Paleoglaciological dynamics in northern Manitoba and the subglacial bed mosaic

by

Michelle Trommelen

A thesis

presented to the University of Waterloo

in fulfillment of the

thesis requirement for the degree of

Doctor of Philosophy

in

Earth Sciences

Waterloo, Ontario, Canada, 2013

AUTHORS DECLARATION

I hereby declare that I am the sole author of this thesis. This is a true copy of the thesis, including any required final revisions, as accepted by my examiners. Included in this thesis are published papers (Chapters 2 and 3), which are co-authored with other researchers.

I understand that my thesis may be made electronically available to the public.

ABSTRACT

During the last glacial maximum (LGM), some 20 ka ago, northern Manitoba was situated beneath 3 to 4.5 km of ice, on the outer fringe of a major ice spreading center of the Laurentide Ice Sheet. The region has also been affected by major paleoglaciological changes linked to multiple source areas, migration of ice centres, and ice-sheet thickening/thinning over multiple glacial cycles. The net effect of this evolution is a very complex geological record, which has major implications for ice-sheet reconstructions and drift prospecting. Theory-based hypothesis for the region suggest initial advance-phase deposition was followed by either net-erosive or cold-based conditions for much of the glacial cycle. In contrast, observation-based reconstructions of ice-sheet behaviour consider the glacial landscape to have been predominately formed by near-complete overprinting during warm-based deglaciation. Some complexity has been recognized in sediment-landform records, but new insights into glacial dynamics and sediment-landscape evolution are needed.

Systematic mapping (remote-sensing) and fieldwork (ice-flow indicators, till composition, ground truthing) in northeastern Manitoba has led to the recognition of spatio-temporal variability in landscape (streamlined-landform event-flowsets) and landform (micro and meso-scale ice-flow indicator records) and till composition inheritance. In particular, analysis of the spatio-temporal characteristics of the subglacial landscape led to the recognition of disjoint zones with internally-consistent assembly histories – termed *glacial terrain zones (GTZ)*. These GTZ were then classified as (1) relict-glacial, (2) palimpsest, or (3) deglacial in nature. Generally, (1) is interpreted as pre-LGM, (2) may include pre-LGM terrain but also LGM to early deglaciation (ice margin still far from study area; ice sheet thinning phase) and (3) was formed during the final ice retreat phase. The resultant surface till composition within relict and palimpsest GTZs is a spatial mosaic interpreted to reflect variable intensities in modification (overprinting) and preservation (inheritance) of a predominately pre-deglacial till sheet. In these regions, streamlined landforms parallel to a known deglacial ice-flow orientation were unable to overprint the underlying inherited glacial sediment composition. Secondly, field investigations (sedimentology, clast fabrics, till composition, near-surface S-wave seismic surveys) have characterized the widespread Rogen moraine terrain. These transverse subglacial ridges are spatially associated with streamlined landforms, are situated on regionally low-lying terrain without topographic constraints and may have small bedrock 'knob' obstacles at their up-ice base. This thesis assesses Rogen moraine formation hypotheses within the

new paleoglaciologic context of northern Manitoba, favours an instability mechanism for formation, and provides important field data against which further formation hypothesis should be tested.

The main insight of this study is not a detailed reconstruction (local history), but rather a series of forms of evidence suggesting that the glacial history of the region is one of prevailing patchy low-erosion conditions which favored preservation of a fragmentary record of non-coeval and sometimes contrasting warm-based (more dynamic) conditions. Despite being near a thick inner-core region of the Laurentide Ice Sheet, where basal conditions are generally considered stable and meltwater availability is low, the hard-bed study area was subject to local spatio-temporal shifts in subglacial conditions that led to generation of a complex palimpsest glacial landscape. Spatial differences in the preservation of older streamlined landforms, variably drumlinized Rogen moraine and the concentrations of inherited subglacial detritus all culminate in a hypothesis that suggests the subglacial landscape was continually evolving and subject to spatio-temporal variations in the intensity of ice-bed processes throughout the last glaciation (subglacial bed mosaic). Based on the new glacial history, and a general lack of ice-marginal landsystems, most warm-based ice-flow phases likely occurred near LGM – with only weak overprinting during late deglaciation.

The idea of landform generation at patches within a transient subglacial bed mosaic now allows for a close association between subglacial drumlins and Rogen moraine ridges, that may have formed by disconnected and not necessarily coeval or related processes. This mosaic, of slow to non-flowing basal ice ('sticky regions') and wet-based flowing-ice patches, also helps to explain preservation of immature landforms (Rogen moraine) and relict or palimpsest terrain. Sticky regions may have formed by at least two different mechanisms: localized heterogeneous switches in basal thermal regime (frozen-bed patches), or within a warm-based subglacial environment from wet to stiff, dewatered till.

ACKNOWLEDGEMENTS

I thank my supervisor Dr. Martin Ross for his guidance in the field, the office and in the academic world; for continued discussions of glacial theory, and for showing how a little Cinderella action can cause bedrock outcrops to tell their stories. I thank Janet Campbell for managing the paperwork side of the project, for numerous debates about ice dynamics and for keeping my mind open during discussions.

I thank Dr. Ahmed Ismail for the gracious interpretation of the collected seismic data. Dr. Isabelle McMartin, Dr. Lynda Dredge and Dr. Harvey Thorleifson are thanked for providing full access to their immense body of research in northern Manitoba and surrounding areas.

I thank the Manitoba Geological Survey and the Geological Survey of Canada, especially Dr. Christian Böhm and Ric Syme, for including me within the Manitoba Far North Geomapping Initiative and for letting me continue this thesis work as 'real work'. Their moral and financial support was invaluable towards the production of this thesis, as was the Natural Resources Canada Research Assistance Program funding. I also received funding from the Northern Studies Training Program to conduct fieldwork in Churchill.

The following paleoglaciological papers provided me with the inspiration to apply these concepts and methods to northern Manitoba: Clark et al. (2000, QSR vol.19, p.1343-1366), Stea and Finck (2001, Geol. Soc of London special pub. 185, p.237-265) and Clarhäll and Jansson (2003, JQS vol.18, p.441-452).

Sincere thanks are also due to the numerous internal and external editors and referees who have helped improve the clarity of the journal publications within, including Mike Smith, Paul Dunlop, Alain Plouffe, Isabelle McMartin, Chris Clark, Neil Glasser, Olav Lian, Arjen Stroeven, Jan Piotrowski and others who have chosen to remain anonymous.

Table of Contents

List of Figures

List of Tables

Chapter 1 Introduction

The impetus for this work was access to the remote region of northeast Manitoba, provided by the Manitoba Far North Geomapping initiative. This thesis uses detailed geological fieldwork and remote-sensed imagery analysis of the glacial landscape of northern Manitoba to investigate paleoglaciological dynamics. During the last glacial maximum, between 26.5 ka and 19 ka (Clark et al. 2009), this area was located beneath 1.5 to 3.0 km of ice (Tarasov and Peltier, 2004) on the outer fringe of a major ice spreading-center of the Laurentide Ice Sheet (LIS). The region has also been affected by major paleoglaciological changes linked to multiple source areas and ice sheet thinning/shrinking until final deglaciation. The three papers herein present different evidence that has led to the development of a subglacial bed mosaic model of landscape evolution for the inner regions of ice sheets. The model provides a paleoglaciologic explanation for the fragmentary palimpsest nature of the subglacial landscape, as well as for the spatial distribution of non-coeval subglacial landform-fields. The research also brings new insight into the long-standing issue of the origin of Rogen moraine.

1.1 NORTHERN MANITOBA

1.1.1 Project area

This thesis focuses on northern Manitoba (58°-60° latitude; 'project area'), with fieldwork in northeastern Manitoba between 58°45′ and 59°45′ latitude and between 95°30′ and 97° longitude (Figure 1-1). Additional work was completed at Churchill, Manitoba. The study area, as referred to in the following chapters, is different for each chapter and delimited as the data demanded. The extent of fieldwork is largely dictated by logistical constraints.

1.1.2 Physiography

Northern Manitoba is bounded by Hudson Bay in the east, and extends west to the Saskatchewan border, rising in elevation to about 500 m a.s.l. (Figure 1-1). In the east, the landscape is gently undulating and consists of subdued (0.5-4 m) subglacial landforms, till plains, bedrock outcrops, and abundant small lakes. In the west, the landscape is one of undulating to hummocky relief consisting of higher-relief (4-50 m) subglacial landforms. Long (100 km), large (30-40 m wide) eskers are present throughout the area, at 8-18 km intervals (Dredge et al., 2007). The eastern and southernmost parts of the area, which were covered by the Tyrrell Sea and Glacial Lake Agassiz, respectively, form extensive relatively flat areas

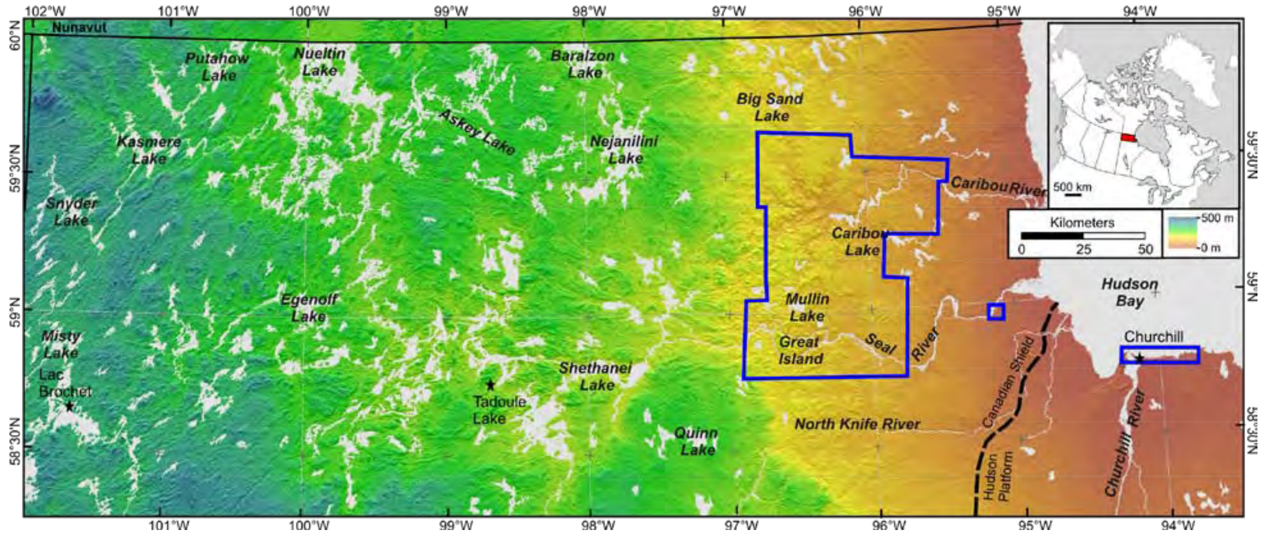


Figure 1-1. Fieldwork areas (blue boxes) in Northern Manitoba (58°-60° latitude, 'project area'). The background image was generated using the radar-derived digital elevation data from the Shuttle Radar Topographic Mission (SRTM) data set (United States Geological Survey, 2002). A hillshade model has been added to enhance the relief.

draped by organic deposits (peat bogs and fens) that often mask glacial landforms (Dredge et al., 1986; Dredge and Nixon, 1992). The marine limit in the study area is around 180 m asl and the lake limits are above 270 m asl (Dredge, 1983; Dredge and Nixon, 1992). Most of the study area is underlain by granitoid and granodiorite rocks of the Archean-Paleoproterozoic Hearne domain, Churchill Province, which is a part of the Canadian Shield (Figure 1-2; Manitoba Geological Survey 2006). Paleoproterozoic

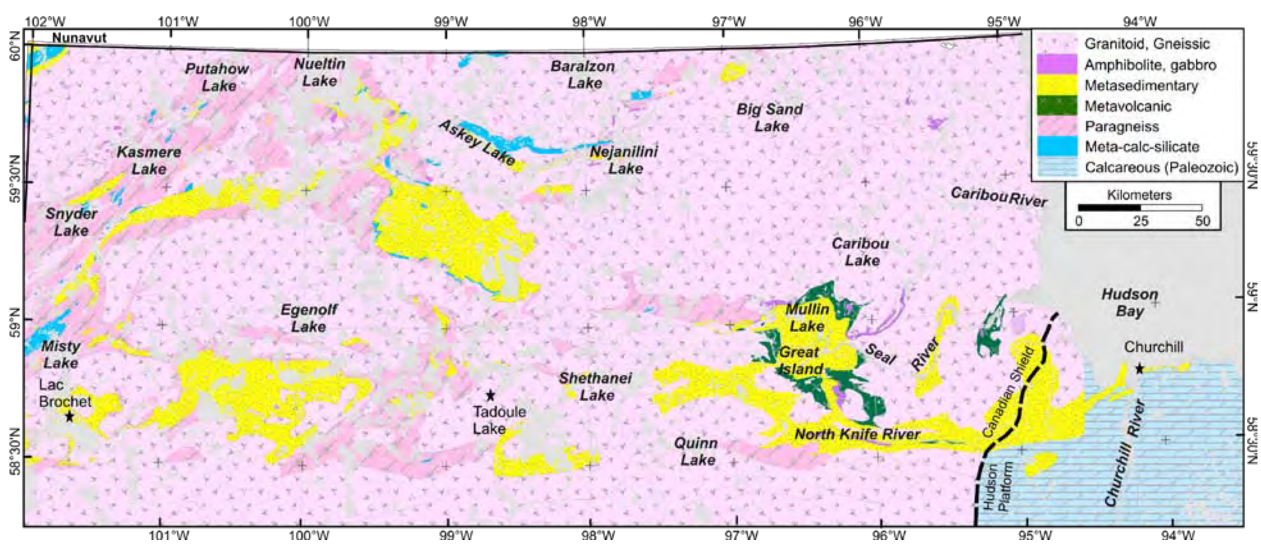


Figure 1-2. Simplified 1:250 000 scale bedrock geology in northern Manitoba classified by major lithology; modified from unpublished compilation mapping.

metasedimentary and metavolcanic rocks also occur throughout the area. In the southeast, the Precambrian basement is overlain by the Paleozoic Hudson carbonate platform (Figure 1-2).

1.2 THEORY-BASED CONDITIONS

There have been numerous theoretical-based models that reconstruct the subglacial conditions of the LIS, usually at ice-sheet wide or grid-cell scales. These schematic continental to regional scale studies make inferences about basal thermal regimes and the prevalence of erosion or deposition over the entire glacial cycle(s). For example, Sugden (1977, 1978) mapped landscapes of 'areal scour' and erosional lake basin density, and correlated this to the configuration of the LIS at its maximum limits (Figure 1-3). According to this reconstruction, the study area should reflect a net-erosional environment that was warm-based at the maximum extent of the LIS. In the late 1990's, Boulton (1996) expanded on this theory of inner ice-sheet erosion (Figure 1-4) by considering patterns of erosion and deposition across an hypothetical ice sheet over time – linked to concepts of ice-flow velocity (zero beneath an ice divide and at the margin and increasing in areas of erosion). He outlined a theory of net deposition during advance ('advance phase till'), net erosion during most of the glacial cycle, and net deposition during retreat ('retreat phase till').

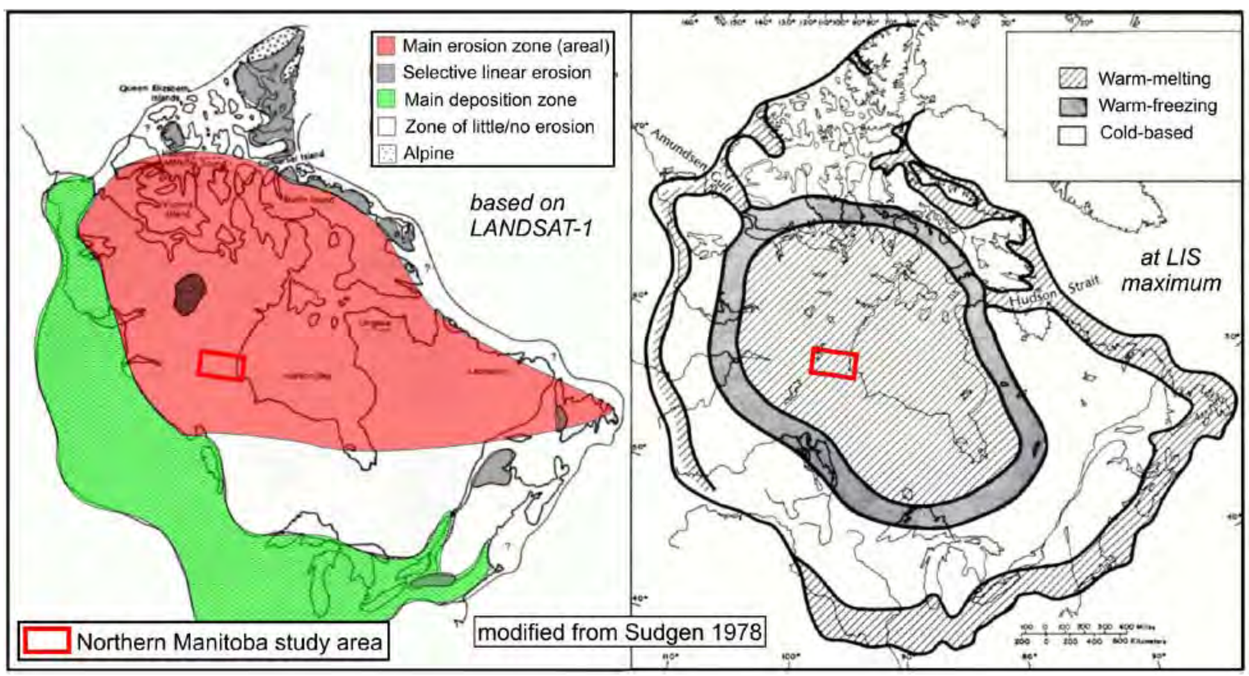


Figure 1-3. Erosion (left) and basal thermal regime (right) reconstructions of the LIS, modified from Sugden (1978). Northern Manitoba is outlined by the red box.

Within the inner core regions of ice sheets, erosion was presumed to be dominant. Of course, this hypothetical ice-sheet does not capture ongoing processes at the base of real ice sheets, such as the influence of meltwater availability (Boulton and Hindmarsh, 1987; Eyles, 2006), sediment supply (Murray, 1997; Stokes and Clark, 2003b), basal thermal regime, sliding vs. deformation, etc. In the early 2000's, after newer imagery identified cross-cutting streamlined landforms and their probable pre-deglacial origin (c.f. Boulton and Clark, 1990; Knight and McCabe, 1997), Boulton refined his theory by suggesting that an inner zone of freezing (expanding and migrating outward during ice sheet growth) would progressively suppress erosion and deposition – leading to preservation/inheritance of older surfaces within the inner core regions of ice sheets (Boulton et al., 2001b). Migration of ice centers would

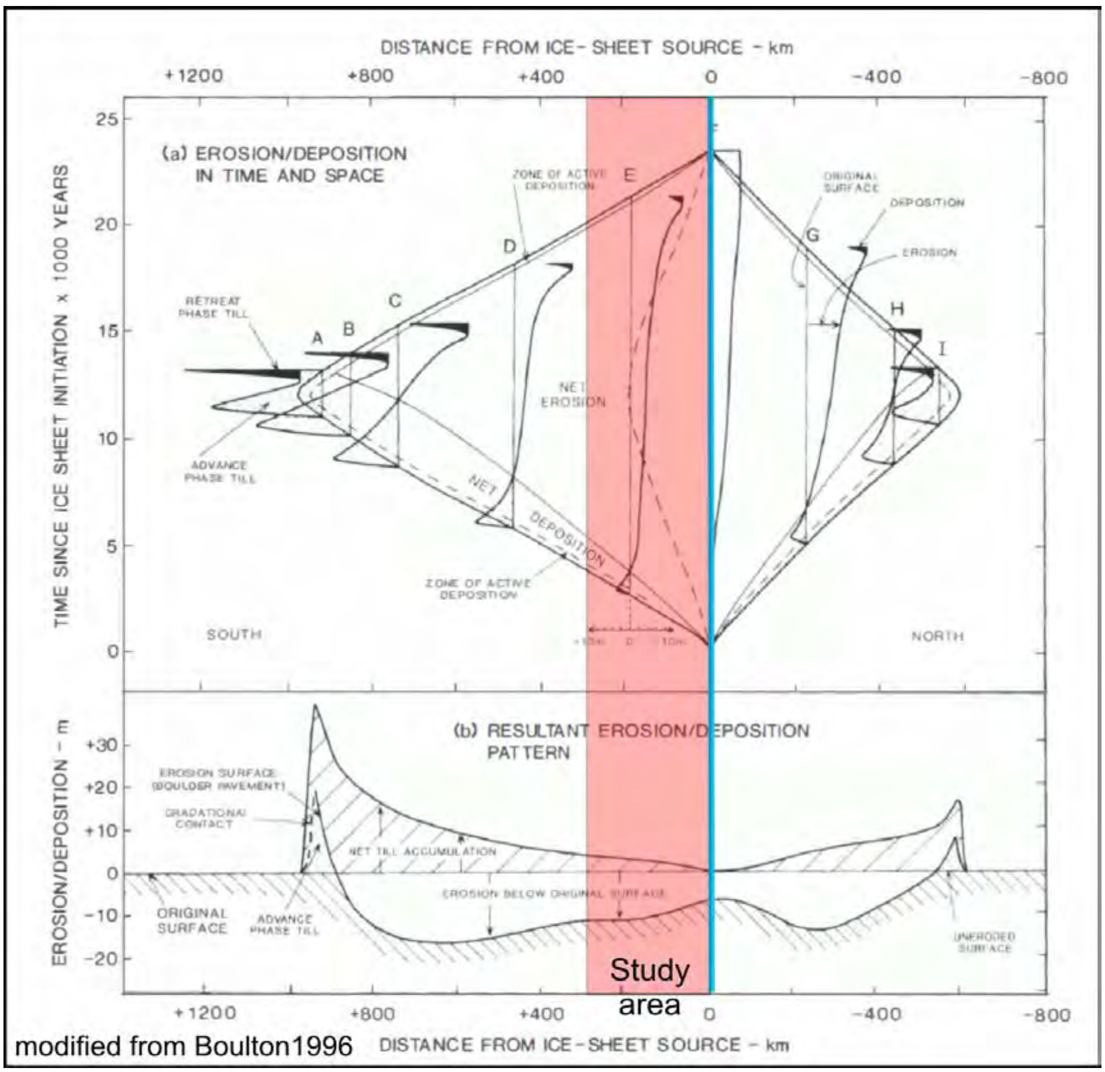


Figure 1-4. Erosion verses deposition over time, with increasing distance from an ice divide; modified from Boulton (1996). The distance between the Late Wisconsinan Keewatin Ice Divde and the northeastern Manitoba study area is outlined by the red box.

have a similar effect (McMartin and Henderson, 2004). Regardless, another large-scale erosion study by Hildes et al. (2004) also suggested that the study area has undergone minimal erosion, and was beneath thick, slow-flowing ice at glacial maximum. In contrast to Sudgen (1977), but perhaps accounting for theoretical low-erosion and high preservation subglacial conditions, Tarasov and Peltier (2007) and Kleman and Glasser (2007) suggested northern Manitoba was cold-based at the Last Glacial Maximum (LGM). Tarasov and Peltier (2007) used their glacial systems model to map the probability of permafrost extent beneath the LIS at LGM (correlative to cold-based ice, Figure 1-5). To test this model prediction, the

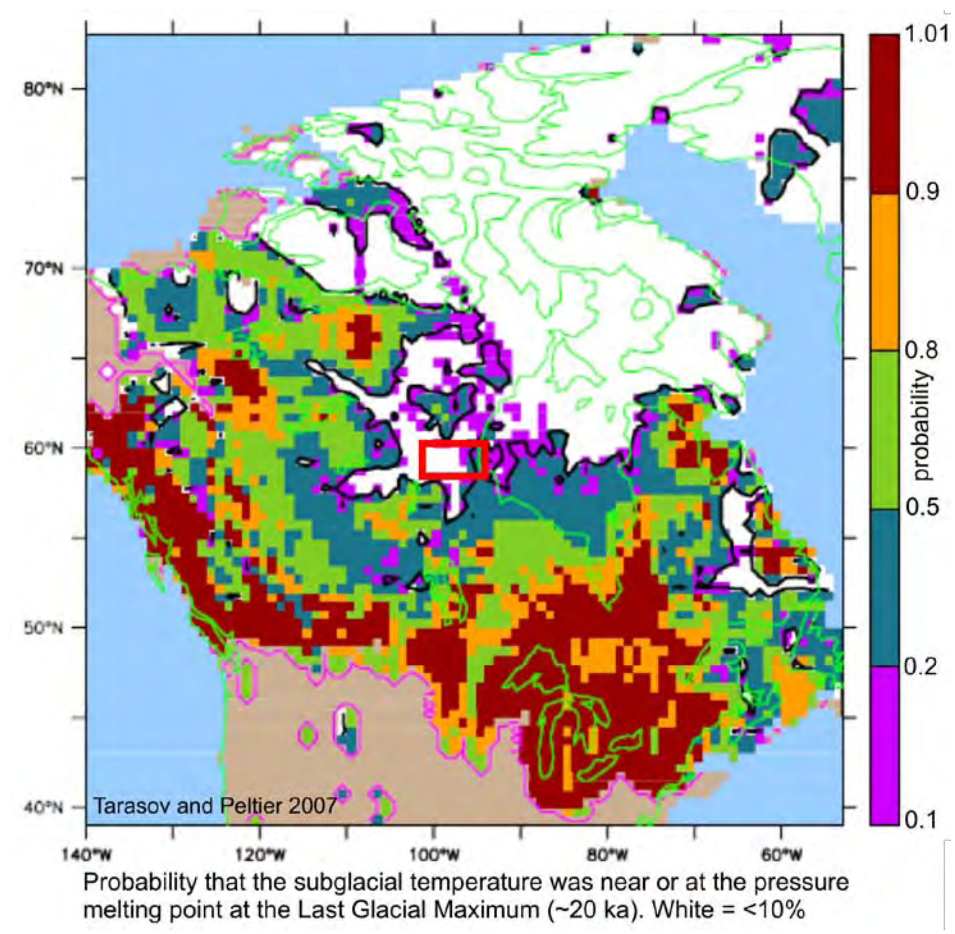


Figure 1-5. Probability that the subglacial landscape of the LIS was at the pressure melting point at the Last Glacial Maximum, modified from Tarasov and Peltier (2007). White areas, which include most of northern Manitoba, have a less than 10% probability and were interpreted as cold-based areas where permafrost was present. Northern Manitoba is outlined by the red box.

authors noted that the occurrence of cold-based ice spatially matches the distribution of ribbed moraine and relict landscapes in Keewatin, Nunavut, but under-estimates the extent of cold-based ice where these landforms/landscapes are present beneath the paleo Quebec/Labrador ice sector. The assumption that ribbed moraine are indicative of cold-based ice was initially proposed by Kleman and Hattestrand (1999) and Hattestrand and Kleman (1999), based on their fracturing theory of how ribbed moraines form. Indeed, the spatial patterning of ribbed moraine and relict landscapes is what Kleman and Glasser (2007) used to interpret widespread cold-based ice at LGM (Figure 1-6). Problematically, the formation and paleoglaciologic implication of Rogen moraine is a question posed throughout the last fifty years that still remains largely unsolved (Lundqvist, 1969; Hättestrand, 1997; Moller, 2006). The current lack of model for the formation for these conspicuous landforms will greatly affect/inform new paleoglaciologic models.

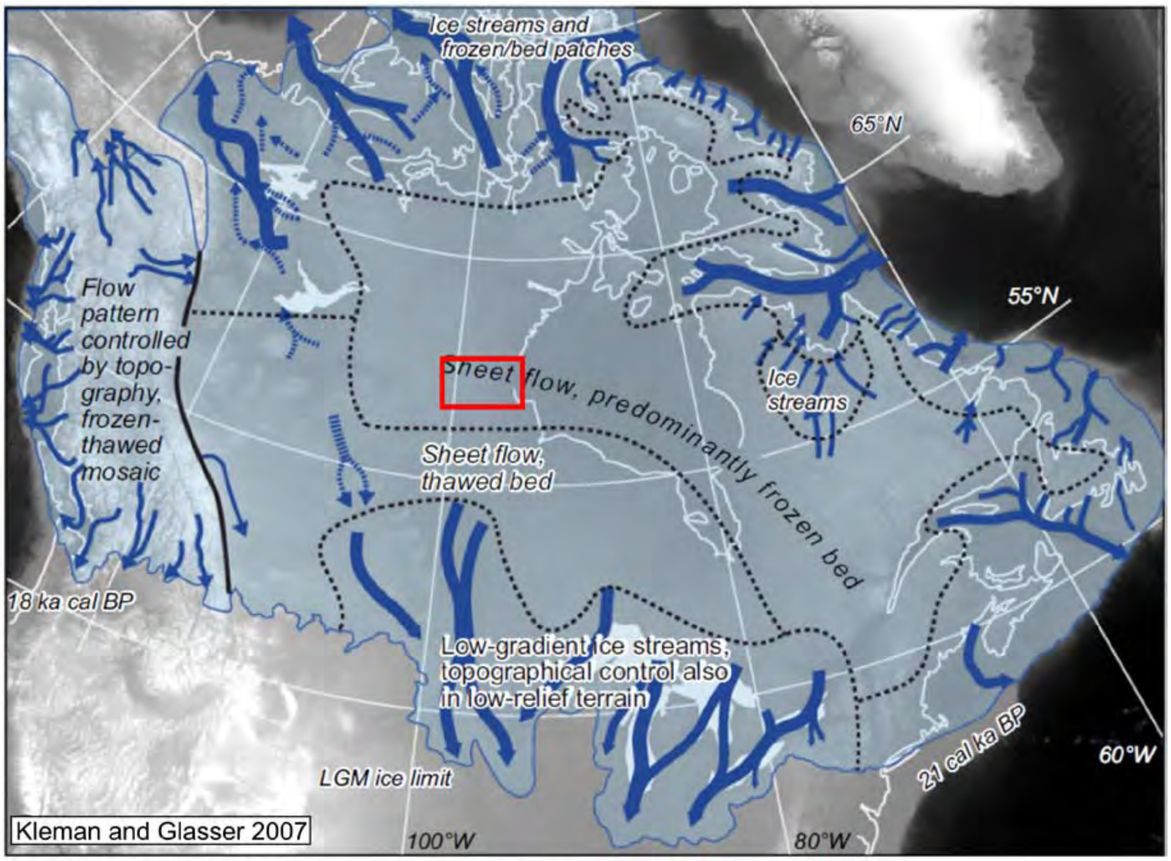


Figure 1-6. Thermal organization of the LIS at LGM, modified from Kleman and Glasser (2007). Northern Manitoba is outlined by the red box. This map was generated using the spatial patterns of mapped paleo-ice streams and presumed frozen-bed patches (ribbed moraine and relict landscape). The inclusion of ribbed moraine as a 'frozen-bed' indicator is still debated (see Chapter 4).

Key predictions, or theory-based hypotheses, for the near-inner core region of the LIS in northern Manitoba suggest that:

1. The glacial landscape should reflect initial advance-phase deposition followed by net-erosion conditions – unless there is a change in basal thermal regime.
2. For most of the glacial cycle, and possibly even at LGM, the study area was presumably cold-based. Little landscape modification processes are predicted during that time interval and the subglacial landscape should reflect this through preservation of older advance-phase landscapes (landform and till composition inheritance). Deposition and streamlined landform formation may have occurred near the ice margin during retreat, dependent on available meltwater.
3. According to the most widely-cited theory, ribbed moraines should form during deglaciation, as the cold-based ice-sheet core transitioned to warm-based near the ice margin and fractured (brittle deformation) the pre-existing soft bed.

1.3 Observation-based conditions

The entirety of northern Manitoba has been the subject of significant, high quality work – generating a large field-based dataset that can be used to test paleoglaciological theories. While this general approach can be applied across glaciated terrains, the glacial geological record of inner core regions of past ice sheets is arguably less well understood than ice marginal settings. The outer zone of major ice spreading centres experienced broad shifts of ice-flow dynamics and meltwater conditions, leading to patterns of subglacial erosion and deposition between ice advance and retreat phases.

1.3.1 Previous work

The region was mapped at 1:250 000 scale (Dredge and Nixon, 1981b, a, c, 1982a, b; Dredge et al., 1982a, b; Nixon et al., 1982; Richardson et al., 1982), and compiled at a 1:500 000 scale (Figure 1-7) (Dredge et al., 1985; Dredge and Nixon, 1986; Dredge et al., 2007). Till was sampled and analyzed for matrix geochemistry and carbonate content at 540 sites (Dredge and Pehrsson, 2006). Following this fieldwork, two significant memoirs were published that thoroughly describe the surficial geology, glacial history, and stratigraphy (Figure 1-8) of the region (Dredge et al., 1986; Dredge and Nixon, 1992). In 2007 158 till samples for parts of the north (NTS 64N, J, O and 54E, F, L) were re-analyzed for matrix geochemistry, utilizing updated technology (Dredge and McMartin, 2007). The region to the southwest was also investigated at 1:125 000 scale (DiLabio and Kaszycki; DiLabio et al., 1986; Kaszycki, 1989; Kaszycki and Way Nee, 1989a, b, d, c, 1990e, d, c, a, b; Lenton and Kaszycki, 2005; Kaszycki et al., 2008), while the area to the south was mapped at a 1:250 000 reconnaissance level (Klassen and Netterville, 1980; Klassen and Netterville, 1985; Klassen, 1986).

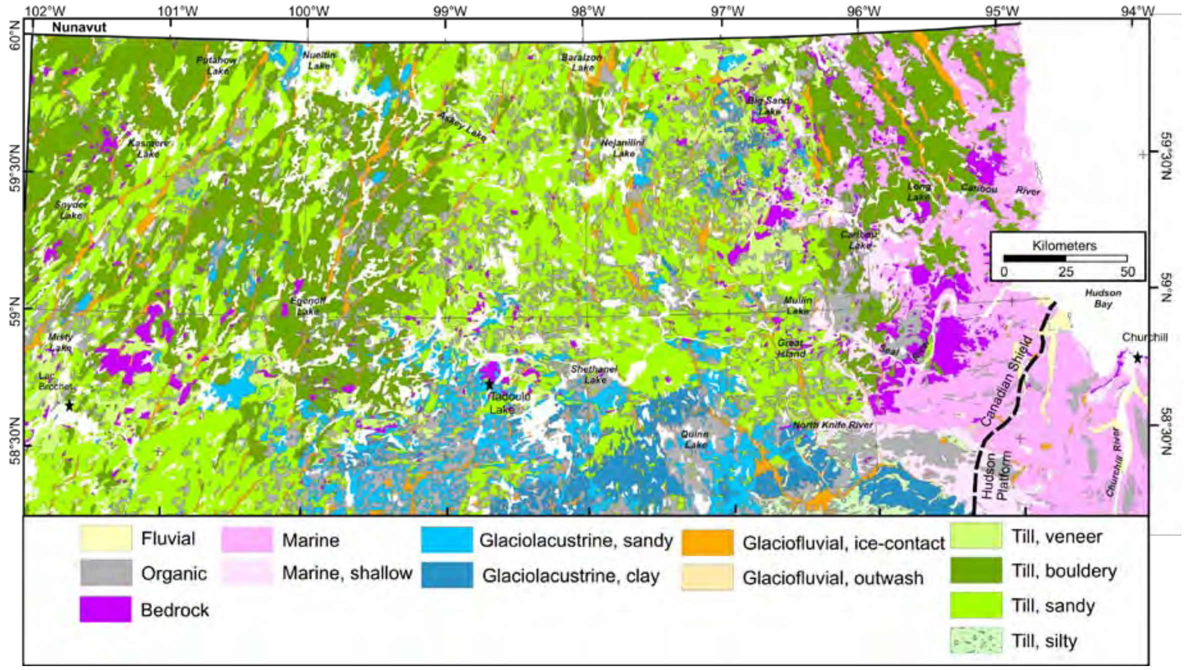


Figure 1-7. Previous surficial map compilation (1:500 000 scale) for northernmost Manitoba, modified after Dredge et al. (2007). The study area is predominately draped by till, which has been mapped according to texture. Bouldery and sandy till are thought to be derived from the Canadian Shield rocks, while a significant component of the silty till is thought to have been derived from carbonate rocks of the Hudson Platform.

On the Canadian Shield of northeastern Manitoba, the landscape is one of rock knobs and thin discontinuous till, with regions of thicker (~1-14 m) drumlinized till plains and fields of Rogen moraines (Aylsworth and Shilts, 1989; Dredge and Nixon, 1992; Dredge et al., 2007). Contrastingly, thick (5 to 35 m) glacial and interglacial sediment packages, deposited within river valleys of the Hudson Bay Lowlands (Pre-Illinoian through to postglacial sediments, Klassen, 1986; Dredge and Nixon, 1992; Nielsen, 2001, 2002a; Dredge and McMartin, 2011), record a longer depositional glacial history. The relationship between this widespread subglacial landscape mosaic and the regional stratigraphy is unknown. Furthermore, fields of Rogen moraine in northern Manitoba and adjacent Nunavut and Saskatchewan, Canada (Prest et al., 1968; Aylsworth and Shilts, 1989; De Angelis, 2007; Kleman et al., 2010), are among the largest in the world, but have rarely been studied (except within the Dubawnt Lake Ice Stream, Stokes et al., 2008). Late Wisconsinan deglaciation has also overprinted this landscape.

1.3.1.1 Stratigraphy

Stratigraphic sections in northeast Manitoba (Figure 1-8), southeast of the map area, have been separated into a Pre-Illinoian Keewatin-derived Sundance Till capped by an interglacial paleosol, overlain by the

Illinoian carbonate-bearing Amery Till, interglacial Sangamonian Nelson River sediments, and Wisconsinan Keewatin-derived sandy till or Quebec/Labrador/Hudson-derived Long Spruce and Sky Pilot silty tills (Nielsen and Dredge, 1982; Klassen, 1986; Nielsen et al., 1986; Dredge et al., 1990; Dredge and Nixon, 1992; Roy, 1998; Nielsen, 2001, 2002a; Dredge and McMartin, 2011). Most sections are capped by sediment deposited within Glacial Lake Agassiz (Dredge, 1983; Klassen, 1983; Thorleifson, 1996; Teller and Leverington, 2004), the post-glacial Tyrrell Sea, or both. In actuality, the stratigraphic data-set is more complex, and reflects a long history of major and minor shifts in ice-sheet behavior throughout several advance and retreat cycles, as well as intervening interglacial and interstadial(?) stages. Nielsen (2002a) conducted a study along the lower Nelson, Hayes, Gods and Pennycutaway rivers, that incorporated analysis of numerous up-section till fabrics and samples (composition, geochemistry). This study documents multiple till units with varying concentrations of north and west-sourced indicator clasts (Nielsen, 2002b) and a large range of till-fabric orientations (Nielsen, 2002a). While significant transitions in the source-area of ice over time are notable, it is very difficult to separate the stratigraphy into four simple till units.

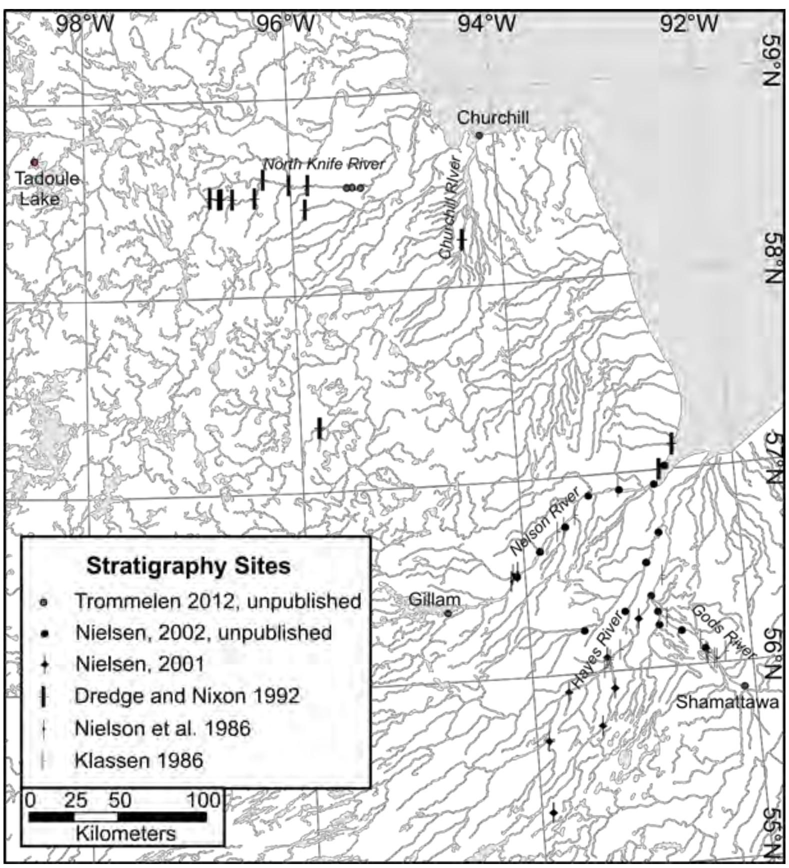


Figure 1-8. Quaternary stratigraphic sections documented along river valleys in northeast Manitoba. Most of these sections are summarized in Dredge and McMartin (2011). The lithological data and site locations from Nielsen (2002a) are currently unpublished, and resides at the MGS. Data for sections visited by this author in 2012 are awaiting publication by the MGS, pending final radiocarbon and till composition results.

1.3.1.2 ICE-FLOW HISTORY

The orientation of ice-flow indicators (striae and roches moutonnées) in northern Manitoba (Figure 1-9) were collected during fieldwork, while the orientation of ice-flow parallel streamlined landforms was mapped using aerial photographs. The majority of ice-flow indicators within the study area are correlated to the Wisconsinan glaciation (Dredge et al., 1986; Dredge and Nixon, 1992), as are all streamlined landforms (Prest et al., 1968; Dyke et al., 1982; Dredge et al., 1986; Dyke and Prest, 1987b; Dyke and Dredge, 1989; Boulton and Clark, 1990; Dredge and Nixon, 1992; Dredge and Pehrsson, 2006; Kaszycki et al., 2008). There are, however, several sites that have rare east- or northeast-trending striations, such as those surrounding Big Sand Lake (Figure 1-9) and at Churchill (Dredge and Nixon, 1992). As first postulated by Dredge and Nixon (1992), these rare east-trending striae are probably inherited and correlate to Pre-Wisconsinan glaciations. Dyke and Prest (1987b), and Dyke and Dredge (1989) interpreted the northeast-trending striae to have been formed during migration of an ice-saddle over the study area.

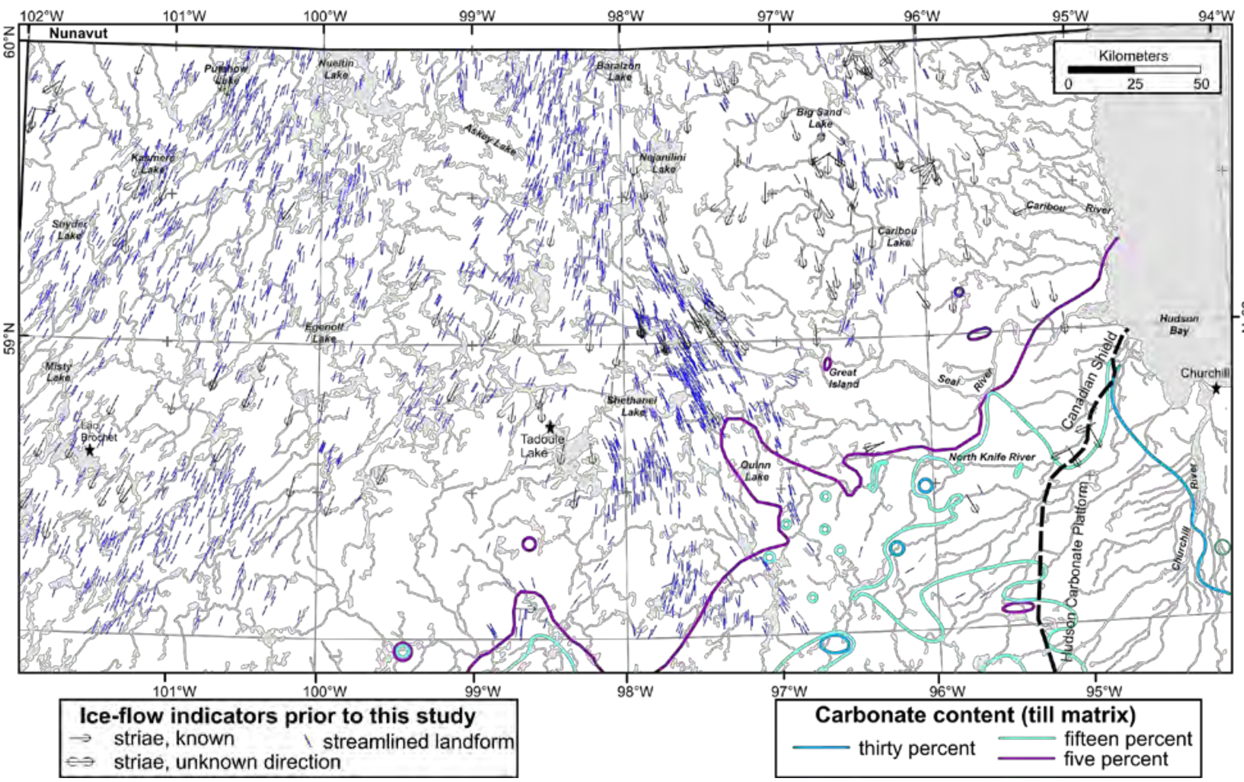


Figure 1-9. Field-based (striae and till-matrix total carbonate) and remotely-sensed (landform) data in northern Manitoba. Ice-flow indicators are predominately sourced from Dredge et al (2007), and supplemented by Dredge and Nixon (1981b), Dredge et al. (1982a) and Matile (2006). Updated carbonate-content is contoured based on data from Dredge and Pehrsson (2006), Chapter 3 and unpublished sites collected in 2012 during work with a company.

