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Ice flow studies in Boothia Mainland (NTS 57A and 57B), Kitikmeot region, Nunavut

T. Tremblay, J.J. Ryan , D.T. James

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Supplementary Geophysical data relating to the Boothia Penninsula Integrated Geoscience Project can be downloaded from the Geoscience Data Repository website:
http://GDR.NRCan.gc.ca

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Abstract

Mapping surficial macroforms and glacial striations in the Boothia Mainland area of central Nunavut has unveiled a complex, four phase ice flow history. From oldest to youngest, the ice flow history consists of: Phase 1 to the north, Phase 2 to the northeast, Phase 3 to the east, and Phase 4 to the north. Glacial transport studies established by petrographic counts of erratics indicate that all four phases had an influence on dispersal trails from up-ice bedrock sources. Erratics of Paleoproterozoic orthoquartzite were particularly useful to illustrate a palimpsest-shaped dispersion train, wherein orthoquartzite eratics were carried more than 175 km from source. These dispersal train studies are vital for mineral exploration programs utilizing kimberlite indicator minerals, kimberlite debris, or till geochemistry in this region.

INTRODUCTION

In the 2005 field season, surficial geology field work was undertaken in the Boothia Mainland area (NTS 57A and 57B, Figure 1), Kitikmeot region, Nunavut. This work was carried out in collaboration with regional bedrock mapping, collectively forming the field-based components of the Boothia Peninsula Integrated Geoscience Project; a joint effort of the Geological Survey of Canada and the Canada-Nunavut Geoscience Office. The surficial studies focussed on geomorphological mapping, esker and till sampling for kimberlite indicator minerals (KIM), glacial striation measurements, marine shell sampling, and petrographic counts in till. This report presents the ice flow record and interpretation of glacial transport history based on petrographic counts in surface till. Understanding the glacial history is imperative for the mineral exploration industry when using till geochemistry and KIMs in the search for metal and diamond deposits.

The main highlights of the Quaternary geology in this region include the evidence for ice flow towards an ice stream in the Gulf of Boothia, the northeast-oriented transport of Paleoproterozoic metasedimentary rocks and their use as indicators of glacial transport, and the elevation limit of marine invasion, which reached 240m above seal level (a.s.l.) in the Pelly Bay region (Figure 1). Through Wisconsinian glaciation, the ice flows in this region were influenced by the Keewatin Ice Divide, located south of study area, and the McClintock ice divide, located west of study area (Dyke, 1984).

Previous work in the study area includes surficial mapping, glacial dispersal and ice flow studies on the Boothia Peninsula by Dyke (1984), based in part on the work of Boydell et al. (1979). South of the project area, and closer to the inferred position of the Keewatin Ice Divide, works by Little et al. (2001) and McMartin et al. (2003a, 2003b) in Committee Bay area provided a regional ice flow history. Adjacent to the southern limit of study area, an ice flow study by Ozyer and Hicock (2006) in Darby Lake links the Committee Bay project area with the Boothia Mainland study area (Figure 1).

REGIONAL SETTING

Physiography

In this paper, the term "Boothia Mainland" refers to that part of mainland Nunavut south of Ithmus of Boothia, and north of 68 degrees north (Figure 1). The area is characterized by two major lowlands (Figure 2) that coincide with areas underlain by Paleozoic rocks of the Arctic Platform (carbonates and sandstones, Figure 3). The Rasmussen Lowlands, which are generally below 90 m a.s.l., consist of a wide, flat plain on the west side of Boothia Mainland, with a few low-elevation hills of Precambrian rocks. The Quaternary sedimentary cover in the Rasmussen Lowlands mainly comprises marine silts and clays, littoral sands, reworked tills and sparse eskers.. Locally, and particularly in the northwestern part of of the Rasmussen Lowlands, till is covered by a pavement of carbonate blocks that may have been concentrated through the wave-washing of till. Draining into Rasmussen Basin, the Murchison, Inglis, and Castor and Pollux rivers flow across the lowlands, which have abundant thermokarst ponds and ice-wedge polygons. The northern part of the lowlands is characterised by a field of large drumlins (streamlined till forms) that extends northeasterly from the coast to the central uplands. Typically, the drumlins are 5 to 10 km long and 400 to 700 m wide.

The Central Plateau (or Wager plateau) is an inverted hull-shaped highland with elevations ranging from 200 to 350 m in the centre, and 50 to 100 m deep depressions in small, ice-gouged valleys. The area is dissected by large linear valleys (generally under 55 m a.s.l.) which are occupied by major lakes and rivers (e.g., Murchison Lake, Simpson Lake and Murchison River). In the uplands, numerous lakes outline the bedrock linear or curvilinear structural grain and the trend of the streamlined forms. Bedrock exposure is abundant here, though covered by thin, dense lichen. Many drumlins and crag-and-tail features occur in areas, where till cover is present.

Physiographic lowlands also occur in the vicinity of Pelly Bay and mainly consist of areas below 100 m a.s.l..

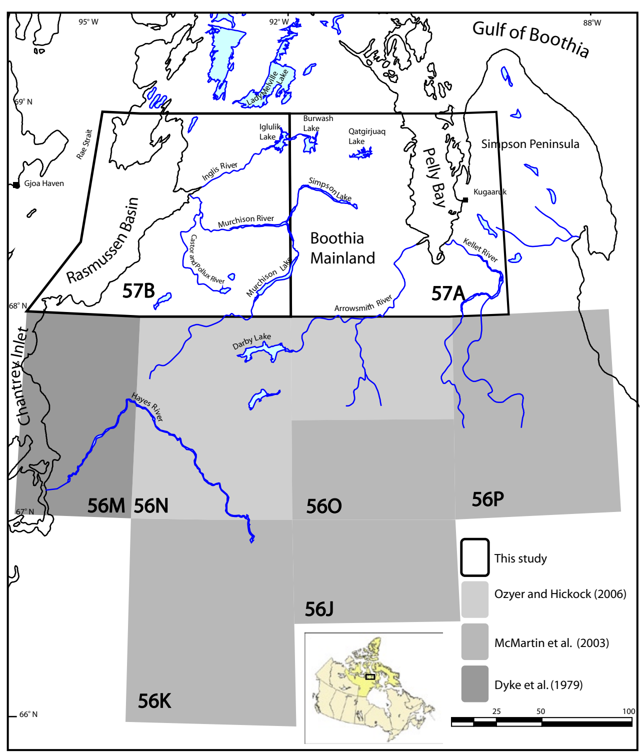


***Figure 1.** Location map of study area.*

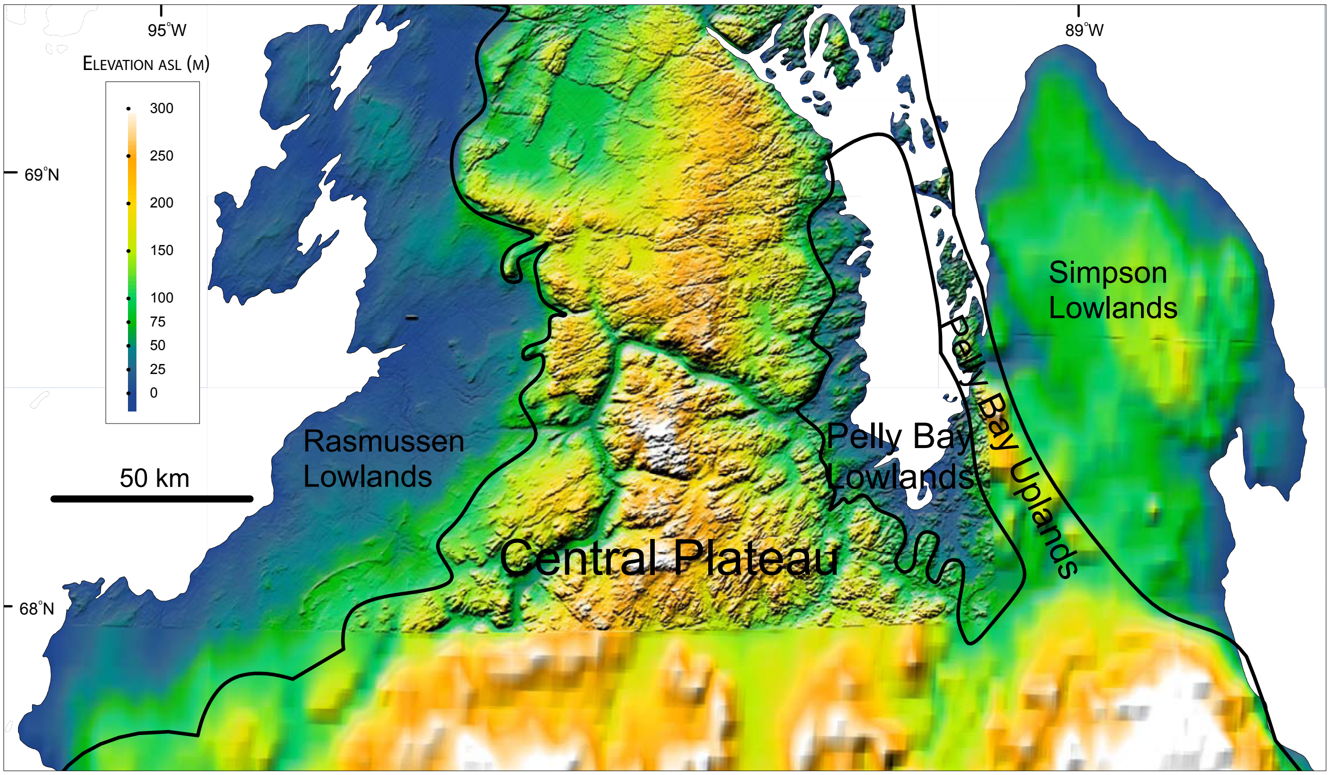


Figure 2. *Digital elevation model and physiographic regions. DEM source : GSC geophysical air survey (90m, unpublished) and GlobeTile (1km).*

