magma in Mg relative to Fe (decrease in Mg#), resulting in less Mg-rich oxides being formed by subsequent crystallisation. This scenario accords well with the different olivine phenocryst abundances in the high- and lowgrade juvenile pyroclasts. That is, more olivine phenocryst crystallisation in the low-grade kimberlite magma would result in relatively lower Mg# for the melt and subsequently crystallising oxide minerals. (4.) The temperatures of crystallisation (e.g. Ballhaus et al., 1991), whereby a higher crystallisation temperature causes a higher Mg# in chromite.

# 5. Conclusions

- 1. Our results show that the groundmass chromites, ilmenites and dolomite laths from the low-grade kimberlite unit all have lower Mg-numbers than groundmass minerals in the high-grade kimberlite unit. Also, the low-grade kimberlite unit is characterised by a higher abundance of olivine phenocrysts.
- 2. This study confirms the conclusion by Webb et al. (2004) that the high- and low-grade kimberlite units in the VNPK pipe formed from two distinct magma pulses. In addition, it shows that these magmas had a different origin in the asthenosphere and/or experienced different crystallisation histories causing a consistent shift in Mg-numbers of most mineral phases.
- 3. The similar oxygen fugacity values for the two magma pulses indicate that differential diamond dissolution cannot be responsible for the grade difference between the high- and low-grade kimberlite units. Thus, the two magmas entrained contrasting amounts of diamond during their ascent towards the surface.
- 4. This validates the geological and emplacement models, which in turn predict the diamond distribution within the pipe (Fig. 1).

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