1.3.1.3 Till composition

Till composition analysis in northern Manitoba is more complex, and in correspondence with the stratigraphic data, has led to interpretation of a complex glacial history and recognition of inheritance within till. Known far-travelled dispersal of north-sourced Proterozoic Dubawnt red volcanic and volcaniclastic rocks, and west-sourced Paleozoic carbonate and red siltstone, Proterozoic Omaralluk greywacke (Figure 1-10; see Chapter 3 for details) indicator clasts (Dredge, 1988; Nielsen, 2001, 2002a; Kaszycki et al., 2008) are the best indicators of older ice-flow phases, incomplete erosion and recycling of older deposits.

Carbonate content in the till matrix (Figure 1-9) is predominately derived from the breakdown of grey to buff carbonate rocks sourced from the Hudson Platform. Within the study area, carbonate detritus has travelled at least 80 km northwest and 140 km west from the source area (Dredge and Pehrsson, 2006). Though carbonate dispersal is palimpsest, there is not a strong correlation between streamlined landform orientation (where present) and carbonate dispersal (Figure 1-9). South of the study area, this carbonate

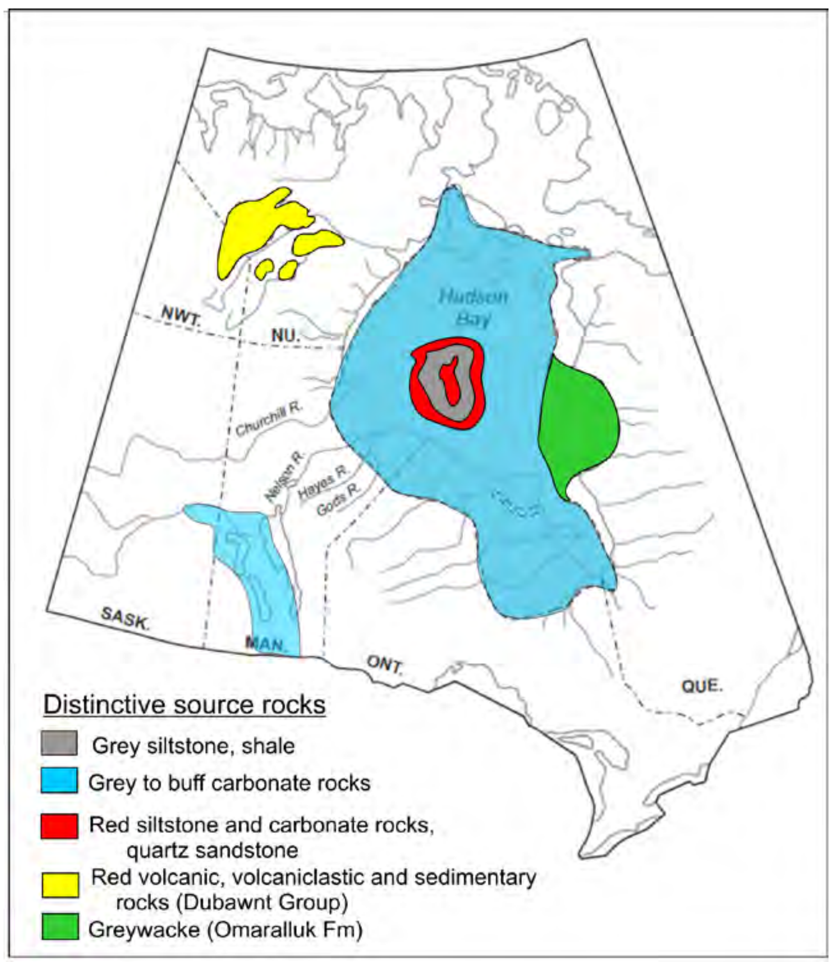


Figure 1-10. Distinctive source rocks that have been transported by glaciers to and/or across parts of northern Manitoba. Modified from Kaszycki et al. (2008) and Dredge and McMartin (2011).

detritus has travelled up to 260 km west across Manitoba (Dredge, 1988; Kaszycki et al., 2008; McMartin et al., in press). This long-distance transport compares well with a compositional analysis of till inland of western Hudson Bay, Nunavut, where Shilts et al. (1979) determined that southeast dispersal trains of Dubawnt erratics extend more than 300 km and were apparently undisturbed by later migration of the Keewatin Ice Divide (KID).

1.3.2 Laurentide Ice Sheet (LIS) configuration

At one time, the LIS was thought to emanate from a central dome situated over Hudson Bay (Sugden, 1977; Denton and Hughes, 1981). The single dome model was slowly replaced by the original multi-dome model (Tyrrell, 1898a, b; Dyke et al., 1982), which acknowledged the spatial variations in isostatic rebound, clast and till geochemistry dispersal trains (Shilts et al., 1979; Shilts, 1980), leading to the portrayal of major ice domes/divides centered over Keewatin (KID), northern Quebec/Labrador (Labradorean), and Hudson Bay. Over the past 30 years, there have been several attempts to reconstruct the ice-flow history in northern Manitoba, as part of a regional picture that often includes Nunavut, Saskatchewan and sometimes northern Ontario. These paleoglaciological reconstruction models have used various proxies for ice flow, including streamlined landforms and/or striations (Boulton and Clark, 1990; McMartin and Henderson, 2004; Kaszycki et al., 2008), clast dispersal trains (Shilts et al., 1979; Shilts, 1980; Dredge, 1988; Dyke and Dredge, 1989; Dredge and Nixon, 1992; Kaszycki et al., 2008), surficial geology mapping (Dredge et al., 1985; Dredge and Nixon, 1986; Kaszycki et al., 2008) and remote-sensing based mapping (Kleman et al., 2010). Field researchers have shown that the KID migrated by as much as 500 km between ice-flow phases and possibly throughout much of the Wisconsinan glaciation (Cunningham and Shilts, 1977; Klassen, 1995; McMartin and Henderson, 2004). Generally, though, regional to continental-scale reconstructions have assumed that most streamlined landforms were generated during the final decay stages of the LIS, such as the radial depiction in Figure 1-11 by Boulton (1987) or the more complex 1:250 000 scale late-deglacial retreat depicted by Dredge and Nixon (1992) (Figure 1-12).

Proximity to the KID must have affected formation of the glacial landscape in northern Manitoba, but there is little suggestion of inherited or palimpsest landforms within the current mapping. In Nunavut, McMartin and Henderson (2004) did identify cross-cutting and palimpsest inherited streamlined landforms. Generally, though in absence of cross-cutting, the paucity of ages (radiocarbon or otherwise; Dyke, 2004) makes it difficult to assign even minimum ages to the landscape – thus limiting recognition of older terrain. This contrasts with old, widespread, pre-Late Wisconsinan subglacial landforms have been mapped beneath core regions of ice sheets in Scandinavia (Hättestrand and Stroeven, 2002; Fabel et al., 2006; Moller, 2006; Goodfellow et al., 2008; Kleman et al., 2008), Scotland (Finlayson et al., 2010), Nova Scotia, Canada (Stea, 1994), and Quebec/Labrador, Canada (Kleman et al., 1994; Clark et al., 2000; Clärhall and Jansson, 2003) and thus would also be expected within the Keewatin Sector of the LIS. Kleman et al. (2010) did identify multiple generations of streamlined landforms across the Keewatin Sector, but their

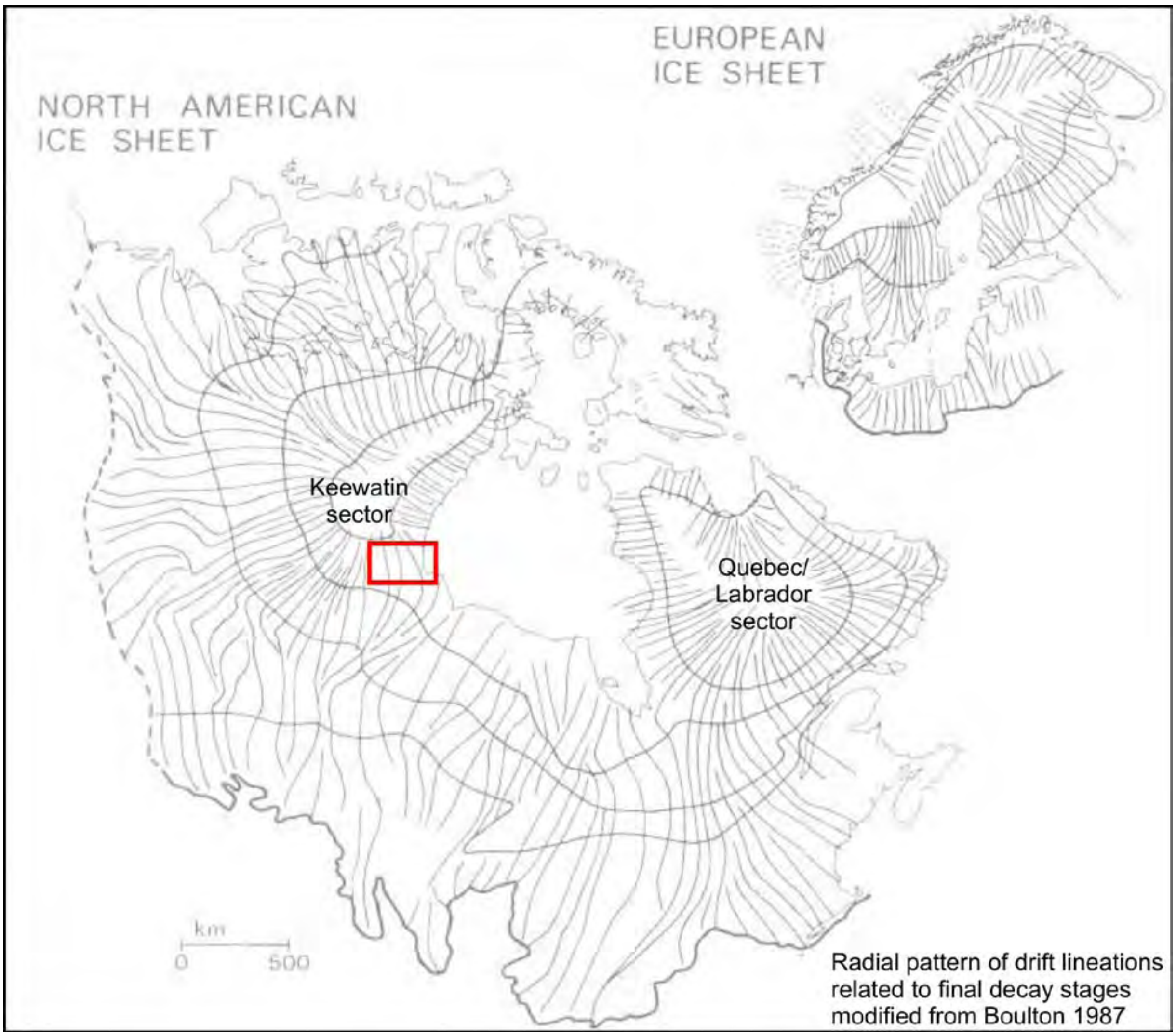


Figure 1-11. Streamlined landforms mapped from remote-sensing and interpreted as 'radial lineations' that depict flowlines related to the final decay stages of the LIS; modified from Boulton (1987). This diagram includes the maximum extent of the LIS, and hence perhaps Boulton thought streamlining either occurred at LGM or that the ice sheet retained the same general shape throughout deglaciation. Northern Manitoba is outlined by the red box.

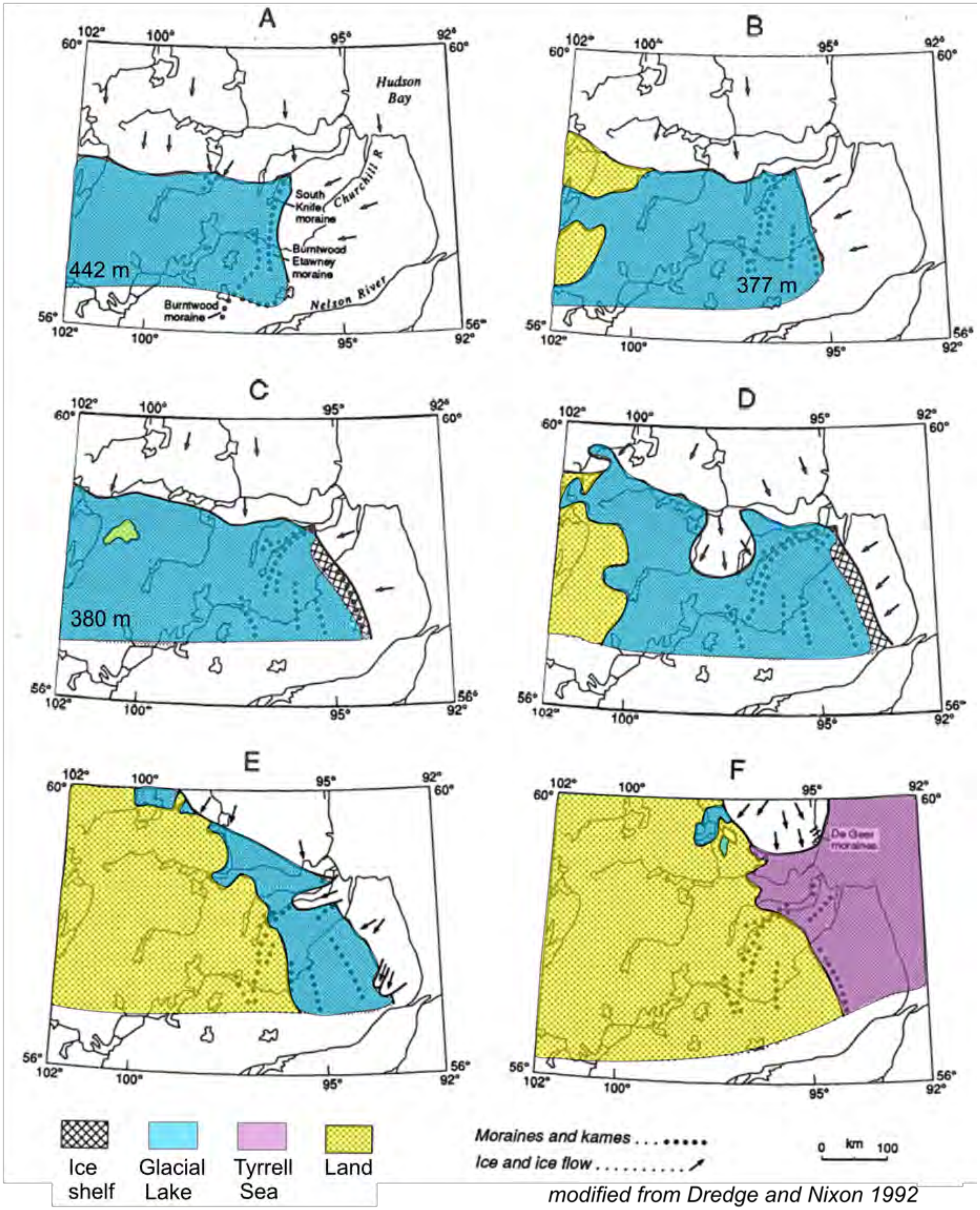


Figure 1-12. Late deglacial ice retreat and expansion of Glacial Lake Agassiz and the Tyrrell Sea within northern Manitoba, modified from Dredge and Nixon (1992). Where denoted, elevation measurements correspond to Lake Agassiz levels. This reconstruction utilizes striae and streamlined landform orientations, as well as the location of ice-marginal landforms and stratigraphy.

first-order remotely-sensed reconstruction is arguably simplistic and lacks input from detailed observation-based data.

Using Dubawnt-erratic dispersal data, Shilts (1980) significantly revised the LGM ice-flow depiction of the LIS (Figure 1-13) – though the concept of radial flow lineations (Figure 1-11) was accepted to at least the late 1980's (Boulton, 1987; Aylsworth and Shilts, 1989). Shilts et al. (1979) also noted that the time required for dispersal of carbonate and Dubawant erratics was much greater than 'late-deglacial' reconstructions allow, and hence suggested a longer time duration for till and landscape generation.

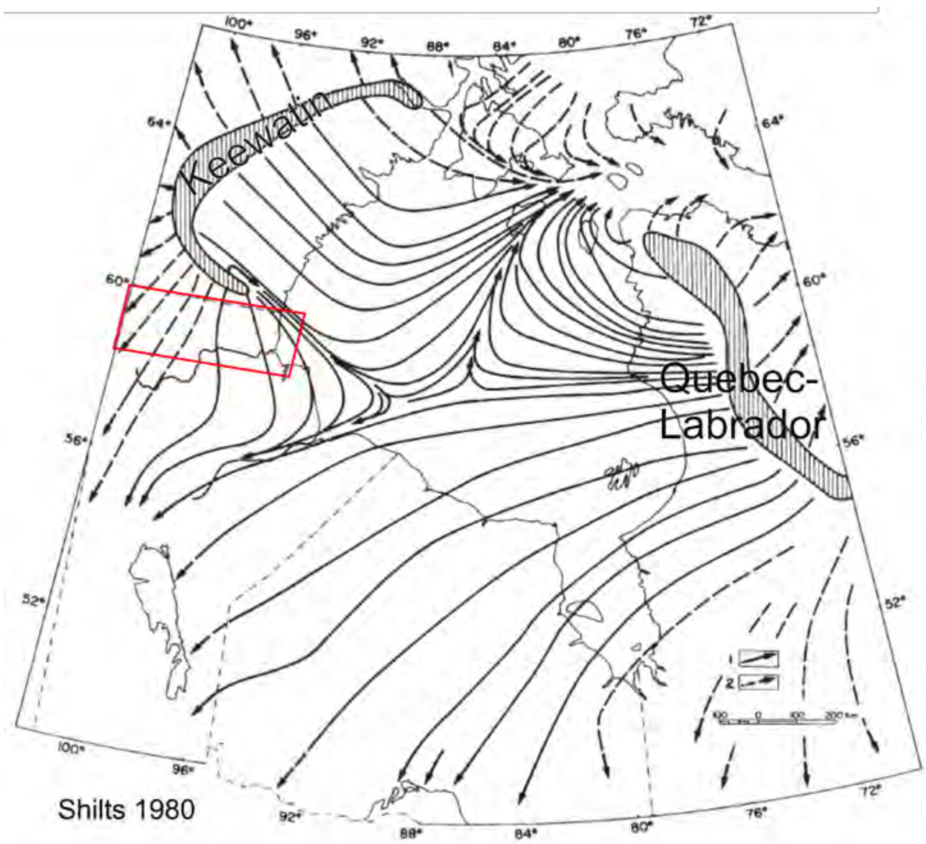


Figure 1-13. Interpreted ice-flow for the Keewatin and Quebec-Labrador sectors of the LIS at LGM, generated using field-based ice-flow indicators and dispersal of carbonate and Dubawnt erratics; modified from Shilts (1980). Northern Manitoba is outlined by the red box.

The multi-dome model of the LIS has continued to evolve, leading to debate over the configuration of ice in northern Manitoba. Possibilities include interaction between the KID and ice flowing [A.] from the Quebec/Labrador sector (Figure 1-12; Shilts, 1980; Boulton and Clark, 1990) [B.] from a Hudson Bay dome (Dyke et al., 1982; Dredge et al., 1986), [C.] from a saddle between the migrating KID and a Quebec/Labrador dome (Figure 1-14; Dyke and Prest, 1987b; Kaszycki et al., 2008), or [D.] from a saddle between the migrating KID and a late deglacial Hudson dome (Figure 1-14; Dyke and Prest, 1987b; Dyke and Dredge, 1989). It is known that Labradorean ice reached northern Manitoba (dispersal of erratics from eastern Hudson Bay, Henderson, 1989; Prest et al., 2000; Johnston and Schreiner, 2011), but debate still exists as to what the suture/ridge/saddle/convergence/divergence zone may have looked like over the

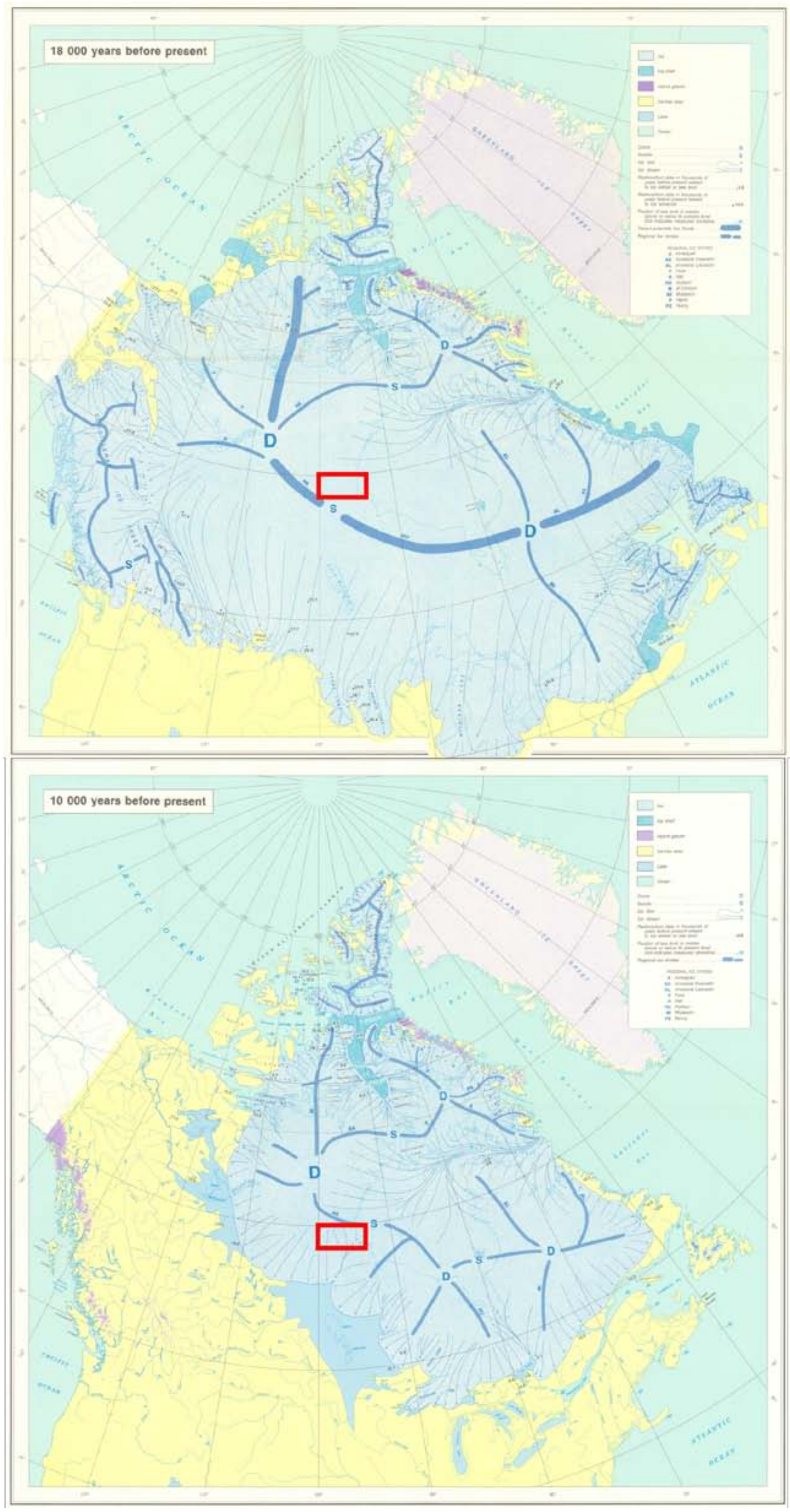


Figure 1-14. Select paleogeography maps of northern North America, modified from Dyke and Prest (1987b). Northern Manitoba (red box) is depicted as underlying a large saddle between the Keewatin and Quebec/Labrador sectors of the LIS near LGM, and eventually beneath a saddle between the Keewatin and Hudson sectors at the start of the Holocene. The reader is directed the full publication and all eleven time-slice maps for the full deglacial reconstruction.

glacial cycle(s) (Dredge and Pehrsson, 2006; Dredge and McMartin, 2007; Kleman et al., 2010). Inferences have been made regarding the Pre-Late Wisconsinan (Dredge and Thorleifson, 1987; Vincent and Prest, 1987; Boulton and Clark, 1990; Kleman et al., 2010) ice-flow history, but without detailed fieldwork there is a lack of data to support concrete interpretations.

More recently, the multi-dome model has incorporated the concept of ice streaming (De Angelis and Kleman, 2005; O Cofaigh and Stokes, 2008; Greenwood and Clark, 2009a, Stokes et al. 2012). Slower or 'steady-state' ice flow is inferred between these ice stream corridors, and new reconstructions include the idea of rapid dynamic changes at short temporal scales. Newer paleoglaciological reconstruction models for Quebec-Labrador (Clark et al., 2000; Jansson et al., 2002; Clärhall and Jansson, 2003), Keewatin (Nunavut, McMartin and Henderson, 2004; De Angelis and Kleman, 2005) and Newfoundland (Blundon et al., 2010) now recognize that glacial and subglacial landforms record a mixture of isochronous (likely inherited) and time-transgressive flow phases. Indeed, in the late 1980's, Dredge (1988) was one of the first to postulate that ice streams crossed northern Manitoba, and were responsible for transport of carbonate material up to 260 km beyond the limit of carbonate bedrock. Since then, the location of an ice stream in northeastern Manitoba was noted by Patterson (1998), but has not been formally named or recognized by later compilations (Winsborrow et al., 2004; Stokes and Tarasov, 2010).

Using observation-based ice-flow reconstructions of the LIS, it is possible to make several inferences regarding the subglacial conditions near this inner core region of the LIS in northern Manitoba:

1. The region was subject to at least two glacial cycles (Illinoian and Wisconsinan river stratigraphy) and 'the lack of evidence' on the subglacial landscape relating to the penultimate glacial cycle indicates that erosion was the dominant process during the Late Wisconsinan.
2. In contrast to the theoretical cold-based interpretation, the glacial landscape was predominately formed by near-complete overprinting during warm-based Late Wisconsinan deglaciation. Importantly, a lack of datable material (or methods) means it is difficult to determine the ages of subglacial landform/striae/dispersal train formation – and thus the default observation-based hypothesis has been 'deglacial'. Scattered rare pockets of regolith, and rare striae trending between E and NE indicate at least some inheritance.
3. Based on the variable concentration of transported north-sourced (Proterozoic Dubawnt) and west-sourced (Paleozoic carbonate and red siltstone, Proterozoic Omaralluk greywacke) clasts within the regional till, some subglacial detritus is inherited and indicates erosion was

incomplete. There is disconnect between transport directions/distances and overlying streamlined landform orientations, again suggesting that in contrast to (1), the retreat-phase may have been weaker than the advance-phase erosion/transport/deposition.

1.4 AIMS OF THESIS

It is clear that the ice-sheet wide theory-based hypotheses for northern Manitoba are in contrast to the regional to continental-scale observation-based hypotheses, as presented above. Both methods attempt to incorporate field-based data, using either top-down (theory to predict what should be generated) or bottom-up (observation data means that x must have occurred) techniques. The resulting paleoglaciological reconstructions, however, can differ greatly depending on the scale of observation-based data that that is incorporated. In addition, field-based observational data is significantly more important than remotely-sensed 'observational data'. Difficulty in up-scaling the large body of detailed Canadian mapping data means that this data has been poorly integrated into the theoretical work completed by academic researchers. This lack of integration may explain the ongoing apparent disconnection between detailed field observations and published concepts and theories (c.f. Tarasov and Peltier, 2004; Kleman and Glasser, 2007; Tarasov and Peltier, 2007; Kleman et al., 2010; Stokes and Tarasov, 2010; Stokes et al., 2012). Theoretical reconstructions, while useful and much-cited within the fields of glacial geology, paleoglaciology and paleoclimatology, are typically oversimplified at a local to regional field-scale. As such, these reconstructions are of limited use for domains such as detailed fieldwork and mineral exploration. Field-based data, especially at a detailed scale, can be used to test the validity of model assumptions, and refine the outcomes (i.e. problem of thick till beneath an ice divide, Kleman et al., 2008). Conversely, theoretical paleoglaciology inferences can be used to better-understand the contradictory field data (i.e. idea of inheritance beneath cold-based ice). When theoretical models do incorporate field evidence, the conclusions for both observation-driven and theory-driven studies are strengthened (Shilts, 1980; Clärhall and Jansson, 2003; Hildes et al., 2004; i.e. Tarasov and Peltier, 2004; Stokes and Tarasov, 2010; Stokes et al., 2012).

Several advances in paleoglaciology have occurred when observational and theoretical methodologies have been combined. For example, while net-erosion may have dominated during certain time intervals as postulated by Sudgen (1978) and Boulton (1996), newer studies in Scotland (Finlayson et al., 2010; Finlayson, 2012) and Ireland (Knight, 2010) suggest erosion may not have been spatially extensive. Secondly, an interpretation paradigm-shift from 'radial' to 'non-contemporaneous' generation of streamlined landforms in numerous paleo-subglacial landscapes (Clark et al., 2000; Greenwood and Clark, 2009b; Ross et al., 2009; Stokes et al., 2009; Clark et al., 2012) has not-yet been applied to northern Manitoba. New interpretation of landscape inheritance would bring this observation-based data in-line with theory-based reconstructions that expect more slow-flowing cold-based ice to preserve/protect the landscape from complete overprinting. As such, a detailed analysis of the subglacial landscape of northern Manitoba may provide a better understanding of the spatio-temporal paleoglaciology for this inner core region of the LIS.

This thesis aims to use detailed field-based data, together with remotely-sensed data, to reconstruct the paleoglaciology of northern Manitoba. Theoretical concepts will be tested, and used to help clarify the numerous contradictions within existing observation-based reconstructions. This thesis has several objectives, with the goal of reconciling theory-based and observation-based hypotheses:

1. Re-examine the inference that the subglacial landscape in northern Manitoba reflects deglacial processes (post-LGM).
2. Investigate the relationship between detailed field-based ice-flow indicators, streamlined landforms and till composition – to assess the amount of inheritance present.
3. Consider how a fragmented mosaic of preservation and overprinting at a landscape and landform-field scale may have developed and evolved over time. This includes an investigation of the conspicuous Rogen moraine in northern Manitoba.

1.5 Thesis format

The data generated during this project is introduced in the relevant chapters.

Chapter 2 presents an analysis of the subglacial landscape and field data in northeast Manitoba. A key aspect of this work is the spatial identification of streamlined landform flowsets and field-based ice-flow indicator data, together with an assessment of the associated temporal ice-flow phases. Spatio-temporal analysis allowed creation of a new glacial terrain zone (GTZ) methodology, which uses detailed field-based and remotely-sensed characteristics to identify internally-coherent landscape-scale zones.

A relative estimation of the degree of inheritance and overprinting is provided for the five zones delineated within northern Manitoba. It was found that the subglacial landscape is best described as a mosaic of variation in the level of inheritance. A subglacial bed mosaic model of landscape evolution is developed to explain the landscape record, whereby slow, sluggish or non-flowing basal ice (termed 'sticky regions') exists amongst areas of higher erosion and faster flowing ice. Mechanisms to generate spatio-temporal variability in sticky regions are discussed. The reader should note that Chapter 2 is now published in *Quaternary Science Reviews* (Trommelen et al. 2012a). The research was designed with the thesis supervisor (second author), but the work was conducted largely by the author of this thesis (who also led the interpretation work). The third author (Janet Campbell) was the main NRCan collaborator for this project component. Lastly, the two co-authors provided assistance in the field, as well as guidance and useful comments on the written manuscript before submission. Subglacial landform mapping for the entirety of northern Manitoba (58°-60°) is also published as a map with an accompanying paper (Thesis Appendix, Trommelen and Ross, 2010). Surficial maps (1:50 000) are co-published by the Geological Survey of Canada and the Manitoba Geological Survey (Trommelen and Campbell, 2012a,b,c,d; Thesis Appendix).

Chapter 3 builds on the GTZ concept developed in Chapter 2 and further examines how the complex ice-flow reconstruction relates to clast dispersal and erosion intensity in the Great Island area. The work shows that the regional surface till is a spatial mosaic reflective of variable intensity in entrainment (overprinting) and preservation (inheritance) of a pre-existing till sheet. Regionally, early phases of high-intensity ice flow likely generated a heterogeneous surface till, which was then followed by low-intensity (long duration) ice flow that variably entrained, diluted and/or comminuted the pre-existing tills while locally preserving numerous high-inheritance (low erosion) areas throughout the Wisconsinan. An example is provided of how till composition, combined with regional knowledge, can be used to map relative spatio-temporal erosion intensity. Chapter 3 is published in *Boreas* (Trommelen et al., accepted). As above, this work was conducted largely by the author of this thesis and the same two co-authors played similar roles as for Chapter 2. Till composition data is presented in Campbell et al. (2012). This report outlines the basic geology of the area, as well as provides an outline of the geochemical analyses.

Chapter 4 investigates the Rogen moraine and streamlined subglacial landform assemblages that dominate portions of northern Manitoba. Advances in the subglacial mosaic model paradigm are also presented. The characterization of Rogen moraine includes mapping, sedimentology, till composition, clast fabrics, and internal architecture as determined from a near-surface S-wave seismic survey. This work has indicated that Rogen moraine in northern Manitoba are likely palimpsest subglacial landforms, generated primarily from pre-existing sediments in areas of low erosion and glacial landscape inheritance near the inner-regions of ice-sheets. Current theories of Rogen moraine formation are reviewed and assessed using the new field-based data, leading to adoption of the instability hypothesis for formation. The chapter also includes a discussion of how and when Rogen moraine may have been emplaced, preserved and/or overprinted over the long glacial history of this inner core region of the Laurentide Ice Sheet (LIS). As of now, Chapter 4 has been submitted to co-authors for initial review, and submission to the journal *Quaternary Science Reviews* is expected. This work was largely conducted by the author of this thesis, though uses tools prepared/interpreted by co-authors. A. Ismail processed the seismic data, helped with interpretation, and contributed to the seismic figure and a discussion of the seismic results.

Chapter 5 provides a brief discussion of the effectiveness of the methods used herein. Chapter 6 summarizes the overall findings and contributions to Quaternary Science, and discusses future work that would build upon the research presented in this thesis.

Chapter 2 Glacial Terrain Zone analysis of a fragmented paleoglaciological record, southeast Keewatin sector of the Laurentide Ice Sheet

2.1 Overview

A highly fragmented subglacial landscape is recognized at the regional to sub-regional scales in northeastern Manitoba, Canada, in the southeast Keewatin Sector (a core region) of the Laurentide Ice Sheet. New field-based ice-flow indicator measurements, mapping of subglacial landforms (remote-sensing and aerial photograph), and a re-examination of previously published data from an 8100 km^2 area in northeastern Manitoba show that the preserved subglacial record reflects a complex and potentially long glacial history. Five streamlined landform flowsets are mapped. A much higher degree of inheritance in the field-based ice-flow indicator data, than previously reported, allows for recognition of multiple ice-flow phases. Analysis of the characteristics of the subglacial landscape combined with a relative-age chronology established with field-based indicators, led to the recognition of disjoint zones with internally-consistent glacial histories – termed *glacial terrain zones (GTZ*). These GTZ were then classified as (1) relict glacial, (2) palimpsest, or (3) deglacial in nature.

Our data suggest that while the southern Keewatin Sector was affected by regional ice-divide translocation, this alone cannot explain the fragmented, high inheritance landscape. We suggest that the subglacial landscape was continually evolving and subject to spatio-temporal variations in intensity of erosion, transportation and/or deposition throughout multiple glacial events (subglacial bed mosaic). Preservation of relict and palimpsest terrain likely occurred under large 'sticky' low-erosion regions. These regions could have formed by at least two different mechanisms: heterogeneous switch from warm-based to cold-based ice or within a warm-based subglacial environment from wet to stiff, dewatered till. Establishment of the regionally extensive (~700 km wide by at least 500 km long) dendritic esker channel-system may have caused rapid spatially-variable dewatering of the substrate far back under the ice sheet. The GTZ approach integrates all available data (e.g. flowsets and other landform data, striations) to advance our interpretation of the spatio-temporal evolution of subglacial dynamics in areas where the degree of landscape inheritance and overprinting is spatially highly variable. This mosaic may be a characteristic net-effect of landscape evolution beneath the core regions of ice sheets.

2.2 Introduction

The geomorphic landscape of glaciated terrain is a unique window into the subglacial environment (Stea, 1994; Kleman and Borgstrom, 1996; Clark, 1999; O Cofaigh and Stokes, 2008). The advent of remote sensing has brought about new ways of investigating this landscape at an unprecedented scale. New imagery, for example, has allowed for recognition of distinct and non-coeval regional groupings (flow stages, Boulton and Clark, 1990; flowsets, Kleman and Borgstrom, 1996; Clark, 1999) of streamlined landforms (Jansson et al., 2002; Greenwood and Clark, 2009b; Stokes et al., 2009). The mapping of these non-coeval ice flowsets across large regions is then used to recognize regional trends and spatio-temporal shifts in ice-flow patterns, which are then integrated in ice-sheet reconstructions (e.g Kleman et al., 1997; Clark et al., 2000; Kleman et al., 2008; Greenwood and Clark, 2009a). In many cases, this approach has led to significant changes in our understanding of ice-sheet dynamics and evolution. Hence while it was once thought that the Keewatin Sector of the Laurentide Ice Sheet (LIS) was a long-lived stable ice centre (Shilts et al., 1979; Shilts, 1980; Aylsworth and Shilts, 1989), satellite imagery has now provided compelling evidence for a dynamic, migrating ice sector (Boulton and Clark, 1990; Kleman et al., 2010). The current flowset approach largely focuses on the characterization of streamlined landforms at vast regional scales (Boulton and Clark, 1990; Kleman et al., 1997; Clark et al., 2000; Kleman and Glasser, 2007; Stokes et al., 2009; Kleman et al., 2010; Smith and Knight, 2011). The success of this methodology has lead to application at smaller regional to local scales, where complications such as palimpsest (overprinted, Greenwood and Clark, 2009b) landforms, or relict (inherited, Finlayson et al., 2010) landforms emerge. Where cross-cutting relationships are difficult to decipher (Greenwood and Clark, 2009b, a), or where flowsets start/end adjacent to, but not crossing, other flowsets (this study), relative age is often difficult to determine. As one aspect of paleoglaciological reconstruction is a timeline of events, researchers need a multi-proxy set of tools (Kleman et al., 1997) to aid spatio-temporal interpretation of detailed, small-scale reconstructions. Other characteristics of the glacial landscape that may be indicative of different glaciological regime or non-coeval glacial records are also important (Greenwood and Clark, 2009a). These may include, for example, distinct clusters of other types of subglacial landforms (e.g. Rogen moraine, end moraines, De Geer moraines), field-based ice-flow indicator data and contrasting sediment dispersal patterns. It is thus critical for paleoglaciological ice-sheet reconstructions to not only recognize and describe distinct flowsets, but also to investigate the spatio-temporal relationships between all available data to identify evidence of inheritance (e.g. relict surfaces, mis-matched orientations, old ice-flow indicators, old sediment dispersal orientations) and/or the intensity of overprinting (e.g. palimpsest landforms, radiating lobate patterns, parallel conformity with esker ridges) (Greenwood and Clark, 2009a).

In this paper we present a spatio-temporal *glacial terrain zone (GTZ)* approach, which combines the typical flowset mapping technique with terrain characteristics and data obtained during detailed fieldwork (e.g. topography, non-streamlined landforms, field-based ice-flow indicator record), and apply it to analyze the subglacial landscape (8100 km^2) within a core region (near-ice divide) of the LIS in northeast Manitoba, Canada. We then present previously unknown aspects of the glacial history of the southeast part of the Keewatin sector of the LIS, identify potential problems and knowledge gaps, and use new insights from this analysis to advance discussion of spatio-temporal subglacial landscape evolution.

2.2.1 Flowsets, landsystems and Glacial Terrain Zones

Flowset mapping began as a way to document complexity in the distribution of subglacial streamlined landforms, recognize consistent patterns, and inform reconstruction of ice-flow history in glaciated regions (Boulton and Clark, 1990; Kleman et al., 1994; Clark et al., 2000; Boulton et al., 2001b; Jansson et al., 2002). A flowset is defined herein as a discrete assemblage of subglacial streamlined landforms based on their pattern and the degree of internal consistency (e.g. parallel landforms with similar morphology), from which inferences about ice-flow history and basal thermal regime can be made (Kleman and Borgstrom, 1996; Kleman et al., 1997; Clark, 1999). Various attempts have been made to identify particular types of subglacial landform patterns and to relate them to specific subglacial conditions or ice-sheet phase (i.e. warm-based, cold-based, event-flow, deglacial) (Greenwood and Clark, 2009a; Stokes et al., 2009). For example, Clarhall and Jansson (2003) recognized that a fragmented glacial landscape in the Quebec-Labrador sector of the LIS, identified by partially superimposed flowsets, is evidence of temporally unrelated landform systems. They then suggested that subglacial landform generation may have occurred in discrete subglacial zones during restricted time periods. Likewise, De Angelis and Kleman (2008) identified several paleo-ice stream onset zones that are geomorphologically distinct from adjacent older palimpsest or relict terrain.

By grouping streamlined landforms into flowsets and by analyzing the spatial relationship between various flowsets, we note that the subglacial landscape may be described as a mosaic of distinct landform patterns (Ross et al., 2009; O Cofaigh et al., 2010) or structural elements (Boulton et al., 2001b). When other characteristics are introduced, such as topography, non-streamlined glacial landforms, field-based ice-flow indicators and sediment composition, it becomes apparent that these are actually sediment-landform assemblages or subglacial zones (Stea and Finck, 2001). We define these subglacial zones as

Glacial Terrain Zones (GTZ), whereby a GTZ is a geologically distinct area that contains an internally consistent assembly history. This GTZ approach is further developed and detailed below (section 3.4) but here we argue that the partitioning of the glacial landscape into GTZ is an important step towards more complete paleoglaciological reconstruction models of past ice sheets. The concept of zonation was first presented by Stea and Finck (2001) for the Nova Scotian glacial landscape beneath Scotian Ice Divide of the Appalachian Ice Complex. These authors suggested each zone recorded a different signature of erosion and deposition, which led to a discrete sediment transport history. Similarly, warm-based fast-flowing ice-stream troughs are distinguished from slow, sluggish, often cold-based upland areas leading to the recognition of distinct zones on the landscape which record contrasting subglacial conditions (Briner et al., 2006; Kleman and Glasser, 2007; Ross et al., 2011). Indeed, paleo-ice stream research has evolved to the point where "paleo-ice stream landsystem" models (Clark and Stokes, 2003) can be identified from the subglacial landscape based on specific geomorphic criteria (O Cofaigh et al., 2002; Clark and Stokes, 2003; Anderson and Oakes Fretwell, 2008; De Angelis and Kleman, 2008; Ross et al., 2009). These glacial landsystems (Evans, 2003b) use geomorphic, sedimentological and stratigraphic data to create process-form models that then relate to specific glaciation styles and/or ice dynamics (Evans, 2003a). However, this type of zonal analysis has rarely been applied to the core regions of ice sheets in areas with subdued topography such as the Keewatin Sector of the LIS. Most landsystem models are concerned with spatial identification of specific units or elements, rather than partitioning of the *entire* glacial landscape into zones. These models are also often viewed as conceptual models synthesizing the diagnostic landscape features of a particular environment (e.g. paleo-ice stream landsystem, ice-marginal landsystem). We suggest that, due to translocation in ice-divide positions (McMartin and Henderson, 2004; Greenwood and Clark, 2009b; Finlayson et al., 2010), potential for high inheritance due to limited erosion, and possible changing thermal regimes (Kleman et al., 1999; Clärhall and Jansson, 2003; Kleman and Glasser, 2007), core regions of ice sheets are likely to be characterized by different zone types reflecting not only different erosion/transportation/depositional processes over short distances and through time, but also distinct subglacial thermal regime histories. GTZ analysis can be used to identify and classify these zones.