No Paleozoic cover sequence is exposed in this area, and the topographic depression is interpreted to be controlled by normal faulting and tilting of unexposed bedrock. This area is characterised by flat-lying marine clays, regressive sands, and interconnected, elongated hills formed by Precambrian bedrock (Figure 4A). In contrast to the Rasmussen Lowlands, where most of the clay zones are flat and covered by grasslands, marine clays in the Pelly Bay Lowlands are commonly unvegetated and gullied and display fluvial erosional patterns forming a spectacular badland landscape (Figure 4B). The east side of Pelly Bay is characterized by a narrow topographic rise, the Pelly Bay Uplands, which commonly reach 250 m a.s.l. and separate the Pelly Bay Lowlands from the Simpson Lowlands. The bedrock ridge closely follows the faulted contact between Precambrian and Paleozoic rocks. The rounded and un-streamlined forms of the Pelly Bay Uplands express a very blocky bedrock topography that is in stark contrast with the streamlined bedrock hills in the Boothia mainland. To the south, this narrow band of hills is connected with the the main plateau, although the plateau is dissected by major valleys occupied by the Arrowsmith and the Kellet Rivers.

The Simpson Lowlands (Figure 2) include an area below 70 m a.s.l.. Till ridges are present in the lowlands, as well as marine beaches and a few, flat-lying areas covered by marine clays. Paleozoic carbonate rock cuestas occur near the faulted contact with the Precambrian rocks.

Throughout the study area, the periglacial features include felsenmeer covering important parts of the bedrocks outcrops, small- to medium-sized polygons and mudboils in till, large ice-wedge polygons in sands and marine clays, pingos and thermokarst in marine clays, and solifluction lobes and stripes on slopes of all sediments.

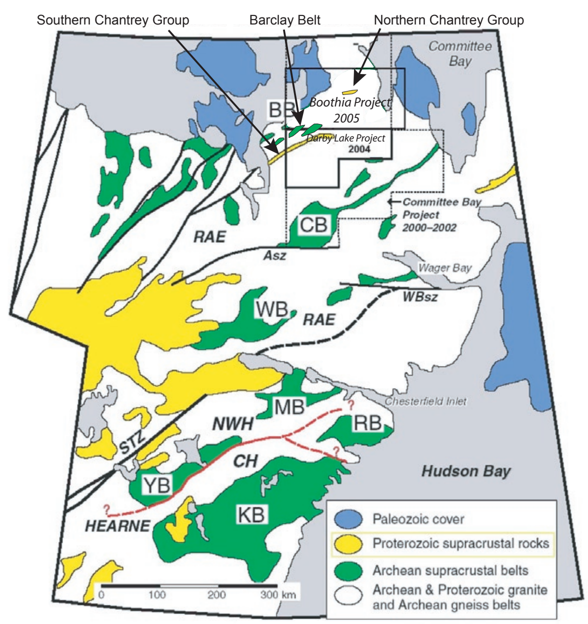
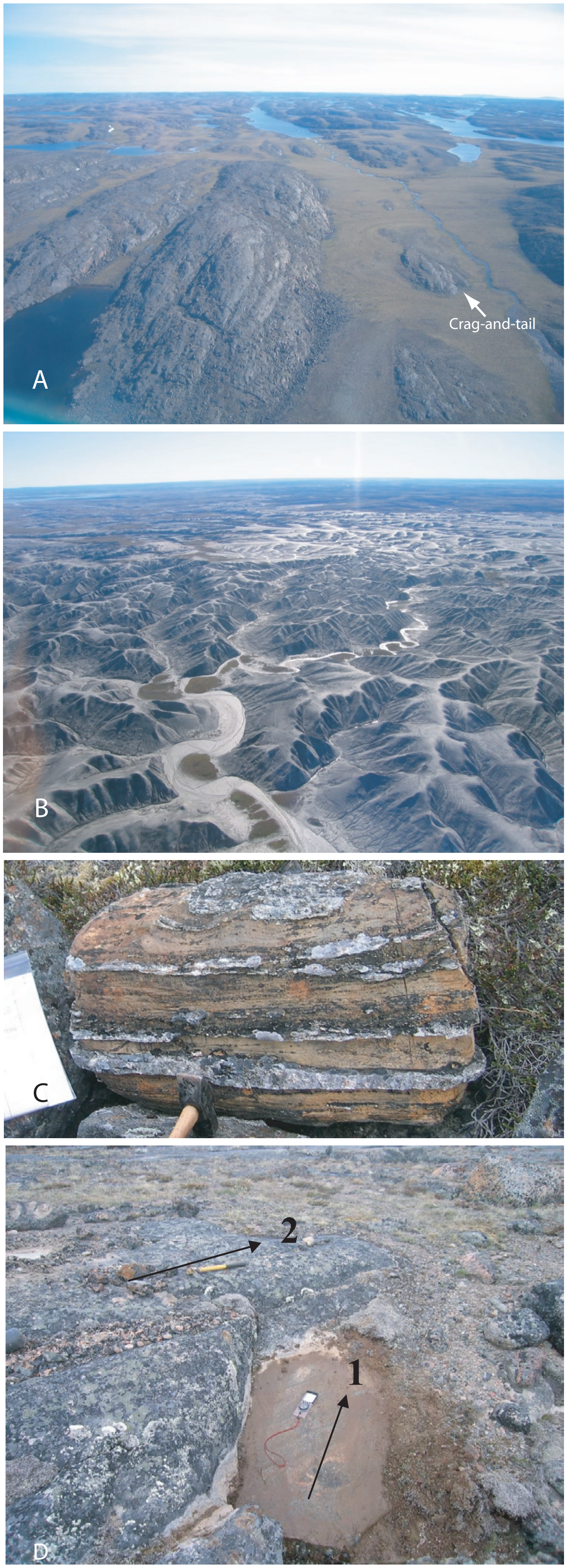


Figure 3. *Regional bedrock map; modified from Sandeman et al. (2005).*



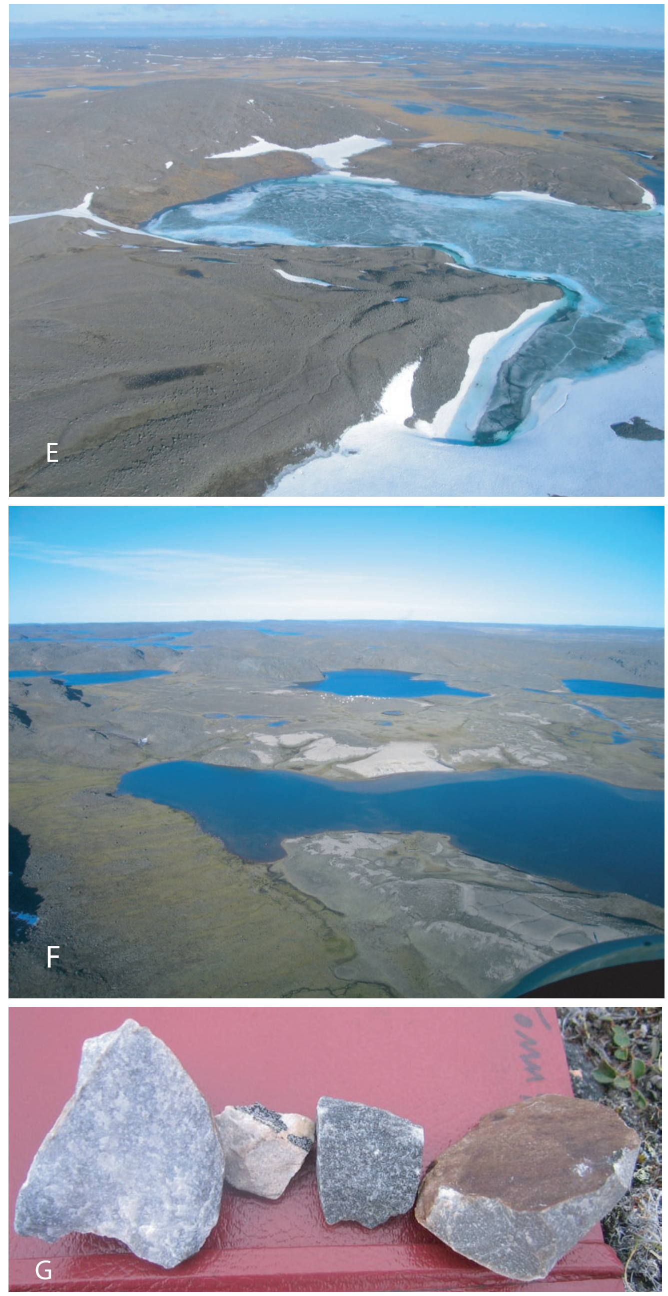


Figure 4. *Photographs.*
A: Ice-moulded bedrock forms and small crag-and-tail; B: Gullied marine clays and sands forming a badland landscape in the Pelly Bay Lowlands; C: Paleozoic carbonate with sandstone layers; D: striations with protected surface chronology indicator; E: Glaciomarine delta (with GSC camp white tents visible in the back); F: Glaciomarine delta. Both photographs E and F display coalescent kettles and ice marginal channels. G: Various colored orthoquartzites.

