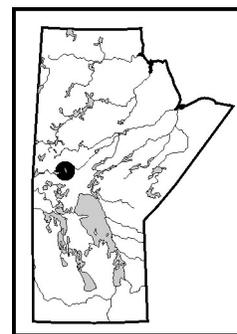


GS-5 **Igneous carbonate-rich rocks from the south Wekusko Lake area, Manitoba (parts of NTS 63K9 and 63J12): a kimberlite-carbonatite dilemma**

by A. Chakhmouradian¹, C.O. Böhm and J.K. Greville¹



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Summary

The petrography, mineralogy and geochemistry of two macroscopically distinct drillcore samples of ‘kimberlitic’ rock from the south Wekusko Lake area (central Manitoba) were examined in detail. The two studied samples share a number of textural, modal and geochemical characteristics which strongly suggests that they were derived from similar sources and underwent broadly similar evolutionary histories. Both examined samples show an inequigranular, uniform to segregation texture arising from the presence of abundant macrocrysts (partly or completely chloritized phlogopite, Mg-Al-Cr-rich spinel and Mg-Cr-rich ilmenite) and coarse-grained segregations in a fine-grained dolomitic groundmass. With the exception of Mg-Al-Cr-rich spinel and Mg-Cr-rich ilmenite macrocrysts, the modal and chemical makeup of the examined samples is inconsistent with the identification of these rocks as either group-I or group-II kimberlite. On the basis of their modal, major- and trace-element composition, the Wekusko Lake samples are tentatively identified as magnesiocarbonatite. The majority of phlogopite macrocrysts are cognate with the hostrock and exhibit compositional variation typical of carbonatitic micas. The spinel and ilmenite macrocrysts are compositionally indistinguishable from their counterparts in kimberlites; the latter are routinely used as diamond-indicator minerals in exploration. This similarity (and the concomitant enrichment of the examined samples in Cr, Ni and Co) implies that the Wekusko Lake rocks were derived from a mantle source similar to those that produce kimberlites. The available mineralogical and geochemical evidence indicates that this source was chemically modified (metasomatized) prior to the melting. Further studies of the Wekusko Lake magnesiocarbonatite and its constituent minerals by isotope-analytical techniques will be required to constrain the source and evolutionary history of these rocks.

Introduction

The forty-plus years of intermittent diamond-exploration activity in Manitoba have led to the recognition of several promising targets across central and

northern Manitoba, where follow-up work is presently being carried out by a number of senior and junior companies. Nevertheless, only one of these targets, at Wekusko Lake in central Manitoba, has so far been confirmed as ‘kimberlite’ (see details below). The lack of reliable petrographic, mineralogical and geochemical data on these rocks prompted a re-examination of the material extracted during the initial exploration phase at Wekusko Lake. The present report is a summary of the preliminary findings.

Geological overview of the south Wekusko Lake area

The south Wekusko Lake area is located approximately 20 km southeast of Snow Lake and is accessible by Highways 392 and 39. Geologically, it forms part of the eastern Flin Flon Belt of the Paleoproterozoic Trans-Hudson Orogen (NATMAP Shield Margin Working Group, 1998; Bailes and Galley, 1999; Figure GS-5-1). Most of the supracrustal components of the Flin Flon Belt range in age from 1.91 to 1.88 Ga (Lucas et al., 1996) and represent the products of ocean floor volcanism and arc magmatism (Syme and Bailes, 1993; Stern et al., 1995a, b). Variations in lithology and geochemical associations demonstrate that these sequences are disparate parts of various ocean-floor and arc assemblages, tectonically juxtaposed during intra-oceanic accretion at ca. 1.88 to 1.87 Ga, and subsequent continental collision at ca. 1.84 to 1.79 Ga (Stern et al., 1995b; Lucas et al., 1996).

The tectonic evolution of the Manitoba-Saskatchewan segment of the Trans-Hudson Orogen is believed to have involved diachronous collision of the Archean Rae-Hearne, Superior and Sask cratons, which was largely complete by 1.8 Ga (Hajnal et al., 2005). The collision between the Paleoproterozoic Flin Flon–Glennie Complex² (FFGC) and the Sask craton began around, or slightly earlier than, 1.83 Ga, and resulted in southwest-vergent thrusting of the FFGC over the Archean–Siderian rocks (ca. 3.10–2.45 Ga) of the Sask plate along the Pelican thrust (or décollement) zone and intense shear-induced deformation, which continued to at least

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² The Paleoproterozoic Glennie Domain in west-central Saskatchewan is geologically similar to the Flin Flon Domain to the east; the two domains are commonly regarded as parts of the Flin Flon–Glennie Complex (Corrigan et al., 2005).

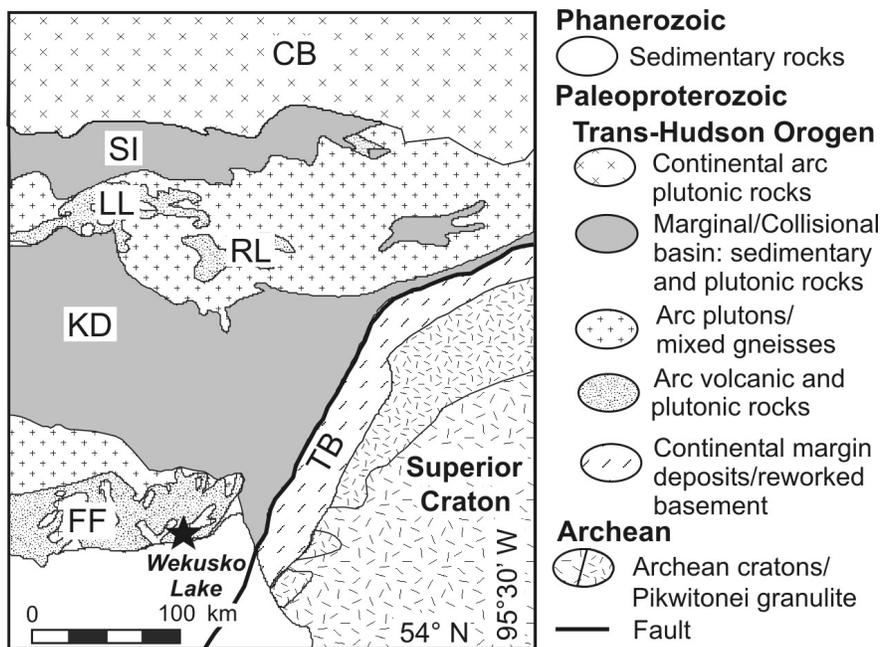


Figure GS-5-1: Simplified geology of the Wekusko Lake study area (outlined) within the Flin Flon–Snow Lake Belt of the Trans-Hudson Orogen (after Bailes and Galley, 1999). Abbreviations: CB, Chipewyan Batholith; FF, Flin Flon–Snow Lake Belt; KD, Kisseynew Domain; LL, Lynn Lake Belt; RL, Rusty Lake Belt; SI, Southern Indian Domain; TB, Thompson Belt.