2.3 REGIONAL SETTING AND PREVIOUS WORK

The study area is located in the northeastern part of Manitoba, on the Canadian Shield, with fieldwork completed in the Great Island-Kellas Lake and Churchill areas (Figure 2-1). Elevation varies from sea level to 330 m above sea level (asl). Local relief is about 30 m high. The northern part of the area is characterized by extensive swaths of bouldery pristine and drumlinized Rogen moraine fields alternating with swaths of streamlined terrain and areas of bedrock outcrops (Trommelen and Ross, 2010). The remaining area is primarily a mix of till blankets and till veneers over bedrock with large regions of peat bogs. Long (100 km), large (30-40 m wide) eskers are present throughout the area, at 8-18 km intervals. The study area has, in part, been wave-washed by both the postglacial Tyrrell Sea and Glacial Lake Agassiz (Dredge, 1983; Dredge and Cowan, 1989). The marine limit in the study area is around 180 m asl and the lake limits are above 270 m asl (Dredge, 1983; Dredge and Nixon, 1992).

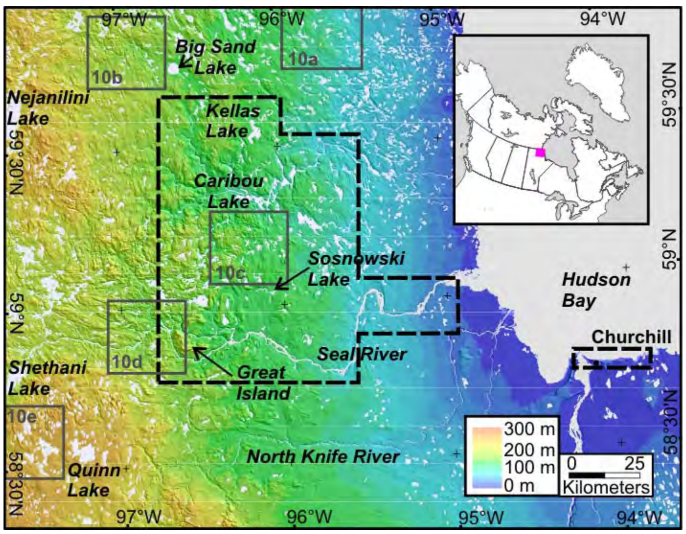


Figure 2-1. The field and study areas in northeast Manitoba. The background image was generated using the radar-derived digital elevation data from the Shuttle Radar Topographic Mission (SRTM). A hillshade model has been added with transparency effect to enhance the relief. Fieldwork and aerial photograph mapping was completed for the large field area, with additional detailed ice-flow fieldwork in the Churchill area. The entire region depicted (study area) has been reviewed in detail and is included herein. Small boxes indicate the location of imagery from Figure 2-10.

Stratigraphic and past ice-flow reconstructions in northeastern Manitoba suggest that the region has been covered at least twice by ice from the Keewatin sector, and at least three times by ice from the Quebec-Labrador sector (Dredge et al., 1986; Klassen, 1986; Dredge and Thorleifson, 1987; Boulton and Clark, 1990; Dredge et al., 1990; Dredge and Nixon, 1992; Kaszycki et al., 2008). Stratigraphic sections in northeast Manitoba broadly suggest the presence, from bottom to top, of a Pre-Illinoian Shield Sundance Till capped by an interglacial paleosol, overlain by the Illinoian carbonate Amery Till, interglacial Sangamonian Nelson River sediments, and Wisconsinan Long Spruce and Sky Pilot tills (Dredge and McMartin, 2011). The study area was situated 200-500 km from the Late Wisconsinan Keewatin Ice Divide (KID) and ~1500 km from the Quebec/Labrador ice divide. Ice was mainly thought to flow south to southeast from the KID, which persisted throughout the Wisconsinan, and occasionally interacted with westward-flowing ice sourced from the east (Hudsonian ice, Dredge and Nixon, 1992; Dredge et al., 2007). At multiple points in time, west or northwest-flowing ice transported and deposited carbonate-bearing clasts/till ~100 km inland (Dredge, 1988) from the source Carbonate Platform in Hudson Bay

(Manitoba Energy and Mines, 1980). The nature of interaction between ice from Keewatin and from Hudson Bay is uncertain. Various reconstructions have depicted Keewatin ice deflected by Hudsonian ice (Shilts, 1980), added yet another ice centre in Hudson Bay (Hudson Dome, Dyke et al., 1982), or delimited a 'saddle' ridge linking the KID and the Quebec-Labrador sector (Dyke and Dredge, 1989; Dyke et al., 2003). Contrastingly, Kleman and Glasser (2007) suggest that most of the region was cold-based (with no interaction) during the Last Glacial Maximum, based on the presence of extensive fields of Rogen moraines (Hättestrand and Kleman, 1999).

Where fieldwork is limited, the age of flowsets cannot be determined directly. The paucity of radiocarbon ages in northern Manitoba (Dyke, 2004) makes it difficult to assign even minimum age to deglacial flowsets. Regional-scale deglacial reconstructions in northern Manitoba (58°-60°) commonly depict a radial pattern, and dominant subglacial landforms are suggested to be of last deglacial age (Prest et al., 1968; Dyke et al., 1982; Dredge et al., 1986; Dyke and Prest, 1987b; Dyke and Dredge, 1989; Boulton and Clark, 1990; Dredge and Nixon, 1992; Dredge and Pehrsson, 2006; Kaszycki et al., 2008). There are several documented sites that show multiple orientations of striations (Dredge et al., 1985; Dredge and Nixon, 1986), which hint at a more complex ice-flow history. Researchers have shown that the KID migrated by as much as 500 km between ice-flow phases and possibly throughout much of the Wisconsinan glaciation (Cunningham and Shilts, 1977; Klassen, 1995; McMartin and Henderson, 2004). Cross-cutting flowsets, however, have not yet been clearly identified in the study area (c.f. Kleman et al., 2010; Shaw et al., 2010). Migration of the KID must have affected formation of the glacial landscape in northern Manitoba (Dredge and Nixon, 1992), and this needs to be further investigated in order to improve paleoglaciological reconstruction. We also note that older, pre-Late Wisconsinan subglacial landforms have been mapped beneath core regions of ice sheets in Scandinavia (Hättestrand and Stroeven, 2002; Fabel et al., 2006; Moller, 2006; Goodfellow et al., 2008; Kleman et al., 2008), Scotland, (Finlayson et al., 2010) and Nova Scotia, Canada (Stea, 1994). To date, the presence of older landforms has been suggested for some areas of the Keewatin and Quebec-Labrador sectors of the LIS (Boulton and Clark, 1990; Clark et al., 2000; Kleman et al., 2010), but not confirmed using detailed analysis of the subglacial landscape.

2.4 METHODS

The two most commonly used proxies for subglacial dynamics are the presence, abundance and orientation of subglacial landforms and field-based ice-flow indicators. We now recognize that there is a very complex interplay between subglacial erosion, till production, and landform development (Boulton et al., 2001a; Christoffersen and Tulaczyk, 2003; van der Meer et al., 2003; Smith et al., 2007; Bradwell et al., 2008; Waller et al., 2008; King et al., 2009; Clark, 2010). It is thus important to integrate all available information, in order to provide a more complete picture of ice sheet evolution at different temporal and spatial scales. Other data used in this study include regional surficial maps (Dredge et al., 1985; Klassen and Netterville, 1985; Aylsworth, 1986; Dredge and Nixon, 1986; Aylsworth and Shilts, 1989; Aylsworth et al., 1990) and publications (Dredge et al., 1986; Klassen, 1986; Dredge and Nixon, 1992; McMartin and Henderson, 2004; Kaszycki et al., 2008; McMartin et al., 2010; Dredge and McMartin, 2011).

2.4.1 Subglacial landform mapping

Using remote-sensing imagery and digital elevation models, Trommelen and Ross (2010) mapped subglacial landforms including ice-flow parallel streamlined landforms (drumlins, mega-flutes, crag-and-tail), ice-flow transverse Rogen moraine, and esker ridges. Imagery used includes Landsat 7 Enhanced Thematic Mapper Plus (ETM+) satellite imagery (15 m resolution, www.geobase.ca) in combination with a Shuttle Radar Topography Mission (SRTM) digital elevation model (90 m resolution; http://srtm.usgs.gov/index.php) and SPOT 4/5 images (10 m resolution; www.geobase.ca). Additional subglacial landforms were mapped from 1:60 000 scale black and white aerial photographs within the field area.

2.4.2 Field-based ice-flow indicators

Detailed field-based surficial geology mapping, including mapping of bedrock ice-flow indicators, was conducted in a 5700 km^2 area northwest of Churchill (Trommelen and Ross, 2009; Trommelen et al., 2010). Additional field-mapped bedrock ice-flow indicator mapping was completed along a 30-km stretch of exposed bedrock in the Churchill area (Trommelen and Ross, 2011a). Erosional bedrock ice-flow indicators (henceforth "field-based") documented in the field areas include non-directional indicators such as striae and grooves and directional indicators such as rattails, chattermarks, gouges and stoss-lee relationships (e.g. Veillette and Roy, 1995; Klassen, 1997). Macroform features encountered in the study area include roches moutonnées and whalebacks (rock drumlins). Visited outcrops were surveyed in detail to record rare and protected field-based ice-flow indicators, in addition to the dominant indicators. Striae clearly indicative of ice deflection around bedrock outcrops were noted, but not used in the regional interpretation. Where cross-cutting patterns were found, the relative ages of flows were determined when possible. Post field-season, field-based ice-flow indicator data was up-scaled to identify regional patterns and analyze spatial and relative-age relationships with the landform record. This study includes sparse, yet highly valuable, regional ice-flow data for areas outside of the field area (Dredge and

Nixon, 1981a, b, 1982b, a; Dredge et al., 1985; Dredge and Nixon, 1986; Dredge et al., 1986; Dredge and Nixon, 1992).

2.4.3 Flowset mapping

By using the common paleoglaciological inversion technique of flowset mapping (Kleman and Borgstrom, 1996; Clark, 1999; Clark et al., 2000; O Cofaigh and Stokes, 2008), we are able to better-use clearly-defined characteristics to identify discrete flowsets thought to represent single ice-flow events. As such, each flowset must exhibit parallel conformity and similar morphology of landforms (Clark, 1999; Clark et al., 2000). Flowset mapping traditionally requires the third requirement of close proximity. However, in areas with incomplete overprinting during successive ice-flow phases, it is possible that isolated but similarly oriented flowsets were formed contemporaneously. As such, we have chosen to lump disjoint but similarly oriented flowsets within the project area into one flowset.

2.4.4 Recognition of glacial terrain zones

It has been demonstrated that in some areas the glacial record is highly fragmented (Clärhall and Jansson, 2003; Finlayson et al., 2010; Knight, 2010). We argue herein that this type of landscape is best analyzed if it is partitioned into coherent "puzzle pieces" that each record a unique glacial history (i.e. a record that is distinct than the one found in adjacent zones). This mosaic landscape could be the product of shifting ice streams and related inter-ice stream areas (Ross et al., 2009; Stokes et al., 2009), or the interplay between ice-marginal surging lobes (Evans et al., 2008). In core regions of ice sheets, a similar patchy mosaic landscape is expected; due to translocation of ice divides and spatio-temporal changes in subglacial conditions, which led to highly variable degrees of landscape preservation (mix of inheritance and overprinting to an unknown depth, Clark, 1999). We envision a subglacial landscape evolution model similar to the till mosaic concept of Piotrowski et al. (2004) but at a much broader (landscape or landform-field) scale. GTZ's then represent the complex end-product of this spatial and/or temporal variability beneath the ice sheets (Figure 2-2).

We partition the landscape into GTZ, based upon:

- the number, distribution, orientation, and possibly type of streamlined landform flowsets (drumlins, flutes, mega-flutes and crag-and tail landforms),
    - The outline of a GTZ thus defines a zone in which the degree of inheritance and overprinting of the landscape is internally consistent and distinct from that of the surrounding terrain.

    - Allows some relative temporal characteristics to be determined (time-transgressive or isochronous, c.f. Greenwood and Clark, 2009b) for the entire zone.
    - Allows analysis of cross-cutting patterns to establish relative age relationships (c.f. Greenwood and Clark, 2009b).
- the variation in orientation of field-based ice-flow indicators and relative-age relationships,
    - Allow comparison and, in some cases, correlation with regional flowsets
    - Helps delimit the complexity of the ice-flow history within each zone
    - In this study, field-based ice-flow indicators have been collected within the field-based area, but not the entire study area
- the relationship between esker orientation and streamlined landform orientation (oblique, sub-parallel, both),
    - This helps determine the temporal relationships between flowsets and extensive channelized meltwater phases
- the presence or absence of Rogen moraine or other features (e.g. felsenmeer, hummocks, end moraines),
    - Association with certain types of landforms may help paleoglaciological interpretation of that particular GTZ (e.g. drumlins ending at an end moraine were most likely formed during a late deglacial surge while Rogen moraine may require cold-based or slow sluggish ice for their formation).
- topographic characteristics such as large and sharp topographic steps or localized distinct landscape roughness,
    - Allows delineation of GTZ boundaries in areas where presence/absence and orientation of streamlined and other subglacial landforms correlates well with topography.
    - In this study large topographic steps are not expected, but some GTZ may be characterized by unique topography (e.g. rough or smooth).
- sediment composition related to ice-flow from a distinctive source area (or areas) (e.g. Shield till vs Carbonate till; (Ross et al., 2009; Ross et al., 2011).

We intend GTZ mapping to be a holistic approach to terrain analysis. This means that as long as the overall glacial history is consistent within a GTZ and different from the surrounding GTZ, each GTZ may be characterized by a different set of characteristics (e.g. various geomorphic elements of an older ice-stream landsystem, or a region with four sets of superimposed streamlined landforms). As such, the

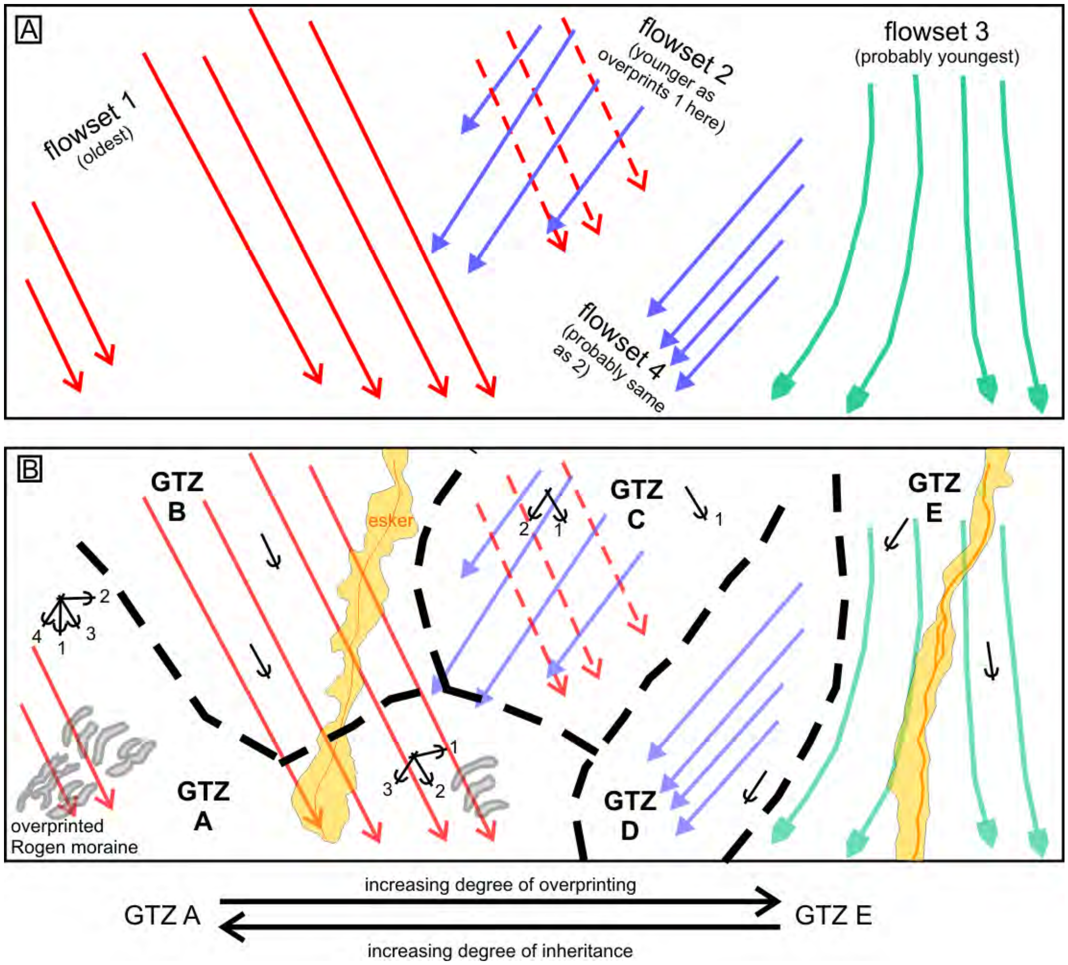


Figure 2-2. Analyzing the subglacial landscape using the widely accepted flowset approach (A), and the GTZ approach (B) which integrates all available data. A more complex record is recognized which contributes to advancing our understanding of paleoglaciological conditions and landscape evolution. Other data such as till composition could be used to refine/test the GTZ boundaries. The subglacial landscape record within a GTZ is representative of an internally-consistent glacial history, which is different from the adjacent GTZ. In this example, the entire study area was glaciated by five differently-oriented ice-flow phases; [1, 2] S and E flow recorded only in outcrop-scale indicators, [3] SE flow, [4] SW flow, [5] lobate flow. The highest inheritance GTZ (A) is defined by the presence of both older streamlined landform orientations, and Rogen moraine, and has not been overprinted by the last ice-flow phase even though the SW-trending phase must have crossed GTZ A (striae evidence). In contrast, the lowest inheritance GTZ (E) is defined by the presence of the youngest lobate streamlined landforms, with no evidence of older ice-flow events. GTZ B, C (palimpsest) and D are intermediate between these two end-members.

boundaries of GTZ can be further tested and redefined when new data becomes available. New data could include till compositional data (e.g. dispersal trains or further delineation of source areas, or perhaps isotope data used to assess subglacial erosion intensity (Briner et al., 2006; Miller et al., 2006; Staiger et al., 2006) across GTZ boundaries. The integration of this type of information is beyond the scope of this paper, though till compositional patterns within the field area are compared with the results of the GTZ analysis in a forthcoming paper.

The defining characteristics are somewhat subjective, in that we do not provide any importance ranking to the above characteristics. Instead, we suggest that partitioning take into consideration the questions asked of the data, as well as the various attributes of a specific study area. For example, we include Rogen moraine as a factor in our GTZ analysis (Section 2.6*)* because we suspect ridge formation involves slow sluggish ice-flow, and that preservation of Rogen moraine is evidence of higher inheritance in those areas. Thus we invoke questions of paleo-ice-flow dynamics and subglacial landscape modification (especially through time). Other researchers may choose to ask questions regarding sediment thickness, landform generation, ice-sheet stagnation, etc; all of which are dependent on the area of study.

2.5 RESULTS

2.5.1 Landforms

Subglacial landforms, in a range of orientations, are present throughout most of the study area (Figure 2-3). Several regions outside of the field areas are depicted without mappable subglacial landforms, as none could be identified in these areas using remote-sensing imagery. The occurrence of small landforms cannot be ruled out completely until detailed air photo mapping is used to provide more detailed delineation at a sub-regional scale (<1:50 000). Subglacial landforms are rare southeast of Great Island, where the post-glacial Lake Agassiz and Tyrrell Sea incursion (marine limit on Figure 2-3) have obscured most of the pre-existing glacial landscape. Because this subglacial landscape is buried or modified beyond recognition, this area is excluded from the GTZ analysis.

2.5.1.1 STREAMLINED LANDFORMS

Streamlined landforms within the project area are oriented southeast, south, and southwest, with rare landforms oriented to the northeast around 97°W and 59°55'N (Figure 2-3). The mapped landforms include drumlins, crag-and-tails and other more elongated ridges with elongation ratios between 5:1 and 10:1. These elongate drumlinoid ridges do not fall within the mega-scale glacial lineation class (Clark, 1993) and are here referred to as megaflutings (c.f. Benn and Evans, 1998). The variously-oriented landforms appear to occur in restricted zones and few cross-cutting landforms were observed (Trommelen and Ross, 2010).

2.5.1.2 ROGEN MORAINES

Rogen moraines within the project area are oriented E-W and ENE-WSW (Figure 2-3). Rogen moraines are characterized by "anastomosing to curved ridges and intervening troughs, all lying transverse to former ice-flow direction" (Lundqvist, 1969, 1989), and often exhibit a gradual up-and/or down ice-flow direction transition to drumlins (Lundqvist, 1969; Boulton, 1987; Bouchard, 1989; Hättestrand and Kleman, 1999) and/or a non-transitional lateral shift to streamlined terrain (c.f. Aylsworth and Shilts, 1989). The region has both pristine and drumlinized Rogen moraine (Figure 2-3; Trommelen and Ross, 2010). The former are ridges characterized by relatively high amplitude and straight crests with a high density of boulders at the surface, whereas the latter are clearly drumlinized, less bouldery, and sometimes segmented, suggesting they have been overridden and modified by actively-flowing ice. The genesis of pristine Rogen moraine is still uncertain (Dunlop and Clark, 2006a; Clark, 2010) and undergoing further study in northeastern Manitoba. The generation of these landforms, however, is generally thought to have occurred under slow, sluggish ice-flow conditions (Aario, 1977; Boulton, 1987; Sollid and Sorbel, 1994; Hättestrand, 1997; Moller, 2006; Dunlop et al., 2008; Finlayson and Bradwell, 2008; Stokes et al., 2008; Clark, 2010; Knight, 2010).

2.5.1.3 ESKERS

Esker ridges within the project area are predominately long (~50-150 km) 'trunk' ridges which are often connected up-flow by smaller eskers (only mapped in field area). The trunk ridges are oriented south, south-southwest and southeast (Figure 2-3). Orientation of smaller ridges is, as expected, more variable, but it is generally oblique to their associated main trunk. For example, they tend to be oriented ESE or SSW if the main trunk is southeast, and SE or SW if the main trunk is south. The trend of trunk esker ridges is assumed to reflect the locations of R-channels in near-marginal zones (Boulton et al., 2007; Boulton et al., 2009), and the smaller eskers may be tributaries to larger channels. Traditionally, esker trend is important for ice-flow reconstruction as most trunk eskers are usually assumed to form parallel to ice-flow direction, concurrent to deglacial retreat (c.f. Kleman and Borgstrom, 1996). In northern Manitoba, however, the dendritic nature and regionally-consistent conduit-system (~700 km wide by at least 500 km long) of esker ridges (Trommelen and Ross, 2010) makes it hard to determine at a detailed scale which esker ridges are 'trunk' eskers (Boulton et al., 2009) that may actually relate to local deglaciation.

Ice-contact crevasse ridge networks are present in the northern half of the study area (Figure 2-3), that trend toward the southeast (150°). These networks consist of dense, rectilinear, hummocky to undulating pebbly sand ridges, photos of which can be found in Dredge and Nixon (1992) and Trommelen et al (2010). These rather odd landforms were visited at a few field sites and generally consist of fine to medium sand that is massive to horizontally-bedded.

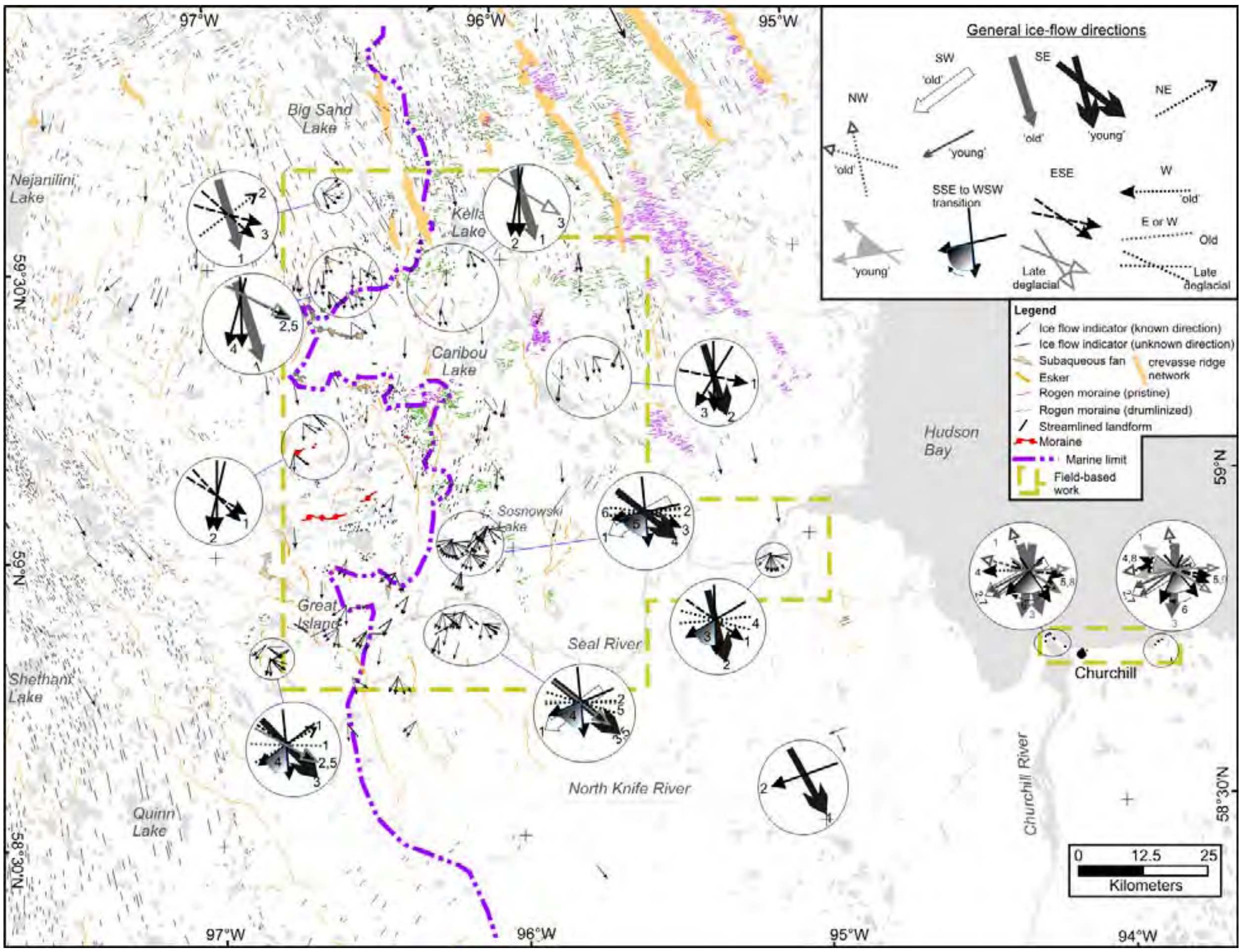


Figure 2-3. Subglacial landforms and field-based ice-flow indicators in the study and field areas. Large circles highlight the region-wide relationship between ice-flow indicators documented during field-mapping and the relative timing of variously-oriented ice-flow phases at each site. The general ice-flow directions box provides a summary of ice-flow orientations for the entire region. Marine limit is provided to illustrate that subglacial landforms may be present to the east of this limit, but progressively buried or re-entrained by marine sediment closer to Hudson Bay.

2.5.1.4 Late Deglacial Features

Late deglacial features are those that can be confidently correlated to the youngest glacial events on the landscape. These features can be used to help distinguish ice-marginal time-transgressive flowsets from isochronous flowsets generated further back from the margin. Deglacial features are relatively rare in the project area (Figure 2-4), and include young, weak striae situated on the tops of outcrops, end moraines, subaqueous fans deposited into the Tyrrell Sea, and De Geer moraines. The marine limit (165-180m) and associated beaches, and three radiocarbon ages (unpublished, Blake, 1982; Dredge and Nixon, 1992) are also depicted. The radiocarbon ages are well below marine limit (160-180 m asl) and are minimum ages for the regression of the Tyrrell Sea. While the large dendritic esker channel system is probably not related to the youngest deglacial event, smaller discontinuous esker ridges trend towards 110° in western Great Island and southwest of Caribou Lake. Features within the field area are taken from Trommelen and Campbell (2012a, b, c, d), and outside the field area from Dredge et al. (2007) and unpublished 2012 fieldwork.

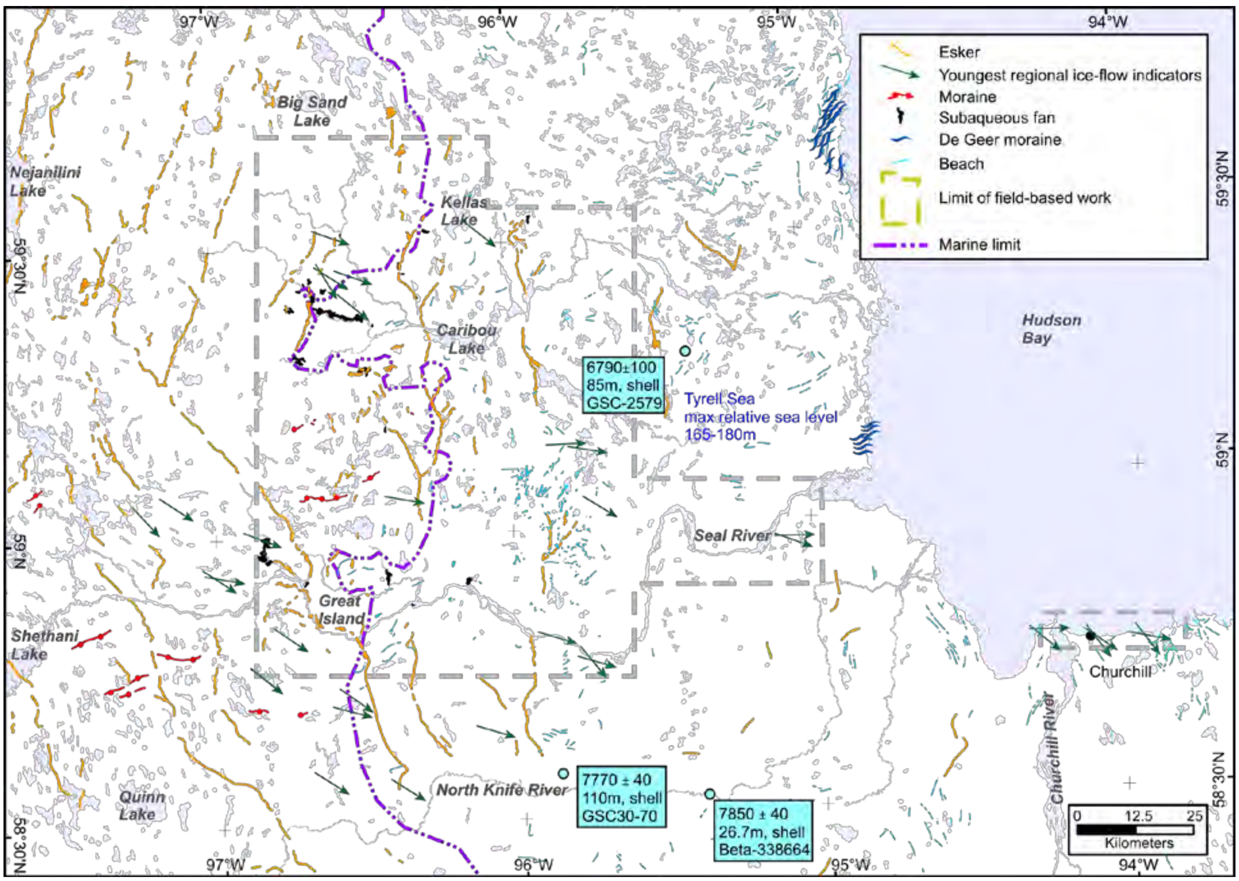


Figure 2-4. Late-deglacial features in northeastern Manitoba, including the youngest mapped striae across the region, end moraines, and features associated with the Holocene Tyrrell Sea (subaqueous fans, beach ridges and De Geer moraines). The Tyrrell Sea may have been coeval with Glacial Lake Agassiz at some point in time, but the paleohydrology is largely unknown, and no obvious lake drainage causeway exists in the region (Dredge, 1983). Features are taken from 1:50 000 scale mapping in the field area, and from 1:500 000 mapping outside of the field area. Eskers are shown, but the dendritic network does not correlate well with known retreat features. Only two radiocarbon ages exists for the area (Dredge and Nixon, 1992). *This figure has been updated since publication of Trommelen et al. (2012a), with the addition of a new unpublished radiocarbon date (Beta-338664), a few moraines and young striae.*

2.5.2 Field-based ice-flow indicators

Field-based ice-flow indicators (106 sites in the field area) are presented in Figure 2-3. Figure 2-5 shows examples of some of the ice-flow indicators encountered in the field. Large circles on Figure 2-3 highlight the relationship between indicators discovered during field-mapping and the relative timing of variously-oriented ice-flow phases at each site. Together, the field-based ice-flow indicators suggest that large-scale deviations of ice-sheet flow direction occurred through time. Localized complexity in the field-based ice-flow record may be explained by local effects such as ice flowing toward a local meltwater tunnel (Punkari, 1997). However, the regional analysis reveals consistent patterns that are best explained by regional-scale glacial dynamic shifts, especially when integrated with the flowset and GTZ analysis described below.

Ice has flowed over a span of 275°, and in the same direction multiple times (Figure 2-3). There is considerable west to east regional variability, as three phases of westward to northwestward ice flow were recorded only in the Churchill area, and one phase of northeastward ice flow was recorded only in the larger field area. Nonetheless, through extensive field study and careful documentation of crosscutting relationships, rarity, and location on individual outcrops, we tentatively subdivide the ice-flow orientations into old, middle and young ice-flow phases (Figure 2-6). Old ice-flow phases, defined mainly by the presence of heavily cross-striated, re-moulded macroforms, trend 230°-240° (Figure 2-5A,B), 160°-170° (Figure 2-5C), and 270°-330° (Churchill only). Northwest-trending ice-flow was the oldest documented at Churchill, while southwest-trending ice-flow was the oldest elsewhere. Rare 45°-60° trending striae are eroded onto 170° trending whalebacks (Figure 2-5C) near Big Sand Lake (Figure 2-3) and are possibly correlative with a well-defined 60° trending chattermark set (oldest at site; Figure 2-5D) southwest of Great Island. It is difficult to establish a regional relationship between NE and SE-trending systems, other than to note that there are both younger and older SE-trending ice-flow phases. Rare, protected, deep striae oriented 085°-265° and 90°-270° were also found at five sites in the Great Island area (Figure 2-3). Because these striae are rare at each of the five sites, and occur in protected locations, we suggest these E-W striae are the oldest at all five sites. Old west-trending grooves and striae also occur on multiple outcrops in the Churchill area (Figure 2-3).

Field-based ice-flow indicators of 'middle' age delimit the main ice-flow direction in the study area. Deep to fine striae, chattermarks, grooves and rare roches moutonnées, trending between 130° and 160° (Figure

2-5E), are widespread across the area covered by fieldwork, and crosscut several different orientations of ice-flow indicators. Next, at almost all sites, field-based ice-flow indicators trend between 170 and 230°. Generally, most field sites indicated a shift in ice-flow orientation over time, from southeast to southwest. Roches moutonnées and numerous smaller-scale ice-flow indicators such as deep to fine striae, grooves, chattermarks and crescentic gouges that trend 175°-210° are also widespread; along with common striae, chattermarks and grooves trending between 200° and 215°.

The younger field-based ice-flow indicators in the study area are striae, grooves, and chattermarks located on the top and some sides of outcrops. The more common 'young' field-based indicators trend between 230° and 260° (Figure 2-5G). This young southwestward ice-flow phase is evidenced in the Churchill area by partial 're-moulding' of some previously streamlined outcrops into small SW-trending roches moutonnées. At two sites, these west-southwesterly striae are cross-cut by fine to very fine striae oriented between 84-264° and 126-306° (Figure 2-5F). The youngest field-based ice-flow indicators in the study area are those clearly associated with the glaciomarine incursion. Southwest of Great Island, north of Caribou Lake and at Churchill, these young striae, grooves and gouges trend between 125° and 130° (Figure 2-5H). North of Caribou Lake, this east-southeast trending ice-flow set is correlative with subaqueous fan sedimentation (Figure 2-4) into the Tyrrell Sea.

2.5.2.1 FLOWSETS

Through careful analysis, we have grouped the streamlined landforms (Figure 2-3) into five distinct flowsets (Figure 2-7). These flowsets (A through E) trend to the NE, SE, S, SW and SE, respectively. Similarly oriented, but isolated, fragments we have chosen to make outliers of a larger flowset. Flowset C, in particular, consists solely of small disjoint patches. Some flowsets also partially or completely overlap (e.g. A and B). There is a scale effect in mapping flowsets in that depending on the size of the study area, not all minor variations can be mapped. Hence, due to a lack of detailed regional mapping, we have chosen to include several differently oriented streamlined landform sets that all roughly trend SE within

Figure 2-5. Some of the ice-flow indicators in northeastern Manitoba. A. Groove trending 244° is older than striae trending 182° that run through the groove. B. Three orientations of striae that occur on the top and lee of an older roche moutonnée. C. Whaleback rock drumlin in the northern study area indicating SSE flow. Younger NE-trending striae are inscribed on the top of these landforms. D. A metre-long chattermark set that trends 060°, in the southeastern study area. E. Rough, deep striae trending 130°, cross-cutting bedrock lineations. F. Deep striae trending 180° are cross-cut by finer striae trending 108° or 288°. G. Chattermark that trends 230°. H. Large bedrock outcrop moulded to trend 180° and overprinted on one side (and top) by younger striae towards 125°.


A
182°
2
244°
1
B
205°
3
1
138°
155°
2
C
170°
D
060°
E
130°
F
2
108-288°
180°
1
G
230°
H
180°
macroform
1
125°
striae
125°
striae
2

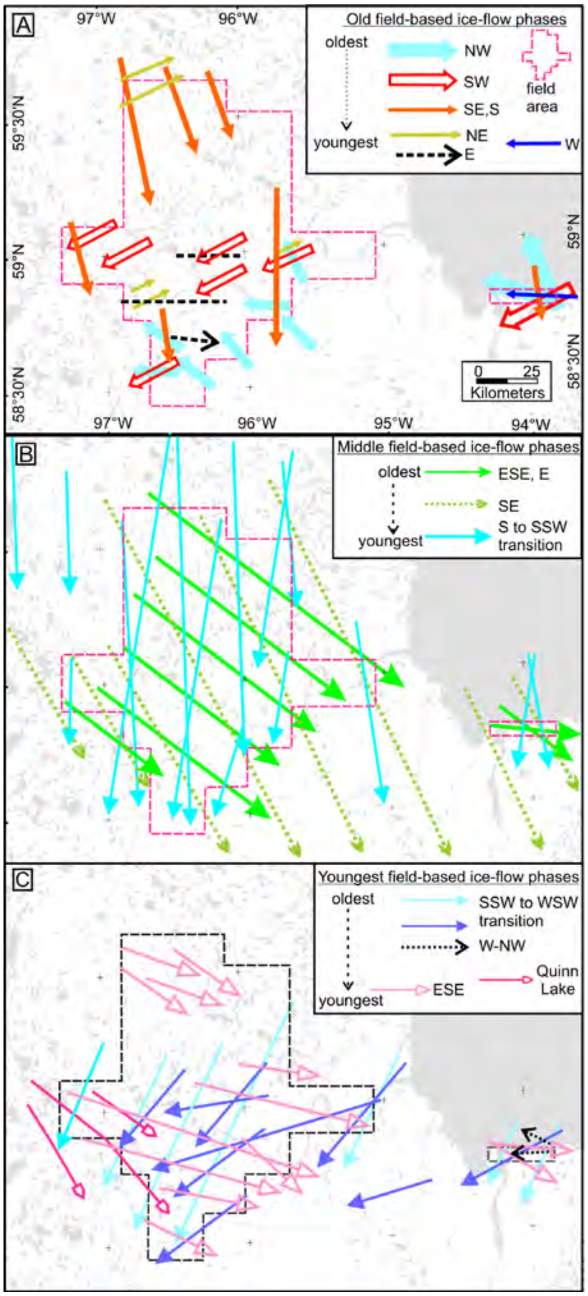


Figure 2-6. Field-based ice-flow indicator phases in northeastern Manitoba, defined based on striae, grooves, chattermarks, crescentic gouges, roches moutonnées and whaleback orientations. Age relationships (old, middle, young) were determined regionally based on integration of cross-cutting relationships at individual sites as well as knowledge of several old buried tills (Dredge and McMartin, 2011). The arrows outline the minimum spatial extent of each phase, where encountered in the field or displayed on regional maps. This figure has been updated from that presented in Trommelen et al. 2012, using unpublished 2012 field data.

one flowset (B). Rogen moraines, esker ridges and a few minor end moraines are shown on Figure 2-7 as part of the subglacial landform assemblage, but are not specifically associated with any one flowset. They are used in the GTZ analysis below to better understand the subglacial landscape as a whole.

In order to best reconstruct the paleoglaciological history of the area, it is advantageous to assign a relative age to each flowset. We can do this by analyzing cross-cutting relationships of streamlined landforms, but cross-cutting is rare in the project area (Trommelen and Ross, 2010). Instead, we attempt to correlate flowset orientation with the established relative-age field-based ice-flow indicator record in the field area (Figure 2-6). Establishing the exact chronological ordering of such a complex record is a challenging exercise and the outcome uncertain. Nonetheless, some flowsets are clearly younger than others. Some direct correlations can be made (e.g. isochronous flowset A is parallel to, and thus coeval with the northeast-trending field-based set; time-transgressive flowset E is parallel to and thus coeval with the late east-southeast-trending field-based set), but the analysis of field-based indicators also suggests that some ice-flow directions occurred multiple times. Hence some regional flowsets may be a composite record of two or more events. For example, southeast-trending flowset B could be correlated to either the younger or older southeast to south-southeast-trending ice-flow phases, or both.

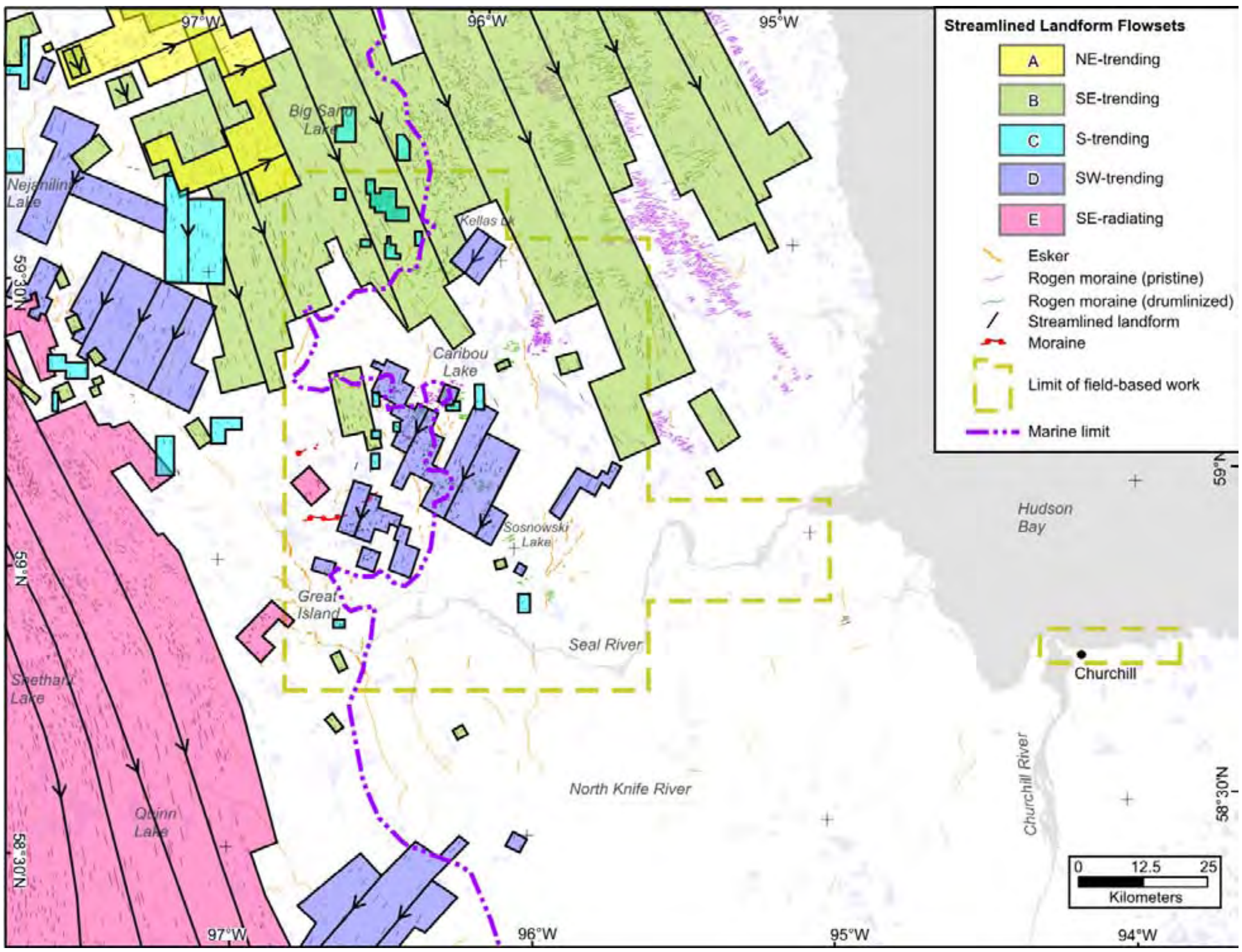


Figure 2-7. Streamlined landform flowsets in northeastern Manitoba. Landforms were mapped using remote sensing data (Trommelen and Ross 2010) and supplemented by aerial photograph mapping within the main field area. Flowsets are lettered (A through E) according to a possible relative-age chronology, as determined by cross-cutting relationships and similar orientation to field-based ice-flow indicator sets. Original landform data is shown as the background layer.

