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New contributions to the ice flow chronology in the Boothia-Lancaster Ice Stream catchment area

Abstract

Within the Boothia-Lancaster Ice Stream (BLIS) catchment area, ice flow patterns were reconstructed based on the synthesis of striation directions and cross-cutting relationships, transport patterns of erratic boulders, glacial landforms, cold-based glacial landsystems, and ice-retreat chronology. New ArcticDEM data, high-definition satellite imagery and multibeam echosounder bathymetric datasets provided increased details on ice flow indicators. Convergent high-velocity ice flows through the BLIS main axis were major, persistent features in the northeastern Laurentide Ice Sheet through the last glaciation, and this study highlights intensity fluctuations and ice flow pattern variations that occurred during that time. Highly contrasting glacial geomorphology, notably in the abundance of moraines, reflects marked differences in ice-margin retreat rates and patterns during deglaciation between the western and eastern sides of the BLIS.

Résumé

Au sein du bassin d'alimentation du Courant de glace Boothia-Lancaster (BLIS), les patrons d'écoulement glaciaire ont été reconstruits à partir de la synthèse des directions et relations de recoupement des stries, des patrons de transport des blocs erratiques, des formes glaciaires, des systèmes terrestres de zone à glace froide, et de la chronologie du retrait glaciaire. Les données récentes de ArcticDEM, des image satellites haute définition et des relevés bathymétriques (échosondeur multi-faisceau) permettent de détailler les indicateurs glaciaires. Des écoulements convergents à haute vitesse au travers de l'axe principal du BLIS furent majeurs et pérennes pour le secteur NE de l'Inlandsis laurentidien lors de la dernière glaciation, durant laquelle cette étude souligne des variations d'intensité et de patrons d'écoulement glaciaires. Durant la déglaciation, la géomorphologie glaciaire contrastée, notamment par l'abondance des moraines, de part et d'autres du BLIS est reliée à des différences dans le taux et les ptrons du retrait glaciaire.

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Introduction

The Boothia-Lancaster Ice Stream (BLIS) was a major glacial dynamic feature of the northeastern Laurentide Ice Sheet (NELIS) and the Innuitian Ice Sheet, forming a zone of fast ice flow channelled in the topographic depression that runs across Committee Bay, Gulf of Boothia, Prince Regent Inlet, Lancaster Sound, and the northwestern part of Baffin Bay (Fig. 1). Many studies have contributed to the reconstruction of ice flow patterns and chronology in the BLIS and surrounding areas (Prest et al., 1968; Dyke and Prest, 1987a; Andrews, 1989; Dyke and Dredge, 1989; Fulton, 1995; Dyke et al., 2002; Dyke et al., 2003), including several that relied dominantly on satellite and digital elevation model (DEM)-based geomorphological interpretations (Boulton and Clark, 1990; Clark and Stokes, 2001; Kleman et al., 2006; De Angelis and Kleman, 2007; Margold et al., 2018). Outside of the study area, a relatively

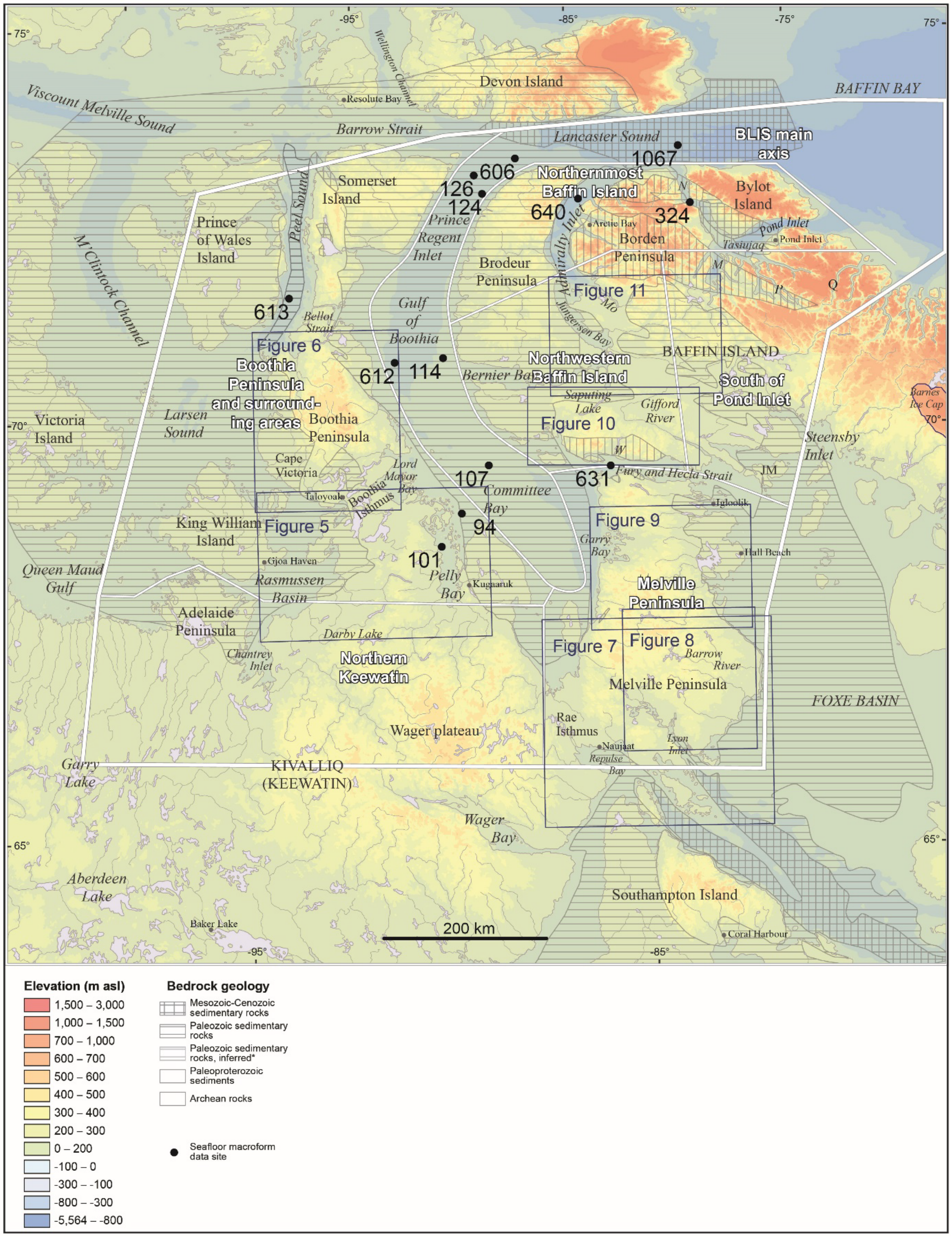


Figure 1. Location map, with geographical regions. The study area is outlined in white. N: Navy Board Inlet; M: Milne Inlet; Mo: Moffet Inlet; P: Paquet Bay; Q: Qiajivik; JM: Jens Munk island; W: Whyte Inlet. Bedrock geology modified from De Kemp et al. (2006) and Wheeler et al. (1996). *Paleozoic carbonate rocks inferred from bathymetric data interpretation. The locations of the seafloor macroform datasets are indicated by a black dot and corresponding number (See Figure 3 for additional information at some of these locations).

small fraction of the BLIS catchment area was part of the Innuitian Ice Sheet, mainly draining ice from Devon Island and the Wellington Channel (England et al., 2006). In general, data on pre-deglacial phases of ice flow within the BLIS catchment area is scarce, while being essential to glacial reconstructions of the NELIS. To update our understanding of the history of the BLIS, there is a need for the integration of historical and new geological and geomorphological observations on land and offshore with recent glaciological studies and numerical ice-sheet modeling results. This paper focuses on the Last Glacial Maximum (LGM; Clark et al., 2009) and deglacial ice flow patterns of the BLIS catchment area within the NELIS, using striation, glacial transport and landform records. New ArcticDEM data, high-definition satellite imagery, multibeam bathymetric datasets, lithological composition of till, and stratigraphy documenting ice retreat, are used to improve reconstructions of the local history of ice flow.

Previous work

The Quaternary geology of the study area (Fig. 1) was summarized at regional or continental scale in several publications (Prest et al., 1968; Dyke and Prest, 1987a; Andrews, 1989; Dyke and Dredge, 1989; Fulton, 1995; Dyke et al., 2002; Dyke et al., 2003). Sangamonian sediments are preserved on Boothia Peninsula and possibly on Prince of Wales Island, with no indication of deglaciation during the Wisconsinan prior to final deglaciation (Dyke and Matthews, 1987). On the coasts of northeastern Baffin and Bylot islands, the regional stratigraphy reflects several ice advances recorded by raised marine sediments dated with radiocarbon and amino acid methods to the Middle Wisconsinan or earlier (Klassen, 1993). The Eclipse Moraine on Bylot Island, thought to be associated with an ice stream in the Gulf of Boothia and Lancaster Sound (e.g. Dyke and Prest, 1987a), was then assigned a Middle Wisconsinan age, based on stratigraphic correlation with the pre-LGM marine sediments of northeastern Baffin and Bylot islands (Klassen, 1993). During the LGM, the presence of an ice shelf, rather than an ice stream, was suggested in the Gulf of Boothia and Lancaster Sound (Dyke and Prest, 1987b). Later, a Late Wisconsinan-Holocene age was reassigned to the Lancaster Sound-Boothia ice stream to reflect Late Wisconsinan (LGM?) ages on the Eclipse and Navy Board moraines on Bylot and Baffin islands (Dyke and Hooper, 2001; Dyke et al., 2002; Dyke et al., 2003).

The first reconstructions of LGM ice flow patterns in the study area were proposed for Melville Peninsula (Sim, 1962), Baffin Island (Ives and Andrews, 1963) and Foxe Basin (Andrews and Miller, 1979). LGM Laurentide Ice Sheet (LIS) ice streams, broadly defined at the ice-sheet scale by Denton and Hughes (1981), were identified on Boothia Peninsula and Prince of Wales Island on the basis on detailed geomorphology and glacial transport patterns by Dyke (1984). Dyke and Morris (1988) later proposed some key considerations behind the ice stream concept, including the convergence of streamlined landforms and the link between fine-grained tills and ice streams. Ice streams were subsequently identified from fieldwork on Melville Peninsula (Dredge, 1995), Rae Isthmus (Dredge, 2002), Bernier Bay (Hooper, 1996) and Steensby Inlet (Dyke, 2008). Deglacial geomorphology established patterns of ice retreat in northwestern Baffin Island (Dyke and Hooper, 2001). A large portion of the region was also covered by surficial geology mapping projects at the scale of 1:100 000 to 1:500 000 (see Kerr et al., Chapter 2 – this volume). Regional satellite and DEM-based geomorphological interpretations (Boulton and Clark, 1990; Clark and Stokes, 2001; Dyke and Hooper, 2001; Kleman et al., 2006; De Angelis and Kleman, 2007; Margold et al., 2018) reviewed the ice flow chronology and the continuity of sets of converging, elongated glacial landforms over large areas indicative of ice streams, notably in the Gulf of Boothia and Lancaster Sound during LGM and deglaciation.

Study area context

The BLIS catchment area is bounded in the southwest by ice divides in Keewatin (Keewatin Ice Divide: KID, Lee et al., 1957; ancestral Keewatin Ice Divide: AK, Dyke and Prest, 1987a) which were connected to the Keewatin Dome (KD) to the southwest, the M'Clintock Ice Divide (MID; Dyke, 1984) in the west, and ice divides and saddles connected to Foxe Dome (FD) in the east (Dyke and Prest, 1987a). FD and KD locations correspond to post-glacial rebound maximums (Dyke and Prest, 1987a; Andrews and Peltier, 1989; Simon et al., 2016a). The complex ice-movement chronology related to KID and AK migrations is treated in several papers (Boulton and Clark, 1990; McMartin and Henderson, 2004; McMartin et al., 2021), and need not be dealt with here. Initially also based on interpretations of the location of the postglacial rebound maximum (using the 10.5 ka cal BP shoreline isobases), the location of the MID during the LGM was positioned over M'Clintock Channel (Dyke, 1984), and later placed over eastern Victoria Island (Dyke and Prest, 1987b), western Prince of Wales Island (Dyke et al., 1992), Boothia Peninsula (De Angelis, 2007a) and Peel Sound (Stokes et al., 2016).

The area between Pelly Bay and western Committee Bay records some of the highest marine limit in the northern LIS, notably southeast of Boothia Peninsula (up to 240 m asl; Dyke, 1984; Giangioppi et al., 2003; McMartin et al., 2015) where it is much higher than on the western side of Boothia Peninsula (Dyke et al., 2005). The differences in marine limit elevations on either side of this peninsula suggests a higher rate of ice retreat and an earlier deglaciation in the BLIS main axis in comparison to the region west of Boothia Peninsula (see Dyke et al., 2003).

Undated with absolute methods, but attributed to LGM based on surface relationships with elongated streamlined landforms and a cover of only one sedimentary (marine) sequence, limits of the grounded ice associated with the BLIS in Baffin Bay are marked by faint grounding line features observed in multibeam bathymetric data at 1300 m depth (Brouard and Lajeunesse, 2017). The occurrence of recurring ice-rafted debris (IRD) layers in Baffin Bay sediments might indicate important fluctuations in the output of glacial material from icebergs calving from ice shelves and fed by ice streams around northern Baffin Bay, including the ice shelf beyond Lancaster Sound linked with the BLIS (Andrews et al., 1998; Simon et al., 2012; Andrews et al., 2014; Simon et al., 2014; Simon et al., 2016b).

Physiography and geology

The region is largely comprised of plateaus and some lowlands, dissected by fault-bounded, deep troughs, over-deepened by glacial erosion (Fig. 1). The plateaus are from 100 to >600 m asl in the west, and 400 to >800 m asl in the east; troughs are typically between 100 and 300 m deep (800 m deep in Lancaster Sound). Glacier-covered mountains and high plateaus, up to 2000 m asl, are found on Bylot Island and eastern Baffin Island. The bedrock (Wheeler et al., 1996; De Kemp et al., 2006) is dominantly crystalline rocks (felsic gneiss, granite and metamorphosed supracrustal rocks) from the Precambrian Canadian Shield; weakly metamorphosed Proterozoic sediments occur around Fury and Hecla Strait and on parts of Borden Peninsula and Bylot Island (de Kemp et al., 2006). Paleozoic carbonate rocks, with sandstones at the base in places, occur on parts of Boothia Peninsula, King William Island, western Baffin Island and eastern Melville Peninsula, as well as on the seafloor of Foxe Basin, Lancaster Sound, Gulf of Boothia and Barrow Strait. Tertiary sedimentary rocks, chiefly composed of poorly consolidated mudstones and sandstones, occur on Bylot Island, northern Boothia Peninsula, Somerset Island, and on the seafloor of Lancaster Sound and Hudson Strait.