1806 Ma (Ashton et al., 1999, 2005; Ansdell, 2005; Bickford et al., 2005). Post-collisional shortening ca. 1.80 to 1.77 Ga, manifested in F3 folds, is typically attributed to the continuing convergence of the Sask and Superior cratons at depth, (i.e., below the FFGC; Ashton et al., 2005). The relative extent of the Archean rocks underplating the FFGC is largely unknown; the recent seismic-reflection and radio-isotope studies indicate that the Sask substructure extends beneath at least 100 000 km² of the Trans-Hudson Orogen, including the FFGC (Bickford et al., 2005; Hajnal et al., 2005).

Bedrock exposures in the south Wekusko Lake area are dominantly metavolcanic and metasedimentary supracrustal rocks intruded by granitoid rocks and overlain by Ordovician dolomite as outlined and described in Gilbert and Bailes (2005; Figure GS-5-2). The supracrustal rocks in the study area are distinct tectonostratigraphic supracrustal components (south Wekusko Lake, Hayward Creek, Burntwood Group; Figure GS-5-2). They form part of a structurally complex collage of Paleoproterozoic volcanic juvenile arc, arc rift, ocean floor and oceanic island assemblages, and sedimentary and minor volcanic deposits. At south Wekusko Lake, the north- to north-northeast-trending Crowduck Bay Fault (Ansdell and Connors, 1993; Gilbert and Bailes, 2005) traverses the south Wekusko Lake ocean floor basalt sequence. A subparallel, possible splay of the fault marks the eastern boundary of a domain, several kilometres wide, of 1.85 to 1.84 Ga (David et al., 1996) Burntwood Group greywacke, siltstone and mudstone turbidite and derived paragneiss (Figure GS-5-2).

The South Wekusko Lake domain is a <4 km wide, east-facing ocean floor volcanic sequence that comprises pillowed, aphyric, tholeiitic basalt flows, related gabbro sills and mafic phyllonite (Gilbert and Bailes, 2005). The Hayward Creek domain (Gilbert, 1994) comprises juvenile arc volcanic rocks including basalt, mafic fragmental rocks, mixed felsic flows and fragmental rocks, and volcanoclastic greywacke and mafic tuff (Gilbert and Bailes, 2005). This lithologically diverse suite is geochemically akin to 1.89 Ga (David et al., 1996) volcanic rocks of the Snow Lake arc assemblage of Bailes and Galley (1999). The Burntwood Group sedimentary rocks at south Wekusko Lake include mainly greywacke, siltstone and mudstone turbidites that are fault-bound between the south Wekusko Lake and the Hayward Creek volcanic domains (Figure GS-5-2).

The 1.84 to 1.83 Ga southwest Wekusko Lake pluton forms an ovoid, mainly tonalite–quartz diorite body that structurally stitches together the Hayward Creek and Burntwood Group domains at the southwest end of Wekusko Lake (Figure GS-5-2; Gilbert and Bailes, 2005). The youngest intrusive rocks in this area include anorthositic gabbro and quartz gabbro (Gilbert and Bailes, 2005) and ‘kimberlitic dikes’, the latter being the focus of this report.

The above bedrock exposures are unconformably overlain by irregularly shaped outliers of flat-lying Ordovician dolomite up to several square kilometres in size (Figure GS-5-2; Gilbert and Bailes, 2005). Exposures of the Precambrian–Phanerozoic boundary are generally

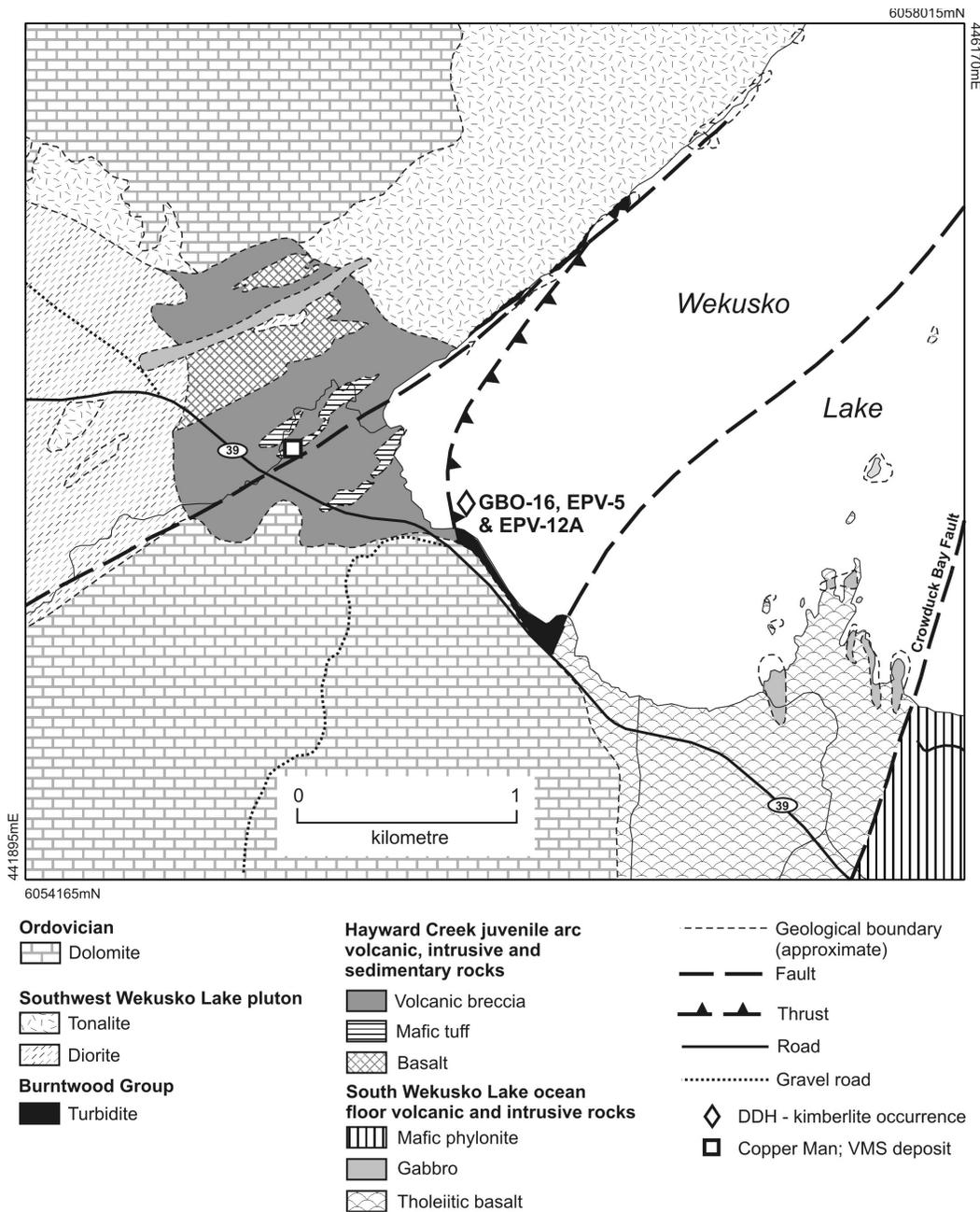


Figure GS-5-2: Simplified geology of the southwest Wekusko Lake area (after Gilbert and Bailes, 2005).

rare in Manitoba, but south Wekusko Lake represents an area where this unconformable boundary is accessible in many places.