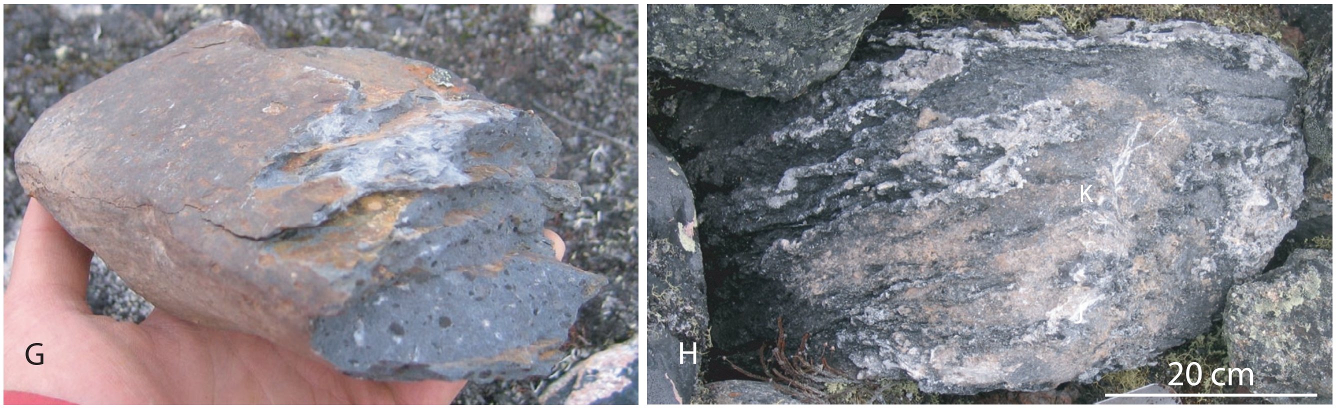


Figure 4 (continued). *Photographs. G: Pelite with biotite\cordierite porphyroblasts, from southern Chantrey Group; H: Proterozoic marble erratic rock, from northern Chantrey Group.*

Bedrock geology

Bedrock geology of the Boothia Mainland area (Ryan et al., in prep., Figure 3) is part the north-central Rae domain of the Churchill province. The area comprises a high-grade gneissic terrain dominated by Neoarchean metaplutonic rocks (Hinchey et al. 2006), lesser Archean and Paleoproterozoic supracrustal sequences, and migmatitic gneiss (Ryan et al., 2006, in prep). The Archean supracrustal rocks (Barclay belt) outcrop as narrow, northeast-striking belts of psammite, semi-pelite, metabasalt, ultramafic schist, and sulphide-bearing (lean) iron formation, metamorphosed at lower to upper amphibolite facies. Based on similarities in rock types, inferred stratigraphy, and the youngest detrital zircon age of 2.76 Ga from a psammitic schist horizon (Hinchey et al. 2007), the Barclay belt is correlated with the pan-Rae Prince Albert Group (Hinchey et al., 2007; Ryan et al., in prep). Granitic rocks, and their gneissic equivalents, are dominated by polyphase, commonly porphyritic, biotite and hornblende-bearing monzogranite to granodiorite. Preliminary zircon ages indicate the intrusive rocks are part of a 2.61-2.59 Ga pan-Rae magmatic event (Hinchey et al., 2006).

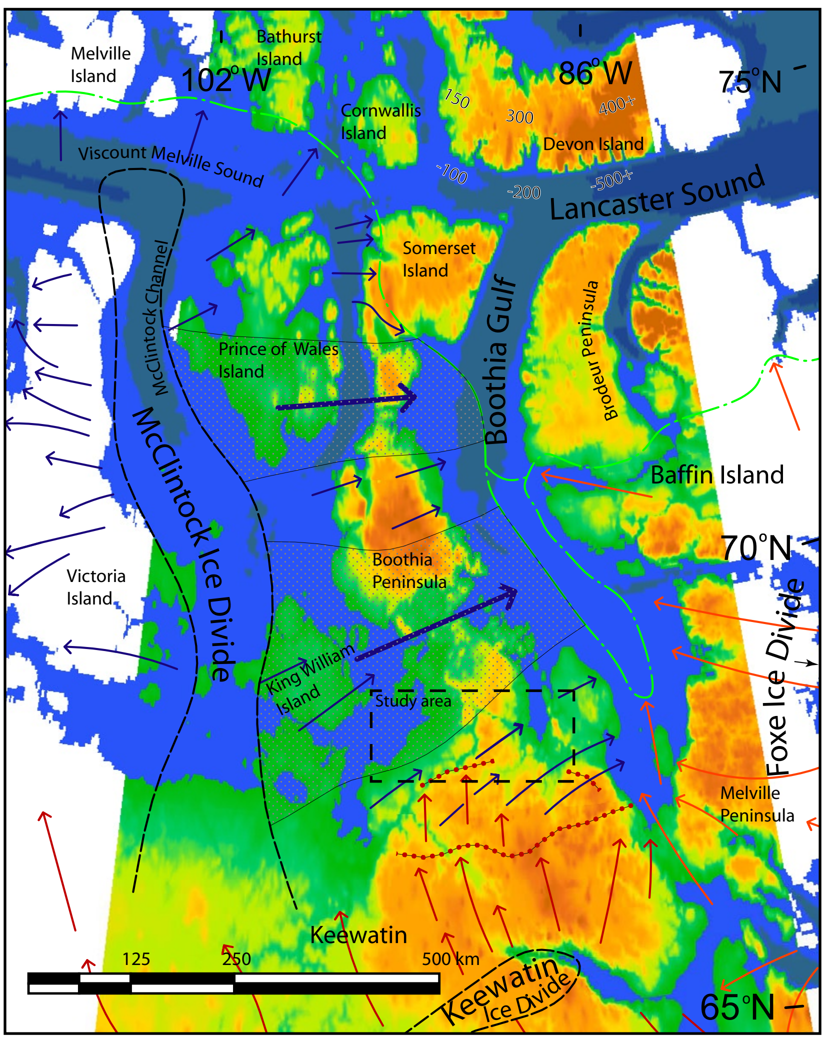


Figure 5. *Regional topographic and bathymetric map (source of DEM : Globetile ; source of bathymetry : Geomatics Canada) and regional ice flow pattern (from Dyke, 1984). Blue lines : ice flows from McClintock Ice Divide. Red lines : ice flows from other ice divides. Thick blue lines : Boothia ice streams.*

In the northern part of the study area, small outliers of a supracrustal sequence dominated by marble, pelitic schist, minor amounts of psammitic schist and quartzite occur as complex infolds in the Archean rocks (Kraft et al., 2005). Detrital zircon studies of the quartzite indicate a Paleoproterozoic age (Hinchey et al., 2007). On the basis of the zircon ages, rock types and stratigraphy, the sequence is correlated with Chantrey Group (Heywood, 1961; Frisch, 2000; Sandeman et al., 2005) which occurs in the Darby Lake approximately 120 km to the south-southwest (Figure 3). For the present purposes, the aforementioned northern and southern occurrences of the Cahntry Group are distinguished by a northern and southern prefix. The Chantrey Group is correlated with other Paleoproterozoic sedimentary sequeneces in the Rae domain including the Amer Piling, and Penrhyn groups (Hinchey et al., 2007; Ryan et al., 2006).

Paleoproterozoic and older rocks are cut by three sets of undeformed diabase dykes. The the oldest, northwest-striking dykes are provisionally assigned to the ca. 1267 Ma Mackenzie swarm, whereas, east and east-northeast striking dykes have not been previously recognized, and have no known correlatives (Nadeau et al., 2005; Ryan et al., 2006, in prep).

Paleozoic sedimentary rocks outcrop in the northwest and east of the study area, but locally form a large proportion of the till lithology. Paleozoic clasts included in the till are inferred to be Cambrian to Ordovician, shallow-water platformal carbonates (Figure 4C), and include yellow to tan, muddy to sandy dolostones, minor quartzose carbonates and quartz arenite. In the study area, the Paleozoic rocks are horizontal to shallowly dipping and unmetamorphosed.

For the glacial transport studies, the most important, uniquely identifiable, indicator lithologies found in the till are derived from the Barclay Belt and the southern Chantrey Group. The southern Chantrey Group is dominantly composed of orthoquartzite (various colour including white, pink, green, violet and grey), with minor amounts of white, tremolite-bearing marble, grey to brown pelite, cordierite + biotite pelite, conglomerate and orange-grey calcsilicate rocks (Heywood, 1961; Frisch, 2000; Sandeman et al., 2005).