Methodology

This chapter integrates field observations and interpretations of ice flow indicators and petrography of till clasts in the study area, and include more than 30 publications relating to fieldwork conducted between 2000 and 2018. Remote sensing-based mapping of geomorphological features at various scales, including glacial macroforms, the patterns of ice retreat and trajectories of subglacial meltwater drainage (eskers), and ice front positions (ice-marginal moraines, ice-marginal channels), were also considered in our ice flow reconstructions. Macroforms comprise streamlined landforms formed by subglacial activity (sliding), such as drumlins, megaflutings, mega-scale glacial lineations (MSGL), crag-and-tail landforms, ice-moulded bedrock forms, and till plumes (carbonate and non-carbonate, as observed on satellite imagery, cf., Tremblay and Lamothe, 2016; McMartin et al., 2021). The strong reliance on field data (including striations on bedrock) for synthesizing the paleo-glacial flow patterns is similar to that of Veillette et al. (1999), for example. The ages presented in this paper are in reservoir effect-corrected calibrated years before present (cal BP), and the ice margin retreat boundaries are from Dalton et al. (2020), which presents a slightly modified version of Dyke et al. (2003) for most of the BLIS catchment area.

Superposition of geomorphological features commonly provided the relative chronology of events. The finest of two cross-cutting sets (within striae or macroforms) was considered as being the youngest. For clarity, we chose to present macroforms (and the interpreted flowsets) as single arrows indicating the general direction of ice flow and relative chronology in a particular area (Fig. 2a). The direction of late deglacial ice flow indicators were variable in some areas due to local relief and for simplicity, we summarized those within 40° or less by a single arrow. Within each deglacial phase (i.e. Phase 3), ice flow indicator arrows do not necessarily represent synchronous events. As the ice front was receding during deglaciation, ice flows varied in direction and were occurring in a time-transgressive manner.

Results

Seafloor mapping

High-definition multibeam bathymetric data collected from the Canadian Coast Guard Ship Amundsen during multiple expeditions (Ocean Mapping Group, 2019) was used to map submarine features, mainly megaflutings and drumlins, in the BLIS main axis, Fury and Hecla Strait, Foxe Basin, Admiralty and Navy Board inlets and in the basins west of Boothia Peninsula. Local marine geology interpretations were incorporated from Lancaster Sound (Li et al., 2011; Bennett et al., 2014), King William Island area (Shaw et al., 2019) and northwest of Baffin Island (Brouard and Lajeunesse, 2017; Brouard and Lajeunesse, 2019). In places landforms were not entirely mapped from the bathymetric data; however, images were generally clear enough to distinguish streamlined features from bedrock structures, other glacial landforms (eskers), or icebergs scour marks. The glacial macroforms on the seafloor indicate the last major ice flow direction and that locally the ice was in contact with the bed at the time — instead of floating. The locations of key, newly interpreted seafloor macroform datasets are presented on Figure 1, and Figure 2 shows the directions of ice flow indicators and ice flow traces inferred from them (and those from other studies). Images of seafloor topography in Navy Board Inlet (Fig. 3a), Peel Sound (Fig. 3b) and Lancaster Sound (Fig. 3c) represent examples of critical evidence for the geomorphological interpretation presented in the following sections.

Mapping of cold-based glacial landsystems

The mapping of various types of cold-based glacial landsystems (Dyke and Savelle, 2000; Dyke and Evans, 2003; McMartin et al., 2021) allows identification of areas where, during most or part of the last glaciation, cold-based ice conditions existed. In the BLIS catchment area, cold-based glacial landsystems were delineated primarily from field observations (weathering of outcrops, presence of regolith, absence or scarcity of striations and roches moutonnées; Figs. 4a-e), aerial photographs, satellite imagery and DEM interpretations (Tremblay and Paulen, 2012; Leblanc-Dumas et al., 2014; Tremblay et al., 2016a; Tremblay, 2017; 2018). Our mapping also incorporated results and concepts from previous geomorphological mapping in the northeastern Canadian Arctic (Dyke, 1993; De Angelis, 2007b), Melville Peninsula (Dredge, 2000), Baffin Island (Sugden and Watts, 1977; Sugden, 1978; Miller, 1980; Dyke et al., 1982; Andrews et al., 1985; Marsella et al., 2000), and concurrent mapping in Keewatin by McMartin et al. (2021).

The majority of the study area consists of warm-based glacial landscapes (see McMartin et al., 2021), displaying strong amounts of glacial erosion, such as streamlined outcrops and bedrock hills (Fig. 4f), linear glacial erosion in fiords and glacial valleys (Fig. 4g), and various glacial macroforms. However the cumulative amount of glacial erosion over the past glaciation(s) is spatially variable (Tremblay and Gosse, 2019), and in some places was not enough to remove the inherited weathering of the bedrock and the regolith. On Figure 2, the "Undifferentiated cold-based" glacial landsystems refer to terrains that were subject to varying duration and extent of cold-based conditions as supported by geomorphological evidence of little to moderate amounts of glacial erosion and moderate to high proportions of weathered bedrock, and are in many cases a simplified version of the mapping presented in the field reports. For example, the regions mapped as undifferentiated cold-based glacial landsystems on the plateau north of Fury and Hecla Strait (Fig. 2) refer to a mix of cold-based glacial landscapes (little to no evidence of glacial erosion; Figs. 4a, 4b and 4c) and intermediate cold-based glacial landscapes (little to moderate evidence of glacial erosion, see McMartin et al. 2021; Figs. 4d and 4e). In the core of a cold-based glacial landscape, there is a general absence of striations, roches moutonnées and glacial macroforms. The ground is typically covered with a mix of regolith (Figs. 4a and 4b), boulders, weathered till, and glaciofluvial sediments often associated with ice-marginal channels (Fig. 4c). Crystalline bedrock is generally moderately to extensively weathered, occasionally with tors (Dyke, 1993) and inselbergs interpreted to have formed before the Quaternary glaciations (Tremblay, 2017). A small amount of glacial dispersal is sometimes observed, for example on Central Baffin Island (Dredge, 2004; Bonham-Carter et al., 2019). Large mountainous regions of Baffin Island were designated as undifferentiated cold-based regions (Staiger, 2005). Nonetheless, locally there may be evidence of warm-based local glaciers in valleys and cirques. Also, in warm-based landscapes, some sectors may have experienced cold-based deglaciation, while in other areas the conditions were cold-based for most of the pre-deglacial ice flow phases. Because of the iterative effects of glacial erosion, fluvial and glaciofluvial activity, and weathering on the landscape (Tremblay, 2018), in many places over the numerous plateaus of the study area, the boundaries between cold-based, intermediate cold-based and warm-based zones are transitional, rarely appearing as sharp boundaries in the landscape.

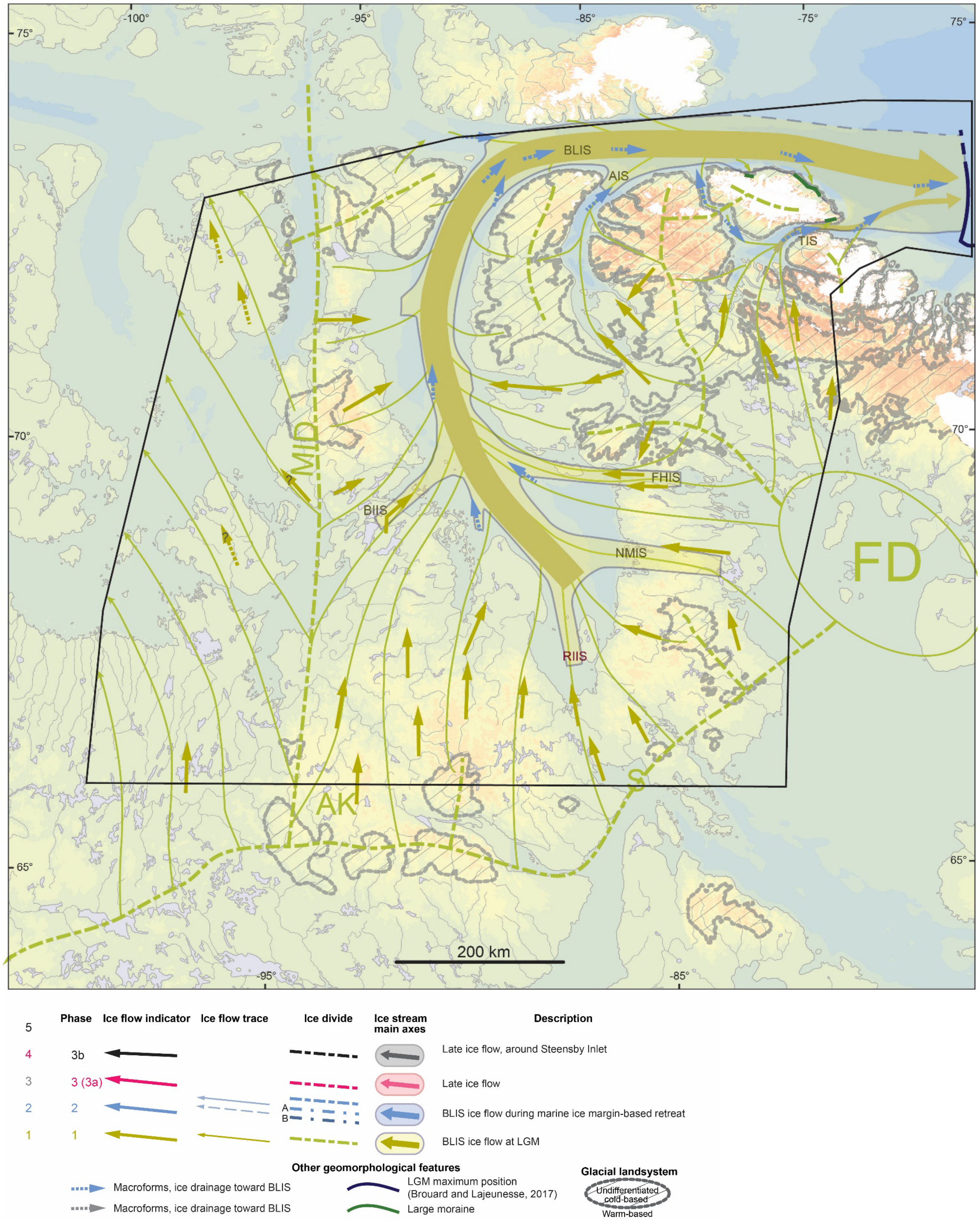
Cold-based glacial landsystems were also detected using cosmogenic isotopic data from northern Baffin Island (Staiger et al., 2006), and Melville Peninsula and Boothia Peninsula plateaus (unpublished; Tremblay and Gosse, 2019), and mineralogical data from Melville Peninsula and southern Baffin Island (Dredge, 2000; Leblanc-Dumas et al., 2015). In cold-based areas, the presence of minerals indicative of weathering (various clay minerals; e.g., kaolinite) in surface diamictons, and higher concentrations of ^{10}Be indicating inheritance in surface diamictons and bedrock samples, helps to emphasize the distinction between more or less glacially eroded terrains (Staiger et al., 2006; Margreth et al., 2016).

Regional ice flow synthesis

From the synthesis of general ice flow indicator directions, glacial transport patterns and cold-based glacial landsystems mapping (Fig. 2a), glacial flow events are parsed into three Phases. Phases 1 and 2,


LGM maximum extent, Brouard and Lajeunesse, 2017
NM
EM
MLM
GM
BM
AM
MM
CM
Dyke et al., 1992
Dyke, 1984
Hooper, 1996
Klassen, 1993
Dyke and Hooper, 2001; Tremblay, 2021
Januszczak 2007
Little et al., 2004
Dyke, 2008
Tremblay and Godbout, 2018
Tremblay, 2017
Dredge, 1995
Dyke and Savelle, 2006
[No reported striations]
Tremblay et al. 2009
Dredge, 1995 Tremblay and Paulen, 2012
Prest et al., 1968
Tremblay et al. 2007
Ozyer, 2011, S of Pelly Bay
Dredge, 2002
Dredge, 2002 Tremblay et al., 2016
Ozyer and Hicock, 2006 56N-W
McMartin et al., 2003 56P
McMartin et al., 2015
McMartin et al., 2013
Little, 2001 McMartin et al., 2003 56K
200 km
Phase
Ice flow indicator
Ice flow trace
Ice divide
Ice stream main axes
Description
5
4 3b
3 3 (3a)
2 2
1 1
A
B
Late ice flow, around Steensby Inlet
Late ice flow
BLIS ice flow during marine ice margin-based retreat
BLIS ice flow at LGM
Other geomorphological features
Macroforms, ice drainage toward BLIS
Macroforms, ice drainage toward BLIS
LGM maximum position (Brouard and Lajeunesse, 2017)
Large moraine
Glacial landsystem
Undifferentiated
cold-based
Warm-based

Figure 2a



-100°
-95°
-85°
-75°
75°
70°
65°
BLIS
AIS
TIS
MID
BIIS
FHIS
NMIS
FD
RIIS
AK
S
200 km
-95°
-85°
Phase
Ice flow indicator
Ice flow trace
Ice divide
Ice stream main axes
Description
5
4
3b
Late ice flow, around Steensby Inlet
3
3 (3a)
Late ice flow
2
2
A
B
BLIS ice flow during marine ice margin-based retreat
1
1
BLIS ice flow at LGM
Other geomorphological features
Macroforms, ice drainage toward BLIS
Macroforms, ice drainage toward BLIS
LGM maximum position
(Brouard and Lajeunesse, 2017)
Large moraine
Glacial landsystem
Undifferentiated cold-based
Warm-based

