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# 1. Introduction

Historically, three industry-led diamond (kimberlite) exploration programs have been conducted on Banks Island, NWT: Monopros (1995-1997; northwest Banks Island, aeromagnetic surveys; Aerodat Inc. 1996, 1997; Wallace and Wood, 1997), Diamonds North Resources Limited (2004-2007; northeast Banks Island, aeromagnetic and stream sediment surveys; Mah, 2005; Kienlen and Vanderspiegel, 2006; Jober and Vanderspiegel, 2008), and Rio Tinto Exploration Canada Incorporated (2010-2011; northeast and southeast Banks Island and northwest Victoria Island, aeromagnetic and stream sediment surveys; Wilson et al. 2012a, b; Fig. 1). These studies by industry succeeded in recovering kimberlite indicator minerals (KIMs), and identifying a number of weak to moderate magnetic anomalies (<5.5 nT). However, they were unable to determine if any of the KIMs and magnetic anomalies related to a kimberlitic source(s) on Banks Island, or whether they represented redeposited KIMs from an extra-Banks Island source(s), and/or were derived from non-kimberlitic magnetic anomalies (note, no anomalies were drill-tested). Coincident with their exploration activities, Diamonds North and Rio Tinto registered numerous Prospecting Permits and Mineral Claims (Fig. 1).

At the time when Industry was active on Banks Island (1995-2008), understanding of the glacial history employed in reconstruction of potential glacial dispersal of KIMs held that during that last glaciation (Late Wisconsinan; ~25-10 ka BP), the continental Laurentide Ice Sheet inundated only the south, east and northern coastal regions, leaving the vast majority of Banks Island unglaciated (Dyke and Prest, 1987). Earlier glaciations were held to be more extensive, and there was a reported complex history of interglacials, transgressions by inland seas and proglacial lakes, and an area of northwestern Banks Island considered never to have been glaciated (Prest et al., 1968; Vincent, 1982, 1983, 1990; Barendregt and Vincent, 1990; Barendregt et al., 1998). Subsequent to industry's exploration activity on Banks Island, research by Dr. John England (University of Alberta) and several of his graduate students (England and Furze, 2008, 2011; England et al., 2009; Lakeman, 2012; Lakeman and England, 2012, 2013, 2014; Vaughan, 2014; Vaughan et al. 2014), and that of Dr. David Evans (Evans et al. 2014), fundamentally revised the glacial history of Banks Island. Collectively, they demonstrated that the entirety of Banks Island had been inundated by Laurentide ice during the Late Wisconsinan glaciation, and that significant differences in geomorphology and sedimentology across northern Banks Island, formerly attributed to past glaciations, often reflected the marked differences and temporal shifts in areas of cold-based vs. warm-based ice cover. The implications of this revised glacial history for devising drift sampling programs and interpretation of past KIM drift exploration results is potentially significant.

All formerly established mineral claims and prospecting permits on Banks Island have now expired, or lapsed. To address identified knowledge gaps and facilitate decisions by the mineral exploration industry to either renew diamond-bearing kimberlite exploration on Banks Island, or determine that this area is unlikely to host such mineral resources, and thus focus exploration activities elsewhere, the Geological Survey of Canada (GSC), as part of the Geo-mapping for Energy and Minerals (GEM-2) Program, conducted a targeted drift, stream sediment, and bedrock sampling program. Kimberlite indicator minerals recovered and analyzed from these surveys, along with geological assessments of various terrains and integration of updated glacial reconstructions, are used to produce new regional geoscience data and increase knowledge on the potential provenance of Banks Island KIMs.

## 1.1 Study Area

Banks Island is the fourth largest island (70 028 km$^2$) in the Canadian Arctic Archipelago. It is part of the Inuvialuit Settlement Region, and the hamlet of Sachs Harbour, located on southwest Banks Island, supports a population of 114 people (2019 NWT Bureau of Statistics). The physiography of Banks Island is characterized by dissected highlands in the north and south, between which lie areas of broad rolling lowlands and outwash plains (Fyles, 1962; Vincent, 1982, 1983). It lies within a region of continuous permafrost, and exhibits bioclimatic zones ranging from low to high-Arctic (reflecting a host of microclimates), including barren to lichen-covered polar desert bedrock uplands and relatively lush dwarf shrub, sedge and peat-covered meadows in extensive interior lowland regions (Maxwell, 1981; Ecosystem Classification Group, 2013).

## 1.2 Bedrock Geology

The bedrock geology of Banks Island has been described by Washburn (1947), Thorsteinsson and Tozer (1959, 1962), Hills (1969), Embry and Klovan (1971, 1974), and exhaustively by Miall (1975, 1976, 1979; Fig. 1). Additional insights come from nine wildcat wells drilled in the early 1970s (cf., Plauchut, 1971; Cassan and Evers, 1973; Jutard and Plauchut, 1973), and from recent structural geology studies (Piepjohn et al., 2018). Harrison et al.'s (2013, 2014) regional geological tectonic assemblage maps and GIS compilations have suggested revision of some former bedrock assemblages and naming conventions. As these have not been formally or conventionally adopted (in all cases), naming conventions employed here follow Miall (1975, 1979) and others, but suggested changes have been noted.

Cherty dolomites and quartzose sandstones of the Proterozoic Glenelg Formation (Nelson Head Formation of Harrison et al., 2013) are the oldest bedrock exposed on Banks Island and outcrop along the southeast coast (Fig. 1). Northeastern Banks Island is comprised of interbedded, cliff-forming Late Devonian pre-Mercy Bay Member (Weatherall Formation of Harrison et al., 2013) sandstone, siltstone and calcareous, siliceous and silty shales and minor coal, and Late Devonian Mercy Bay Member and younger (Parry Islands Formation of Harrison et al., 2013) fluvial and deltaic sandstones with minor siltstone, shale, and coal. The Mercy Bay Member sandstones are predominantly quartzose, containing rare plagioclase feldspar grains and sparse greenish biotite mica, clastic dolomite and clastic limonite. The Late Devonian shales are dark grey and carbonaceous to medium grey, silty, and micaceous with pyrite as a minor constituent. Rare, reddish brown shale is also found in the Devonian strata, as are ironstones (Miall, 1976). A prominent, 61 m thick limestone unit (Mercy Bay Member; Embry and Klovan, 1971) is exposed near the top of the coastal bluffs on northeastern Banks Island (Dm2 – Fig. 1).

More than 80% of the island is underlain by generally weak to poorly consolidated Mesozoic and Cenozoic strata (Fig. 1). Early Cretaceous Isachsen Formation is comprised of grey to yellow, fine to mainly coarse-grained sandstone and pebble-conglomeratic sandstone. This formation also includes thin laminae of coal, clay-ironstone and marcasite concretions (Thorsteinson and Tozer, 1962), and the upper parts of this unit are largely a poorly consolidated cross-bedded, white, quartzose sandstone with rare pebble beds (quartz, chert, silicified carbonate) with boulders up to 34 cm diameter, and wood fragments (Miall, 1979). Early Cretaceous Christopher Formation conformably overlies Isachsen beds, and is predominantly comprised of black to brown to grey shale and inter-bedded light grey, fine-grained sandstone. Red and yellow-weathering calcareous siltstones and mudstones, clay ironstones, marcasite concretions, and diagnostic calcite concretions (glendonites) are also found in the Christopher beds (Thorsteinson and Tozer, 1962; Miall, 1979). The Hassel Formation consists of sandstone with minor amounts of carbonaceous shale and coal; glauconitic sands are present at the base of the formation. The base of Late Cretaceous Kanguk Formation exposes a distinctive dark grey to black, bituminous, sulphurous shale member containing bentonitic clay layers and tuff beds (Miall, 1979). The rest of the Kanguk is comprised of shale with minor silty beds. The shale is in places burnt (bocannes), forming a brick-red shale and cindery material (klinker).

The lower unit of the Eureka Sound Formation (shale member; Eureka Sound-West assemblage, Strand Bay Formation of Harrison et al., 2013) is a Paleocene marine deposit characterized by light grey, fine-grained sandstone, coal, and carbonaceous shale. Above this, the deltaic Eureka Sound Formation (cyclic member; Eureka Sound-West, Iceberg Bay Formation of Harrison et al., 2013) is comprised of interbedded sand (unconsolidated), light grey shale, silty sand, silt, siltstone, sandstone, lignitic coal, soil beds, carbonaceous shale, and clay ironstone (Miall, 1979). The carbonaceous shale contains abundant plant stems and wide fronds, and in places carbonized fossil wood up to 60 cm in diameter. Eureka Sound Formation wood is reported to be diagenetically distinguished from the overlying Beaufort Formation wood by its more carbonized and compressed appearance (Thorsteinsson and Tozer, 1962). Sand beds in the upper cyclic member generally display a mottled “salt and pepper” coloration owing to the abundant dark chert grains (Miall, 1979). Exposures of this upper cyclic member in southern Banks Island are much coarser, and while unconsolidated fine to coarse-grained sand predominates, pebble and cobble conglomerate beds, containing clasts up to 12 cm occur; clast types include silicified carbonate sediments, quartzite, black, brown and grey chert, vein quartz, rare diabase, and rare pink granites (Miall, 1979). Thicknesses of the upper Eureka Sound cyclic member across Banks Island range from 60 – 939 m (Miall, 1979). The  Miocene Beaufort Formation (Beaufort assemblage, Ballast Brook and Beaufort formations of Harrison et al. (2013)) unconformably overlies the Eureka Sound Formation, and is characterized by a succession of unconsolidated sand, gravel, clay and peat, with abundant wood (Miall, 1979). The Beaufort Formation type section is found on Prince Patrick Island (Tozer, 1956; Fyles, 1990). On northwestern Banks Island, the Beaufort Formation is assigned a Middle to Late Miocene age (Hills and Fyles, 1973). Across much of central and northern Banks Island, the Beaufort Formation deposits occur only as a thin veneer capping an elevated, dissected plateau (Miall, 1979). Mapping of the Beaufort Formation contact is reported to be readily apparent over much of the western island as it imparts a darker colour and smoother texture on the surface than does the Eureka Sound Formation (Miall, 1979). Gravel in the Beaufort Formation is characterized by an abundance of orange and brown chert, which contrasts that of the Eureka Sound Formation which contains typically grey and black chert (Miall, 1979). The Beaufort Formation is noted to also contain rare granites (Miall, 1979). The Beaufort strata thicken westward, reaching 100 m at Ballast Brook (Hills, 1969) and possibly as much as 260 m in the Storkerson Bay A-15 well and 276 m in the Bar Harbour E-76 well (Fig. 1; Miall, 1979). Miall (1979) has also suggested that because deposits of the Beaufort Formation can be found far inland, much of Banks Island was probably, at one time, covered by these sediments. Their seeming absence from much of central and eastern Banks Island would then be explained as a product of both fluvial and glacial erosion.

## 1.3 Structural Geology

Miall (1975, 1976, 1979) working from a basis of petroleum well sub-surface lithostratigraphic and gravity anomaly data characterized Banks Island as an “Unstable Craton Margin” and described a series of small, broad pericratonic basins, platforms and structural highs. Several periods of minor epirogenic uplift and rifting from Jurassic to Tertiary, produced horst and graben topography, including features such as the Cape Crozier Anticline, Northern and Central Banks basins and the Storkerson Uplift (Fig. 1; Miall, 1979; Jefferson et al., 1988). Little in the way of faulting is identified on the Banks Island geology maps reflecting their pre-dating of much of the seismic exploration activity on Banks, or at least access to these data. Recently, Piepjohn et al., (2018) have demonstrated an extensive array of normal faulting across northern Banks Island, and small-scale, local and isolated deformed zones within otherwise generally undeformed Cretaceous and Paleogene strata. They relate much of the faulting to extension during opening of the Canada Basin (Hadlari et al., 2016), and Eocene Eurekan deformation and possible strike-slip motion along the North American Arctic continental margin.

## 1.4 Quaternary Glacial History and Surficial Geology

There is a long history of observations and reconstructions of the Quaternary glacial history on Banks Island. Much has centered on the extent and timing of continental glaciations, and has for most of its record argued for both restricted Late Wisconsinan ice cover, multiple pre-Wisconsinan glaciations, transgressions by inland seas and proglacial lakes, and the existence of unglaciated terrain. The implications of this complex array of depositional and erosional environments is that the drift records of Banks Island are assumed to be complex and difficult to reconstruct.

The original Glacial Map of Canada (Wilson et al., 1958) summarized the earliest observations of Hobbs (1945), Washburn (1947), Porsild (1950), Jenness (1952) and Manning (1956) and indicated that during the last glaciation (Late Wisconsinan), glaciers inundated the eastern and southern margins of Banks Island, leaving the west and north unglaciated. The revised 2$^{nd}$ edition of the Glacial Map of Canada (Prest et al., 1968) updated the glacial history based on research by Craig and Fyles (1960, 1965) and Fyles (1962, 1965), and indicated that the southwest, south and east coastal regions had been glaciated during the Late Wisconsinan, as was the northern coastal regions bounding M'Clure Strait. The rest of Banks Island was indicated to have been glaciated during pre-Wisconsinan time, although no actual ice limits or margin retreat positions were affixed to this earlier glaciation.

Vincent was the first to focus aerial photograph interpretation and helicopter-based fieldwork traverses (conducted in 1974 and 1975) on a broad-based study of the glacial history throughout much of Banks Island (Vincent, 1978 a-d, 1980 a, b, 1982, 1983, 1984, 1989; Vincent and Edlund, 1978). Magnetostratigraphy and amino acid racemization dating of pre-Wisconsinan interglacial and glacial deposits, first identified by Fyles (Craig and Fyles, 1965) and Kuc (1974), was also undertaken and progressively expanded by Morris and Vincent (1979), Vincent (1982, 1983, 1990, 1992), Vincent et al. (1983, 1984), Barendregt and Vincent (1990), and Barendregt et al. (1998). Glacial reconstructions based on this research suggested a conformable succession of three distinct glaciations (Banks, Thomsen, and Amundsen; Fig. 2; Vincent, 1982, 1983), dating from >780 000 to ~20 000 years ago; these were separated by two interglacials (Cape Collinson and Morgan Bluffs), and preceded by possibly a third interglacial or the pre-glacial Worth Point Formation. The oldest glaciation (Banks) was considered to have been the most extensive, inundating all of Banks Island save for an area in the northwest that apparently contained no till or glacial erratics and was thus considered to have always remained unglaciated (Fig. 2). Successive glaciations to the Banks Glaciation were progressively less extensive. The Amundsen Glaciation included both an Early Wisconsinan continental ice advance (M'Clure Stade – Bar Harbour, Mercy, Jesse, Carpenter and Sachs tills), and a very restricted Late

Wisconsinan continental ice advance (Russell Stade – Viscount Melville lobe; Fig. 2). This configuration of restricted ice margins was used to argue that much of Banks Island formed a refugium during the last glacial maximum (Vincent, 1982, 1983; Harrington, 2005; MacPhee, 2007).

Based on interpretations of stratigraphy exposed in coastal bluffs, geomorphology, and amino-acid racemization dating of surface shell collections, Vincent (1982, 1983) also proposed a series of ice-marginal/ice-dammed lakes and postglacial marine transgressions and regressions that were associated with the various glaciations and till sheets. In some cases, these water bodies represented small to large topographic impoundments, whereas others, such as the post-Thomsen glacial Big Sea, transgressed almost half the island from 60 and up to 215 m above sea level. Thus, in light of the more restricted Amundsen Glaciation extents (Early and Late Wisconsinan), much of the surficial geology west of the Jesse Till (Fig. 2) would potentially constitute raised marine sediments overlying till, or washed till. In the M'Clure stade (Early Wisconsinan) of the Amundsen Glaciation, Lake Ivitaruk is suggested to have formed throughout the Thomsen River valley, impounded between the Prince of Wales lobe to the southeast, and the Prince Albert lobe to the north. Post-M'Clure stade marine transgressions included the East Coast Sea (~120 m) formed along the east and southeast Banks Island coastal periphery, the Investigator Sea (~30 m) along the northern coast, and the Meek Point Sea (~20 m) along the western coast.

The glacial reconstructions of Vincent (1982, 1983) were largely accepted, with one chief modification by Dyke (1987) and Dyke and Prest (1987) that placed the Late Wisconsinan ice limit (18 $^{14}$C ka BP) in configuration with the Jesse Till (Fig. 3). This revised configuration was made in consideration of glacial reconstructions and minimum glaciation levels on adjacent islands, and the isostatic correction of gradients on what Vincent (1982, 1983) had considered to be glacial moraines that were subsequently reinterpreted as ice shelf moraines (Hodgson and Vincent, 1984; Hodgson, 1994). That the majority of Banks Island remained ice free during the Late Wisconsinan, however, remained entrenched in the glacial history and modeling literature.

Recent detailed and methodical aerial photograph and field-based mapping, exploration, chrono- and litho-stratigraphic, and facies investigations on Banks, Melville, and Eglinton islands has fundamentally revised the glacial history of Banks Island, rejecting or reinterpreting the majority of Fyles (1962), Vincent (1982, 1983), Barendregt and Vincent (1990), Barendregt et al. (1998) and others' conclusions. The implications of this collective new research for interpretation of drift stratigraphy on Banks Island are significant, and are herein summarized. This discussion is not meant to be an exhaustive treatment of the material, for which readers are instead referred to England and Furze (2008, 2011), England et al. (2009), Lakeman (2012), Lakeman and England (2012, 2013, 2014), Nixon (2012), Evans et al. (2014), Nixon and England (2014), Vaughan (2014), and Vaughan et al. (2014).