2.6 GLACIAL TERRAIN ZONES IN NORTHEAST MANITOBA

To gain further insights into the paleoglaciological evolution of this previously glaciated region, we apply our GTZ approach. Using criteria outlined in the methodology, as depicted in Figure 2-8, we delimit five discrete GTZ with internally common characteristics that together define a unique glacial history. GTZ boundaries are better defined where detailed (<100 000 scale) surficial maps, including ice-flow indicators, were available. The regions overlain or significantly reworked by postglacial water (Glacial Lake Agassiz or the Tyrrell Sea) were not assigned a GTZ as there are only rare subglacial landforms at the surface.

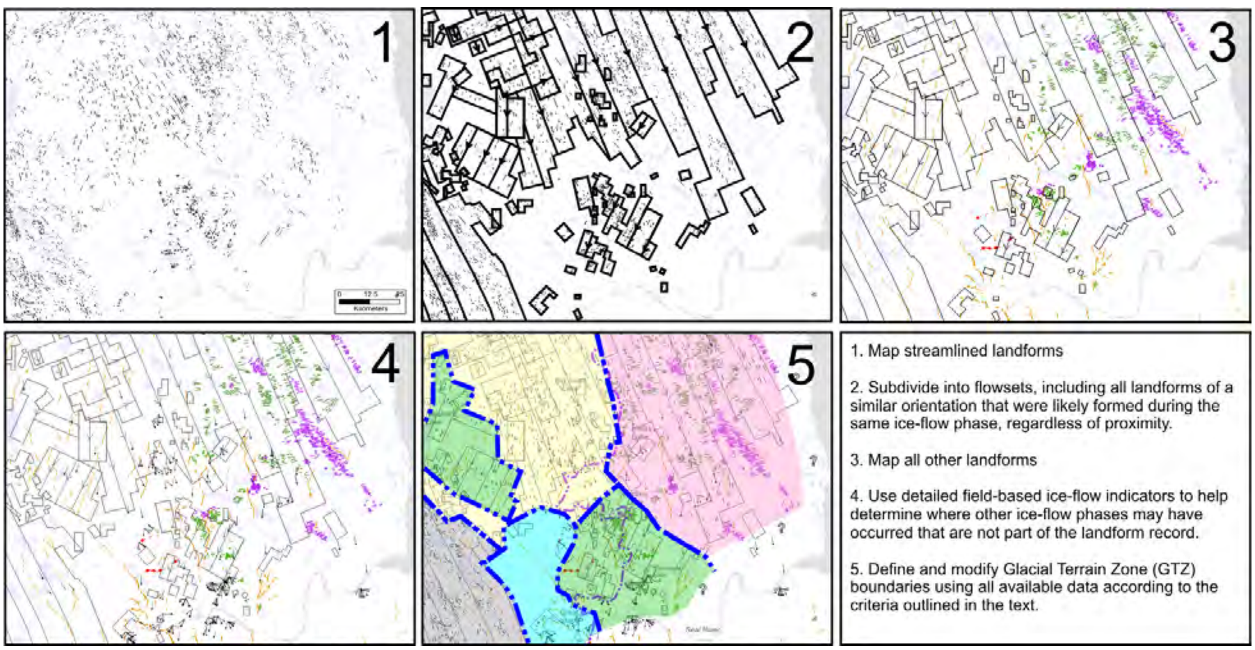


Figure 2-8. Glacial Terrain Zone (GTZ) methodology for the study area. GTZs are generated by incorporating all available data, including streamlined landform orientation (summarized by flowsets), other subglacial geomorphology (rogen moraine, esker, end moraines) and field-based ice-flow indicators.

2.6.1 Caribou River GTZ

The Caribou River GTZ (Figure 2-9), in the northeast-most corner of Manitoba, was delimited based on swaths of southeast-trending streamlined landforms (flowset B) that alternate with abundant Rogen moraine swaths. Flowset B extends beyond the Caribou GTZ but it is the lack of overprinting, association with Rogen moraine, and the presence of extensive crevasse ridge networks Figure 2-10A), that defines the Caribou GTZ. Field-based ice-flow indicators mapped in the southwest part of this GTZ (n=9 sites) trend towards 100° (oldest phase in this GTZ), 160°, 180°, 205° and a late, weak 125°. Flowset B within the Caribou River GTZ trends between 140° and 160°, which may correlate to either the older or younger (or both) regional southeast-trending ice-flow phases (Figure 2-6). The limited fieldwork in this region does not resolve this. Further north in Nunavut, the southeast-trending Rogen-streamlined landform swaths are overprinted by east-trending streamlined landforms (De Angelis, 2007), indicating that flowset B is not the youngest regional flowset. Ribbed moraine that may have formed near the ice margin are associated with hummocky moraine or De Geer moraine, unlike those of the CR GTZ, It is generally accepted that Rogen moraine are subglacial landforms, and newer research suggests that formation most likely occurred further back beneath a thicker ice sheet (Moller, 2006; Dunlop et al., 2008; Stokes et al., 2008; Finlayson et al., 2010; Knight, 2010) – not at the ice margin.

Eskers were only mapped in the southern half of the GTZ, and trend 180°, 200° and 140°. Where present, esker ridges obliquely crosscuts flowset B. The marine limit for the entire project area is around 165-180 m asl. Both the subaqueous fan west of Caribou Lake and De Geer moraines on the shores of Hudson Bay (Figure 2-4) suggest the ice margin retreated to the west in this area. Thus while at one time it was thought that the streamlined landforms within this GTZ (flowset B) were generated during the youngest retreat-phase (Dyke and Prest, 1987a, b), new mapping conclusively shows that flowset B is *not* part of an ice-retreat landsystem.

2.6.2 Big Sand Lake GTZ

The Big Sand Lake GTZ (Figure 2-9), in the upper northwest of the study area, is different from the Caribou Lake GTZ based on two criteria: 1) it lacks Rogen moraine ridges and 2) there is more landscape inheritance, evidenced by overlapping or superimposed streamlined landform flowsets (A, B and C) (Figure 2-10B). Fieldwork was limited to the southeast side of this GTZ, and at least six distinct ice-flow phases have been identified on the basis of well-preserved field indicators (above marine limit). The main field-based ice-flow indicators trend towards 160° (old and young phases), 45-60°, 125°, 180°, and minor more variable and younger indicators ranging 110-135°. Based on the cross-cutting relationship between megaflutes near Big Sand Lake (Trommelen and Ross, 2010, Figure 3), flowset A is older than flowset B. However, the field-based ice-flow indicator record is more complex (Figure 2-6), with both older and younger SE-trending ice-flow phases. This complexity demonstrates both overprinting and inheritance within this GTZ. Flowset A is reported for the first time in the area and extends 37 km into southern Nunavut (Aylsworth et al., 1990). South-trending flutes or megaflutes (flowset C) are present as intact outliers. There is currently only one southward-trending ice-flow phase documented in the field-based ice-flow indicator record (Figure 2-6) and we correlate flowset C with this southerly ice-flow. This

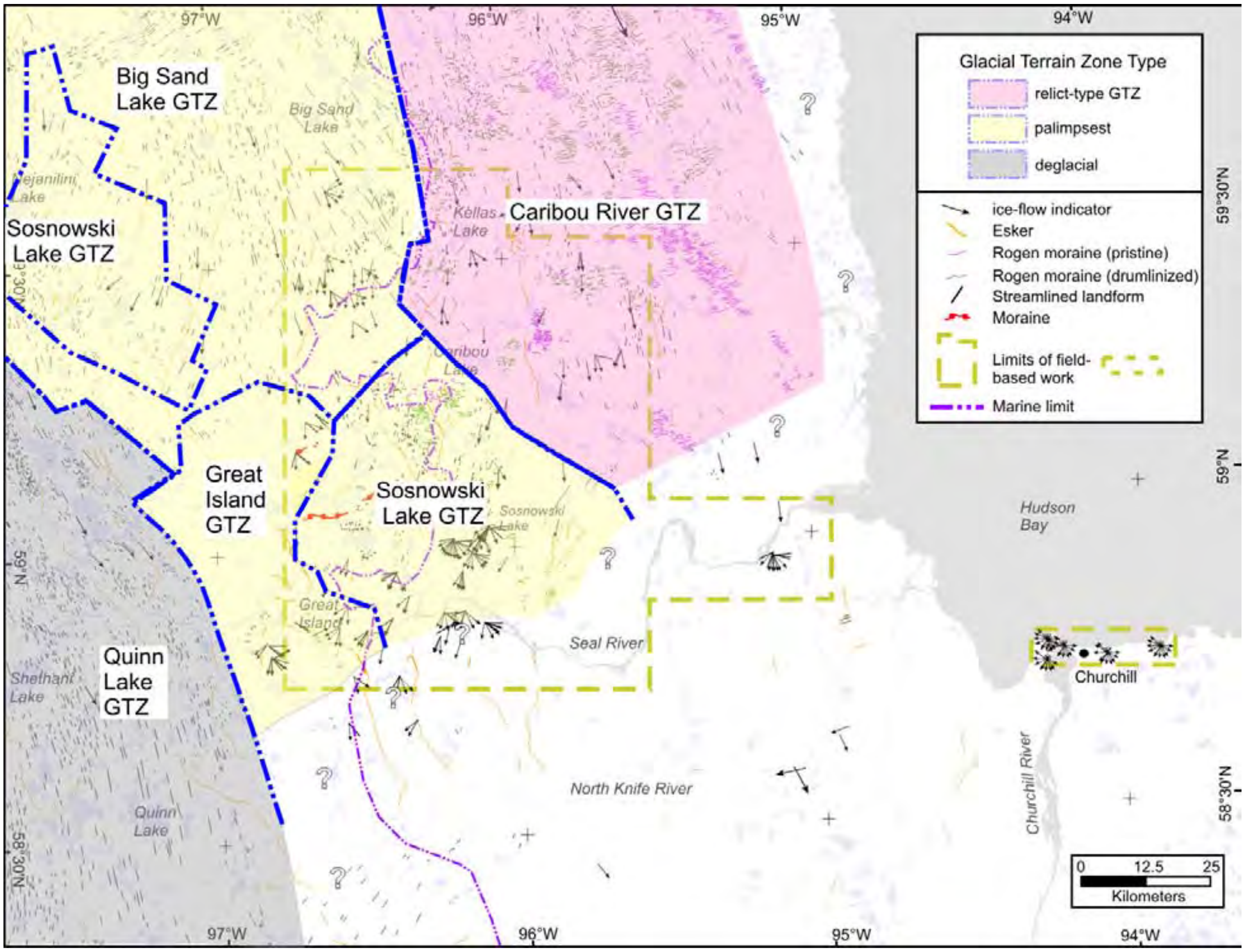


Figure 2-9. The five established Glacial Terrain Zones (GTZ) in northeastern Manitoba, classified by paleoglaciological type. Each of these GTZ delimits a portion of the subglacial landscape with internally consistent glacial histories and different levels of inheritance and overprinting.

correlation is tentative, as older ice-flow phases to the SE and SW are present and it is likely that ice has flowed southward multiple times. Even so, based on field-based ice-flow indicator relationships, flowset C is likely younger than flowset B. Minimal overprinting during the younger (S and SE) ice-flow phases has resulted in high inheritance of older streamlined landforms and this is a defining characteristic of the Big Sand Lake GTZ.

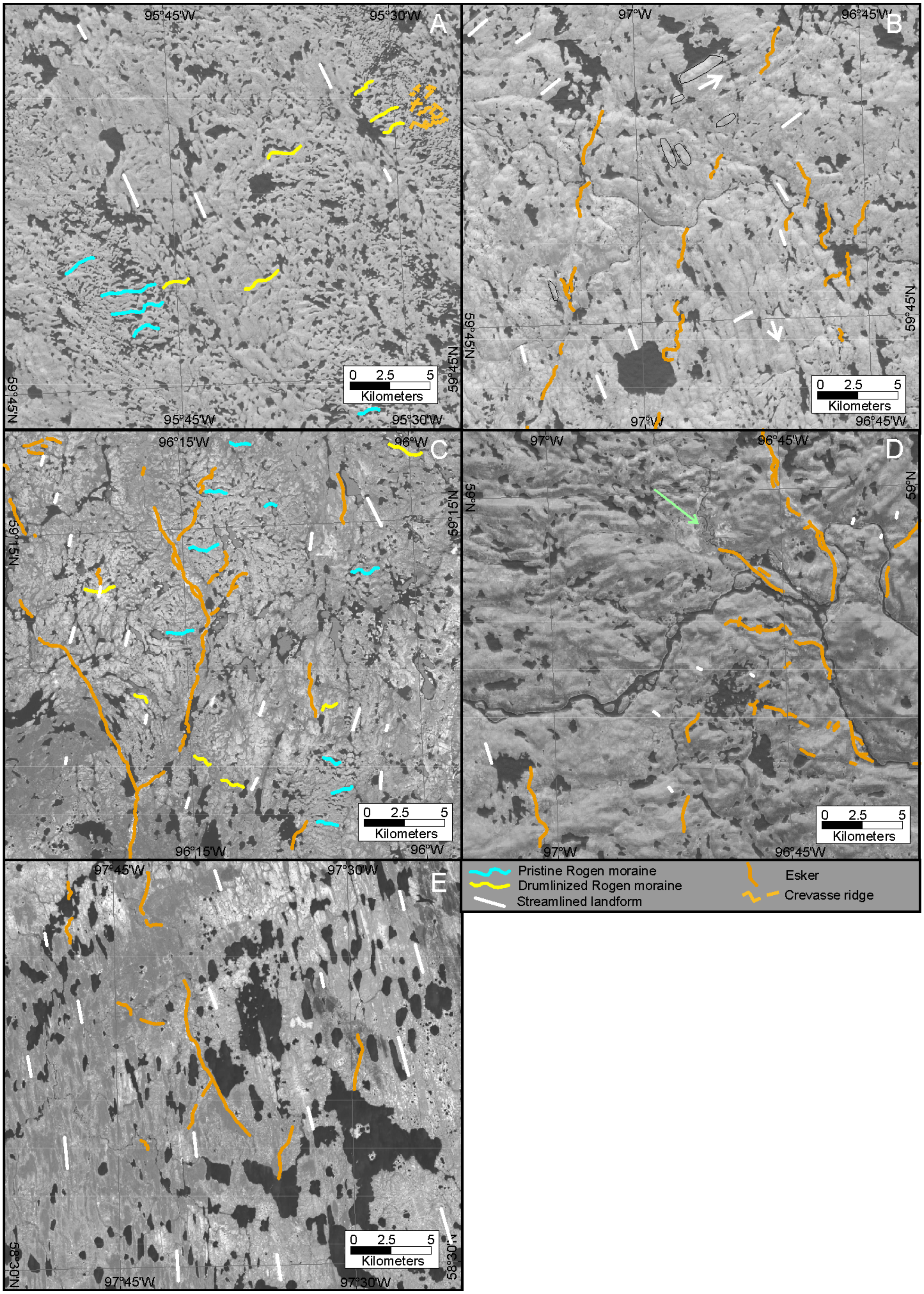
Eskers are mapped throughout most of this GTZ, and trend 180°, 200° and 140°. These esker orientations are consistent regionally, and are oblique to, and crosscut, all subglacial streamlined landforms in the

GTZ. This relationship suggests that the streamlined landforms in that GTZ, like the ones in the Caribou GTZ, formed prior to the development of the regionally extensive channelized meltwater system. Weak striae situated at the tops of outcrops in the field portion of this GTZ trend to the ESE (Figure 2-6), and are correlated with the deposition of subaqueous fan sediments into the Tyrrell Sea near the marine limit (Figure 2-4). The oblique orientation of all flowsets in the Big Sand GTZ to late-deglacial features shows that the subglacial landscape within this GTZ is also not part of an ice-retreat landsystem.

2.6.3 Sosnowski Lake GTZ

The Sosnowski Lake GTZ (Figure 2-9), in the middle of the study area, is defined by: 1) SW-trending flowset D, 2) variably drumlinized Rogen moraine, and 3) a complex field-based ice-flow record (85° to 270° range). The largest streamlined landform flowset D in this GTZ trends 190-200°, making this a unique GTZ where the most ubiquitous and well-preserved landforms are indicative of southwest-trending ice-flow dynamics. However, the flowset is weak and consists of patchy flutes and drumlins with low elongation ratios (1:1 to 1:3), associated with variably drumlinized fields of Rogen moraine (Figure 2-10C). Numerous small streamlined landform outlier areas were denoted, using landforms mapped from air photos, that trend to 160° and 180° (lumped with flowsets B and C, respectively). Flowset D is parallel to the common SSW-trending field-based ice-flow indicators in the Sosnowski Lake area. Numerous field sites document a transition from southeast to south-southwest-trending ice flow (Figure 2-3) and thus we suggest flowset D is part of this younger south-southwest-trending ice-flow phase. Southeast-trending field-based ice-flow indicators are also present throughout the Sosnowksi Lake GTZ (Figure 2-3), are generally rare and found in protected areas of outcrops, and are tentatively correlated with flowset B. Because of the well-preserved common field-based ice-flow indicator

Figure 2-10. Examples of the subglacial landscape in each GTZ, as seen on satellite imagery (Landsat EMT+ and SPOT4 (C, E)), at the same scale. Representative streamlined landforms and Rogen moraine (and all eskers) are outlined according to the legend. See Figure 2-1 for locations. A) Caribou River GTZ, showing common SE-trending swaths of Rogen moraine, streamlined landforms and crevasse ridges. Drumlinized Rogen moraine is present within some of the streamlined landform swaths, indicating that drumlinization is a secondary process. B) Big Sand Lake GTZ, showing streamlined landforms trending NE, interspersed with streamlined landforms that trend SE. Eskers clearly cross-cut both flowsets. C) Sosnowski Lake GTZ, showing the weak SW-trending streamlined landforms that variably overprint Rogen moraine ridges. Eskers clearly cross-cut all other subglacial landforms. D) Great Island GTZ, showing the rough topography and the ‘nonglacial’ appearance of the landscape. The main esker system converges into a topographically-confined valley where the Seal River now flows. The few streamlined landforms present are bedrock-cored. A late-deglacial subaqueous fan (green arrow) is also situated within this GTZ, and likely formed at the ice-margin into Glacial Lake Agassiz. E) Quinn Lake GTZ, showing the dense spacing of drumlins within this GTZ. An esker trunk and its tributaries crosscuts the streamlined landforms.


A
95°45'W
95°30'W
59°45'N
B
97°W
96°45'W
59°45'N
C
96°15'W
96°W
59°15'N
D
97°W
96°45'W
59°N
E
97°45'W
97°30'W
58°30'N
0 2.5 5
Kilometers
Pristine Rogen moraine
Drumlinized Rogen moraine
Streamlined landform
Esker
Crevasse ridge

transitions from southeast to south-southwest-trending, we suggest that the south-southwest-trending flowset D is also younger than the southeast -trending flowset B. In contrast to the Big Sand Lake GTZ, the younger (south-southwestward) ice flow has resulted in significant overprinting, with only small relict (high inheritance) areas of older streamlined landforms (southeast and south-trending). The presence of a very complex field-based ice-flow indicator record, however, indicates high inheritance and may signify that the south-southwest-trending ice-flow phase was weak.

The main esker ridges trend towards 180° and 190°, with tributary eskers that trend to 140° and 200°. Eskers are sometimes situated within larger meltwater corridors, and both are regionally oblique to or cross-cut, all streamlined landforms Figure 2-10C). This is important because it may indicate that despite the younger age of flowset D relative to flowset B and C, the SW-trending glacial dynamics also does not reflect late deglacial flow contemporaneous to esker formation. No clearly defined late-deglacial features (Figure 2-4) trend SW either.

2.6.4 Great Island GTZ

In contrast to the other GTZ, the Great Island GTZ (Figure 2-9) is characterized by: 1) rare subglacial landforms, 2) thin discontinuous till cover that leads to a rough "bedrock-controlled' topography (Figure 2-10D), with hummocky till interspersed with glaciofluvial sediment in the topographic low and 3) a complex field-based ice-flow indicator record. The few streamlined landforms present are small (seen only on air photos), and are mostly bedrock-cored. These landforms trend 125-155°, and 180-190°, and are mapped as outliers of flowsets E and C, respectively. Fieldwork within this GTZ has identified a complex pattern of ice-flow indicators (range 60° to 230°). These include strong, common field-based ice-flow indicators (S to SW) as well as rarer protected east or west, east-southeast and southeast indicators and the only northeast-trending indicators (Figure 2-5D) in the southern half of the field area. Similar to the Sosnowski Lake GTZ, small relict outlier flowsets and a very complex ice-flow indicator record suggest high inheritance and incomplete overprinting. A lack of thick till or subglacial landforms also suggests this area may have been subject to different subglacial dynamics than the adjacent GTZs.

One trunk esker trends southeast across this GTZ (Figure 2-10D). This esker is parallel to the Quinn Lake landsystem. Small discontinuous eskers that trend 110°-120° are also present and may have formed during the late deglaciation (Figure 2-4).

2.6.5 Quinn Lake GTZ

The Quinn Lake GTZ, in the southwest corner of the study area, is delimited based on the presence of one internally consistent curvilinear streamlined landform flowset (E) that trends between 140° and 180° (Figure 2-9). Flowset E is actually part of a larger radiating streamlined landform flowset that extends 45 km west to Shethani Lake (Figure 2-1) and south 25 km past North Knife Lake (Trommelen and Ross, 2010). Flowset E contains a high-density suite of streamlined landforms (Figure 2-10E) that includes parabolic drumlins and spindle drumlins. This GTZ also lacks Rogen moraines. The remote-sensing analysis hence suggests strong landscape-scale overprinting. However, field-based ice-flow indicator data (Dredge and Nixon, 1982a) is sparse in this GTZ and thus the complexity of this record is unknown. Detailed sediment composition mapping would also help to determine the intensity and spatial extent of overprinting.

The trunk eskers trend sub-parallel the lobate pattern of flowset E, suggesting these eskers were actively forming during generation of flowset E. There are also smaller, discontinuous esker ridges that trend E, SE and SW, which may be tributary eskers or a younger deglacial imprint.

The Quinn Lake GTZ can be mapped far outside the study area, for a total of 14,100 km^2. This regional GTZ contains drumlins and drumlinoid ridges (Dredge and Nixon, 1982a; Dredge et al., 1982b; Trommelen and Ross, 2010) with elongation ratios less than 6:1, that terminate in Glacial Lake Agassiz and are overlain by DeGeer moraines (Dredge et al., 1986). This GTZ may be an ice-marginal terrestrial landsystem (landsystem B, Colgan et al., 2003).

2.7 Discussion and Implications

GTZ analysis was developed during this project as a tool for paleoglaciological reconstruction to help 'untangle' the complex, and often contradictory, data mapped in the study area. This approach does not replace flowset analysis, but together with flowset analysis helps to answer questions such as temporal relationships, degree of inheritance and/or overprinting, and continuity of ice-flow regimes within a specific area. For example, in Figure 2-2, GTZ analysis suggests that the forth southwesterly phase of ice-flow must have affected GTZ B, even though this phase was not documented in the field-based or subglacial landform records. If GTZ B was a large area, and encompassed the entire project area, this may be a revelation. Questions are also raised about subglacial ice dynamics in the area, including how GTZ B was preserved as relict terrain during the youngest known deglacial ice-flow phase. Additionally, in

Figure 2-2 there are only patch cross-cutting relationships between landforms in one spot. GTZ analysis helps to confirm the likely time-dependant regional ice-flow history by incorporating field-based ice-flow indicators.

GTZ analysis represents the first step towards piecing together spatio-temporal subglacial landscape evolution, rather than simply denoting end-result (i.e. warm-based, cold-based, mixed). This technique may be particularly useful in regions of complex paleo-subglacial evolution where there is no clear division between end-member criteria. For example, lateral sliding (Dyke, 2008) or regolith/till (Kleman and Borgström, 1994) boundaries provide obvious evidence for a change from cold-based to warm-based faster-flowing ice. But in other areas, perhaps within a frozen-thawed mosaic (Kleman et al., 1999; Kleman and Glasser, 2007), spatio-temporal switches in basal thermal regime may be harder to distinguish. With local spatio-temporal variation, it is possible that a single flowset may cross GTZ boundaries (Figure 2-2), as it is the cumulative combination of characteristics that allows for zone partitioning. There is an obvious scale effect, and different information could be learnt from regional, local and even property-scale projects. The following is a discussion of GTZ analysis in northeast Manitoba, as it applies to paleoglaciology and regional ice-flow history.

2.7.1 Glacial Terrain Zones (GTZ) and Paleoglaciology

A significant outcome of the GTZ analysis is the recognition that four of the five GTZ contain subglacial landscapes with significant inheritance and varied levels of overprinting. Only the Quinn Lake GTZ is best explained by strong overprinting during a late-deglacial ice-flow event. In order to expand upon these interpretations, we have classified the regional GTZ according to their relative degree of inheritance vs overprinting as: (1) relict, (2) palimpsest, and (3) deglacial. These categories are quite broad and further subdivision is to be expected with continued application of the GTZ approach to different study areas and subglacial histories.

2.7.1.1 Relict, warm-based glacial

Relict terrain is commonly defined as a region where the ground surface and landforms (tors, periglacial sorted circles, boulder fields, etc) are essentially unmodified by the last ice sheet (Dyke and Morris, 1988; Kleman and Borgström, 1994; Kleman and Hättestrand, 1999; Hättestrand and Stroeven, 2002). Herein, we consider preservation of old glacial features to be another type of relict terrain (relict landscapes, Kleman and Hättestrand, 1999). In order to preserve older features, subglacially-active (warm-based, and with high effective pressure (e.g. Denis et al., 2009) ice that created the subglacial landscape, later transitioned to a subglacially-inactive cold-based (e.g. Boulton et al., 2001b) or stiff, dewatered (e.g. Meriano and Eyles, 2009) environment. This high inheritance glacial landscape generally consists of one or more isochronous flowsets whose orientation best matches early, older, ice-flow orientations, with little or no evidence of overprinting. Eskers, if present, obliquely crosscut flowsets and are clearly associated with an unrelated later-stage ice-flow phase or stagnant conditions. A range of ice-flow indicators may be present, but their orientations are oblique to regionally young indicators. Relict glacial landscapes have been recognized in glaciated terrains chiefly in the uplands of central Fennoscandia (Kleman et al., 1994; Hättestrand and Stroeven, 2002; Fabel et al., 2006; Goodfellow et al., 2008). Other examples of Canadian relict warm-based glacial terrain include intersecting drumlin fields on Prince of Wales island (Dyke and Morris, 1988), subglacial assemblages of the Maskwa corridor in Saskatchewan (Ross et al., 2009) and unidirectional dispersal trains associated with the core region of the Scotian Ice Divide in Nova Scotia (Stea and Finck, 2001).

The Caribou Lake GTZ, in this study area, is an example of relict warm-based glacial terrain (Figure 2-9). A single regionally extensive landscape-scale flow phase (SE-trending flowset B) epitomizes this GTZ, which is crosscut by trunk eskers and overprinted only locally by younger field-based ice-flow indicator features. The lack of landform overprinting by younger ice-flow phases (except in Nunavut to the north) suggests the Caribou Lake GTZ contains a relict landsystem that somehow survived the later events. These same SE-trending streamlined landforms are present in both adjacent GTZ, but are more strongly overprinted. The regional relative-age chronology established using field-based ice-flow indicators (Figure 2-6), combined with known late-deglacial features (Figure 2-4), suggests that after southeasterly-flowing ice generated streamlined landforms within this GTZ, ice flowed to the south, southwest and west-southwest before finally flowing east-southeast into the Tyrrell Sea at the marine limit. Rogen moraine is also common in this GTZ. While the interpretation of Rogen (ribbed) moraine ridges remains contentious, these landforms are often associated with slow, sluggish ice-flow. In summary, we suggest this relict-type GTZ likely contains a high level of inheritance from the early SE-trending event, which should be reflected in the surface till composition. Deglaciation of this area likely occurred under sticky area of stiff dewatered till (Christoffersen and Tulaczyk, 2003; Piotrowski et al., 2004), or frozen-bed conditions, as indicated by the general absence of eskers parallel to flowset B, lack of overprinting and preservation of Rogen moraine.

2.7.1.2 Palimpsest

Palimpsest landscapes are common near the core regions of ice sheets where subglacially active warm-based ice experienced a shift in flow-direction over time that resulted in the incomplete re-moulding of surface landforms such as drumlins and Rogen (ribbed) moraine (Stea and Brown, 1989; Kleman, 1992; Stea, 1994; Stea and Pe-Pier, 1999; Clark and Meehan, 2001; Livingstone et al., 2008; Finlayson et al., 2010; Knight, 2010; Plouffe et al., 2011b). This incomplete overprinting forms a distinct GTZ type characterized by the presence of two or more distinct isochronous flowsets with cross-cutting relationships or a mosaic-like fragmented imprint on the landscape. Eskers, if present, were formed during deglaciation and will be oriented oblique to the overprinted ice flowsets in the region.

In northeastern Manitoba, the Big Sand Lake, Sosnowski Lake and Great Island GTZs (Figure 2-9) are all different, but have strong palimpsest records with different levels of inheritance and overprinting. A complex field-based ice-flow record indicative of a long glacial history is a defining characteristic of all three GTZs. Sediment may be thin or thick, but young ice flow phases seem to have had limited erosional effect on the landscape. These GTZs also contain small isolated fragments of different flowsets, and cross-cutting streamlined landforms are preserved in parts of the Big Sand Lake GTZ. At a landform-field scale, the isolated flowset fragments could also be considered relict-type GTZ, but at a regional-scale, we have included these as fragments within the larger palimpsest-type GTZ. Esker ridges within these GTZ trend SE, S and SW, though a few minor ridges also trend ESE. The S-trending esker ridges are oblique to, and cross-cut, all flowsets. The SW-trending esker ridges are sub-parallel to flowset D, and appear to converge with SE-trending esker ridges (sub-parallel to flowset E). This convergence may indicate that the regionally-extensive dendritic esker network was present far back from the ice sheet margin during deglaciation (Boulton et al., 2009), and that the esker trends do not correlate well to deglacial ice-flow orientations (Figure 2-4). The minor ESE-trending esker ridges, contrastingly, are likely late-deglacial.

The regional relative-age chronology established using field-based ice-flow indicators (Figure 2-6), combined with cross-cutting relationships and mapped late-deglacial features (Figure 2-4), suggests that the Big Sand Lake GTZ is a region of high inheritance and landscape-elements formed during the oldest landscape-forming events (flowset A and B). The Sosnowski Lake GTZ, contrastingly, appears to be a region of lower relative inheritance with overprinting during a younger SW-trending ice-flow phase. The complex field-based ice-flow indicator record, however, suggests the SW-trending ice-flow phase may not have been very intensive or long-lived. The age of the Great Island GTZ is uncertain, though fragments of flowsets B, C and E are present. NE-trending field-based ice-flow indicators are also present, which may be coeval to flowset A. If so, the composite-age would be similar to the Big Sand Lake GTZ. Because the orientation of flowset D is oblique to the orientation of late-deglacial features in the region (Figure 2-4), all three palimpsest-type GTZs are considered older than the final stage of ice retreat. Again, sticky conditions (due to stiff, dewatered till or frozen-bed conditions) are required to preserve the older landscape fragments of these GTZ.

We expect the till within the palimpsest-type GTZs should consist of a mix of inherited and overprinted characteristics.

2.7.1.3 Deglacial, warm-based

Deglacial GTZs can include any number of different deglacial elements or ice-marginal landsystems (Evans, 2003b), including time-transgressive flowsets ('warm-based deglacial flowset' (Stokes et al., 2009) or 'time-transgressive retreating margin flowset' (Greenwood and Clark, 2009b)) or mega-scale glacial lineations associated with push, thrust or overridden moraines (Evans et al., 2008, ice-marginal assemblages such as ribbed and De Geer moraine), surge-type drumlins and end-moraines (Waller et al., 2008; Johnson et al., 2010), controlled moraines (Evans, 2009) and other geomorphic associations that may include hummocky moraine, and glaciofluvial, glaciolacustrine or glaciomarine sediments.

In northeastern Manitoba, the Quinn Lake GTZ is classified as a deglacial-type GTZ (Figure 2-9). This GTZ consists of one isochronous well-defined flowset (E), associated with subparallel esker ridges, that was likely formed during a late-stage ice-flow event, possibly into Glacial Lake Agassiz (Dredge and Nixon, 1992). Till within this GTZ is likely related to the source area of the surge, with little inherited characteristics.

2.8 Key insights for regional paleoglaciological reconstruction

The landscape analysis described above has identified a number of terrain types each recording different or partly overlapping portions of the regional glacial landscape, which provides clues or insights to the regional glacial history. GTZ analysis can also help identify knowledge gaps or problems with previously published reconstructions.

When combining the landform flowset and field-based ice-flow indicator phases, it appears that several ice-flow directions, especially between SE and SW were widespread across the entire area at multiple times. Older SW and SE ice-flows could be correlative to similarly-oriented striae and till fabrics associated with the Amery (SE, S, SW-W trends; Illinoian) and Sundance (SW, SE trends; Pre-Illinoian) tills along the Nelson, Hayes, Gods and Pennycutaway rivers in northeast-central Manitoba (Nielsen et al., 1986; Dredge et al., 1990; Dredge and Nixon, 1992; Nielsen, 2001, 2002a). Southeast to southwest - trending ice-flow phases are thought to be regionally extensive, as they are documented in numerous regional studies from the surrounding areas in Manitoba, northeast Saskatchewan and the Keewatin area of Nunavut (Dredge et al., 1985; Klassen and Netterville, 1985; Dredge and Nixon, 1986; Campbell, 2001, 2002; McMartin and Henderson, 2004; Smith, 2006; Smith and Kaczowka, 2007; Kaszycki et al., 2008; Avery, 2010; McMartin et al., 2010; Trommelen, 2011b). Therefore, and despite the complexity of the landscape as described in this study, it is possible to identify coherent features that are likely to be the product of a limited number of regional ice-flow phases. For example, it is less complicated to suggest a regionally extensive southeast to southwest ice-flow shift to explain our observations in the Great Island and Caribou Lake GTZ, rather than to present the ice-flow history as a series of coeval, convergent yet migrating southeast and southwest-trending ice-flow phases (Kaszycki et al., 2008). This time-transgressive southeast to southwest rotation of ice-flow orientation could be related to the previously proposed Late Wisconsinan migration of the KID eastward from the Northwest Territories and into Nunavut north of the study area (Dyke and Prest, 1987a; Klassen, 1995; McMartin and Henderson, 2004). The GTZ analysis also leads us to suggest that the common SE-trending subglacial landforms in northeastern-most Manitoba were generated earlier in the Late Wisconsinan and preserved as a relict glacial terrain throughout deglaciation, instead of reflecting late deglacial ice dynamics as suggested in the literature (Dyke and Prest, 1987a). These landforms may be a part of flow stage B (early or Pre-Wisconsinan), as presented in Boulton and Clark (1990), though the scale of their figures makes this correlation uncertain.

Several new ice flow directions are recognized in the striation record, especially in the Great Island GTZ, which was the focus of the field-based work. Our mapping effort shows a regionally consistent southwest (e.g. flowset D) to west-southwest ice-flow phase that was not previously recognized as an important phase in the region and does not appear in regional reconstructions. The exact significance and timing of this phase is unclear but it is tempting to relate it to an ice ridge (C1 of Boulton and Clark, 1990?; or as in Tarasov and Peltier, 2004) in southern Hudson Bay. Thick ice is indeed necessary in that area until late into deglaciation, in order to keep Lake Agassiz from draining into the rapidly deglaciating Hudson Bay before 8.4 cal ka BP (Dyke et al., 2003; Tarasov and Peltier, 2004). Flowset D may thus be important geomorphological evidence of the existence of such a thick ice ridge in southern Hudson Bay after the Last Glacial Maximum.

Newly identified old NE and E trending ice-flow phases will also need to be taken into account in regional ice-sheet reconstructions (flow stage E? from Boulton and Clark, 1990). We note that no evidence was found in this study of the 'Churchill Swarm' reported by Kleman et al. (2010), suggesting it might have been mis-located in their paper. The closest flowset that could potentially match the Churchill Swarm orientation would be our flowset A. However, flowset A clearly trends to the NE, and not the SW as proposed for the Churchill Swarm.

Overall, this outer fringe area of the Keewatin Sector of the LIS appears to have experienced major changes in subglacial conditions. The level of landscape complexity is arguably comparable to areas that have undergone rapid spatio-temporal shifts of highly contrasting conditions such as those that evolved from ice streams to inter-ice stream conditions (Ross et al., 2009; Stokes et al., 2009; O Cofaigh et al., 2010). In the case of northeast Manitoba, however, this complexity likely took more time to develop. With further detailed mapping, we suspect more relict and palimpsest GTZ will be documented. Regions with slow, sluggish and/or cold-based ice patches need to be better documented spatially and through time to fully understand the effect of LIS evolution on the glacial landscape, as well as provide a better understanding of subglacial hydrology and rheology. By careful mapping of GTZ using multiple data types, compilation of this regional data will allow for more accurate paleoglaciological reconstruction models throughout the Quaternary. Detailed sediment dispersal studies are also needed to further assess the intensity of ice-flow phases to determine which ice-flow phases can be correlated with sediment erosion, transportation and deposition, and which ice-flow phases may simply have overprinted the surface of the glacial landscape.

2.9 SUBGLACIAL LANDSCAPE EVOLUTION

Recognition of high inheritance areas, along with a regional field-based ice-flow indicator chronology, may be the first step towards piecing together spatio-temporal subglacial landscape evolution. The main insight of this study is not a detailed reconstruction (history), but rather a series of forms of evidence suggesting that the glacial history of the region is one of prevailing patchy low erosion conditions which favored preservation of a fragmentary record of non-coeval and sometimes contrasting warm-based (more dynamic) conditions (patch sheet flow, Clark, 1999). Altogether, the various GTZ preserve a record of multiple flowsets and ice-flow indicator phases, likely formed during multiple glaciations. The preservation of older relict and/or palimpsest landforms, surrounded by younger landforms has important implications for subglacial ice-flow dynamics. Similar relict areas were denoted in Scotland (Finlayson et al., 2010), where the authors also questioned how these regions were preserved when in the path of several younger phases of relatively fast glacier flow. Additional GTZ analysis beneath the core areas of ice sheets will likely identify more of these regions for future study. These regions of low erosion and presumably slow, sluggish or non-flowing basal ice (termed 'sticky regions') exist amongst areas of higher erosion and faster flowing ice. In this manner, we envision that processes beneath the ice sheet must have occurred as part of a subglacial landscape evolution model similar to the till mosaic concept of Piotrowski et al. (2004) but at a much broader (landscape or landform-field) scale (see also Finlayson et al., 2010). GTZs would then represent the complex end-product of this spatial and/or temporal variability beneath the ice sheets. We envision that subglacial landscape evolution processes were not ice-sheet wide, but instead extremely variable over time and space, leading to spatio-temporal variations in erosion, transportation and deposition intensities (subglacial bed mosaic model of landscape evolution). We suggest spatio-temporal formation and preservation of relict and palimpsest GTZ, beneath sticky regions, could occur by at least two different mechanisms.

Firstly, a locally changing glaciological regime from warm-based to patchy cold-based (e.g. Clärhall and Jansson, 2003) may explain the formation of a fragmented subglacial landscape and preservation under cold-based 'sticky' areas. In this case, sticky areas would be akin to the frozen-bed patches highlighted in Kleman and Glasser (2007), termed the frozen-bed mosaic (Kleman et al., 1999), that typically occur on the upland areas between faster-flowing lowland troughs. If so, our interpretation of an older glacial landscape partially preserved by cold-based ice would lend some credence to the cold-based Keewatin reconstruction presented by Kleman and Glasser (2007). A subglacial permafrost model (Tarasov and Peltier, 2007) and the latest deglacial basal velocity model (Stokes and Tarasov, 2010) also suggests the study area was cold-based during the Last Glacial Maximum (~26.5 to 19 ka). Similar frozen-bed conditions have been suggested for parts of the Fennoscandian Ice Sheet at deglaciation (Kleman et al., 1997). Perhaps even though end-member cold-based geomorphic landscapes (regolith, tors, lack of till, etc) are mostly absent (small patch of regolith in northwest Manitoba, Trommelen, 2011b) in northern Manitoba, cold-based conditions were much more widespread than previously thought. Preserved relict and palimpsest subglacial landscapes, together with a lack of deglacial landsystems, may then be evidence of a change from earlier warm-based conditions to later cold-based conditions.

As conclusive geomorphic evidence of cold-based ice is generally lacking in northern Manitoba (Dredge et al., 1986; Dredge and Cowan, 1989; Dyke and Dredge, 1989; Dredge and Nixon, 1992), and most of the subglacial landscape shows evidence of warm-based conditions, another 'sticky' mechanism is also probable. Considering the importance of subglacial meltwater availability (Boulton et al., 2009) to ice-flow velocity and subglacial deformation/sliding, perhaps establishment of the regionally extensive (~700 km wide by at least 500 km long) dendritic esker channel-system caused spatially variable dewatering of the substrate far back under the ice sheet. Significant dewatering, over a short time-scale, would have resulted in patchy stiffening of the substrate and slowing of subglacial deformation and/or basal sliding (Boulton et al., 2009). These new subglacial conditions, occurring along with climatic warming and episodic surges during deglaciation, would have resulted in slow-flowing sluggish basal ice ('sticky' conditions) in non-surging areas. Indeed, late-deglacial ice-flow in the field area, including weak flowset D, was unable to completely overprint and obliterate older subglacial landscape elements (palimpsest-type GTZs). In some areas, rapid dewatering may have led to basal freeze-on (e.g. Christoffersen and Tulaczyk, 2003) that in-turn preserved relict-type GTZs. We envision that if water availability was a significant landscape modifier, this process was not ice-sheet wide, but instead extremely variable over time and space, leading to spatio-temporal variations in erosion, transportation and deposition intensities (subglacial bed mosaic model of landscape evolution).

2.10 CONCLUSION

A highly fragmented glacial landscape is recognized at the regional to sub-regional scales in northeastern Manitoba, in the southeast corner of the Keewatin Sector, a core region of the Laurentide Ice Sheet (LIS). There is significant spatial disjoint between mainly isochronous streamlined landform flowsets in the field areas. The glacial ice-flow record is also clearly much more complex than suggested by recent 'mega-geomorphology' mapping effort and paleoglaciological inversion models. We have documented old glacial terrains surrounded by younger ones in relict or palimpsest GTZ, highly variable degree of inheritance and overprinting in palimpsest GTZ, and strong overprinting in what is interpreted as a young deglacial GTZ. The nature and relative-age of flowsets within a GTZ should be used to inform regional correlation during paleoglaciological reconstructions, and highlights the need for regional detailed studies in addition to low-resolution continental scale analyses.

The GTZ methodology can be applied to similar glaciated landscapes where spatio-temporal variations in the subglacial environment, including ice-flow orientation, subglacial water availability, ice-flow velocity, and intensity of erosion, transportation and deposition, occurred. This method is particularly useful for ice-divide and near ice-divide (core) regions of ice sheets, where significant palimpsest and relict terrain may occur. Important points of this study include:

- Various degrees of landscape overprinting and inheritance are present within the subglacial landscape, suggestive of uneven spatio-temporal changes in ice-flow direction, erosion intensity and even basal thermal regime, across large areas of this core region (Keewatin Sector) of the Laurentide Ice Sheet.
- More inheritance is preserved in the field-based ice-flow indicator record than in the streamlined landform (flowset) record.
- Rogen moraine present in the northern portion of the GTZ is variably overridden by more active ice, resulting in patchy drumlinization of Rogen moraine ridge surfaces. Preservation of these landforms throughout deglaciation requires low erosion and/or locally cold-based or very sluggish ice patches (as suggested by Knight, 2010).
- There is no obvious deglacial landsystem in northeastern Manitoba. There are very few ice-marginal landforms (e.g. end moraines) or subglacial landforms that can be correlated to the late-deglacial retreat of ice over the study area. The eskers in the area are part of a very large dendritic esker network, and do not appear to be related to local ice-marginal dynamics.
- GTZ analysis suggests that the subglacial landscape may have evolved with considerable spatio-temporal variation, termed herein as the subglacial bed mosaic.
- Further delineation of GTZ, based on detailed field mapping that includes ice-flow indicators, is required to advance our understanding of the subglacial bed mosaic beneath the inner (core) regions of ice sheets.

