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Publications in this series have not been edited; they are released as submitted by the author.

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ABSTRACT

Till (n=196) and stream sediment (n=60) samples were collected in the area south and west of Great Slave Lake, Northwest Territories (NTS 85B, C, F, and G), over the course of 3 summer field seasons. Samples were processed to recover kimberlite and other indicator minerals. This report summarizes results of the kimberlite indicator mineral (KIM) studies, including measures of KIM mineral types, abundances, and chemistry (major, trace, and rare earth elements). KIMs were present in 24% of the samples collected, and only 183 KIM grains in total were recovered, of which Cr-pyrope garnets were the most abundant (65.6%). Chemical analyses revealed strong similarities to the Drybones Bay and Mud Lake kimberlites which are situated 50 to >100 km to the northeast, roughly aligned with prominent glacially streamlined landform flowsets in this field area. Results suggest there is little evidence for undetected kimberlite outcrop or sub-crop in the study area.

1.0 INTRODUCTION

This Open File presents the major oxide, trace element, and rare earth element (REE) mineral chemistries of kimberlite indicator minerals (KIMs) recovered from stream sediment and till samples collected in the southern NWT region, south and west of Great Slave Lake (Fig. 1). Fieldwork and sample collection were undertaken in the summers of 2017, 2018, and 2019, as part of Natural Resources Canada's Geo-mapping for Energy and Minerals program, Southern Mackenzie Surficial activity (Paulen et al., 2017, 2019; Day et al., 2018). A total of 256 samples (196 till samples, 60 stream sediments samples; including 8 field duplicates) were collected as part of a larger investigation of potential undiscovered economic mineralization in the region (NTS sheets 85B, C, D, E, F, G; 95B, G, H). Fieldwork was also conducted in support of regional 1:100 000 scale surficial geology mapping of NTS sheets 85C and 85F, and 1:125 000 scale mapping of 85G. This research follows upon earlier surficial geology, mineral prospecting, and KIM studies conducted west and south of the focus study area (Gal and Lariviere, 2004; Day et al., 2005, 2007; Hannigan, 2006; Plouffe et al., 2006, 2007, 2008; Huntley et al., 2008; Mills, 2008; Pronk, 2009; Watson, 2011a, b, 2013; McClenaghan et al., 2012; Pitman, 2014; Poitras et al., 2018), and in the Drybones Bay, eastern Great Slave Lake region (Carbno, 2000; Kerr et al., 2000; Carbno and Canil, 2002; Kaminski et al., 2013; Sheng, 2016; Reguir et al., 2018). Standard Geological Survey of Canada (GSC) project and sample metadata are reported in Appendix 1.

1.1 Location and Physiography

The study area is situated south and west of Great Slave Lake, largely within the core study area of NTS map sheets 85B (north half), 85C, 85F, and 85G (Fig. 1). The area is characterized as a broad, low-relief plain, overlying gently westward-dipping Phanerozoic bedrock (Okulitch, 2006). In the south, the prominent Cretaceous bedrock Cameron Hills upland rises ~600 m above the boreal plains. Bedrock outcrops along the northern edge of the prominent Devonian limestone escarpment that trends westward, paralleling the south shore of Great Slave Lake and the upper Mackenzie River. Elsewhere, bedrock is exposed in the headwalls of slumps and along deeply dissected stream channels draining off Cameron Hills, and along northward-draining streams that incise the Devonian limestone platform (e.g. Hay and Kakisa rivers).

1.2 Glacial History

The study area was inundated by the Laurentide Ice Sheet (LIS) during the Late Wisconsinan glaciation (~24-10 ^{14}C ka BP; Dyke and Prest, 1987; Dyke, 2004). Glacially streamlined landforms and bedrock striae record three phases of ice flow; first to the southwest (230°), then westward (250°), and finally northwest (305°). During ice retreat, drainage was extensively ponded within ice-marginal topographic and glacioisostatically-depressed basins, which eventually merged to form the expansive glacial Lake McConnell (Fig. 2; Lemmen et al., 1994; Smith, 1994). Ice is considered to have retreated from the study area between 11-10 ^{14}C ka BP (Dyke, 2004). However, optical stimulated luminescence ages from eolian dunes and beach ridges that formed after parts of glacial Lake McConnell had drained, indicate that ice retreat occurred up to 500 years earlier (Hagedorn et al., 2019). By 8.5 ^{14}C ka BP, as the eastward retreating LIS exited the Mackenzie drainage basin, glacial Lake McConnell had shrunk to form three disconnected basins (Great Bear, Great Slave and Athabasca; Fig. 2). Drainage of these basins continued through the Holocene largely through a process of decantation and glacioisostatic uplift (cf. O'Neill et al., 2019).

1.3 Surficial Geology

Around the historic Pine Point mining district (Fig. 1), till thicknesses range from 1 m in the east to >25 m in the west (Lemmen, 1990; Rice et al., 2013). Across the study area, data from seismic shothole drillers' logs (Smith and Lesk-Winfield, 2010a), exploration diamond drill holes, and petroleum well log records (Smith et al., 2019) have been used to identify extensive till blankets and undifferentiated drift thicknesses >8 m, and up to 150 m in areas north of the Devonian escarpment. Tills characteristically have a silty-clay matrix (1:3 sand to silt-and-clay ratio), with around 15% clast content. Clasts are largely of local Paleozoic bedrock lithologies and some further travelled Precambrian Canadian Shield clasts (Plakholm et al., 2019). In places, eolian sand sheets and dunes have formed, typically in association with large glaciolacustrine deltas, outwash plains, and areas of coarse glaciolacustrine sediment accumulation (Lemmen et al., 1998a, b; Wolfe et al., 2009; Hagedorn et al., in press – a, b; Smith et al., in press; Paulen and Smith, in press). While much of the low-lying area was inundated by glacial Lake McConnell (Fig. 2; Lemmen et al., 1994), there is scant depositional evidence of its occurrence. Instead, across much of the area in NTS 85G, eastern 85F and central 85C, there is a conspicuous but sporadic cobble-boulder surface lag sitting atop <50 cm of littoral glaciolacustrine sediments/washed till. Fine sediments winnowed from areas of till cover appear to have either been re-deposited into deeper basins (e.g. areas now submerged by Great Slave Lake), or were decanted down the Mackenzie River outflow. Much of the field area is blanketed by bog and fen, reflecting the low relief topography, presence of discontinuous permafrost (Hegginbottom et al., 1995), and the relative impermeability of underlying fine-grained till. Thicknesses of peat in these terrains are estimated from seismic shothole drillers' logs to average 2.76 m (considered an over-estimate), and range up to 9.1 m (Smith and Lesk-Winfield, 2010b).