Chronology of the Wekusko Lake ‘kimberlite’ core samples

Falconbridge Nickel Mines Ltd. first intersected a brecciated rock of uncertain petrographic affinity during their 1983 drilling program focused on the Copper Man VMS deposit (Golden Bounty option) in the southwest section of Wekusko Lake. The problematic rock was intersected in drillhole GBO-16 within Burntwood Group

greywacke, just offshore from the VMS deposit (Figure GS-5-2). Kimberlite is not reported in the original drill log; however, the interval in question (263–280 ft.) was described as “breccia, fine to medium grained, light grey matrix with up to 10% granule and pebble-sized fragments of dolomite, siltstone and mudstone. Some of the fragments are angular to sub-angular; others are subrounded to rounded. The matrix is soft and crumbly. Core is extremely broken. Contains occasional pyrope garnets” (Assessment File 70569, Manitoba Science, Technology, Energy and Mines, Winnipeg). This core was stored at Falconbridge’s Anderson Bay core storage facility in Snow Lake, and was retrieved by D. Prouse

(Manitoba Geological Survey) in the summer of 2003.

In the winters of 1992 to 1994, European Ventures Inc. (now Firestone Ventures Inc.) drilled a series of holes in the southwest Wekusko Lake area, following up on the reported garnet-bearing breccia intersected in the Falconbridge Nickel Mines drillhole GBO-16. Three of the European Ventures drillholes (EPV-5-93, EPV-12A-94 and EPV-17-94) intersected a ‘kimberlitic’ rock. The EPV-12A drillhole intersected a short section of ‘kimberlitic’ rock and showed weak conductivity, which was attributed to chloritic muddy gouge on both walls. The drillhole logs for EPV-12A and EPV-17 are attached to Assessment File 73676 (Manitoba Science, Technology, Energy and Mines, Winnipeg) and the core for EPV-12A was submitted to The Pas core storage facility. The drillhole log for EPV-5 was not attached. The log description at 106.5 to 107.3 m in drillhole EPV-12A mentions “kimberlite dike — breccia, fine to medium grained, brown, vuggy”. The 0.5 m wide dike intersection at 43.8 m in drillhole EPV-17 is described as “brown-pitted, slight red tinge — diopside crystals” and is identified as being “visually identical to the dike intersected by holes EPV-5 and 12A” (A.F. 73676).

A European Ventures press release of July 26, 1993, refers to a report by Gurney and Zweistra (1993) stating that “G10, G9 and possible DI (diamond inclusion) eclogitic garnets have been identified by microprobe analyses of concentrate from a 5.2 kg kimberlite sample intersected near Snow Lake” (European Ventures Inc., 1993). This was followed up by a brief note on The Diamond Page in The Northern Miner from March 28, 1994, reporting that “European Ventures (VSE) has intersected a narrow kimberlite dike on Wekusko property near Snow Lake, Man. A 5 kg sample containing indicator minerals has been sent for microdiamond analysis” (The Northern Miner, 1994).

The small remaining portions of the ‘kimberlitic’ interval of the original Falconbridge core (GBO-16) were recovered and studied by the Manitoba Geological Survey (2003). Based on thin sections, the small samples present were interpreted as carbonatite and were subsequently processed for U-Pb perovskite age dating at the University of Alberta Radiogenic Isotope Facility. Two mineral fractions were analyzed but likely contained rutile and/or chromite together with, or instead of, perovskite. Consequently, the U-Pb and Pb-Pb isotopic analyses are dominated by common lead (^{204}U) and therefore provided no clear age result (Böhm and Heaman, unpublished data, 2004).

In spring 2005, P. Holden (Gossan Resources Ltd.) provided R. Syme (Manitoba Geological Survey) and subsequently the authors of this report with additional

material from the Wekusko Lake drillcore. This material is believed to have come from the intersection of kimberlitic rock in the European Ventures drillhole EPV-12A-94 (A.F. 73676). The material studied in this work included two fragments of the EPV-12A-94 drillcore (20 g each) and several fragments of GBO-16 drillcore (ca. 60 g in total).

Petrography

The petrography of samples EPV-12A-94 and GBO-16 was studied in conventional and polished thin sections using polarizing optical microscopy, back-scattered electron (BSE) imaging and energy-dispersive x-ray spectrometry (EDS). The two samples differ in macroscopic colour (brownish grey and greyish green, respectively), but share a number of textural and modal characteristics, including

- inequigranular, slightly vesicular uniform to segregation texture consisting of macrocrysts³ (0.2–2.5 mm) set in a fine-grained carbonate matrix (Figure GS-5-3);
- the preponderance of teal green (emerald green in hand specimen) chlorite (\pm its precursor phlogopite) among macrocrysts, followed by chromian spinel⁴ and magnesian ilmenite;
- the bulk of the groundmass being composed of turbid carbonate material with indiscernible grain boundaries, which locally grades into larger (up to 300 μm in size) clear rhombohedral crystals lining vesicles and macrocryst boundaries;
- the presence of fragmented clastic material (predominantly quartz); and
- ubiquitous reaction textures (e.g., replacement of phlogopite by chlorite).

Textural evidence indicates that chlorite is not a primary mineral in these rocks, and that it developed at the expense of phlogopite macrocrysts (Figure GS-5-3c). Within a relatively small area, all gradations from unaltered phlogopite (yellowish orange, bright interference colours) to complete pseudomorphing of phlogopite by chlorite (teal green, anomalous interference colours) can be found (Figures GS-5-3a, d), suggesting a very localized nature of replacement reactions. Although fresh phlogopite was not observed in EPV-12A-94, the chlorite macrocrysts are identical in both samples and were probably formed through the same process. Sample GBO-16 contains a higher modal percentage and a greater variety of macrocrysts, including phlogopite, zircon and fluorapatite. Zircon is nearly as common as ilmenite in that sample, and is readily recognizable by its high

³ In this report, the term ‘macrocryst’ is preferred to such genesis-specific terms as microphenocryst, xenocryst, etc., because the potential source (or sources) of this material remains undetermined.

⁴ Consistent with the conventionally used kimberlite terminology, the name ‘spinel’ here refers to spinel-group minerals *sensu lato*.