Chapter 3 Inherited clast dispersal patterns: implications for paleoglaciology of the southeast Keewatin Sector of the Laurentide Ice Sheet

3.1 OVERVIEW

The net effect of ice-flow shifts resulting in dilution or reworking of clasts on a single preserved till sheet is often unknown yet has major implications for palaeoglaciology and mineral exploration. Herein, we analyze variations in till clast lithologies from a single till sheet, within palimpsest-type Glacial Terrain Zones in northeast Manitoba, Canada, to better understand sediment-landform relationships in this area of high landform inheritance. This near-ice divide area is known to consist of a highly fragmented subglacial landscape, resulting from spatio-temporal variations in intensity of reworking and inheritance throughout multiple glacial events (subglacial bed mosaic). We show that a seemingly homogenous 'Keewatin' till sheet is comprised of local (>15 km) and continental-scale (~100 km long carbonate train and 350-600 km long Dubawnt red erratic train) fan, amoeboid and irregular or lobate palimpsest dispersal patterns. Local dispersal is more complex than the preserved local landform flowset(s) record, but appears consistent with the overall glacial history reconstructed from regional flowset and striation analyses. The resultant surface till is a spatial mosaic interpreted to reflect variable intensities in modification (overprinting) and preservation (inheritance) of a predominately pre-existing till sheet. A multi-faceted approach integrating till composition, regional landform, ice-flow indicator, and stratigraphic knowledge is used to map relative spatio-temporal erosion/reworking intensity.

3.2 INTRODUCTION

The subglacial geomorphic record of palaeo-ice sheets is often a detailed series of 'snapshots' that provide a window into palaeo-subglacial conditions and basal processes. Previous research envisioned largely 'steady-state' subglacial processes, assuming relatively uniform conditions beneath an ice-sheet, where at a large scale changes in erosion and deposition rate varied with distance from an ice divide and the external climatic regime (Boulton, 1996). It is now recognized that the subglacial environment was one of significant spatio-temporal variability and evolution throughout the Quaternary (Boulton and Clark, 1990; Kleman et al., 1997; Kleman and Stroeven, 1997; Stroeven et al., 2002; Stroeven et al., 2006; Bradwell et al., 2008; Goodfellow et al., 2008; Kleman et al., 2008; Greenwood and Clark, 2009a; Finlayson et al., 2010; Hall and Migon, 2010; Refsnider and Miller, 2010; Smith and Knight, 2011), owing to fluctuations in hydrology, basal thermal regime, ice thickness, shear stress, and sediment rheology (Glasser and Bennett, 2004; Eyles, 2006; Meriano and Eyles, 2009).

In order to produce the most accurate ice sheet models, it is advantageous to procure detailed information about the evolution of local- to regional-scale subglacial conditions (Tarasov and Peltier, 2004, 2007; Greenwood and Clark, 2009b; Kleman et al., 2010; Refsnider and Miller, 2010). Spatio-temporal analysis of complex streamlined landform records, combined with field data such as ice-flow indicators and till composition, has led to the recognition of a fragmented glacial landscape that consists of discrete landform assemblages (Kleman et al., 1997; Stea and Finck, 2001; Ross et al., 2009; Finlayson et al., 2010). We propose to use these discrete landform assemblages, such as the palimpsest or relict-type Glacial Terrain Zones (GTZs) of Chapter 2 (Figure 3-1), to piece together the relative age of various subglacial 'structural elements' (Kleman et al., 1997; Boulton et al., 2001b). Then, it is possible to infer the degree of overprinting (partial to complete modification, Stea and Finck, 2001) or inheritance (preservation of relict surfaces), which in turn can be used to interpret the net effect of spatio-temporal changes in intensity of subglacial erosion, transportation and deposition throughout glaciation(s) (cf. Knight, 2010). By identifying separate erosional or depositional phases leading up to formation of the subglacial landscape, we can then start to interpret spatio-temporal variations in palaeo-subglacial conditions (Hicock and Lian, 1999; Lian and Hicock, 2000; Smith et al., 2007; Stokes et al., 2008; Lian and Hicock, 2010) – rather than just the end-member (i.e. warm-based or cold-based) conditions.

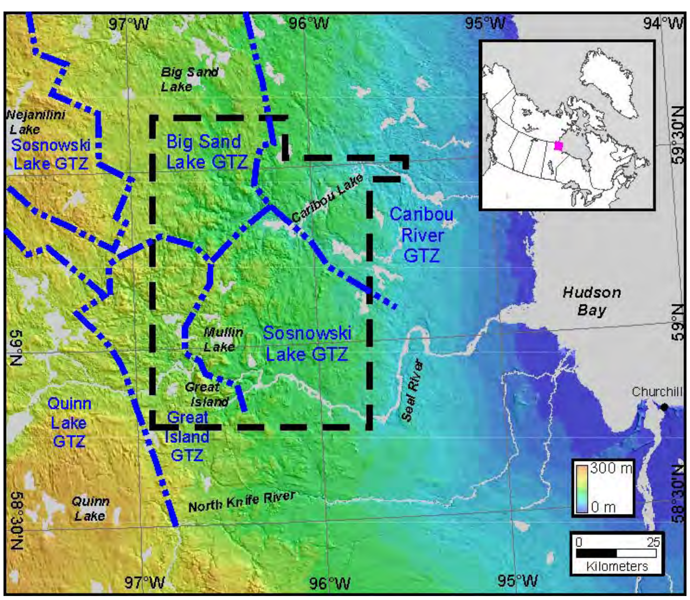


Figure 3-1. Field area (dashed box) in northeast Manitoba, with glacial terrain zones (GTZ) from Chapter 2. The background image was generated using the radar-derived digital elevation data from the Shuttle Radar Topographic Mission (SRTM) data set (USGS). A hillshade model has been added to enhance the relief.

A measure of erosion intensity, for example, is often desired in order to interpret basal thermal regime or to use as a proxy for ice-flow velocity or ice-sheet configuration (Hall and Glasser, 2003; Hildes et al., 2004; Henderson and McMartin, 2008; Kleman et al., 2008; Knight, 2010). In warm-based areas, methods using streamlined landform characteristics (density (Hättestrand et al., 1999; Bradwell et al., 2008) or elongation ratio (Hart and Smith, 1997)), drift thickness versus glacially-scoured bedrock lake basin density (Henderson and McMartin, 2008; Kleman et al., 2008), and cosmogenic nuclide ratios (Fabel et al., 2002; Stroeven et al., 2002; Briner et al., 2006; Harbor et al., 2006; Miller et al., 2006; Staiger et al., 2006; Refsnider and Miller, 2010) have been introduced to help quantify erosional intensity and ice-flow velocity. These are difficult approaches, however, to interpret the evolution of erosional intensity, especially where till cover is widespread, scoured lake basins are rare, and subglacial landform flowsets are disjoint, fragmented, rare or heavily overprinted.

An alternative, but often-overlooked proxy for the evolution of erosion intensity is till composition. When a distinctive source is known, indicator clasts within till can delimit transport distances (Dreimanis and Vagners, 1971; Shilts, 1980; Dredge, 1988; Charbonneau and David, 1993; Parent et al., 1995; Boulton, 1996; Kjaer et al., 2003; Hildes et al., 2004; Plouffe et al., 2011a). The strength of a clast dispersal pattern largely depends on the net erosion intensity up-ice in the source area(s) (Hildes et al., 2004; Kleman et al., 2008), as well as on the duration of transport in one or more direction(s) relative to the source area (Kleman et al., 2008). If these parameters are paired with information pertaining to roughness and/or hardness of local relief (i.e. Philips et al., 2010), and other subglacial landscape elements such as geomorphology, the evolution of subglacial conditions such as basal thermal regime and mechanical conditions can also be estimated (Dreimanis and Vagners, 1971; Aario and Peuraniemi, 1992; Evans et al., 2006).

Near the inner core regions of palaeo-ice sheets, where fragmented palimpsest and relict-type GTZ (Chapter 2) are present, assessing till provenance, erosion and transportation history is challenging. This is because ice-divide areas of the Laurentide Ice Sheet (LIS) were characteristically regions with low ice-flow velocities, related to fluctuating basal thermal regimes (Bouchard, 1989; Clärhall and Jansson, 2003; Kleman and Glasser, 2007) and low subglacial meltwater availability (Boulton, 1996; Clark, 1999; Boulton et al., 2001a; Clarke, 2005; Evans et al., 2006), resulting in low subglacial erosion intensities (Andrews and Miller, 1979; Boulton, 1987; Stea and Finck, 2001) and thus high inheritance. At small distances from the inner ice-divide regions, subglacial erosion intensity was likely highly variable owing to additional factors such as ice-divide migration and ice advance and retreat cycles (Kleman et al., 2008), which may locally involve development of ice-streams (e.g. Dubawnt Lake Ice Stream, Stokes and Clark, 2003a) and complex time-transgressive overprinting of variable durations (Andrews and Miller, 1979; Boulton, 1996; Glasser and Bennett, 2004; Bradwell et al., 2008). Irregular multi-lobate till dispersal patterns (amoeboid, Figure 3-2.; Turner and Stea, 1987; Stea et al., 1989; Shilts, 1993; Parent et al., 1996; Stea and Finck, 2001) that indicates dispersal in several, often opposing directions, is often a diagnostic net effect for the occurrence of this spatio-temporal variability in subglacial conditions. In these cases, a higher amount of inheritance is suspected, and younger ice-flow events may have been unable to fully overprint the subglacial records from older ice-flow events.

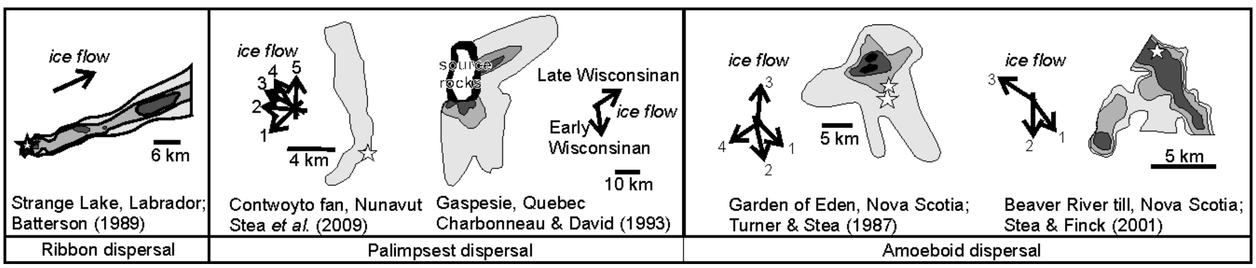


Figure 3-2. Example of common dispersal patterns in till, from various detailed studies in Canada. Unless otherwise indicated, stars represent the bedrock source.

In this paper, we assess how the final composition of thin surface till is related to the evolution of a glacial landscape in northeastern Manitoba, Canada, produced by the Keewatin Sector of the LIS. We first compare clast dispersal patterns to spatial subglacial landform and field-based ice-flow indicator data, to assess the overall degree of inheritance and overprinting on the landscape. An example of how this information can then be used to construct palaeoglaciological models – including spatio-temporal evolution of relative erosion intensity, is shown.

3.3 Regional Setting

The Great Island area in northeastern Manitoba (Figure 3-1) is an ideal location to study glacial sediment dispersal because the low-lying region has distinctive low metamorphic grade supracrustal rocks, surrounded by felsic intrusives (Anderson et al., 2009a; Anderson et al., 2010b), that were recently mapped (Figure 3-3) at a scale appropriate for tracing of clast provenance. A mantle of glacial and post-glacial sediments drape the area (Figure 3-4), but scattered bedrock outcrops provide good control on bedrock geology maps. From outward appearances, the regional till seems relatively homogenous throughout the study area. There are no sections or drill holes in the study area, though multiple till sequences are noted south of the North Knife River (Dredge and Nixon, 1992) and in the Hudson Bay Lowlands (Klassen, 1986; Nielsen, 2001, 2002a; Dredge and McMartin, 2011).

The study area was glaciated by ice flowing from multiple, migratory, domes of the LIS in the Late Wisconsinan and presumably previous glaciations as well (Dredge and Thorleifson, 1987; Vincent and Prest, 1987; Thorleifson et al., 1992; Dyke et al., 2002; Kleman et al., 2002; Kleman et al., 2010). Recent detailed work in the area (Chapter 2) has revealed a long and complex ice-flow history. A simplified regional generalization (insets on Figure 3-3; some data on Figure 3-4) indicates that ice has flowed between east-southeast and west-southwest multiple times throughout the Quaternary (Chapter 2). Additional rarer northeastward and northwestward ice-flow phases are also documented ((Chapter 2). These ice-flow phases are based on ice-flow indicators such as striae, grooves, chattermarks, gouges, crescentic fractures, roches moutonnées and whaleback drumlins identified in the field (Campbell et al., 2012), and flowsets (Chapter 2) generated from streamlined landforms mapped using aerial photographs and satellite imagery (Trommelen and Ross, 2010).

3.3.1 Inheritance

Northeastern Manitoba contains a subglacial landscape that likely formed over multiple glaciations from different migratory source areas (Chapter 2). Elements of a deglacial ice-marginal landsystem (Colgan et al., 2003; Evans, 2003b) are mostly absent from the study area (Chapter 2), even though deglaciation of the area was not complete until around 7770±40 ^{14}C BP (GSC30-70, Dredge and Nixon, 1992). Instead, the region consists of high inheritance areas (Caribou River relict-type GTZ) that contain old streamlined landform flowset fragments, or, high inheritance areas interspersed with overprinting younger streamlined landform flowsets (Big Sand Lake, Great Island and Sosnowski Lake palimpsest-type GTZs) (Figure 3-1). The Quinn Lake GTZ is a warm-based deglacial-type, and is not discussed herein as it falls outside of the study area.

3.3.2 Clast provenance

Most subglacial detritus in the till is locally derived. The study area (Figure 3-3) consists of both Archean basement granitoid and orthogneiss rocks and rarer Archean supracrustal rocks, all overlain by scattered Archean and Palaeoproterozoic siliciclastic rocks (Anderson et al., 2009a; Anderson et al., 2010a, b).

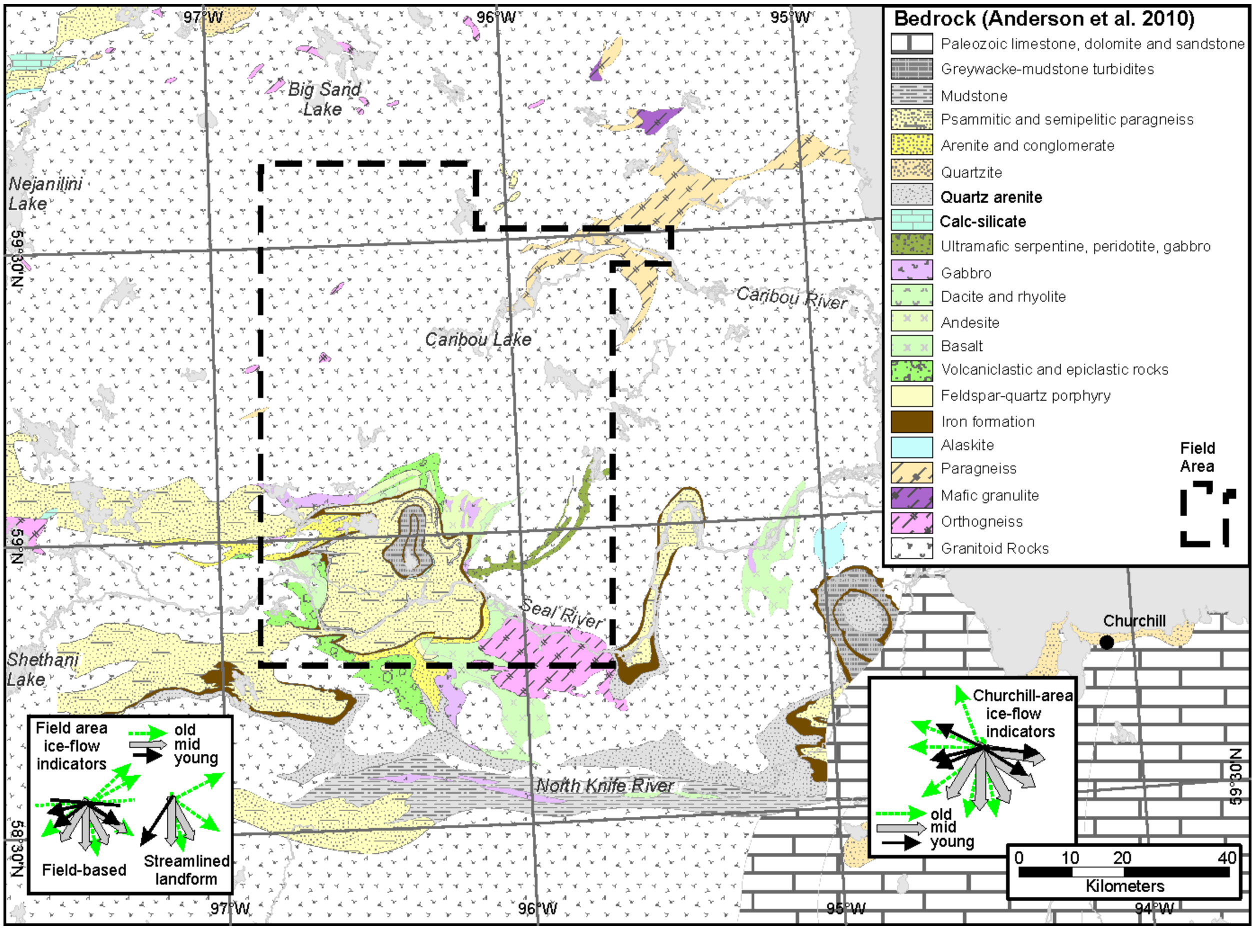


Figure 3-3. Regional bedrock lithologies from Anderson et al. (2010b). Relative chronology of field-based ice-flow indicator data (striae, grooves, chattermarks, gouges, crescentic fractures) is also depicted, for both the field area and the Churchill area (Trommelen et al. 2010; Trommelen & Ross 2011a). Streamlined landform orientation is provided for the study area (Trommelen et al. 2012).

Siliciclastic (greenschist-grade) rocks of the Seal River Domain (Anderson et al., 2010b) are present throughout the southern field area and include oxide-silicate-facies iron formation, arenite, conglomerate, quartz arenite, varicoloured mudstone (Figure 3-5A and Figure 3-5B), greywacke, andalusite-bearing paragneiss and a minor pink to white marble (Figure 3-3). These lithologies are generally not resistant to erosion, can be broken apart easily, and are commonly angular to sub-angular. The study area also contains the only known occurrences of volcanic rocks in Manitoba's far north (Anderson et al., 2010b, a), which includes basalt and andesitic flows (Figure 3-5C), minor dacite and rhyolite and thick packages of volcaniclastic rocks (Figure 3-5D) of the Sosnowski Lake Assemblage (Seal River Domain, Anderson et al., 2010b). Gabbro and ultramafic (serpentine and peridotite) rocks are also present. Parts of the region

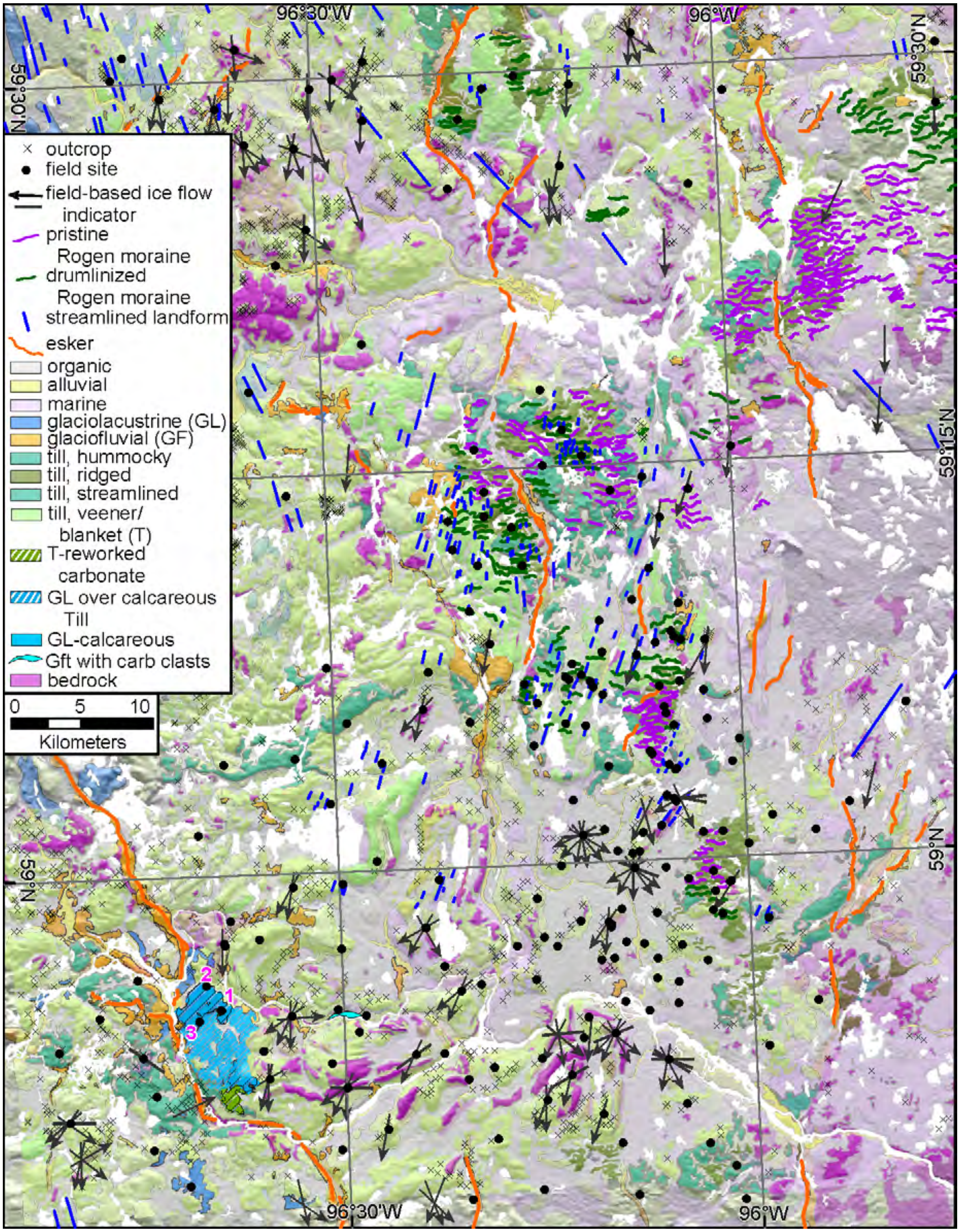


Figure 3-4. Simplified surficial geology in the study area, based on new data from Trommelen and Campbell (2012a, b, c, d) within the field area and supplemented by regional data from Dredge et al. (2007). Carbonate-bearing sediment includes till (samples 1-3), glaciolacustrine silty-clay with weathered carbonate granules and a glaciofluvial terrace (GFt) with common rounded dolostone cobbles.

are draped by thick sediment/bog cover with sparse bedrock exposure, but mapping inaccuracy is mitigated by detailed aeromagnetic and gamma-ray spectrometric geophysical surveys (Fortin et al., 2009), grid-pattern foot-traverse and helicopter-assisted field mapping, and sparse post-fire vegetation. The bedrock mapping is considered to be sufficiently accurate for the purposes of this study (R. Syme, pers comm. 2011), though there is always a possibility that minor 'hidden' granulite-grade supracrustal rocks may exist beneath regions of thick glacial or postglacial sediment to the north. There are scattered sedimentary and volcanic clasts, within till, in the northern part of the study area, and the source-area for these is more uncertain. The next-closest known source of sedimentary and volcanic rocks (metamorphosed Hearne Craton) crops out approximately 150 km north of the study area, in Nunavut, Canada (Tella et al., 2007).

Indicator lithologies of continental-scale include both Dubawnt Supergroup red erratics (red/purple, Table 3-1), sourced 350-600 km north to northwest of the top of the study area (Peterson, 2006; Tella et al., 2007), and Palaeozoic carbonate derived from the Palaeozoic Hudson Platform in the Hudson Bay Lowlands (Manitoba Energy and Mines, 1980), southeast of the field area (Figure 3-3). Palaeozoic carbonate rocks also crop out beneath the waters of Hudson Bay (Grant and Sanford, 1988). Transported Dubawnt Supergroup lithologies (Figure 3-5E and Figure 3-5F) include red metasandstone (Thelon Formation), and volcanic porphyries with white phenocrysts (Pitz Formation) or phlogopite phenocrysts (Christopher Island Formation).

3.4 METHODOLOGY

Representative surface till samples (25-130cm depth) were collected during surficial mapping (1:50 000 scale; Trommelen and Campbell, 2012b, a, c, d) of the study area (Figure 3-4). Till showing evidence of possible post-depositional winnowing by marine or glaciofluvial processes (e.g. coarse matrix with silt-coated clasts) was avoided during sampling. By removing the possibility of fluvial or marine transport, we assume dispersal patterns result exclusively from glacial transport. Post-depositional break-down of a single large transported clast may skew pebble counts (most clasts are sub-angular to angular and only partially abraded). This has been taken into account in the interpretation of isolated outlier values, but it is assumed that this process has not been widespread enough to affect the overall results.

3.4.1 Clasts counts

Sample preparation was completed at the University of Waterloo in 2009 and at the GSC Sedimentology Laboratory in 2010. Clasts were sieved from a portion of each till sample collected (n=230), and further separated into clast-size fractions of 2-4 mm, 4-8 mm and 8-80 mm (Figure 3-6). Then, granules or pebbles within each clast-size fraction were separated according to lithology under a binocular microscope. Additional surface clast counts (35 sites) were taken from mud boils, or from within the sample hole when no mud boils were present. Clasts were separated into 23 detailed rock-types (Campbell et al., 2012). For interpretation, the clasts were then grouped into eight simplified classes (sedimentary; volcanic; granitoid; quartzite; mafic and ultramafics; red/purple; Palaeozoic carbonate; and exotic) to reduce lithological identification errors. The breakdown of clast-types included in the eight classes is presented in Table 1. Fuchsite-rich quartz-porphyry is not a sedimentary rock but derives from a small group of outcrops in the middle of sedimentary rocks, so it has been included in this class for dispersal analysis.

Table 3-1. Simple and detailed lithological classes for till sample clasts.

	Simplified classes							
	Sedimentary	Volcanic	Granitoid	Mafic and Ultramafic	Quartzite	Carb	Red/Purple	Exotic
Detailed classes	mudstone / pelite / phyllite sandstone / psammite / schist fuchsite-rich quartz-porphyry* marble	basalt / andesite rhyolite / dacite volcani-clastic	red granite tonalite granodiorite gneissic orthogneiss	gabbro amphibolite ultramafic	quartzite quartz-arenite	lime dolo	meta-sediment meta-volcanic meta-conglomerate	chert white granitoid

*Not a sedimentary rock but derives from a small group of outcrops in the middle of sedimentary rocks, so it has been included in this class for dispersal analysis.

3.5 RESULTS

Sedimentary, volcanic, red/purple and Palaeozoic carbonate clast dispersal is discussed herein, as the source areas for the other rock types are largely unknown. Because the effects of comminution on each rock type are unknown, or may not be comparable over all classes, we chose to group the clast-size fractions (Table 3-2) together for spatial analysis. The results are depicted as count-percentage bubble plots (ArcGIS), for both the sieved and field (not collected at every site) clasts, and discussed below. The occurrence of multiple bedrock source areas, the bedrock configuration and the non-grid sampling pattern, in this case, limits the usefulness of automatic contouring by a software program. This is not a

Figure 3-5. Distinctive clast-lithologies encountered within tills from the field area. Includes greywacke mudstone (A), friable mudstone (B), andesite (C), volcaniclastic (D), Dubawnt Supergroup (E, F), fossiliferous limestone (G) and carbonaceous clasts (H).


A
B
C
D
E
F
G
H
crinoid
ossicle
carbonate
granule
Manitoba Geological Survey

problem, however, given that we are more interested in the spatial distribution of values at sampled locations, rather than in estimated values between points (interpolation).

Figure 3-6. Sorted clasts from a till sample, taken from a site southeast of Great Island. The clasts have been separated by lithology and grain-size (8+mm, 4-8 mm and 2-4 mm sieved size-fractions). The location of site 10MT149 is marked on Figure 3-7 and Figure 3-8.

Table 3-2. Total count-percent of clasts per grain-size fraction, for each simplified lithology.

Clast Fraction	Sedimentary	Volcanic	Red/purple	Carbonate	Granitoid	Mafic and Ultramafic	Exotic	Quartz	Sum
2-4 mm	13.46	1.39	0.45	0.17	71.78	0.81	0.24	11.7	125274
4-8 mm	16.19	0.8	0.19	0	74.65	2.2	0.55	5.41	7262
8+ mm	13.83	1.78	0.39	0.22	80.53	0.31	0.13	2.82	6394
Field 20-80 mm	12.68	7.89	0.69	0.27	74.43	0.54	0.07	3.43	5916

3.5.1 Local to regional supracrustal clasts

The spatial distribution of sedimentary (Figure 3-7) and volcanic (Figure 3-8) clasts from sampled tills is complex, producing an amoeboid pattern (Figure 3-2). The concentration of each clast-type is highest in till that overlies the corresponding mapped bedrock and decreases to zero within 5 to 9 km northwards. Dispersal of clasts beyond the mapped bedrock extends at least 15 km toward the NE and SE for sedimentary rocks (Figure 3-7) and 10 km to the NE, 15 km to the SE and 8 km to the SSE for volcanic rocks (Figure 3-8). The rare volcanic clasts in the till northwest of Mullin Lake may indicate northwest dispersal at some point in time. Given that southeast to southwest ice flow is known to have occurred at least twice in the study area (Chapter 2), significant clast dispersal is also expected to

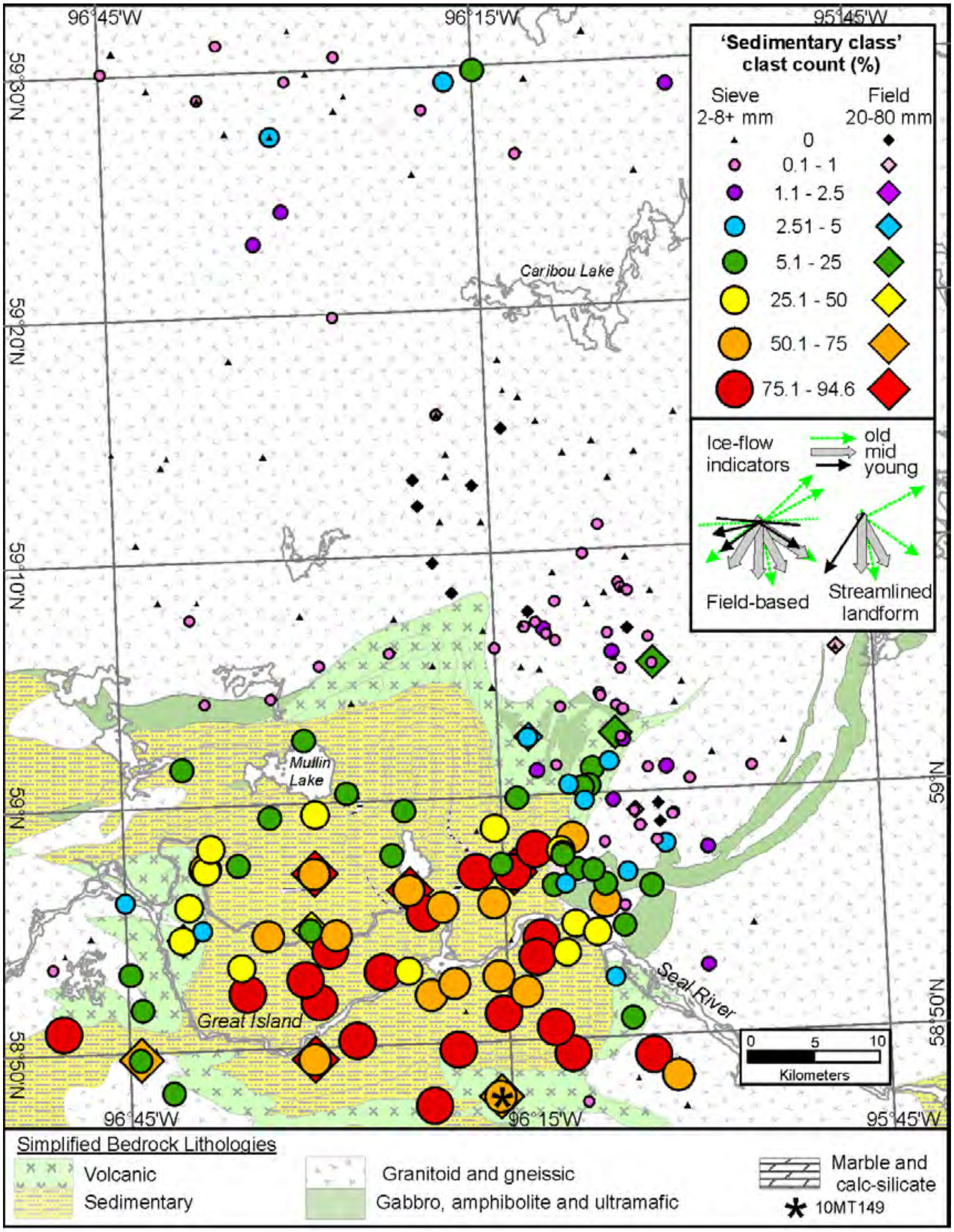


Figure 3-7. Sedimentary and metasedimentary clast dispersal in till, Great Island, northeast Manitoba. Bubble plots represent count-percent, based on number of clasts/sample. Underlying simplified bedrock geology is from Anderson et al. (2010b). Relative chronology of field-based ice-flow indicator data (striae, grooves, chattermarks, gouges, crescentic fractures) is also depicted, for both the field area and the Churchill area (Trommelen and Ross, 2011a; Trommelen et al., 2012a). Streamlined landform orientation is provided for the study area.

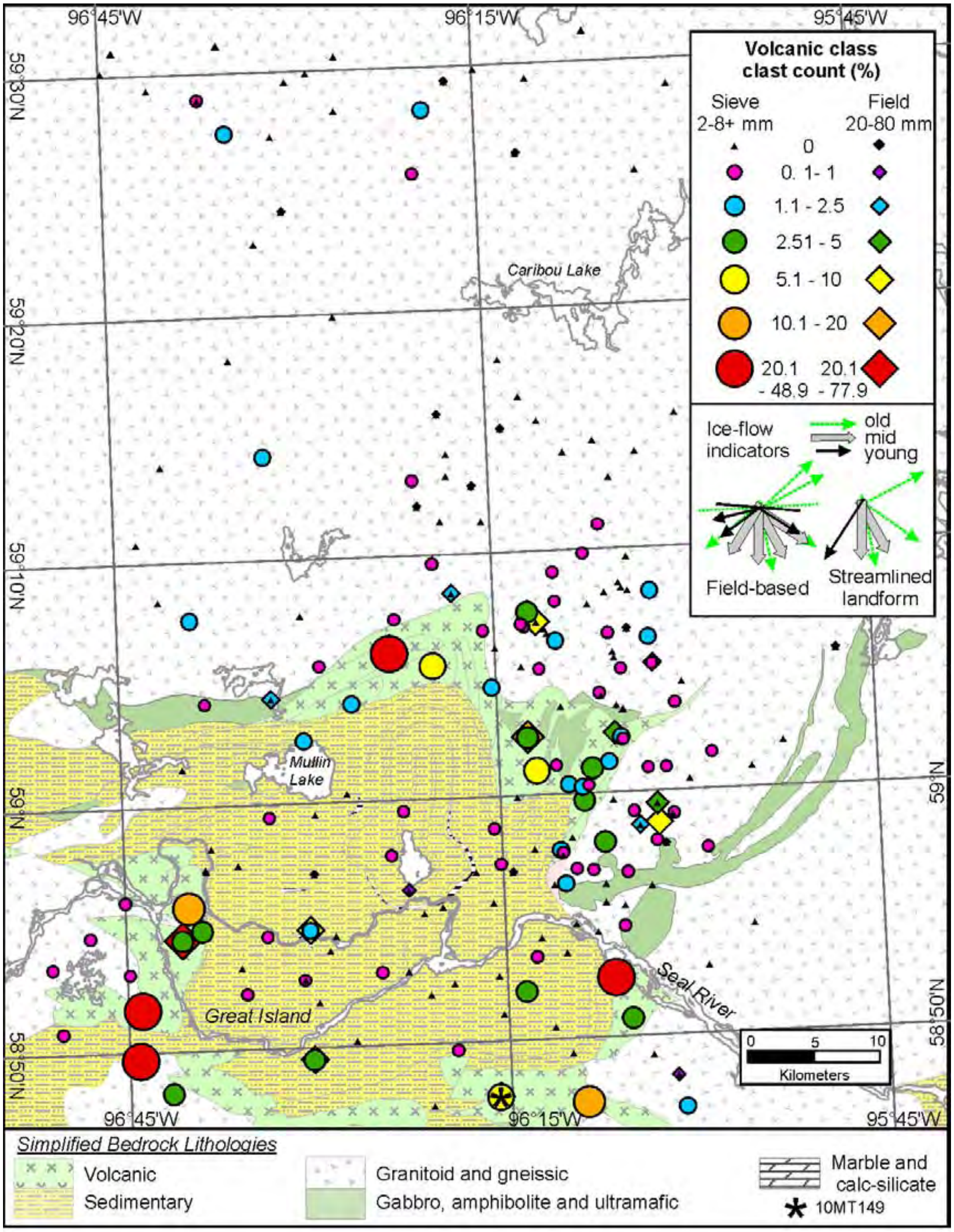


Figure 3-8. Volcanic (basalt, andesite, rhyolite, dacite, volcaniclastic) clast dispersal in till, Great Island, northeast Manitoba. Bubble plots represent count-percent, based on number of clasts/sample. Underlying simplified bedrock geology is from Anderson et al. (2010b). Relative chronology of field-based ice-flow indicator data (striae, grooves, chattermarks, gouges, crescentic fractures) is also depicted, for both the field area and the Churchill (Trommelen and Ross, 2011a; Trommelen et al., 2012a). Streamlined landform orientation is provided for the study area.

extend beyond the study area to the south and southwest. Volcanic concentrations are low within 11-15 km SW of the northern outcrops, though this may not indicate maximum SW transport distance but instead could signify dilution due to entrainment of the more friable sedimentary clasts. Finally, there is a slight increase in concentration of both clast types approximately 35-40 km north of the mapped rocks. The source of these clasts is unknown, but may indicate the presence of buried supracrustal bedrock near, or north of, the Manitoba- Nunavut border.

3.5.2 Continental-scale erratics

The distribution of low concentrations of Dubawnt red/purple clasts (Figure 3-9) in tills across the study area is indicative of the tail-end of a continental-scale dispersal fan. Dubawnt concentration generally decreases from north to south, though the highest concentration of red/purple clasts was actually sampled from at a site on the west side of Great Island (site 1, Figure 3-4). Here, an upper till sample (15-25 cm depth) contained 48 (5.6%) red/purple clasts, while a lower (60-70 cm depth) sample at the same site contained 19 (1.6%). The concentration of red/purple clasts in both samples at this site is more than two standard deviations (std) higher than the concentration in the surrounding till (Table 3-3).

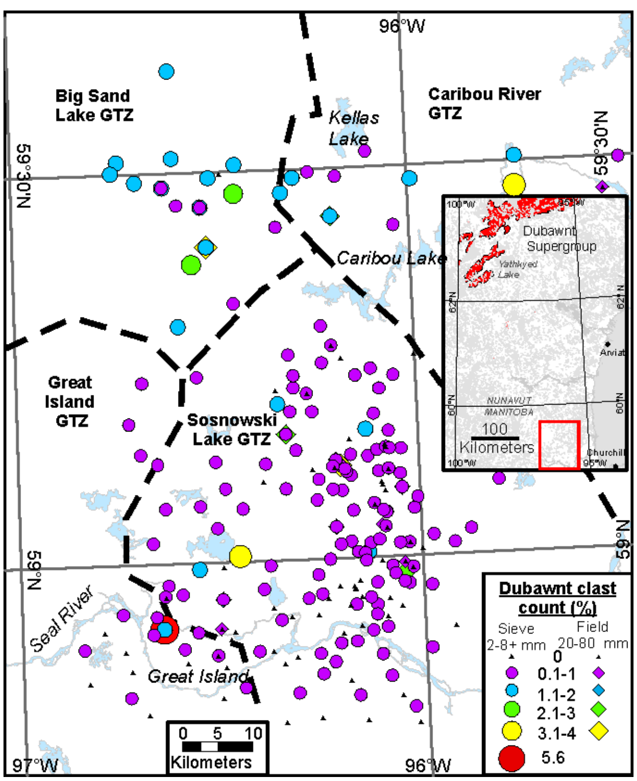


Figure 3-9. Dubawnt Supergroup (red/purple erratic) clast dispersal in till. Bubble plots represent count-percent, based on number of clasts/sample. Bedrock geology inset is from Peterson (2006), showing the source area 350-600 km north-northwest of the study area.

Carbonate clast distribution in the study area is combined with regional data (Dredge and Pehrsson, 2006) (Figure 3-10), and results indicate that the study area may be outside of any recognizable Palaeozoic dispersal pattern. Rare carbonate clasts (0.1-0.4 count-percent) were noted across the field area, up to 80-90 km inland from Hudson Bay. Higher concentrations of carbonate clasts were measured from a ~11 km^2 area of carbonate till (Figure 3-4), 35 km NNW of the closest previously mapped calcareous silt-rich till (Figure 3-10) (Dredge et al., 2007). Where encountered, the calcareous till at Great Island underlies 15–50 cm of calcareous silt and clay (Figure 3-4) and/or Keewatin till (gradual contact). A helicopter-assisted ground-truthing survey determined the surrounding till, within an 8 km radius, contains no Palaeozoic clasts (Figure 3-10).

Lithologic analysis has revealed that the till with carbonate-clasts at Great Island cannot be characterized as one till type, but is actually quite heterogeneous. Though the texture (Table 3-3) and granitoid concentration is similar (~54%) for the till sampled at sites 1 and 3 (Figure 3-4), the count-percent of carbonate clasts within these till samples varies greatly. At site 1, the proportion of carbonate in the lower (60-70 cm depth) till sample is considerably elevated (17.6%, 13 std) even in relation to the upper (15-25 cm depth) till sample at the same site (1.1%, 1 std). Both samples at site 1 contain elevated concentrations of Dubawnt clasts, relative to not just the surrounding region, but the entire study area. The sample at site 2 contains considerably less sand and more silt than any other samples (Table 3-3). As such, this sample may not be till but instead a glaciolacustrine flow diamicton.

Table 3-3. Simplified lithology and texture from sieved samples of carbonate-bearing till (locations of sites 1-3, Figure 3-4).

Sample Site	Clast Lithology								Matrix texture		
	Carbonate		Supracrustal (sed and volc)		Granitoid		Dubawnt		Sand	Silt	Clay
	#	%	#	%	#	%	#	%	%	%	%
1 - upper	9	1.10%	306	36%	492	57%	48	5.60%	61.4	36.6	2
1 - lower	206	17.60%	275	23%	673	57%	19	1.60%	65.6	32.7	1.8
2	27	8.10%	234	70%	71	21.20%	2	0.60%	36.6	58	5.4
3 - sieved	36	4.20%	383	44.10%	446	51.30%	2	0.20%	50.2	48.5	1.3
3 - field	15	22.40%	26	38.80%	26	38.80%	0	0%			
Field area median	*0*	*0%*	*68*	*10.50%*	*514*	*83.00%*	*2*	*0.20%*	*63.9*	*31.4*	*4.7*
Field area standard deviation	*14*	*1.30%*	*302*	*43.00%*	*305*	*28.40%*	*5*	*0.70%*	*13*	*10.8*	*3.6*

The association between carbonate-bearing till and calcareous silt and clay (glaciolacustrine?) sediments in the Great Island area is important. At two field sites, the clayey-silt sediments contained up to five percent carbonate and red mudstone granules (likely derived from Hudson Bay), with no supracrustal or granitoid clasts. These fine-grained sediments are present throughout western Great Island (Figure 3-4) up to the highest point in the field area (between 190 and 260 m a.s.l.; Trommelen and Campbell, 2012a). To the west of Great Island, non-calcareous coarser-grained sediments assumed to have been deposited within late deglacial Lake Agassiz (Dredge, 1983) are situated at 180-220 m a.s.l. (Trommelen and Campbell, 2012a).