It is now recognized that the continental Laurentide Ice Sheet inundated all of Banks Island during the Late Wisconsinan glaciation. Absolute limits of ice are unknown, but based on stratigraphy and minimum age constraints provided by radiocarbon dates on ice-transported and *in situ* shells, the margins are demonstrated (England et al. 2009; Lakeman and England, 2013) to have lain west of the outer-most islands along Banks Island's west coast, presumably terminating somewhere along the continental shelf (Fig. 3). Numerous maximum and minimum-limiting radiocarbon ages of ~31 $^{14}$C ka BP and ~13.75 cal ka BP, respectively, collected from areas previously mapped as Bernard Till (Fig. 2), and from ice-contact deltas emanating from ice descending off of Banks Island into deglacial higher sea levels, indicate that the uppermost till, found across Banks Island correlates to a Late Wisconsinan advance and retreat (England and Furze, 2008; England et al., 2009; Lakeman and

England, 2012, 2013). This is perhaps the most stark revision of Vincent's (1982, 1983) work, as it simultaneously erases the notion of limited Late Wisconsinan ice cover, and rejects the chronostratigraphic linkage between interpretations of bluff exposures along the west (Worth Point), southwest (Duck Hawk Bluffs) and east (Morgan Bluffs) Banks Island coasts, and the proposed co-eval Banks Glaciation Bernard Till that was considered to have mantled most of the central and western regions of the island (Fig. 2; Vincent, 1982, 1983, 1990; Vincent et al. 1983, 1984; Barendregt et al., 1998).

Lakeman's (2012) and Lakeman and England's (2014) detailed lithofacies and stratigraphic reconstruction of the Morgan Bluffs succession demonstrates how previous reconstructions have erred in their dependence on single exposures, their correlation by deposit type without consideration of lithofacies and facies assemblages, and their inability to recognize the significance of glaciotectonism (cf., Evans et al. 2014; Vaughan et al. 2014). By fully considering lateral in addition to vertical successions, Lakeman (2012) and Lakeman and England (2014) were able to reject many of the previously labeled tills as debris flow, turbidites and glaciomarine deposits, which had previously been interpreted as a succession of discrete thin tills and interglacials correlated with the Thomsen Glaciation (Vincent, 1982, 1983). Similar reinterpretations have been made from detailed lithofacies and stratigraphic analysis of the Duck Hawk Bluffs and Worth Point sections (Evans et al., 2014; Vaughan et al., 2014). What remains consistent is the identification of a lowermost, magnetically reversed till, constrained as >780 ka, and underlain by proposed Jaramillo Subchron normal sediments (Barendregt et al., 1998), hence dating between 0.99-0.78 Ma. Subsequent to this, overlying sediments are largely ascribed to fluvial/marine deposits, changes in which may reflect different climatic and tectonic drivers, and not as evidence of glaciation (Lakeman, 2012; Lakeman and England, 2014).

On the issue of past marine transgressions and glacial lakes postulated by Vincent (1982, 1983), Lakeman and England (2013, p.102) report that "Despite exhaustive field surveys throughout northern and eastern Banks Island, England et al. (2009) and Lakeman and England (2012) did not observe any of the raised marine deposits and shorelines reported." Similarly, no sedimentary evidence of Lake Ivitaruk, extending >300 km along the Thomsen River valley was found (Lakeman and England, 2013), although they did note multiple glaciolacustrine deposits of rhythmically-bedded silt conforming to separate, small ice-dammed lakes in the Thomsen River valley (Fig. 3). Lakeman and England (2012, 2014) also describe small ice-dammed lakes that formed within what they now termed the "Jesse moraine belt" (Jesse Till of Vincent, 1982, 1983; Fig. 2) impounded between ice lobes, or against the slope and the retreating Prince of Wales Strait ice lobe (Fig. 3). Marine limits recorded by Lakeman and England (2013) in Mercy and Castel bays of 41 and 37 m, respectively (dated to ~13.75 cal ka BP), identify a significant amount of glacioisostatic unloading of the Late Wisconsinan northwestern Laurentide Ice Sheet, thusly accommodating the geomorphic observations supporting a thicker and pervasive ice cover on Banks Island. Marine limits on the west Banks Island coast range from ~13-40 m, while on the east coast, they range from 27 m in southeastern Jesse Bay, up to 60 m at the northeastern Parker Point (Lakeman and England, 2013). All of these elevations represent minimums of past glacioisostatic depression since the entire coastline of Banks Island is currently being transgressed, reflecting its position beyond the zero isobase (cf., Andrews and Peltier, 1989; Lajeunesse and Hanson, 2008; Nixon, 2012, Grasby et al., 2013). Extensive sandurs and glaciofluvial outwash in major west-draining valleys (e.g., Bernard, Storkerson, and Big rivers) extend from the Jesse moraine belt dated at ~12.75 cal ka BP (Lakeman and England, 2012) to a base level similar to the modern sea level on the west coast of Banks Island. This suggests that the modern west coast shoreline actually equates to the ~13 cal ka BP shoreline. These broad and extensive sandurs are today occupied by small misfit streams, indicating their dynamic link to the retreating Late Wisconsinan

Laurentide Ice Sheet and abandonment once ice had retreated off of the east coast of Banks Island (post-Jesse moraine belt).

Under the revised glacial reconstruction, following a pervasive Late Wisconsinan ice cover, cold-based ice (cf. Dyke, 1993) retreated eastward during what is now termed the Beaufort Phase, as a series of digitate lobes oriented within major valleys (Fig. 3; Lakeman and England, 2012, 2013). Lateral meltwater channels record the topographically conformable retreat of this ice, indicating that ice cover was thin. Cold-based ice conditions are implied by the virtual absence of moraines, other than rare ice thrusted frozen sediments along former margins, and by the minor scattering of erratics and the very thin, discontinuous and largely absent till veneer (Lakeman and England, 2013). On central and western areas of Banks Island (equating to Vincent's (1982, 1983) Bernard Till), clasts and rare boulders of mafic erratics derived from western Victoria Island occur sparsely; sandstone and granite (Canadian Shield origin) erratics are also found ubiquitously, albeit rarely (England et al., 2009; Lakeman and England, 2012, 2013, 2014; Smith and Farineau, 2015). Upland surfaces are characterized by *in situ* weathered bedrock and colluviated material, and even where sediment exists that is interpreted as till, it largely consists of reworked local poorly-consolidated bedrock material (Lakeman and England, 2012, 2013). Large and extensive glaciofluvial valley trains began forming as the ice retreated eastward, depositing and remobilizing largely local bedrock and sedimentary material.

By ~14 cal ka BP, ice had retreated to the island's central interior, forming a prominent curvilinear margin (Fig. 3), distinct from the previous lobate flow pattern; this ice is considered to have remained cold-based (Lakeman and England, 2013). Lateral meltwater channels draining northward along this curvilinear ice front and down the Thomsen River valley into Castel and Mercy bays contacted marine limit deltas the age of which are constrained by mollusc shell dates of ~13.75 cal ka BP (England and Furze, 2008; England et al. 2009, Lakeman and England, 2012, 2013). This northward phase of ice flow down the Thomsen River valley converged with thick (>600 m) westward flowing ice in M'Clure Strait and is labeled the Thomsen Phase (Fig. 3; Lakeman and England, 2013). Subsequent ice retreat occurred southwards and east towards Prince of Wales Strait. The Jesse moraine belt (Fig. 3), marks a prominent stillstand or readvance of the Laurentide Ice Sheet on eastern Banks Island (Prince of Wales Phase) forming an ice lobe that occupied Prince of Wales Strait and terminated in M'Clure Strait as a tidewater trunk glacier (Lakeman and England, 2012, 2013). Abundant mollusc shell dates constrain the age of deposition of the Jesse moraine belt as occurring between ~13.75 – 12.75 cal ka BP (Lakeman and England, 2012). Morphometry of the Jesse moraine belt is distinctly different from any other deposits on Banks Island, and signals the change from cold-based ice to polythermal conditions (Lakeman and England, 2012). Abundant ice-cored controlled moraines (cf., Evans, 2009), sharp-crested ice thrust moraines, and nested end moraines are found within the Jesse moraine belt, as is foliated, clast-rich buried ice (presumed to be buried glacial ice; French, 1974; Lakeman and England, 2012, 2013; Smith and Farineau, 2015; Rudy et al. 2017). Till thicknesses of 0.5 to 2 m within the Jesse moraine belt are also dramatically different from the general absence or scant veneer of till elsewhere across the island. These deposits are associated with a deglacial landsystem (cf., Dyke and Evans, 2003; Evans, 2009) that reflects the strong advection of material from areas of warm-based flow to peripheral cold-based margins (i.e., polythermal glaciers) where it is variously deposited in proglacial debris aprons or becomes adfrozen to the glacier base and is thrust and stacked as a series of controlled moraines.

# 2. Methods

## 2.1 Sample Collection

Samples for KIMs and other heavy minerals were collected on Banks Island over two summer field seasons (Smith, 2015; Smith et al., 2016). In 2015, a basecamp was established at Johnson Point (72°46.3'N; 118°29.3'W) and a helicopter was used to access sites along eastern coastal and central interior regions (Fig. 4). In 2016, working with a helicopter based at Sachs Harbour, samples were collected at Duck Hawk Bluffs (SW Banks Island; 71°58.7'N; 125°29.9'W) and in the Nelson Head area (SE Banks Island; 71°14.5'N; 123°05.6'W) before moving north to a basecamp at Polar Bear Cabin (74°08.3'N; 119°59.7'W) within Aulavik National Park, where collections from central interior and northern Banks Island were made (Fig. 4). In both summers, to varying degrees, helicopter mechanical and weather issues significantly limited the number and geographic distribution of samples that could be collected.

The selection of sample sites was informed by previous industry collections, and was designed to both test and expand these surveys, particularly as they related to comprehending potential glacial dispersal histories under the new model of Late Wisconsinan glaciation (cf., England et al., 2009; Lakeman and England, 2012). Sample collection sites included areas within Aulavik National Park (established 1992), a region industry would have been excluded from operating. Samples were also collected from pre-Late Wisconsinan till and glaciofluvial deposits identified in Duck Hawk Bluffs (Evans et al., 2014) and Morgan Bluffs (Lakeman and England, 2014) in order to assess potential earlier glacial KIM dispersal that may subsequently have been reworked by Late Wisconsinan ice flow. The Jesse moraine belt was revealed to be extensively underlain by buried glacial ice (French, 1974; Lakeman and England, 2012, Smith and Farineau, 2015), within which retrogressive thaw slumps are now exponentially being triggered and expanding, contributing significant sediment loads to streams and rivers (Rudy et al., 2017; Lewkowicz and Way, 2019). Several samples specifically targeted material from these slumps in order to test potential differences in KIM contents between what was considered more englacial till/debris capping thick buried glacial ice (>25 m; Smith and Farineau, 2015), from other areas considered more likely to be basal till.

Samples of unconsolidated Lower Cretaceous fluvial and alluvial fan Isachsen Formation strata, comprised dominantly of quartz arenite and lithic arenite (also limestone, shale, and chert) derived from local and farther-travelled Proterozoic Glenelg and Shaler Group (which is of first order or multi-cycle derivation from the Canadian Shield; Miall, 1979) were also collected to determine if they might host KIMs. Finally, unconsolidated fluvial Beaufort Formation sediments, which comprise the uppermost bedrock unit on Banks Island were sampled for KIMs. Previously, the extent of Beaufort Formation strata on Banks Island was delimitated to western and central areas (Fig. 5; Miall, 1975; Vincent, 1983, 1990). During the course of fieldwork in 2015, deposits of conspicuously orange-coloured gravel containing deeply weathered rounded quartzites and abundant black and grey chert (with minor red and green chert) were identified capping upland terraces at numerous sites in northeastern Banks Island (Fig. 5). These neither correlate with Vincent's (1983) glaciofluvial or ice-contact deposits, nor Vincent's (1990) Beaufort Formation deposits. Visually, these deposits strongly suggest great antiquity and a non-glacial origin (e.g., clast sizes, absence of striae, lithological compliment, degree of pebble rounding and oxidation/weathering), as does a clear geomorphic pattern of isolated plateau remnants and former peneplains and associated incised fluvial terraces that can be traced 10s of kilometres across the landscape (Fig. 5). For these reasons, they are herein interpreted as Beaufort Formation deposits, *sensu lato*. Given the location of these sites within catchments where industry surveys had previously recovered KIMs, and the fact that the Beaufort Formation is reported to contain rare granites (Miall, 1979; hence, a potential Canadian Shield origin), focus was paid to collecting samples of these and other mapped Beaufort Formation sediments throughout the field area over both summer field seasons in order to test for the potential of bedrock inheritance of KIMs.

Several strategies were employed in sample collection, and these followed established GSC protocols (Spirito et al., 2011; McClenaghan et al., 2013, in press). Bulk samples (n=18) were typically collected from unconsolidated Beaufort Formation and Isachsen Formation sediments, tills and glaciofluvial deposits (Appendix 1A). At surface collection sites, holes were dug ~50-70 cm deep (accordant with depth of the active layer at time of sampling), and then a 19 litre (5 gallon) plastic pail was filled with the basal sediments (15-30 kg), discarding cobbles (>64 mm), while retaining the pebble fraction (2-64 mm) for later lithic, provenance, and shape analysis. At exposed section locations (e.g., Duck Hawk Bluffs), vertical faces were cleaned with a shovel until all slumped material had been removed, and then the sample was collected.

Sieved samples (<2.38 mm (#8 mesh); n=30) were collected from stream sediments, Beaufort Formation deposits and one till sample (Appendix 1A). Samples were predominantly wet-sieved directly into plastic 19 litre pails while standing in the stream environment and followed NGR stream sediment sampling protocols (McCurdy et al., 2012). Stream sediment samples were collected at sites where stream flow would naturally trap heavy minerals, such as at the head of mid-channel bars, and behind boulder traps (cf., Prior et al., 2009; Smith, 2015). In cases of dry, ephemeral stream beds, and in coarser clastic, sand-poor sediments, material was dry-sieved to produce a sand concentrate.
A total of 48 samples (including 2 duplicates; 15SUV019, 15SUV055) were collected on Banks Island as part of the GSC's KIM investigations: 30 stream sediment samples, 7 Beaufort Formation deposits, 5 till samples, 4 glaciofluvial sediment samples, and 2 bedrock samples (unconsolidated Isachsen Formation sandstone; Appendix 1A, 1B).

## 2.2 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. 6; McClenaghan et al., 2013, in press; Plouffe et al., 2013). Received samples were disaggregated and sieved at 2 mm. The <2 mm fraction was then passed across a shaker table to pre-concentrate the heavy fraction, which 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 preconcentrates which were then sieved at 0.25 mm. The 0.25 to 2.0 mm concentrate was then refined using a heavy liquid separation (3.2 specific gravity (SG)). Concentrates were then ferromagnetically separated. The non-ferromagnetic heavy mineral fractions were 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 >3.2 SG fraction was subjected to paramagnetic separations, producing <0.6 amp (strongly paramagnetic), 0.6-0.8 (moderately paramagnetic), 0.8-1.0 amp (weakly paramagnetic) and >1.0 amp (non-paramagnetic) fractions 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. Six samples from Beaufort Formation deposits were processed in a similar manner for KIM recovery in the 0.18-0.25 mm heavy mineral fraction; this required separate sieving, paramagnetic separation and oxalic acid cleaning. In support of broader reconnaissance geological sampling objectives and to further constrain mineral dispersal studies, ODM's metamorphosed or magmatic massive sulphide indicator minerals (MMSIM®; Averill, 2001) and other heavy minerals were also picked. Visual identification of a limited number of KIMs and MMSIM grains were verified by ODM using an EDS-equipped SEM.

GSC quality-control checks on sample processing included the insertion of two KIM-spiked blanks in each of the sample year batches (Appendix 1A, 1B, 2; 15SUV017, 15SUV029, 16SUV021, 16SUV029). In order to mimic the range of sample sediment types being submitted to ODM for processing, the blanks included bulk samples of both the GSC's Linton till and Bathurst weathered granite (grus; mimicking a fluvial sediment), each of which is known to contain no KIMs (Plouffe et al., 2013). Based on results from previous industry sampling programs, it was presumed that KIM concentrations in the GSC Banks Island samples were likely to be low or zero. In addition to using the blanks as a means of monitoring carry-over of grains between samples, the four GSC blanks were spiked with known numbers and sizes of different KIMs previously picked by ODM from Buffalo Head Hills (Alberta) till and glaciofluvial samples (Smith and Paulen, 2016) as a means of testing KIM recovery. Preference was given to those grains confirmed by ODM using SEM, in which case, these grains were washed with acetone and then inspected to ensure that any adhering SEM stub glue was removed. Only euhedral chromites were selected, and the forsterite grains came from a particularly olivine-rich sample, estimated by ODM to contain >60 000 grains. All spike KIM grains were photographed before they were inserted into the blanks in order to facilitate their later confirmation after sample processing and KIM picking. Two duplicate field sample were also collected in 2015 and submitted as a blind duplicate pairs – samples 15SUV018 and 15SUV019 were collected at the base of individual dug pits from the same upland gravel deposit, approximately 10 m apart, and sample 15SUV031 and 15SUV055 were collected from the same dug pit.