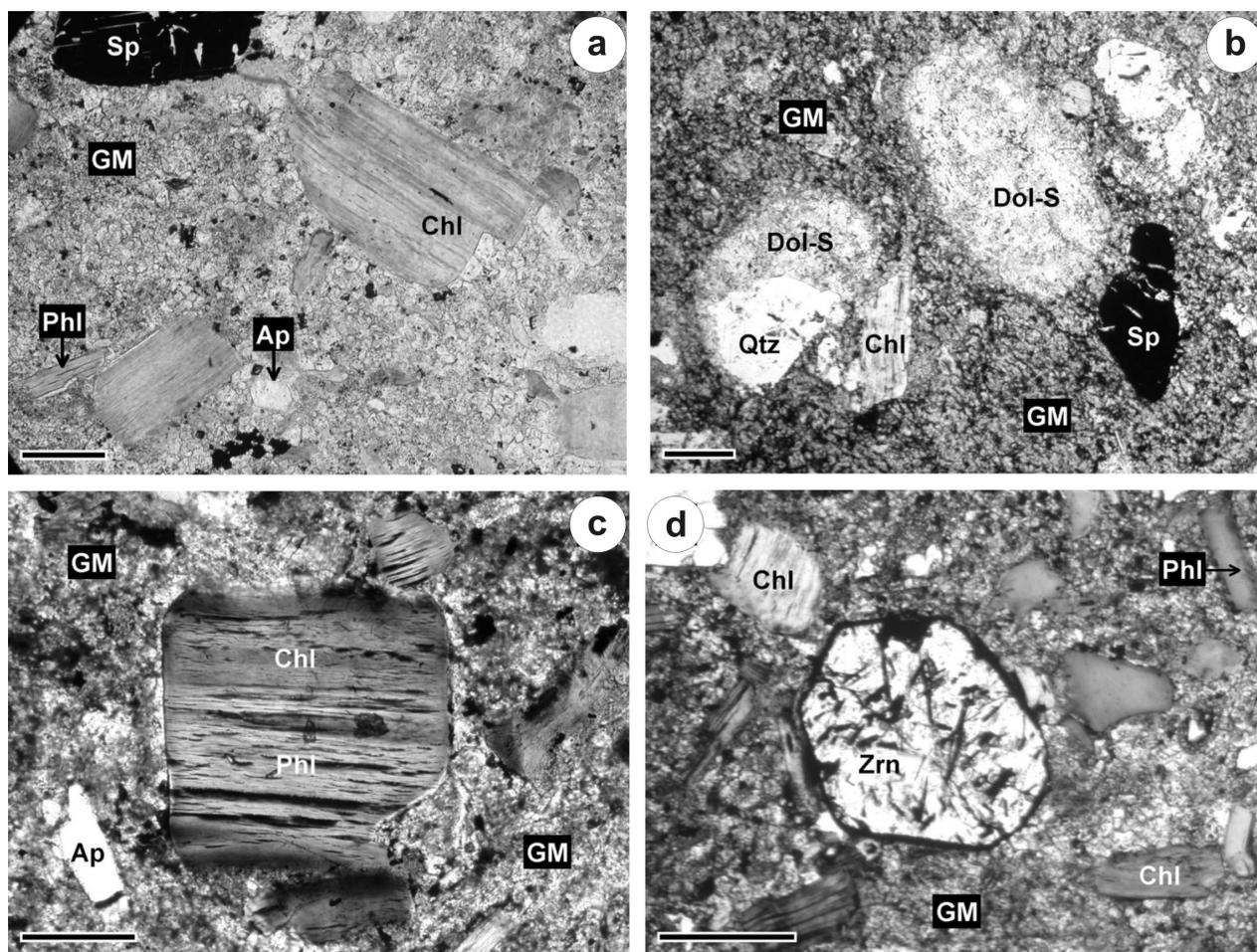


Figure GS-5-3: Textural characteristics of samples GBO-16 (a, c, d) and EPV-12A-94 (b), as seen in thin section in plane-polarized light. Scale bar is 200 μm in (a) and (c) and 500 μm in (b) and (d). Note the presence of chlorite (Chl), phlogopite (Phl), spinel (Sp), fluorapatite (Ap) and zircon (Zrn) macrocrysts in a predominantly dolomitic groundmass (GM). Sample EPV-12A-94 also features ovoid segregations of relatively coarser-grained dolomite (Dol-S) and quartz clasts (Qtz).

birefringence and a thin opaque rim (Figure GS-5-3d). In addition to the macrocrysts, sample EPV-12A-94 contains abundant ovoid segregations (0.5–2.5 mm across) of coarser-grained dolomite and angular fragments of quartz (Figure GS-5-3b) and Al-Ti-rich biotite. Dolomite segregations are also present in GBO-16, but these invariably host numerous crystals of (chloritized) phlogopite.

Mineral chemistry

The chemical compositions of all minerals were determined in polished thin sections by wavelength-dispersive x-ray spectrometry (WDS) using a fully automated Cameca SX100 electron microprobe. The bulk of both studied samples is made up of dolomite with variable proportions of an ankerite component in its composition (3–22 mol % $\text{CaFe}(\text{CO}_3)_2$). Variations in the Mg/Fe ratio result in intricate growth zoning detectable only in BSE images (Figure GS-5-4a). The highest

FeO content (8.3 wt. %) is observed in the core of large crystals from GBO-16, which also show appreciable levels of SrO (0.3–0.6 wt. %). The low-Fe dolomite compositions overlap in the two samples. The MnO content is systematically low (≤ 1.4 wt. %), and Ba is below the detection limit of WDS (500 ppm).

Phlogopite macrocrysts and crystals from segregations exhibit a wide compositional range ($\text{mg}\#^5 = 0.64\text{--}0.85$, $\text{Al}_2\text{O}_3 = 8.8\text{--}14.9$ wt. %, $\text{TiO}_2 = 0.1\text{--}1.9$ wt. %, $\text{Cr}_2\text{O}_3 \leq 0.06$ wt. %, $\text{BaO} \leq 0.5$ wt. %, $\text{MnO} = 0.08\text{--}0.22$ wt. %, $\text{F} = 0.6\text{--}1.4$ wt. %), indicating that they equilibrated at different P-T conditions or were derived from different sources. None of these igneous compositions, however, overlap with that of Al-Ti-rich biotite xenocrysts observed in sample EPV-12A-94 ($\text{mg}\# \approx 0.35$, $\text{Al}_2\text{O}_3 \approx 18.0$ wt. %, $\text{TiO}_2 \approx 3.4$ wt. %, $\text{F} \approx 0.4$ wt. %). Similar to quartz, the biotite xenocrysts clearly represent crustal material, but it remains to be determined whether this