Regional ice flow

The study area is located more than 200 kilometers north of the Keewatin Ice Divide and was strongly influenced by ice flowing from Keewatin and McClintock ice divides Laurentide Ice Sheet during the last glaciation (Figure 5; Dyke, 1984). The regional scheme of ice flow was also influenced by an ice shelf located in Lancaster Sound and later in the Gulf of Boothia.

Dyke (1984) proposed that the resulting effect of this ice shelf on the interior of the ice sheet was to drain the ice and pull down the ice profile in a converging way towards the ice shelf area where the glacier ice was downwasted as icebergs. The ice flow was locally increased and possibly enhanced by a mobile bed under the ice sheet, and Dyke (1984) identified northeastward ice streams on the basis of long dispersal trains having sharp lateral contacts and consisting of carbonate debris. The initiation of the Boothia/Lancaster ice stream is probably younger than LGM, according to a Late Wisconsinian-Holocene age for the Navy Board Inlet moraines in Baffin Island (Dyke and Hooper, 2001; Dyke et al., 2003). The existence of a northward oriented ice-flow in Committee Bay area is probably dating from LGM (McMartin, 2003 a,b, 2005), and is followed by northeast ice flow directed toward Boothia ice stream.

During deglaciation, the McClintock Ice Divide was separating the ice flow towards the Lancaster/Boothia ice stream, and the ice flow towards the Beaufort Sea. During further deglaciation, a final northward ice movement emanating from the a glacial ice divide position is present in the southern part of study area (Dyke, 1984).Therefore, the regional glacial context suggests that shifting ice movement directions may be related both to shifts in the position of the Keewatin Ice Divide position itself, and to the proximity and influence of the calving bay in the Gulf of Boothia.

Till sedimentology

In the study area, till matrix texture varies considerably from south to north. In the southern part of the area, the terrain surface is covered with an almost continuous veneer of boulders, especially in the marine-inundated lands and meltwater channellized surfaces, whereas north of Simpson Lake, blocks are much more sparse on the till surface. In the south, the till contains a sandy-silty, non-calcareous matrix and a high proportion of clasts. In the north, the till matrix is siltier as compared to the south, and the pebble proportion decreases northward. These variations reflect the glacial transport of Paleozoic platform carbonates eroded from the Rasmussen Basin area, which are easily crushed to silt-size fraction during glacial transport.

Marine invasion

Elevation of the marine limit, as interpreted by Dyke (1984), ranges from over 240 m a.s.l. in the Pelly Bay area to around 180 m a.s.l. in the western part of the study area. Wave-washed outcrops, gravel beaches and ice-rafted debris are common in marine-inundated lands. The features are better developed on hills facing open paleo-sea, that were well exposed to the effects of waves during sea invasion (Figure 4C). In the Pelly Bay region, numerous large ice-rafted Paleozoic carbonate blocks (figure 4B) litter the surface, over till that contains no carbonate clasts. Marine shells are abundant below marine limit. Notably, marine shells are prominent between 205 and 180 m a.s.l. in the central plateau.

FIELD WORK AND METHODOLOGIES

Surficial geology field work was aimed at providing detailed information about the directions of ice flow and glacial transport direction of indicator lithologies. Striation measurements were collected in the main upland area as well as throughout most of the lowland areas. Despite plentiful bedrock outcrops in the area, collection of reliable striation measurements was hampered by three difficulties. Firstly, the rock type is dominantly coarse-grained granite and gneiss having a variable amount of mafic rafts. Typically, these rocks do not readily preserve glacially polished surfaces due to chemical and physical weathering of exposed surfaces. The best unaltered, glacially polished and striated surfaces were found by physically removing sediments at the edges of outcrops, especially on resistant K-felspar rich rocks, to expose fresh surfaces. The second problem is that the gneissosity and foliation in the Precambrian rocks is commonly parallel with the main, Phase 2 (to the northeast; see below) ice flow direction. Thus, striations and grooves on tops of bedrock surfaces are often deceiving, and require careful inspection. Also, stoss-side protected surfaces are rare and of poor quality because of the geometry of slab quarrying. Because of this problem, the strategy was to spend more time searching for protected surfaces on fewer, more favourable outcrops (Figure 4D). The third problem was that many outcrops can be severely wave-washed, and therefore physically eroded due to the elevated limit of marine invasion in the study area. Judicious examination of wave-protected outcrops sometimes yielded useful surfaces.

Lithological counts in surface till were undertaken at 26 sites, where 5-15 cm sized pebbles were counted mostly from mudboils in the surface of till. Counts were also completed at 17 sites in eskers. From 100 to 400 (in most cases more than 200) pebbles were counted at each site. Lithologic counts followed a consistent procedure, from determination of a counting area of about 2 to 10 square meters to the identification of rock types. Several different lithologies were considered, but the main focus was on unique indicator lithologies such as southern Chantrey Group rocks and Archean supracrustal rocks.

Field work also included esker (n=20) and till (n=10) sampling for heavy mineral analysis, marine shell collection (n=40) to determine sea level timing and variation, surficial mapping at detailed scale at critical sites, and examination of stratigraphic sections along the Arrowsmith River. The results of these field activities will be discussed in a subsequent Current Research report.

ICE FLOW HISTORY

Glacial geomorphology

The glaciated landscape is characterised by a variety of geomorphological features (Figure 6) from which it is possible to interpret the history of the last glaciation and deglaciation. The Quaternary geology map of the Boothia Peninsula region (Dyke, 1984), which coverered an area four times that of the present study, is based on field work, Landsat images interpretation and photo-interpretation by Boydell et al. (1979). To construct the geomorphological map of the Boothia Mainland (Figure 6), most of the already mapped geomorphological features were utilized from Dyke's 1984 map. However, a number of glacial features were added (including streamlined forms south of the study area), and in some places, minor corrections were made to Dyke's interpretations, especially in areas having sparse or equivocal glacial flow indicators.

Geomorphological features indicative of ice flow direction include drumlins, flutings, crag-and-tail features, ice-moulded bedrock forms, and rare occurrences of Rogen moraines. Glacial geomorphological features indicative of late-glacial or deglacial history include eskers, glaciofluvial transverse forms (rare), similar to those observed by Utting et al. (2002) south of the study area, end moraines (rare), ice-marginal glacio-marine deltas, and De Geer moraines. Combined with Dyke's geomorphology map, the detailed record and synthesis of micro-forms observed in the field (striae, grooves, rat tails and roches moutonnées) permitted reconstruction of the ice flow sequence. The remainder of this section details the macroscopic features in the area, and our interpretation of the ice flow history is portrayed in the subsequent section.

The streamlined forms and ice-moulded bedrock forms map ("macroforms" map, Figure 6), allow interpretation of the last important phases of glacial ice flow in the study area. Over most of the Central Plateau and the Rasmussen Lowlands, the macroforms indicate ice flow trending from 040˚ in the eastern part of study area and the extreme north-west part, to 070˚ in the north-central plateau, and approximately 060˚ in the Rasmussen Lowlands. North and east of Pelly Bay, groups of macroforms indicate flow towards NNE at 020˚. In the southern region, macroforms with 000˚ (N) and 020˚ (NNE) trends end abruptly and appear to locally crosscut flow to the NE (around 060˚).

Fields of streamlined forms (drumlins, flutings and crag-and-tail features) are scattered across the study area, and the largest field is located in the northeastern part of the area (Figure 6). Other fields are scattered in the Rasmussen Lowlands where till is not covered by marine clays, and in the central plateau where till cover is present in depressions between large bedrock outcrops. Scattered forms, mostly crag-and-tails, occur in places covered by discontinous till