## 2.3 Mineral chemistry

Selected KIM and other heavy indicator minerals 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 $\varphi$ ($\rho Z$) 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 garnet, olivine, clinopyroxene, spinel, Mg-ilmenite and rutile grains were determined by laser ablation inductively coupled mass spectrometry (LA-ICP-MS) 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 ($M/\Delta M$=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 was acquired using the rapid peak-hopping multichannel mode of the ICP-MS (cf., Poitras et al., 2018). Data was 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 secondary standards (cf., Liu et al., 2018; Poitras et al., 2018). Garnet trace element concentrations are normalized to $^{43}Ca$; olivine trace element concentrations are normalized to $^{29}Si$.

Magnesian-ilmenite Hf isotope compositions have been used to infer kimberlite emplacement ages with varying levels of accuracy and precision (cf., Nowell et al., 2003; Poitras et al., 2018). As kimberlites on Victoria Island, Parry Peninsula and elsewhere in the western Canadian Arctic have a disparate range of ages, Hf isotope compositional analysis was undertaken on Banks Island samples as a further means of discriminating potential KIM sources. Preparation of Mg-ilmenite grains follows that outlined by Poitras et al. 2018 (their *Supplemental Material 2*). Individual grains were dug out of chemical analytical mounts, inspected, and then rigorously cleaned before being dissolved in a mixture of 1:3 concentrated $HNO_3$:HF, evaporated to dryness, and then re-dissolved in concentrated HCl. Dissolved Mg-ilmenite solutions were then evolved chromatographically, and purified solutions were analyzed for Hf isotopic signatures by multi-collector (MC-)ICP-MS according to the instrument analytical protocols of Nowell et al. (2003).

# 3. Results

## 3.1 KIM Sample Processing Quality Assurance – Quality Control

Results of KIM recovery in spiked blanks and duplicate samples 15SUV018 and 15SUV019, and 15SUV031 and 15SUV055 are provided in Appendix 2. In 2015, 19 and 16 known KIMs that were 1-2 mm or 0.5-1.0 mm in size were inserted into blank samples 15SUV017 and 15SUV029, respectively, and submitted blindly to ODM for processing. Recovery of the spike KIM grains was 90 and 81%.

In the till sample (15SUV017), only the two 1-2 mm chromite grains were not recovered. In the 15SUV029 sample, one additional pyrope garnet was picked and 1 of 2 chromites were not recovered in the 1-2 mm size fraction, while 2 pyrope garnets and 1 chromite were not recovered from the 0.5-1.0 mm size fraction. It is well documented that both types of GSC in-house blanks do not contain any KIMs (Plouffe et al., 2013; GSC unpublished data). Visual comparisons of spiked grains and picked grains indicated that the additional 1-2 mm sized pyrope garnet was one of the 0.5-1.0 mm spike pyrope grains, and thus is simply a sieving artifact. It has been indicated (Averill and McClenaghan, 1994) that in the case of orange garnets, it is much more difficult to discriminate and select kimberlitic versus crustal garnets purely on the basis of colour. Often there may be considerable abundances of orange crustal almandine and spessartine grains, and thus ODM picks up to 30 percent ambiguous orange grains that may or may not be kimberlitic in order to ensure that no true kimberlitic grains are missed. ODM did pick 4 almandine grains, which they confirmed by SEM analysis, from sample 15SUV029. Visual comparison between the pyrope spike grains and these almandine grains recovered by ODM indicate they do not include the missing 0.5-1.0 mm size pyrope. In the duplicate samples, 15SUV018 yielded 3 chromites in the 0.25-0.5 mm size fraction, 15SUV019 yielded 2 pyrope garnets in the 0.25-0.50 mm size fraction, 15SUV031 yielded a single chromite grain in the 0.25-0.5 mm size fraction, while no KIMs were recovered in sample 15SUV055. The difference between these is considered insignificant, and also indicates a likely absence of carry-over between successively tabled samples.

In 2016, spike grains inserted into the blank samples were 0.5-1.0 or 0.25-0.5 mm in size in order to better reflect actual KIM size ranges in the 2015 routine samples. Twenty and 21 grains were inserted in blanks 16SUV021 and 16SUV029, respectively (Appendix 2). Picking results from these two blanks is considered poor. Spiked KIM grain recoveries in the 0.5 – 1.0 mm size fraction are better than the finer 0.25-0.5 mm fraction. In the 0.5-1.0 mm fraction, 1 of 2 Mg-ilmenites were not recovered in sample 16SUV021, while 1 of 1 Mg-ilmenites were not recovered in sample 16SUV029. In the 0.25-0.5 mm fraction of sample 16SUV021, 0 of 4 Cr-diopsides, 0 of 2 Mg-ilmenites, and 2 of 4 chromites were not recovered. An eclogitic garnet (GO), that was indicated by ODM to be SEM-confirmed was picked, but no such grain was included in the spike. Curiously, while 5 forsterite grains were inserted in the blank sample, 20 grains were recovered by ODM, of which 15 were picked out of the sample, 5 of which were confirmed by SEM. Grain fracture and splitting does not explain this disparity, nor does carry-over from the previously processed sample (16SUV020) which had no KIMs. In the 0.5-1.0 mm size fraction of sample 16SUV029, 0 of 1 Mg-ilmenites were not picked, while in the 0.25-0.5 mm size fraction, 0 of 1 chromium diopsides, 0 of 5 Mg-ilmenites and 2 of 5 forsterites were not picked.

## 3.2 Kimberlite Indicator Mineral – Abundances

For archival purposes, the unedited data files reported by ODM for sample processing are reported in Appendix 3A (year 2015) and 3B (year 2016). Amended and compiled data for the two years of sampling are presented in Appendix 4A-4J. A total of 48 samples (including 2 duplicates; 15SUV019 and 15SUV055) were collected on Banks Island as part of the GSC's KIM investigations. These included 30 stream sediment samples, 7 Beaufort Formation deposits, 5 till samples, 4 glaciofluvial deposits, and 2 bedrock samples (unconsolidated Isachsen Formation sandstone; Appendix 4A). KIM abundances are reported on an "as-picked" (raw data) basis in three size fractions (1.0-2.0 mm, 0.5-1.0 mm, 0.25-0.5 mm; Appendix 4C). These KIM identifications and numbers were then confirmed and where necessary, corrected, based on EPMA chemistry (discussed below; Appendix 4C). These corrected KIM abundances (Appendix 4C) were then normalized to a 10 kg <2 mm table feed weight (Appendix 4D), and also normalized to a 50 g nonferromagnetic heavy mineral concentrate (HMC) 0.25-0.5 mm picking fraction weight (Appendix 4E). While simple presence/absence of KIMs has significance, use of normalized abundance data is important when comparing disparate sample types, varying sample masses, sampling programs, and for removing biases for natural sediment density-concentrating mechanisms present in stream environments (cf., McClenaghan et al., 2002; Prior et al., 2009; McClenaghan, 2011; Day et al., 2018). KIMs were recovered in 32 of the 48 samples, yielding a total of 334 grains. No KIMs were found within the 1.0-2.0 mm size range, 6.0% were within the 0.5-1.0 mm size, and 94.0% in the 0.25-0.5 mm size. Normalizing the KIM data to a 10 kg table feed (<2 mm) weight does not dramatically change the percentage compositional data from the raw data. Chromites were the dominant KIM recovered (62.5%), followed by pyrope garnets (14.1%), Mg-ilmenites (15.6%), and forsterite (olivine; 7.0%); eclogitic garnets (0.7%) and Cr-diopsides (0.6%) are rare (Appendix 4D). When normalized to 50 g of the nonferromagnetic 0.25-0.5 mm picking fraction though, chromites remain the most abundant KIM, but decline to 46.6%, pyrope garnets increase significantly to 31.6%, followed by Mg-ilmenites (17.1%), forsterite (3.3%), eclogitic garnet (0.7%) and Cr-diopsides (0.7%; Appendix 4E).

Six of the 7 Beaufort Formation samples were also examined for KIMs in the 0.18-0.25 mm fraction in order to further characterize the KIM content in these ancient fluvial deposits, and to test for potential KIM comminution. The time-consuming nature, and hence cost of picking this fine fraction, means this is a step rarely included in KIM surveys, but can be added to bolster individual KIM numbers when trying to qualify aspects of geochemistry (e.g., McClenaghan et al., 2008). Results show generally higher KIM abundances in the 0.18-0.25 mm fraction as compared to the coarser fractions, along with an increase in the number of Mg-ilmenite grains (18.0 to 23.5%) and a decrease in pyrope garnets (48.0 to 34.6%; Appendix 4D). It should be noted that both Mg-ilmenites and chromites were not picked to completion in sample 16SUV023. This sample contains significantly higher KIM abundances than all other samples, particularly garnets, and may be skewing results.

None of the pre-Late Wisconsinan till samples from Morgan Bluffs (15SUV022, -023; Lakeman and England, 2014) or Duck Hawk Bluffs (16SUV019, -020; Evans et al., 2014) returned any KIMs (Appendix 4C). The "old" (?>0.99 Ma; Barendregt et al., 1998) preglacial gravel from Morgan Bluffs (15SUV024; lithofacies 3 of Lakeman and England (2014)) returned a diverse, low abundance KIM assemblage (Appendix 4C), and is one of only two samples that contained an eclogitic garnet. Only 1 of the 2 basal glaciofluvial outwash samples from Duck Hawk Bluffs (16SVU018) had any KIMs; 4 chromites (Appendix 4C). The only other till sample collected as part of this study (a Late Wisconsinan till from within the Jessie moraine belt) was from an area west of Johnson Point (15SUV054; Fig. 4c; Appendix 1A), and it returned two chromite grains (Appendix 4C). Neither of the two Isachsen Formation sandstone samples yielded KIMs (16SUV022, -024; Appendix 4C). Interestingly, the four Morgan Bluffs and Duck Hawk Bluffs till samples, the two Duck Hawk Bluffs glaciofluvial samples, and the two Isachsen Formation bedrock samples all had extremely low nonferromagnetic HMC 0.25-0.5 mm sample masses (Appendix 4B, 4F), representing between 7.7 and 71.4 times less than the 50 g mass used to standardize grain counts to. The one other till sample collected (15SUV054) has a nonferromagnetic HMC 0.25-0.5 mm mass of only 10.6 g (Appendix 4B, 4E). Five of the 7 Beaufort Formation samples recorded large nonferromagnetic HMC 0.25-0.5 mm mass deficiencies compared to the 50 g standard (2.1-19.2 times; Appendix 4E), including sample 16SUV023 which had the highest number of garnets and the highest overall number of KIMs in the raw picked data. When the 50 g picking fraction standardization is factored in, the KIM numbers for sample 16SUV023 become exceptional (Appendix 4E), indicating that its high numbers in the raw picked sample is not the simple product of nonferromagnetic heavy mineral concentration in a fluvial environment. Similarly, even when the mass of the total HMC 0.25-2.0 mm (SG>3.2) fraction, normalized to a 10 kg Table Feed (<2 mm) is considered (Appendix 4B), Beaufort Formation samples are generally half that of the stream sediment samples (average of 31.6 and 74.6, respectively). This suggests that there is a relative deficiency of heavy minerals within Beaufort Formation samples compared to modern stream sediment samples, although to what degree this could reflect post-depositional weathering is uncertain. Both modern stream and Beaufort Formation sediment samples have significantly higher HMC concentrations than the 5 till samples from this study (avg. 5.9 g/10 kg Table Feed (<2 mm); Appendix 4B).

## 3.3 Kimberlite Indicator Mineral – Grain Morphology

All the KIM grains picked by ODM were inspected and photographed using paleontological microscopes at GSC Calgary, prior to their being mounted for chemical analysis. Photomicrographs are reproduced in Appendix 5. KIM grain morphology has been used to help discriminate near- from far-travelled grains. The absence of kelyphitic rims, increased grain rounding, and the absence of fractures can be proxies for transport-induced physical weathering (cf., Mosig, 1980; Afanas'ev et al., 1984; Garvie and Robinson, 1984; McCandless, 1990; Averill and McClenaghan, 1994; Dredge et al., 1996; McClenaghan and Kjarsgaard, 2001, 2007; Cummings et al., 2014). Inherent properties of individual kimberlites, their emplacement history, and individual mineral physical properties may also be significant in controlling grain morphology independent of actual glacial/fluvial/eolian transport, abrasion, and weathering (cf., Pokhilenko et al., 2010; Jones and Russell, 2018). No kelyphitic rims, or remnant coatings were seen on any of the recovered garnets from Banks Island. However, the sub-kelyphitic "orange-peel" texture was seen on 11 of 23 pyrope garnets (excluding the 16SUV023 sample; c.f., samples 15SUV026 and 16SUV030 – Appendix 5, where it is prominently displayed), and on 4 of the 6 eclogitic garnets. In sample 16SUV023, the "orange-peel" texture was seen on 10 of 25 pyrope garnets in the 0.25-0.5 mm fraction, and in 6 of 18 pyrope garnets in the 0.18-0.25 mm fraction. Recovered pyrope garnets are generally sub-angular to sub-rounded, often displaying conchoidal fracture surfaces. Within sample 16SUV023, the pyrope garnets tend to be more sub-angular to angular, and the angularity is higher in the finer 0.25-0.18 mm fraction than the 0.5-0.25 mm fraction. Conchoidal fracture surfaces are seen in many of the 16SUV023 pyrope garnets, and more of these mineral grains have jagged edges than are seen with the other Banks Island pyrope garnets. These differences may reflect the relative transport/depositional systems operating in deposit 16SUV023 (Beaufort Formation fluvial) and the fracture of grains along pre-existing internal weakness/fracture planes versus most of the others which were recovered from primary and reworked glacial deposits (e.g., Averill and McClenaghan, 1994; Cummings et al., 2014). The Mg-ilmenite grains display a range of surface characteristics, including pitted surfaces, conchoidal fracture, and prismatic overgrowths of what may be microcrystalline ilmenite (e.g., Appendix 5 – 16SUV014, upper right of three grains). Chromite grains largely display varying degrees of resorbed octahedral crystal faces, although distorted, elongated and subhedral crystal shapes are also seen (Appendix 5), which are suggested by Yannan and Matsyuk (1991) and Lee et al. (2003) as being characteristic of kimberlitic or lamproitic sources. Note, in the sample 15SUV009 photographs, 15 of the 20 grains picked by ODM as chromites were determined by EPMA to be Mg-ilmenite grains instead.

## 3.4 Other Heavy Minerals and MMSIMs

Non-kimberlitic heavy indicator minerals (from heavy mineral concentrate (HMC) ≥3.2 SG) picked by ODM are described in Appendix 4F (Picking Remarks) and Appendix 4G, tabulated in Appendix 4H, and summarized in Table 1.

For the 0.25-0.5 mm size fractions of those samples selected for ODM's MMSIM screening (Averill, 2001), results are provided in Appendix 4F and 4G and are included in the heavy mineral tabulations of Appendix 4H. Generally, there were few non-KIM heavy minerals identified/picked in the 0.5-1.0 mm fraction of the Banks Island samples. The most abundant non-KIMs in the 0.25-0.5 mm fraction were pyrite, chalcopyrite, sphalerite and vesuvianite. Pyrite abundance in the 0.25-0.5 mm fraction of the subset of samples submitted for MMSIM is highly variable, ranging from ~300 000 grains to zero (Appendix 4G, 4H). High pyrite abundances are found in Beaufort Formation and stream sediment samples, but these types of sediment samples can also have low or zero concentrations, so there is no discernible trend. Next to pyrite, chalcopyrite is the most common of the identified/picked grains in the 0.25-0.5 mm size fraction (30 of 48 samples), with up to 31 individual grains picked in a single sample. Chalcopyrite is absent, however, in 6 of the 7 Beaufort Formation samples. The 0.25-0.5 mm fraction of 10 samples contained sphalerite and vesuvianite grains.

The vesuvianite grains are potentially a useful indicator of glacial dispersal from western Victoria Island. Contact metamorphism between intrusive gabbro sheets and silty dolostones produced vesuvianite which outcrops on Victoria Island in the Neoproterozoic Shaler Supergroup (Nabelek et al., 2013). Vesuvianite grains are not found in any of the Beaufort Formation samples. Other minerals described by Nabelek et al. (2013) as having formed in the Shaler Supergroup during contact metamorphism and fluid flow include diopside, garnet (grossular and andradite), phlogopite, serpentine, chlorite, and Fe-sequestered sulphides (pyrite). Chlorite, phlogopite, and serpentine have specific gravities ≤3.3 and may thus have been too low to have been recovered in the heavy liquid separation employed in this study. Diopside was prevalent in 28 of the heavy mineral concentrates which were characterized by ODM as largely augite-diopside, or augite-goethite-diopside assemblages (Appendix 4F-H), pointing to the high abundance of these heavy minerals (but not picked). None of the Beaufort Formation samples contained these diopside-rich assemblages, and only 3 diopside grains from 2 Beaufort Formation samples were picked reflecting their rarity (Appendix 4G, 4H). Andradite was recovered in 5 of the Banks Island samples (none of the Beaufort Formation samples). Stream sediment surveys in dolomitized Shaler Supergroup terrains at the head of Minto Inlet on western Victoria Island (McCurdy et al., 2012, 2017), recovered both sphalerite and galena grains, pointing to the presence of potential MVT Pb-Zn mineralization (unknown localities). The sphalerite and galena grains recovered in the Banks Island samples could thus be derived from this area, and glacially transported westward. Isotopic and geochemical analyses of recovered sphalerite and galena grains may provide a mechanism for correlating with Victoria Island mineral grains/source, and is currently being explored.