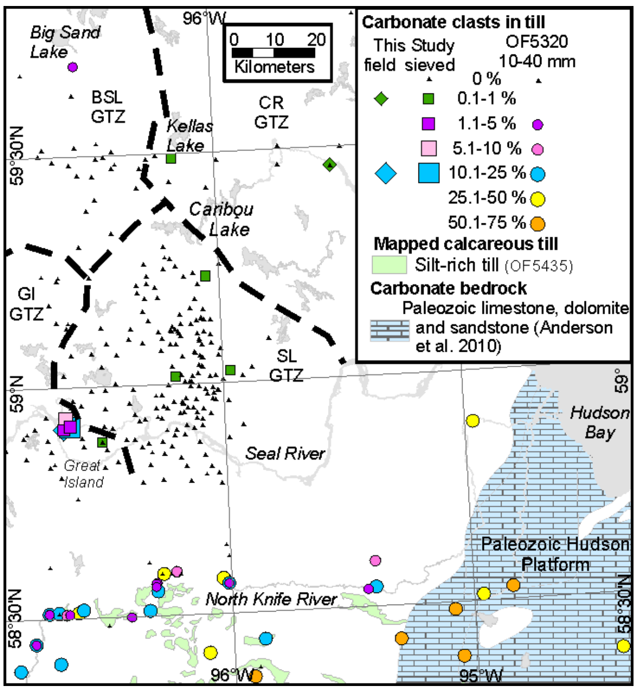


Figure 3-10. Palaeozoic clast dispersal in till, depicted with data from this study inside the field area and from Dredge & Pehrsson (2006) outside. Carbonate bedrock is sourced from the Palaeozoic Hudson Platform southeast of the field area, and is also situated beneath Hudson Bay.

3.6 REGIONAL PALEOGLACIOLOGICAL INTERPRETATION

This near-ice divide area in northeastern Manitoba is known to consist of a highly fragmented subglacial landscape subject to spatio-temporal variations in intensity of overprinting and inheritance, throughout multiple glacial events (subglacial bed mosaic; Trommelen et al., 2012a). As such, the area presents an opportunity to compare and contrast various proxies of ice flow over a long glacial history. What follows is an examination of streamlined landform flowsets, Rogen moraine and esker subglacial landform characteristics, field-based ice-flow indicator data and clast dispersal (patterns, transport distances, and source areas) from a small portion of each of the study area GTZs.

3.6.1 Sosnowski Lake area, Sosnowski Lake Glacial Terrain Zone

The Sosnowski Lake region (Figure 3-11) of this palimpsest-type GTZ (Figure 3-1) likely has the lowest landform inheritance relative to other palimpsest- and relict-type GTZs in the area (Chapter 2), but overall high inheritance is still suspected due to the presence of complex and contrasting striations. This region is dominated by a weak young south-southwest-trending streamlined landform flowset (Flowset D, Trommelen et al., 2012a) that consists of patchy flutes and drumlins (Figure 3-11A) with low elongation ratios (1:1 to 1:3). There are also several small fragments of older south-trending and southeast-trending (~160°) streamlined landform flowsets (Figure 3-11A), which indicates that patchy low-erosion conditions were present for their preservation. Three Rogen moraine fields, consisting of pristine (high amplitude and straight-crested) and more subdued variably drumlinized Rogen moraine are present in the region (Trommelen and Ross, 2010), with ridge axes oriented roughly east-west. One branched esker and three smaller esker segments are mapped and all follow a general north-south orientation (oblique to the youngest streamlined landforms) with an inferred palaeoflow to the south.

The regional field-based ice-flow indicator data (Campbell et al., 2012; Trommelen et al., 2012a) is complex (Figure 3-11B), which further indicates that the young southwest-trending ice-flow phase was weak (short duration/low intensity) and unable to obliterate pre-existing striae. Rare ice-flow indicators such as protected and thus older striae and grooves, or large macroforms, exhibit a wide range of orientations, from 110° to 250°. Intermediate relative-age ('mid'; presumably Late Wisconsinan) striae and grooves show a transgressive swing in orientation from east-southeast to southwest over time. Mid-aged southwest-trending ice-flow indicators are likely coeval to the south-southwest-trending streamlined landform flowset. The youngest (presumably late deglacial) field-based ice-flow indicators are fine, small striae and grooves that occur on the tops of outcrops; oriented west-southwest and then east-southeast.
Locally sourced clasts are dispersed within the till (Figure 3-7 and Figure 3-8) to the northeast and southeast away from the mapped bedrock boundaries (Figure 3-11C) in the Sosnowski Lake area. While the amoeboid dispersal patterns likely resulted from multi-cyclic glacial erosion and transport, several important conclusions about transport direction can be made. Firstly, there is regional south-southwestward clast dispersal (Figure 3-7 and Figure 3-8) extending some unknown distance outside the field area – parallel to the dominant south-southwest-trending flowset. In the Sosnowski Lake area, however, some clast dispersal trains (Figure 3-11D) indicate transport towards the southeast, and

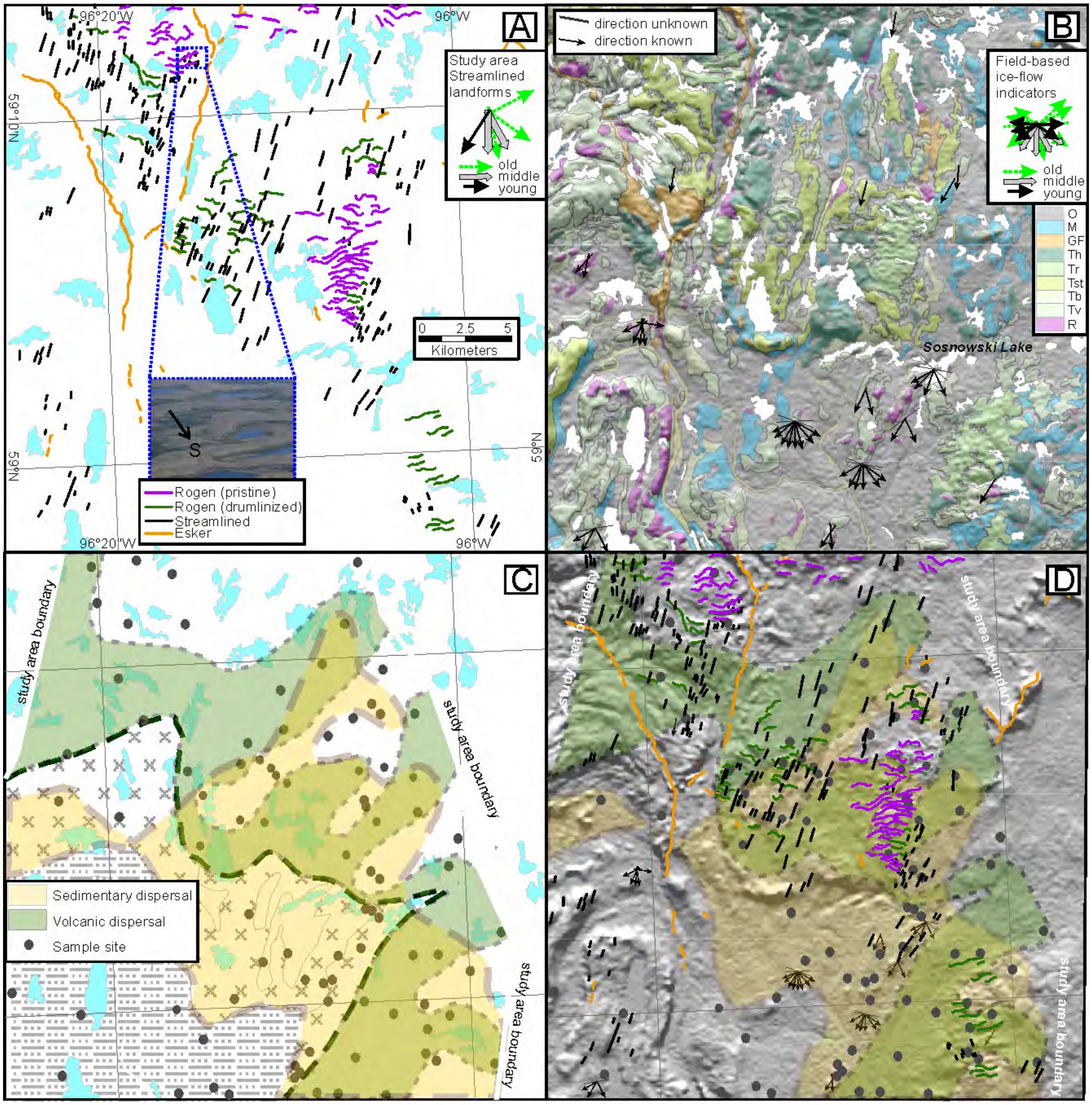


Figure 3-11. Multi-proxy data in the Sosnowski Lake GTZ, including subglacial landforms (A), field-based ice-flow indicator data, 90 m resolution SRTM hillshade and surficial materials (1:50 000 scale) (B), and local clast dispersal (C) are combined (D) to interpret the relationship between erosional and depositional ice-flow indicators. The compiled data indicates the surface till preserves an inherited record of dispersal to the northeast and southeast - even within Rogen moraine ridges, and areas overprinted by young streamlining to the southwest. The green dashed line in (C) is the bedrock contact with granitoid rocks (Fill patterns: X = volcanic rocks and horizontal dots and lines = sedimentary rocks). O – organic; M – marine; GF – glaciofluvial; Th – hummocky till; Tr – rogen moraine; Tst – drumlinized till; Tb – till blanket; Tv – till veneer; R – bedrock.

possibly to the east and/or northeast, beneath fields of Rogen moraine and *within* fields of streamlined landforms indicative of ice-flow to the south-southwest.

3.6.1.1 Palaeoglaciological implications

Southeastward dispersal of local clasts within the till, and possible remnant eastward and/or northeastward dispersal, must be a record of inherited subglacial detritus as ice-flow in these directions is thought to have occurred prior to the south-southwest streamlining ice-flow phase (Figure 3-11). Southeast-trending ice-flow indicators are locally present, but east and northeast-trending ice-flow indicators are much rarer (Figure 3-4) (Chapter 2). Regionally, though, there is a fragmentary record of northeast-trending streamlined landforms near Big Sand Lake (80 km to the NW; Figure 3-1) (Trommelen et al., 2010; Trommelen et al., 2012a), and 40-60° trending striae, grooves or chattermarks (40 km to the SW and 65 km to the NNW; Trommelen et al., 2012a). It appears that, although generally subtle, an inherited signature of this old northeast dispersal may still be recognizable in the regional surface till. If so, the fact that northeast-travelled subglacial detritus is present at Sosnowski Lake indicates that the northeast ice-flow phase must have covered most, if not all, of the study area - something that was suspected but was not considered to be definitive based solely on the landform and striation records. Inherited clast dispersal indicates that the mid-age south-southwest ice-flow phase must have been sliding-controlled with limited quarrying, or involved relatively weak/short-lived subglacial deformation that moulded (or eroded?) the surface but did not affect the underlying till composition. This interpretation implies that a pre-existing till sheet may have been common throughout the study area prior to the Late Wisconsinan glaciation (as would be likely according to the regional river stratigraphy; Dredge and McMartin, 2011; and others).

3.6.2 Western Great Island, Great Island Glacial Terrain Zone

In contrast to the Sosnowski Lake GTZ, the Great Island GTZ (Figure 2-9) consists of thin, generally featureless till overlying bedrock, but sparse fragments of streamlined landform flowsets trending southeastward (younger and older sets) and southward, and a wide range of orientation in field-based ice-flow indicators (60-235°) are still present. GTZ analysis indicates that it is difficult to delimit the relative degree of inheritance within this GTZ owing to a general lack of flowsets, but that the complex field-based ice-flow indicator data suggests inheritance may be high (Chapter 2).

Of particular interest within this palimpsest-type GTZ is the patch of carbonate-bearing till and calcareous glaciolacustrine (?) sediments overlying western Great Island (Figure 3-4). The carbonate clasts are presumed to be derived from the Palaeozoic Hudson Platform (Figure 3-10), which is a minimum distance of 90 km to the east-southeast, 105 km to the east or 130 km to the northeast. Possible fast ice-flow (ice stream) Boothia-type (Dyke and Morris, 1988) westward dispersal of carbonate clasts across Manitoba has been invoked south of the study area (Dredge, 1988; Kaszycki et al., 2008). While Boothia-type dispersal could explain how a till in the Great Island area has a concentration of almost 18% soft, friable carbonate clasts, it is unlikely that these isolated sites represent young Late Wisconsinan transport. Though current ice-flow mapping to the south and southeast is poor, there are no confidently-mapped regional streamlined landforms (Trommelen and Ross, 2010) or young field-based ice-flow indicators oriented to the west or west-northwest in the study area or anywhere west of Churchill, Manitoba (Klassen and Netterville, 1985; Dredge and Nixon, 1986; Klassen, 1986). There is a strong old southwest-trending ice-flow phase that formed roches moutonnées in the field area (Campbell et al., 2012; Trommelen et al., 2012a), but if carbonate dispersal were attributed to this phase the minimum transport distance increases from 90 to 130km.

3.6.2.1 Palaeoglaciological Implications

The data seem to indicate carbonate-bearing till in Great Island is a remnant, inherited, outlier of a penultimate till sheet, evidenced by a lack of corresponding geomorphological ice-flow indicators, and the lack of mapped carbonate-bearing till closer to the source area in any direction (Dredge and Nixon, 1981a, 1982a; Trommelen and Campbell, 2012a, c). These sediments were likely preserved (inherited) within a low-erosion sticky region (Chapter 2), with only minor reworking throughout subsequent differently-oriented ice-flow phases that are known to have affected the area (Chapter 2). The elevated concentration of Dubawnt red/purple erratics at till sample site one (Figure 3-4), within the carbonate-bearing till, indicates either overprinting of a pre-existing till high in Dubawnt erratics, or downward overprinting by a younger high intensity/long duration ice-flow phase from the north-northwest. This area of high-inheritance till verifies the palimpsest-type, high inheritance classification of the GI GTZ which was based on the 'pristine' nature of bedrock topography, lack of subglacial landforms and complex field-based ice-flow indicator record (Chapter 2). Much of the presumably more-extensive older till sheets, with both higher carbonate and Dubawnt clast concentrations are presumed to have been completely overprinted.

3.6.2.2 Calcareous glaciolacustrine sediments

The unusual presence of calcareous clayey-silt sediments at high elevation (Figure 3-4) in this area also requires an explanation. (Trommelen et al., 2010). All surficial glaciolacustrine sediments in the study area have previously been interpreted as belonging to glacial Lake Agassiz (Dredge, 1982, 1983; Klassen, 1983). In the Great Island area, however, we suggest that the sediment composition and isolated spatial character challenge this current view. If the calcareous finer-grained sediments were deposited within the late deglacial Lake Agassiz, one must question why this type of young sediment is not widespread in the study area. Furthermore, to support that these sediments were actually deposited in glacial Lake Agassiz, it would be necessary to invoke a remarkably different late deglacial ice-margin configuration, oblique to local esker orientations, than has been currently published (Dredge, 1982, 1983; Klassen, 1983). Instead, another possible interpretation may be that the calcareous clayey-silt sediments are a remnant isolated deposit – related to an older glacial lake for which the ice margin and glacial lake configuration would have favored delivery of carbonate sediments into the area. The minimum extent of the LIS during the Mid-Wisconsinan is largely unknown (Dredge and Thorleifson, 1987), and though evidence is lacking, it remains possible that the area was deglaciated.

3.6.3 Caribou River and Big Sand Lake Glacial Terrain Zones

The surface till in the Caribou River (warm-based relict-type) and Big Sand Lake (palimpsest-type) GTZs (Figure 3-1) contains a strong landform imprint indicative of old SE-trending ice-flow dynamics and an even older northeast-trending flowset that has been subsequently modified in isolated patches/spots by southward-trending streamlined landform flowset fragments (Chapter 2). GTZ analysis indicated that these regions likely contain a high level of inheritance from early ice-flow phases, which should be reflected in the surface till composition. Indeed, these regions contain elevated concentration of Dubawnt clasts (Figure 3-9). While this may simply signify that these GTZs are closer to the source area, the highest concentration of Dubawnt clasts 45-50 km further south (Figure 3-9) implies that, at an earlier point in time, the Dubawnt clast concentration may have been higher across the region. If so, the 'removal' of Dubawnt clasts from the Great Island and Sosnowski Lake GTZs required some amount of entrainment, comminution or dilution. Though the final Dubawnt concentrations in the surface till, as seen today, are probably palimpsest, the transport distance of 350-600 km indicates that at one time southeast ice flow was vigorous and/or sustained over a long period of time (Shilts et al., 1979; Kjaer et al., 2003; Hildes et al., 2004).

3.7 SECONDARY MODIFICATION OF A PRE-EXISTING TILL SHEET

The degree of compositional overprinting, inheritance or formation of a new till type (entrainment, comminution, dilution) is largely dependent on the evolution of basal conditions over time (

Figure 3-12). Near the inner core regions of ice sheets, as in northeastern Manitoba, we have suggested that changes in basal conditions were not widespread, but instead extremely variable over time and space at a local scale, leading to patchy low-erosion conditions ('sticky regions') which favored preservation of a fragmentary record of non-coeval and sometimes contrasting warm-based (deforming) conditions (Chapter 2). A fragmentary subglacial record was also noted in Scotland (Finlayson et al., 2010), where the authors envisioned a mosaic of deforming sediments and stable spots (*sensu* Piotrowski et al., 2004) enabled preservation of older sediment-landform assemblages amongst younger assemblages. The 'regional till

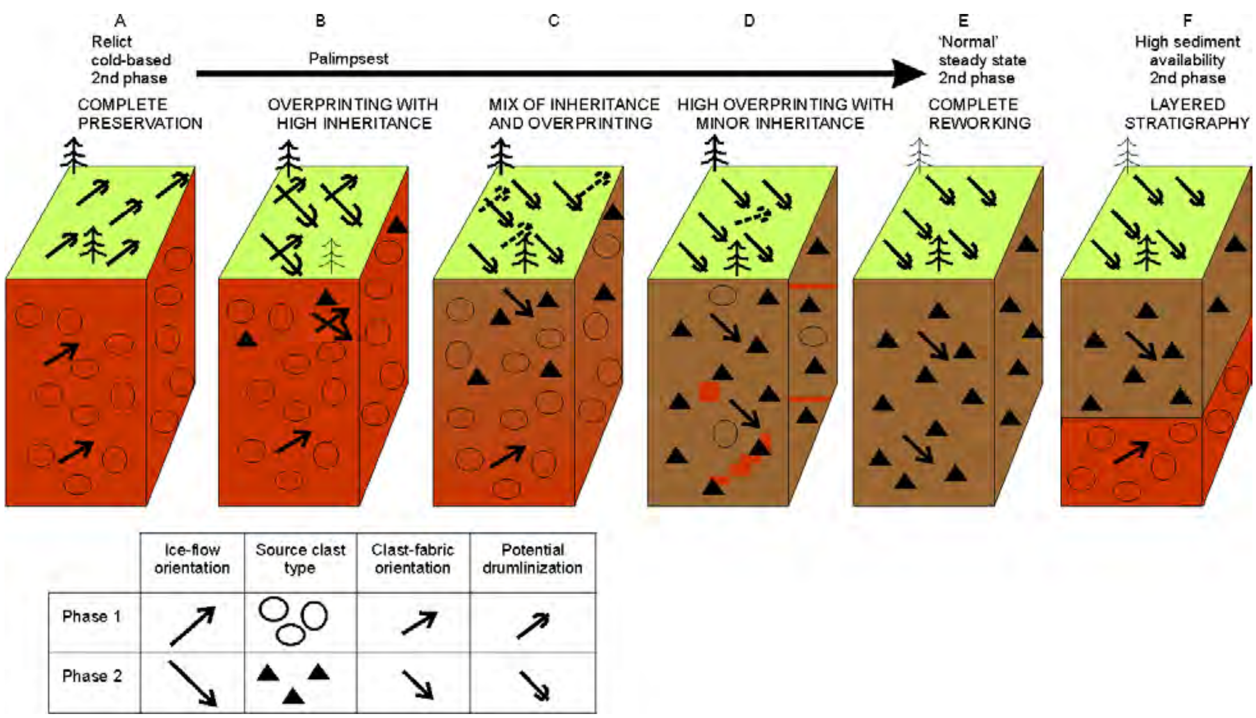


Figure 3-12. Possible combinations of composition, clast-fabric orientation (a-axis) and subglacial landform orientation that may arise when an ice-sheet advances over a pre-existing till sheet. This re-advance could occur during a new glacial cycle, or due to a deflection in ice-flow orientation over time within one glacial cycle. If the second ice-flow phase is cold-based, a relict till-type (A) may be preserved (all inherited characteristics) (relict landscapes, Harbor et al., 2006; Kleman et al., 2008). Contrastingly, if the second ice-flow phase is warm-based, moving at a sufficient ice-flow velocity and has sufficient sediment availability and effective water pressure, a new homogenous till-sheet (E) may form (complete overprinting due to comminution and/or dilution). The intermediate till-types (B-D) are transitional hybrid tills that may form under conditions between the two end-member tills. A 'layered' till stratigraphy (F) could develop in areas where a high effective porewater pressure in fine-grained tills (Lian et al., 2008)

led to fast-flowing ice streams with long-distance transport of clasts, such as a possible ice stream presumed to have transported carbonate >470 km across the Canadian Shield in Manitoba, Canada (Dredge, 1988; Kaszycki et al., 2008).

sheet' in the study area exhibits significant heterogeneity owing to variable levels of inheritance and overprinting in accordance with the subglacial bed mosaic proposed for the area (Chapter 2). Surface landforms may be related to either the younger or older glacial phases (Kleman and Stroeven, 1997; Stea and Finck, 2001; Kleman et al., 2008; Ross et al., 2009; Knight, 2010) or may be palimpsest (Stea, 1994; Clark and Meehan, 2001; Finlayson et al., 2010). Each till type (Figure 3-12) may be considered a 'hybrid till' (Lian and Hicock, 2000; Stea and Finck, 2001; Lian and Hicock, 2010; Seaman et al., 2011), meaning that previously-deposited tills experienced some level of reworking during subsequent ice-flow events, resulting in a till with a mixture of inherited and overprinted characteristics (e.g. Seaman et al., 2011; Tremblay et al., 2011). The hybrid nature of the till is most evident at western Great Island, where the carbonate-bearing till (hybrid-till type C; Figure 3-12) contains subglacial detritus from far northwest (Dubawnt; >350 km) and east or southeast (carbonate; >100 km) of the study area, in addition to locally-derived clasts. In most of the Sosnowski Lake palimpsest-type GTZ, the till is heterogeneous with patches of hybrid till type C and D (Figure 3-12) intermixed. Outside of the study area, near Big Sand Lake (Figure 3-1) where northeast-trending crag-and-tail features and other older streamlined landforms are preserved (Chapter 2), we would expect to find patches of till type B (Figure 3-12). In the Caribou Lake relict-type GTZ, where only one non-deglacial southeast-trending streamlined landform flowset is present oblique to esker ridges (Chapter 2), we would expect to find mostly till type A.

3.8 THE RELATION TO EROSION INTENSITY

Till composition, combined with regional knowledge (source areas, subglacial landform characteristics and field-based ice-flow indicators) (Figure 3-13A) as a proxy for the evolution of erosion intensity, may help to identify transitory sticky and stable spots (subglacial bed mosaic; Trommelen et al., 2012a) beneath an ice sheet. An understanding of the transport direction and nature (local vs. continental) of these indicator clasts can be correlated to relative ice-flow phases in the study area, and combined with the spatial pattern of subglacial landforms – allowing for creation of a relative erosion intensity map (Figure 3-13B). This is a somewhat subjective process, but follows a few key guidelines. Firstly, increased concentrations of inherited subglacial detritus necessitate a lower level of erosion relative to adjacent areas with no inheritance. Secondly, inherited dispersal directions correlated to progressively older ice-flow phases also necessitate a progressively lower level of relative erosion.

Applying this methodology to the field area, very low erosion is interpreted to have occurred at the calcareous sediment outlier in western Great Island, post the oldest SE, SW and NW ice-flow phases. Low erosion, post a major SE ice-flow phase, is interpreted for the areas of higher Dubawnt concentration in the Big Sand Lake and Caribou River GTZs. Low erosion, post a NE ice-flow phase and during a subsequent migration of ice-flow from ESE to WSW, is interpreted for the region of NE to E local clast dispersal in the Great Island GTZ, where a presumably regionally-extensive higher concentration of Dubawnt erratics have been mostly overprinted, but inherited dispersal to an old ice-flow orientation is still present. Areas of Rogen moraine are also assigned as low relative erosion, owing to the presence of inherited clasts (Campbell et al., 2012) that indicate formation from a pre-existing sediment sheet. Lastly, relatively moderate erosion during the Late Wisconsinan is interpreted for the remainder of the field area

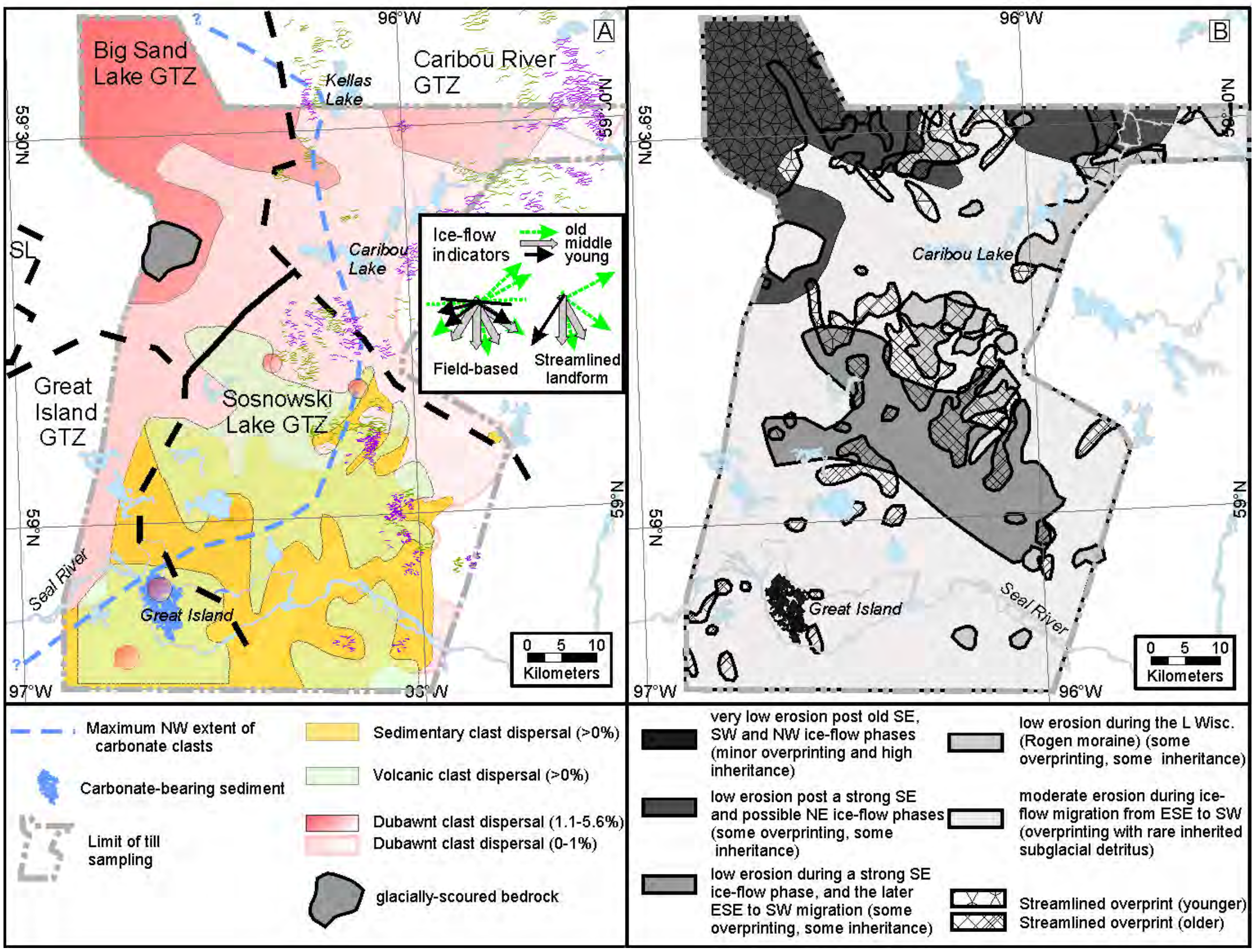


Figure 3-13. Till composition map (A) and relative erosion intensity map (B). Till dispersal patterns include short-distance amoeboid dispersal of locally-derived sedimentary and volcanic clasts, and long-distance dispersal of Dubawnt Supergroup and Palaeozoic carbonate clasts. As predicted, till composition varies across the field area and is different within each GTZ (delimited using

streamlined landform and field-based ice-flow indicator characteristics, Trommelen et al., 2012a). Relative erosion intensity is mapped based on the relative concentration and dispersal directions of clasts within the till, combined with ice-flow phases presented in the text.

where comminution/dilution is more extensive and the till has lower concentrations of Dubawnt and carbonate clasts (Figure 3-9, Figure 3-10). If the strength of overprinting is known, streamlined landform phases could also be added to this map.

3.8.1 Implications

While previous researchers have interpreted the subglacial landscape in northeastern Manitoba as deglacial (Dyke and Prest, 1987a; Dyke and Dredge, 1989; Dredge and Nixon, 1992; Dyke, 2004), the preservation of transitional hybrid tills likely required slow, sluggish ice-flow ('sticky) regions during the Late Wisconsinan deglaciation, to allow for low erosion and/or minimal reworking. Patchy inheritance surrounded by areas of higher reworking/erosion may have occurred as part of a fragmented spatio-temporal subglacial deformation mosaic (van der Meer et al., 2003; Piotrowski et al., 2004; Piotrowski et al., 2006; Knight, 2010; Trommelen et al., 2012a) at a landform scale. Hybrid tills have been commonly recognized beneath the Appalachian Ice Complex in Nova Scotia and New Brunswick (Stea and Finck, 2001; Seaman et al., 2011), but to date have not been identified within the Keewatin Sector. Our work supports the regional low-erosion designation by Sugden (1978), and suggests that the presence of a fragmented patchy glacial landscape with different glacial histories, ages and inheritance levels has likely been under-estimated for core regions of the LIS. Alternatively, high inheritance (low erosion) patches may be evidence for localized cold-based ice, as part of a frozen-thawed mosaic as proposed by Kleman and Borgström (1994) and Kleman *et al.* (1999; 2008).

3.9 Conclusion

Local and continental-scale clast-dispersal analysis has been conducted in northeastern Manitoba, Canada, to help address how the final composition of a surface till, overlying low-relief bedrock, is related to the evolution of palaeo-subglacial conditions in areas of complex ice flow near the core regions of ice sheets. The field area preserves evidence of at least two contrasting styles of dispersal: i) long distance dispersal of carbonate clasts (~100 km long) and Dubawnt red erratic (~350-600 km long), and ii) shorter distance amoeboid dispersal of locally-derived clasts in the Great Island area (>15 km) that includes inherited NE to SE dispersal, despite later deflection and migration of ice-flowing ESE through to WSW during the late deglaciation. Dispersal of till clasts is heterogeneous within palimpsest-type Glacial Terrain Zones (GTZs), and inherited dispersal does not locally match the expected glacial history gleaned from streamlined landform flowsets in the field area. This highlights the importance of clast dispersal in tills as an additional proxy necessary to fully reconstruct palaeo-ice sheet dynamics.

Clast-dispersal analysis, integrated with Glacial Terrain Zone (GTZ) partitioning, can be used to further strengthen the reconstruction of palaeoglaciological evolution. We suggest there was/were a pre-existing till sheet(s) in northern Manitoba, which underwent variable spatio-temporal reworking (comminution and dilution), resulting in a spatial mosaic of hybrid tills with different levels of inheritance and overprinting. High inheritance patches were likely preserved, or variably re-worked, under transitory low-erosion ‘sticky regions’, perhaps as part of a subglacial bed mosaic. If spatio-temporal variations in erosion, transportation and deposition intensities occurred at the ice-bed interface throughout the evolution of an ice-sheet, the regional surface till should reflect this. A fragmented mosaic of varying till composition can then be used as part of a regional dataset to make spatial erosion/reworking intensity maps. If this subglacial landscape mosaic can be identified in other core regions of ice sheets, its analysis will greatly improve models of regional spatio-temporal palaeoglaciological evolution.

Chapter 4 Rogen Moraine in northern Manitoba, Canada: characteristics and formation as part of a subglacial bed mosaic near the core regions of ice sheets

4.1 Overview

Rogen moraine are enigmatic glacial landforms for which different models, with contrasting paleoglaciologic implications, have been proposed to explain their formation. Some of the largest fields of Rogen moraine overlie low-relief topography of the Canadian Shield, in northern Manitoba, Canada, where they have been the least studied. We investigate the characteristics of these transverse-to ice-flow ridges at landscape-scale (mapping and spatial analysis) and landform-scale (internal structure using high-resolution shear wave (S-wave) seismic reflection surveys, sedimentological characteristics, and clast-fabric analyses). Two main types of Rogen moraine are recognized: 'pristine', high amplitude straight-crested ridges and modified subdued 'drumlinized' ridges. Rogen moraine in northeast Manitoba consist of massive, matrix-supported till at surface, which is similar in matrix texture and composition to the regional till sheet. Compositional inheritance in the till, as well as in the landscape record at all scales, indicates Rogen moraine likely required a sufficiently thick pre-existing sediment sheet from which to form. The orientation of Rogen moraine in northern Manitoba suggest these landforms were likely generated during a sticky, slow-flowing but warm-based 'mid' SSE- to SSW-trending phase of regional ice flow, and then subsequently preserved during deglaciation. Lastly, two seismic profiles reveal subparallel-to surface layered stratigraphy with minor folding but no internal stacking, unconformities or faulting.

Given the regionally-complex ice-flow history, palimpsest subglacial landscape, short-distance amoeboid dispersal patterns, and variation in Rogen ridge a-axis orientation and internal clast-fabric data, we suggest the Keewatin Laurentide Ice Sheet Rogen moraine are likely palimpsest subglacial landforms, generated primarily from pre-existing sediments in areas where subsequent low-erosion phases limited till production and led to high glacial landscape inheritance. The impetus for Rogen ridge generation may have been subglacial instability due to transient velocity/stress regimes within a landscape-scale subglacial bed mosaic, combined with rough bedrock topography (landform scale). Some sticky regions (cold-based or dewatered patches resisting basal stress) may have remained throughout deglaciation, while others expanded during ice-sheet thinning in early deglaciation - allowing for preservation of ‘pristine’ Rogen ridges. As part of this interior subglacial bed mosaic, where the availability of subglacial meltwater increased during deglaciation, the mosaic-scale may have decreased. Subsequently more local changes beneath the progressively more-active warm-based ice may have initiated patchy drumlinization of some of the Rogen moraines. Where ice-flow velocity increased substantially, the pre-existing landscape was overprinted and in some areas streamlined terrain was generated. We confirm a link between Rogen moraine and widespread cold-based subglacial dynamics, but for reasons of preservation and not formation.

4.2 INTRODUCTION

The formation and paleoglaciologic implication of Rogen moraines is a question, posed throughout the last fifty years, that still remains largely unsolved (Lundqvist, 1969; Hättestrand and Kleman, 1999; Moller, 2006). These anastomosing to curved subglacial transverse ridges are present in Canada (Prest et al., 1968; Aylsworth and Shilts, 1989; Clark et al., 2000; Jansson, 2005; Dunlop and Clark, 2006b; De Angelis, 2007; Trommelen and Ross, 2010), Fennoscandia (Sollid and Sorbel, 1984, 1994; Hättestrand, 1997; Sarala, 2006), Ireland (Knight and McCabe, 1997; Clark and Meehan, 2001; Knight, 2010) and Scotland (Finlayson and Bradwell, 2008; Finlayson et al., 2010). Numerous competing hypotheses have been proposed for the formation and preservation of Rogen moraine, some of which imply a single formative mechanism (Fisher and Shaw, 1992; Hättestrand, 1997; Hättestrand and Kleman, 1999; Chapwanya et al., 2011), while others invoke multiple stages of development dependent on the landscape association and morphologic expression (Lundqvist, 1997; Moller, 2006; Stokes et al., 2008).

Rogen moraine, a geomorphic term without genetic connotation, was first defined by Lundqvist (1969) to include only transverse ridges that have a gradual up-and/or down-ice flow-direction transition to drumlins and/or a non-transitional lateral shift to streamlined terrain (c.f. Alysworth and Shilts, 1989). While this subglacial landscape association was noted in northern Manitoba, we suggest that it is a lack of association with ice-marginal landforms that differentiates Rogen moraine from other types of transverse or ribbed moraine that are only associated with hummocky moraine (Fisher and Shaw, 1992; Marich et al., 2005), ice-marginal proglacial landforms (de Geer moraines, Linden et al., 2008; controlled moraines, Moller, 2010), or that are valley-confined (Finlayson and Bradwell, 2008; Linden et al., 2008). This definition does not inherently imply that all ribbed moraine are produced by different processes (as per Moller, 2006; Linden et al., 2008; Moller, 2010), as there may indeed be similar underlying mechanics (c.f. Dunlop and Clark, 2006a; Dunlop et al., 2008; Chapwanya et al., 2011), but simply that each scenario should be studied individually first. As such, this paper uses the term Rogen moraine, defined as ribbed moraine ridges that are part of a subglacial landscape assemblage and are situated on regionally low-lying terrain without topographic constraints.

The presence and orientation of Rogen moraine ridges have been used to reconstruct paleo-ice flow dynamics or to infer basal conditions in numerous studies (Kleman et al., 1997; Knight and McCabe, 1997; Kleman et al., 1999; Jansson et al., 2002; Raunholm et al., 2003; Van Landeghem et al., 2009). As ridge formation is still contentious (Hättestrand and Kleman, 1999; Dunlop and Clark, 2006a; Moller, 2006; Chapwanya et al., 2011), inferences about basal thermo-mechanical conditions (e.g. link to widespread cold-based ice, Hättestrand, 1997; Hättestrand and Kleman, 1999) remain uncertain and largely untested. To date, most field studies have focused on small areas of valley-confined or ice-marginal ribbed moraine ridges (Fisher and Shaw, 1992; Lundqvist, 1997; Marich et al., 2005; Moller, 2006; Linden et al., 2008; Moller, 2010; Sutinen et al., 2010). Only a few studies have looked at regionally-extensive Rogen moraine situated in areas that are not topographically-constrained, in Finland (Sarala, 2006) and in Quebec (Cowan, 1968; Bouchard, 1989) and Nunavut (Stokes et al., 2008), Canada. Some of the largest fields of low-relief topography Rogen moraine in the world are situated within close proximity (<250 km) to the paleo Keewatin Ice Divide (KID), in Nunavut and northern Manitoba, Canada (Prest et al., 1968; Aylsworth and Shilts, 1989; De Angelis, 2007; Kleman et al., 2010); yet they have not been studied. Field investigations in this area thus hold great potential for advancing our understanding of this unique landscape.

In this paper, we present the first study that combines remotely-sensed mapping of subglacial landforms across northern Manitoba, Canada, with field-based data from Rogen moraine ridges in northeast Manitoba. We then review current theories of Rogen moraine formation and assess how our new data supports or challenges these theories. Incorporation of Rogen moraine data within a regional paleoglaciological context (Chapters 2 and 3) then allows us to consider how and when Rogen moraine may have been emplaced, preserved and/or overprinted.

4.2.1 Current debate

Hypotheses to explain the generation of Rogen moraine vary in terms of ice-flow dynamics (extensive or compressive), basal thermal regime (cold/warm interface or warm-based), water availability (sliding ice to sticky ice) and theorized obstacles at the base of the ice sheet (rough bedrock topography, basin and hill). Most hypotheses attempt an explanation for the formation of Rogen moraine, as well as the overprinting and spatial relationship to streamlined (drumlin and flute) landforms. In this manner, a continuum of subglacial deformation landforms from ribbed moraine (slow velocity) to drumlins (medium velocity) to flutes (fast velocity) is often denoted (Aario, 1977; Boulton, 1987; Rose, 1987; Dyke and Morris, 1988; Lundqvist, 1989, 1997; Knight et al., 1999; Moller, 2006; Bradwell et al., 2008; Fowler, 2009). Contrastingly, mathematically modeled Rogen moraine generation (Dunlop et al., 2008; Clark, 2010; Chapwanya et al., 2011) does not easily explain a transition between subglacial landform-types, but instead better-addresses fundamental questions regarding wave-length consistency, ridge spacing and relief amplification. Chapwanya et al (2011) were able to demonstrate that the theory can generate ribbed moraine-like features, when adding a small subglacial water flux, but that drumlins are harder to generate. Similar bed-ribbing hypotheses are frequently used in mathematics to explain instability-related ripple and dune formation and washboard formation on gravel roads (Taberlet et al., 2007; Fourriere et al., 2010), but remains largely untested using field-based geology. The formation of Rogen moraine has also been included as part of the subglacial megaflood hypothesis (Shaw, 2002).

Of all the hypotheses proposed for Rogen moraine formation, at least three are still frequently used in the recent literature. The first involves *fracture and extension* of the sediment sheet at the cold/warm-based interface, where sediment 'ribs' are detached from the cold-based non-sliding zone and displaced down-ice through warm-based sliding and or folding (Hättestrand, 1997; Hättestrand and Kleman, 1999; Sarala, 2006). The second hypothesis suggests Rogen (ribbed) moraine are formed by warm-based *subglacial modification* of pre-existing ridges, which may have been preserved by a long period of cold-based glaciation (Lundqvist, 1969; Boulton, 1987; Lundqvist, 1989, 1997; Moller, 2006; Finlayson et al., 2010; Moller, 2010), or of the sediment sheet itself arising from a natural instability (Aario, 1977; Hindmarsh, 1998; Dunlop et al., 2008; Fowler, 2009; Clark, 2010; Chapwanya et al., 2011). The third hypothesis suggests flowing warm-based ice locally undergoes compressive flow, and Rogen (ribbed) moraine are formed by *shearing and stacking* as the ice-flow velocity slows due to an adverse basal topographic slope (Shaw, 1979; Bouchard, 1989), patches of dewatered stiff till (Boulton, 1987; Linden et al., 2008; Stokes et al., 2008) or cold-based ice (Sollid and Sorbel, 1984; Dyke and Morris, 1988; Dyke et al., 1992; Finlayson and Bradwell, 2008; Linden et al., 2008).

Table 4-1 presents the requirements, conclusions and a summary of each hypothesis. Depending on the chosen hypothesis, interpretations of the paleoglaciological evolution (ice-flow dynamics, thermal regime, hydrology, and rheology) may be quite diverse.

Table 4-1. Hypotheses to explain the formation of Rogen moraine

Hypothesis	**Fracture and Extend**	**Subglacial Modification - pre-existing ridges or hummocks**	**Subglacial Modification - natural instability**	**Shear and Stack - dewatered stiff till patches or cold-based ice patches**
Authors	Hattestrand and Kleman 1999, modified by Sarala 2006	Boulton 1987, Lundqvist 1997, 1989, Knight and McCabe 1997, Moller 2006, Finlayson et al. 2010	Aario 1977, Hindmarsh 1998, Dunlop et al. 2008, Fowler 2009, Chapwanya et al. 2011	Shaw 1979, Bouchard 1989, Linden et al. 2008; Stokes et al. 2008
Thermal Regime	Cold/Warm interface	Warm	Warm-based	Warm-based
Glacial Stage of Formation	deglaciation - as interface between warm- and cold-based ice moves inward	over multiple glacial episodes	any time	any time
Ice Flow Velocity	increases down-ice	increases down-ice	constant/sheet flow	decreases at Rogen site
Ice Movement	Fast-flow near margins causes extensive drawdown and sliding movement	Extensive/sliding	deformation and sliding as guided by till rheology and effective pressure	Compressive/sticky spot
Sediment Source	pre-existing drift sheet together with the lower part of the ice sheet	pre-existing ridges or rough topography	pre-existing drift sheet; all sediment was effectively transmuted to the ridges	pre-existing drift sheet OR continued subglacial deformation conveyor-belt stacking of debris-rich ice
Synopsis	At the transition from proximal frozen (non-sliding) conditions to distal melting (sliding) conditions, **high tensional stresses and extensional ice flow** will occur, as the basal ice velocity increases across the boundary of basal thermal regime. These tensional stresses lead to detachment and **'boudinage-like' fracturing** of a pre-existing drift sheet into ribbed moraine.	1. **Pre-existing transverse moraine ridges,** originally deposited from ice-cored moraines, streamlined landforms or thrust ridges 2.**Reshaping**: sheath folds and shear bands indicate **high strain at deformation"** "compressional shortening of sediments would be combined with horizontal flattening and extension, gradually transforming a transverse element into longitudinal landforms	Generated during basal sliding as a **natural instability** is initiated **at the ice/substrate boundary**. The instability takes the form of transverse two-dimensional rolls, which cause the resultant **finite-amplitude waveforms** (ridges). Factors that may initiate instability include substrate thickness, effective water pressure, textural variability, velocity variability, etc.	An obstacle to glacier flow* causes the development of **shear planes** in the basal part of the glacier through **compressional ice flow** and these, in turn, lead to the **stacking of slices** of debris-laden ice. Early models suggest this process repeats itself near the ice margin, where the stagnating margin creates an up-ice expansion of the obstacles to glacier flow *whereby the obstacle could be warm/cold interface, dewatered stiff till, bedrock topography
Sediment Provenance	Sediment within may have no relation to the landform	Sediment within may have no relation to the landform; composed mainly of ablation till, or proglacial sediments. May have a till carapace.	Sediment sourced from up-ice of the landform (transverse to ribbed moraine)	Sediment sourced from up-ice of the landform (transverse to ribbed moraine)
Relationship to other landforms	Variably overprinted by drumlins; not possible to transition back to ribbed.	Transitional to drumlins which are in turn transitional to ribbed moraines, etc	Transitional to drumlins	Variably overprinted by drumlins; overprints MGSL in the Dubawnt Ice Stream

We suggest that current questions surrounding Rogen moraine formation include:

- ***What*** is the sediment source (pre-existing till/regolith/glaciofluvial sheet or locally-derived directly up-ice during ridge formation)?
- ***How*** is thick till generated in Rogen moraine fields (Dunlop et al., 2008)?
- ***Where*** do Rogen moraine form underneath ice sheets?
- ***When*** does Rogen moraine generation(s) occur during the regional glacial history?
- ***How*** are Rogen moraine ridges generated, sometimes modified and ultimately preserved to various degrees during deglaciation?
- ***Why*** do Rogen moraine fields form (or are preserved) in some areas and not in others?
- ***What*** are the basal thermo-mechanical conditions necessary for their formation?