Figure 2b

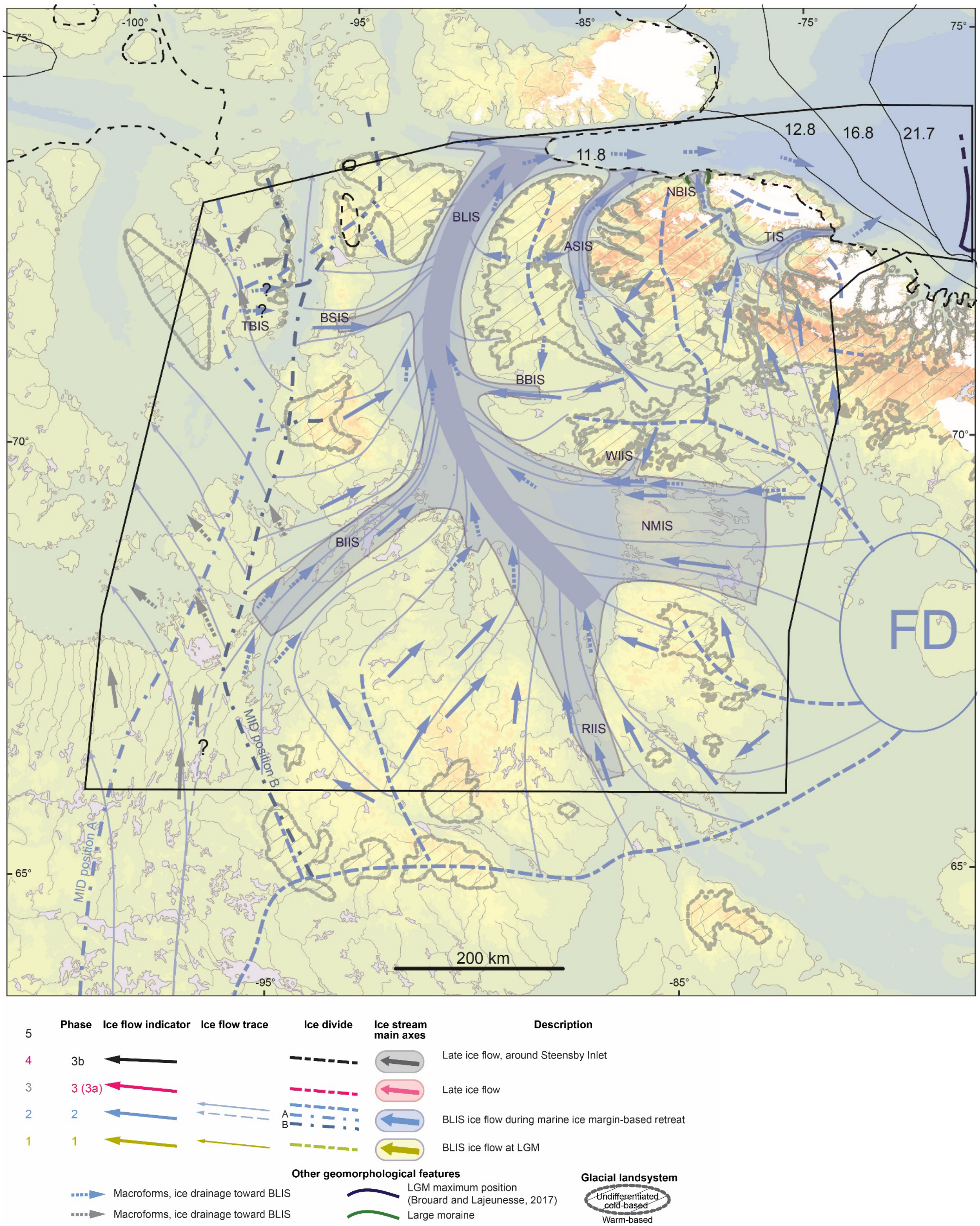
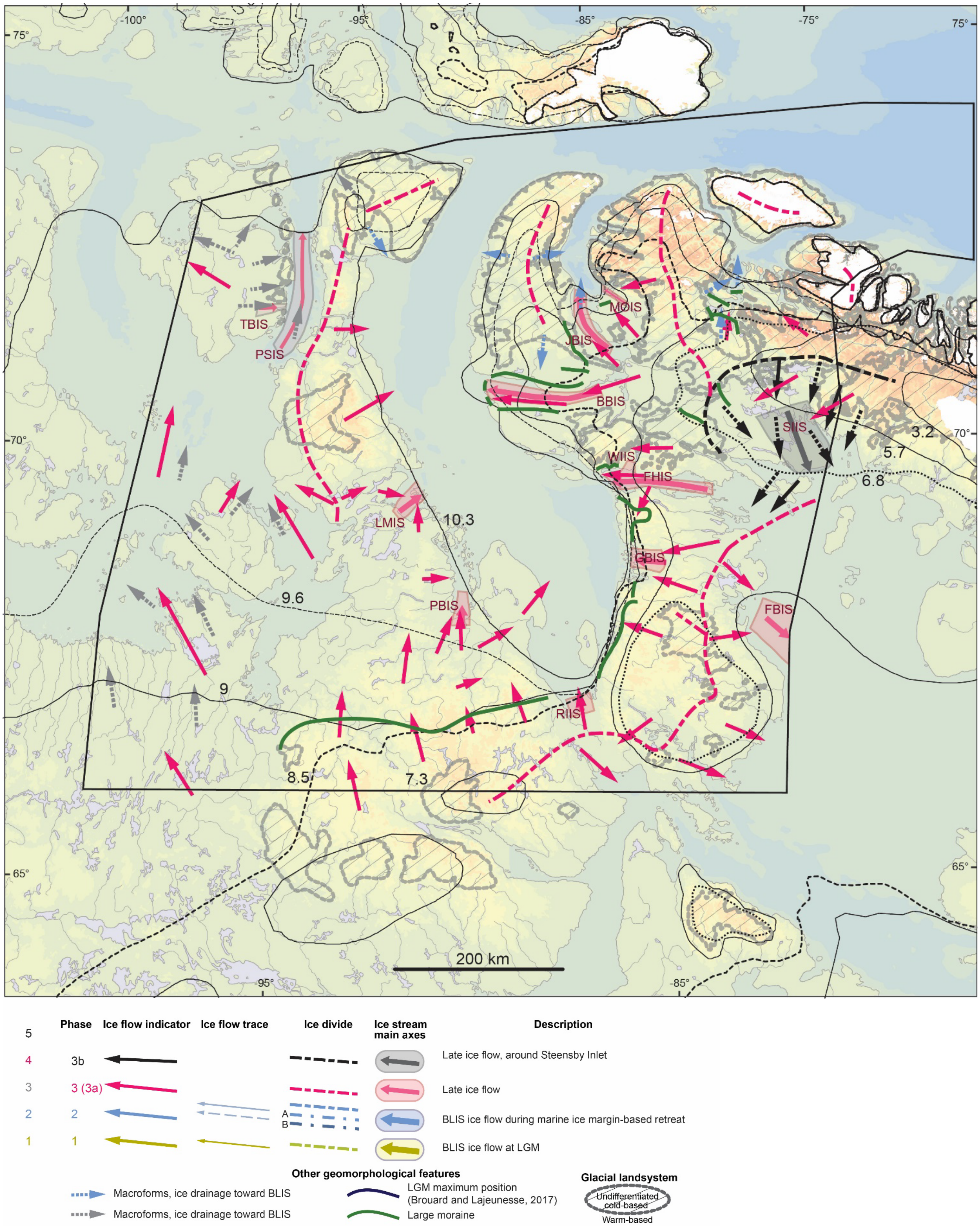

-100°
-95°
-85°
-75°
75°
70°
65°
12.8
16.8
21.7
11.8
NBIS
BLIS
ASIS
TIS
BSIS
TBIS
BBIS
WIIS
NMIS
BIIS
FD
RIIS
MID position B
MID position A
200 km
Phase
Ice flow indicator
Ice flow trace
Ice divide
Ice stream main axes
Description
5
4
3b
Late ice flow, around Steensby Inlet
3
3 (3a)
Late ice flow
2
2
A
B
BLIS ice flow during marine ice margin-based retreat
1
1
BLIS ice flow at LGM
Other geomorphological features
Macroforms, ice drainage toward BLIS
Macroforms, ice drainage toward BLIS
LGM maximum position (Brouard and Lajeunesse, 2017)
Large moraine
Glacial landsystem
Undifferentiated
cold-based
Warm-based

Figure 2c



-100°
-95°
-85°
-75°
75°
70°
65°
TBIS
PSIS
MOIS
JBIS
BBIS
SIIS
WIIS
FHIS
LMIS
GBIS
PBIS
FBIS
RIIS
10.3
9.6
9
8.5
7.3
3.2
5.7
6.8
200 km
Phase
Ice flow indicator
Ice flow trace
Ice divide
Ice stream main axes
Description
5
4
3b
Late ice flow, around Steensby Inlet
3
3 (3a)
Late ice flow
2
2
A
B
BLIS ice flow during marine ice margin-based retreat
1
1
BLIS ice flow at LGM
Other geomorphological features
Macroforms, ice drainage toward BLIS
Macroforms, ice drainage toward BLIS
LGM maximum position
(Brouard and Lajeunesse, 2017)
Large moraine
Glacial landsystem
Undifferentiated
cold-based
Warm-based

Figure 2d

Figure 2. Regional synthesis of complex ice flow chronology from terrestrial field observations (mostly from striations, lithological contents of the till pebble fraction, macroforms and stratigraphic records) in the Boothia-Lancaster Ice Stream (BLIS) catchment area and from offshore datasets. Dashed ice flow directions represent directions of macroforms only (onshore and offshore). The Phases are in general order of chronology relative to their time of initiation. The deglaciation ice margin chronology mapped as thin black lines (with various line types) is in calibrated years, with ^{14}C years corrected for reservoir effect, taken from Dalton et al. (2020); this updated ice margin chronology presents an almost unmodified version of Dyke (2003) for most of the BLIS catchment area. AK: Ancestral Keewatin Ice Divide, FD: Foxe Dome, MID: M'Clintock Ice Divide, S: saddle section of ice divides. a) Summary of regional ice flow chronology indicators from multiple studies; b) Phase 1 ice flow patterns; c) Phase 2 ice flow patterns. Phase 2 MID alternative configurations: A. maximum BLIS ice catchment area comprising the area between Garry Lake and Adelaide Peninsula on the mainland, and eastern Prince of Wales Island; B. minimum BLIS ice catchment area; d) Phases 3 ice flow patterns. AM: Autridge Bay Moraine; ASIS: Amiralty Inlet IS ; BBIS: Bernier Bay Ice Stream ; BIIS: Boothia Isthmus Ice Stream ; BLIS: Gulf of Boothia - Lancaster Sound Ice Stream ; BM: Bernier Bay Moraine; BSIS: Bellot Strait Ice Stream; CM: Chantrey Moraine; EM: Eclipse Moraine; FBIS: Foxe Basin Ice Stream; FHIS: Fury and Hecla Ice Stream ; GM: Gifford Moraine; LMIS: Lord Mayor Bay Ice Stream; MLM : Milne Inlet Moraines; MM: Melville Moraine; MOIS: Moffet Inlet Ice Stream; NBIS: Navy Board Inlet Ice Stream; NM: Navy Board Moraine; NMIS: Melville Ice Stream ; PBIS : Pelly Bay Ice Stream; PSIS: Peel Sound Ice Stream ; RIIS: Rae Isthmus Ice Stream; TBIS : Transition Bay Ice Stream; TIS: Tasiujaq Ice Stream; WIIS: Whyte Inlet Ice Stream.

the only phases for which flowlines (i.e. ice flow traces) are presented (Figs. 2b, 2c), refer to the entire study area, whereas Phase 3 is generally more local and occurred under a restricted ice cover during the late stages of deglaciation. For the purpose of this synthesis, the study area was divided into 7 geographical regions sharing glacial history similarities (see Fig. 1 for boundaries of these regions).

Lancaster Sound and Gulf of Boothia (BLIS main axis)

Phases 1 and 2

In the BLIS main axis, submarine macroforms, often MSGLs, are mapped in various locations from the seafloor topography data (Figs. 2a and 3a). Seabed morphology clearly indicates a spatial continuity of macroform directions from Prince Regent Inlet to Lancaster Sound; the continuity of macroforms is much less apparent from Barrow Strait to Lancaster Sound. Ice flow within or toward the main BLIS axis is also identified by occasional striations and terrestrial macroforms on the adjacent shores (Fig. 2a). Landforms perpendicular to ice flow on the seabed demarcate the LGM limit of the BLIS grounded ice in Baffin Bay (Brouard and Lajeunesse, 2017), about 100 km further offshore than the limit shown in Dyke et al. (2003) and Dalton et al. (2020). On Bylot Island, Dyke and Hooper (2001) and Dyke et al. (2002) interpreted the Eclipse Moraine as a frontal moraine, or lateral moraine, indicating diverging ice flow patterns during LGM at the limit of the ice stream located in Lancaster Sound (Dyke and Hooper, 2001; Dyke et al., 2002).

During Phase 1, ice streaming conditions existed in the deep marine trough between Committee Bay and Lancaster Sound, which acted as a channel for ice flowing out of the NELIS in a quasi-stable or growing mass-balance state, presumably during LGM (Figure 2b). Ice streaming conditions continued during periods of rapid, marine-based ice margin retreat within the BLIS main axis throughout Phase 2 (Fig. 2c), and until the ice retreated to the coasts of Boothia Peninsula and Committee Bay at approximately 10.3 ka (Phase 3; Fig. 2d). Figure 2c illustrates ice flow directions during ice retreat of the BLIS marine ice front through Lancaster Sound at approximately 11.8 ka. The sea bottom morphology suggests that the dominant source of ice flow within the Lancaster Sound part of BLIS was transiting through the Gulf of Boothia, with a minor contribution from subsidiary ice flows in Barrow Strait and the Innuitian Ice Sheet.

Boothia Peninsula and surrounding areas

Phase 1

South of Boothia Isthmus, the oldest striations are directed to the north and found on surfaces protected from a main northeastern ice flow (Fig. 2b and 5; Ozyer and Hicock, 2006; Tremblay et al., 2007; Tremblay et al., 2009). Additionally, evidence for an early northward glacial transport of Paleoproterozoic sedimentary clasts in till, prior to glacial transport and re-entrainment to the northeast during a later phase, is observed east of Rasmussen Basin (Tremblay et al., 2007). The striation record indicates that northward ice flow eventually merged with northeast (east) ice flow over Boothia Isthmus, where the early northward ice flow is no longer observed (Fig. 5) (Tremblay et al., 2009; Tremblay, 2017). Phase 1 ice flow predates the main northeastern ice flow associated with an enhanced ice stream activity over Boothia Isthmus at the beginning of ice retreat within the BLIS main axis (see Phase 2 below).