## 3.5 KIM Chemistry

EPMA data of all Banks Island KIMs and other heavy minerals picked by ODM are presented in Appendix 6A, and then separately by mineral type in Appendix 6B to 6J. Chemical classifications of KIMs and other heavy minerals picked are listed in Table 1. Based on EPMA data, 22 of the 356 KIM grains visually identified and picked by ODM were rejected as KIMs, representing 6.2% of the total. Additionally, 20 picked KIMs were re-classified; 16 chromites were geochemically re-classified as ilmenites/Mg-ilmenite, 2 eclogitic garnets were re-classified as Cr-pyrope garnets, and 2 Mg-ilmenites were re-classified as chromites. Total mineral identification variance from what ODM picked is 11.8% (n=42).

### 3.5.1 Garnet

Of the 122 garnets picked (and/or confirmed by EPMA; excluding replicate samples), 66 were Cr-pyropes (wt%$Cr_2O_3$≥1.0; Appendix 6B). Garnets were classified using the scheme of Grütter et al. (2004), derivation of which and final classifications are shown in Appendix 6B and plotted in Figure 7. There are forty G0 (unclassified), 5 low-Cr megacrystic G1 garnets, 9 eclogitic G3 garnets, 4 eclogitic G4 garnets ($Na_2O$>0.07 wt%), 41 lherzoilitic G9 garnets, 12 harzburgitic G10 garnets (3 of which plot within the diamond stability field and are classified as G10Ds), 9 high $TiO_2$ peridotitic G11 garnets, and 2 wehrlitic G12 garnets. Of the 3 G10D's, 1 occurs in the Beaufort Formation sample 16SUV023 (as do 6 of the 12 G10 graphite phase), and the other two occur in stream sediment samples 15SUV014 and 16SUV025.

A plot of $Na_2O$ vs $TiO_2$ (wt%; Fig. 8) indicates that most (n=39) of the Cr-pyrope garnets plot within the websteritic/pyroxenitic/type 2 eclogitic zone, with a lesser abundance of megacrystic garnets (n=27), and no diamondiferous type 1 eclogitic grains. Eclogitic garnets (Cr2O3 wt% <1.0; G0, G1, G3, G4) were also analyzed following the statistical discrimination methodology of Hardman et al. (2018a; Appendix 6B); results are displayed on a graphical plot of ln(Mg/Fe) versus ln(Ti/Si) (Fig. 9). This plot differs from the Ca# (molar Ca/[Ca+Mg]) vs Mg# discrimination of Schulze (2003), and is considered to better separate mantle versus crustal-sourced eclogitic garnets (Hardman et al., 2018a, b;

Poitras et al., 2018). Regardless of which discrimination scheme is used, a chemical compositional overlap between the two sources remains. Results for 34 low-Cr garnets (excluding all andradite garnets; ln(Ti/Si) >-3.0; Appendix 6B) suggest only two of the Banks Island low-Cr garnets are conclusively mantle derived (16SUV023, G4; and 16SUV028, G1). However, numerous grains plot within the 5 and 10% confidence interval bounds of Hardman et al.'s (2018a) calibration dataset (and their full 10% discrimination error), suggesting that while the vast majority of low-Cr garnets are likely crustal-derived, more than simply the two that lie beyond the 10% crustal confidence interval (CI) may be mantle-derived (Fig. 9).

Trace and rare earth element (REE) compositions provide sensitive tracers of metasomatic processes and have been increasingly used to characterize and qualify different populations of garnets and Cr-pyropes (cf., Griffin et al., 1996; Stachel et al., 1998, 2004; Cox and Barnes (2005); Banas et al., 2009; Viljoen et al., 2014; Hardman et al., 2018b; Liu et al., 2018; Poitras et al., 2018). Analytical results of trace and REE analysis are presented in Appendix 7, and by individual mineral grain types in Appendix 7A through 7H. Garnets (Appendix 7A) were further subdivided to isolate the Cr-pyrope G9, G10, G11, and G12s (n=39, 12, 8, and 2, respectively; excludes high Ca-andradites and one lost G11 grain), and their elemental concentrations were normalized to the C1 carbonaceous chondrite values of McDonough and Sun (1995; Appendix 7A). Results for each mineral grain were plotted and then grouped by populations of (1) heavy rare earth element (HREE)-enriched (n=27), (2) sinusoidal (n=25), (3) middle rare earth elements (MREE)-depleted (n=4), and (4) others (n=6; Fig. 10). All but 1 of the HREE-enriched grains have lanthanum (La) concentrations below chondrite levels, and 14 (of 27 total) also have cerium (Ce) below chondrite levels. Concentrations 10x chondrite levels and greater for the HREE-enriched grains generally occur from samarium (Sm) upwards, and there is little difference in the average REE composition of the G9 versus G11 grains, other than a slight enrichment of the heaviest REE in the G11 grains (Appendix 7A, Fig. 10A). There is quite a bit of variation within the sinusoidal group, including the degree of curvature (particularly with the HREE), elemental concentrations above and below chondrite values, and the elements where peak and low concentrations occur (Appendix 7A, Fig. 10B). The sinusoidal group includes 35% of G9s, 67% of G10s (including 2 of 3 G10Ds), 13% of G11s and 100% of G12s (Fig. 10B). All of the sinusoidal group G9s have La concentrations below chondrite levels, as does the 1 G11, but only 2 of the 8 G10s; all other grains, including the 2 G12s have La concentrations above chondrite level. Highest elemental abundances above chondrite levels in the sinusoidal curves occur in the LREE; neodymium (Nd) in 4 of 14 G9s, 6 of the 8 G10s, and both G12s; Sm in 5 G9s, 1 G10, 1 G11, and 1 G10; and europium (Eu) in 4 G9s. Lowest abundances in the HREE (close to chondrite levels) occur typically in holmium (Ho), erbium (Er), and thulium (Tm), but show a wide variation (particularly in the G9s), including gadolinium (Gd), terbium (Tb), ytterbium (Yb) and lutetium (Lu); excluding La, Tm is the lowest HREE in 6 of 14 G9s and 2 of 8 G10s; Ho in 4 of 8 G10s and 1 of 2 G12s; Er in the 1 G11, 2 of the G10s, and 1 of each of the G9 and G12s. The degree of sinuosity is higher for the G10, G11 and G12s, then it is for the G9s, and differences between the two sets are most pronounced in the HREE fractions where the G10, G11 and G12s trend towards chondrite normal levels and the G9s show greater enrichment (Fig. 10B).

Metasomatic depletion and enrichment of REE in Cr-pyropes is similarly reflected in Y and Zr concentrations (Griffin and Ryan, 1995; Griffin et al. 1999; Banas et al., 2009; Smit et al., 2014; Stachel et al., 2018). All but two of the hartzburgitic (G10) Cr-pyropes from this study exhibit depleted mantle source characteristics, as do all the low-Ti ($TiO_2 \leq 0.02$ wt%) lherzolitic Cr-pyropes, and all the MREE-depleted and LREE-depleted (normal to slightly humped; Fig. 10, Appendix 7) Cr-pyropes (Fig. 11). Most lherzolitic (G9) Cr-pyropes from this study are high-Ti ($Ti0_2 \geq 0.06$ wt%), and display characteristics of high temperature melt-metasomatism (Fig. 11). The spread of higher Y and

Zr values follows the trend of grain core to rim characteristics that are reflective of zoning and secondary replacement rims (Griffin et al., 1999).

Ni-in-garnet geothermometry determinations were made on the Cr-pyropes, and are displayed in Figure 12 in a combined histogram and cumulative probability plot. Ni-in-garnet temperature reconstruction using the methodologies of Ryan et al. (1996), Griffin et al. (1989), and Canil (1999) are included in Appendix 8 for comparative purposes, and are used for modelling geotherms in P (kbar) and conversions to depth (km) of projections to modelled geotherms. The Ni-in-garnet geothermometer is based on the strong temperature dependence of the partitioning of Ni between garnet and olivine, and provides a means of reconstructing the thermal state of the lithosphere from which they were derived (Ryan et al., 1996). This provides a means of characterizing the sample garnet population as a reflection of those which may have been derived from the diamond stability window, considered to occur between 900-1200°C. While sample numbers are low (n=61; 15 from stream sediment samples and 46 from Beaufort Formation samples), it appears to suggest that some Cr-pyropes from both sample media display potential of being derived from diamondiferous kimberlitic magma. The stream sediment samples appear to exhibit characteristics of a cooler geotherm, while those from Beaufort Formation sediments appear to contain garnets from a wide range of temperatures, a third of which exist within a hot geotherm where no diamonds would either exist within the lithosphere, or survive the transport path to the surface.

REE plots of Cr-pyrope garnets were also grouped according to derived Ni-in-garnet geothermometry temperatures (following methodology of Ryan et al. (1996); Fig. 13). The 800-950°C Ni-in-garnet temperature group contains 7 strongly sinusoidal traces, 1 MREE-depleted grain (15SVU026_r2g9 (G10)) and one HREE-enriched grain (15SUV028_r2g10 (G9)). Almost all LREE are above chondrite levels, while MREE and HREE trend towards chondrite-normal levels. Both of the wehrlitic G12 grains occur in this Ni-in-garnet temperature grouping. Fifteen grains are in the 950-1100°C Ni-in-garnet range, 1 of which is strongly MREE-depleted (16SUV025_r3g7 (G9)), 3 of which are low sinuosity (15SUV014_r2g3 (G10); 15SUV014_r2g4 (G10D); 16SUV025_r3g6 (G10D)), and the rest (n=11) HREE-enriched, with MREE and HREE trending towards 10x chondrite levels. The 1100-1250°C Ni-in-garnet group exhibits the greatest variability in REE trends, including two G10 grains (16SUV023_r2g4 and 16SUV023_r2g5m3) and a G9 (16SUV015_r1g6) which are the most divergent (HREE-enriched, high sinuosity, and normal, respectively), otherwise the rest of grains in this Ni-in-garnet temperature grouping (n=13) are weakly sinusoidal to HREE-enriched at levels, at or below those in the 950-1100°C. The 1250-1400°C group (n=8) are all G9 grains, and are generally the most HREE-enriched of all temperature groupings, and the most consistent in terms of REE trends. The 1400-1650°C group (n=10) is also generally HREE-enriched, but generally at or below that of the 1250-1400°C group, and with a higher variability in REE trends.

### 3.5.2 Spinel – Chromite

Chromites were the most abundant visually identified indicator mineral (n=215), and were present in 29 of 47 samples (Appendix 4C). Fifty-four of the grains were identified as euhedral, the rest as subeuhedral (resorbed) to distorted crystal shapes (Appendix 5), which as indicated by Averill (2011) can provide some discrimination between non-kimberlitic (angular, well defined crystal faces) and kimberlitic (resorbed surfaces, less obvious crystal faces) sources. Chemical compositions of the chromite grains are provided in Appendix 6C (EPMA) and 7B (ICPMS). Mantle-derived chromites exhibit wide variations in both major and minor element concentrations, and a number of geochemical indices have been developed to discriminate different 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 inclusion compositions (Fig. 14A-E; Sobolev, 1977; Fipke et al., 1989, 1995; Griffin et al., 1997;

McClenaghan and Kjarsgaard, 2007; Roeder and Schulze, 2008). Overall, only 1-3 of the Banks chromite-spinels plot within the variously defined diamond inclusion, or diamond intergrowth chromite fields (Fig. 14A, B, C). While 3 grains in the $TiO_2$ – $Cr_2O_3$ division of Fipke et al., 1995 (Fig. 14B) do fall within diamond inclusion and intergrowth fields, it is noted that in their analysis of kimberlitic, lamproitic, and non-diatreme sources, the range of values is significant, and there is a great deal of overlap between the populations, making discrimination uncertain. The spinel Mg# - Cr# plot (Fig. 14D; after Roeder and Schulze, 2008; Liu et al., 2018; El Dien et al. 2019) suggests approximately half of the Banks grains fall within the bounds of values from cratonic peridotites, although there is also significant overlap with non-cratonic peridotite compositions. It is further noted that the Banks Cr# values tend to be on the lower end of the cratonic spectrum, which is lower than most of those associated with till-derived spinels from central Victoria Island samples (cf., Liu et al., 2018). The $Cr_2O_3$ – $Al_2O_3$ and $Cr_2O_3$ – $TiO_2$ indices of Sobolev (1971, 1977; see also McClenaghan and Kjarsgaard, 2007; Fig. 14E) provide further discrimination of the chrome spinels, similarly indicating an absence of chromites from diamond inclusion fields, and a mixed population of potentially kimberlitic and non-diatreme derived grains.

### 3.5.3 Mg-Ilmenite

During the KIM picking process, ODM visually identified 99 Mg-ilmenite grains in the GSC Banks Island samples, 71 of which were classified as crustal Mg-ilmenite. Using EPMA, 4 of the crustal Mg-ilmenite grains were reclassified as pseudorutile, and 2 were reclassified as chromite. EPMA also reclassified 17 picked chromites as Mg-ilmenites, producing a total of 110 Mg-ilmenite grains (Appendix 6E). Two methods were used to discriminate kimberlitic from non-kimberlitic (i.e., crustal) Mg-ilmenites. In the bivariate plot of MgO versus $Cr_2O_3$ (Fig. 15A) non-kimberlitic ilmenites are identified as those with <4.0 wt% MgO. Of the total 110 ilmenites, 83 are classified as non-kimberlitic (includes 2 Beaufort Formation grains that were misidentified as ilmenite (non-crustal)), while 27 (~25%) were classified as kimberlitic (i.e., non-crustal; includes 2 stream sediment and 1 Beaufort Formation samples that were misidentified as crustal ilmenites). Using the $TiO_2$ versus $Cr_2O_3$ classification of Wyatt et al. (2004), 26 kimberlitic versus 84 non-kimberlitic grains are depicted (Fig. 15B). The Beaufort Formation samples yielded 12 kimberlitic versus 4 non-kimberlitic Mg-ilmenite grains (75%), although it is noted that 11 of the 12 kimberlitic Mg-ilmenite grains came from sample 16SUV023; only one other Beaufort Formation sample yielded a kimberlitic Mg-ilmenite grain (15SUV030), and sample 16SUV023 contained only 3 non-kimberlitic ilmenite (Appendix 6E). By comparison, only 16 kimberlitic Mg-ilmenites (from 7 different samples) were recovered from the 94 stream sediment sample ilmenite grains (17%; Appendix 6E).

### 3.5.4 Olivine

Visually identified olivine grains that were analyzed by EPMA include forsterite, fayalite, diopside and bronzite (Appendix 6F). Olivines can be derived from kimberlite, but can also be sourced from basaltic rocks; resolving between the two is important. Considering only the forsterite grains (n=24), the Banks Island samples yielded lower Mg# (average 0.83) than those typically associated with kimberlite xenocrysts and diamond inclusions (Mg# 0.89-0.94; Bussweiler et al., 2015, 2017; Stachel and Harris, 2008; Fig. 16). Similarly, the Banks Island olivines generally have elemental compositions greater than those associated with kimberlitic olivines which generally have chemical compositions of: CaO $\leq$ 0.1wt%, MnO $\leq$ 0.15 wt%, an $dCr_2O_3 \leq 0.03$ wt% (Appendix 6F; Fipke et al., 1995; Bussweiler et al., 2015, 2017). Only 1 forsteritic olivine was recovered from a Beaufort Formation deposit (15SUV030), and only 1 forsterite grain from stream sediment sample 15SUV020 plots within the bounds of Fipke et al.'s (1995) kimberlite inclusion field (Fig. 16).

### 3.5.5 Cr-diopside

There were 26 clinopyroxene grains picked by ODM from the Banks Island KIM samples (Appendix 6G). These included 2 Cr-diopsides, 20 low Cr-diopsides, and 4 diopside grains. Six of the low Cr-diopsides were recovered from 2 Beaufort Formation samples. The clinopyroxene grains are generally Mg-poor with Mg# (100Mg/(Mg+$Fe^{2+}$)) <88, with only 4 considered to be Mg-rich (Mg# ≥88). Concentrations of $Cr_2O_3$ vary between 0.33 – 1.28 wt% (Fig. 17). Cr-rich diopside 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). None of Banks Island Cr-diopside plot within these parameters. However, kimberlites also contain diopside with <1.5 wt% $Cr_2O_3$, and a variety of discrimination plots have been utilized to characterize these (cf., Schulze, 1987; McCandless and Gurney, 1989; Nimis, 1998; Morris et al., 2002). Based on these various plots, only 1 of the Banks Island Cr-diopside grains (15SUV014; a stream sediment sample) plots within the bounds of peridotite mantle assemblages (Fig. 17). The other Cr-diopside (15SUV024; a fluvial gravel deposit) with a Mg# of 88 and $Cr_2O_3$ wt% of 0.6, is interpreted to be from mantle lherzolite (McClenaghan et al., 2006). The low Cr-diopside grains largely plot within, or above, the websteritic assemblage, with 4 grains in the megacryst assemblage.