1.4 Bedrock Geology

The study area is situated within the Western Canada Sedimentary Basin (WCSB), which consists of a western-thickening wedge of Phanerozoic sedimentary rocks overlapping the Precambrian craton to the east, and abutting the deformed belt of the Cordillera to the west (Fig. 3; Porter et al., 1982; Okulitch, 2006). Sedimentary successions of the WCSB have been affected by cratonic arches and faulting in the underlying crystalline Proterozoic basement, which has produced the prominent northeast-southwest trending Great Slave Lake Shear Zone, and a series of subparallel fault zones (e.g. Tathlina and Rabbit Lake fault zones; Fig. 3; Eaton and Hope, 2003; Hannigan, 2006).

The study area is directly underlain by Middle to Upper Devonian reef complexes and platform carbonates, which in places have undergone dolomitization to form the Presqu'ile barrier complex (Kyle, 1981; Rhodes et al., 1984). Significant Mississippi Valley-type Pb-Zn mineralization occurs in the historic world-class Pine Point mining district (Fig. 1). Similar Pb-Zn mineralization is reported from dolomitized limestone in diamond drill holes and petroleum well cores south of Great Slave Lake at core depths of 300-400 m below uppermost bedrock contact (e.g. Gulf Minerals Canada Ltd., 1979; Cominco, 1980). Lead-zinc mineralization in dolomitized and karst limestone also occurs at/near surface around Windy Point and Qito, west of Great Slave Lake (Fig. 3).

1.4.1 *Kimberlites*

Olivut Resources Limited's HOAM project, using airborne and ground magnetic surveys, airborne gravity surveys, drilling, and stream sediment and till surveys, identified 29 kimberlites (2 diamondiferous) south and west of Fort Simpson, NWT (Fig. 3; Pitman, 2014). The KIM chemistries of these kimberlites differs from those from previously conducted regional sediment surveys (which have high proportions of G10 garnets and Mg-rich picroilmenites), suggesting that additional, potentially diamondiferous kimberlites remain to be identified (Pitman, 2014; Poitras et al., 2018). The Horn Plateau stream sediment survey (Fig. 3; Day et al., 2005, 2007) yielded >3300 KIMs from 324 samples, and included a high proportion of Cr-pyrope garnets. This survey also yielded a single dull-grey diamond from the 0.5 – 1.0 mm fraction of one sample — the first such diamond ever recovered in a GSC National Geochemical Reconnaissance (NGR) survey. As part of a regional paleo-environmental reconstruction study, 25 small (2-10 kg) surface sediment samples were collected and analyzed for KIM content (Mills, 2008). Despite their small size, these yielded 61 KIMs, including 43 peridotitic pyrope garnets. The kimberlitic source of the Horn Plateau KIMs remains unknown, but cannot have been derived from the known HOAM kimberlites that are 30 to >100 km glacially down-flow. With the establishment of the Edéhzhíe Protected Area most of the Horn Plateau region has now been excluded from future mineral exploration and development considerations.

East of the study area, within the Precambrian Slave Craton, >300 kimberlites have been identified in the central Slave Craton – Lac de Gras field alone (Fig. 3; Giuliani and Pearson, 2019). Smaller clusters of kimberlites have been identified in the west and southeast Slave Craton regions (Fig. 3). The most proximal known kimberlites to our study area are the Drybones Bay kimberlite pipe and Mud Lake kimberlitic dyke (4 km to the south of Drybones Bay), situated along the eastern shore of Great Slave Lake (Fig. 3). The Drybones Bay kimberlite is only 50 km northeast of the easternmost land in NTS 85G, and 220 km northeast of Fort Providence. The Drybones Bay kimberlite was discovered in 1994 and erupted at 445.6 ± 0.8 Ma ($^{207}Pb/^{235}U$ Concordia age, 2σ standard error, measured on zircon marcrocrysts; Kretschmar, 1997; Reguir et al., 2018). It was originally classified as juvenile lapilli-bearing volcaniclastic kimberlite (VK), with no evidence of tuffisitic kimberlite (TK), tuffisitic kimberlite breccia (TKB), or hypabyssal kimberlite (HK) facies (Field and Scott Smith, 1999). Subsequent research indicated that the Drybones Bay pipe is an elongated intrusion (900 m long, 400 m wide), comprising 2 or 3 separate intrusive phases, and includes TK (similar to Gahcho Kué), pyroclastic kimberlite (PK) and resedimented volcaniclastic kimberlite (RVK) facies, and may well host a HK root zone (inferred from autoliths; Sheng, 2016). It is diamond-bearing, and contains abundant serpentinized olivine, ilmenite, pyrope and chromite mineralogy (Kretschmar, 1997; Kerr et al., 2000; Carbno and Canil, 2002; Kaminski et al., 2013; Sheng, 2016). A 10.1 tonne bulk sample of Drybones Bay kimberlite yielded 96 macro-diamonds (one dimension >0.50 mm), and grades up to 39.2 cpht were determined on a 1025 kg sample of the central unit kimberlite cluster (Kretschmar, 1996). Kimberlite bulk chemistry and indicator mineral chemistries of the Drybones Bay kimberlite facies led Sheng (2016) to suggest that its diamond potential is likely to be greatly underestimated. The Mud Lake kimberlite is a hypabyssal kimberlitic dyke, with an eruption $^{207}Pb/^{235}U$ Concordia age (2σ) determined on zircon macrocrysts, of 469.6 ± 9.7 Ma (Sheng, 2016). Only 2 macro-diamonds were recovered from the Mud Lake drill core, and news releases by Snowfield Development Corporation (as reported in Sheng, 2016) indicated that the first 100 tonnes of material excavated from the Mud Lake kimberlite, processed for diamond evaluation, were not promising. Mineral chemistries for KIMs in the kimberlite that were reported by Kretschmar (1997) and Sheng (2016) suggest diamond resorption was likely in the Mud Lake kimberlite.

2.0 METHODS

Field research, stream sediment, and till sample sites were accessed by truck, helicopter and boat, over the course of three summer field seasons. Basecamps were established in Hay River and Fort Providence, and wildlife monitors from local First Nation and Métis communities (Kátł'odeeche (Hay River), Ka'a'gee Tu (Kakisa), Deh Gáh Got'ie Koe (Fort Providence), Deninu Kue (Fort Resolution), and the Fort Resolution Métis Council), principally accompanied helicopter crews when we were working in their traditional territory.