⁵ $\text{mg}\# = \text{Mg}/(\text{Mg}+\text{Fe}_2)$

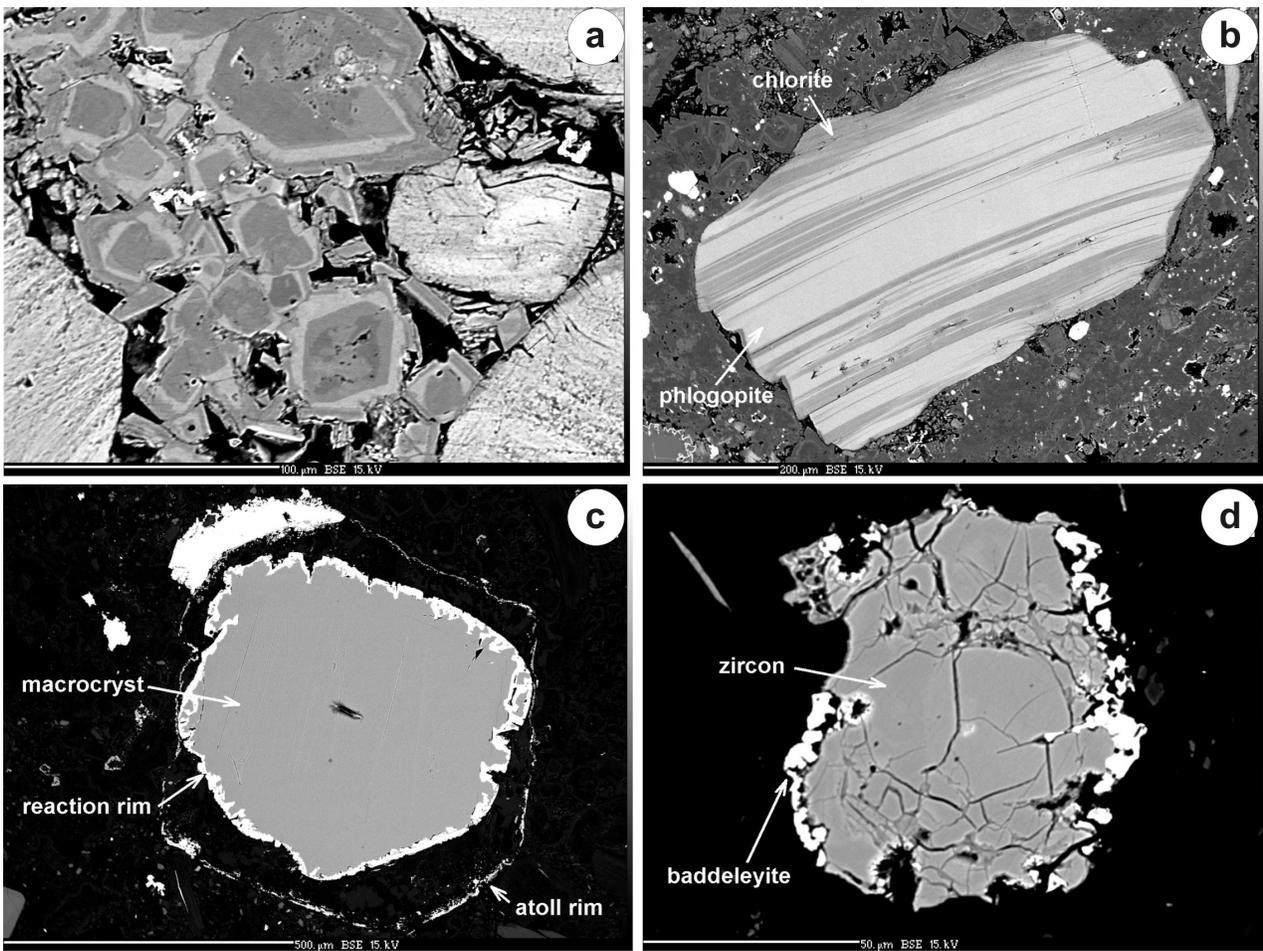


Figure GS-5-4: Mineral interrelations in sample GBO-16 imaged in back-scattered electrons: **a)** zoned rhombohedral dolomite crystals and chloritized phlogopite form segregations; **b)** phlogopite macrocryst partly replaced by chlorite (dark grey); **c)** atoll-like spinel macrocryst showing a Mg-Al-Cr-rich core (dark grey) and a Fe-Ti-rich rim (white); **d)** fractured zircon crystal (dark grey) mantled by Zr oxides (white).

material was derived from the surrounding metasedimentary rocks or deeper crustal levels. Chlorite, developed after the primary phlogopite (Figures GS-5-3c, 4b), covers a range of compositions that is remarkably similar to the phlogopite range in terms of key chemical parameters: $mg\# = 0.64$ to 0.84 , $Al_2O_3 = 10.8$ to 14.3 wt. %, $TiO_2 = 0$ to 1.4 wt. %, $F = 0.6$ to 1.2 wt. % (cf. phlogopite above). Chlorite macrocrysts from sample EPV-12A-94 are generally similar, but have slightly higher $mg\#$ (0.68 – 0.88). These data indicate that most, if not all, of the chlorite in the examined samples is the product of phlogopite replacement, whereas low-grade metamorphic chlorite is characteristically lacking.

Spinel and **ilmenite** are both important minerals from the standpoint of mineral exploration (see below). Spinel macrocrysts are dark brown to opaque in plane-polarized light and show a relatively simple zoning pattern in BSE images; some crystals also exhibit a well-defined atoll texture (Figure GS-5-4c). The core of large macrocrysts is enriched in MgO , Al_2O_3 and Cr_2O_3 (10.0 – 13.4 , 11.1 – 21.9 and 46.7 – 55.6 wt. %, respectively), and low in

TiO_2 (≤ 1.0 wt. %, but undetectable in the most Mg-Al-Cr-rich compositions). Back-scattered electron (BSE) imaging reveals that many macrocrysts are fragments of larger crystals showing variations in MgO , Al_2O_3 and TiO_2 contents. In these cases, enrichment in TiO_2 is invariably accompanied by depletion in MgO and Al_2O_3 . In comparison with the core, the narrow outer rim of the large macrocrysts (Figure GS-5-4c) and smaller grains are depleted in MgO and especially Al_2O_3 (4.6 – 7.4 and ≤ 1.2 wt. %, respectively), but enriched in TiO_2 (up to 7.0 wt. %) and $(FeO+Fe_2O_3)$. The earliest compositions to crystallize are ternary spinels with roughly equal proportions of $MgAl_2O_4$, $MgCr_2O_4$ and $FeCr_2O_4$, but with a general predominance (35 – 45 mol %) of the latter component. The later-crystallized Ti-rich Mg-Al-depleted compositions contain nearly equal proportions of $MgCr_2O_4$, $FeCr_2O_4$ and $FeFe_2O_4$ (20 – 40 mol %) at low to moderate $TiFe_2O_4$ and low $MgAl_2O_4$ contents. Ilmenite shows little variation in chemistry. It is characteristically rich in MgO and Cr_2O_3 (9.3 – 10.0 and 1.6 – 2.6 wt. %, respectively), but poor in MnO and Nb_2O_5 .

(0.4–0.5 and ≤ 0.2 wt. %, respectively). All of the analyzed grains contain an appreciable proportion of oxidized iron ($\sim 20\%$ of all Fe is Fe^{3+} , or 8 mol % Fe_2O_3).

Fluorapatite (i.e., apatite, in which F is the dominant monovalent anion) contains small but measurable concentrations of Na, Sr and rare earth elements (up to 0.5 wt. % Na_2O , and 1.1 wt. % SrO and $\text{Ce}_2\text{O}_3 + \text{La}_2\text{O}_3 + \text{Nd}_2\text{O}_3$). Other elements of petrogenetic significance (Si, S, Ba and Cl) are either at or below the detection limit of WDS.