On Prince of Wales Island, an earlier north-northwest ice flow is indicated by the presence of Precambrian erratics over Paleozoic bedrock and north-northwest-aligned macroforms (Dyke et al., 1992). On King William Island where no striations are reported on surficial maps (Helie, 1984), Ozyer

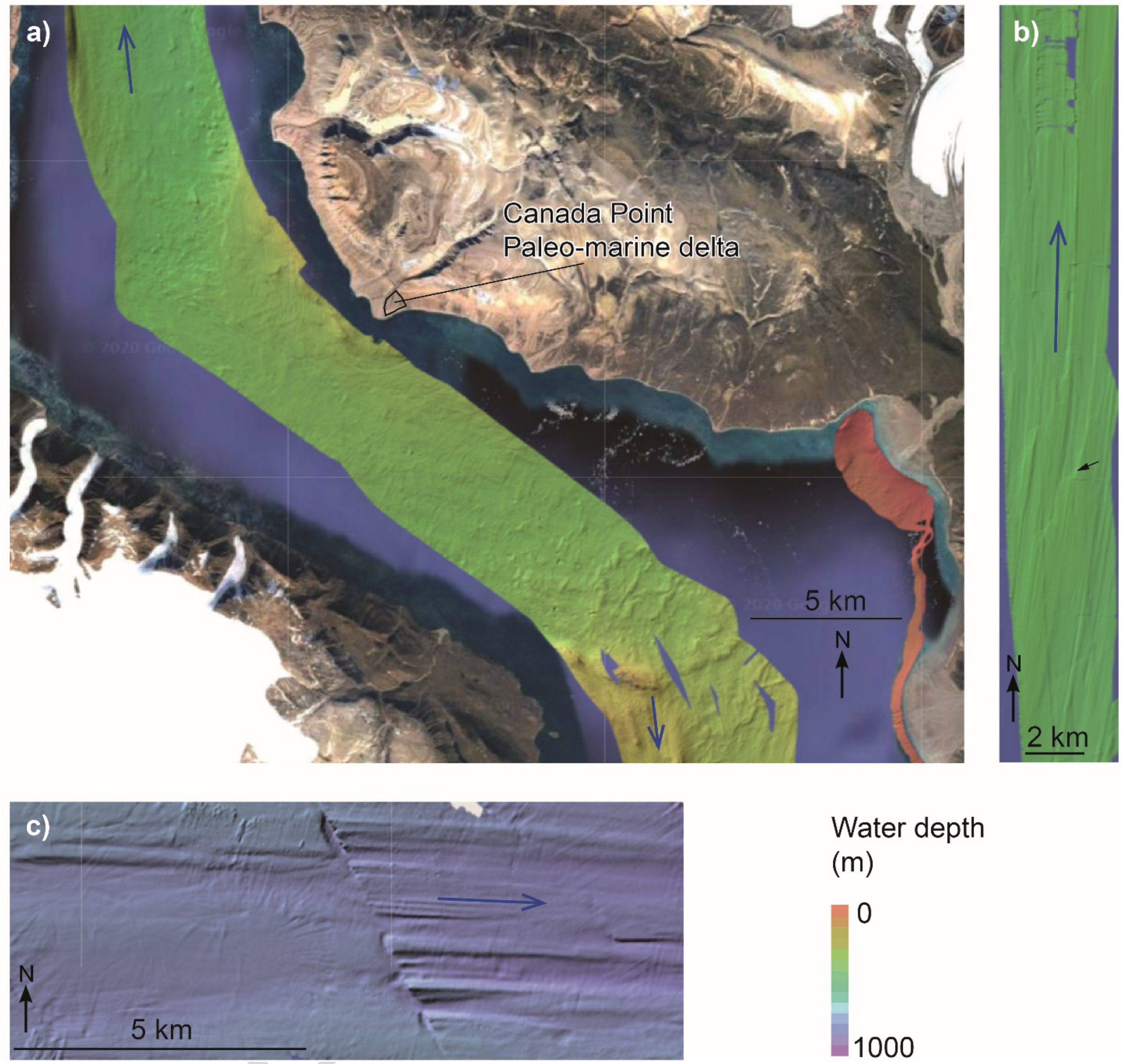


Figure 3. Seafloor features interpreted from multibeam bathymetric data (Ocean Mapping Group, 2019) : a) Central portion of Navy Board Inlet (site 324) showing an absence of streamlined landform indicative of cold-base conditions in the middle, also associated with the preserved pre-Wisconsinan Canada Point Paleo-marine delta (Klassen, 1993; Tremblay et al., 2020a), and streamlined landforms starting on either side of the cold-based zone (blue arrows); b) Peel Sound (site 613) with megaflutings (blue arrow) over flat seafloor probably underlain by Paleozoic carbonates (with cuestas?, shown by black arrow); and c) Lancaster Sound (site 1067) showing megaflutings and crag-and-tails (blue arrows).

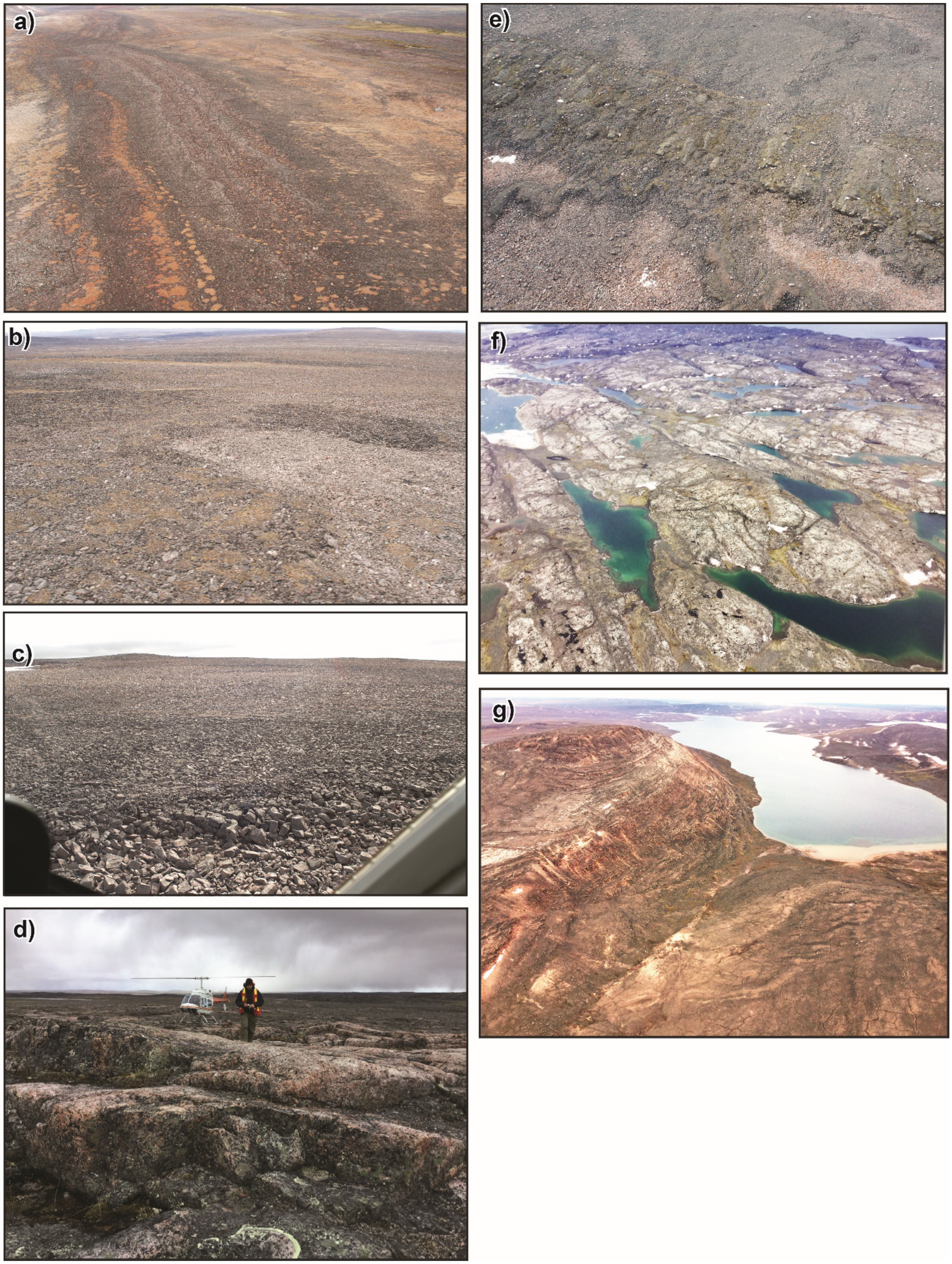


Figure 4. Field photographs typical of glacial landscapes from north of Fury and Hecla Strait: a) regolith and some regolith-till mix over Mesoproterozoic sedimentary rocks (CB); b) regolith and regolith-till mix over Archean felsic and mafic intrusive rocks (CB); c) regolitic bedrock exposed in

glaciofluvial channel, with regolith and regolith-till mix in the background, over Archean felsic granitic gneiss rocks (CB); d) Archean felsic granitic gneiss rocks outcrop, with no preferential erosion features (such as roches moutonnées) or striations (IB); e) roches moutonnées on gabbro dyke surrounded by till and regolith mix (warm-based to intermediate cold-based glacial landscape); f) ice-moulded bedrock form in Archean granitic rocks (WB); g) linear glacial erosion in Mesoproterozoic sedimentary rocks (WB). CB: cold-based glacial landscape; IB: intermediate cold-based glacial landscape.; WB: warm-based glacial landscape.

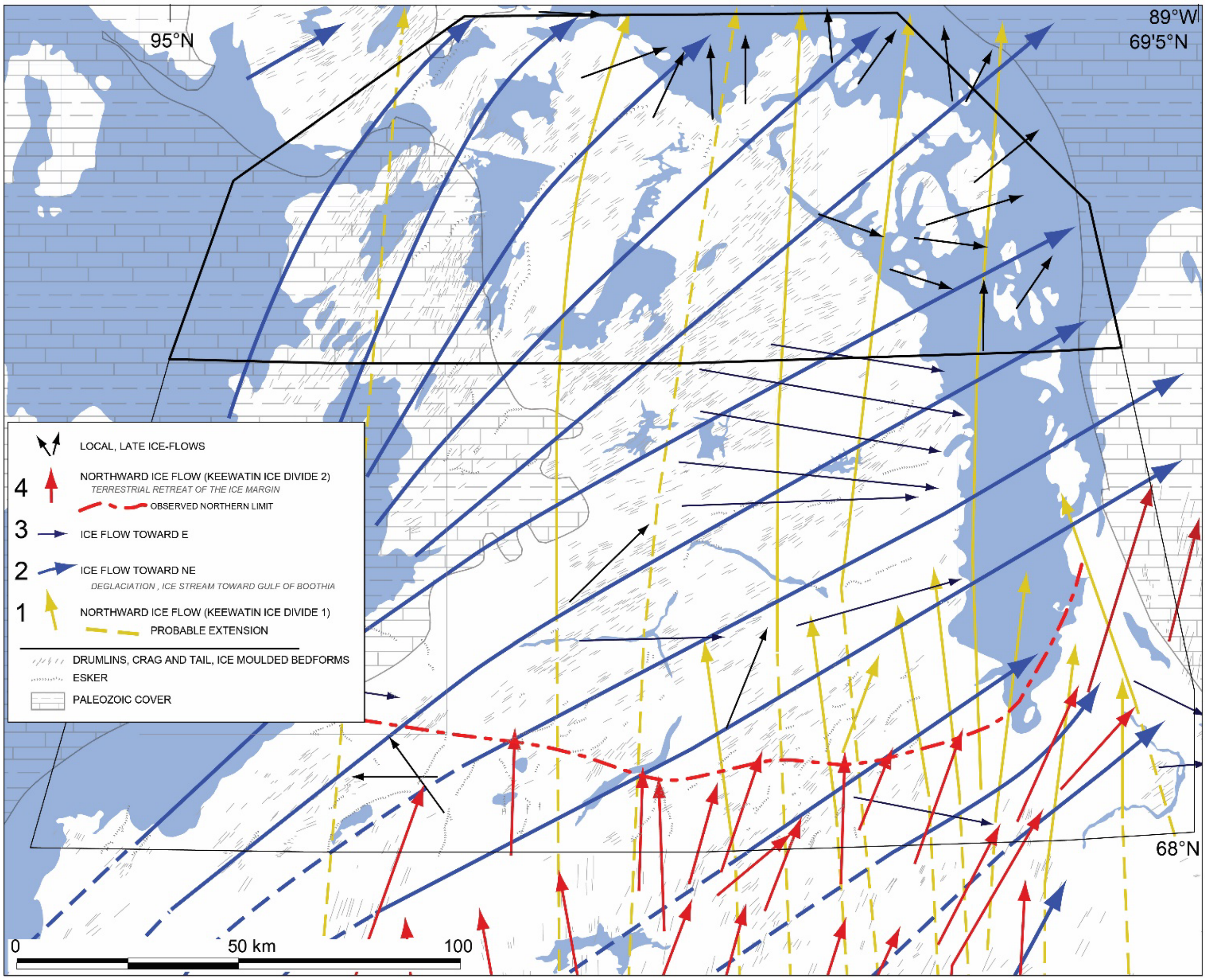


Figure 5. Ice-flow lines in northern Kivalliq (south of Boothia Peninsula), from Tremblay et al. (2009), with macroforms modified from Dyke (1984).

(2011) suggested that similar northwest-aligned macroforms, also seen on Fulton (1995), might date from an early ice flow.

The early northward ice flow south of Boothia Isthmus, also observed in northern Keewatin (e.g. Ozyer and Hicock, 2002; McMartin et al., 2003), and the Boothia Isthmus subsidiary ice stream draining into the BLIS, are attributed to ice sheet build-up and LGM. Although it is uncertain when the early northward ice flow commenced given the available data, it was probably active during LGM because of the relatively high abundance of associated striations and the evidence of northward glacial transport. During this Phase, the ice south of Boothia Isthmus was not directed toward the topographic lows in the

Gulf of Boothia, but instead flowed northward in a normal direction from the KID, supporting the idea of relatively thick ice not significantly influenced by topography during that period (McMartin et al., 2003).

Based on the mapping of cold-based terrains and the ice-flow patterns north of Boothia Isthmus and on Prince of Wales Island, we suggest that the MID central axis was anchored over central Boothia Peninsula and western Somerset Island, Peel Sound and eastern Prince of Wales Island during Phase 1, close to where Margold et al. (2018) placed the ice divide. During that time, the presence of active ice streams in Viscount Melville Sound and M'Clintock Channel (England et al., 2009) implies that the MID was located east of Prince of Wales Island (Stokes, 2008; Stokes et al., 2012; Margold et al., 2018).

The attributed LGM age for Phase 1 differs from that of Dyke et al. (1992) who assigned an early Wisconsinan age for the north-northwest ice flow pattern on Prince of Wales Island, and a location of the MID over M'Clintock Channel at LGM (Dyke and Prest (1987b). Alternatively, the absence of absolute dating on the older striae south of Boothia Isthmus could be indicative of an early northward ice flow formed during the Middle Wisconsinan, maybe during inception toward LGM (see discussion).

Phase 2

South of Boothia Isthmus, striations record ice flow to the northeast (Figs. 2c, 5 and 6), cross-cutting the previous northward flow (Tremblay et al., 2007; Tremblay et al., 2009). Paleoproterozoic erratics also indicate a strong northeast transport episode after the northward dispersion, forming a palimpsest glacial dispersal train (Tremblay et al., 2007). Based on striations and glacial transport patterns of kimberlite indicator minerals (KIMs) in glacial sediments south of Pelly Bay, Ozyer (2011) separated the northeast ice flow indicators into three separate ice flow events, which we all attribute to our Phase 2. We also attribute the northeast macroforms on northern Pelly Bay seafloor to Phase 2.

On Boothia Isthmus, a major ice flow event is associated with convergent ice flow toward the northeast and east, as indicated by striations, elongated macroforms (MSGLs) developed in fine-grained carbonate till, abundant glacially-moulded bedrock forms, and long distances of glacial transport (Dyke, 1984). A large dispersal train of carbonate till, approximately 200 km wide, indicates glacial transport of more than 100 km from source and is associated with the Boothia Isthmus Ice Stream (BIIS; Fig. 2c)(Dyke, 1984; Tremblay et al., 2009). The Bellot Strait Ice Stream (BSIS) forms another important zone of convergent ice flow further north on the topographic saddle between northernmost Boothia Peninsula and adjacent Somerset Island. There, east-northeast ice flow indicators are recorded by striations, macroforms and the glacial transport of Paleozoic carbonate rocks over Precambrian basement (Dyke, 1983; Dyke, 1984).