2.1 Till Sampling

Till samples were collected at 10-15 km spacing throughout the northern half of NTS 85B, and throughout NTS 85C, F, and G (Fig. 1, Appendix 2). Additional samples were collected along the Mackenzie #1 Highway across to Highway #3 and south past Fort Liard to the NWT/BC border. These sites were included in order to establish longer-distance erratic clast transport vectors, provide overlap and context with previous regional sediment surveys, and support base metal exploration studies and links to known mineralization occurrences in proximity to the Liard River (e.g. Dudek, 1993). Till samples (12-37 kg) were mostly collected from shovel-dug holes, within relatively unaltered parent material (soil Cca horizons), following GSC standard methodologies (cf. Spirito et al., 2011, McClenaghan et al., 2013, 2020; Plouffe et al., 2013; Paulen et al., 2017, 2019). Sample depths generally exceeded 70 cm, and in rare cases were limited by contact with shallow frozen ground (permafrost and/or seasonal freezing). Till samples were also collected along road cuts, and in borrow pits and quarries. Other than removing rootlets, cobbles and boulders, samples were collected as found with no field processing or concentration. Stratigraphic samples collected from a single site were labelled with the same sample number and then used letters A, B, C (etc.) to identify different samples, e.g. 17-SUV-033-A, 17-SUV-033-B (Appendix 2). Duplicate samples collected from the same site were given different sample numbers, e.g. 17-PTA-019 and 17-PTA-020 (Appendix 2).

2.2 Stream Sediment Sampling

Stream sediment sampling followed standard GSC NGR protocols (Friske and Hornbrook, 1991; Paulen et al., 2017; Day et al., 2018, 2019, in progress), and involved the collection of 8-20 kg samples of mostly sand-sized sediments, wet sieved through a #10 mesh (2.0 mm) screen while standing in the stream. Where possible, stream sediment samples were collected at sites where stream flow would naturally trap heavy minerals, such as at the head of mid-channel and lateral bars, and behind boulder traps (cf. Prior et al., 2009). Stream sediment sampling was limited owing to the low relief and extensive bog/fen throughout much of the field area (Fig. 1, Appendix 2). High water levels in 2018 also restricted collections, necessitating a return to the field in 2019 in order to collect samples in key areas and as follow up to significant heavy mineral results from the 2017 and 2018 sample collections.

2.3 KIM Sample Processing and Quality Control

Field samples for KIM and other heavy mineral analysis were shipped to Overburden Drilling Management Limited (ODM; Ottawa, ON) for processing using established GSC protocols (Fig. 4; Appendix 1, 3; Plouffe et al., 2013; McClenaghan et al., 2020). Till samples were disaggregated and sieved at 2 mm (stream sediment samples had been pre-sieved in the field). The <2 mm fraction was passed across a shaker table to pre-concentrate the heavy fraction. The pre-concentrate was then micropanned to recover fine-grained gold, sulphide and other indicator minerals. These micropanned grains were recorded and described and then returned to the pre-concentrates which were then sieved at 0.25 mm. The remaining 0.25-2.0 mm concentrate was then refined using a heavy liquid separation of methylene iodide diluted with acetone to a specific gravity (SG) of 3.2 to produce a heavy mineral concentrate (HMC). Ferromagnetic minerals were removed from the heavy mineral concentrate using a hand magnetic. The non-ferromagnetic heavy mineral fraction was then sieved into <0.25 (archived), 0.25-0.5, 0.5-1.0 and 1.0-2.0 mm size fractions. The 0.25-0.5 mm HMC was subjected to paramagnetic separations to assist counting and picking of indicator minerals in this fine fraction. The 0.25-0.5 mm fraction was also cleaned with oxalic acid to remove oxidation stains as a further aid to mineral identification and picking. Visual identification of a limited number of KIMs was verified by ODM using an EDS-equipped SEM. Sample metadata and processing weights of the selected size fractions are listed in Appendix 2.

Quality-control checks on sample processing, and tests of sample medium homogeneity, included the use of GSC sample blanks, and field duplicates. We employed 11 Bathurst blanks (a weathered Silurian-Devonian granite (grus) resembling a well-sorted beach sand; collected from the Miramichi Highlands in New Brunswick), 3 Linton till blanks (a Canadian Shield-derived till collected from a borrow pit at Almonte, ON; Plouffe et al., 2013), and 8 blind (unknown to the lab) field duplicates (6 till, 2 stream sediment samples). Sample blanks and field duplicates are listed along with their results in Appendix 2. Sample processing order at ODM was predetermined by the GSC (Appendix 2A) in order to intermix blanks and duplicates within the sample batches. Processing order also ensured that samples with anticipated higher concentrations of sulphide minerals (another focus of this study – cf. Day et al., in progress; Paulen et al., in progress) were processed at the end of a batch, and/or had blanks inserted before/after specific samples to test for potential carry-over between samples. Raw, unedited data reported by ODM are listed in Appendix 3A-3E.

2.4 Mineral Chemistry

All KIM grains picked and removed from the samples were mounted, polished, and then analyzed to determine major and minor elements by electron probe microanalysis (EPMA) at the University of Alberta's Arctic Resources Geochemistry Laboratory. EPMA data were acquired for 10 elements (Ti, Na, K, Si, Fe, Cr, Mg, Ca, Al, Mn, O) with a JEOL JXA-8900R electron microprobe, fitted with 5 wavelength-dispersive spectrometers. Wavelength dispersive spectroscopy (WDS) was employed at a 40° takeoff angle, operating at a 20 kV accelerating voltage, a 20 nA probe current, a 2 µm beam diameter, and count times on peaks and backgrounds ranged from 20 to 60 s. A variety of natural minerals (silicate and oxide) were used for standardization. Data reduction used Probe for EPMA software (Donovan et al., 2015), while the X-ray intensity data were reduced based on the ϕ (ρZ) method of Armstrong (1995). EPMA geochemistry was used to test the identity of visually-picked (and in some cases SEM-based EDS) KIMs, and when required, eliminate or reclassify individual grains.

Trace and rare earth element (REE) concentrations in all garnet and Mg-ilmenite grains were determined by laser ablation inductively coupled mass spectrometry (LA-ICP-MS) at the University of Alberta's Arctic Resources Geochemistry Laboratory. Analyses were conducted using a Resonetics Resolution LR50 193 nm laser coupled to a ThermoScientific Element 2XR ICP-MS. The mass spectrometer was operated in low mass resolution mode (M/ΔM=ca. 300) with a power setting of ~1300 W and a torch depth of ~3.6 mm. Spot size ablation craters of 60-130 µm were used depending on the mineral grain type, and data were acquired using the rapid peak-hopping multichannel mode of the ICP-MS (cf., Poitras et al., 2018). Data were reduced offline using Iolite v3.32 software (Woodhead et al., 2007; Paton et al., 2011). Concentrations were calibrated with reference to the NIST SRM 612 glass standard and internal mineral-based secondary standards (cf., Liu et al., 2018; Poitras et al., 2018). Garnet trace element concentrations were normalized to ^{43}Ca.

3.0 RESULTS

3.1 KIM Sample Processing Quality Assurance – Quality Control

Linton blank sample 17-SUV-001 was the only one of the 14 blanks employed in this study that was found to contain KIMs (Appendix 2D). The single olivine grain recovered from this sample is within acceptable concentrations of olivine that have been previously recovered from the Linton blank (Plouffe et al., 2013). Thus, we conclude that no KIM carry-over between successively tabled samples, or laboratory contamination was detected from the blanks in this study.