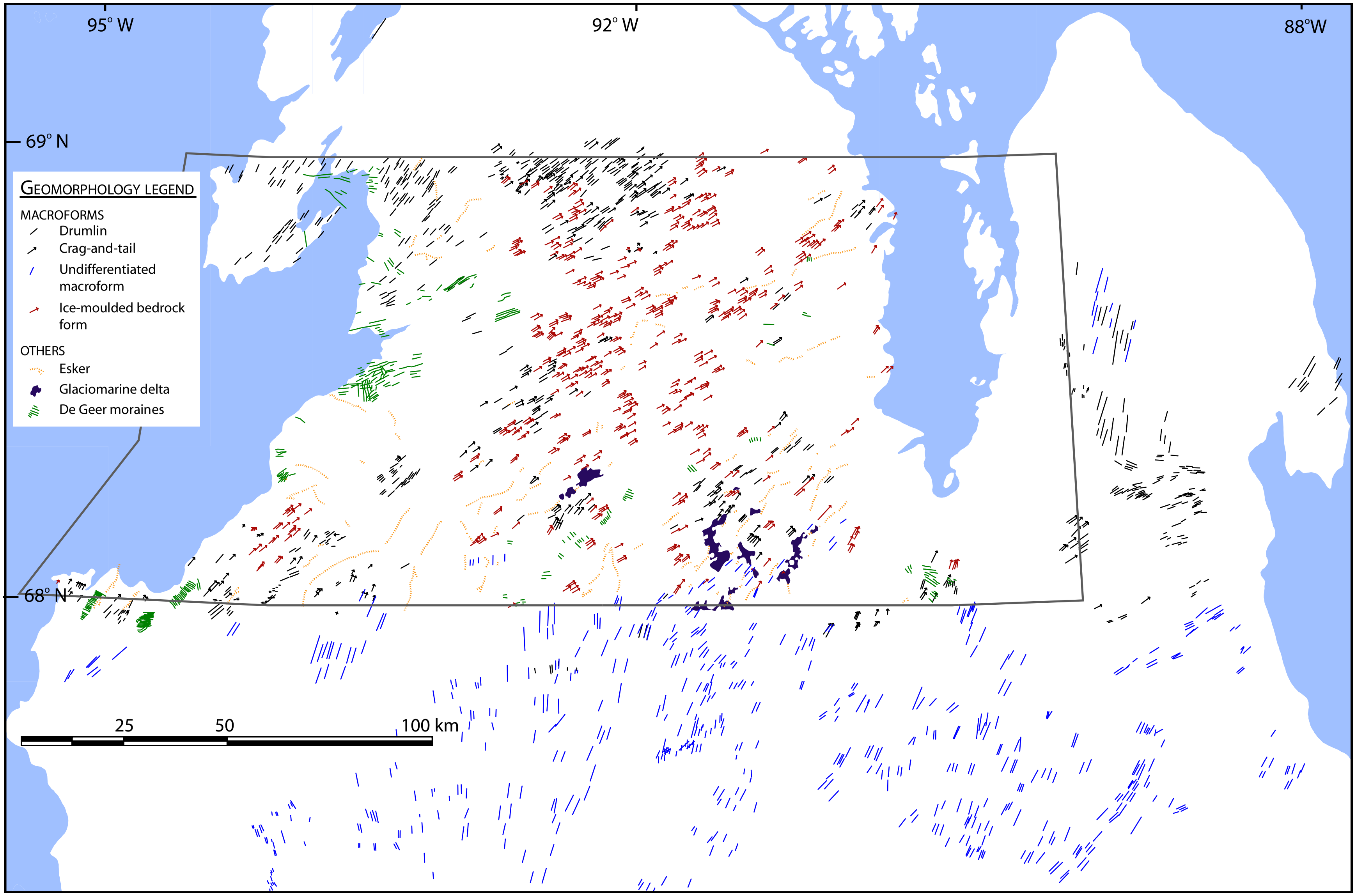


Figure 6. *Glacial geomorphology map, modified from Dyke (1984).*

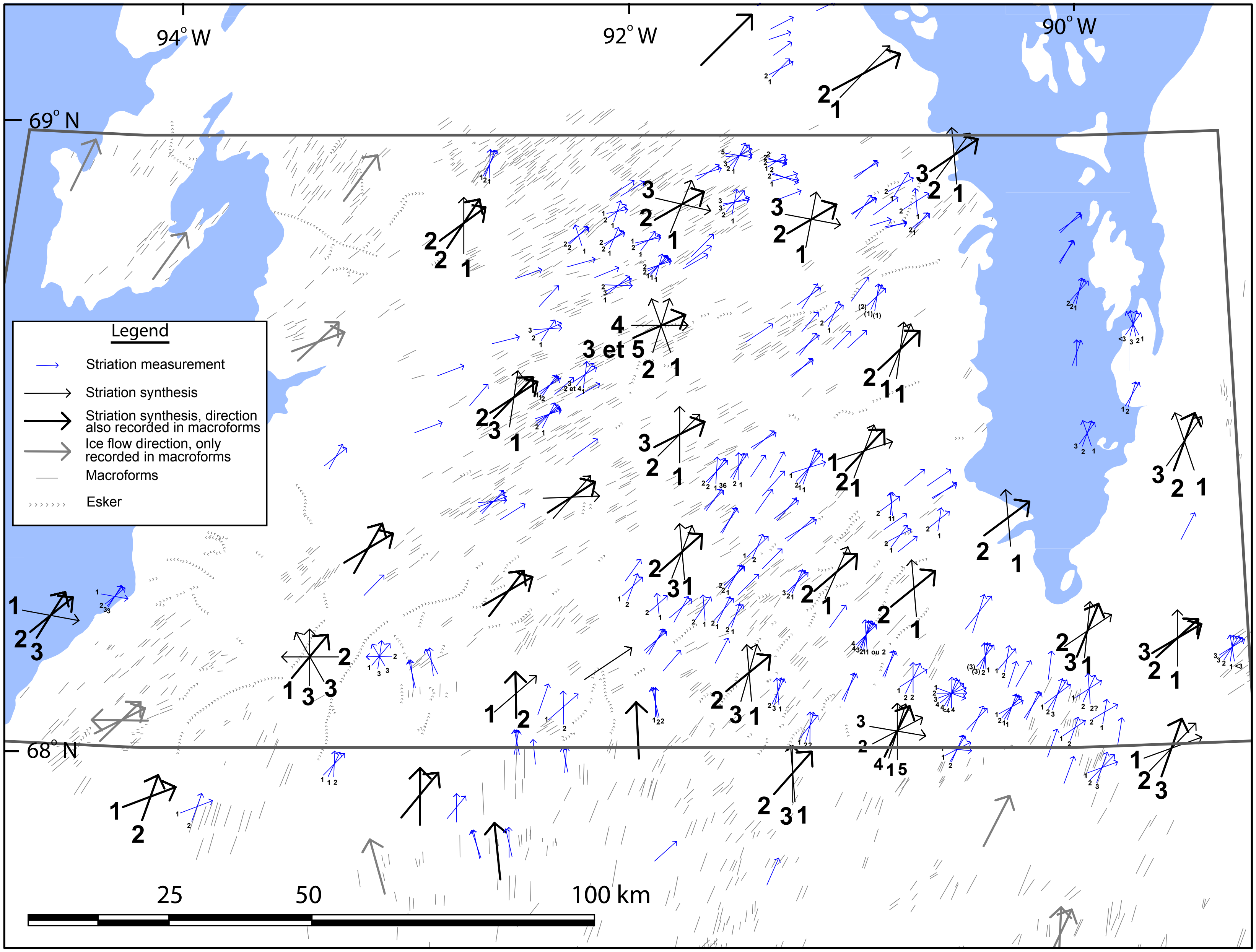


Figure 7. *Striations measurements map. Geomorphology modified from Dyke (1984).*

veneer. In most areas, the direction of streamlined forms in till is consistent with that of ice-moulded bedrock forms.

Ice-moulded bedrock forms, including rock drumlins, large roches moutonnées, or crag-and-tails (Figures 4A) are commonly identified from aerial photographs. However, in some areas, the structural 'grain' of the Precambrian rock controls hill forms, especially in areas having prominent map-scale fold structures. Thus, bedrock topography reflects the influence of structural grain and the main ice flow direction to the Northeast. In most of the study area, these two components are commonly parallel. The depth and development of scouring of bedrock forms suggests erosion by multiple glacial episodes during the Quaternary.

In the study area, eskers are more abundant in the south than in the north. Eskers in the south, some of which are more than 40 km long, are connected with a well developed esker system that reaches the interior of the Keewatin sector. Eskers trend north in the southern part of the study area, gradually shifting to NE in the central part of the study area. An exception to this occurs in major valleys in which eskers follow the topographic lows. Locally, and especially in the southern region, some bouldery glaciofluvial channels are linked to eskers. In the extreme west, the eskers trend west-northwest, reflecting the very last position of the ice front during deglaciation, which appears to have faced Rasmussen Basin. No esker was observed in the extreme eastern reaches of the study area (south and west of Pelly Bay), in part due to burial by thick cover of marine clays.

De Geer moraines are common in the Rasmussen Lowlands. The moraines are mainly oriented SW-NE, E-W and NW-SE and on this basis are interpreted to reflect a southeastern, southern or southwestern ice retreat. It is known that De Geer moraines can form by subglacial reworking of glacial sediments in ice crevasses under a floating ice front (Elson, 1957), or by glaciofluvial filling of basal crevasses (Beaudry and Prichonnet, 1991). In the Murchison Lake area of the central plateau, most of the De Geer moraines occur below 70 m a.s.l.. A few De Geer moraines occur in the lowlands south of Pelly Bay, and are interpreted to mark a southwestern ice retreat.

Glaciomarine deltas at an elevation of about 205 m a.s.l. occur as a group and mark the marine limit in the south-central part of the study area (Figure 6). The dimensions of the deltas range from 3 to 14 km long and 2 to 6 km wide. The total area covered by this group of 14 glacio-marine deltas is approximately 117 km^2, possibly indicative of a standstill of the ice front in this area. The deltaic surfaces are mostly flat, but are locally cut by kettles (rounded or coalescent), anastomosed channels, ice-contact channels, pebbly kames, ice-contact gravelly deltas and lateral terraces (Figures 4E and 4F). In some places, upstream eskers are directly connected with the downstream glaciomarine deltas.

Ice flow sequence

A complex, four stage ice flow sequence (Figure 8) is documented from the striation record (Figure 7) and the mapped macroforms (Figure 6). The southern limit of the ice flow study was extended to northern parts of NTS 56M, 56N, 56O and 56P for a more comprehensive ice flow history of the region (Figure 7). A map from Dyke et al. (1979) was also used to compile the macroforms near Chantrey Inlet (NTS 56M). From these data, it is possible to compare results from the Boothia Mainland with data from map areas NTS 56N and the northern part of NTS 56O (Ozyer and Hicock, 2006), and NTS 56P and the southern part of NTS 56O (McMartin et al., 2003a, b).