Zircon macrocrysts contain only minor amounts of HfO_2 (1.1–1.4 wt. %); other potential substituents (Ca, Fe, Ti, Th, U, Nb, Ta, Y and Al) were sought, but not detected. The opaque rim, clearly visible in plane-polarized light, is discontinuous (Figure GS-5-4d) and consists of two or more Zr oxides. The size of individual grains is too small for these minerals to be confidently identified, but the bulk of this material is probably baddeleyite (ZrO_2).

Geochemistry

Random off-cuts of samples EPV-12A-94 and GBO-16 (~ 7 g each) were cleaned, pulverized and analyzed for major and trace elements at Activation Laboratories Ltd. (Ancaster, Ontario). ‘Randomization’ was deemed impractical because of the fine-grained nature of the samples (*see* ‘Petrography’ section) and scarcity of material. Conventional lithium–borate fusion and inductively coupled plasma techniques were used for all major and trace elements except CO_2 (determined by

coulometry).

Both studied samples have the same major- and trace-element signature:

- high MgO , CaO and CO_2 (>20 wt. % each), moderate SiO_2 and $(\text{Fe}_2\text{O}_3)_\Sigma$ (3–8 wt. %), and low TiO_2 , Na_2O , K_2O and MnO contents (<1 wt. %); the Al_2O_3 and P_2O_5 values are low in EPV-12A-94, but appreciable in GBO-16 (2.6 and 1.5 wt. %, respectively) owing to the much greater abundance of phlogopite, chlorite and fluorapatite in the latter
- high concentrations of compatible trace elements (42–98 ppm Sc, 42–87 ppm V, 28–112 ppm Co, 910–990 ppm Cr and 380–610 ppm Ni)
- high concentrations of ‘incompatible’⁶ trace elements, particularly Sr (173–674 ppm), light rare earth elements ($\text{LREE}_\Sigma = 351\text{--}878$ ppm), Nb (162–189 ppm), Zr (270–290 ppm), Th (60–100 ppm) and U (19–22 ppm)
- high Nb/Ta and Zr/Hf ratios (>21 and 48, respectively), but a low Zr/Nb ratio (1.5–1.7) relative to the primitive-mantle composition (McDonough and Sun, 1995)
- primitive Y/Ho ratio (25–28)

The trace-element compositions of the two rocks are remarkably similar: both show moderate enrichment in LREE relative to heavy REE, strong negative Rb–Ba, K, Zr–Hf and Ti anomalies and moderate-to-strong Pb and Sr anomalies (Figure GS-5-5).

⁶ The term ‘incompatible element’ is misleading when used in the context of small-volume mantle-derived rocks (e.g., kimberlite, carbonatite, etc.), because most of these elements behave compatibly with respect to the primary minerals crystallizing from such magmas (phlogopite, perovskite, pyrochlore, etc.).

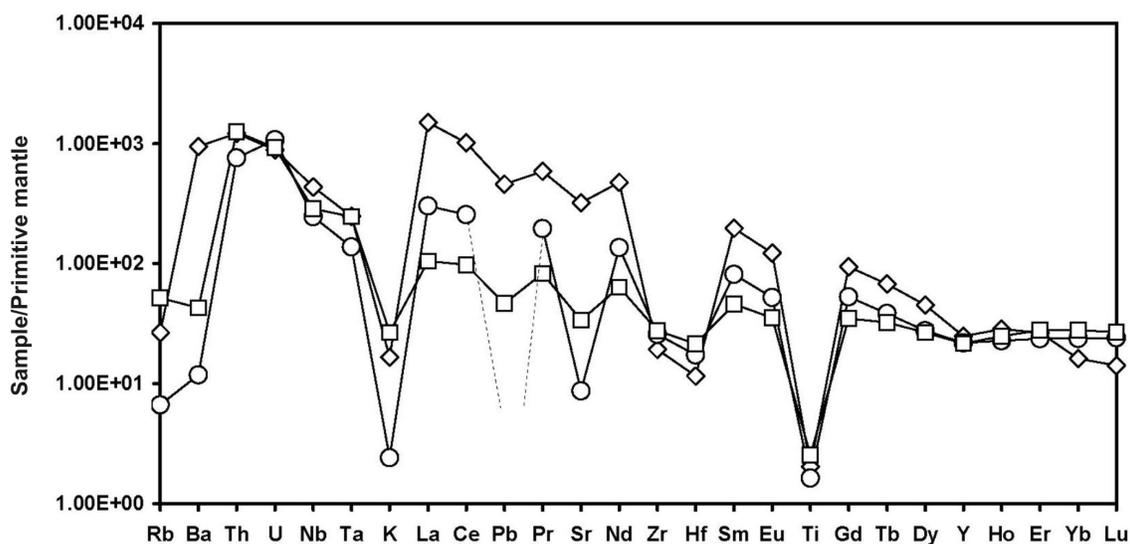


Figure GS-5-5: Abundances of selected ‘incompatible’ trace elements in the Wekusko Lake rocks (EPV-12A-94: circles; GBO-16: squares) normalized to the primitive mantle of McDonough and Sun (1995). The ‘average carbonatite’ pattern (Chakhmouradian, unpublished data) is shown for comparison (diamonds).

Discussion

The accumulated petrographic, mineralogical and geochemical evidence indicates that samples EPV-12A-94 and GBO-16 were derived from similar sources and underwent broadly similar evolutionary histories. The observed differences in texture and modal mineralogy are difficult to explain in terms of a single petrogenetic process (e.g., crystal fractionation or vapour release), and it is more likely that the two rocks are different (but related) intrusion phases rather than different facies derived from the same magma batch. In either case, EPV-12A-94 is the more evolved of the two, as indicated by the greater abundance of carbonate segregations, lower abundance of macrocrysts, complete conversion of phlogopite to chlorite, and higher Zr/Hf, Nb/Ta and LREE₂ values in this sample. At this point, any attempt to constrain the exact source and evolutionary paths of the magmas that produced the Wekusko Lake 'kimberlitic' rocks would be premature. The presently available data, however, will allow some important conclusions to be drawn about the nature of these rocks and their economic potential.

Kimberlite-carbonatite dilemma

It has long been recognized that several types of mantle-derived volatile-rich and silica-undersaturated rocks (e.g., ultramafic lamprophyre, silicocarbonatite and melilitite) are texturally, modally and chemically akin to kimberlite. The similarity in texture reflects the similarity of processes involved in the emplacement of alkali- and volatile-rich silica-undersaturated magmas at shallow crustal levels. The similarity in mineralogy and/or whole-rock chemistry is only apparent and may be readily refuted upon a detailed examination and comparative analysis. Further discussion of this subject is beyond the scope of the present report; interested readers are referred to the discussions in Mitchell (1986), Mitchell (1995), Mitchell et al. (1999), Taylor and Kingdom (1999) and Tappe et al. (2005). It is important to note that, not only kimberlite (*sensu lato*), but also such mantle-derived rocks as lamprophyre and carbonatite may contain significant concentrations of diamond (e.g., Djuraev and Divaev, 1999; Kornilova et al., 1999; Litvin et al., 2001; Tappe et al., 2004; De Stefano et al., 2006).