Also, within an acceptable level of uncertainty, the results for the 8 field duplicates confirm precise detection of KIMs by ODM (Appendix 2D). Five of the six till duplicate sample pairs yielded zero KIMs in each sample pair. One of the till duplicate pairs (17-PTA-048, -049) was found to contain 2 KIMs, while the other contained zero. At such low sample KIM concentrations, it is possible that inherent KIM-content variability within the sample medium at any one location is enough to explain this difference.

Stream sediment field duplicates 095G-2017-1004 and -1005 each contained 6 garnets, while 095G-2017-1005 also contained 1 Mg-ilmenite – these results are considered equivalent. We also note that till sample 17-PTA-055 was collected from the stream cut-bank, right beside where stream sediment samples 095G-2017-1004, -1005 were collected, and recovered zero KIMs (Appendix 2). In the other stream sediment duplicate pair (085F-2019-1004, -1005), there is a significant difference between the two samples; 085F-2019-1004 reported zero KIMs, while 085F-2019-1005 contained 6 KIMs (3 garnet, 2 Mg-ilmenite, 1 forsterite; Appendix 2). In light of the similarity of all other results, it is assumed that the difference in KIM content between these two samples represents inherent variability in the original sample medium, or differences between sample locations within the same stream depositional environment that the material was collected.

3.2 Kimberlite Indicator Mineral – Abundances

For archival purposes, the original unaltered data files reported by ODM are reported in Appendix 3 (separated by years and geologists (Day, Paulen (PTA), and Smith (SUV)); Appendices 3A, 3B, 3C, 3D, 3E). Of the 270 samples processed by ODM, 14 were blanks, leaving 256 samples (196 till and 60 stream sediment) collected over the 3 years of fieldwork (Appendix 2). A total of 183 KIM grains were recovered from the sediment samples (excluding the 1 forsterite grain recovered in Linton blank 17-SUV-001). Only 1 KIM was found in the 1.0-2.0 mm size fraction, 18 in the 0.5-1.0 mm fraction, and 164 (~90%) in the 0.25-0.5 mm fraction (Appendix 2D). In terms of KIM types and abundances, ODM identified and picked 120 Cr-pyrope garnets, 0 eclogitic garnets, 6 Cr-diopsides, 23 Mg-ilmenites, 31 chromites, and 2 forsterite grains. ODM also picked 27 grains that they classified as "crustal ilmenite." Following EPMA analysis (Appendix 4; discussed below), removing the 1 forsterite grain from Linton blank sample 17-SUV-001, and adding the Cr-diopside from an interlocked grain that was originally listed only as 1 Cr-pyrope (18-PTA-011), these totals were revised to 120 garnets, 7 Cr-diopsides, 24 Mg-ilmenites, 30 chromites, and 2 forsterite grains (Appendix 2D). KIMs were present in 17.9% of the till samples, and 43.3% of the stream sediment samples (Appendix 2D).

Stream sediment samples were largely collected from the same areas that till samples were (acknowledging an absence of suitable stream environments through much of central NTS 85C and 85F; Fig. 5). Catchment weathering and erosion of a regional till cover is considered the principal source material for stream sediments, and therefore, we would assume in the absence of differences in weathering (including mechanical breakdown) and erosion, that the overall percentages of each KIM mineral type should be the same in both sample media. This does not appear to be the case. Instead, there are double the number of Cr-pyrope garnets recovered from the stream sediment samples, and 1.7 and 3.3 times more Mg-ilmenite and chromite grains, respectively, recovered from the till samples. Forsteritic olivine (n-2) and Cr-diopside (n=7) grain numbers are approximately equal, but grain abundances are considered too low to allow meaningful comparisons.

While simple presence/absence of KIMs is instructive for glacial dispersal studies (e.g. Fig. 1), the presence of multiple KIM mineral types in the same sample is considered greater evidence of a kimberlitic source, as opposed to potentially being derived from other ultramafic rocks. Ten of the 35 till samples containing KIMs had 2 or more KIM mineral types (29%), while 8 of the 26 stream sediment samples containing KIMs had 2 or more KIM mineral types (31%; Appendix 2D).

The use of normalized abundance data is important when comparing different sample types, sample masses, sampling programs, and for removing bias for natural stream density-concentrating mechanisms (cf. McClenaghan et al., 2002, 2020; Prior et al., 2009; McClenaghan, 2011; Day et al., 2018). The KIM data was first normalized to a 20 kg table-feed (<2 mm) weight (Fig. 5a; Appendix 2E), yielding KIM abundances of 66% garnet, 15.8% chromite, 13.4% Mg-ilmenite, 3.9% Cr-diopside, and 0.9% forsterite (Appendix 2E). The average normalization factor for converting the till sample table-feed weights was 0.94, while for the stream sediments, it was 1.46. The KIM compositional makeup was largely the same in the normalized 20 kg table-feed (<2 mm) sample weights, as it was in the raw sample data (Appendix 2D, 2E). Relative abundances of different KIM grain mineral types are similar to the original values (Appendix 2D, 2E), although Cr-pyrope garnets are almost 3x more abundant in the stream sediments, Mg-ilmenite abundances are only slightly elevated in the till samples, and chromites are 2.5x more abundant in the till samples.

KIM abundance data was also normalized to a "50 g non-ferromagnetic 0.25-0.5 mm HMC picking fraction weight" (abbreviated hereafter to "50 g HMC weight"; Day et al., 2018; Appendix 2F). The average normalization factor for converting the till sample 50 g HMC weights is 7.28, while for the stream sediments, it was 1.0 (Appendix 2F). The 50 g HMC normalization changed KIM compositional abundances more substantively (recognizing that this only pertains to KIMs within the 0.25-0.50 mm size fraction), yielding 57.0% garnet, 22.1% Mg-ilmenite, 15.8% chromite, 4% Cr-diopside, and 1.1% forsterite (Appendix 2F). This may suggest that locally, garnets are more likely to survive post-depositional weathering and erosion than Mg-ilmenites. Plotting the KIM normalized abundance data using both the 20 kg table-feed (Fig. 5a) and 50 g HMC weight (Fig. 5b) reveals, perhaps, important differences. Clearly, the two sample media (till versus stream sediments) cannot be considered the same, although it is the catchment weathering and erosion of a regional till blanket that is the main source of the stream sediments themselves. The 20 kg <2 mm table-feed normalized data tends to over-emphasize the KIM abundance in the stream sediment samples when compared directly to the till samples, while the opposite occurs when data is normalized to the 50 g HMC fraction. Consider stream sediment sample 085G-2018-1003 (n=49.1 KIMs) and till sample 18-PTA-011 (n=19.4 KIMs; Fig. 5a) versus n=98.1 and 202.4, respectively, for the same samples in the 50 g HMC data (Fig. 5b).