# 4. Discussion

## 4.1 Beaufort Formation and the Potential for KIM Bedrock Inheritance

Six of the 7 Beaufort Formation (*sensu lato*) samples contained a total of 131 KIMs of different mineralogy, many of which display what are interpreted as kimberlitic chemical compositions (Appendix 4; Fig. 7-15). The presence of KIMs in these samples indicates that during the time at which the fluvial Beaufort Formation sediments were being deposited by ancient rivers, crossing what was then a contiguous arctic coastal plain (i.e., no arctic islands or inter-island channels existed), exposed kimberlites, or secondarily reworked kimberlitic materials were being actively eroded and redeposited. Beaufort Formation deposits on Banks Island are reported to have a general westerly azimuth, and contain rare granite clasts (Hills and Fyles, 1973; Miall, 1979). These characteristics were similarly observed in outcrops and searches of surface deflation lags at all such sites during the course of this project's fieldwork (Fig. 5). The granite erratics are clearly far travelled; the closest granitic outcrop to Banks Island is a granodiorite in southern Hadley Bay on northern Victoria Island (Thorsteinsson and Tozer, 1962; 450 km east of 16SUV023), while granitic terrain of the Canadian Shield outcrops extensively on western Somerset Island, Boothia Peninsula, and throughout the continental mainland fringing southern Victoria Island and extending westward as far as Great Bear Lake (Fig. 5 – see inset Canada map). Kimberlites are known to occur both within and beyond margins of exposed Canadian Shield granitic terrain on western Somerset Island (850 km away), central and southern Bathurst Peninsula (>1000 km), southeastern Victoria Island (550 km away), areas around Great Bear Lake (>700 km; including 13 on Parry Peninsula; 500 km away), and abundantly further south through the Slave Craton (Stubley and Irwin, 2019).

While locations of known kimberlites and granite outcrops coincide, it seems unlikely that KIMs in Beaufort Formation deposits on Banks Island were derived from such distant kimberlite sources. KIMs in the Beaufort Formation deposits are small, e.g., no KIMs were recovered in the 1-2 mm size fraction, and are rare (6.3%) in the 0.5-1.0 mm size fraction; the same, however, holds true for all stream and glacial sediment samples collected on Banks Island. While transport-related grain abrasion and comminution would result in smaller KIM grain sizes, the travel distances to known kimberlites appears too great to account for the mineralogical suite and abundances of KIMs recovered (cf., Averill and McClenaghan, 1994; Dredge et al., 1996; McClenaghan, 2005; McClenaghan and Kjarsgaard, 2007; Cummings et al., 2014). Consider that the combined glacial and fluvial dispersal trains of pyrope garnets in Attawapiskat (ON) and Lac de Gras (NWT) kimberlite fields are ~300 and 180 km, respectively (McClenaghan, 2005). The site at which sample 16SUV023 was collected (surface dug pit) is a prominent Beaufort Formation fluvial peneplain, that unconformably overlies Eureka Sound Formation sediments, extends 0.4 x 2 km, and for which lateral equivalent surfaces can be seen extending northward on the east side of Thomsen River valley, as a series of discontinuous, uplands capped by conspicuously orange-covered sandy-gravel (Fig. 5, 18).

The 16SUV023 sample is particularly noteworthy. It contains the highest abundance of KIMs of any sample collected on Banks Island by either the GSC or industry (n=140; more than twice that of the next highest sample; Fig. 4C; Appendix 4). While most KIM samples on Banks Island are dominated by chromites, this sample contains a majority of garnets, at abundances more than an order of magnitude greater than any other sample. The potential bias of fluvial concentration of heavy minerals also does not appear to be the source of this anomalous abundance. This sample has markedly low HMC weights, and when normalized to a nonferromagnetic HMC 0.25-0.5 mm picking fraction weight of 50 g, its dominance is only magnified to 6 times greater than the next most abundant sample (Appendix 4E). It is noted though, that this sample contains the most gold grains (9) of any of the GSC Banks Island samples (Appendix 4I), suggesting some degree of heavy mineral concentration. Cummings et al. (2014) in their comminution experiments indicate that the tendency for kimberlitic pyrope garnets to fracture may increase their relative abundance compared to other KIMs down flow from a source. As discussed previously, the recovered pyrope garnet grains do exhibit a sub-angular to sub-rounded morphology, but the prominence of sub-kelyphitic "orange-peel" texture in sample 16SUV023 (10 of 25 garnets in the 0.25-0.5 mm fraction and 6 of 18 garnets in the 0.18-0.25 mm fraction; Appendix 5) suggests a relative lack of abrasion, and hence a proximal source, not one several hundreds to >1000 km away. It is suggested, therefore, that sample 16SUV023 records the presence of a proximal Banks Island kimberlite source. Intentions to collect additional samples of Beaufort Formation deposits from the area were lost due to helicopter mechanical issues. It is useful to note though that 9 stream sediment samples collected by Rio Tinto in areas 30-50 km south of 16SUV023 returned zero KIMs (Fig. 1, 4C).

The presence of KIMs within unconsolidated fluvial Beaufort Formation deposits, and the identification of previously undocumented erosional remnant Beaufort Formation outliers on northeastern and central Banks Island clearly represents a potential of bedrock inheritance of KIMs that must be considered in models of subsequent glacial erosion and dispersal, and the interpretation of regional sample collections. These and other considerations of glacial history on Banks Island are discussed below.

## 4.2 Glacial History and Potential KIM Dispersal

During the last glaciation (Late Wisconsinan), the Laurentide Ice Sheet dispersed flow westward across the western Canadian arctic from a dome centered in the Keewatin area of northwest Hudson Bay. At maximum glacial extent, the prominent M'Clintock Ice Divide formed on eastern Victoria Island and as depicted by Dyke and Prest (1987; in accordance with the glacial reconstructions of Vincent (1983) and others) was suggested to feed flow westerly across all of Victoria Island onto the eastern rim of Banks Island (Fig. 19a). Limited extent ice streams occupied central Amundsen Gulf and M'Clure Strait, terminating in ice shelves. The only known kimberlite this maximum glacial extent ice flow reconstruction would have encountered is the Snow Goose cluster on east-central Victoria Island. Generalized ice flow patterns would not have carried this north of Johnson Point, Banks Island (Fig. 19a). On Banks Island, this flow pattern would potentially entrain and disperse locally sourced bedrock and sediments in a west-northwest fashion, largely perpendicular to the coastline (Fig. 19a), but would not correlate with the majority of positive KIM samples that were collected west of the Jesse moraine belt (Fig. 4).

During deglaciation, Dyke and Prest (1987) depicted progressively increasing topographic confinement of flow on Victoria Island, but most of this did not extend onto Banks Island, and only then along the very periphery of the eastern coastal region. Ice flow histories for Vincent's (1982, 1983) conjectured older (pre-Late Wisconsinan) glaciations and tills (Fig. 2) are unknown in terms of source area, although they can be presumed to have drawn from ice sheets similarly centered in the Keewatin, central Canadian arctic mainland, or arctic islands (cf., Barendregt and Duk-Rodkin, 2004). It seems likely that regardless of source of former ice sheets, the Amundsen Gulf basin would have served as an effective draw down for ice westward towards the Beaufort Sea, rather than advancing perpendicularly northwards across it and onto Banks Island. This would then rule out redistribution of KIMs from Parry Peninsula kimberlites, or those in and around Great Bear Lake and the central and southern Slave Craton. Domes and ice divides from earlier glaciations forming east of the Late Wisconsinan M'Clintock Ice Divide, could potentially redistribute KIMs from sites on Somerset Island, Boothia Peninsula, and the northern Slave Craton. A very limited collection of four samples of the lowermost (oldest, ? >0.78Ma) till at the base of Morgan Bluffs (Lakeman and England, 2014) and Duck Hawk Bluffs (Evans et al., 2014) returned no KIMs, and only 1 of 3 associated glaciofluvial deposits (Morgan Bluffs) yielded any KIMs (3 garnets and 1 Cr-diopside) which do not geochemically differ significantly from other Banks Island KIMs (Appendix 4, 6).

The recent revisions of Late Wisconsinan glacial history on Banks Island (England et al., 2009; Lakeman and England, 2012, 2013) clearly document pervasive Laurentide Ice Sheet cover over Banks Island (Fig. 19B), and prominent ice streams in M'Clure Strait and Amundsen Gulf out onto the continental margin. This establishes the potential for last glacial redistribution of KIMs from sites across Victoria and Banks islands. Hypothetical ice flow patterns at maximum glacial extents sees ice flow across the Victoria Island Snow Goose kimberlite cluster (550 km away) that is carried across central Banks Island (Fig. 19B). More importantly, it is recognized (Stokes et al., 2009; Lakeman and England, 2012) that during deglaciation, a shifting series of local topographic divides began to emerge along the Shaler Mountains and peripheral highland areas on Victoria Island at times when a contiguous ice flow across the eastern half of Banks Island continued (Lakeman and England's (2012) Thomsen Phase ice flow; Fig. 3). This progressive shift in ice flow patterns provides for several potential geographical sources of erratic material from Victoria Island to be carried onto central, northern and southern Banks Island (Fig. 19C, D).

Lakeman and England (2012) indicate that much of the last glacial cover of Banks Island occurred under cold-based ice. Under such conditions, quarrying and entrainment of debris is limited. However, while this may describe the situation under full-glacial conditions (Lakeman and England's (2012) Beaufort Phase ice flow; Fig. 3), the presence of glacially streamlined landforms and flow sets of these features on Victoria Island (Stokes et al., 2009) clearly identifies warm-based glacier conditions at times when westward ice flow off the Shaler Mountain divide(s) extended across eastern Banks Island, conforming to the deglacial landsystem of Dyke and Evans (2003). This provides a mechanism for erosion and entrainment of potentially unknown kimberlitic sources on the Prince Albert Peninsula of northwestern Victoria Island, and their dispersal westward onto Banks Island. Lakeman and England (2012) characterize the Thomsen Phase of ice flow as largely cold-based, but progressively lobate flow points to warm, or at least polythermal conditions, and entrained debris within the ice would have been deposited as a thin, discontinuous till cover, outwash deposits, or controlled moraines. Areas of Vincent's (1982, 1983) Baker Till (Fig. 2), extensively reside within Lakeman and England's (2012, 2013) Thomsen phase of ice flow, and are reported to have tills comprised of sand, silt and clay derived from underlying Cretaceous bedrock, with rare erratic clasts of quartzite, granite and limestone (ranging from pebbles to cobbles and rare boulders; Lakeman and England, 2012). As ice continued to retreat under polythermal conditions, increasing opportunities for local, Banks Island-sourced bedrock erosion and entrainment of debris may have occurred. However, this may have occurred only during the waning stages of regional deglaciation (between 14 – 12.75 ka cal yr BP; Lakeman and England, 2012), providing both a short temporal window, and more importantly in terms of potential KIM redistribution from a hypothetical Banks Island kimberlitic source, across a relatively short (<40 km) dispersal path. Most of the KIM samples on northeastern Banks Island lying beyond the limits of the Jesse moraine belt occur beyond the limits of Lakeman and England's (2012) mapping. Field observations from this study recorded similar muted glacial landforms (controlled moraines) and outwash deposits within 20 km of the Jesse moraine belt, and then little evidence beyond a scattering of erratics over much of the Devonian carbonate uplands (areas of Vincent's (1982, 1983) Plateau and southeastern Baker tills; Fig. 2).

Field observations also record the presence of abundant kames in the Thomsen River valley and eastern Banks Island (beyond and within the Jesse moraine belt). These depositional features commonly form along steep-fronted margins of polythermal glaciers, where supraglacial streams incised debris-rich bands of ice at the glacier margin, forming conspicuous conical piles of sorted sediments, in places >10 m high (Fig. 20). Field investigations of magnetic anomalies in Diamonds North's aeromagnetic survey, and their identification of ~150 magnetic "features" (Jober and Vanderspiegel, 2008), demonstrated that many of these corresponded to kames and associated outwash deposits. Presumably the fluvial nature of the kames is providing a mechanism for concentrating iron-rich minerals beyond background concentrations in regional low magnetic signature bedrock or discontinuous till cover. Several of the higher KIM abundance stream sediment samples are found in proximity to individual and clusters of kame deposits (e.g., Fig. 4C, D, west of Johnson Point; industry's 49 chromite, 7 pyrope garnet, 5 Mg-ilmenite sample, and the adjacent GSC sample 16SUV028).

As discussed, the Jesse moraine belt is conspicuously different from all other terrain on Banks Island. It has the thickest till, a dramatic increase in abundance and size of erratic boulders (particularly mafic basalt and gabbro derived from the Minto Inlier and Shaler Mountains on Victoria Island; Lakeman and England, 2012), and preserves widespread buried glacial ice (Smith and Farineau, 2015; Rudy et al., 2017). The Jesse moraine belt is evidence of a fundamentally different style of glaciation, and as suggested by Lakeman and England (2012), correlates to a prominent stillstand or re-advance of warm-based ice northwards up Prince of Wales Strait. It is also likely the first time during the Late Wisconsinan glaciation that ice draining from the Minto Inlet area of western Victoria Island was advected onto northeastern Banks Island (north of Johnson Point; Fig. 19D). The importance of these observations is that characteristics of KIMs or glacial sediments from samples collected inside the Jesse moraine belt, cannot be extended to glacial deposits beyond its limits (excepting those of the glaciofluvial outwash deposits in the major rivers that cross Banks Island and whose headwaters emanate from the edge of the Jesse moraine belt). Further, any potential kimberlitic-bearing terrain or bedrock along eastern Banks Island that may have been glacially eroded and redeposited westward prior to the formation of the Jesse moraine belt, now lies largely concealed below its cover – i.e., it would be unlikely to find a dispersal train leading from deposits within the Jesse moraine belt to those situated west (glacially down flow) of its limits.

### 4.3 Bedrock Geology and Kimberlite Preservation Potential

Some consideration needs to be given for the types of bedrock and their degree of lithification in terms of what the topographic expression a kimberlite erupting through these rocks might have in the Banks Island landscape today. As discussed, >80% of the island is covered by poor to weakly lithified Mesozoic and Cenozoic sediments (Fig. 1). Any kimberlite erupting prior to the Mesozoic would likely be deeply buried in all areas of the island except for the northeastern quadrant where the Devonian Mercy Bay strata extensively outcrops (Fig. 1). If a kimberlite had erupted through the well-lithified Mercy Bay strata then it could either form a positive relief feature (more resistant to weathering/erosion), or a recessive feature (i.e., lake/pond; less resistant to weather/erosion). Elsewhere on Banks Island, any kimberlite outcropping at surface would have to be Cretaceous or younger, and in most areas would still likely be extensively buried by the generally poorly lithified Paleocene-Eocene Eureka Sound Formation strata (Fig. 1). Where the Eureka Sound Formation cover has been eroded away (fluvial/glacial), then it would be expected that given the poor to weakly-lithified nature of underlying Cretaceous bedrock, a kimberlite would tend to have positive relief (more resistant to weathering/erosion). Given how generally subdued the landscape is across most of Banks Island, a conspicuous positive topographic landform, of a possible kimberlitic rock type, has never been identified throughout its historical exploration, nor in Diamond North's extensive survey of magnetic anomalies in the northeastern quadrant (excepting what are now recognized as glacial kames; Fig. 20). Only in areas of thick glacial cover in the Jesse moraine belt and glaciofluvial outwash in the Thomsen and other major river valleys could modern sediment conceal a kimberlite.

## 5. Conclusions

Abundant chemical and morphological evidence indicates the heavy mineral suite picked from both stream sediment and Beaufort Formation samples on Banks Island contain indicator minerals that are kimberlitic in origin. While the Ni-in-garnet geothermometry suggests the potential for entrainment of minerals and diamonds by kimberlitic magma within the diamond stability window, the wide range of paleotemperatures, and particularly the higher spread of temperatures in the Beaufort Formation chrome pyropes (>1200°C) suggests that were all such garnets to be entrained by a single kimberlitic source, than much of the diamond potential could be reduced, or at least have a diminished capacity for preservation of larger diamonds. If multiple kimberlitic sources existed (i.e., those sampled by Beaufort Formation fluvial environments), then different eruption histories and diamond potential could also exist. Similarly, kimberlites sampled by Beaufort Formation fluvial environments may differ spatially from those eroded and redeposited by later glaciations.

In the context of resolving potentially different kimberlitic sources, the relatively small number of samples and recovered KIMs may complicate attempts at discerning geochemical differences between KIMs collected from different sample media (Beaufort Formation vs. stream sediment vs. glacial sediments), and from different locations on Banks Island. Integration of this study's data with data sets from industry, including those from the Snow Goose kimberlite cluster on Victoria Island and those from Parry Peninsula (Liu et al., 2019), will address this small number of samples issue in a subsequent publication. Clearly, the abundance of KIMs collected by Rio Tinto on Prince Albert Peninsula on northwestern Victoria Island (Fig. 4C) suggests that this is an area of interest for future diamond exploration as it may signal the presence of unknown proximal kimberlites, there or to the east, and may be a potential source for fluvially or glacially-dispersed KIMs on northeastern Banks Island. At the same time, a different kimberlitic source remains to be identified for KIMs recovered on southeastern Banks Island (Nelson Head). As discussed, the anomalous 16SUV023 Beaufort Formation sample suggests a proximal kimberlite source(s) on Banks Island. The 2 clusters of G10D Cr-pyrope garnets on northeastern Banks Island (GSC and industry data; Fig. 4C), and the presence of usually fast-weathering forsterite grains argues for the potential of diamondiferous kimberlite source(s) either locally on eastern Banks Island, or perhaps, from somewhere on northwestern Victoria Island.

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## 7. References

Aerodat Inc., 1996. Assessment Report on a fixed wing high sensitivity magnetic survey, Banks Island, N.W.T., NTS 98 D/10, 15 and 98 E/2, 6, 7, 10. Monopros Limited. 15 pages + figures. NWT Assessment Report #083984

Aerodat Inc., 1997. Report on a fixed wing high sensitivity magnetic survey, Ballast Brook, Banks Island, N.W.T., Tahoe Lake, Victoria Island, N.W.T. Monopros Limited. 24 pages + figures. NWT Assessment Report #083984

Afanas'ev, V.P., Varlamov, V.A., and Garanin, V.K., 1984. Dependence of the abrasion of kimberlite minerals from conditions and distance of transportation. Geologiya i Geofizika, n. 10, p. 119-124.