It is quite clear that the two samples studied in the present work are *not* group-I kimberlite, because they completely lack macrocrystic and (micro)phenocrystic olivine (as well as serpentine or any other products of olivine alteration), and are composed largely of dolomite (70–85% by weight). In group-I kimberlite, calcite is the principal carbonate mineral present as (micro)phenocrysts, groundmass grains, segregations and secondary alteration products (Dawson and Hawthorne, 1973; Mel'nik et al., 1982; Podvysotskii, 1985; Armstrong et al. 2004). Primary dolomite is much less common and, where present,

forms (reaction) mantles and overgrowths on earlier crystallized calcite (e.g., Chakhmouradian and Mitchell, 1999; Armstrong et al., 2004). Even evolved macrocryst-poor and carbonate-rich varieties of kimberlite (sometimes incorrectly referred to as carbonatite) are composed predominantly of calcite (Dawson and Hawthorne, 1973; Beard et al., 2000). Late-stage dolomite, developed at the expense of olivine (and, to a lesser degree, other Mg-Fe minerals) contemporaneously with serpentine, is much more common (e.g., Mel'nik et al., 1982), but texturally distinct from the groundmass and segregation dolomite observed in the Wekusko Lake samples. To the best of the authors' knowledge, the only reported example of kimberlite with predominantly dolomitic groundmass is a dike in Keith Township (Ontario), described by Watson et al. (1978). It is noteworthy, however, that 1) the identification of carbonate in that rock as dolomite was based solely on x-ray powder diffraction data, and 2) the modal composition of the Keith Township dike is not entirely consistent with that of typical kimberlite. Also noteworthy is the absence in the present samples of poikilitic groundmass phlogopite, monticellite, perovskite, and Mg-poor ulvöspinel-magnetite (titanomagnetite), the 'hallmark' minerals of group-I kimberlite (Mitchell, 1986, 1995).

Group-II kimberlite, also referred to in the literature as micaceous kimberlite or orangeite (Mitchell, 1995), is a rare type of ultrapotassic volatile-rich (primarily, H₂O-rich) rock ranging in composition from unevolved ultramafic varieties with <38 wt. % SiO₂, <4 wt. % Al₂O₃ and >21 wt. % MgO to evolved varieties containing >41 wt. % SiO₂, >4 wt. % Al₂O₃ and <21 wt. % MgO. Evolved group-II kimberlite is compositionally similar to olivine lamproite (except for somewhat higher Ba and Zr in the latter), whereas unevolved varieties can be distinguished from olivine lamproite by their lower SiO₂, and higher CaO and CO₂ contents. It is more difficult to discriminate between group-I and group-II rocks on the basis of whole-rock chemistry (Mitchell, 1995). Unevolved group-II kimberlite is generally richer in K and poorer in Ti, although there is much overlap between the two rock types below 1.5 wt. % K₂O and TiO₂. The enrichment of group-II kimberlite in K and depletion in Ti are reflected in the mineralogy of these rocks, including the presence of ubiquitous macrocrystic, microphenocrystic and groundmass phlogopite, and paucity of Mg-Ti-rich spinel. Group-II rocks may also contain an assortment of exotic minerals (e.g., aegirine, Ti-Zr garnet, Ba and REE carbonates) that are not observed in either group-I kimberlite or olivine lamproite. Thus far, true orangeite has been observed only in southern Africa. Most of these occurrences are restricted to the Kapvaal craton, and range from 110 to 200 Ma in age (Skinner, 1989). The only economically diamondiferous (0.4–0.5 ct/tonne) group-II kimberlite is the Finsch pipe, 165 km west of Kimberley, South Africa.

Group-II kimberlite occurrences from the Bellsbank

field in South Africa have been reported to contain a high proportion of dolomite, to the extent where it is the principal constituent in the groundmass (Mitchell, 1995, p. 245). It is important to note, however, that this groundmass dolomite is believed to have replaced serpentine, and fresh samples contain more serpentine than dolomite (Mitchell, 1995, p. 245). A similar mechanism has been proposed for shallow dolomite-rich portions of the Finsch pipe (Ruotsala, 1975). In contrast, the authors found no indication that the groundmass dolomite in the Wekusko Lake samples developed at the expense of serpentine or any other mineral. The composition of macrocrystic phlogopite in the two studied samples partly overlaps with the compositional range of mica from group-I and group-II kimberlite; however, phlogopite compositions with $mg\# < 0.75$ are virtually absent from group-I rocks and restricted to extremely Al-depleted Fe ($\pm Ti$)-rich mica in group-II kimberlite ('orangeite tetra-ferriphlogopite trend' of Mitchell, 1995; *see also* Taylor and Kingdom, 1999; Kaminsky et al., 2004; Masun et al., 2004). The compositional range of spinel macrocrysts from the Wekusko Lake rocks overlaps that of macrocrystic spinel from group-I and group-II kimberlites (Mitchell, 1995; Schulze, 2001; Kaminsky et al., 2004; Masun et al., 2004; Skinner et al., 2004). Note, however, that chemically similar spinel is found in a broad range of rocks, including mantle peridotite (e.g., Boyd et al., 1999; Lee and Rudnick, 1999); that is, these minerals "have no compositional characteristics permitting their assignment to a particular paragenesis" (Mitchell, 1995, p. 195). Magnesium-poor ($mg\# < 0.25$) Fe-Ti-rich compositions approaching the chromite-ulvöspinel/magnetite side of the spinel prism (orangeite group-B spinel of Mitchell, 1995) are absent in the rocks examined in the present study. Ilmenite is an important indicator mineral that exhibits paragenesis-dependent compositional characteristics. The compositions obtained in this study plot in the compositional field of picroilmenite from group-I kimberlite, and are distinct from ilmenite in group-II rocks in containing much lower Mn and Nb contents (Mitchell, 1995; Kaminsky et al., 2004).

The whole-rock geochemistry of the Wekusko Lake samples is, perhaps, most difficult to reconcile with the identification of these rocks as kimberlite. None of the fresh kimberlite analyses reported in the literature approach the major-element composition of the two studied samples. Using the whole-rock chemistry as the primary classification criterion, the rocks in question should be termed magnesio-carbonatite (Le Maitre, 2002, p. 10). A comparison of trace-element abundances sheds further light on the identity of the Wekusko Lake rocks. Although their V, Cr, Co, Ni, Sr, Ba, Nb, Zr and REE abundances are similar to those of group-I and group-II kimberlite, they contain significantly lower levels of Sc, Th and U, and have lower Zr/Hf and Nb/Ta ratios (typically, < 45 and 20 , respectively), relative to the values reported in the present study (e.g., Mitchell, 1995;

O'Brien and Tyni, 1999; Kaminsky et al., 2004; Nowicki et al., 2004). The strong Ti anomaly (Figure GS-5-5) is also atypical of kimberlite (e.g., Mitchell, 1995; O'Brien and Tyni, 1999; Price et al., 2000). Further comparison demonstrates close similarity between the 'incompatible' trace-element budget of samples EPV-12A-94 and GBO-16 and the 'average carbonatite' (Figure GS-5-5). In common with the Wekusko Lake rocks, carbonatite shows superchondritic Nb/Ta and Zr/Hf, but primitive Y/Ho values, which has important implications for the interpretation of carbonatite petrogenesis (Chakhmouradian, 2006; A. Chakhmouradian, work in progress, 2006). Although the Cr, Co and Ni abundances are significantly lower in the 'average carbonatite' (39, 16 and 29 ppm, respectively; Chakhmouradian, unpublished data) than the present samples, (silico)carbonatite with comparable levels of compatible trace elements has been described in the literature (e.g., Bagdasarov, 1981; Stoppa and Woolley, 1997; Ripp et al., 2006). It is important to note that such carbonatite rocks are not associated with any alkaline silicate rocks and are believed to have formed directly from mantle-derived carbonatitic melts.