Various studies have undertaken to answer these questions, but many uncertainties persist. Here, we argue that a multi-faceted approach involving fieldwork in these large fields, in combination with remote sensing and laboratory analyses is necessary to test existing models, and that this is a critical step if we are to develop a unifying theory of Rogen moraine.

4.3 STUDY AREA

Northern Manitoba (58°-60, Figure 4-1) is bounded by Hudson Bay in the east, and extends west to the Saskatchewan border, rising in elevation to about 500 m above sea level (a.s.l.). The entire region forms a uniform, gently-sloping plain with subdued topographic features. The northeast region falls within the continuous permafrost zone, while the remainder of the areas is within the extensive discontinuous permafrost zone (Sladen, 2011). Rogen moraine is common in northeast and northwest Manitoba, but absent in the north-centre of the province. Both pristine and drumlinized Rogen moraine are mapped (Figure 4-1) (Trommelen and Ross, 2010), the latter that have been overridden by actively-flowing ice, resulting in partial streamlining of the ridge surfaces.

The Great Island field area (Figure 4-2) is low-lying and elevation varies mainly from 140 to 220 m. The region is a mix of till veneers, blankets, streamlined, ridged and hummocky terrain, with scattered organic bogs. Below 180 m a.s.l. the area has been briefly inundated by the post-glacial Tyrrell Sea (Trommelen et al., 2010), which has caused minor surface erosion and masking of subglacial sediment below ~160 m a.s.l. Linear networks of dense conjugate ridges are also present within the Caribou River GTZ (Figure 4-1 and 4-3), and appear to be spatially-associated with swaths of Rogen moraine and streamlined landforms. The composition of these networks is variable (Dredge and Nixon, 1992) and poorly-studied. At the one site visited during this study, we encountered a boulder-covered (0.5 to 4 m diameter) low-lying (<8 m) hummocky to ridged moderately-sorted diamict landscape (Figure 4-3). There

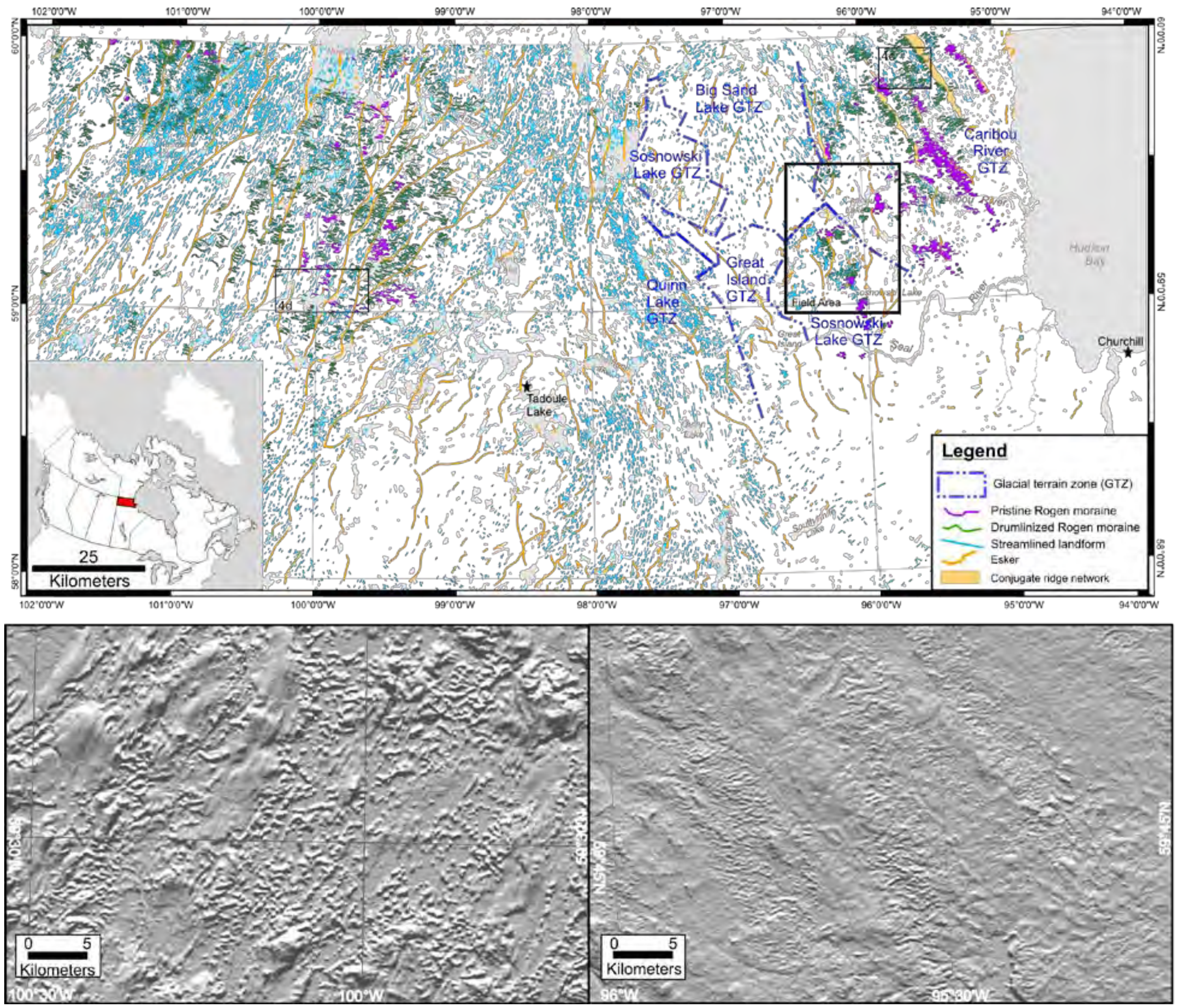


Figure 4-1. Rogen moraine, streamlined landforms and esker ridges in northern Manitoba, as mapped from Landsat EMT+ 7 and SRTM in Trommelen and Ross (2010) and updated using SPOT 4/5 imagery. Glacial terrain zones (GTZ) have been delimited in northeastern Manitoba (Trommelen et al., 2012b) whereby each GTZ reflects a different subglacial landscape evolution and assembly history from the adjacent terrain. Crevasse ridge networks are depicted as mapped by Dredge and Nixon (1986). Shuttle Radar Topography Mission (United States Geological Survey, 2002) hillshades, at the same 1:250 000 scale, shown below depict the different character of the subglacial landscape in the west, and east, respectively. Rogen moraine in the west are interspersed amongst drumlins and drumlinoid ridges of varying scale. Rogen moraine in the east are situated in alternating swaths with drumlinoid ridges. In both cases, we consider streamlining to be a secondary process.

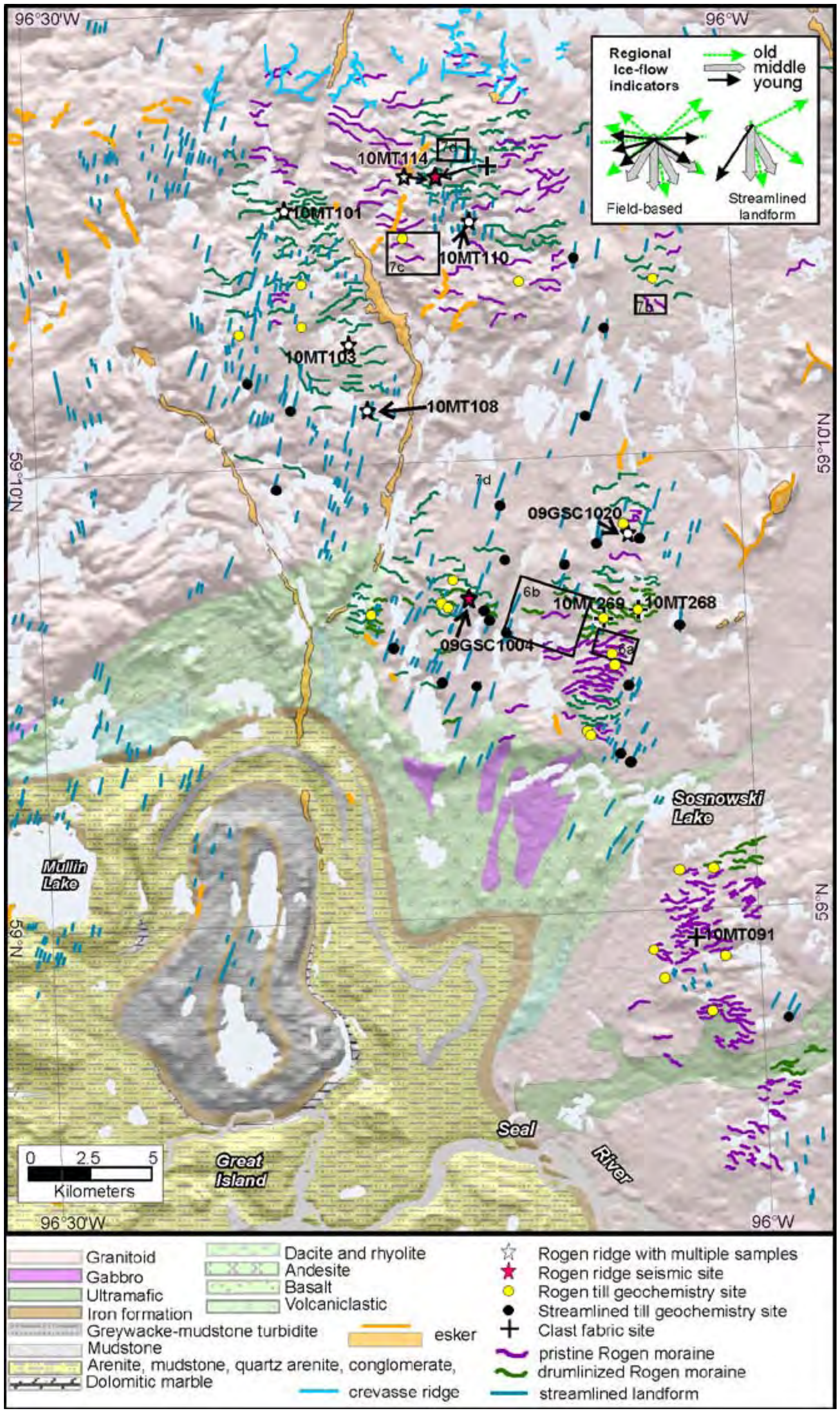


Figure 4-2. Great Island field area, which includes the southernmost Rogen moraine in northeast Manitoba. Fieldwork completed in the area includes hand sampling, shallow seismic on ridge axes and clast fabric analyses from lee-slopes. The background delimits the 1:75 000 scale bedrock map (Anderson et al., 2009) draped on a 90m resolution hillshaded SRTM DEM (United States Geological Survey, 2002). Seismic site MT213 is located another 12 km NW of this figure, and is denoted on Figure 4-4.

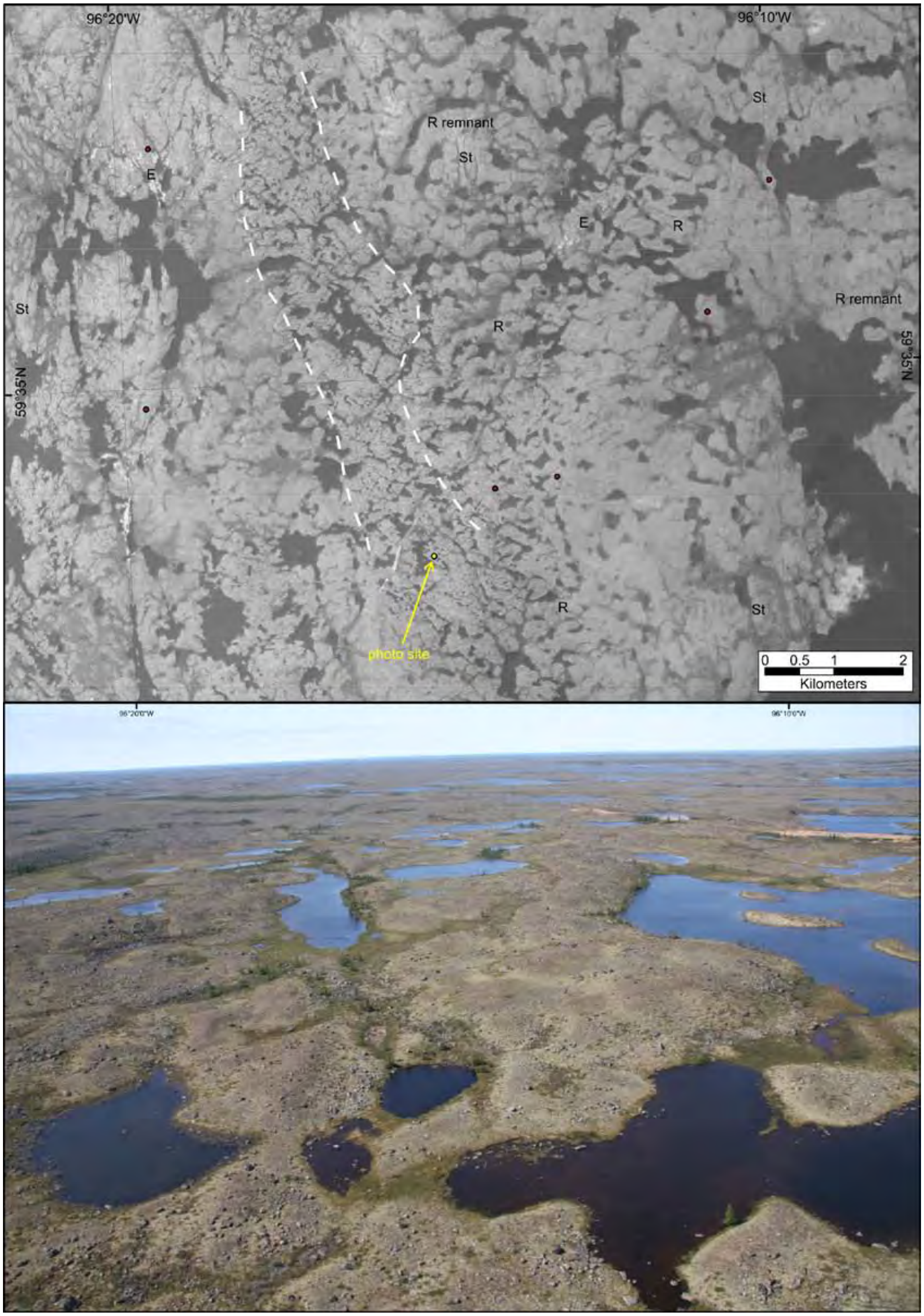


Figure 4-3. Satellite image (SPOT4, Geobase, 2005-2010) of Rogen moraine fields (R), overprinted Rogen moraine ridges (R remnant), streamlined terrain (St), eskers (E) and a conjugate ridge network in northeastern-most Manitoba (within the white dashed lines). The lower photograph is taken from a helicopter, at the location denoted. A younger esker (white on SPOT, orange on photo) overprints the network.

is no obvious ice-marginal landsystem in the field area, though a large lobate streamlined landform flowset outlines the deglacial Quinn Lake Glacial Terrain Zone 25 km to the west (Chapter 2). Only a few ice-marginal end moraines and subaqueous fans are scattered throughout the Great Island area (Figure 2-4). Eskers are present across northern Manitoba (Figure 4-1), and are part of a large (up to 450 km long) dendritic esker network that may be indicative of large-scale organization of the subglacial drainage pattern beneath this region of the LIS (Boulton et al., 2009), that extended over a significant period of time far back beneath the ice sheet.

The bedrock geology largely consists of Archean basement granitoid rocks (Anderson et al., 2009a; Anderson et al., 2010b). In the Great Island area (Figure 4-2), the Neoarchean Garlinski Lake greenstone belt is comprised of mafic-intermediate volcanic rocks (Sosnowski Lake assemblage) and an unconformably overlying clastic succession (Omand Lake assemblage) that are in turn unconformably overlain by the Paleoproterozoic Great Island Group (predominately sandstones).

4.3.1 Ice-flow history

The glacial record in northeastern Manitoba is complex and reflects multiple ice-flow phases (Chapter 2). A simplified regional generalization (inset on Figure 4-2) indicates that there have been at least three phases of ice flow to both the southwest and southeast and two phases each to the east, east-southeast, south, and west-southwest (Chapter 2). Old rare ice-flow indicators also suggest ice-flow to the northeast (field area) and northwest (documented 40 km to the south) occurred at some point in time. In the Great Island region, Late Wisconsinan ice-flow was deflected clockwise, from southwest (~210°) to the west-southwest, due to migration of the KID. Young, fine striae inscribed at the top of some outcrops are oriented east and east-southeast, and correlate with subaqueous fans deposited during glaciomarine incursion in late deglaciation. These ice-flow phases are based on ice-flow indicators such as striae, grooves, chattermarks, gouges, crescentic fractures, roches moutonnées and whaleback drumlins identified in the field (Trommelen and Ross, 2009; Trommelen et al., 2010; Trommelen and Ross, 2011a, b; Campbell et al., 2012), and streamlined landform flowsets (Chapter 2).

4.3.2 Palaeoglaciology

The subglacial landscape of northeast Manitoba has recently been described as patchy and highly fragmented, characterized as disjoint landscape pieces with internally-consistent glacial histories – termed *glacial terrain zones (GTZs*; Chapter 2). These GTZs were likely formed by multiple phases of ice flow with different flow directions, velocities and intensities - reflecting not only different erosion/transportation/depositional processes over short distances and through time, but possibly also distinct subglacial thermal regime histories. In northeast Manitoba, Rogen moraine ridges were mapped within the relict-type Caribou River GTZ and the palimpsest-type Sosnowski Lake GTZ, but not within the two adjacent GTZ (Figure 4-1). Relict-type GTZ (Chapter 2) is defined as a high inheritance glacial landscape that generally consists of one or more isochronous flowsets whose orientation best matches early, older, ice-flow orientations, with little or no evidence of overprinting. Chapter 2 suggested that these relict-type GTZs likely formed during a subglacially-active (warm-based, with low effective water pressure (e.g. Denis et al., 2009)) phase and then transitioned to sticky regions post initial formation and throughout deglaciation (final preservation). Sticky regions are defined as areas of low erosion and/or reworking created by subglacially-inactive cold-based ice (e.g. Boulton et al., 2001b) or a stiff, dewatered (e.g. Christoffersen and Tulaczyk, 2003; Piotrowski et al., 2004; Meriano and Eyles, 2009) environment. This interpretation is thought to best explain the occurrence of a widespread landscape that does not fit the younger deglacial record of the region. Thus the sediment-landform assemblages within these GTZ are 'inherited' from older ice-flow phases with little modification (overprinting/reworking) by younger and contrasting phases, which are better recorded in other adjacent GTZ. Palimpsest-type GTZ (Chapter 2) may include regions of relict-type subglacial landscape, but also reflects spatio-temporal variations in warm-based ice-flow dynamics throughout the entire glacial history. Sediment-landform assemblages within this type of GTZ contain a mix of older inherited terrain (Chapter 3), and younger partial to completely overprinted terrain. The fact that Rogen moraine are situated within both relict and palimpsest-type GTZ indicates that these landforms may be palimpsest, as was also noted in Ireland (Clark and Meehan, 2001).

4.4 METHODS

4.4.1 Mapping and characterization

The length and orientation of subglacial landforms (pristine and drumlinized Rogen moraine, streamlined landforms, eskers) in northern Manitoba were mapped (Trommelen and Ross, 2010; Trommelen et al., 2012b) using remotely-sensed imagery (Landsat 7 ETM+ and SPOT4/5 satellite imagery, www.geobase.ca), to provide a detailed georeferenced and updated view of the regional subglacial landscape. Further surficial mapping on air photos (1:60 000) was completed for the field area (Trommelen and Campbell, 2012a, b, c, d) and other select areas where digital air photos were available. These maps provide crucial information on the geomorphological shape of Rogen ridges, their spatial distribution, the relationship between Rogen moraine and other subglacial landforms, and relative drift thickness.

4.4.2 Rogen field investigation

If drumlinoid ridges and Rogen moraine are formed under different ice-flow velocity regimes, and possibly by different subglacial dynamics, it is possible these landform types would contain subglacial detritus sourced from different distances up-ice. To address this question, we collected surface till samples (25 to 130 cm depth) from the crests and steeper lee-sides of individual Rogen moraine ridges, as well as from different geomorphic expressions (pristine Rogen, drumlinized Rogen and streamlined terrain; Fig. 2). No natural sections were encountered, and the presence of permafrost, combined with the remoteness of the region, precluded deeper sample excavation.

These till samples then were assessed for variability within a single ridge, within a Rogen moraine field, and as part of a larger regional dataset (Campbell et al., 2012; Chapter 3). Clasts (2-80 mm) were sieved from a portion of each till sample collected, and separated according to lithology under a binocular microscope. The regional geology is quite complex (Figure 4-2) and clasts were grouped into eight simplified classes (sedimentary; volcanic; granitoid; quartz-rich; mafic and ultramafics; Dubawnt Supergroup; Paleozoic carbonate; and exotic) to reduce lithological identification errors. Most of the clast-types are presumed to be locally-sourced (Chapter 3), with the exception of Dubawnt Supergroup clasts, which crop out 350-600 km to the north-northwest (Paul et al., 2002). Carbonate clasts are sourced from the Paleozoic Hudson Platform ~100 km to the east at, and beneath, Hudson Bay (Manitoba Energy and Mines, 1980). 'Exotic' includes chert and white granitoid clasts that are faceted and bullet-shaped. The source of these clast-types is unknown, but they are thought to have travelled in from outside of the field area. A portion of each till sample was also dried and underwent major/trace-element geochemical analysis (Aqua regia partial digestion ICP-ES and ICP-MS, <63 µm fraction) and grain size analysis of the <2 mm size fraction (Campbell et al., 2012) as outlined by Girard et al. (2004).

4.4.2.1 Clast fabric analyses

Rogen moraine ridges are thought to form transverse to ice flow (Lundqvist, 1997; Hättestrand and Kleman, 1999; Moller, 2006). Formational stresses are then assumed to be parallel to ice-flow, regardless of the model chosen to explain ridge generation. To test the validity of this assumption, pebble-sized clast fabric measurements were completed within the steep lee-slope of four Rogen ridges (Figure 4-2), using both horizontal and vertical surfaces to allow access to an appropriate number of clasts. A-axis orientation were measured at all sites, and a/b plane orientation was measured at sites MT268 and MT269 following standard methods as outlined in Evans et al. (2007). The sample size varied from 24-50, due to the bouldery nature of sites and the availability of suitable clasts (a:b axial ratios of ≥1.5). The orientation data were statistically evaluated according to the eigenvalue method of Mark (1973) and Benn (1994), and graphically manipulated with Rockware® StereoStat. A/B plane data was evaluated based on methodology set out in Evans et al. (2007).

At each of the four sites, clast-fabrics were undertaken at a depth of at least 30 cm. No nearby active mudboils or cryoturbation of the soil horizons (albeit poorly-developed) were noted. Regardless, because the field area is within the extensive discontinuous permafrost zone, there is a potential that sub-surface clast orientation may have been affected by frost-churning and frost-heaving – leading to weak and/or disrupted clast fabrics. As such, strong fabrics are assumed to indicate little to no post-glacial reworking and are used less-tentatively herein.

4.4.2.2 High resolution S-wave seismic reflection data ACQUISITION AND ANALYSIS

Due to the lack of natural sections and the impossibility of trenching in such a remote northern region, geophysical techniques were used to profile the internal architecture of Rogen moraine in northeast Manitoba. Ground-penetrating radar (GPR) test profiles showed significant signal attenuation, most likely due to the higher clay and moisture content in the surficial layer. As such, GPR surveys were abandoned and our efforts focused on high resolution S-wave seismic reflection surveys. We chose to use S-waves, rather than P-waves, as S-wave propagation is not affected by variable saturation levels. In the field, three S-wave seismic survey lines (Figure 4-2 and Figure 4-4) were run perpendicular to three Rogen moraine ridge axes, and one parallel to a ridge axes.

The data were acquired using a Geode seismographic (Geometrics) with 24 horizontally polarized geophones and a geophone spacing of 1.5 m. Seismic waves were generated by striking a 5 kg sledge hammer horizontally on a well grounded H-beam at one meter offset from the geophone array and in-between geophones. Each spread consisted of 6 shots. The first six geophones were then moved to the end of the line prior to the next spread. The geophone array was kept fixed during each spread while the

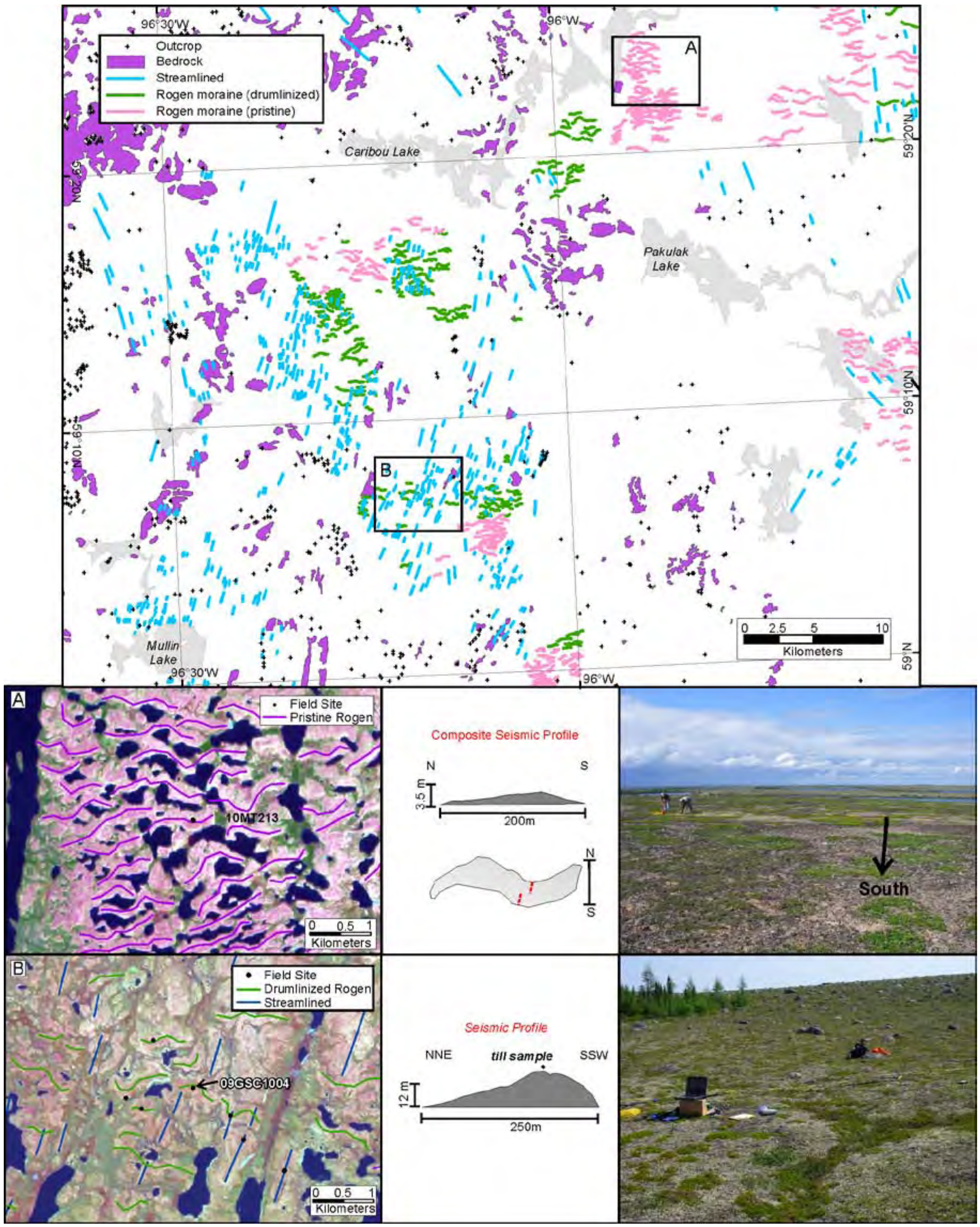


Figure 4-4. Near-surface S-wave seismic surveys were completed perpendicular to three Rogen moraine ridge axes, but only two profiles are discussed herein – due to processing difficulties. (A) low-relief pristine Rogen moraine at site 10MT213. Photo is from the flat crest. Another Rogen ridge is visible in the background. (B) Site 09GSC1004, at a concave down-ice higher-relief drumlinized Rogen ridge. Photo is looking up-slope towards the SSW. Note the somewhat bouldery nature of the ridge top which has been partially winnowed by wave-washing near the maximum relative sea level of the Tyrrell Sea.

source was moved along the geophone array at 1.5 m distance interval. Topography was measured along the four seismic survey lines using a metre-stick, and was utilized during the seismic data processing. Only two line profiles are presented herein, due to processing difficulties.

The shear-wave reflection field records were processed using Landmark Graphics' ProMax Software. The processing started with assigning the field geometry to the data file headers. Careful attention was given to all of the reflection events appearing in the field records in order to preserve all of them in the stacked profiles. A band pass filter with corners (10-20-70-80 Hz) was applied to suppress the unwanted high-frequency and low-frequency noise. An automatic gain control (AGC) with a sliding time window of 50 ms was applied to balance the trace amplitude and enhance the resolution potential by minimizing the difference in the amplitudes along the traces. A detailed stacking-velocity analysis was conducted by evaluating the velocity spectra resulting from plotting the amplitude after stacking the common depth point gathers (CMP) using varying velocities versus two-way travel-time (semblance). The produced stacking velocity functions were used to correct the CMP gathers for normal moveout and to convert the reflection profiles from time to depth. Both profiles were processed to the same datum.

4.5 RESULTS

4.5.1 Rogen moraine characteristics

It is essential to thoroughly describe the characteristics of Rogen moraine in the study area before any conclusions can be drawn. In accordance with general summaries presented by past researchers (Alysworth and Shilts, 1989; Bouchard, 1989; Lundqvist, 1989; Hättestrand and Kleman, 1999; Dunlop and Clark, 2006a), Rogen moraine in northern Manitoba:

- form fields at a wide range of scales – one km^2 to several thousand km^2;
- can be extensive and continuous, elongated ribbons and narrow tracks, densely packed or dispersed, and/or occur as isolated fields;
- can contain ridges with varying lengths and heights;
- usually have a high level of parallel conformity, and do not cross-cut;
- have a spatial association with drumlin and fluting fields, and individual ridges are commonly streamlined, both up-ice and down-ice as well as abrupt lateral shifts (Figure 4-5);
- no Rogen ridges were located that overlie streamlined landforms, with the exception of a small area of ridges in the northwestern corner of Manitoba that overlie old crag and tail landforms (Figure 4-5);

- overlie granitic bedrock of the Archean Canadian Shield (Anderson et al., 2010b) and do not extend down-ice beyond the contact with volcanic and metavolcanic rocks (Figure 4-2);
- are not associated with any particular topographic expression; and
- are locally crosscut by eskers.

The characteristics of individual ridges studied in northeast Manitoba are as follows:

- ridges are 0.1 to 9.3 km long (mean 1.8 km; n=3584), and 70 to 200 m wide;
- visited ridges were an average of six metres high (3 to 13 m, median=4.5, n=11)
- exhibit complex plan view morphologies classified as jagged, anastomosing, broad arcuate, downstream curving, and barchan shaped, following Dunlop and Clark (2006a) terminology;
- exhibit undulating longitudinal crest profiles with multiple subcrests – like ripples or dunes;
- nearly always asymmetric (cross-section) and usually have steep down-ice sides (Trommelen and Ross, 2010); and
- have surfaces littered with angular to sub-angular blocky granitic boulders (0.4 to 3 m diameter).

Furthermore, pristine ridges have been variably overridden by actively flowing ice, resulting in spatially variable drumlinization of their surfaces (Figure 4-6; 69% of ridges). There are down-ice, up-ice and lateral shifts to drumlinized Rogen moraine, with no obvious pattern (Trommelen and Ross, 2010). As such, we prefer to use the terms 'pristine' and 'drumlinized' Rogen moraine (Trommelen and Ross, 2010), as opposed to the numerous types of ribbed moraine presented in Dunlop and Clark (2006a). The geomorphic record clearly indicates that drumlinization, and the various intermediate forms produced, is a secondary distinctive process (as in Knight, 2010) and ribbing is the predominate pattern (c.f. Clark, 2010).

Most Rogen ridges encountered in the field have surfaces studded with large boulders. Based on visual observations, boulder concentration is clearly much higher at the surface of "pristine" Rogen moraines than on drumlinized moraines. In addition, boulder concentration on Rogen ridges is significantly higher than on adjacent drumlinoid ridges (Figure 4-5b). Many boulders appear to be partially buried at the surface and are abundant in the lee-slope, indicating that boulders are not just a supraglacial meltout drape or surface lag. The boulders are invariably granitoids, consistent with the surrounding bedrock. These non-distinctive hard rock lithologies could be sourced from a large up-ice area (at least 150 km north) and are not necessarily locally-derived.

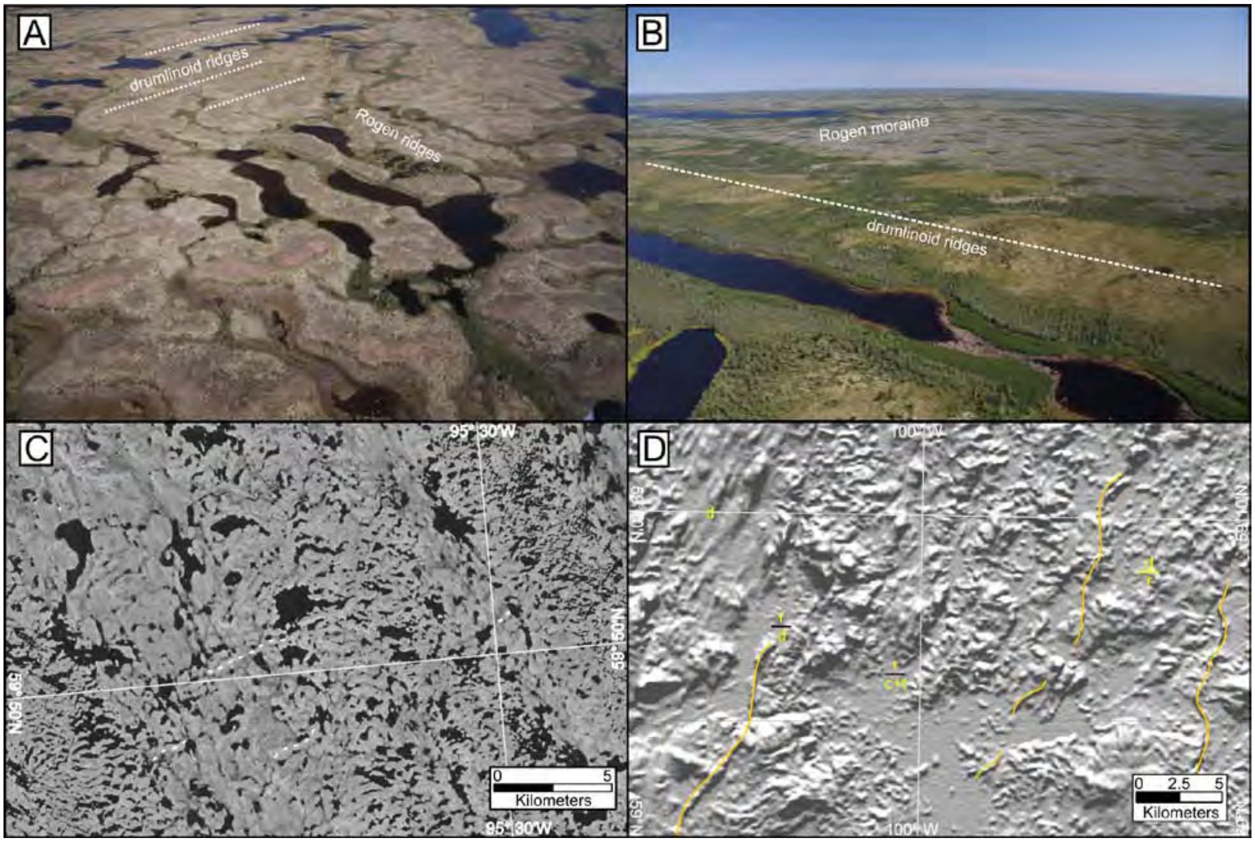


Figure 4-5. The complex relationship between streamlined landforms and Rogen moraine in northern Manitoba. Drumlinoid ridges may occur amongst (A) or adjacent to (B) Rogen moraine; or even overprint the moraine in large swaths (C). The surface of Rogen moraine ridges are littered with locally-derived granitic boulders (grey-color (B)), leading to a distinct visual contrast with streamlined landforms. It appears that Rogen moraines were more widespread, and are overprinted to obliterated beneath some swaths of streamlined landforms (white lines (C)). In northwest Manitoba, there are a few instances where Rogen moraine (r) appear to overprint older crag-and-tail features (c+t) or drumlinoid ridges (D), while nearby drumlinoid ridges clearly overprint or have completely re-worked pre-existing Rogen moraine. See Figure 4-2 for location.

Subglacial landforms are depicted with eskers, and overlain on the 1:75 000 scale bedrock geology (Figure 4-2), to assess the relationship between spatial landform location, bedrock geology (Anderson et al., 2009a) and meltwater availability.

4.5.2 Where do Rogen moraine form

In northeastern Manitoba, Rogen moraine are situated within at least ten fields, that cover ~5800 km^2, and extend south-southeast 120 km, within relict- and higher inheritance palimpsest-type GTZ (Figure 4-1).

In northwestern Manitoba, remotely-sensed mapping of Rogen moraine is more difficult, owing to extensive secondary erosion in large meltwater corridors (e.g. Trommelen, 2011b), but the ridges are situated within an area of ~17,300 km^2, extending up to 185 km south-southwest. Rogen moraine are not situated everywhere, but rather form as fields 14-2000 km², interspersed amongst streamlined terrain and bedrock-controlled terrain (Figure 4-1). Rogen moraine mapped south of 59° latitude are rare, though several fields of ribbed moraine ridges have been mapped in northeastern Manitoba just north of the 54° parallel near Red Sucker Lake (Trommelen et al., 2012b).

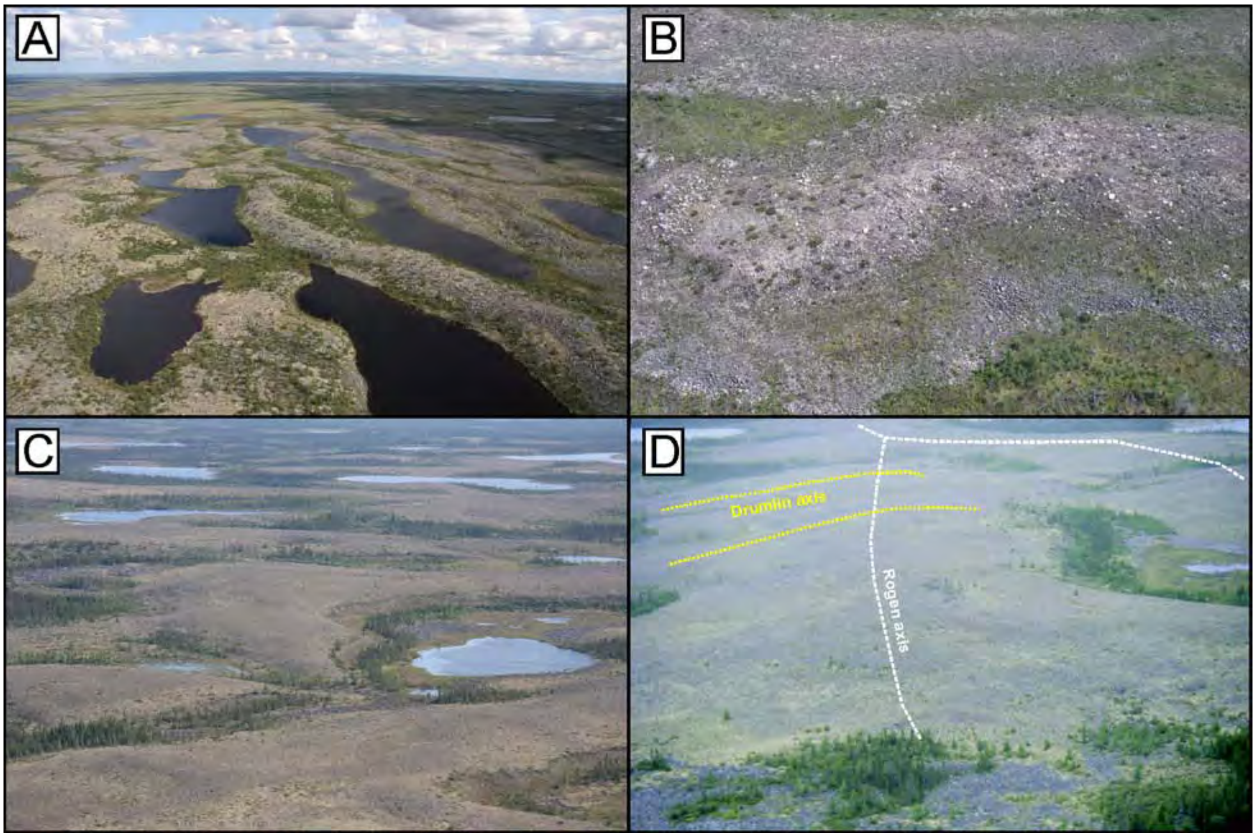


Figure 4-6. Pristine Rogen moraine (A and B) and drumlinized Rogen moraine (C and D) photographed in the field in northeast Manitoba. Significant wave-washing by the Tyrrell Sea (B) is responsible for the removal of fines in the lows between ridges. See Figure 4-2 for location.

While the majority of the study area is underlain by hard crystalline granitoid and gneissic rocks of the Canadian Shield, the southwestern field area (Figure 4-2) is underlain by less resistant, more permeable metavolcanic and metasedimentary rocks that exhibit significantly smoother surface topography than in the north. It appears that the concentration of both streamlined landforms and Rogen moraine changes abruptly at or near the contact between granitoid and these non-granitoid rock types (Figure 4-2), where there is also a slight drop (20-30 m) in elevation. Indeed, Rogen moraine ridges do not overlie the metavolcanic and metasedimentary rocks, but are situated just to the north and east and appear to mimic the shape of the curvilinear northern bedrock contact. However, as the study area contains the southernmost Rogen moraine in northeast Manitoba (Trommelen and Ross, 2010), it is difficult to define the significance of this spatial relationship.

4.5.3 Sedimentology

Large, stepped, profiles were dug into the lee-slope of six pristine Rogen moraine ridges and four drumlinized Rogen moraine ridges. The surfaces of both ridge-types are covered in granitoid and gneissic boulders (pristine: 35 to 80%, 40 to 300 cm diameter; drumlinized: 20-40%, 40 to 150 cm diameter). Boulders are present within the lee-slope of ridges at a similar concentration, and are sometimes stacked. Boulders at the surface are not perched, but are partially set into the till surface. At most sites, sediment encountered is massive, matrix-supported till, with a fine sandy-silt matrix and ~10 to 35% angular blocky to faceted clasts. Typically, a thin layer (20 to 40 cm) of coarser, more oxidized sandy till (ablation) drapes this finer-grained till. At one site (09GSC1020, Figure 4-2) there was a tabular cobble that dipped 55° towards 357°. There were also two shear planes above the cobble, within a 5 cm vertical distance, that align along the same a-b plane as the cobble. Within the lee-slope at another site (10MT108, Figure 4-2), a 3D surface of a bed within a layer of crudely stratified sand and silt was measured to dip 25° towards 185°.