Overall, there appears to be a southwest declining vector of KIM abundances in both till and stream sediment samples from northern NTS 85G into northern and central NTS 95F, roughly aligned with the orientation (~250°) of prominent glacially fluted terrain in these map areas (Fig. 5). The Drybones Bay and Mud Lake kimberlites are situated on the eastern shore of Great Slave Lake, only 50-100 km northeast of the two highest abundance KIM samples (Fig. 3). Slightly elevated abundances north and south of Kakisa may reflect stream concentration factors, as these are not generally reflected in surrounding till samples. Similarly, samples to the west along streams feeding into the Mackenzie River appear to be simple inflations of low background till KIM concentrations of principally garnet. Most of the area surveyed is largely barren of KIMs suggesting an absence of evidence for the potential presence of unknown local kimberlites. Elevated KIM abundances and an increase in chromites along the Liard River samples may reflect dispersal from the Fort Simpson area HOAM kimberlites, or from other unknown kimberlites detected by the Trout Lake survey samples.

3.3 KIM Chemistry

EPMA chemistry data of the KIMs picked by ODM are presented in Appendix 4, separated into individual worksheets by mineral type (Appendix 4A to 4E). Each KIM grain was analyzed twice (3 times for grains recovered from samples collected in 2019), and both the individual spot measurements, and average values are presented (note, average values are used in all subsequent geochemistry plots). For samples that contained more than one grain of the same mineral, these grains were labelled in the Appendices with the same sample number and then "Mount #" - "Row #" - "Grain #" were used to indicate individual grains, e.g. sample 17-PTA-045-1-5-1 (Appendix 4, 5). There were 6 KIM grains lost during the grain mounting and polishing processes; 3 garnets, 2 chromites, and 1 Cr-diopside; unfortunately, this included a rather spectacular interlocked pyrope garnet – Cr-diopside grain shown in Figure 6. One of the KIMs picked and classified by ODM as a chromite was reclassified by EPMA as a "crustal" ilmenite (Appendix 4C); two picked "crustal ilmenites" were reclassified as Mg-ilmenites; one picked Mg-ilmenite was reclassified as a "crustal" ilmenite; all other ODM visual classifications were confirmed by EPMA to determine their mineral chemistry. KIM grains were further classified according to various chemical criteria listed (e.g. "crustal" ilmenite vs Mg-ilmenite; Appendix 4).

LA-ICP-MS trace element and rare earth element chemistry data of the KIMs are presented in Appendix 5, separated by mineral type into individual worksheets (Appendix 5A to 5G). As with the EPMA chemistry data, each KIM grain was analyzed twice (3 times for the grains recovered from samples collected in 2019), and both the individual spot measurements and averaged values are presented (only the averaged values are used in subsequent geochemistry plots). One additional garnet was lost from the mounts during polishing and re-coating in preparation for LA-ICP-MS analysis; a G10 garnet from sample 18-PTA-011 (grain # 1-3-1; Appendix 5A).

3.3.1 Garnet

Of the 117 garnets picked and analyzed by EPMA (excluding the three lost during mounting/polishing), all were Cr-pyropes (wt% $Cr_2O_3 \geq 1.0$; Appendix 4A). These Cr-pyropes were further classified according to Grütter et al.'s (2004) scheme: 65 lherzolitic G9 garnets, 3 hartzburgitic G10 garnets, 6 high TiO_2 peridotitic G11 garnets, and 43 wehrlitic G12 garnets (Fig. 7; Appendix 4A). The absence of eclogitic garnets (wt% $Cr_2O_3<1.0$) may reflect the lower total number of garnets recovered in this study, but also mirrors low eclogitic garnet abundances in the other regional KIM surveys (Day et al., 2005, 2007; Mills, 2008; Pitman, 2014; Sheng, 2016; Poitras et al., 2018). While two of the three G10s were recovered from northeastern NTS 85G, there does not appear to be any obvious geographic sorting of the different Cr-pyrope types.

Trace and rare earth element (REE) chemistry are considered to provide sensitive tracers of metasomatic process and have been increasingly used to characterize and qualify populations of Cr-pyropes and other minerals (cf. Stachel et al., 1988, 2004; Griffin et al., 1996; Cox and Barnes, 2005; Banas et al., 2009; Viljoen et al., 2014; Hardman et al., 2018; Liu et al., 2018; Poitras et al., 2018). Analytical results for grains recovered from till and stream sediment samples are presented in Appendix 5, separated by mineral grain type (e.g. garnets – Appendix 5A-C). Results include individual spot determinations (Appendix 5A), averages of the two spot measurements for each grain (Appendix 5B), and elemental concentrations normalized to the C1 carbonaceous chondrite values of McDonough and Sun (1995; Appendix 5C). The chondrite normalized values for element abundances in samples from the main study area (i.e. excluding the 18 Cr-pyropes from NTS 95B and G, down ice flow of the HOAM kimberlite field) were plotted in order of increasing atomic number (La through Lu) and then visually grouped according to different plot shapes. The first of these are **heavy REE enriched** (HREE-enriched; Fig. 8a, b, c; n=41), subdivided into **HREE-enriched – high middle REE** (MREE; at or above 10x chondrite levels from samarium (Sm) through lutetium (Lu)), **HREE-enriched – lower MREE** (generally <10x chondrite levels), and **Sinusoidal – MREE-enriched** (peak abundances around 10x chondrite levels between praseodymium (Pr) through terbium (Tb; Fig. 8c). The HREE-enriched populations include 72% of all G9 garnets, 100% of G10s and G11s, and 4% of G12s. Twenty-two of the Cr-pyropes have a **Sinusoidal – REE-normal** distribution exhibiting only slight enrichment/depletion (Fig. 8d; 9 – G9 (21%); 13 – G12 (30%)). The final grouping are those labeled **Sinusoidal – MREE-depleted** (Fig. 8e) and **LREE/MREE-depleted** (Fig. 8f); these account for 7% of the G9s and 66% of the G12s. When grouped simply by the Grütter et al. (2004) garnet classification (Fig. 9), plots of the G9 and G12s exhibit a wide range of chondrite normalized REE abundances. In general, though, the G9, G10 and G11 garnets exhibit HREE and MREE enriched patterns (and lower lanthanum (La) abundance), while the G12s have conspicuously more MREE-depleted patterns, and often higher LREE (La, cerium (Ce), Pr; Fig. 9).

Metasomatic depletion and enrichment of REE in Cr-pyropes is also reflected in yttrium (Y) and zirconium (Zr) concentrations (Fig. 10; Griffin and Ryan, 1995; Griffin et al. 1999a; Banas et al., 2009; Smit et al., 2014; Stachel et al., 2018). The two G10s from this study have high concentrations of both Zr and Y, and trend along the high temperature melt – metasomatism line of Griffin et al. (1999a), as do the lower concentration G11s (Fig. 10). With the G9s, 44% plot within the depleted zone, the rest exhibit a wide scatter, with points trending along all three metasomatic trends (Fig. 10). The G12s almost universally (98%) plot within the depleted category, 93% within a tight cluster near the origin at very low Zr and Y abundances (Fig. 10). The spread of higher Y and Zr abundances follows the trend of grain core to rim characteristics that Griffin et al. (1999a) considered reflective of zoning and secondary replacement rims.