Paleozoic carbonate erratics found in the cold-based zone of the Boothia Peninsula plateau (Tremblay, 2017) could have been transported for a short distance under cold-based northeastward ice flows across Boothia Peninsula during Phase 2 (Fig. 6). This resembles the short-distance transport of Paleozoic carbonate erratics across cold-based terrain in central Baffin Island (Dredge, 2004). On the central Somerset Island plateau, the presence of sparse but widespread Precambrian erratics over Paleozoic carbonate bedrock (Dyke, 1983) suggests that the plateau was overridden by eastward or northeastward ice flows prior to or during the LGM (Phase 1), or as part of this early deglacial phase (Phase 2).

On southeastern King William Island, high-elongation ratio macroforms (MSGLs) in carbonate till oriented northeast on land and in seafloor data (Shaw et al., 2019) are associated with the strong ice stream flow on Boothia Isthmus. Seafloor data suggest that the northeastward macroforms are continuous between southeastern King William Island and Boothia Isthmus.

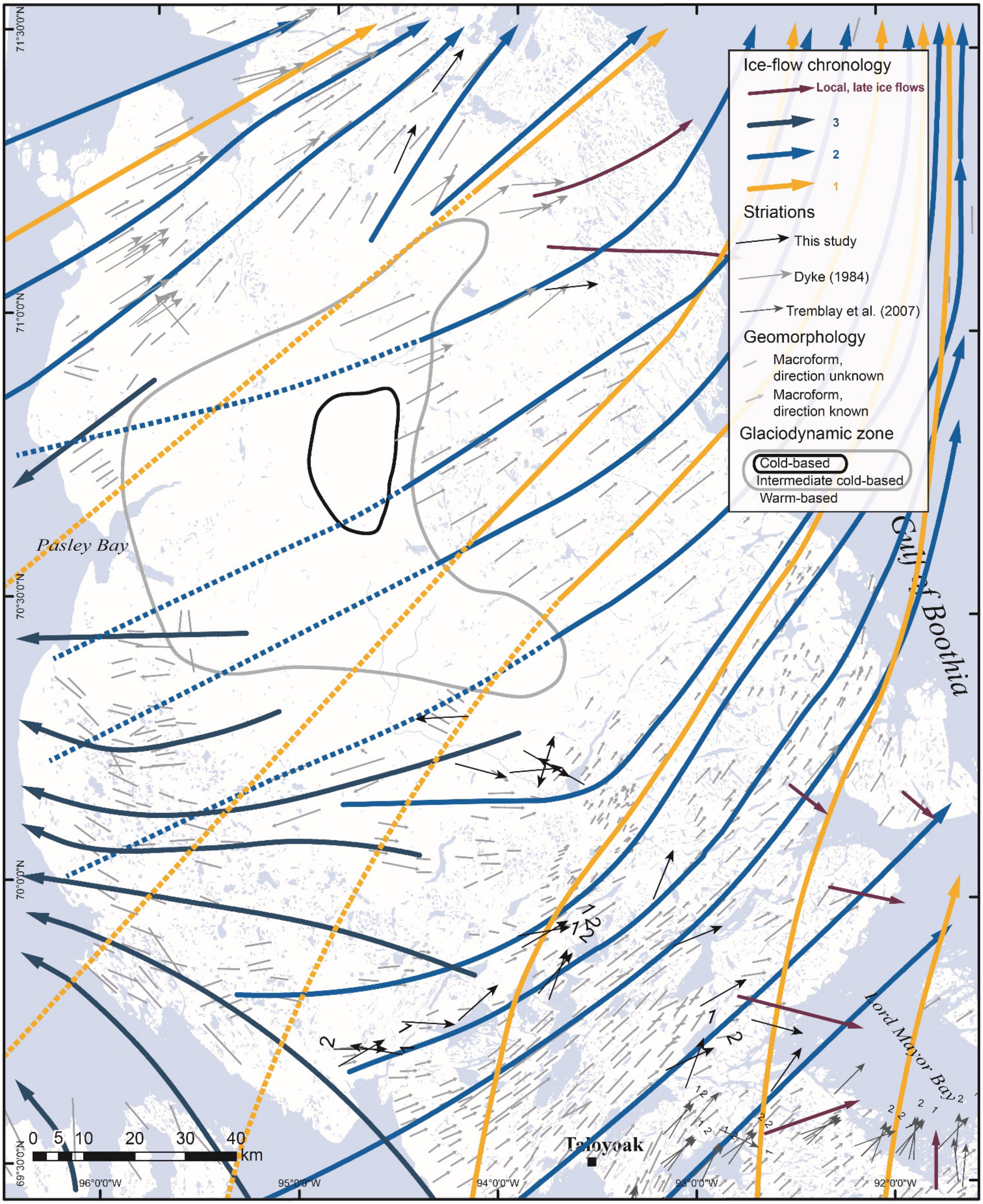


Figure 6. Striations (with age relationships), ice-flow lines, simplified geomorphological features (modified from Dyke,1984) and glaciodynamic zones of Boothia Peninsula, modified from Tremblay (2017). Dotted ice-flow line indicates weaker ice flow.

On the eastern portion of Prince of Wales Island, several eastward converging ice flow patterns are recognized (Fig. 2c). Among those, the Transition Bay Ice Stream (TBIS; Margold et al. 2015) comprises converging macroforms in carbonate till cross-cutting earlier northwest macroforms. The converging macroforms are constricted between patches of cold-based landscapes over the hilly terrain along the coast characterised by a lack of macroform, a scarcity of carbonate erratics (from data in Dyke et al. , 1992) and carbonate plumes (satellite images), dendritic river systems and a low lake density. The connection of these eastward ice flow patterns with eastward striations and macroforms in the Bellot Strait area of northern Boothia Peninsula and southern Somerset Island appears to be interrupted by northward-aligned macroforms in Peel Sound (Fig. 2a). If connected, the glacial transport of carbonate erratics over the Bellot Strait area implies that the onset zone of the Bellot Strait Ice Stream must have reached Paleozoic carbonate outcrops between Peel Sound and Larsen Sound, possibly during Phase 2 ice flow within a maximum BLIS ice catchment area (MID configuration A; Fig. 2c). Contrary to the geological map of Wheeler et al. (1996), bathymetric data indicate flat topography with subdued cuestas in Peel Sound (Fig. 3b), suggesting that sedimentary rocks may cover a large part of this marine through. Alternatively, basal entrainment from MID position B located in Peel Sound could also explain the glacial transport of carbonate till as part of the BSIS flow. Although the exact position of the MID remains unknown due to lack of field data (between hypothetical position A or B), the MID probably reached its most westward position on King William Island during this Phase (Dyke, 1984; Ozyer, 2011).

On Boothia Peninsula northwest of Boothia Isthmus, northwest striations cross-cut northeast striations (Tremblay, 2017). This late ice flow is associated with small northwest (west) macroforms on land (Dyke, 1984; Tremblay, 2017) and on the adjacent seafloor (Shaw et al., 2019). The overprint of these late NW-trending ice flow indicators on northeast striations suggests the beginning of the eastward migration of the MID over northern Boothia Peninsula at the end of Phase 2.

On Prince of Wales and King William islands, a complex system of north to northwest macroforms are recorded on land (Prest et al., 1968; Netterville et al., 1976; Dyke et al., 1992; De Angelis and Kleman, 2005; Margold et al., 2015) and on the seabed (new data; Shaw et al., 2019). Several macroforms are directly linked with ice streams, or onset zones of ice streams, flowing toward the northwest (De Angelis and Kleman, 2005).

Phase 2 corresponds to ice flows connected with an ice stream in M'Clintock Channel (Clark and Stokes, 2001; De Angelis, 2007b).

Phase 3

On the east coast of Boothia Peninsula, the latest striations indicate northeastward and eastward converging patterns toward Lord Mayor Bay (Tremblay et al., 2009; Tremblay, 2017) and Pelly Bay (Tremblay et al., 2007; Ozyer, 2011) which formed as part of late, relatively small ice streams during Phase 3 (Fig. 2d). The last remnant of the Boothia Isthmus Ice Stream was probably reduced to the smaller Lord Mayor Bay Ice Stream when the BLIS ice flow was waning following its former marine-based retreat to a terrestrial position. Local, late eastward ice flows toward Pelly Bay are also linked with minor re-entrainment of NE dispersal trains of Paleozoic carbonates and Paleoproterozoic marbles (Tremblay et al., 2009). East of Darby Lake, late, fine northward striations cross-cut northeast striations from Phase 2 (Tremblay et al., 2007). South of Pelly Bay, a significant portion of the north to northeast glacial transport of KIMs (Ozyer, 2011) are thought to have occurred during Phase 3.

Phase 3 occurred from 10.3 ka until deglaciation of the sector around 9 ka. Locally, Phase 3 is characterised with sudden ice deflections, influenced by topography on land and coastal bathymetric slopes, normal to the shoreline. A complex network of short eskers and moraines, and the occurrence of ice-dammed glacial lakes on the plateaus, indicate a general southwestward ice retreat on southern Boothia Peninsula (Dyke, 1984; Tremblay, 2017). At the latest stages of deglaciation, southwest and southeast retreating ice margins were towards an ice divide positioned in the middle of Boothia Peninsula and extending north of Somerset Island (Fig. 2d).

Northern Keewatin region

Phase 1

Over the Wager plateau, early northward striations are commonly found on protected lee surfaces of outcrops with prevalent northeastern striae (Little, 2001; McMartin et al., 2003; Ozyer and Hicock, 2006; McMartin et al., 2015). This early northward ice flow was correlated with Phase 1 south of Boothia Isthmus (see previous section), and attributed an LGM age. An earlier southwest flow was recognized to the south of the study area (Taylor, 1956; Boulton and Clark, 1990; McMartin and Hendersen, 2004; De Angelis and Kleman, 2005; McMartin and Dredge, 2005; Hodder et al., 2016) and attributed to the Early-Middle Wisconsinan, or a previous glaciation.

Phase 2

Between Wager plateau and Darby Lake, northeast striations and macroforms cross-cut the earlier northward striae found on protected surfaces mentioned above and are attributed to Phase 2. Between Chantrey Inlet and Darby Lake, northwest striations also crosscut the earlier northward indicators (Little, 2001; Ozyer and Hicock, 2006) and therefore also assigned to Phase 2, and are crosscut by later, northward striae associated with the Chantrey Moraine System (see Phase 3 below). Ozyer (2011) suggested that the western extent of this northeastern ice flow (and therefore equivalent to the MID position at the time) was located beyond Adelaide Peninsula (hypothetical position A – see Fig. 2c). Although northeastward-aligned macroforms are not present over the Adelaide Peninsula, the possibility that they have been eroded or obliterated by a later more northwesterly flow at the end of Phase 2 implies that position A is still plausible. This is supported by the presence of north-northeast trending MSGLs south of Adelaide Peninsula, superimposed by northward macroforms curving to the northwest over the peninsula (McMartin et al., 2021).

During Phase 2, similar to the region south of Boothia Isthmus (see above), ice was drained toward the regional topographic lows (i.e. Committee Bay). Figure 2c shows a subsidiary ice divide separating the northeast ice flow toward the BLIS main axis from northwest ice flows feeding the Boothia Isthmus subsidiary ice stream. In contrast, McMartin and Henderson (2004) attributed this subsidiary ice divide to a later ice flow phase, separating ice flow towards opening marine waters in Committee Bay from those in Chantrey Inlet, which would correspond to our Phase 3.

Phase 3

In northern Keewatin, Phase 3 late deglacial ice flows commenced when the ice margin was near the coastline or in terrestrial positions (starting around 10.3 ka), and lasted until complete deglaciation of the KID sector around 7 ka (Fig. 2d). Northward ice flows are linked with the occurrence of a group of glaciomarine deltas near the marine limit north of Darby Lake (Tremblay et al., 2007), and to ice retreat to the Chantrey Moraine System north of Wager plateau (Prest et al., 1968; McMartin and Henderson, 2004).

Melville Peninsula

Phases 1 and 2

In northern and western Melville Peninsula, west and northwest ice flow directions converge toward the main axis of the BLIS as indicated by the directions of striations and macroforms (Figs. 7 and 9). Glacial transport of Paleozoic carbonate debris (Dredge, 2001) and clasts of Proterozoic marbles and Archean supracrustal rocks (Tremblay and Paulen, 2012) is associated with this converging westerly ice flow.

In the southeastern part of Melville Peninsula, evidence of ice crossing over the interior plateau from Foxe Basin is ambiguous. Only rare, early striations toward the northwest and west are present near the coast (Dredge, 2001). Cold-based terrains cover a portion of the southern Melville Peninsula plateau (Fig. 7) so Dredge (2001) presumed that any ice crossing over southern Melville Peninsula plateau would have been largely cold-based. A few, even earlier, southwest striae and drumlins occur in the same area but these are associated with a radial flow pattern around Melville Peninsula during the Early Wisconsinan ice build-up phase (Dredge, 2002). Dredge (2001) suggested that an early, subsidiary ice divide from FD (Melville Ice Divide) ran across the southern part of the peninsula during LGM and early deglaciation (i.e., our Phase 1 and Phase 2), and possibly during earlier Wisconsinan glaciations.

On the other hand, Paleozoic carbonate boulders are found over Precambrian bedrock of the southern Melville Peninsula plateau (Dredge, 2002; Tremblay et al., 2016b) indicating that some westerly glacial transport from Foxe Basin occurred across this plateau. Because of the absence of any fine-grained carbonate matrix associated with these boulders, Dredge (2001) suggested that the few lone Paleozoic carbonate boulders might have been deposited during pre-Sangamonian glaciations, and that the calcareous matrix component had since weathered away. Around Barrow River however, glacial sediments contain Paleozoic carbonate pebbles and calcareous fine-grained material in two zones (Tremblay et al., 2016b). Paleozoic carbonate clasts also occur in till forming a narrow band along the southeastern coast of Melville Peninsula (Dredge, 2002; Tremblay et al., 2016b). On this basis, Tremblay et al. (2016a) proposed that Paleozoic carbonate tills observed in valleys around Barrow River were deposited from a nearby unmapped Paleozoic carbonate outlier, therefore not invoking ice flow out of Foxe Basin. Alternatively, the Barrow River carbonate tills could also represent remnants of a former carbonate till plume that crossed southern Melville Peninsula from Foxe Basin. Although the exact timing of this carbonate dispersal remains uncertain, it would represent the only area of southern Melville Peninsula with evidence of glacial transport from Foxe Basin during the Wisconsinian glaciations.

In summary, neither Dredge (2001) nor Tremblay et al. (2016a) showed unequivocal evidence of ice flow crossing from Foxe Basin over the southern Melville Peninsula plateau during Phase 1 and Phase 2. Based on this, we suggest that a portion of the ice divide (saddle) between the KD and FD (Dyke and Prest, 1987b) could have been located over southern Melville Peninsula during Phase 1 and that it migrated to the southeast of Melville Peninsula during Phase 2. On the basis of assumptions relative to the increasing intensity of BLIS during Phase 2, a minimum BLIS catchment area is sketched for Phase 1, and a maximum BLIS catchment area is drawn for Phase 2 (Figs. 2b and 2c).