Andrews, J.T. and Peltier, W.R., 1989. Quaternary geodynamics in Canada; *in* Chapter 8 of Quaternary Geology of Canada and Greenland, (ed.) R.J. Fulton; Geological Survey of Canada, Geology of Canada, no. 1, p. 541-572. doi:10.4095/127905

Armstrong, J.T., 1995. CITZAF: a package of correction programs for the quantitative Electron Microbeam X-ray-analysis of thick polished materials, thin-films, and particles; Microbeam Analysis, v. 4, p. 177-200.

Averill, S.A., 2001. The application of heavy indicator mineralogy in mineral exploration with emphasis on base metal indicators in glaciated metamorphic and plutonic terrains; *in* Drift Exploration in Glaciated Terrain, (eds.) M.B. McClenaghan, P.T. Bobrowsky, G.E.M. Hall, and S.J. Cook; Geological Society of London, Special Publication 185, p. 69-81.

Averill, S.A., 2011. Viable indicator minerals in surficial sediments for two major base metal deposit types: Ni-Cu-PGE and porphyry Cu; Geochemistry: Exploration, Environment, Analysis, v. 11, p. 279-291. doi:10.1144/1467-7873/10-IM-022

Averill, S.A. and McClenaghan, M.B., 1994. Distribution and character of kimberlite indicator minerals in glacial sediments, C14 and Diamond Lake kimberlite pipes, Kirkland Lake, Ontario; Geological Survey of Canada, Open File 2819, 52 p. doi:10.4095/193980

Banas, A., Stachel, T., Phillips, D., Shimizu, N., Viljoen, K.S., and Harris, J.W., 2009. Ancient metasomatism recorded by ultra-depleted garnet inclusions in diamonds from DeBeers Pool, South Africa; Lithos, v. 112S, p. 736-746. doi:10.1016/j.lithos.2009.04.043

Barendregt, R.W. and Duk-Rodkin, A., 2004. Chronology and extent of Late Cenozoic ice sheets in North America: a magnetostratigraphic assessment; in Quaternary Glaciations – Extent and Chronology, Part II: North America, (eds.) J. Ehlers and P.L. Gibbard; Elsevier, Amsterdam; p. 1-7.

Barendregt, R.W. and Vincent, J.-S., 1990. Late Cenozoic paleomagnetic record of Duck Hawk Bluffs, Banks Island, Canadian Arctic Archipelago; Canadian Journal of Earth Sciences, v. 27, p. 124-130. doi:10.1139/e90-010

Barendregt, R.W., Vincent, J.-S., Irving, E., and Baker, J., 1998. Magnetostratigraphy of Quaternary and late Tertiary sediments on Banks Island, Canadian Arctic Archipelago; Canadian Journal of Earth Sciences, v. 35, p. 147-161. doi:10.1139/e97-094

Bussweiler, Y., Foley, S.F., Prelević, D., and Jacob, D.E., 2015. The olivine macrocryst problem: new insights from minor and trace element compositions of olivine from Lac de Gras kimberlites, Canada; Lithos, v. 220, p. 238-252. doi:10.1016/j.lithos.2015.02.016

Bussweiler, Y., Brey, G.P., Pearson, D.G., Stachel, T., Stern, R.A., Hardman, M.F., Kjarsgaard, B.A., and Jackson, S.E., 2017. The aluminium-in-olivine thermometer for mantle peridotites – experimental versus empirical calibration and potential applications; Lithos, v. 272-273, p. 301-314. doi:10.1016/j.lithos.2016.12.015

Canil, D., 1999. The Ni-in-garnet geothermometer: calibration at natural abundances; Contributions to Mineralogy and Petrology, v. 136, p.240-246. doi: 10.1007/s004100050535

Cassan, J.P. and Evers, J.H., 1973. The Cretaceous Isachsen, Christopher and Hassel Formations on Banks Island, N.W.T. – a sedimentological interpretation; Geological Association of Canada and Canadian Society of Petroleum Geology, Symposium Geology of the Canadian Arctic, Program and Abstracts, Saskatoon, Canada.

Cox, R. and Barnes, S-J., 2005. LA-ICP-MS analysis of garnet: Trace-element characterization and single-mineral thermometry applied to diamond exploration. Projet DIVEX SC17, Rapport Final. 14 p.

Craig, B.G. and Fyles, J.G., 1960. Pleistocene geology of Arctic Canada; Geological Survey of Canada, Paper 60-10, 21 p. doi:10.4095/101191

Craig, B.G. and Fyles, J.G., 1965. Quaternary of Arctic Canada; *in* Antropogenovye period v Arktike i Subarktike; Naucho-Issl. Inst. Geol. Arktiki, Moscow, Trudy 143, p. 5-33 (in Russian with English summary).

Cummings, D.I., Kjarsgaard, B.A., Russell, H.A.J., and Sharpe, D.R., 2014. Comminution of indicator minerals in a tumbling mill: implications for mineral exploration; Geological Survey of Canada, Open File 7111, 18 p. doi:10.4095/293467

Day, S.J.A., Paulen, R.C., Smith, I.R., and King, R.D., 2018. Heavy-mineral and indicator-mineral data for stream sediments in southwest Northwest Territories: new potential for undiscovered mineralization. Geological Survey of Canada, Open File 8362, 24 p. doi:10.4095/306587

Deer, W.A., Howie, R.A., and Zussman, J., 1982. Rock Forming Minerals, Volume 1A, Orthosilicates, Second Edition. Longman Group Ltd., New York, USA, 579 p.

Donovan, J.J., Kremser, D., Fournelle, J.H., and Goemann, K., 2015. Probe for EPMA: Acquisition, automation and analysis, version 11: Eugene, Oregon, Probe Software, Inc., www.probesoftware.com

Dredge, L.A., Ward, B.C., and Kerr, D.E., 1996. Morphology and kelyphite preservation on glacially transported pyrope grains; *in* Searching for Diamonds in Canada, (eds.) A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson; Geological Survey of Canada, Open File 3228, p. 197-204. doi:10.4095/211755

Dyke, A.S., 1987. A reinterpretation of glacial and marine limits around the northwestern Laurentide Ice Sheet; Canadian Journal of Earth Sciences, v. 24, p. 591-601. doi:10.1139/e87-058

Dyke, A.S., 1993. Landscapes of cold-centred Late Wisconsinan ice caps, Arctic Canada; Progress in Physical Geography, v. 17, p. 223-247. doi:10.1177/030913339301700208

Dyke, A.S. and Evans, D.J.A., 2003. Ice-marginal terrestrial landsystems: northern Laurentide and Innuitian ice sheet margins; *in* Glacial Landsystems, (ed.) D.J.A. Evans; Arnold, London, p. 143-165.

Dyke, A.S. and Prest, V.K., 1987. Late Wisconsinan and Holocene history of the Laurentide Ice Sheet; Géographie physique et Quaternaire, v. 41, p. 237-263. doi :10.7202/032681ar

Ecosystem Classification Group, 2013. Ecological regions of the Northwest Territories – Northern Arctic; Department of Environment and Natural Resources, Government of the Northwest Territories, Yellowknife, NT, Canada. 157 p. + insert map

El Dien, H.G., Arai, S., Doucet, L-S., Li, Z-X., Kil, Y., Fougerouse, D., Reddy, S.M., Saxey, D.W., and Hamdy, M., 2019. Cr-spinel records metasomatism not petrogenesis of mantle rocks; Nature Communications, 10:5103. 12 p. doi:10.1038/s41467-019-131177-1

Embry, A.F. and Klovan, J.E., 1971. A Late Devonian reef tract on northeastern Banks Island, N.W.T.; Bulletin of Canadian Petroleum Geology, v. 19, p. 730-781. doi: 10.35767/gscpgbull.19.4.730

Embry, A.F. and Klovan, J.E., 1974. The Devonian clastic wedge of the Canadian Arctic Archipelago; Geological Society of America Annual Meeting, Abstracts with Programs, p. 721-722.

England, J.H. and Furze, M.F.A., 2008. New evidence from the western Canadian Arctic Archipelago for the resubmergence of Bering Strait; Quaternary Research, v. 70, p. 60-67. doi:10.1016/j.yqres.2008.03.001

England, J.H. and Furze, M.F.A., 2011. The Viscount Melville Sound Ice Shelf of the NW Laurentide Ice Sheet: a story befitting the seminal contributions of Art Dyke; 41st International Arctic Workshop, Program and Abstracts, p. 94-96; Université du Québec à Montréal, Montréal, Canada, March 2-4.

England, J.H., Furze, M.F.A. and Doupé, J.P., 2009. Revision of the NW Laurentide Ice Sheet: implications for paleoclimate, the northeast extremity of Beringia, and Arctic Ocean sedimentation; Quaternary Science Reviews, v. 28, p. 1573-1596. doi:10.1016/j.quascirev.2009.04.006

Evans, D.J.A., 2009. Controlled moraines: origins, characteristics and palaeoglaciological implications; Quaternary Science Reviews, v. 28, p. 183-208. doi:10.1016/j.quascirev.2008.10.024

Evans, D.J.A., England, J.H., LaFarge, C., Coulthard, R.D., Lakeman, T.R., and Vaughan, J.M., 2014. Quaternary geology of the Duck Hawk Bluffs, southwest Banks Island, Arctic Canada: a re-investigation of a critical terrestrial type locality for glacial and interglacial events bordering the Arctic Ocean; Quaternary Science Reviews, v. 91, p. 82-123. doi:10.1016/j.quascirev.2014.03.011

Fipke, C.E., Gurney, J.J., Moore, R.O., and Nassichuk, W.W., 1989. The development of advanced technology to distinguish between diamondiferous and barren diatremes; Geological Survey of Canada, Open File 2124, 90 p. doi:10.4095/130799

Fipke, C.E., Gurney, J.J., and Moore, R.O., 1995. Diamond exploration techniques emphasizing indicator mineral geochemistry and Canadian examples; Geological Survey of Canada, Bulletin 423, 80 p. doi:10.4095/205728

French, H.M., 1974. Active thermokarst processes, eastern Banks Island, western Canadian Arctic; Canadian Journal of Earth Sciences, v. 11, p. 785-794. doi:10.1139/e74-078

Fyles, J.G., 1962. Physiography in Banks, Victoria and Stefansson islands, Arctic Archipelago; Geological Survey of Canada, Memoir 330, p. 8-17. doi:10.4095/100554

Fyles, J.G., 1965. Surficial Geology, western Queen Elizabeth Islands; Geological Survey of Canada, Report of Activities, Paper 65-01, p. 3-5. doi:10.4095/121948

Fyles, J.G., 1990. Beaufort Formation (Late Tertiary) as seen from Prince Patrick Island, Arctic Canada; Arctic, v. 43, p. 393-403.

Garvie, O. and Robinson, D., 1984. The formation of kelyphite and associated sub-kelyphitic and sculptured surfaces on pyrope from kimberlite; *in* Proceedings of the 3rd Kimberlite Conference, (ed.) J. Kornprobst, Elsevier, Amsterdam. Developments in Petrology, v. 11A, p. 371-382.

Grasby, S.E., Smith, I.R., Bell, T., and Forbes, D.L., 2013. Cryogenic formation of brine and sedimentary mirabilite in submergent coastal lake basins, Canadian Arctic; Geochimica et Cosmochimica Acta, v. 110, p. 13-28. doi: 10.1016/j.gca.2013.02.014

Griffin, W.L. and Ryan, C.G., 1995. Trace elements in indicator minerals: area selection and target evaluation in diamond exploration; Journal of Geochemical Exploration, v. 53, p. 311-337. doi:10.1016/0375-6742(94)00015-4

Griffin, W.L., Cousens, D.R., Ryan, C.G., Sie, S.H., and Suter, G.F., 1989. Ni in chrome pyrope garnets: a new geothermometer; Contributions to Mineralogy and Petrology, v. 103, p. 199-202. doi:10.1007/BF00378505

Griffin, W.L., Smith, D., Ryan, C.G., O'Reilly, S.Y, and Win, T.T., 1996. Trace-element zoning in mantle minerals: metasomatism and thermal events in the upper mantle; Canadian Mineralogist, 34(6): 1179-1193.

Griffin, W.L., Fisher, N.I., Friedman, J.H., and Ryan, C.G., 1997. Statistical techniques for the classification of chromites in diamond exploration; Journal of Geochemical Exploration, v. 59, p. 233-249. doi:10.1016/S0375-6742(97)00015-0

Griffin, W.L., Shee, S.R., Ryan, C.G., Win, T.T., and Wyatt, B.A. 1999. Harzburgite to lherzolite and back again: metasomatic processes in ultramafic xenoliths from the Wesselton kimberlite, Kimberley, South Africa; Contributions to Mineralogy and Petrology, v. 134, p. 232-250. doi:10.1007/s004100050481

Grütter, H.S., Gurney, J.J., Menzies, A.H., and Winter, F., 2004. An updated classification scheme for mantle-derived garnet, for use by diamond explorers; Lithos, v. 77, p. 841-857. doi:10.1016/j.lithos.2004.04.012

Hadlari, T., Midwinter, D., Galloway, J.M., Dewing, K., and Durbano, A.M., 2016. Mesozoic rift to post-rift tectonostratigraphy of the Sverdrup Basin, Canadian Arctic; Marine Petroleum Geology, v. 76, p. 148-158.

Hardman, M.F., Pearson, D.G., Stachel, T., and Sweeney, R.J., 2018a. Statistical approaches to the discrimination of crust- and mantle-derived low-Cr garnet – Major-element based methods and their application in diamond exploration; Journal of Geochemical Exploration, v. 186, p. 24-35. doi:10.1016/j.gexplo.2017.11.012

Hardman, M.F., Pearson, D.G., Stachel, T., and Sweeney, R.J., 2018b. Statistical approaches to the discrimination of mantle- and crust-derived low-Cr garnets using major and trace element data; Mineralogy and Petrology, v. 112 no. 2, supp., p. S697-S706. doi:10.1007/s00710-018-0622-7

Harrington, C.R., 2005. The eastern limit of Beringia: mammoth remains from Banks and Melville islands, Northwest Territories; Arctic, v. 58, p. 361-369.

Harrison, J.C., Ford, A., Miall, A.D., Rainbird, R.H., Hulbert, L.J., Christie, R.L., and Campbell, F.H.A., 2013. Geology, Tectonic assemblage map of Aulavik, Banks Island and northwestern Victoria Island, Northwest Territories; Geological Survey of Canada, Canadian Geoscience Map 35 (preliminary), scale 1:500 000. doi:10.4095/292833

Harrison, J.C., Lynds, T., Ford, A., and Miall, A.D., 2014. Geology, Tectonic assemblage map of M'Clure Strait, Prince Patrick, Eglinton, and surrounding islands, Northwest Territories; Geological Survey of Canada, Canadian Geoscience Map 31 (preliminary), scale 1:500 000. doi:10.4095/292828

Hills, L.V., 1969. Beaufort Formation, northwestern Banks Island, District of Franklin; *in* Report of Activities Par A, April to October 1968; Geological Survey of Canada, Paper 69-1A, p. 204-207. doi:10.4095/106000

Hills, L.V. and Fyles, J.G., 1973. The Beaufort Formation, Canadian Arctic Islands (abs.); *in* Canadian Society of Petroleum Geology and Geological Association of Canada, Program and Abstracts, Symposium on the Geology of the Canadian Arctic, Saskatoon, Canada.

Hobbs, W.H., 1945. The boundary of the latest glaciation in Arctic Canada; Science, v. 101, no. 2631, p. 549-551. doi: 10.1126/science.101.2631.549

Hodgson, D.A., 1994. Episodic ice streams and ice shelves during retreat of the northwesternmost sector of the late Wisconsinan Laurentide Ice Sheet over the central Canadian Arctic Archipelago; Boreas, v. 23, p. 14-28. Doi:10.1111/j.1502-3885.1994.tb00582.x

Hodgson, D.A. and Vincent, J-S., 1984. A 10,000 yr B.P. extensive ice shelf over Viscount Melville Sound, Arctic Canada; Quaternary Research, v. 22, p. 18-30. doi:10.1016/0033-5894(84)90003-6

Jakobsson, M., Mayer, L.A., Coakley, B., Dowdeswell, J.A., Forbes, S., Fridman, B., Hodnesdal, H., Noormets, R., Pedersen, R., Rebesco, M., Schenke, H.-W., Zarayskaya, Y., Accettella, A.D., Armstrong, A., Anderson, R.M., Bienhoff, P., Camerlenghi, A., Church, I., Edwards, M., Gardner, J.V., Hall, J.K., Hell, B., Hestvik, O.B., Kristoffersen, Y., Marcussen, C., Mohammad, R., Mosher, D., Nghiem, S.V., Pedrosa, M.T., Travaglini, P.G., and Weatherall, P., 2012. The International Bathymetric Chart of the Arctic Ocean (IBCAO) Version 3.0; Geophysical Research Letters, v. 39, L12609. doi:10.1029/2012GL052219

Jefferson, C.W., Scoates, R.F.J., and Smith, D.R., 1988. Evaluation of the regional non-renewable resource potential of Banks and northwestern Victoria Islands, Arctic Canada; Geological Survey of Canada, Open File 1695, 63 p. doi:10.4095/130491

Jenness, J.L., 1952. Problem of glaciation in the western islands of Arctic Canada; Bulletin, Geological Society of America, v. 63, p. 939-952. doi:10.1130/0016-7606(1952)63[939:POGITW]2.0.CO;2

Jober, S. and Vanderspiegel, R., 2008. Banks Island Project, Airborne Geophysical Survey and Till / Stream Sediment Sampling; Diamonds North Resources Ltd. 26 p. NWT Assessment Report #085305

Jones, T.J. and Russell, J.K., 2018. Attrition in the kimberlite system; Mineralogy and Petrology, v. 112 no. 2, supp., p. S491-S501. doi: 10.1007/s00710-018-0580-0

Jutard, G. and Plauchut, B.P., 1973. Cretaceous and Tertiary stratigraphy northern Banks Island; *in* Geological Association of Canada and Canadian Society of Petroleum Geology, (eds.) J.D. Aitken and D.J. Glass, Proceedings, Symposium, Geology of the Canadian Arctic, Saskatoon, Canada, p. 203-219.