3.3.2 Mg-Ilmenite

ODM visually identified and picked 49 ilmenite grains during the sample processing, 23 of which they categorized as Mg-ilmenite, and the rest were considered "crustal ilmenite". Chemical analysis by EPMA of all the ilmenite grains picked by ODM reclassified two "crustal ilmenites" as Mg-ilmenites, one picked Mg-ilmenite as a "crustal" ilmenite, and 1 chromite as a "crustal" ilmenite ("crustal" ilmenite < 4 wt% MgO < Mg-ilmenite; Appendix 4B). Standard MgO – Cr_2O_3 and MgO – TiO_2 wt% plots of the EPMA chemistry data are used to discriminate non-kimberlitic from kimberlitic Mg-ilmenites from this study, and to compare them with other regional survey Mg-ilmenite data (Fig. 11; Appendix 4B). Using the MgO – TiO_2 classification of Wyatt et al. (2004), 24 are considered "kimberlitic" and 26 are "non-kimberlitic." Magnesium, chrome, and titanium concentrations in the kimberlitic Mg-ilmenites tends to be fairly low (~6, 0.75, and 44 wt%, respectively; Fig. 11).

The trace element and REE concentrations in the Mg- and crustal ilmenite grains are shown in Fig. 12 and reported in Appendix 5D. In order to facilitate comparisons with published studies, element concentrations are normalized to both the C1 carbonaceous chondrite (C1; McDonough and Sun, 1995) and Primitive Mantle (PM – silicate earth; Sun and McDonough, 1989; Fig. 12). The Mg-ilmenites display a wide range of concentrations, particularly within the LREE, and are largely well below C1/PM concentrations. A plot of the average concentration display a shallow concave MREE-depleted pattern, however median values better reflect 14 of the 24 grains analyzed, and display a strongly linearly increasing trend from very low LREE to higher HREE. The crustal ilmenite samples are most strikingly distinguished from the Mg-ilmenites by the pronounced increase in HREE, particularly dysprosium (Dy) through Lu, where concentrations of many samples exceed C1 levels (Fig. 12). This trend is strongest in 16 of the 26 crustal ilmenite samples, but even the other 10 display the same general HREE-upward trend, and both the average and median values closely mirror each other. Where Ashchepkov et al. (2017) studying ilmenite megacrysts from the Russian Dalnyaya kimberlite noted a positive inflection in europium (Eu) and gadolinium (Gd), these are barely discernible in the southern Mackenzie samples, and indeed, there is a more prominent negative inflection in the crustal ilmenite median Eu concentrations (Fig. 12). For comparative purposes, spider diagrams of elemental concentrations normalized to PM (Sun and McDonough, 1989) were also constructed for the ilmenites (Fig. 12), although several of the elements used by Ashchepkov et al. (2017) were not determined by this study (e.g. Cs, Rb, Th, U, Ta, Pb). Peak and average hafnium (Hf) and Zr concentrations for the Mg-ilmenites and crustal ilmenites are comparable and well above PM levels (~100), however, the Mg-ilmenite grains exhibit much tighter clustering of concentrations and the crustal ilmenites range down to PM levels.

Niobium (Nb) concentrations in the Mg-ilmenites tend to be much higher than the crustal ilmenite samples, and both are highly enriched (~100-10 000), while barium (Ba) levels are similarly low (~0.1-0.01). The Mg-ilmenites have far lower Lu/Hf ratios than the crustal ilmenites (Fig. 12; Appendix 5D), possibly because of equilibration with another high Lu phase such as garnet (if the Mg-ilmenites are megacrystic). Crustal ilmenites from basalt would not have equilibrated with garnet.

3.3.3 Spinel – Chromite

There were 30 chromites recovered by ODM (3 were lost during mounting/polishing; 1 was reclassified by EPMA as a "crustal" ilmenite). Chromites were subdivided based on the chemical criteria employed by Poitras et al. (2018):

$$\text{spinel} < Cr_2O_3/Al_2O_3 = 1.5 < \text{Cr-spinel}$$
$$\text{chromite} < 11 \text{ wt\% MgO} < \text{magnesio-chromite}$$
$$\text{chromite} < 3 \text{ wt\% } TiO_2 < \text{Ti-chromite}$$

This classification yielded 16 spinel; 2 chromite, and 8 magnesio-chromite (Appendix 4C). Mantle-derived chromites exhibit a wide range of major and minor element concentrations. Several geochemical indices have been developed to discriminate different chromite populations; Cr-spinel with >61 wt% Cr_2O_3, 10-16 wt% MgO, <0.50 wt% TiO_2, <8 wt% Al_2O_3, and <6 wt% Fe_2O_3 have been correlated with diamond chromite inclusion compositions (Fig. 13a-d; Sobolev, 1977; Fipke et al., 1989, 1995; Griffin et al., 1997; McClenaghan and Kjarsgaard, 2007; Roeder and Schulze, 2008). None of the Cr-spinels from this study plot within what are considered diamond-inclusion, or diamond intergrowth fields. The spinel Mg# - Cr# plot (Fig. 13d; after Roeder and Schulze, 2008) suggests all but 2 of the Cr-spinels plot within the bounds of values for cratonic peridotites. The Cr_2O_3 – TiO_2 and Cr_2O_3 – Al_2O_3 indices of Sobolev (1971, 1977; see also McClenaghan and Kjarsgaard, 2007; Fig. 13b, c) similarly indicate an absence of chromites from diamond inclusion fields, and mixed population of potentially kimberlitic and non-diatreme derived grains. Trace and REE data are listed in Appendix 5E.

3.3.4 Olivine

Only 3 olivine (forsterite) grains were recovered from the KIM samples, and one of these was in a GSC blank, and is considered background and therefore disregarded. Both of the two remaining olivine grains have Mg#'s of 91 and 92 (Appendix 4D), but neither contain any detectable Cr_2O_3, placing them outside Fipke et al.'s (1995) Mg# - Cr_2O_3 wt% diamond inclusion field (Fig. 14). All of the previous regional KIM surveys also report low abundances and occurrences of olivine (Kerr et al., 2000; Day et al., 2005, 2007; Mills, 2008; Pitman, 2014; Poitras et al., 2018). Locally, the Drybones Bay and Mud Lake kimberlites are reported to contain abundant serpentinized olivine and carbonates (after olivine and ilmenite; Kretschmar, 1997; Carbno and Canil, 2002; Sheng, 2016). If these 2 kimberlites were the contributing sources to the Southern Mackenzie survey KIMs, it may explain the paucity of recovered olivine grains. At depth, Kretschmar (1997) did report olivine to be the most abundant macrocryst in the Drybones Bay kimberlite, however, the distinction is made between relatively fresh material from deeper diatreme facies from the uppermost volcaniclastic tuff strata (which would have been actively eroded by glaciers) in which olivine serpentinization is complete. Trace and REE data are listed in Appendix 5F.