Phase 1: Northerly flow

Striations indicating a northward flow were commonly measured, and overprinting relationships in the central and northern part of the study area indicate that this flow was anterior to the main ice flow to the northeast (Phase 2). Notably, no macroforms related to phase 1 ice flow were observed anywhere in the study area, as it was overidden and obliterated by later major ice flows.

Phase 1, the earliest recorded ice flow, is interpreted to be related to ice flowing from a divide in the Keewatin region to the south, undisturbed by the occurrence of an ice stream in the Gulf of Boothia. Because the ice divide was located directly south of the study area, the flow was towards the north (Figures 7 and 8), ranging from NNW (350˚) to NNE (010˚). This ice flow probably lasted for a significant period of time, and occured possibly during LGM (Last Glacial Maximum) as suggested by Little (2001), McMartin et al. (2003a, b).

Phase 2: Northeasterly flow

Phase 2 NE-flow is the predominant event observed in the Boothia Mainland area. It is recorded from the vast majority of macroforms, except in the south, and from northeast-trending striations present on the top of almost every glacially polished outcrop surface. With the exception of the southern part of the area, the phase 2 flow represents the last important glacial flow direction. Flow direction indicated by striations and macroforms shifts from 040˚ in the eastern part and the extreme northwest Rasmussen Lowlands, to 070˚ in the north-central plateau, and 060˚ in the Rasmussen Lowlands. At several locations, the NE-trending striations cross-cut the older phase 1 northerly striations (figures 7 and 8).

Phase 2 is interpreted to be caused by the occurrence of an ice stream in the Gulf of Boothia identified by Dyke (1984). At many locations, the striations record a transition from Phase 1 (N) to Phase 2 (NE). This transition may be related to the strengthening period of this ice stream in the Gulf of Boothia. This important ice flow phase extends to the south of the study area (shown by the dotted lines in Figure 8). Phase 4

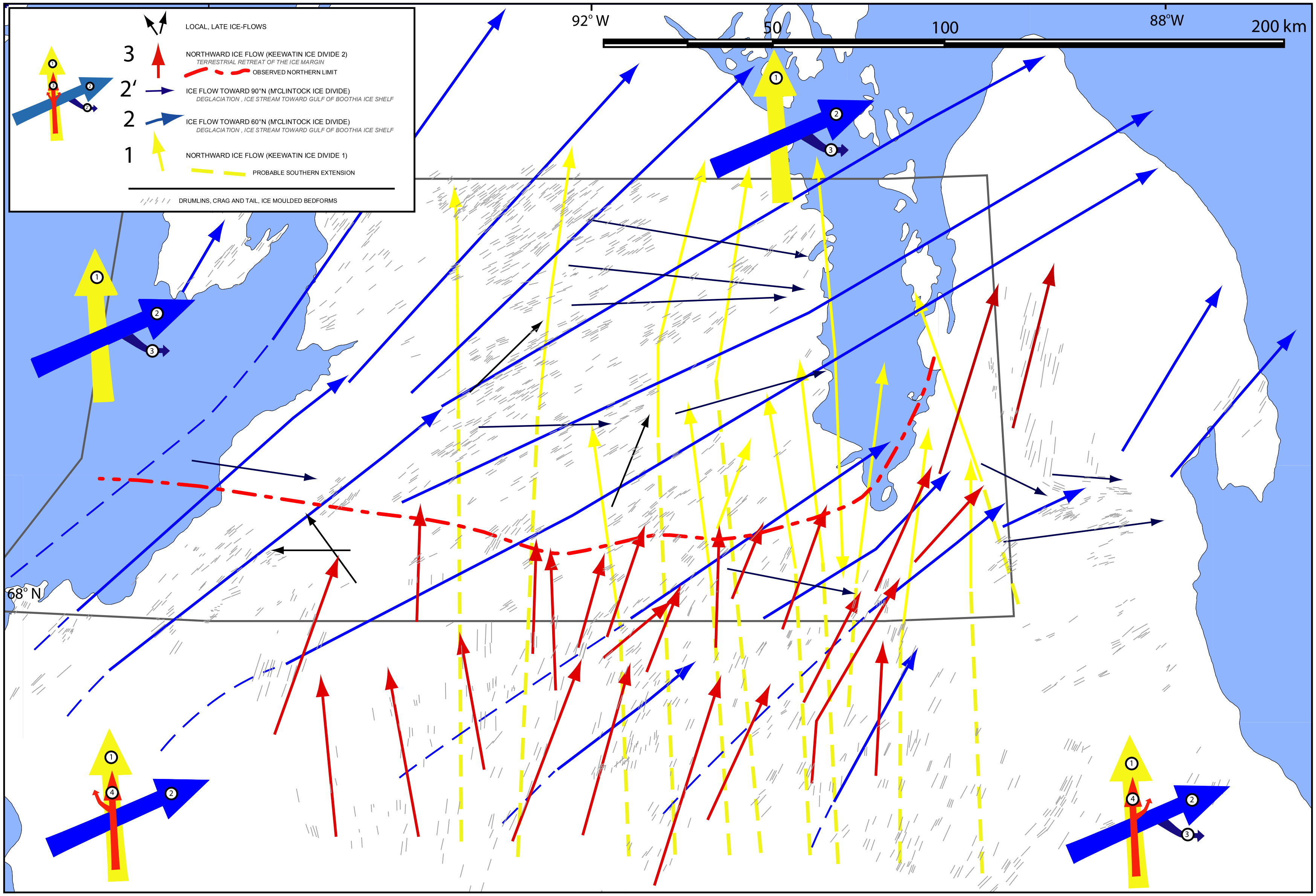


Figure 8. *Ice flow lines, with macroforms modified from Dyke (1984).*

Phase 3: Easterly flow

Phase 3 is broadly a subset of phase 2, deviating from a northeast (040° to 070°) trends to an easterly trend (085° to 100°). On Boothia Mainland, this ice flow phase was short in duration, because very few polished surfaces recorded it, and no macroforms are associated with that ice flow. While in most places it is the last ice flow recorded, the striation chronology at a few localities suggests a return back to a NE ice flow after Phase 3.

Phase 3 ice flow existed probably while the calving bay in Committee Bay was dragging the ice eastward (see McMartin et al., 2003a,b). This phase occurred at the very end of the glaciation, probably close to deglaciation.

On Simpson Peninsula, Phase 3 is observed in macroforms only, because no new field work was done in this area hence no striations were measured. It is not known whether the flow towards the east or the flow towards the NNE occured last on Simpson Peninsula, as no good relative chronology indicator is available.

Phase 4: Northerly flow in the south

Phase 4 is a late stage ice flow recorded by diverging NNW to NNE striations and macroforms (Figure 8), and whose northerly limit lies across the southern part of the present study area. This northward ice flow cross-cuts the NE flow (Phase 2) at many localities.

In the Simpson Lowlands, the Phase 4 NNE ice flow seems to be well represented with the occurrence of numerous 000° and 020° oriented macroforms. Those macroforms extend the limit of Phase 3 flow far to the north compared with the rest of the study area. The cause for this later ice flow in Simpson Lowlands is difficult to explain, although these macroforms may be originating from relicts ice flows.

Locallised late-stage ice flows

The effects of late-stage ice flows are observed at a few localities throughout the study area. These late-stage flows may be related to topography or position of the ice front during deglaciation. Furthermore, these sites are not related to significant development of macroforms, and on this basis are considered to reflect minor shifts in ice flow directions.

Two sites located in the centre of the Central Plateau record a late ice flow shift from 060° toward 040° and from 045° toward 020°. Very few macroforms are associated with this movement and therefore it is not included in the phase 3 ice movement.

In the southwestern corner of the study area, NW- and W-trending striations are present that chronologically follow the NE flow (Phase 2), but predate the N flow (Phase 4). These may be related to a shifting ice flow towards a calving bay located in Rasmussen Basin during some stage of ice retreat.

Glacial transport of erratics

Outcrops of Chantrey Group Proterozoic rocks and Archean supracrustal rocks (Figures 2 and 9) located up-ice (south) of the study area permitted study of the dispersion of glacial erratics through petrographic counts, and evaluation of the direction and distance of glacial transport. We report the results of two distinct lithologic groups: Chantrey Group orthoquartzite rocks (Figure 9A) and the remainder of the supracrustal rocks, which we refer to as *non-orthoquartzites* (Figure 9B). A model of glacial dispersal from the successive ice flow directions is presented for the orthoquartzites (Figure 10).