Phase 3

On the Committee Bay coast of Melville Peninsula, late ice flows varying between NNW and SSW are contemporaneous with the establishment of the Melville Moraine (Dredge, 1995). On western Melville plateau, ice flow directions during this later phase are only slightly different from those of Phase 2, but

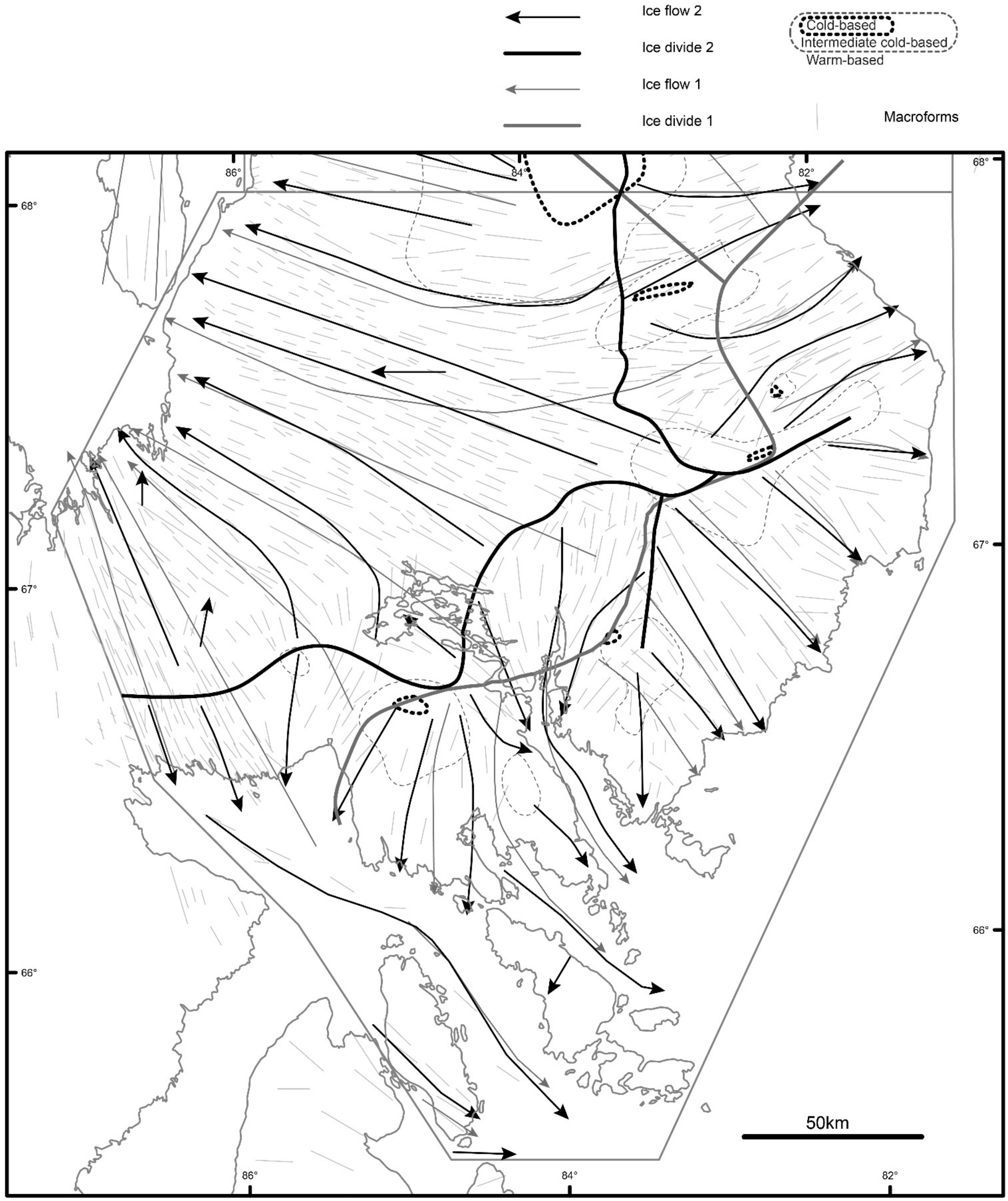


Figure 7. Ice flow lines in southern Melville Peninsula area, from Tremblay et al. (2016b), with glacial geomorphology modified from Dredge (1995, 2002).

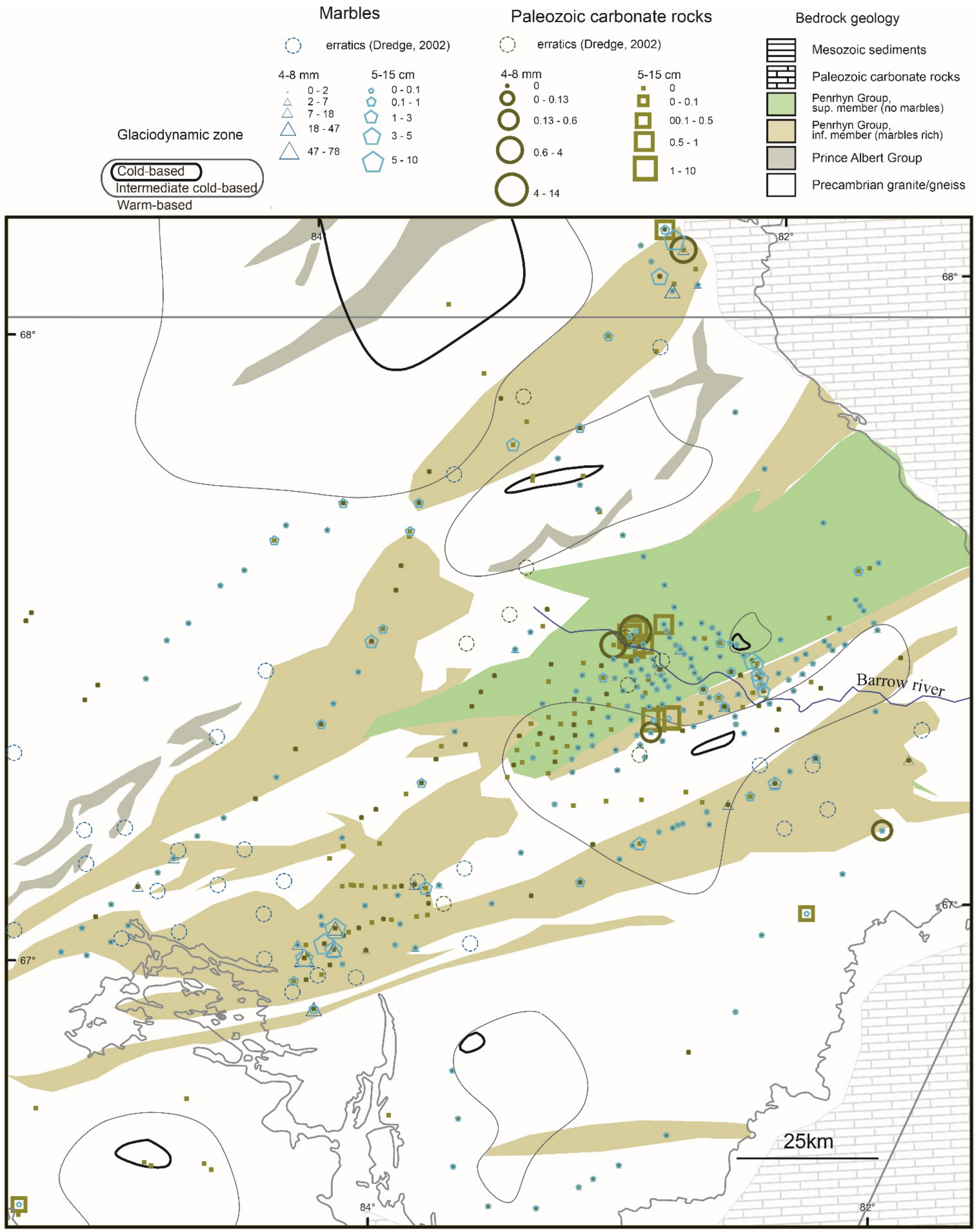


Figure 8. Petrographic counts in till pebbles and erratic boulders observations, for Paleozoic carbonate rocks and Penrhyn Group marbles (Tremblay et al., 2016b).

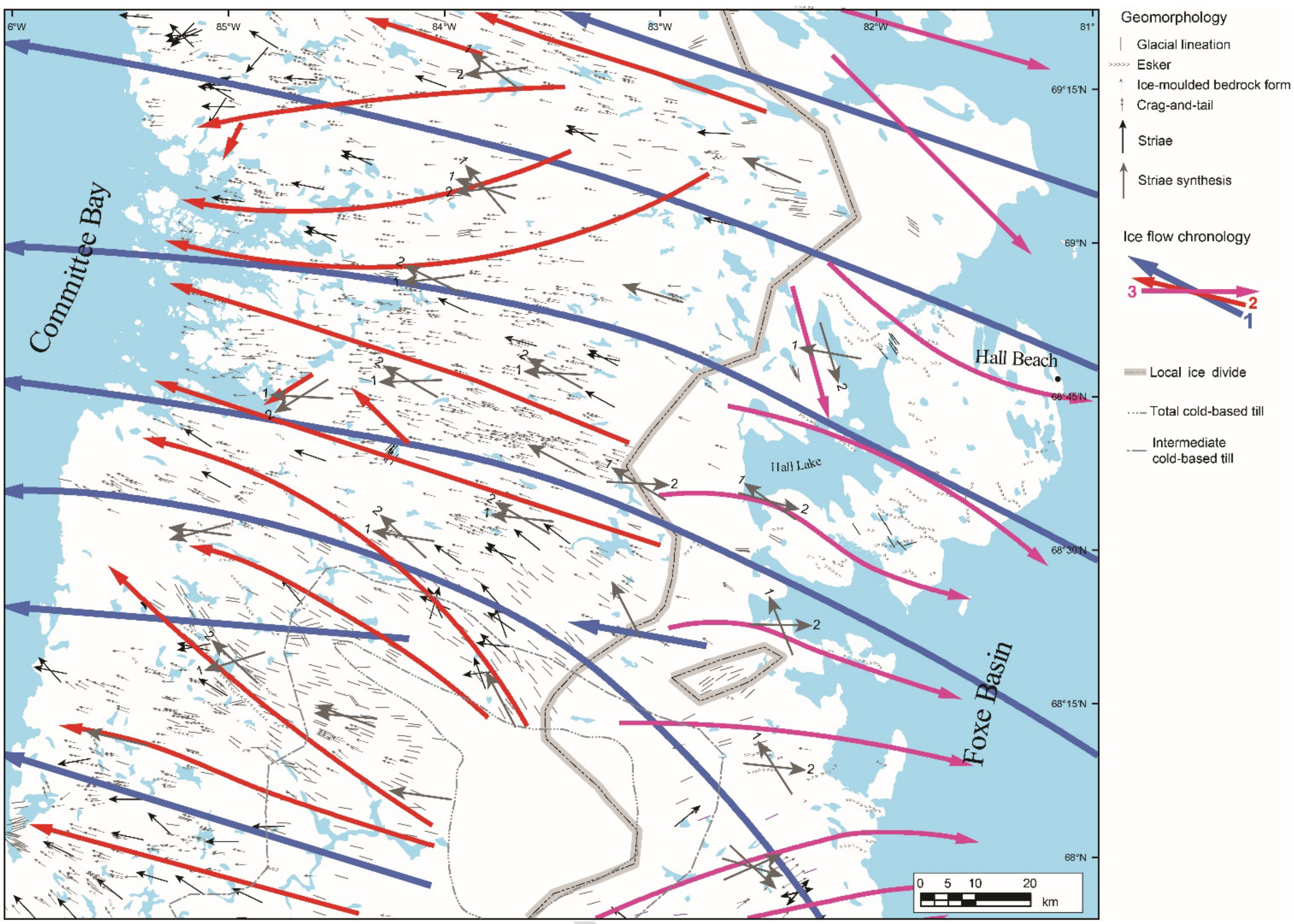


Figure 9. Ice flow lines in Northern Melville area (from Tremblay and Paulen (2012), with glacial geomorphology (including new satellite image interpretation, and elements from Dredge 1995, 2002; and De Angelis, 2007).

suggest an increasing influence by local topography. A convergent ice flow pattern, indicative of a waning Northern Melville Ice Stream, is observed in Garry Bay (Garry Bay Ice Stream). On the northernmost tip of Melville Peninsula, late south and southwestward striations and SW-oriented glacial transport of Mesoproterozoic quartz arenite are observed, heading toward an east-west oriented section of the Melville Moraine (Dredge, 2001).

Ice flows during Phase 3 started around 10.3 ka, when the ice margin in contact with marine waters retreated near the coastline, and continued to be active along a stillstand position during the deposition of the Melville Moraine, from 9.6 ka until about 7.5 ka (Dredge, 2001).

On Melville Peninsula, easterly indicators cross-cut westerly ice flow indicators locally, reflecting a northwestern migration of the ice divide to the centre of the Peninsula (Fig. 2d), from a position in Foxe Basin during the previous Phase 2. On the east coast of Melville Peninsula, ice flow towards the southeast, east and northeast is indicated by striations, glacial transport of Precambrian crystalline rocks and chromite grains over Paleozoic bedrock, and macroforms (Figs. 7 and 9; Tremblay and Paulen, 2012). Notably, several SE ice flow striations cross-cut the earlier WNW ice flow indicators (Phase 2; Fig. 2a). In central Melville Peninsula (North Arrow, 2018) and Rae Isthmus (Kupsch and Armstrong, 2013), KIM dispersal trains indicate late glacial transport toward Foxe Basin. Late ice flows in south-central Melville Peninsula are indicated by complex striations (W, S, E, NE) toward Repulse Bay and Foxe Basin that cross-cut northwest to NNW ice flow indicators converging into Committee Bay. A major, converging ice flow-reversal during late deglaciation toward Repulse Bay and Roes Welcome Sound in adjacent Keewatin documented by McMartin et al. (2015) can be associated with the latest part of Phase 3 when marine waters began to open in northern Hudson Bay. Late southeast, east and northeast ice flows toward ice streams draining Foxe Basin via Hudson Strait were major features on the east coast of Melville Peninsula. The ice divide separating easterly ice flows from northwesterly ice flows continued to migrate northwestward to its final position shown on Figure 2d.

On southeastern Melville Peninsula, the latest part of Phase 3 would have commenced when the ice flows toward BLIS were waning (before or at 10.3 ka), and the Hudson Strait catchment area began to capture some of the BLIS catchment area. Phase 3 lasted until ice retreated near the coastline slightly before 7.3 ka. Terrestrial ice retreat toward small residual ice caps occurred between 7.3 ka to about 6 ka (Dredge, 2001).