3.3.5 Clinopyroxene

There were 19 clinopyroxene (CPX) grains picked by ODM from the Southern Mackenzie samples, 17 of which were from the 0.25-0.5 mm size fraction, and the other 2 from the 0.5-1.0 mm size fraction; all but 2 of the grains came from stream sediment samples (Appendix 4E). The picked CPX grains included 1 diopside (0 wt% Cr_2O_3), 13 low-Cr diopsides (<1 wt% Cr_2O_3), 4 Cr-diopsides (>1% Cr_2O_3; <1.4 wt% Cr_2O_3), and 1 high-Cr diopside (>1.4 wt% Cr_2O_3). The clinopyroxene grains exhibit a wide range of Mg# values (0.82-0.94), and concentrations of Cr_2O_3 up to 1.45 wt% (Appendix 4E). Cr-rich diopsides with >1.5 wt% Cr_2O_3 are only associated with kimberlites and mantle xenoliths (Deer et al, 1982, Fipke et al., 1989, 1995; McClenaghan and Kjarsgaard, 2007), and may include the high-Cr diopside from sample 17-SUV-109 (Appendix 4E). Kimberlites also contain clinopyroxene with <1.5 wt% Cr_2O_3, and several discrimination plots have been utilized to characterize these (cf. Schulze, 1987; McCandless and Gurney, 1989; Nimis, 1998; Morris et al., 2002). Using the classification scheme of McClenaghan and Kjarsgaard (2006), nine of the grains plot within the bounds of peridotite mantle assemblages, while the other 10 are classified as non-peridotitic, plotting within megacrystic and websteritic fields (Fig. 15). Trace and REE data are listed in Appendix 5G.

4 DISCUSSION

4.1 Geothermometry

Ni-in-garnet geothermometry calculations based on the methodologies of Ryan et al. (1996), Griffin et al. (1989) and Canil (1999) are provided in Appendix 5A and 6, and illustrated in Figure 16 (using an average of the Griffin et al. (1989) and Canil (1999) measures). The Ni-in-garnet geothermometer is based on the strong temperature dependence of the partitioning of Ni between garnet and olivine. It is used to model geothermometers in P (kbar) and conversions to depth (km) of projections of modelled geotherms, and provides a means of reconstructing the thermal state of the lithosphere from which the Cr-pyropes were derived, including within the context of the 900-1200°C diamond stability window (Ryan et al., 1996). When plotted against Carbno and Canil's (2002) average Slave Craton mantle xenolith geotherm (after Boyd and Canil, 1997; Kopylova et al., 1999; MacKenzie and Canil, 1999), 42 of 44 G12 garnets, and 29 of 63 G9 garnets (within the 600-1200°C geotherm) plot above their graphite – diamond transition curve (Fig. 17). While Carbno and Canil (2002) note that geotherms based on xenolith thermobarometry vary between Slave Craton localities (cf. Kopylova et al., 1999; Pearson et al., 1998; MacKenzie and Canil, 1999), Poitras et al. (2018) illustrate a strong similarity between central Slave Craton and Horn Plateau Cr-diopside geotherms. Based on the Ni-in-garnet distribution, it appears that half of the G9s, two of the G11s, and the two G10s from sediment samples in this study display potential of being derived from diamondiferous kimberlitic magma. Most of the G12s appear to come from kimberlite intruded in a zone with too cool a geotherm, and the G11s are largely from too hot a geotherm (i.e. >1200°C, Fig. 16).

The REE data for Cr-pyrope garnets was also plotted according to Ni-in-garnet geothermometry groupings (using average values calculated from the Griffin et al. (1989) and Canil (1999) methodologies; Fig. 18; Appendix 5A). The 800-900°C Ni-in-garnet temperature group contains mostly (82%) wehrlitic G12 Cr-pyropes, with pronounced MREE-depleted traces. The 900-1000°C Ni-in-garnet geothermometry group contains a 3:2 ratio of G12 to G9 Cr-pyropes. The G12 traces within this temperature group are largely MREE-depleted and Sinusoidal: REE-normal, while the G9s trend from Sinusoidal: REE-normal to Sinusoidal: MREE-enriched (n=3) and HREE-enriched (n=3). The wider spread of the G9 and G12 REE compositions is reflected by the MREE departure between the trace of the average values for this group and that of the median (Fig. 18). As Ni-in-garnet temperatures rise, the groups increasingly display MREE and HREE-enriched trends, including the final grouping (>1200°C) which contains 4 of 6 peridotitic G11 garnets (Figs. 16, 18; Appendix 5A).

4.2 Indicator Minerals

Plots of Cr-pyrope CaO vs Cr_2O_3 wt% for grains from this study compared to plots for other regional KIM surveys and kimberlite studies reveal a striking similarity to those of the Drybones Bay kimberlite, and a marked dissimilarity with the regionally adjacent Horn Plateau and Trout Lake KIM surveys (Fig. 7). Comparisons with the Mud Lake kimberlite are limited by the few samples of it that have been analyzed (cf. Kerr et al., 2000; Sheng, 2016). Our KIM samples appear to have an unusual abundance of wehrlitic (G12) garnets (37%), suggesting these, at least, were derived from a kimberlite with low diamond potential, as G12s are rarely found within diamond inclusions (Gurney, 1984; Grütter et al., 2004). The Y versus Zr plot of Cr-pyropes (Fig. 10) similarly depicts an ultra-depleted magma source for almost all the G12s, and many of the G9 garnets, possibly from a location with a more elevated geotherm. These data, along with the cool Ni-in-garnet temperatures of the G12s (Fig. 16), are similar to the shallow layer (120 – 150 km) chemistry of the mantle beneath the central Slave Craton kimberlites (cf. Griffin et al., 1999b; Menzies et al., 2004; Poitras et al., 2018).

We see the same chemical separation between populations of ilmenite grains recovered from this study and from Drybones Bay/Mud Lake kimberlites, compared with those of the Horn Plateau and Trout Lake surveys, and West and Central Slave Craton kimberlites (Fig. 11). Ilmenites from this study (like the Drybones Bay kimberlite) are characterized by low Cr_2O_3 and MgO concentrations, distinguishing them from the generally higher MgO concentrations reported from West and Central Slave kimberlite samples, and the overall higher, and much wider spread of the Horn Plateau sample ilmenites (Fig. 11a). Similar distinctions are seen with the TiO_2 concentrations, with generally lower values in this study's ilmenites, as well as those of the Drybones Bay samples (Fig. 11b). The majority of this study's ilmenites (82%) have Zr/Nb concentrations >0.37 (Appendix 5), which Carmody et al. (2014) used to characterize those with higher diamond potential. However, Castillo-Oliver et al. (2017), based on studies of kimberlitic ilmenites, including those found as inclusions in diamonds, indicate that complex ilmenite petrogenesis does not support the Zr/Nb model for diagnosing diamond potential of a kimberlite pipe. Low MgO and inferred high Fe_2O_3 concentrations of this study's ilmenites (indicative of oxidizing conditions; Fig. 11b), also suggests that diamonds within the kimberlite that these were derived from, may not have survived transport to the surface.