Orthoquartzites (Figure 4G) dominate the eastern portion of the southern Chantrey Group, are less prevalent in the middle part, and are rare in the western part (Sandeman et al., 2005; Frisch 2000). The non-orthoquarzite supracrustal rocks are dominated by metapsammite and metapelite compositions, with common biotite, garnet or andalousite porphyroblasts (Figure 4H). Such compositions are reported to outcrop abundantly in the central zone, and more sporadically in the remainder of the southern Chantrey Group (Frisch 2000; Sandeman et al. 2005). Erratics of metasedimentary and metavolcanic rocks from the Barclay belt located north of the Chantrey Group belt comprise fine-grained metamorphic rocks that are very similar in appearance to the porphyroblastic metapelites of the southern Chantrey Group, resulting in a source area that is quite large for porphyroblastic pelites. There are no known occurences of orthoquartzite within the Archean greenstone belts in this area (Frisch, 2000; Sandeman et al. , 2005 Ryan et al. , 2006). As a result, orthoquartzite is a better lithology than the non-orthoquartzites for the study of glacial transport. Erratics of marble from the southern Chantrey Group were found (Figure 4I), but marble is not abundant in the area and therefore is not considered a good glacial transport indicator.

Glacial dispersal of orthoquartzites

The distribution of orthoquartzite clasts in till display a NE-oriented dispersal trail that is more than 175 km long and 100 km wide (Figure 9A). The main section, displaying the highest concentrations (over 1%), is ribbon-shaped relative to the outcrops of southern Chantrey Group orthoquartzite, but displaced further to the north than would be expected if ice flow responsible for the dispersal train was only directed towards the NE.

Based on the history of ice flow interpreted from the macroforms and glacial striae, a glacial dispersal model for the orthoquartzites is presented in Figure 10. Phase 1 flow first transported the orthoquartzites in a northward direction to an uncertain northerly extent (± 70km) as shown on Figure 10-1. Secondly, phase 2 flow to the NE re-entrained the earlier dispersal train and eroded new material from the orthoquartzite outcrops in the southern Chantrey Group (Figure 10-2). Finally, late transport of orthoquartzites to

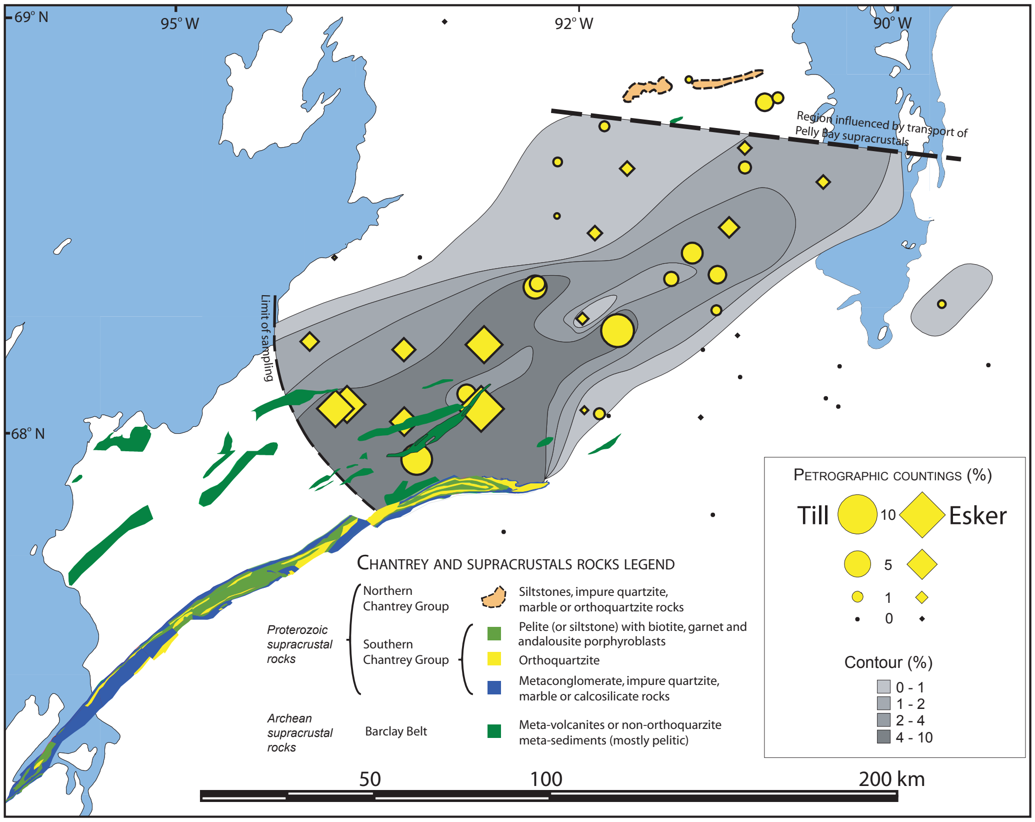


Figure 9A. *Orthoquartzite erratics in 5-15 cm fraction of till and esker samples (n= 100 to 400)*

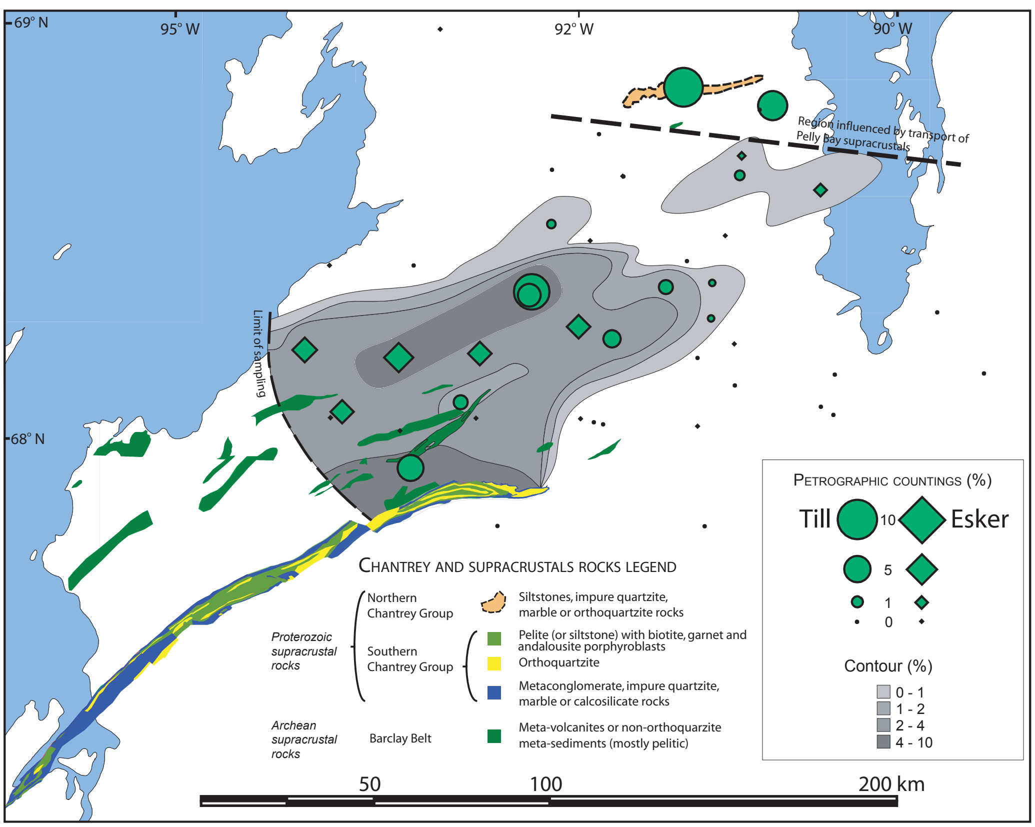


Figure 9B. *Non-orthoquartzite supracrustal rocks erratics % in 5-15 cm fraction of till and esker samples (n= 100 to 400).*