Kienlen, B. and Vanderspiegel, R., 2006. Banks Island Project, Airborne Geophysical Survey and Stream Sediment Sampling; Diamonds North Resources Ltd. 135 p. NWT Assessment Report #085105

Kuc, M., 1974. The interglacial flora of Worth Point, western Banks Island; *in* Report of Activities, part B, November 1973 to March 1974; (ed.) R.G. Blackadar; Geological Survey of Canada, Paper 74-1B, p. 227-231. doi:10.4095/104761

Lajeunesse, P. and Hanson, M., 2008. Field observation of recent transgression on northern and eastern Melville Island, western Canadian High Arctic; Geomorphology, v. 101, p. 618-630. doi:10.1016/j.geomorph.2008.03.002

Lakeman, T.R., 2012. Quaternary glaciation of central Banks Island, NT, Canada. unpublished Ph.D. thesis, Department of Earth and Atmospheric Sciences, University of Alberta, 218 p.

Lakeman, T.R. and England, J.H., 2012. Paleoglaciological insights from the age and morphology of the Jesse moraine belt, western Canadian Arctic; Quaternary Science Reviews, v. 47, p. 82-100. doi:10.1016/j.quascirev.2012.04.018

Lakeman, T.R. and England, J.H., 2013. Late Wisconsinan glaciation and postglacial relative sea-level change on western Banks Island, Canadian Arctic Archipelago; Quaternary Research, v. 80, p. 99-112. doi:10.1016/j.yqres.2013.02.001

Lakeman, T.R. and England, J.H., 2014. Facies and stratigraphical analyses of glacial and interglacial sediments at Morgan Bluffs, Banks Island, Canadian Arctic Archipelago; Boreas, v. 43, p. 895-913. doi:10.1111/bor.12069.

Lee, D.C., Maddren, J., and Griffin, B.J., 2003. The importance of chromite morphology in diamond exploration. *in* 8$^{th}$ International Kimberlite Conference, Extended Abstract, p. 1-5.

Lewkowicz, A.G. and Way, R.G., 2019. Extremes of summer climate trigger thousands of thermokarst landslides in a High Arctic environment; Nature Communications, v. 10, no.1329, 11 p. doi:10.1038/s41467-019-09314-7

Liu, J., Brin, L.E., Pearson, D.G., Bretschneider, L., Luguet, A., van Acken, D., Kjarsgaard, B., Riches, A., and Mišković, A., 2018. Diamondiferous Paleoproterozoic mantle roots beneath Arctic Canada: a study of mantle xenoliths from Parry Peninsula and central Victoria Island; Geochimica et Cosmochimica Acta, v. 239, p. 284-311. doi:10.1016/j.gca.2018.08.010

MacPhee, R.D.E., 2007. Mammoths in the insular Nearctic? Some constraints on the existence of a Pleistocene megafaunal refugium in the Canadian Arctic Archipelago; Quaternary International, v. 169-170, p. 29-38. doi:10.1016/j.quaint.2006.08.007

Mah, D., 2005. Banks Island Project, Prospecting and Stream Sediment Sampling. Diamonds North Resources Ltd. 35 p. NWT Assessment Report #084932

Manning, T.H., 1956. Narrative of a second Defence Research Board expedition to Banks Island, with notes on the country and its history; Arctic, v. 9, p. 2-77.

Maxwell, J.B., 1981. Climatic regions of the Canadian Arctic Islands; Arctic, v. 34, p. 224-240.

McCandless, T.E., 1990. Kimberlite xenocryst wear in high-energy fluvial systems: experimental studies; Journal of Geochemical Exploration, v. 37, p. 323-331. doi:10.1016/0375-6742(90)90106-K

McCandless, T.E. and Gurney, J.J., 1989. Sodium in garnet and potassium in clinopyroxene: criteria for classifying mantle eclogites; *in* Kimberlites and Related Rocks, (ed.) J. Ross; Geological Society of Australia, Special Publication 14, p. 827-832.

McClenaghan, MB., 2005. Indicator mineral methods in mineral exploration; Geochemistry: Exploration, Environment, Analysis, v. 5, p. 233-245. doi:10.1144/1467-7873/03-066

McClenaghan, M.B., 2011. Overview of common processing methods for recovery of indicator minerals from sediment and bedrock in mineral exploration; Geochemistry: Exploration, Environment, Analysis, v. 11, p. 265-278.

McClenaghan, M.B. and Kjarsgaard, B.A., 2001. Indicator mineral and geochemical methods for diamond exploration in glaciated terrain in Canada. In Drift Exploration in Glaciated Terrain, (eds.) M.B. McClenaghan, P.T. Bobrowsky, G.E.M. Hall, and S.J. Cook; Geological Society, Special Publication 185, p. 83-123. doi:10.1144/1467-7873/10-IM-025

McClenaghan, M.B. and Kjarsgaard, B.A., 2007. Indicator mineral and surficial geochemical exploration methods for kimberlite in glaciated terrain: examples from Canada; *in* Mineral Deposits of Canada: A Synthesis of Major Deposit Types, District Metallogeny, the Evolution of Geological Provinces, and Exploration Methods, (ed.) W.D. Goodfellow, Geological Association of Canada, Mineral Deposits Division, Special Publication no. 5, p. 983-1006.

McClenaghan, M.B., Ward, B.C., Kjarsgaard, I.M., Kjarsgaard, B.A., Kerr, D.E., and Dredge, L.A., 2002. Indicator mineral and till geochemical dispersal patterns associated with the Ranch Lake kimberlite, Lac de Gras region, NWT, Canada; Geochemistry: Exploration, Environment, Analysis. v. 2, p. 299-319. doi:10.1144/1467-787302-004

McClenaghan, M.B., Kjarsgaard, I.M., Kjarsgaard, B.A., and Russell, H.A.J., 2006. Indicator mineralogy of kimberlite boulders and sand samples from the Lac Baby and Sharp Lake eskers, Lake Timiskaming kimberlite field, western Quebec and northeastern Ontario; Geological Survey of Canada, Open File 5050, 24 p. doi:10.4095/221567

McClenaghan, M.B., Kjarsgaard, I.M., Paulen, R.C., and Eccles, D.R., 2008. Indicator mineralogy of a kimberlite cobble from an esker on the southeastern flank of the Buffalo Head Hills, northern Alberta. Geological Survey of Canada, Open File 5646, 33 p. doi:10.4095/224970

McClenaghan, M.B., Plouffe, A., McMartin, I., Campbell, J.E., Spirito, W.A., Paulen, R.C., Garrett, R.G., and Hall, G.E.M., 2013. Till sampling and geochemical analytical protocols used by the Geological Survey of Canada; Geochemistry: Exploration, Environment, Analysis, v. 13, p. 285-301. doi:10.1144/geochem2011-083

McClenaghan, M.B., Spirito, W.A., Plouffe, A., McMartin, I. Campbell, J.E., Paulen, R.C., Garrett, R.G., Hall, G.E.M., Pelchat, P., and Gauthier, M.S., (in press). Geological Survey of Canada till sampling and analytical protocols: from field to archive: 2020 update. Geological Survey of Canada, Open File 8591, 60 p.

McCurdy, M.W., McNeil, R.J., and Day, S.J.A., 2012. Geochemical, mineralogical and kimberlite indicator mineral data for silts, heavy mineral concentrates and waters from two geochemical surveys (2010 and 2011) on Victoria Island, Northwest Territories (NTS 87G, 87H, 88A and 88B); Geological Survey of Canada, Open File 7198, 16 p. doi:10.4095/291509

McCurdy, M., Rainbird, R., and McNeil, R., 2017. Exploring for lead and zinc using indicator minerals with stream silt and water geochemistry in the Canadian Arctic Islands: an example from Victoria Island, Northwest Territories; *in* New Frontiers for Exploration in Glaciated Terrain, (eds.) R.C. Paulen and M.B. McClenaghan; Geological Survey of Canada, Open File 7374 (revised), p. 65-70. Doi:10.4095/300293

McDonough, W.F. and Sun, S.-S., 1995. The composition of the Earth; Chemical Geology, v. 120, p. 223-253. doi:10.1016/0009-2541(94)00140-4

Miall, A.D., 1975. Geology, Banks Island, District of Franklin; Geological Survey of Canada, Map 1454A, scale 1:1 000 000. doi:10.4095/109177

Miall, A.D., 1976. Proterozoic and Paleozoic geology of Banks Island, Arctic Canada; Geological Survey of Canada, Bulletin 258, 77 p. doi:10.4095/103964

Miall, A.D., 1979. Mesozoic and Tertiary geology of Banks Island, Arctic Canada: the history of an unstable craton margin; Geological Survey of Canada, Memoir 387, 235 p. doi:10.4095/105620

Morris, T.F., Sage, R.P., Ayer, J.A., and Crabtree, D.C., 2002. A study in clinopyroxene composition: implications for kimberlite exploration; Geochemistry: Exploration, Analysis, Environment, v. 2, p. 321-331. doi:10.1144/1467-787302-037

Morris, W.A. and Vincent, J-S., 1979. Magnetostratigraphy of Pleistocene sediments of Banks Island, Northwest Territories: a feasibility study; *in* Current Research, Part B; Geological Survey of Canada, Paper 79-1B, p. 301-306. doi:10.4095/119849

Mosig, R.W., 1980. Morphology of indicator minerals as a guide to proximity of source; in Kimberlites and Diamonds, (eds.) J.E. Glover and D.I. Groves; p. 81-88.

Nabelek, P.I., Bédard, J.H., Hryciuk, M., and Hayes, B., 2013. Short-duration contact metamorphism of calcareous sedimentary rocks by Neoproterozoic Franklin gabbro sills and dykes on Victoria Island, Canada; Journal of Metamorphic Geology, v. 31, p. 205-220. doi:10.1111/jmg.12015

Nimis, P., 1998. Evaluation of diamond potential from composition of peridotitic chromian diopside; European Journal of Mineralogy, v. 10, p. 505-519.

Nixon, F.C., 2012. The glacial, sea level, and sea ice histories of Melville and Eglinton islands, western Canadian High Arctic: Last Glacial Maximum to present; Unpublished Ph.D. thesis, University of Alberta, Edmonton, Canada, 219 p.

Nixon, F.C. and England, J., 2014. Expanded Late Wisconsinan ice cap and ice sheet margins in the western Queen Elizabeth Islands, Arctic Canada; Quaternary Science Reviews, v. 91, p. 146-164. doi:10.1016/j.quascirev.2013.10.036

Nowell, G.M., Pearson, D.G., Ottley, C.J., Schweiters, J., and Dowall, D., 2003. Long-term performance characteristics of a plasma ionisation multi-collector mass spectrometer (PIMMS): the ThermoFinnigan Neptune; *in* Plasma Source Mass Spectrometry: applications and emerging technologies; (eds.) G. Holland and S.D. Tanner. Cambridge, Royal Society Ch., p. 307-320.

Paton, C., Hellstrom, J., Paul, B., Woodhead, J., and Hergt, J., 2011. Iolite: Freeware for the visualisation and processing of mass spectrometric data; Journal of Analytical Atomic Spectrometry, v. 26, p. 2508-2518. doi:10.1039/C1JA10172B

Piepjohn, K., Atkinson, E., Dewing, K., Smith, R., and Galloway, J., 2018. Cenozoic structural evolution on northern Banks Island, N.W.T. Canada; Arktos, v. 4, no. 15, 19 p. doi:10.1007/s41063-018-0046-y

Plauchut, B.P., 1971. Geology of the Sverdrup Basin; Bulletin Canadian Petroleum Geology, 19: 659-679. doi:10.35767/gscpgbull.19.3.659

Plouffe, A., McClenaghan, M.B., Paulen, R.C., McMartin, I., Campbell, J.E., and Spirito, W.A., 2013. Processing of glacial sediments for the recovery of indicator minerals: protocols used at the Geological Survey of Canada; Geochemistry: Exploration, Environment, Analysis, v. 13, p. 303-316. doi:10.1144/geochem2011-109

Poitras, S.P., Pearson, D.G., Hardman, M.F., Stachel, T., Nowell, G.M., and Cairns, S., 2018. Evidence for a 200 km thick diamond-bearing root beneath the Central Mackenzie Valley, Northwest Territories, Canada? Diamond indicator mineral geochemistry from the Horn Plateau and Trout Lake regions; Mineralogy and Petrology, 112, no. 2, sup., p. S719-S736. doi:10.1007/s00710-018-0641-4

Pokhilenko, N.P., Afanas'ev, V.P., and Vavilov, M.A., 2010. Behavior of kimberlite indicator minerals during the formation of mechanical dispersion halos in glacial settings; Lithology and Mineral Resources, v. 45, n. 4, p. 324-329.

Porsild, A.E., 1950. A biological exploration of Banks and Victoria islands; Arctic, v. 3, p. 45-54.

Prest, V.K., Grant, D.R., and Rampton, V.N., 1968. Glacial Map of Canada; Geological Survey of Canada, Map 1257A, scale 1:5 000 000. doi:10.4095/108979

Prior, G.J., McCurdy, M.W., and Friske, P.W.B., 2009. Stream sediment sampling for kimberlite indicator minerals in the Western Canada Sedimentary Basin: the Buffalo Head Hills survey, north-central Alberta; *in* Application of Till and Stream Sediment Heavy Mineral and Geochemical Methods to Mineral Exploration in Western and Northern Canada; (eds.) R.C. Paulen and I. McMartin; GAC Short Course Notes, v. 18, p. 111-124.

Roeder, P.L. and Schulze, D.J., 2008. Crystallization of groundmass spinel in kimberlite; Journal of Petrology, v. 49, p. 1473-1495. doi:10.1093/petrology/egn034

Rudy, A.C.A., Lamoureux, S.F., Kokelj, S.V., Smith, I.R., and England, J.H., 2017. Accelerating thermokarst transforms ice-cored terrain triggering a downstream cascade to the ocean; Geophysical Research Letters, v. 44, p. 11,080-11,087. doi:10.1002/2017GL074912

Ryan, C.G., Griffin, W.L., and Pearson, N.J., 1996. Garnet geotherms: pressure-temperature data from Cr-pyrope garnet xenocrysts in volcanic rocks; Journal of Geophysical Research, v. 101, no. B3, p. 5611-5625. doi:10.1029/95JB03207

Schulze, D.J., 1987. Megacrysts from alkali volcanic rocks; *in* Mantle Xenoliths, (ed.) P.H. Nixon, John Wiley & Sons Ltd., New York, USA, p. 433-452.

Schulze, D.J., 2003. A classification scheme for mantle-derived garnets in kimberlite: a tool for investigating the mantle and exploring for diamonds; Lithos, v. 71, p. 195-213. doi:10.1016/S0024-4937(03)00113-0

Smit, K.V., Pearson, D.G., Stachel, T., and Seller, M. 2014. Peridotites from Attawapiskat, Canada: Mesoproterozoic reworking of Palaeoarchaean lithospheric mantle beneath the northern Superior superterrane. Journal of Petrology, v. 55, p. 1829-1863. doi:10.1093/petrology/egu043

Smith, I.R., 2015. Report of activities for the assessment of kimberlite indicator mineral sources on northeastern Banks Island, Northwest Territories: GEM 2 Western Arctic Margins project; Geological Survey of Canada, Open File 7972, 19 p. doi:10.4095/297304

Smith, I.R. and Paulen, R.C., 2016. Testing the efficacy of a field-portable spiral helix sediment concentrator for capturing kimberlite indicator minerals and gold grains from unconsolidated glacial and non-glacial sediments; Geological Survey of Canada, Open File 8092, 21 p. doi:10.4095/29912

Smith, I.R. and Farineau, A., 2015. A reconstruction of drift lithostratigraphy on Banks Island, Northwest Territories, based on seismic shothole drillers' logs and lithogeochemical sample records; Geological Survey of Canada, Open File 7322. doi:10.4095/296215

Smith, I.R., Dewing, K., Galloway, J., and Piepjohn, K., 2016. Report of Activities for the GEM-2 Western Arctic Margins project, Banks Island, Northwest Territories; Geological Survey of Canada, Open File 8150, p. 18. doi:10.4095/299294

Sobolev, N.V., 1971. On mineralogical criteria of a diamond potential of kimberlites; Geologika et Geophysika, v. 12, p. 70-78 (in Russian).

Sobolev, N.V., 1977. Deep-seated inclusions in kimberlites and the problem of the Upper Mantle composition; Translated from Russian by D.A. Brown and F.R. Boyd (ed.), American Geophysical Union, Washington, 259 p.