Few chromite grains were recovered from sediment samples in this study (n=10), or were reported in published Drybones Bay/Mud Lake and Trout Lake studies (Fig. 13; Kerr et al., 2000; Pronk, 2008; Sheng, 2016). Using various chemical discriminating plots (Fig. 13), this study's chromites appear dissimilar from those of Drybones Bay/Mud Lake, often with lower Cr_2O_3 contents. This may suggest that chromite grains are representative of further-travelled KIMs. The lower Cr contents are in keeping with the scarcity of high-Cr G10 garnets and indicates a modest level of mantle depletion compared to the shallow “ultra-depleted” layer of the central Slave Craton, so transport from kimberlites located there seems unlikely.

The few clinopyroxene grains recovered in this study provide little opportunity to assess chemical trends in comparison to regional studies with greater numbers of samples (Fig. 15). Only 5 Cr-diopside grains (including 1 high-Cr diopside; >1.5wt% Cr_2O_3) were found, and these generally plot within the same broad peridotitic fields as the central and southeast Slave Craton and Horn Plateau samples do (Fig. 15). They broadly overlap 3 of the 7 Drybones Bay clinopyroxene grains, but only those with lower Cr_2O_3 concentrations (Fig. 15).

4.3 Diamond Potential

There appears to be little evidence of undiscovered kimberlite deposits within the core field area of this study (NTS 85B, C, F, G). Only 183 KIMs were recovered in 61 of the total 256 samples collected, representing positive results in only 17.9% of the till samples and 43.3% of the stream sediment samples (23.8% of all samples; Appendix 2D). By comparison, Watson’s (2011b) collection of till and a variety of other sediment sample types from the Ka’a’gee Tu candidate protected area (Fig. 3; largely using a Dutch auger, as opposed to this study’s mostly shovel-dug holes), recovered KIMs in only 6 of 93 samples (6% recovery; 5 samples had only single grains, 1 sample had 2 grains). KIM recovery was much greater in Watson’s (2011a) Sambaa K’e Candidate Protected Area survey (Fig. 3), where 67 of 163 samples (41%) contained KIMs (albeit, most samples contained only a single KIM grain). In stream sediment samples in the Horn Plateau region (Day et al., 2005, 2007; Fig. 3), >3300 KIMs were recovered from 324 bulk stream sediment samples (94% of samples contained KIMs). These included a particularly high abundance of pyrope garnets with a high proportion of G10, high-Cr, and low-Ca garnets. Even in areas potentially down-ice flow from known HOAM kimberlites south of Fort Simpson (where ice flow would have been largely west-southwest; Dyke and Prest, 1987; Bednarski, 2008), we see only modest to low KIM abundances in 11 of the 18 samples collected (Fig. 5; Appendix 2).

The source of the abundant KIMs in the Horn Plateau surveys has not been linked to any actual kimberlite(s) occurrence. Furthermore, these KIMs are unlikely to relate to the HOAM kimberlites (estimated to be Early to Late Devonian; Poitras et al., 2018) south of Fort Simpson as the HOAM kimberlites lie 30 to >100 km down-ice flow of the Horn Plateau. Instead, the abundant Horn Plateau KIMs are most likely related to glacial dispersal from unknown (possibly proximal) kimberlite source(s), as opposed to being from reworked KIMs hosted in the Cretaceous bedrock of the Horn Plateau itself (Day et al., 2005, 2007; Huntley et al., 2008; Mills et al., 2008). Poitras et al. (2018, and their Supplemental Material 4) have speculated that Horn Plateau KIMs could, in part, relate to Phanerozoic fluvial erosion of southeast or western Cambro-Ordovician Slave Craton kimberlites (>275-450 km to the east of Horn Plateau; Fig. 3), particularly in the context of the largely absent Paleozoic cover from the Slave region (Patchett et al., 1999; Ault et al., 2009, 2013). However, from Late Jurassic through Pleistocene times, former westerly transport directions shifted to eastward-directed fluvial systems originating from the rising Cordillera (Duk-Rodkin and Hughes, 1994; Duk-Rodkin and Lemmen, 2000). Poitras et al. (2018) further suggest that a combined glacial and fluvial mechanism could also have transported western Slave Craton kimberlite material >275 km to the west or southwest, even though typical maximum glacial transport distances are considered to be far less (<100 km; McClenaghan et al., 2002; Armstrong and Kjarsgaard, 2003; McClenaghan, 2005; Armstrong, 2009).

Our KIM results, and the prevalence of samples with none or very low abundances (Fig. 5; Appendix 2D), indicate that there is no evidence of glacial dispersal of kimberlitic material either from, or across the area to the southeast of Horn Plateau (NTS 85B, C, F, G). Glacial ice flow reconstructions for the last (Late Wisconsinan) glaciation, based on streamlined bedforms, bedrock striae, and till clast fabrics indicate an early northwest flow across NTS 85F and 85G map areas (extending down-flow to an encircling and over-riding flow across Horn Plateau). This northwest flow was then cross-cut by the regionally much more prominent southwest flow that is recorded in extensive mega-scale glacial lineation flowsets (Dyke and Prest, 1987; Margold et al., 2018; Paulen et al., 2019; Paulen and Smith, in press; Hagedorn et al., in prep). It is this southwest flow that most of our KIM till samples are likely to have been derived from.

5 CONCLUSION

The very low contents to absent KIMs in stream sediment and till samples in this Southern Mackenzie survey indicate that there are no local (unknown) kimberlite outcrops or subcrops within the terrestrial extents of our main field area (NTS 85B, C, F, G). Based on pronounced similarities in mineral chemistry, we suggest that the two Cr-pyrope and ilmenite rich samples (085G-2018-1003; 18-PTA-011) in northwest NTS 85G, and a discontinuous southwestward vector of low abundance Cr-pyropes across NTS 85F (Fig. 5), correspond to long distance glacial dispersal of kimberlitic material from the Drybones Bay and/or Mud Lake kimberlites (Fig. 3). It is also possible that the 2-3 kimberlite intrusive phases in the Drybones Bay kimberlite (Sheng, 2016), produced other proximal kimberlite occurrences that have yet to be detected (e.g. below Great Slave Lake). Glacial erosion and transport of KIMs from the Drybones Bay area (Kerr, 2006), would have been aligned with the most prominent southwest-oriented glacial flowsets crossing NTS 85F and G.

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