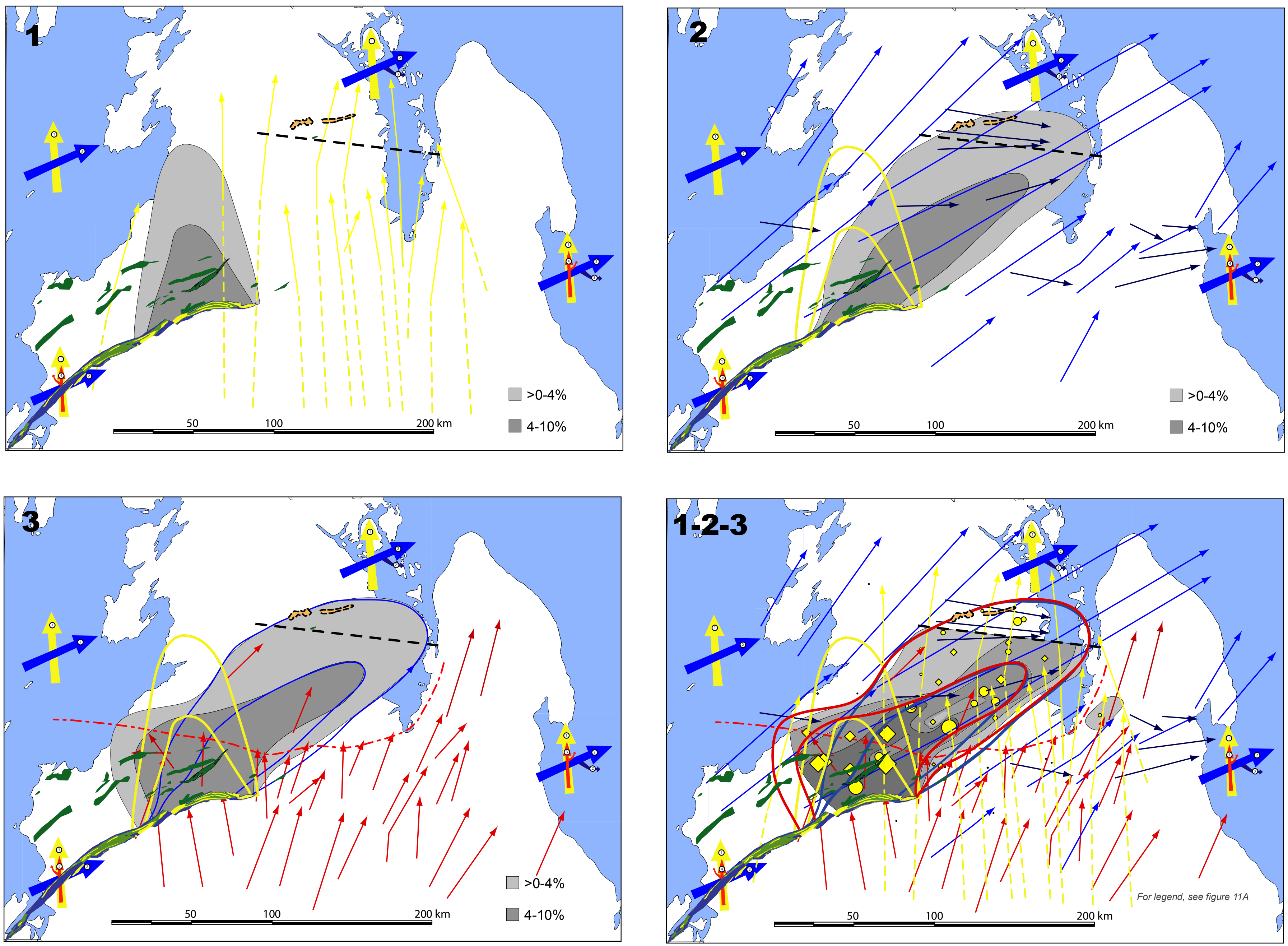


Figure 10. Glacial transport dispersal model for orthoquartzites, in chronological order from 1 to 3, and synthesis figure featuring original lithological counts and contours.

the north (Phase 4 flow) is also apparent in the Rasmussen Lowlands, as shown by the northward deflection of the western part of the dispersal train in Figure 10-3. The final dispersal train thus represents a palimpest train as described by Parent et al. (1996).

Glacial dispersal of non-orthoquartzites

The non-orthoquartzites (mostly pelites) display a NE-oriented dispersal train that is 100 km long and 50-75 km wide (Figure 9B). Clast concentrations of non-orthoquartzites are lower, and the dispersal train is shorter, relative to the orthoquartzite dispersal train. The glacial transport distance of the massive and hard orthoquartzites is greater than that of the fine-grained pelites, reflecting differences in hardness and resistance to abrasion. The shape of many pelite pebbles in till is a typical bullet shape, contrasting with the prismatic and rounded shapes of granitic gneiss erratics and the spherical and usually well-rounded shapes of the orthoquartzite erratics.

Glacial transport of marbles as an analog to kimberlite glacial dispersal

A map showing the transport of block-sized marble erratics (Figure 4I) from the northern Chantrey Group from north of Simpson Lake is provided in Figure 11. A ground traverse revealed the location of 15 to 200 cm-large marble erratics on the till surface along a 6 km transect. The location of these erratics reflects glacial transport in the direction of the main flow at 060° in this area. This map also gives an idea of the minimum distance of glacial transport (3 km) for blocks of a rather non-resistant lithology outcroping in a small area. Glacial erosion, transport and deposition of marbles can be an analog to the dispersal of kimberlite rocks, although kimberlites commonly form depressions rather than bedrock hills. Nevertheless, glacial transport in this area should be qualified as significant (at least 3 km for marbles blocks). As shown earlier, the southern Chantrey Group rocks have been dispersed great distances, over 175 km for orthoquartzites. Dyke (1984) used similar maps showing dispersal trains of Paleozoic carbonates in till to illustrate very long transport distances: "[…] as much as 70% of the till on [the carbonate dispersal train located north of the study area] came from a distance of 100 km or more. The actual distance of transport of material in the trains could be several hundred kilometres" (Dyke, 1984, p.13).

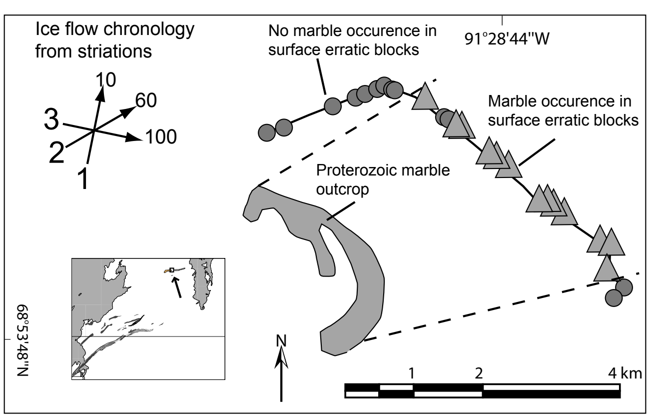


Figure 11. *Transport of block-sized marble erratics from a portion of the northern Chantrey Group. Location indicated by arrow in inset map.*

DISCUSSION

An ice flow history with 4 phases is observed in the Boothia Mainland area. It is possible to correlate, at least in part, this ice flow history with previous work in areas located further south (up-ice) of the study area (McMartin et al., 2003a and b; Ozyer and Hicock, 2006).

The ice flow chronology presented herein is in good agreement with the study of Dyke (1984), although we present microform evidence for an early northward flow (Phase 1). Dyke (1984) alleged that although previous authors supposed its existence (e.g. Hughes et al., 1977), he found no evidence for early northerly flow through field or air-photo interpretation. Recently, a Late Wisconsinian-Holocene age was attributed to the Lancaster/Boothia ice stream, as a Late Wisconsinian-Holocene age was indicated for the Navy Board Inlet moraines on Baffin island (Dyke and Hooper, 2001; Dyke et al., 2003), rather than Mid-Wisconsinian. From field observations in the Keewation ice divide area, a early northward ice movement from an E-W oriented Keewatin ice divide, prior to the establishment of the Boothia ice stream, was proposed by Little (2001). It was further substantiated by McMartin et al. (2003a, b; their Phase I ice movement) in the Committee Bay area, south of the present study area (see Figure 1). Ozyer and Hicock (2006) observed an early northward flow in the Committee Bay North area that diverted to the NNE at the southern limit of the present study area. Because numerous old N- and NNE-trending striations were found in the current study area, we extend the phase 1 N-flow of McMartin et al. (2003a and b) much further north.

The phase 2 NE ice movement in Boothia Mainland seems to correlate very well with the ice movement "Phase II" of McMartin et al. (2003 a, b) in the Committee Belt area. In the Committee Bay North area, Ozyer and Hicock (2006) presented a NNE ice movement (their Phase II) that is similar to Phase 2 ice flow in the present study, however their Phase II flow does not seem to bend to the NE but to the NNE at the southern limit of the present study area.

The Phase 3 easterly ice flow has correlatives within the extreme eastern part of the Committee Bay map area (McMartin et al., 2003a, b). It occurs as a local, late stage eastward ice flow. This ice flow has no equivalent in the Committee Bay North area studied by Ozyer and Hickock (2006).

Phase 4 was observed in the eastern part of the Committee Bay project area (020° direction, McMartin et al., 2003a, b), but not in the central and western parts, where it was probably obliterated by a younger NNW flow (Phase III of McMartin et al. 2003a, b). However, in NTS 56O-South sheet it is present at a few sites. In the Darby Lake area, this ice flow appears to be recorded as the northward, Phase III of Ozyer and Hicock (2006), altough that ice flow is not described to be as diverging as in Boothia Mainland.

The occurrence of an important group of glacio-marine deltas near the northern limit of northward ice flow probably indicates that Phase 4 occurred during a standstill of the glacier front around Murchinson Lake. Therefore, we propose to name this deglaciation episode "Murchison Lake retreat phase". Dyke (1984, p.13) shows this "moraine" on his map, but he does not discuss it in the text.

CONCLUSIONS

Based on the mapping of macroforms and striations, a complex, four phase ice flow history is interpreted for the Boothia Mainland area: 1) Phase 1 N flow, 2) shifting to Phase 2 flow to the NE, 3) a less prevalent Phase 3 to the E, and 4) a return to a Phase 4 flow to the north in the southern part of the study area.

All the regional ice flow phases (especially 1, 2 and 4) did impacted on the final distribution of glacial erratics in till, especially for the orthoquartzite rocks from the southern Chantrey Group. Erratics of these Paleoproterozoic rocks illustrate a palimpsest-shaped dispersal train in the area. Distance of transport is observed to be greater than 100 km for pelites and 175 km for orthoquartzites. Dispersal train studies will be useful for the study of glacial transport of kimberlite debris and minerals in till.

The northern limit of Phase 4 (as indicated by macroforms and striations) coincides with the location of a major group of glacio-marine deltas in the Murchison Lake area. We refer to this standstill of the glacier front as the "Murchison Lake Retreat Phase", and is slightly older than the Chantrey Moraine located 100 km south.

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