Northwestern Baffin Island

Phases 1 and 2

The glacial geology of northwestern Baffin Island is characterized by the presence of several converging features indicated by macroforms and striations and forming ice stream landsystems (Figs. 1, 2, 10 and 11): a) Bernier Bay (converging toward the Gulf of Boothia, (Hooper, 1996)), b) Fury and Hecla Strait (converging toward Committee Bay, Tremblay and Godbout, 2018), c) Whyte Inlet (converging toward Fury and Hecla Strait (Tremblay and Godbout, 2018), d) Admiralty Inlet (converging toward Lancaster Sound) and e) Moffet Inlet. Abundant elongated streamlined glacial landforms are present on the seafloor of the Fury and Hecla Strait (Tremblay and Godbout, 2018) and Bernier Bay (Fig. 2a). Numerous ice-moulded bedrock hills indicating a converging westerly flow were recorded on the shores of Fury and Hecla Strait (Fig. 10). Important glacial transport distances (up to 100 km) were reported within the Bernier Bay Ice Stream (Hooper, 1996), Admiralty Inlet Ice Stream (Dyke and Hooper, 2000; Tremblay, 2021), and Fury and Hecla Ice Stream (Hooper, 1996; Tremblay and Godbout, 2018; Tremblay et al., 2020b).

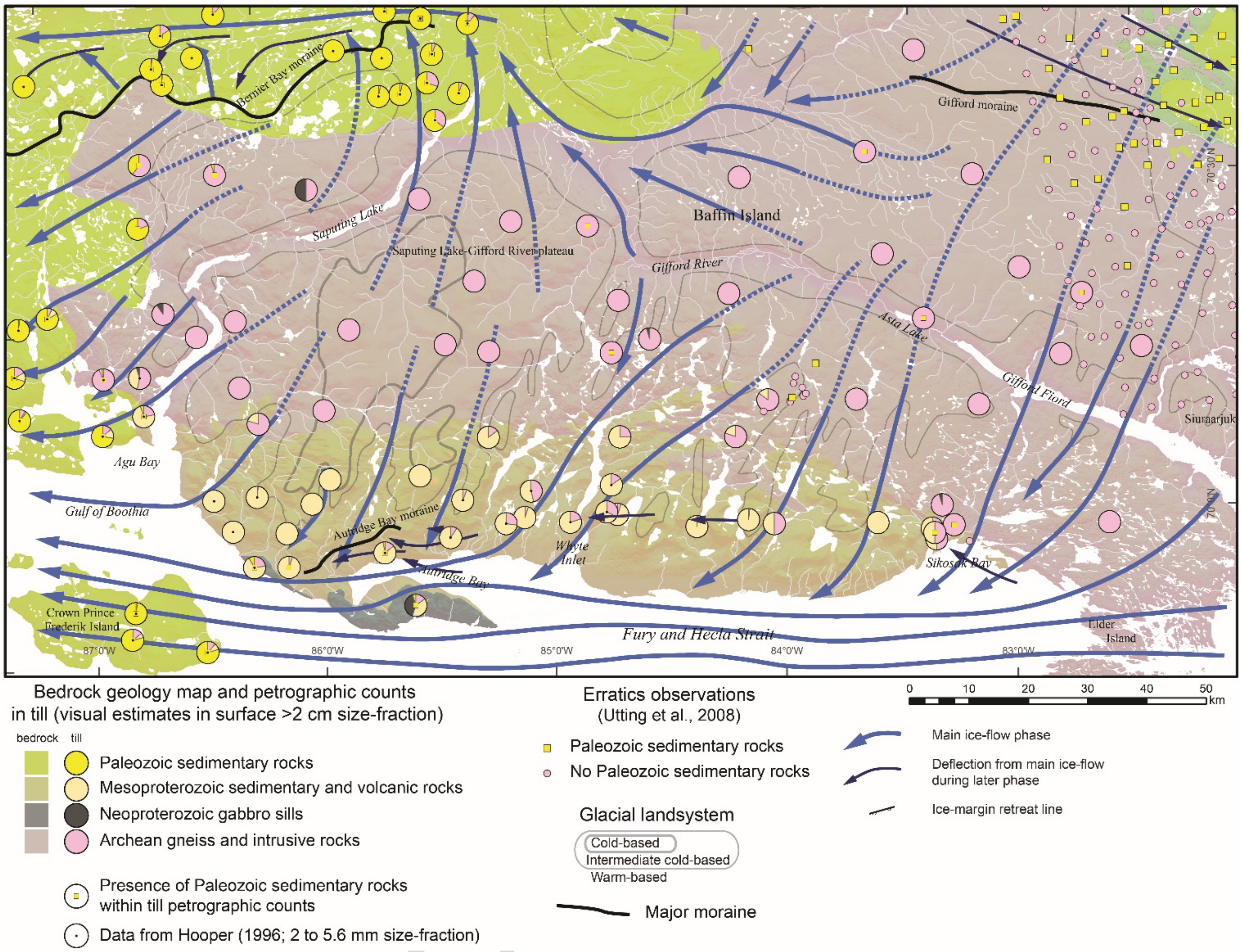


Figure 10. Glacial events north of Fury and Hecla Strait, from striations, glacial macroforms and petrographic counts in till (from Tremblay and Godbout, 2018). Surficial geology modified from Hooper (1996) and Dyke and Hooper (2001), DEM from Porter et al. (2018), and bedrock geology from de Kemp and Scott (1998).

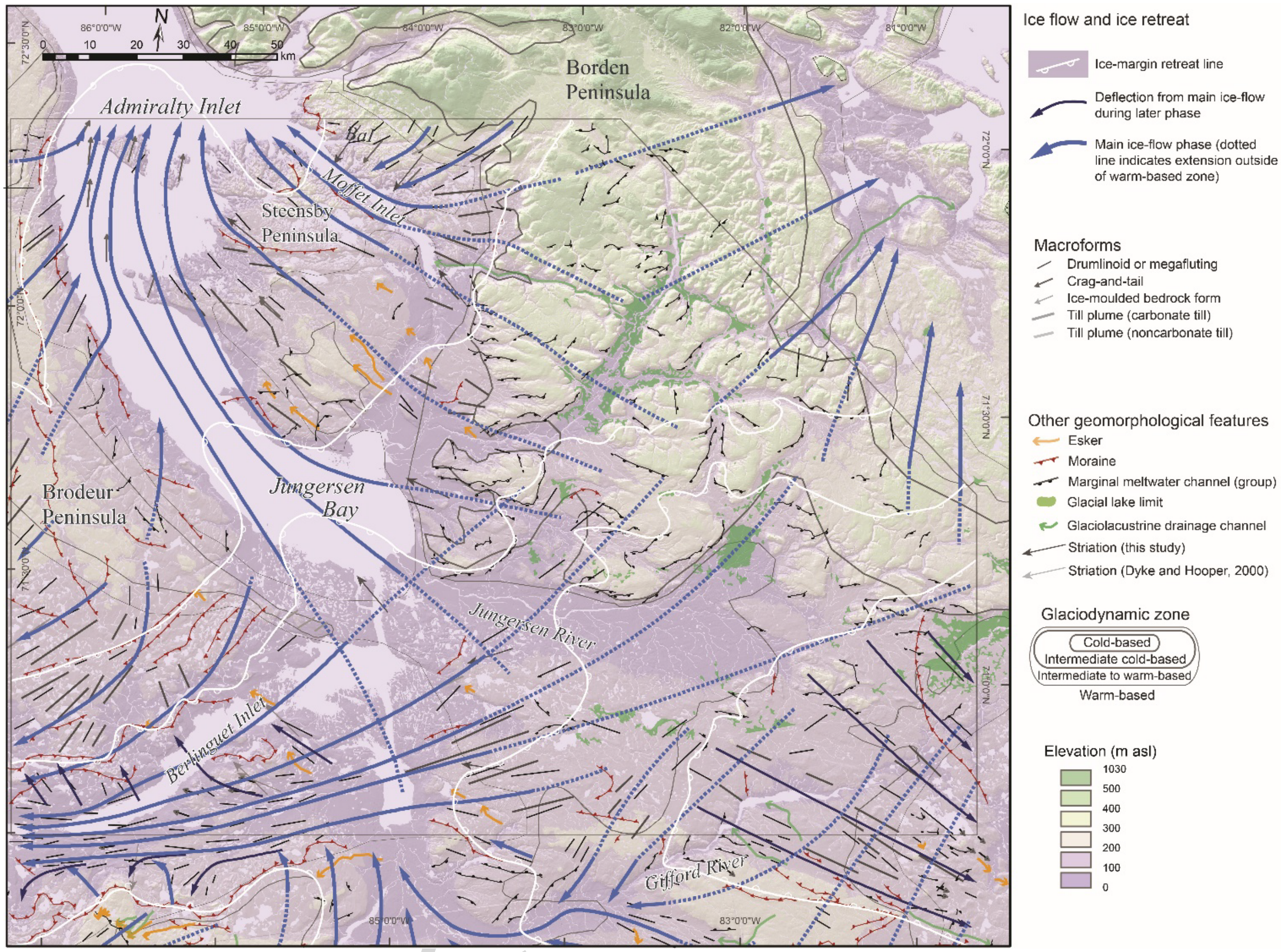


Figure 11. Ice-flow lines and ice-margin retreat lines with glacial geomorphology from the Jungersen River area (from Tremblay, 2021). Digital elevation model from Canadian digital elevation model (CDEM; Natural Resources Canada, 2016). Study area in gray outline.

The ice streams are separated by plateaus over which cold-based conditions persisted during at least Phases 1 and 2. The Saputing Lake-Gifford River plateau, mostly covered by cold-based terrains, separates the northern Melville Peninsula area from the rest of Baffin Island. It is strategically located to detect glacial transport from a radial flow out of northern Melville Peninsula versus a flow into the BLIS because cold-based zones can preserve glacial erratics for a long period of time due to slow erosion rates. Figure 10 (from Tremblay and Godbout, 2018) shows the absence of Paleozoic carbonates and Mesoproterozoic sedimentary rocks (chiefly indurated quartz arenite) in the clast fraction of glacial sediments over the Saputing Lake-Gifford River plateau, and their abundance south of the plateau. This supports the idea that the ice did not flow northward to cross the cold-based Saputing Lake-Gifford River plateau, and that the BLIS was active during the LGM and possibly during previous Wisconsinan glaciations (Dyke and Prest, 1987b; De Angelis, 2007a; Margold et al., 2018), in contrast with earlier reconstructions (Denton and Hughes, 1981; Mayewski et al., 1981). East of southern Admiralty Inlet, no Precambrian erratics were found over the core of the Paleozoic carbonate plateau (Tremblay, 2021), suggesting ice did not flow northeastward across Borden Peninsula.

Phase 1 occurred during late ice build-up to LGM (Fig. 2b), and Phase 2 from LGM to around 13 ka when the BLIS was at its most active phase of retreat (Fig. 2c), ending around 10.3 ka when marine ice margins retreated near the coastline (Fig. 2d).

Phase 3

When the ice front was near the coastline, the Bernier Bay Ice stream, Admiralty Inlet Ice Stream, Moffet Inlet Ice Stream and Fury and Hecla Strait Ice Stream were still active, and formed numerous frontal and lateral moraines (Fig. 2d). The Bernier Bay Moraine was deposited by lateral divergent ice flows, from the down-ice segment of warm-based ice within the Bernier Bay Ice Stream. Similarly, the Autridge Bay Moraine (Fig. 10) and the associated northernmost, NW-SE trending portion of the Melville Moraine were deposited by down-ice divergent warm-based flow within the Fury and Hecla Strait Ice Stream system (Fig. 2a). The Whyte Inlet Ice Stream was active as a tributary to the Fury and Hecla Strait Ice Stream, until the ice retreated around 9 ka.

On the Saputing-Gifford plateau, as ice receded over the land gradually from 10.3 ka to 6.8 ka, the orientation of the ice margin was often different from directions suggested during Phase 2, as indicated by eskers, moraines, marginal channels and proglacial lake configurations (Dyke and Hooper, 2001; Dyke, 2008; Tremblay and Godbout, 2018). The frequent absence of striations and macroforms associated with the pattern of ice retreat was probably associated with sluggish, cold-based ice flow, in some places accounting for the deposition of cold-based moraines (e.g., Gifford Moraine).

South of Pond Inlet

Phases 1 and 2

Between Steensby Inlet and Milne Inlet (site M, Fig. 1), early north-northwest striations (Januszczak, 2007) and weak, but noticeable glacial transport of Paleozoic carbonate rocks (Dyke, 2008) indicate early northward (NNE, NNW) ice flow toward Milne Inlet. East of Jens Munk Island (site JM, Fig. 1), the easternmost early striation (Dyke, 2008) and macroforms are oriented westward toward Fury and Hecla Strait. On the plateau east of Milne Inlet, early striations indicate NNW ice flow toward Paquet Bay (site P, Fig. 1; Little et al., 2004).

The ice flow indicators described above are interpreted to have occurred during Phase 1 and Phase 2. There is no indication that ice flow patterns would have varied much during the interval of LGM and earliest deglaciation. The north-northwest striations reported by Januszczak (2007) and the westward striations by Dyke (2008) (above) are interpreted as the inland limits of warm-based ice flow during Phase 2. Hypothetically during Phase 1, these sites could have been cold-based. Over the area around Steensby Inlet and east of Jens Munk Island, the conditions are inferred to be mostly cold-based during Phase 1 and Phase 2. Carbonate erratics found east of Steensby Inlet (Andrews and Sim, 1964; Andrews and Miller, 1979; Dyke, 2008) indicate glacial transport during an indeterminate period, possibly during LGM (Phase 1) or pre-LGM times.

Phase 1 occurred during LGM, and possibly prior to LGM during late ice build-up, and Phase 2 lasted from LGM to about 9 ka, when ice margins in contact with the sea retreated to near the coastline around Milne Inlet.

Phase 3a

The Phase 3a ice flow for most of the area south of Pond Inlet occurred in continuity with that from northwestern Baffin Island region. Late ice flows toward a terrestrial margin are indicated by moraines, ice marginal channels, eskers and proglacial lake configurations (Dyke, 2000a; b). Late ice flow events towards the north are also indicated by submarine moraines in the Tasiujaq (Eclipse Sound) and Milne Inlet areas (Brouard and Lajeunesse, 2017), concordant with moraines related to northward flow on land (Klassen, 1993; Dyke, 2000a; b).

The Phase 3a ice flow pattern indicates a progressive, mostly cold-based retreat, as the ice margin abandoned the last marine inlets and became fully terrestrial around 6 ka over the Steensby Inlet area, and earlier on the plateau around Gifford River (continuing terrestrial margin ice retreat from northwestern Baffin Island area at 6.8 ka). Eventually, early Barnes Ice Cap features on central Baffin Island were forming on the plateau around 6.8 ka. On the eastern highlands of Baffin Island, the ice retreated gradually toward the modern position of Barnes Ice Cap and some of the other plateau ice caps (Dyke and Hooper, 2001; Gilbert et al., 2017), while retreat toward modern mountain glaciers might have occurred earlier in the Holocene (Margreth et al., 2014; Pendleton, 2019).

Phase 3b

Over the area around Steensby Inlet, later striations, macroforms and glacial transport features indicate southeastward convergent-divergent ice flow toward Steensby Inlet and Jens Munk Island (Steensby Inlet Ice Stream; Dyke, 2008). This ice flow was probably active during only a few centuries around 7 ka (Dyke, 2008). Near the ice divide between Milne Inlet and Steensby Inlet, southward striations cross-cut earlier Phase 2 north-northwestward striations (Januszczak, 2007). On Jens Munk Island, late southwestward striations cross-cut earlier striations trending westward toward Fury and Hecla Strait (Phase 2; Dyke and Savelle, 2006; Dyke, 2008). Southward dispersal trains of Paleozoic sediments (carbonate and sandstone; Dyke and Savelle, 2006; Dyke, 2008; Utting et al., 2008) and KIMs (Januszczak, 2007) are observed in this area and distances of glacial transport increase from the ice divide toward the coast, reflecting the increased velocity of ice flow towards the terminus of the ice stream.