Spirito, W.A., McClenaghan, M.B., Plouffe, A., McMartin, I., Campbell, J.E., Paulen, R.C., Garrett, R.G., and Hall, G.E.M., 2011. Till sampling and analytical protocols for GEM projects: from field to archive; Geological Survey of Canada, Open File 6850, 85 p. doi:10.4095/300281

Stachel, T. and Harris, J.W., 2008. The origin of cratonic diamonds – constraints from mineral inclusions; Ore Geology Reviews, v. 34, no. 1-2, p. 5-32. doi:10.1016/j.oregeorev.2007.05.002

Stachel, T., Viljoen, K.S., Brey, G., and Harris, J.W., 1998. Metasomatic processes in lherzolitic and harzburgitic domains of diamondiferous lithospheric mantle: REE in garnets from xenoliths and inclusions in diamonds; Earth and Planetary Science Letters, v. 159, p. 1-12. doi: 10.1016/S0012-821X(98)00064-8

Stachel, T., Aulbach, S., Brey, G.P., Harris, J.W., Leost, I., Tappert, R., and Viljoen, K.S., 2004. The trace element composition of silicate inclusions in diamonds: a review. Lithos, v. 77, p. 1-19. doi:10.1016/j.lithos.2004.03.027

Stachel, T., Banas, A., Aulbach, S., Smit, K.V., Wescott, P., Chinn, I.L., and Kong, J., 2018. The Victor Mine (Superior Craton, Canada): Neoproterozoic lherzolitic diamonds from a thermally-modified cratonic root. Mineralogy and Petrology, v. 112, supplement 1, p. S325-S336. doi:10.1007/s00710-018-0574-y

Stokes, C.R., Clark, C.D., and Storrar, R., 2009. Major changes in ice stream dynamics during deglaciation of the north-western margin of the Laurentide Ice Sheet; Quaternary Science Reviews, v. 28, p. 721-738. doi:10.1016/j.quascirev.2008.07.019

Stubley, M.P. and Irwin, D., 2019. Bedrock Geology of the Slave Craton, Northwest Territories and Nunavut. NWT Open File 2019-01, scale 1:750 000

Thorsteinsson, R. and Tozer, E.T., 1959. Western Queen Elizabeth Islands, District of Franklin, Northwest Territories; Geological Survey of Canada, Paper 59-1, 11 p. doi:10.4095/101235

Thorsteinsson, R. and Tozer, E.T., 1962. Banks, Victoria, and Stefansson Islands, Arctic Archipelago. Geological Survey of Canada, Memoir 330, 83 p. doi:10.4095/100554

Tozer, E.T., 1956. Geological reconnaissance, Prince Patrick, Eglinton and western Melville Islands, Arctic Archipelago, Northwest Territories; Geological Survey of Canada, Paper 55-5, 32 p. doi:10.4095/101290

Vaughan, J.M., 2014. The glacial and relative sea level history of southern Banks Island, NT, Canada. Unpublished Ph.D. thesis, University of Alberta, 295 p.

Vaughan, J.M., England, J.H., and Evans, D.J.A., 2014. Glaciotectonic deformation and reinterpretation of the Worth Point stratigraphic sequence: Banks Island, NT, Canada; Quaternary Science Reviews, v. 91, p. 124-145. doi:10.1016/j.quascirev.2013.11.005

Viljoen, K.S., Harris, J.W., Ivanic, T., Richardson, S.H., and Gray, K., 2014. Trace element chemistry of peridotitic garnets in diamonds from the Premier (Cullinan) and Finsch kimberlites, South Africa: Contrasting styles of mantle metasomatism; Lithos, v. 208-209, p. 1-15. doi:10.1016/j.lithos.2014.08.010

Vincent, J-S., 1978a. Lithostratigraphy of the Quaternary sediments east of Jesse Bay, Banks Island, District of Franklin, N.W.T.; Current Research, Part A, (eds.) R.G. Blackadar, P.J., Griffin, H. Dumych, and E.J.W. Irish; Geological Survey of Canada, Paper 78-1A, p. 189-193. doi:10.4095/103887

Vincent, J-S., 1978b. Surficial Geology of Banks Island, District of Franklin, N.W.T.; Geological Survey of Canada, Open File 577 (NTS map areas 97G, H, 98A-F, parts of 88B-D). doi:10.4095/129490

Vincent, J-S., 1978c. Limits of ice advance, glacial lakes, and marine transgressions on Banks Island, District of Franklin: a preliminary interpretation; Current Research, Part C, (eds.) R.G. Blackadar, P.J. Griffin, H. Dumych and E.R.W. Neale; Geological Survey of Canada, Paper 78-1C, p. 53-62. doi:10.4095/104591

Vincent, J-S., 1978d. Glacial limits on Banks Island with special emphasis on the Wisconsinan limit and events. Geological Society of America, Joint Annual Meeting, Toronto, ON, Abstracts with Program, v.10, no.7, p. 509-510.

Vincent, J-S., 1980a. Surficial Geology – Banks Island, Northwest Territories; Geological Survey of Canada, Map 16-1979 (North half), scale 1:250 000. doi:10.4095/109647

Vincent, J-S., 1980b. Surficial Geology – Banks Island, Northwest Territories; Geological Survey of Canada, Map 17-1979 (South half), scale 1:250 000. doi:10.4095/109698

Vincent, J-S., 1982. The Quaternary history of Banks Island, N.W.T., Canada; Géographie physique et Quaternaire, v. 36, p. 209-232. doi:10.7202/032478ar

Vincent, J.-S., 1983. La géologie du Quaternaire et la géomorphologie de l'Île Banks, Arctique Canadien. Geological Survey of Canada, Memoir 405. 118 p. doi.10.4095/119517

Vincent, J-S., 1984. Quaternary stratigraphy of the western Canadian Arctic Archipelago; *in* Quaternary Stratigraphy of Canada – A Canadian Contribution to IGCP Project 24, (ed.) R.J. Fulton; Geological Survey of Canada, Paper 84-10, p. 87-100. doi:10.4095/119761

Vincent, J-S., 1989. Quaternary geology of the northern Canadian Interior Plains; *in* Quaternary Geology of Canada and Greenland, (ed.) R.J Fulton; Geological Survey of Canada, Geology of Canada Series no. 1,  p. 100-137. doi:10.4095/127955

Vincent, J.-S., 1990. Late Tertiary and Early Pleistocene deposits and history of Banks Island, southwestern Canadian Arctic Archipelago; Arctic, v. 43, p. 339-363. doi:10.14430/arctic1630

Vincent, J-S., 1992. The Sangamonian and early Wisconsinan glacial record in the western Canadian Arctic; *in* The Last Glacial-Interglacial Transition in North America, (eds.) P.U. Clark and P.D Lea; Geological Society of America, Special Paper, v. 270, p. 233-252. doi:10.1130/spe270-p233

Vincent, J-S. and Edlund, S.A., 1978. Extended legend to accompany preliminary surficial geology maps of Banks Island. Geological Survey of Canada, Open File 577. doi:10.4095/129490

Vincent, J-S., Occhietti, S., Rutter, N., Lortie, G., Guilbault, J-P., and De Bourtray, B., 1983. The Late Tertiary-Quaternary record of the Duck Hawk Bluffs, Banks Island, Canadian Arctic Archipelago; Canadian Journal of Earth Sciences, v. 20, p. 1694-1712. doi:10.1139/e83-160

Vincent, J-S., Morris, W.A., and Occhietti, S., 1984. Glacial and non-glacial sediments of Matuyama paleomagnetic age on Banks Island, Canadian Arctic Archipelago; Geology, v. 12, p. 139-142. doi: 10.1130/0091-7613(1984)12<139:GANSOM>2.0.CO;2

Wallace, C. and Wood, B.D., 1997. Monopros Limited, Banks Island Property Geophysics. Work Completed from January, 1996 until December 1997. NTS map sheets: 98D/06, 07, 10, 11, 12, 13, 14, 15; 98E/02, 03, 04, 05, 06, 07, 10. 45 pages + figures. NWT Assessment Report #083984

Washburn, A.L. 1947. Reconnaissance geology of portions of Victoria Island and adjacent regions, Arctic Canada; Geological Society of America, Memoir 22, 142 p.

Wilson, H., Anonby, L., and Sacco, D., 2012a. Assessment Report, Rio Tinto Exploration Canada Inc., Banks Island Property, Banks Island, NT, Stream Sediment Sampling. 35 p. NWT Assessment Report #085765.

Wilson, H., Lyon, E., and Anonby, L., 2012b. Assessment Report, Rio Tinto Exploration Canada Inc., Banks Island Property, Banks Island, NT, Airborne Geophysical Report of Activities. 50 p. NWT Assessment Report #085747.

Wilson, J.T., Falconer, G., Mathews, W.H., and Prest, V.K., 1958. Glacial Map of Canada; Geological Association of Canada, scale 1:5 000 000.

Woodhead, J., Hellstrom, J., Hergt, J., Greig, A., and Maas, R., 2007. Isotopic and elemental imaging of geological materials by laser ablation inductively coupled plasma-mass spectrometry; Journal of Geostandards and Geoanalytical Research, v. 31, p. 331-343. doi:10.1111/j.1751-908X.2007.00104.x

Wyatt, B.A., Baumgartner, M., Anckar, E., and Grütter, H., 2004. Compositional classification of "kimberlitic" and "non-kimberlitic" ilmenite. Lithos, v. 77, p. 819-840. doi:10.1016/j.lithos.2004.04.025

Yannan, B. and Matsyuk, S.S., 1991. Compositions of kimberlite spinels from Shangtung and Liaoning provinces in China; Geochemistry International, v. 28, p. 89-96.

# 8. List of Figures

Formation sample 16SUV023 indicated. Pink shading on Canada index map identifies extents of the exposed Canadian Shield (granite source).

**Figure 6.** Flow chart illustrating the processing steps of glacial sediment samples for indicator minerals as per GSC protocols (adapted from Plouffe et al., 2013; McClenaghan et al. 2013, in press). Samples targeted for KIM recovery are passed twice across the shaking table. While the <0.25 mm sample is generally not picked for KIMs, several samples from Beaufort Formation deposits had this step added to capture more of the total KIM population.

**Figure 7.** $Cr_2O_3$ vs CaO (wt%) composition for all GSC Banks Island garnets (Appendix 6B; minus analytical repeats and the Ca-member grossular and andradite grains). Garnet classification according to Grütter et al. (2004); dashed red line is the graphite-diamond constraint line.

**Figure 8.** $TiO_2$ vs $Na_2O$ (wt%) composition for GSC Banks Island garnets (Appendix 6B; only those grains with $TiO_2$<1.2 wt% are shown).

**Figure 9.** Classification schema of Hardman et al. (2018a) used to discriminate the GSC Banks Island low chrome garnets ($Cr_2O_3$ wt%<1.0) into possible mantle or crust origins.

**Figure 10.** Plots of GSC Banks Island garnet, C1 chondrite-normalized (McDonough and Sun (1995)), rare earth element (REE) concentrations for those classified by Grütter et al. (2004) as G9, G10, G11 or G12 (Appendix 7A). **A)** Cr-pyrope garnets with C1 chondrite-normalized heavy-REE enriched profiles. Heavy red and black lines illustrate the average concentrations for the G9 and G11 garnets, respectively; **B)** Cr-pyrope garnets with C1 chondrite-normalized sinusoidal REE profiles. Heavy red, blue, and purple lines illustrate the average concentrations for the G9, G10 and G12 garnets, respectively; **C)** Cr-pyrope garnets with C1 chondrite-normalized middle-REE depleted profiles (n=4), along with other anomalous profiles of normal, slightly-humped, and humped (cf., Banas et al., 2009). Heavy blue line illustrates the average concentrations for the three G10 garnets.

**Figure 11.** Yitrium (Y) versus Zirconium (Zr) (ppm) for Banks Island Cr-pyrope garnets. Points in red are Beaufort Formation samples, blue are stream sediment samples. Fields and metasomatic trends from Griffin et al. (1999). Low-Ti is TiO2≤0.02 wt%; high-Ti is TiO2>0.06 wt%.

**Figure 12.** Ni-in-garnet geothermometry (Ryan et al., 1996) for GSC Banks Island Cr-pyrope garnets (n=61) between 450 and 1800°C. G9 garnets are unlabelled, all others are as indicated. Dark line is the relative probability curve of the distribution. Diamond stability window typically sits between 900 and 1200°C.

**Figure 13.** Plots of GSC Banks Island Cr-pyrope garnet rare earth element (REE) concentrations grouped by Ni-in-garnet geothermometry temperature determinations (calculated according to Ryan et al. (1996)), using only those garnets classified by Grütter et al. (2004) as G9, 10, 11, or 12 (Appendix 6B). Solid black line averages exclude 2 outliers in the 800-950°C group, 1 outlier in the 950-1100°C group, and 3 outliers in the 1100-1250°C group. Data points illustrated by circle symbols are from Beaufort Formation samples; triangle symbols are from stream sediment samples.

**Figure 14.** Plots for Cr-spinel (chromites) from GSC Banks Island samples (n=215). **A)** $Cr_2O_3$ versus MgO bivariate plot (diamond inclusion and intergrowth fields from Fipke et al. (1995)); **B)** $Cr_2O_3$ versus $TiO_2$ plot with the lamproite, diamondiferous kimberlite, non-kimberlite and non-lamproite sources of Fipke et al. (1995); **C)** $Al_2O_3$ versus $Cr_2O_3$ bivariate plot with diamond inclusion field from Sobolev (1977); **D)** Mg# versus Cr# of spinel; fields of global non-cratonic and cratonic peridotites from Liu et al. (2018); **E)** $TiO_2$ versus $Cr_2O_3$ plot after McClenaghan and Kjarsgaard (2007).

**Figure 15**. GSC Banks Island ilmenite grain chemistry plots. **A)** $Cr_2O_3$ versus MgO bivariate plot (Mg-ilmenites >4 wt% MgO); **B)** MgO versus $TiO_2$ (wt%) plot discriminating kimberlitic from non-kimberlitic ilmenite grains; dashed lines are percentage $Fe_2O_3$ contours (after Wyatt et al., 2004). Circle symbols are those grains identified by ODM as Mg-ilmenite; triangle symbols are those grains identified by ODM as crustal ilmenite.

**Figure 16.** Bivariate plot of MAGNUM (Mg#; Mg/(Mg+Fe)) versus $Cr_2O_3$ wt% for GSC Banks Island forsterite grains. Red line outlines the diamond inclusion field compositions from Fipke et al. (1995) – greatly distorted to accommodate low Mg# of these Banks Island samples.

**Figure 17.** $Cr_2O_3$ versus Mg# (100Mg/(Fe+Mg)) in Cr-diopside, low-Cr diopside and diopside grains from Banks Island, NWT (classification after McClenaghan et al., 2006).

**Figure 18.** Location of sample 16SUV023. **A)** Prominent Beaufort Formation peneplain gravelled surface outlined by red dashed line (0.4 x 2.0 km). View to southwest. Absence of ice wedge polygons points to coarse nature of deposit in comparison to adjacent finer Eureka Sound Formation strata. **B)** Pebble-gravel surface exhibiting orange-oxidized, rounded quartzite, and dominant sub-rounded to sub-angular black and grey chert, with lesser tan, green, and red chert; clast sizes average 1-3 cm diameter. **C)** Looking north along peneplain surface from sample site. Note absence of glacial erratic cobbles or boulders, and conspicuous bright orange-colour.

**Figure 19.** Laurentide Ice Sheet Late Wisconsinan ice flow reconstructions. Red stars represents position of Snow Goose kimberlite cluster on Victoria Island, and the kimberlite cluster on Parry Peninsula; green square is Beaufort Formation sample 16-SUV-023; distance to Snow Goose cluster is ~550 km. Basemap image from IBCAO topographic and bathymetric model (Jakobsson et al. 2012). **A)** Maximum glacial extents (18 ka ($^{14}C$ yr BP)) of Dyke and Prest (1987) illustrating position of M'Clintock Ice Divide on eastern Victoria Island, restricted ice margins on eastern Banks Island (accordant with the Jesse Moraine of Fyles (1962) and Vincent (1982, 1983)), and grounded and floating ice shelves in M'Clure Strait and Amundsen Gulf. **B)** Revised last glacial maximum extents proposed by England et al. (2009) and Lakeman and England (2012) illustrating complete inundation of Banks Island by Laurentide Ice Sheet and prominent ice streams in M'Clure Strait and Amundsen Gulf, terminating on the continental shelf. **C)** Lakeman and England's (2012) 14 cal ka BP (~12.6 $^{14}C$ yr BP) Thomsen Phase cold-base ice margins, and the emergence of a local ice divide on the Shaler Mountains. **D)** Lakeman and England's (2012) 13.25 cal ka BP (~12.2 $^{14}C$ yr BP) ice margin coincident with the prominent outer margin of the Jesse moraine belt (Prince of Wales Phase), and the emergence of increasingly topographically-defined ice divides and ice flow paths on Victoria Island.

**Figure 20.** Examples of kame deposits associated with aeromagnetic anomalies identified by Diamonds North (Jober and Vanderspiegel, 2007; their Appendix 4). **A)** Prominent 8 m high (60 m across) pyramidal kame pile formed at the margin of a retreating polythermal glacier. This was classified as a moderate anomaly (<3 nT). **B)** Example of the imbricated beds of sorted sediments within another kame, here exposed by a slump. Exposed section is the upper 2.5 m of a 5 m high kame); glacier margin would have been towards the viewer, while supraglacial streamflow draining off the steep-fronted glacier margin extended away, towards the figure in the photograph.

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