Thus, the Cr, Co and Ni enrichment of the Wekusko Lake rocks can be explained by the entrainment of mantle-derived xenocrysts (magnesian aluminous chromite and magnesian ilmenite) in ascending carbonatitic magma. The essentially dolomitic makeup of these rocks is consistent with the composition of carbonate melts in equilibrium with upper-mantle peridotite (Wallace and Green, 1988; Sweeney, 1994). Spinel and ilmenite xenocrysts of composition similar to those described in this report have been documented in kimberlite and other mantle-derived silica-undersaturated rocks (e.g., Boyd et al., 1999; Hearn, 1999; Lee and Rudnick, 1999; Grégoire et al., 2002; Aulbach et al., 2004).

Mica compositions from the Wekusko Lake samples are distinct from typical mantle phlogopite in their higher Fe and Mn, but lower Cr content. Mica from metasomatized peridotite in both on- and off-craton settings spans a wide range of compositions, but normally has $mg\# > 84$, $MnO < 0.1$ wt. % and $Cr_2O_3 > 0.1$ wt. % (Dawson, 1999; Lee and Rudnick, 1999; Moine et al., 2001; van Achterberg et al., 2001; Grégoire et al., 2002). The macrocrysts analyzed in the present work plot in the $mg\#$ and Al_2O_3 range of carbonatitic mica and contain comparable levels of Ti, Mn, Na and Ba (e.g., Seifert et al., 2000; Chakhmouradian and Zaitsev, 2002; Zaitsev et al., 2004). It is noteworthy that evolved Al-deficient mica transitional to tetra-ferriphlogopite $[K(Mg,Fe)_3(FeSi_3O_{10})(OH,F)_2]$ was not observed in the Wekusko Lake rocks. The minor-element chemistry of dolomite (Fe, Mn, Sr) and fluorapatite (Sr, REE) from samples EPV-12A-94 and GBO-16 is also consistent with a carbonatitic origin (e.g., Dawson et al., 1996; Seifert et al., 2000; Dawson and Hinton, 2003; Zaitsev et al., 2004).

Economic considerations

Accurate petrographic characterization of the Wekusko Lake rocks is primarily crucial for determining the nature and extent of melting in the underlying mantle and secondarily for evaluating the economic potential of this specific rock suite. The assessment of this region's diamond potential requires constraining such parameters as the depth of magma generation, the composition of mantle source(s) involved in the melting and the degree of melting. In common with kimberlite, carbonatitic magmas can potentially transport diamonds from the mantle. The alleged incompatibility of elemental and oxidized carbon is a misconception: diamond readily forms in natural and synthetic carbonate systems (e.g., Litvin et al., 2001; Pal'yanov et al., 2002; Shatsky et al., 2005). Both graphite and carbonate minerals can serve as the source of carbon for the nucleation and growth of diamonds at pressures in excess of 6 GPa (ca. 180 km depth).

In the absence of diamond-count data, the chemistry of indicator minerals and, to some extent, whole-rock geochemistry can be used for the assessment of diamond potential. Spinel macrocrysts of composition identical to the cores of the Wekusko Lake macrocrysts have been recovered from diamondiferous kimberlite in many areas around the world, including the Kaapvaal craton, Buffalo Hills and Attawapiskat kimberlite fields (Kong et al., 1999; Schulze, 2001; Hood and McCandless, 2004). The most Cr-rich compositions (55.1–55.6 wt. % Cr_2O_3) plot just outside the diamond-inclusion field of Gurney and Zweistra (1995). The enrichment of macrocryst cores in Al ([Cr/(Cr+Al)] = 0.6–0.8) and variations in Mg, Al and Ti contents observed in some of the grains indicate metasomatic processes in the mantle source (Schulze, 2001), which is in agreement with the trace-element geochemistry of the Wekusko Lake rocks (see above). Mantle metasomatism (by carbonate-rich melts or fluids) has been recognized as a potential driving force behind diamond formation in the mantle (e.g., Navon, 1999; Cartigny et al., 2004). The largely dolomitic composition and K-Rb-Ba-depleted trace-element signature of the examined material indicate very low degrees of melting and the retention of refractory phlogopite in the source. It is not inconceivable that the most Mg-rich phlogopite macrocrysts ($\text{mg}\# = 0.80\text{--}0.85$), also showing the lowest Al and Mn contents, may represent xenocrystic material. Unfortunately, the relationship between the chemistry of xenocrystic mica and diamond potential has not been studied in adequate detail.

The compositions of magnesian ilmenite plot in the upper middle part of the well-defined kimberlitic trend and well away from the reference curve for 'non-kimberlitic' ilmenite (Wyatt et al., 2004). It is noteworthy that, although ilmenite compositions from some alnöitic rocks

straddle the reference curve (Wyatt et al., 2004), few of them contain as much MgO and Cr_2O_3 as the ilmenite macrocrysts examined in this study. The low Fe_2O_3 values of the Wekusko Lake ilmenite indicate relatively reducing (and therefore 'diamond-friendly') crystallization conditions.

The absence of garnet and other indicator minerals in the examined samples (cf. Gurney and Zweistra, 2003) makes further assessment of the economic potential of the Wekusko Lake rocks difficult. Additional information about the sources and processes involved in the petrogenesis of these rocks can be extracted from trace-element and isotopic data (see below).

Further work

A more detailed petrogenetic interpretation of the Wekusko Lake suite of 'kimberlitic dikes' will require the analysis of whole-rock and single-crystal isotope data. The stable-isotope (C, O) and radiogenic-isotope (Nd, Sr, Pb) compositions of these rocks will be used to determine the nature of their mantle source and the extent to which this primary isotopic signature was modified by magma evolution in the crust. The C-O isotope composition of carbonate minerals is sensitive to fractionation and degassing processes in the magma, as well as interaction of the magma with crustal materials (e.g., Bell and Rukhlov, 2004; Demény et al., 2004). The whole-rock radiogenic-isotope data have been successfully used for constraining the mantle sources of carbonatites and kimberlites, and the nature of processes that generate carbonatitic and kimberlitic melts in the mantle (e.g., Harmer, 1999; Bell and Rukhlov, 2004). The age of the Wekusko Lake rocks will be determined by combining single-mineral (U-Pb on zircon and Rb-Sr on phlogopite) and whole-rock geochronology. In order to aid diamond-exploration efforts in the region, the mineral-chemistry database for the Wekusko Lake rocks will be expanded and compared with the data for locally derived indicator minerals.

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