Phase 3b corresponds to a brief but powerful ice flow episode, commencing when the Steensby Inlet Ice stream flowed south toward a glacial margin near the coast, south of Jens Munk Island, until ice finally retreated north from Steensby Inlet (Dyke, 2008).

Northernmost Baffin Island

Phases 1 and 2

On northern Brodeur and Borden peninsulas, ice flow indicators radiate outward from mostly cold-based terrains. Over these terrains, ice flowed from local ice divides, independent of the rest of the LIS, and merged with ice streams in the deep marine channels (BLIS main axis, Admiralty Inlet, Navy Board Inlet, Tasiujaq and Milne Inlet). Bathymetric data indicate macroforms along those marine channels as evidence for strong ice stream flow (Bennett et al., 2016; Brouard and Lajeunesse, 2019). At LGM, the ice flow toward Milne Inlet was channeled through Tasiujaq and Pond Inlet, toward Baffin Bay, joining with the BLIS in Baffin Bay. On Bylot Island and the area around Pond Inlet, the Eclipse Till and Eclipse Moraine formed at the margin of warm-based ice from regional ice flow depositing Paleozoic and Proterozoic sedimentary erratics. There are currently two opposing ideas for the age of the Eclipse Moraine (and Till). The first is a pre-LGM age on the basis of stratigraphy, TCN exposure dating of boulders and stratigraphic correlations with radiocarbon and amino-acid –dated units (Klassen, 1993; McCuaig, 1994); the landform was preserved under cold-based ice from LGM to deglaciation. The second is a LGM age (Dyke and Hooper, 2001; Dyke et al., 2002) based on the Holocene age of the Navy Board Inlet Moraine and the correlation of these moraines with the largest ice extent during the entire Pleistocene (Dyke et al., 2002; Brouard and Lajeunesse, 2017).

In central Navy Board Inlet, a minor ice divide separated ice flowing toward the BLIS from glacial flow toward Tasiujaq (formely known as Eclipse Sound). This ice divide position is interpreted from a lack of glacial macroforms on the seafloor (Fig. 3a) and the presence of the perfectly preserved pre-LGM Canada Point Paleo-delta on the shore of Navy Board Inlet (Klassen, 1993; Tremblay and Lamothe, 2019). After ice retreated from the adjacent Lancaster Sound, the divergent, down-ice portion of northern Navy Board Inlet Ice Stream deposited moraines on both sides of the channel (Dyke and Hooper, 2001; Bennett et al., 2016). Phase 1 was active from late ice build-up to LGM, and Phase 2 progressively ended at about 10.9 ka, when the majority of the coastline was deglaciated.

Phase 3

As the ice gradually retreated from inland, the orientation of moraines, marginal channels and proglacial lakes configurations indicate that the ice front was retreating in a concentric pattern, with cold-based ice flowing quasi-radially, and twists and deviations according to local topographic peculiarities (Dyke and Hooper, 2001).

During this phase, cold-based ice, and locally polythermal valley glaciers, retreated toward modern glaciers and ice caps on Brodeur Peninsula, Borden Peninsula, Bylot Island and near the southern coast of Pond Inlet.

Discussion

Relative ice flow chronology data and possible reconstructions

While the synthesis of regional ice flow histories for the BLIS catchment area produced a reconstruction of three main ice flow phases, the fragmented and poorly time-constrained nature of the data implies that differing interpretations can be compatible with the various ice flow datasets. By nature, the ice flow record is more spatially continuous for macroforms and discontinuous for striations, and for both some of the record may have been removed due to erosion by subsequent ice flow events or covered by later sedimentary sequences. Therefore, the older the striae set (or other paleoflow indicator), the more the interpretation can vary, in part because of the impossibility to date glacial striations with an absolute method. For example, as an outcome of this drawback, it could be argued that the older, northward striations south of Boothia Isthmus, part of Phase 1 and tentatively assigned to the LGM, could alternatively be attributed to pre-LGM glaciation(s). In this scenario, Phase 2 would have started at LGM instead of during deglaciation. Throughout this synthesis, although we have tried to present the most coherent interpretation using all available data, alternative interpretations are impossible to dismiss entirely. Ultimately, because multiple solutions are likely to result from multi-scale, truncated datasets, there is a need for a data compilation platform that will allow evaluation of datasets, comparison of different scenarios, and the assessment of the magnitude of the uncertainty for each proposed reconstruction. Such a structured data platform in which multiple-scale glacial datasets could be consolidated for analysis was recently developed for a large region west of Hudson Bay in Nunavut that includes the southwestern part of the BLIS catchment area (Behnia, 2020; McMartin et al., 2021). Expansion of this glacial data platform to include the entire BLIS area datasets holds the capacity for multiple interpretations and could be useful in the production of different working hypothesis in glacial modelling and for future field work.

New implications

New evidence for both stability and fluctuations in the intensity and patterns of ice flow are presented and evaluated in the BLIS catchment area. Building on glacial reconstructions of LGM to deglaciation, such as Dyke and Prest (1987), Dyke and Hooper (2001), De Angelis (2007a) and Margold et al. (2018), the synthesis of field data and new mapping on land and offshore show that convergent high-velocity ice flows through the BLIS main axis was a major, persistent feature in the NELIS throughout the Late Wisconsinan glaciation. As the ice flow velocities changed (i.e. generally increased) from LGM to deglaciation, the patterns of ice flow were transformed, reflecting the channelling of ice toward varying regional topographic lows connected to the deep depression occupied by the BLIS. New tributary ice streams are recognized (WIIS, FHIS, FBIS, GBIS, LMIS, MOIS, JBIS) and different names for previously established ice streams (BLIS, BSIS, BIIS) are proposed to better reflect their location and extent from LGM until deglaciation.

The data and interpretations presented in this Chapter have implications for glaciodynamic constraints on ice flows, as expressed by the updated mapping of low-velocity zones in cold-based landscapes and high-velocity zones in ice streams. Based on a number of recent field studies on the high plateaus of southeastern Baffin Island (Johnson et al., 2013; Tremblay et al., 2016a), Melville Peninsula (Tremblay and Paulen, 2012) and Wager Bay (McMartin et al., 2021), the transition from warm-based to cold-based glacial landscapes generally appears to be gradational over a zone several tens of kilometers wide. The incremental impact of spatially variable subglacial thermal conditions over time is probably the cause of these subtle transition zones.

Numerous striation and till clast data have improved our understanding of successive ice flows in the BLIS catchment area, particularly on the mainland western portion of this area. Notably, variable ice flow patterns and palimpsest glacial transport dispersal trains were identified south of Boothia Isthmus to the KID during LGM (Phase 1) until early deglaciation (Phase 2). Phase 1 switched to Phase 2 in this area because ice flow accelerated within the BLIS main axis, enlarging the footprint of the Boothia Isthmus Ice Stream, and depressing the ice profile, thus speeding up ice flow within the region. As a result, the catchment area of the BLIS may have been larger during Phase 2 relative to Phase 1 (for example on southern Melville Peninsula). North of Boothia Isthmus, the presence of carbonate erratics in the core zone of the cold-based plateau confirmed that a weak, but discernible glacial transport occurred under cold-based ice flow conditions. On Melville Peninsula, a precise mapping of a late deglaciation ice flow reversal during Phase 3 was determined based on new striae, macroforms and pebble data. On the southern part of this peninsula, new data on the distribution of Paleozoic carbonate clasts in till indicated a possible local source for those clasts on the southeastern shores, with implications that ice did not cross over from Foxe Basin and further west into Committee Bay.

To the east beyond Nunavut's mainland, our understanding of ice flow patterns has also improved. In the region straddling Baffin Island and Melville Peninsula, we showed evidence that the Autridge Bay Moraine and the northern Melville Moraine were part of an end moraine formed during retreat of the ice stream system in the Fury and Hecla Strait (FHIS), as opposed to the limit of ice flow descending south from the hills located north of Fury and Hecla Strait (Dyke and Prest, 1987b; Dredge, 2001). In the central part of Navy Board Inlet, new bathymetric datasets showed the absence of macroforms on till surface, as well as the presence of a pristine pre-LGM delta close to the shoreline, indicating lasting flow separation and cold-based ice conditions under a subsidiary ice divide.

Warm-based interior of the NELIS?

The data synthesized here suggest that ice streams in the NELIS were active during the LGM until deglaciation. This work also presents new evidence for the absence of ice flowing northward over the cold-based Saputing-Gifford plateau on Baffin Island during the LGM and possibly during pre-LGM glaciations. This evidence provides a clear support for glacial reconstructions (Dyke and Prest, 1987b; De Angelis, 2007a; Margold et al., 2018) in which westward ice streaming in the Fury and Hecla Strait and northern Melville Peninsula, feeding an important inner core of the BLIS catchment area, was prevalent during LGM and until deglaciation of marine-based ice margins (10.3 ka). In the southwestern part of the BLIS catchment area south of our study area, glacial stratigraphy, till provenance and mapping of glacial landsystems suggest that ice streams might have reached the interior of the KD during the marine-based ice margin deglaciation phase of the BLIS (Hodder et al., 2016; McMartin et al., 2021). These assertions deal with the important paleoglaciological question of whether warm-based ice flow extended into the interior of the NELIS during the LGM and early deglaciation, or whether cold-based conditions prevaled. In the work presented here, although cold-based landscapes were recognized over high plateaus and under former ice divides, most evidence points toward a well-developed ice stream network being active during the last glaciation. This warm-based interior of the NELIS upholds reconstructions presented by Dyke and Prest (1987), Stokes (2012), De Angelis (2007a) and Margold et al. (2018), but with a slightly more extended footprint of auxiliary ice stream activity within the BLIS catchment area during LGM. The contrasting view by Kleman and Glasser (2007) that warm-based ice flows could be highly diachronic, occurring only on the outer rims of ice sheets, and that the northern LIS was mostly cold-based during the LGM, is not supported by the evidence presented here.

Future work

This Chapter is a first step in the synthesis of ice flow indicators around Foxe Basin and Baffin Island in a large part of the BLIS catchment area, focusing especially on striations, macroforms and glacial transport of Paleozoic carbonate erratics. The reconstruction of the three ice-flow phases presented here is certainly limited and could be expanded into more time slices and with more details using appropriate and more complete datasets. For example in the area around Steensby Inlet, the full picture of the successive ice flows was not captured in this generalized reconstruction. Other regions also deserve closer scrutiny, including in Pelly Bay, where new bathymetric, till composition and striae datasets suggest a much more complex series of events that those depicted here. Similarly, although new evidence on the floor of Peel Sound is constraining the provenance of carbonate clasts in the Bellot Strait area to a closer source, the complexity of cold-based and warm-based ice flows around Prince of Wales Island, Peel Sound and Bellot Strait needs to be re-examined in detail.

In many other areas, additional field datasets would help increase the resolution of the glacial reconstructions. Late during deglaciation, local topography played an increasingly important role in the dynamics of ice draining into the marine troughs in and around Chantrey Inlet, Rasmussen Basin, Adelaide Peninsula and King William Island. Relatively pronounced changes in ice flow directions occurred within short distances in these regions, compared to surrounding flatter terrains on land. New fieldwork around this region where the ice flow record is scarce (Prest et al., 1968) could provide key elements for the ice stream chronology in the western BLIS catchment sector. In addition to increasing the ice flow record, the regional stratigraphic context of the entire BLIS catchment area needs to be updated, notably with new geochronological measurements (TCN, IRSL) in the northeastern Baffin Island and Bylot Island sectors (Klassen, 1993; Tremblay et al., 2020a). Particularly, the age of the Eclipse Moraine is a major issue and should be studied in more in detail.

Finally, numerical ice-sheet modeling experiments could attempt to reproduce the ice flow phases proposed here. For example, the models could try to account for the ice flow pattern variability in the area between Boothia Isthmus and the KID, as well as between Boothia Peninsula and the MID, and for the absence of N or NW ice flows over the Saputing-Gifford plateau. In Stokes et al. (2012), the modeled ice flow from 65 ka to 25 ka generally shows a northwest ice flow direction over the Saputing-Gifford plateau and Fury and Hecla Strait, which contradicts the petrographic study of till clasts in this area (Tremblay and Godbout, 2018). In order to solve this disagreement, the model could extend areas of enhanced basal sliding to where ice stream footprints were mapped, including over northern Melville Peninsula (a mostly hard-bedded ice stream in this case) and Fury and Hecla Strait.

Conclusion

Regional ice flow patterns and chronology were reconstructed in the BLIS catchment area by synthesizing results and interpretations from previous work and from new striation measurements, glacial transport studies, and macroform mapping using ArcticDEM, satellite images and multibeam bathymetric data. Three main ice flow phases were reconstructed: Phase 1) active ice flow within the BLIS, at LGM; Phase 2) increasingly active and far-reaching ice flow within the BLIS, between LGM and 10.3 ka; and Phase 3) local ice flows in various tributary ice stream catchment areas, toward retreating terrestrial or coastal ice margins after 10.3 ka. Generally, the synthesized field data are supporting the presence of ice streams and variable ice flow patterns, which are generally in accord with most recent reconstructions primarily based on remote sensing (De Angelis, 2007a; Margold et al., 2018). However we suggest that the footprints of auxiliary ice streams within the BLIS catchment area were slightly more extended during deglaciation (Phase 2) than during LGM (Phase 1). Importantly, new bathymetric observations on the seafloor of major depressions bring fresh elements to the ice flow history. In the BLIS main axis, seafloor geomorphological interpretations indicate strong continuity of macroforms from Prince Regent Inlet to Lancaster Sound; the continuity is much less apparent from Barrow Strait to Lancaster Sound, probably because the ice flow between Barrow Strait and Lancaster Sound was volumetrically less important than from Prince Regent Inlet to Lancaster Sound. Between Committee Bay and Prince Regent Inlet, the seafloor data indicates the continuity of macroforms, interpreted as marking the footprint of the BLIS. Also, new sea bottom morphology evidence indicates that a cold-based ice divide was situated within Navy Board Inlet between Bylot and Baffin islands. The mapping of cold-based glacial landscapes is also a key element in the proposed reconstruction of glacial dynamics and ice flow events. In many regions, this compilation can be used as a tool to study changes in glacial transport and glacial patterns during the last glaciation.

An integrated GIS-based data platform is necessary to combine and make accessible field data and remote sensing data, used to test and strengthen competing glacial flow reconstructions, especially for the early pre-LGM phases where evidence of older ice flows is equivocal. Glacial modeling studies can use the glacial history interpretations presented here as a driver for refining the constraints on ice flow chronology and patterns in different geographic settings and at different times. In return, the numerical models can provide feedback on the processes at work in the BLIS catchment area and in the NELIS in general.

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