4.5.4 Compositional variation

Textural and clast lithology analyses (Campbell et al., 2012) of till samples collected from the toe, mid, upper and/or crests of five Rogen moraine ridges, as well as from different subglacial landforms, indicate that the till throughout the region is generally similar (Table 4-2). There is not a significant variation (<5 standard deviation (st dev)) in texture or major clast lithology within any single ridge, except perhaps in

Table 4-2. Till sample data used to assess variability within a Rogen ridge.

Site	Texture			Clast Lithology				
	Sand (st dev)	Silt (st dev)	Clay (st dev)	Granitoid (st dev)	Quartz (st dev)	Gabbro/Amp[a]	Exotic[a]	Dubawnt[a]
10MT103	1.34	0.9	0.44	2.15	2.35	-	2,6	0,2
10MT108	4.98	3.69	1.4	2.24	2.19	0,1,1	-	3,1,4
10MT110	4.61	4.27	0.35	0.76	0.98	2,0	-	-
10MT114	3.31	1.72	2.11	3.65	3.51	0,2,4,2	-	6,1,0,0
09GSC1020	0.34	*	*	4.79	4.73	-	1,0	0,1

[a] Actual clast count, lowermost to uppermost sample
- signifies no clasts of that type were found
*silt vs clay was only calculated for one sample

the concentration of rare gabbro/amphibolites, exotic and Dubawnt clasts. At these low values, it is difficult to say whether any 'true' differences in clast lithology exist. The minor variation in clay content between different landform-types is likely due to post-glacial wave-washing at higher topographic points (all samples are below maximum sea level), rather than to a difference in sediment source or subglacial dynamics.

Table 4-3. Average grain-size for till samples from different subglacial landforms.

Landform	Texture			
	Sand (%)	Silt[a] (%)	Clay[b] (%)	n
Rogen (pristine)	67.43	29.57	3.00	6
Rogen (drumlinized)	67.86	28.17	3.98	22
Streamlined	65.34	29.69	4.97	22
%rsd	3.02	2.9	24.73	-

[a] 63 to 4 μm
[b] < 4 μm

Variations in clast lithology from surface till samples in northeast Manitoba are presented for the Great Island area in Chapter 3, without regard to geomorphic expression. In the Sosnowski Lake area (Figure 4-2), inherited southeast and northeastward dispersal of local sedimentary and volcanic clasts within till was documented, even within south-southwest trending streamlined landforms. This inherited subglacial detritus was also documented within Rogen moraine till samples (Chapter 3). Sampled streamlined terrain in the Great Island area has the highest maximum concentration of these lithologies (Table 4-4),

Table 4-4. Clast lithology (count-percent) of till samples from different geomorphic expressions.

Subglacial Landform		Clast Lithology							
		Exotic	Dubawnt	Sediment	Quartz	Volcanic	Granitoid	Gabbro_Amp	Carbonate
pristine Rogen	av	0.41	0.21	0.26	6.1	0.12	92.21	0.69	0.006
	max	1.38	0.69	1.6	12.94	0.73	98.11	4.35	0.07
	min	0	0	0	1.75	0	86.88	0	0
drumlinized Rogen	av	0.31	0.26	0.31	8.07	0.25	90.61	0.19	0
	max	2.31	1.11	5.97	21.3	3.89	100	1.16	0
	min	0	0	0	0	0	77.72	0	0
streamlined	av	0.33	0.24	0.3	7.57	0.22	91.01	0.32	0.002
	max	2.05	2.82	40.52	31.51	6.85	96.69	2.5	0.17
	min	0	0	0	2.49	0	43.87	0	0

but generally the mean values are consistent regardless of geomorphic expression. Rare carbonate clasts (Chapter 3) were encountered in surface samples from one easternmost-sampled pristine Rogen ridge and one streamlined landform, but not from within the landforms. Because both landform-types contain inherited subglacial detritus, it is likely that they formed partly, if not largely, from a pre-existing till sheet. The time of significant till production in the area is not well-constrained, but required either ice-marginal warm-based deposition (pre-LGM advance or retreat phases c.f. (Boulton et al., 2001b)?) or significant warm-based conditions and good till production beneath the thick inner-core region of the LIS (unlikely according to Chapter 2).

4.5.5 Geochemical data

The source of individual elements is often more difficult to determine than the source of clasts within till, and as such, for most elements no significant dispersal pattern was observed (Campbell et al., 2012). Arsenic (As), however, is an element that does show interesting dispersal patterns (Figure 4-7), with background (granitoid/gneissic) concentrations of less than 1.9 ppm and elevated concentrations up to 61.4 ppm. This element is confidently sourced to the volcanic and sedimentary rocks of the Great Island area (Anderson et al., 2009a, b), including within arsenopyrite at three known mineralized occurrences. There is a fan-shaped southwest-trending dispersal train (at least 12 km long) sourced from occurrence 1, and southeast and south-trending dispersal trains (3.5-5.5 km) sourced from occurrences 2 and 3. Relevant to this paper, however, are the two clusters of moderately-elevated As concentration situated 8 to 10 km north of the mapped volcanic and sedimentary rocks (Figure 4-7). Seven of the elevated (5.21 to 61.4 ppm) sites are samples from the crest and lee-slopes of Rogen moraine ridges, while the other four sites are from the crest of 205° trending low-lying drumlinoid ridges. It is unlikely that As was derived from the underlying granitic and granodiorite rocks, and hence reasonable to conclude that the As-bearing till was glacially-transported to these sites. There is no correlation between elevated concentrations of As and increased oxide concentration (such as Fe_2O_3, Campbell et al., 2012), meaning that the elevated concentration of As within Rogen moraine ridges is likely primary, and not simply related to the oxidation of arsenopyrite under better drainage conditions.

While it is tempting to simply delimit 15° to 20° -trending dispersal trains from all three arsenopyrite—bearing occurrences to these northern outlier areas (Figure 4-7), the ice-flow indicator record in the study area does not show evidence of such a flow direction. Instead, we suggest that As sourced from

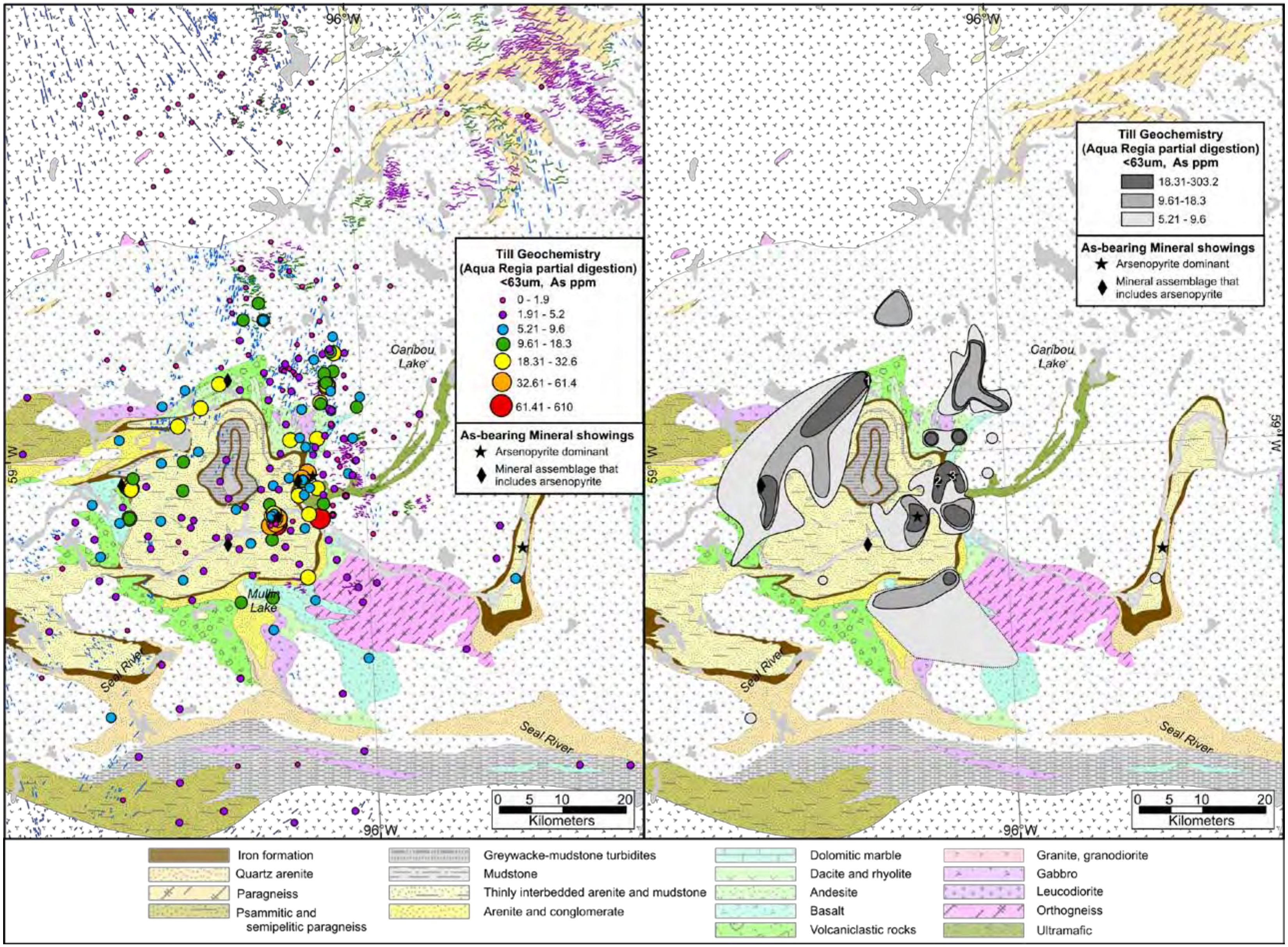


Figure 4-7. Arsenic (As) concentrations within the till matrix (<63 μm, ICP-MS, partial digestion), separated using natural breaks (jenks) in ArcMap. Data is from Nielsen (1987), Campbell et al. (2012), and unpublished 2012 field samples. Bedrock geology is sourced from Anderson et al. (2009a).

occurrence 1 may have been transported ENE to ESE (between 70° and 105°) into the region north of Sosnowski Lake. This is supported by old and young (weak fine striae) ENE to ESE-trending ice-flow indicators mapped across the study area (Campbell et al., 2012). Additionally, Chapter 3 documented southeast, and possibly east, dispersal of sedimentary and volcanic clasts within till in the same area. The location of the second cluster, north of Larocque Lake, is harder to explain. Old 338° ice flow is evidenced by chattermarks mapped 44 km to the ESE, while old 320°-trending chattermarks and grooves were documented 62 km SSE of this second cluster (unpublished data, M. Trommelen). Together, with the northwest or westerly transport of carbonate-bearing erratics into the Great Island area (Chapter 3), this weak evidence indicates possible NNW transport into the region. While it is always possible that buried un-mapped As-bearing volcanic/sedimentary rocks exist near the northern end of these interpreted dispersal trains, the bedrock map is well-constrained by geophysics and numerous outcrops. Thus if there is an unknown buried source, it has to be small localized subcropping mineralization.

The regional As-dispersal patterns are complex, and the source rocks are not confirmed. Nonetheless, this complexity cannot easily be explained using paleoglaciologic scenarios of the Late Wisconsinan. However, compositional inheritance from incomplete recycling of pre-existing till could lead to complex patterns like this. Indeed, I mapped several clusters of till with increased concentrations of inherited subglacial detritus in this same area (Chapter 3). If our interpretation of dispersal orientations is correct, it would indicate that while most of this NW(?) to ESE-transported material has since been removed, remnant outliers of higher-inheritance till remain *nearby or within* areas of Rogen moraine (Figure 4-7).

4.5.6 Clast fabric data

To assess the possibility of clast-fabric inheritance, and to test whether the till within was subject to stresses perpendicular to the ridge axes, clast fabric analyses were undertaken at the lee-slope of four Rogen moraine ridges (Figure 4-2 and Figure 4-8). Sites 10MT091 and 10MT114 exhibit strong A-axis clast fabrics (low isotropy and high elongation) while sites 10MT268 and -269 exhibit weak A-axis clast fabrics (high isotropy and low elongation) (Table 4-5 and Figure 4-8). It has been suggested that even under a single stress regime, smaller particles may display a systematic variation in orientation relative to larger particles (Kjaer and Kruger, 1998; Carr and Rose, 2003). As such, there may have been a sample bias leading to a stronger fabric in the former sites, given the longer mean A-axes relative to the latter sites (4.8

Table 4-5. Statistical data on clast macrofabrics

Site	Axis	Ridge Type	# clasts	S1	S2	S3	Isotropy (S3/S1)	Elongation 1-(S2/S1)	Mean Dip(°)	V1 (°)
10MT091	A-Axis	pristine	24	0.705	0.22	0.075	0.106	0.688	25°	30°
10MT114		drumlinized	39	0.596	0.265	0.139	0.234	0.555	31°	165°
10MT268		drumlinized	43	0.45	0.333	0.217	0.483	0.26	35°	~270° to 290°
10MT269		pristine	50	0.468	0.342	0.19	0.406	0.269	35°	~200° to 230°
10MT268	Poles to the A/B plane	drumlinized	39	0.53	0.263	0.207	0.309	0.503	42°	279°
10MT269		pristine	50	0.431	0.299	0.27	0.628	0.306	35°	241°

cm vs. 2.9 cm). In order to avoid the problem of transverse orientation with smaller clasts, Evans et al. (2007) proposed clast fabric analysis using A/B plane dip and orientation rather than just A-axis. A/B plane data was only collected for sites 10MT268 and -269 (Table 4-5). At site 10MT269, the poles to the A/B plane (the C-axis) cluster roughly in the southwest quadrant and exhibit moderate dips. The poles to the A/B plane at site 10MT268 exhibit bimodal clustering in the northwest and southwest quadrants, with moderate dips. Poles-to-plane data for sites 10MT268 and 10MT269 tend to cluster at moderate to steep dips, which perhaps indicates that shear was not common during till formation at these sites (Benn, 1995; Benn and Evans, 1996; Evans et al., 2007). The weak A- and A/B-plane fabric data for site 10MT268 indicates possible permafrost re-working, meaning this site was not used during interpretation.

4.5.6.1 Field area ice-flow data

The field area contains numerous ice-flow indicators (which include field-based striae and roches moutonnées), parts of five differently-oriented streamlined landform flowsets (Chapter 2) and amoeboid dispersal of locally-derived clasts (Chapter 3). The regional (~9000 km^2) field-based ice-flow indicator data suggests that ice flowed between 045° and 285° at various times during at least the last glaciation (Chapter 2), with intensive older and younger ice-flow phases to the southeast, south and southwest. Rogen moraine are presumed to be transverse-to ice-flow landforms, and hence the assumed ice-flow direction during initial ridge generation would have varied between 160° and 200°. Perpendicular to oblique streamlining of some drumlin ridges also ranges between 160° and 200°. Within the Rogen moraine ridges, sedimentological indicators of ice-flow include 185° down-ice dipping lee-side sand and silt beds (site 10MT108), and a 357° up-ice dipping tabular cobble (site 09GSC1020; Figure 4-9).

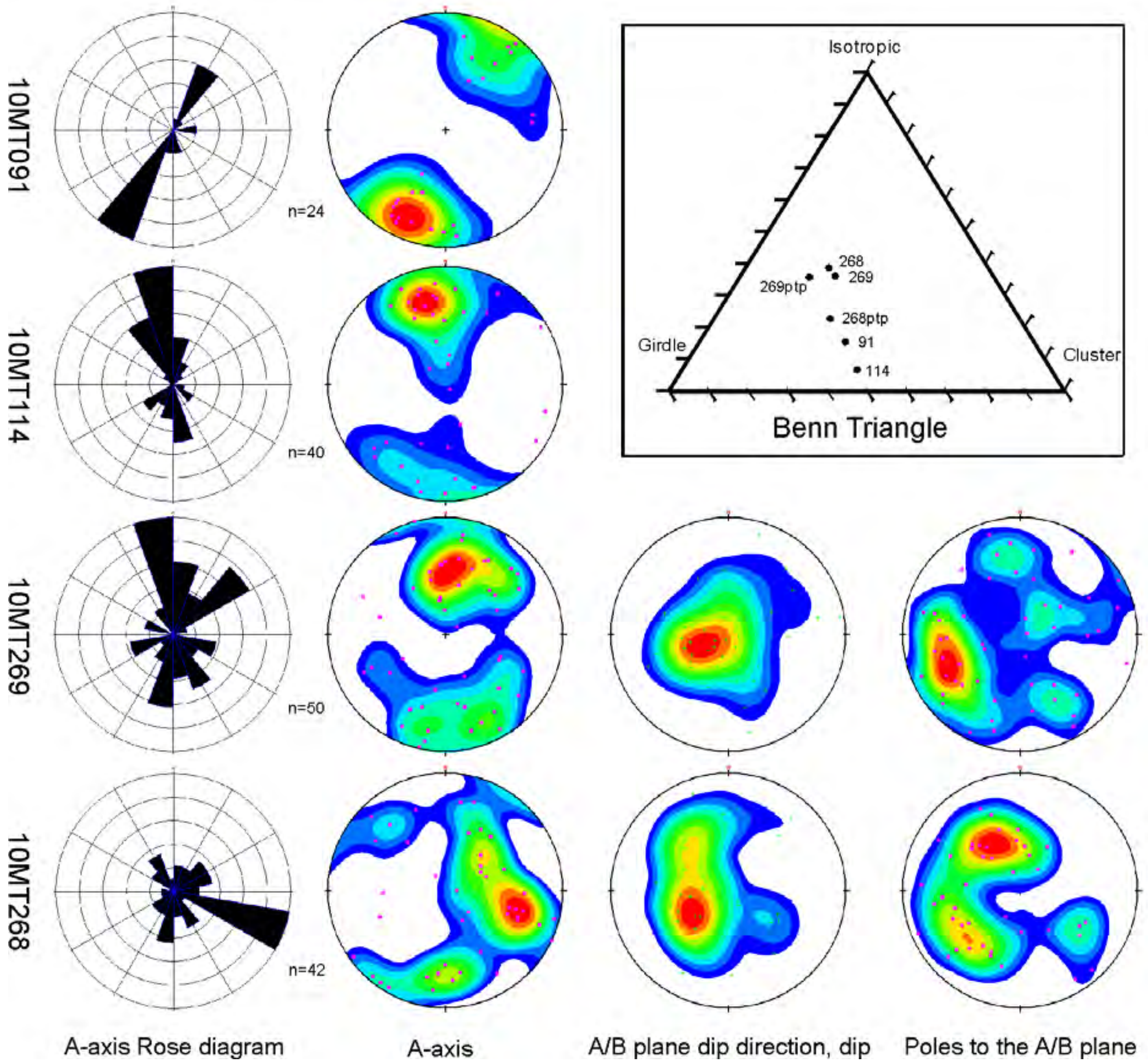


Figure 4-8. Clast fabric data from the lee-slope of four Rogen moraine ridges (see Figure 4-2 for locations). Clast macrofabrics are contoured as equal area lower hemisphere Schmidt nets using the Rockware Steronet program. Data is presented as A-axis dip-direction (rose diagram and steronet), A/B plane dip and dip direction, and as poles to the A/B plane (ptp).

4.5.6.2 INTERPRETATION

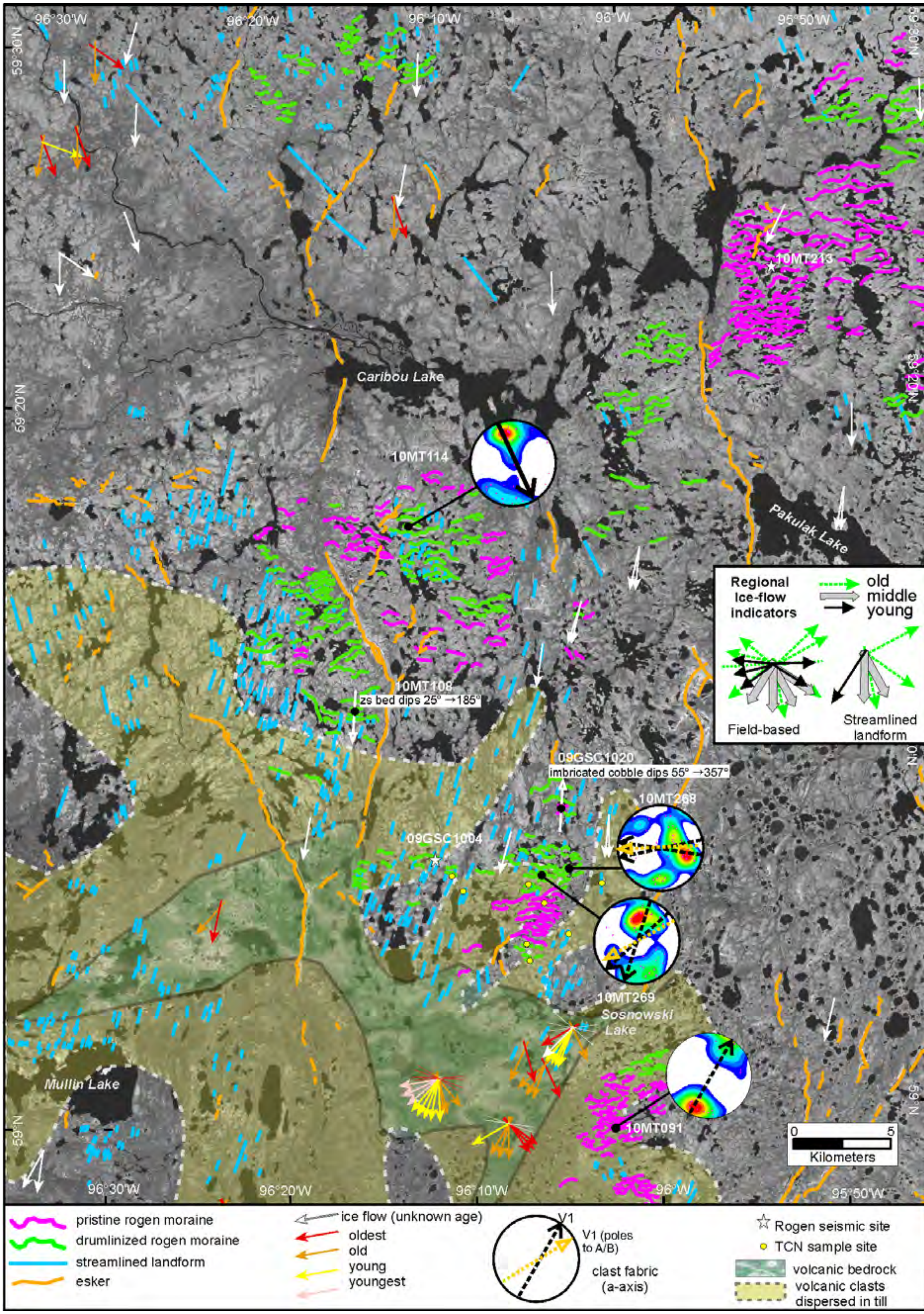
Clast fabric analysis at sites 10MT91 and 10MT114, with high S1 and low isotropy, indicate clasts are preferentially aligned, with interpreted up-dip ice-flow toward 30° and 165°, respectively. At site 114, this orientation is oblique to both the ridge axes and the secondary streamlining, but within the regional range of orientations. The clast fabric at site 10MT91 could be interpreted in several ways. A correlation with an older northeast ice-flow phase (2 striae trend 60°, 48 km to the southwest and 1 site with striae trending 45° to 60°, 65 km to the north-northwest) cannot be excluded as subtle northeastward dispersal of local clasts has been recognized (Chapter 3). Alternatively, the mean dip of clasts is 25°, and if the clasts dip down-ice instead of up-ice, we could interpret an ice-flow orientation of 210°. This would suggest that the clasts within the lee-side till at site 10MT91 are preferentially oriented parallel to the youngest south-southwest streamlined landforms. The a-axis clast fabrics at site 269 is more isotropic, but with additional A/B plane data it is possible to make inferences about ice flow direction (Evans et al., 2007). Ice likely flowed between 200° and 230° at site 10MT269 (V1=241°), substantiated by A-axis that dip towards the NNE and a clustering of A/B plane dip directions/dip in the SW (Figure 4-8).

In summary, ice-flow indicators associated with Rogen moraine generation are preferentially aligned towards the mid-phase of regional ice flow (inset on Figure 4-9). While the similar range of orientation between streamlined landforms and Rogen moraine in the field area may indicate coeval subglacial processes, in most cases it is clear that Rogen moraine ridges were generated first, and that some ridges were later partially to completely reworked into streamlined landforms (as in Figure 4-5C). Given the variation in clast-fabric data, it is likely that the landforms were generated, preserved and partially re-worked through complex switches in basal conditions – rather than generated just once at the same time across the entire area. Weak a-axis clast-fabric measurements from ribbed moraine in Labrador, Canada (Cowan, 1968), were also parallel, perpendicular and oblique to the dominant ice-flow direction – suggestive of a complex history.

4.5.7 Internal architecture

Processed high-resolution S-wave reflection profiles perpendicular to the axes of two Rogen moraine ridges are shown in Figure 4-10 and Figure 4-11, without and with interpretation guides. Several strong sub-horizontal reflectors are revealed in the first 15 to 25 metres. These reflectors are generally sub-parallel to surface, and exhibit minor undulations along profile – some of which can be interpreted as

Figure 4-9. Rogen moraine in the field area, as spatially-associated with the local and regional ice-flow indicators, including subglacial landforms (streamlined, eskers), field-based ice-flow indicators (striae, grooves, chattermarks, roches moutonnées), a-axis (91 and 114) and poles to the a-b plane (268, 269) clast fabric analysis from the lee slope of four Rogen moraine ridges, sedimentological data from two Rogen moraine sites, and regional clast dispersal of local volcanic bedrock in till (0% contour). The inset presents the regional ice-flow chronology as determined in Trommelen et al. (2012b).


96°30'W
96°20'W
96°10'W
96°W
95°50'W
59°30'N
59°20'N
59°N
Caribou Lake
Pakulak Lake
Sosnowski Lake
Mullin Lake
10MT213
10MT114
10MT108
zs bed dips 25° →185°
09GSC1020
imbricated cobble dips 55° →357°
10MT268
09GSC1004
10MT269
10MT091
Regional ice-flow indicators
old
middle
young
Field-based
Streamlined landform
0
5
Kilometers
pristine rogen moraine
drumlinized rogen moraine
streamlined landform
esker
ice flow (unknown age)
oldest
old
young
youngest
V1
V1 (poles to A/B)
clast fabric (a-axis)
Rogen seismic site
TCN sample site
volcanic bedrock
volcanic clasts dispersed in till

small folds (circled on Figure 4-10 and arrows on Figure 4-11). The sub-horizontal reflector (Figure 4-10) and the lowest reflector (Figure 4-11) at ~40 m depth are interpreted as the top of the granitoid bedrock. This reflector is gently undulating on the northernmost profile (Figure 4-10) and outlines three significant bumps on the southern profile (Figure 4-11). The bedrock reflector on Figure 4-11 is uncertain and was picked based on changes in the roughness of the reflector. We acknowledge that the change in properties between till and granitoid bedrock should have generated a stronger reflector, and thus it is theoretically possible that bedrock is even deeper. However, given that bedrock outcrops within 3 to 10 km in any direction, even 40 m depth is unexpected.

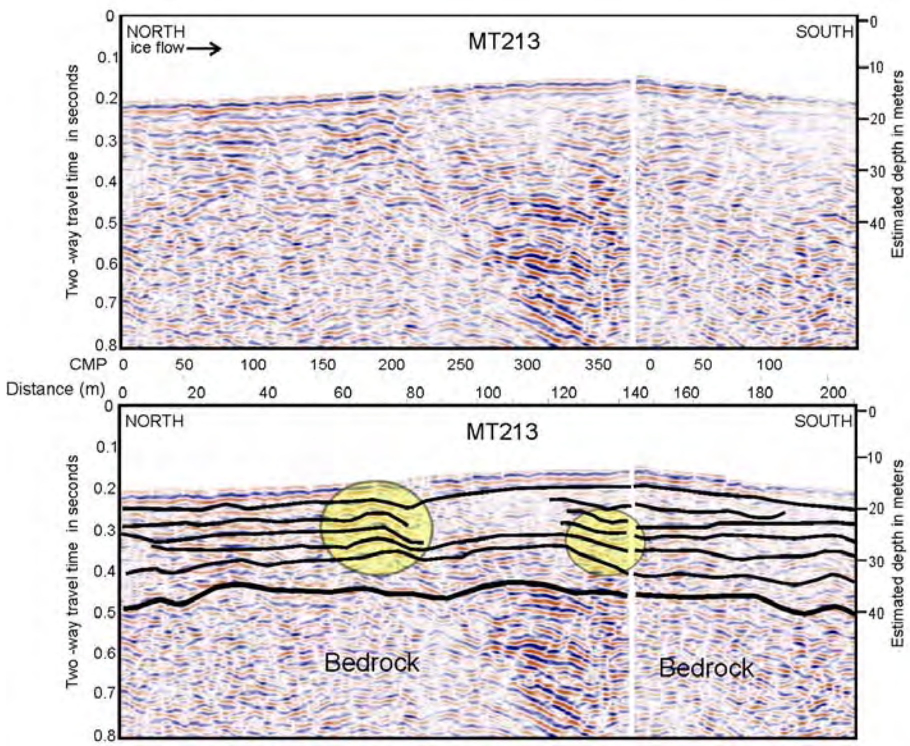


Figure 4-10. Seismic profile across the Rogen moraine ridge at site MT213. There is a 40 m offset between the left and right panels (see survey outline on Figure 4-4), and both profiles were shot up-slope. Layered reflectors, interpreted as subparallel-to-surface bedding, overlie granitoid bedrock (thickest black line). There are several folds within this stratigraphy (highlighted by yellow ovals).

It is reasonable to assume that there are no major sediment changes within the two profiled ridges, because large boulders (>1m) or beds of clay/sand should easily be detected by refractions or attenuated/increased signal, respectively. Till was present at the surface of both ridges, and we suggest the layered reflectors, above the bedrock, should be interpreted as stratified till that sometimes exhibits internal shallow folds. The subparallel-to-surface nature of the upper reflections in these perpendicular profiles indicate that there is a lack of normal faulting or cut-off beds - suggesting that the ridges are primary depositional features and not the result of boudinage (Sarala, 2006), stacking (Shaw, 1979) or thrusting and folding near an ice margin (Linden et al., 2008). The undulating to bumpy reflectors at depth may indicate that Rogen ridge initiation/amplification is linked to a bumpy substrate (precursor obstables; Boulton, 1987; Knight and McCabe, 1997; Clark, 2010). Unfortunately, the reflector interpreted as bedrock-surface is weak and not all that distinctive, meaning we cannot be certain about this interpretation.

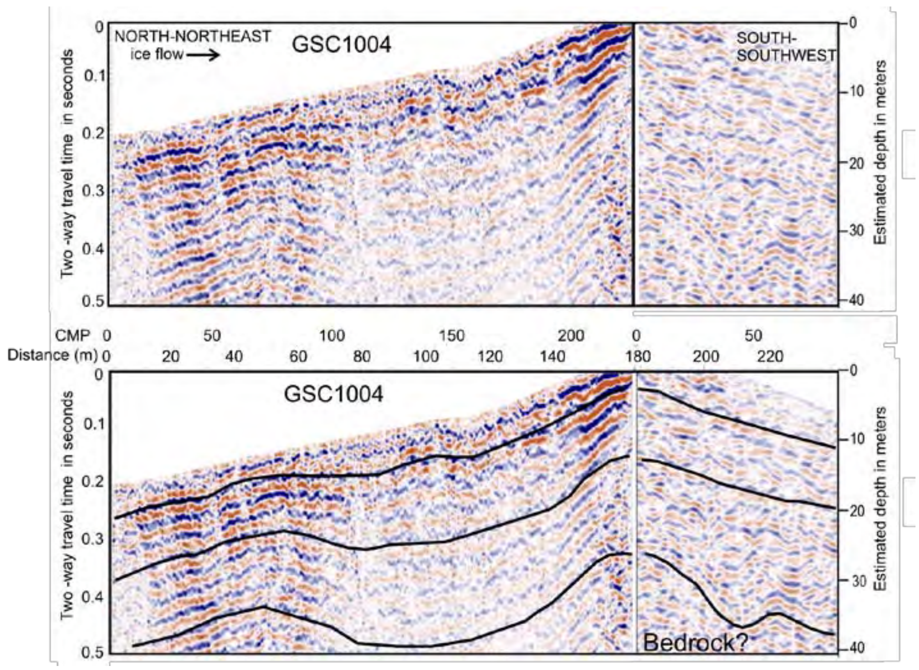


Figure 4-11. Seismic profile across the drumlinized Rogen moraine ridge at site GSC1004. There is a 16 m offset between the left (shot up-slope) and right (shot down-slope) panels. Layered reflectors, interpreted as subparallel-to-surface bedding, overlie bumpy granitoid bedrock (lowest black line). There are several shallow folds within this stratigraphy. The bedrock pick on this profile is uncertain and could be deeper.

4.6 DISCUSSION

Rogen moraine are an enigmatic subglacial landform around which numerous paleoglaciological questions revolve. To answer questions about formation, modification and preservation of Rogen moraine, it is important to consider not only individual ridges and fields of ridges, but also landform assemblages at the landscape scale (1000 km^2) to ensure consistency within the context of broader ice-sheet evolution. In this manner, we discuss landform assemblages, and then individual ridge characteristics.

4.6.1 Where do Rogen moraine form?

Most large fields of Rogen moraine are situated within 500 km of the inner core areas of ice sheets (Aylsworth and Shilts, 1989; Hättestrand and Kleman, 1999; Kleman and Hättestrand, 1999; Clark et al., 2000; Kleman and Glasser, 2007; Stokes et al., 2008; Greenwood and Clark, 2009b), including the ones mapped in Manitoba during this study. We acknowledge, however, that the apparent association between Rogen moraine and inner core regions of ice sheets may be one of preservation (c.f. Boulton 1987) rather than a strict requirement for their formation. The ridges may simply be better preserved in relict/palimpsest landscape assemblages whereby sluggish, sticky conditions were prevalent throughout deglaciation. Rogen moraine have also been associated with ice stream imprints (Dunlop, 2004) in northern Canada and interpreted as evidence of ice stream initiation shut down (Stokes et al., 2008) or onset (Dyke et al., 1992).

4.6.2 When do Rogen moraine form?

Evidence of compositional inheritance in the till in northeast Manitoba, as well as in the landscape record at all scales, suggests that considerable time was necessary to produce the study area sediment-landform assemblages under predominately low-erosion conditions. As the Rogen moraine ridges are part of this assemblage, and likely required a sufficiently thick sediment sheet from which to form relief, it is unlikely that Rogen moraine formed over a hard substrate during initial ice-sheet advance. Numerical experiments also suggest that initial build-up phases are characterized by thickening of a cold-based high-viscosity ice-mass that advances primarily by viscous creep (Stokes et al., 2012). Rogen moraine are thus more likely to have developed at a later time. Inherited eastward and southeastward (and possibly northeast and northwestward) dispersal of arsenic and volcanic clasts (Chapter 2) within the Rogen moraine and streamlined terrain also indicate that Rogen moraine were formed after a number of earlier ice-flow phases.

The orientation of Rogen moraine in northern Manitoba suggest these landforms were likely generated during the 'mid' SSE- to SSW-trending phase of regional ice flow, and then subsequently preserved during deglaciation. The timing of these early, mid and young ice-flow phases is unknown, though a thick ice-sheet is essential for basal temperatures to attain pressure melting point in inner-core regions. Thick ice also provides the required mass and thermal inertia to enable these widespread ice-flow phases far inside the ice sheet. Consequently, if these major south-trending ice-flow phases reflect ice dynamics at a time of large ice-sheet areal extent, it is possible that Rogen moraines formed shortly after the LGM or during a pre-LGM large-volume phase (e.g. 65ka; OIS 4). Alternatively, they could have formed during an interstadial retreat-phase when the study area would have been within the wet-based outer zone of the ice sheet (e.g. 40 ka, OIS 3; Kleman et al., 2010; Stokes et al., 2012). Similar pre-Late deglaciation scenarios have been suggested for Rogen moraine formation in Scotland (Finlayson et al., 2010) and Ireland (Knight and McCabe, 1997; Clark and Meehan, 2001; Greenwood and Clark, 2009b; Knight, 2010).

4.6.3 Regional basal thermal regime

Numerical modeling experiments of the LIS show that basal temperatures reached the pressure melting point in this area only during brief periods (Marshall et al., 2000; Tarasov and Peltier, 2004, 2007; Stokes et al., 2012). Though these low-resolution evolution models of the LIS are still simplified at a local-scale, predominately stiff, sticky and/or frozen-bed conditions probably prevailed for most of the last glacial cycle. Indeed, Chapters 2 and 3 have demonstrated that non-sticky wet-based conditions are over-represented in the deglacial geomorphic record of northern Manitoba, and that most warm-based ice-flow phases occurred before the last ice retreat-phase. Thus, while previous researchers suggested Canadian Rogen moraine formed within warm-based ice-marginal environments (Cowan, 1968; Shaw, 1979; Aylsworth and Shilts, 1989; Bouchard, 1989), this contrasts with the preservation of high-inheritance regions of the palimpsest spatio-temporal mosaic of landforms and ice-flow indicators (Chapter 2), sediment (Chapter 3) and relict landscapes (Chapter 2) in northern Manitoba. The general lack of ice-marginal landsystems (c.f. Dyke and Evans, 2003) is also interesting. Thus while widespread cold-based conditions were likely not present at the Last Glacial Maximum (e.g. Kleman and Glasser, 2007), as postulated in Chapter 2, sticky protective regions could form within stiff dewatered sediment under conditions of low subglacial meltwater availability (Boulton et al., 2009), or beneath frozen-bed patches (Kleman et al., 1999). As such, we tentatively confirm the often-cited association between Rogen moraine and widespread cold-based subglacial dynamics (Hättestrand, 1997; Kleman et al., 1997; Knight and McCabe, 1997; Hättestrand and Kleman, 1999; Kleman and Hättestrand, 1999; Jansson et al., 2002; Raunholm et al., 2003; Sarala, 2006; Van Landeghem et al., 2009), but for reasons of preservation and not formation.

4.6.4 What are the ridges made of?

Hand-dug holes into surface of Rogen moraine in northeast Manitoba indicate that the ridges have at least a carapace of bouldery till with minor lee-cavity sand and silt drapes on the down-ice side. The sampled till composition and matrix texture are indistinguishable from till within adjacent streamlined terrain, though patches within Rogen moraine fields contain higher inheritance of arsenic dispersal that the surrounding regions. Ribbed moraine studies conducted in similar low-relief areas of ribbed moraine in Quebec, Labrador, and Newfoundland, Canada, not associated with deglacial landsystems, have found the same general similarity between till from ribbed ridges and from surrounding terrain (Henderson, 1959; Drummond, 1965; Cowan, 1968; Marich et al., 2005). Deposition by meltout (Shaw, 1979; Moller, 2010) is precluded by the lack of internal meltwater-related features and absence of subglacial melt-out till, though lee-cavity sediments (c.f. Linden et al., 2008) were encountered at one site. The variety of interpreted ice-flow parallel stress orientations determined from the four clast-fabric pits (Figure 4-8) indicates that the sediment within the ridges were likely not sourced/deformed by one phase of ice-flow perpendicular to the ridge axes. Shallow folds (Figure 4-10 and Figure 4-11) are likely present within some northern Manitoba Rogen moraine ridges. This supports an extensive flow (increasing ice-flow velocity) rogen moraine generation model.

Regionally, the Rogen moraine fields are situated within relict and palimpsest-type GTZs (Chapter 2), which are associated with preservation of multiple ice-flow phases and greater glacial landscape inheritance. Thus in northeast Manitoba, evidence from multiple scales indicate Rogen moraines were formed by sediment sourced from a pre-existing regionally heterogenous till sheet. The angular to subangular boulders and cobbles at the surface of Manitoban Rogen moraine are lithologically similar to the up-ice granitoid and gneissic bedrock (Figure 4-2) – possibly indicative (consistent bedrock for >150 km north) of local provenance as was the case in Finland (Sarala, 2006) and Quebec, Canada (Bouchard, 1980). Till or lakes are found between ridges in northern Manitoba, and it is difficult to identify a clear plucking/short-transport relationship of surface boulders as posited by these authors. Till thicker that the ridge amplitude is also suggested by seismic data.

It should be noted that the remoteness of the study area (and extensive permafrost) precludes trenching, and that we only penetrated the upper one metre of sediment along the ridges. Thus whether the sedimentology is the same throughout the entire ridge thickness is still uncertain. Nonetheless, seismic data do not show major changes into deeper zones suggesting the ridges mainly consist of a single sediment body (as opposed to stacked stratigraphy).

4.6.5 How did the ridges get so thick?

With the exception of Rogen moraine (~40 m thick on seismic profiles) and hummocky till areas, the till is thin (0.3 m to <6 m) throughout the study area. At a few sites, bedrock outcrops (Figure 4-4) are situated just outside of Rogen moraine (observed crest heights of 3-13 m) fields and within streamlined terrain, which further suggests that regional till sheet is thin and that localized subglacial processes are needed to generate and or preserve thicker till within Rogen moraine (c.f. Dunlop et al., 2008). The preferred Rogen moraine formation hypothesis in northeast Manitoba must take into account this spatial discrepancy in till thickness.

Patchy shearing and stacking (Linden et al., 2008; Stokes et al., 2008) is one potential way to thicken subglacial sediment within a Rogen field, but it is difficult, if not impossible, to understand how a fracture and extend/fold model (Hättestrand, 1997; Hättestrand and Kleman, 1999; Sarala, 2006) could generate spatially variable till thickness. The instability model of ribbed moraine formation is part of a larger instability theory (Schoof, 2007; Dunlop et al., 2008; Fowler, 2009; Clark, 2010; Chapwanya et al., 2011) to explain streamlined landform (flutes, drumlinoid ridges, drumlins) and Rogen moraine generation. Clark (2010) specifically addresses the initiation and amplification of bumps by dividing drumlins into emergent drumlins, drumlin clones (pre-existing stiff patches) and obstacle drumlins (pre-existing rough topography). Emergent drumlins are described as drumlin forms that are instability-generated from a perfectly flat pre-existing till sheet where initial relief is created through shear thickening (dilatancy) or shear thinning (decreased viscosity and increased flow with increasing shear stress) (Smalley and Unwin, 1968; Clark, 2010). Abstractly, it may be possible to have the same trifecta in Rogen moraine, whereby the ridges could be initially sourced from 1) pre-existing stiff patches (dewatered (Boulton, 1987; Linden et al., 2008; Stokes et al., 2008) or cold-based (Sollid and Sorbel, 1984; Dyke and Morris, 1988; Dyke et al., 1992; Finlayson and Bradwell, 2008; Linden et al., 2008)), 2) obstacles (pre-cursor ridge (Lundqvist, 1969; Boulton, 1987; Knight and McCabe, 1997; Lundqvist, 1997; Moller, 2006; Finlayson et al., 2010; Knight, 2010)), or 3) as 'emergent' Rogen moraine.

4.7 Rogen moraine formation near an inner region of the Laurentide Ice Sheet

Rogen moraine in northeast Manitoba overlie the low-lying hard-bed Canadian Shield, are not confined by topography, are situated within high-inheritance relict- and palimpsest-type GTZ (Chapter 2) and are likely formed from a pre-existing regionally-extensive till sheet with a component of inherited subglacial detritus (Chapter 3). The ridges were likely generated, preserved and partially re-worked (drumlinized) during a warm-based SSE to SSW migration of ice-flow orientation (near or pre-LGM) – rather than generated during just one phase across northern Manitoba. Supraglacial meltout (Shaw, 1979; Moller,

2010) is thus negated as a formation model within this subglacial non-ice-marginal environment. Simple fracture and extension (Hättestrand and Kleman, 1999; Kleman and Hättestrand, 1999; Sarala, 2006) at a cold to warm-based thermal regime transition is also an unlikely formation model given the need for amplification of relief and preservation of ridge forms during deglaciation. Shallow S-wave reflection profiles of two Rogen moraine ridges indicate subparallel-to-the-surface layered bedding, with minor internal folds, that overlie gently-undulating to bumpy bedrock surfaces. Insufficient internal exposure of Rogen moraine ridges in Manitoba limits an assessment of the shear and stack hypothesis (
Table 4-1), though the seismic data obtained does not show the predicted compressional stacked layers, internal unconformities or reverse faults. Stacked till layers (~5° slope) were noted in the field by Shaw (1979), Bouchard (1980) and Lundqvist (1997), but were lacking in the internal excavations conducted by Cowan (1968) and Sarala (2006). Thus while there is no evidence for pre-cursor obstacles (ridges or hummocks; Boulton, 1987; Lundqvist, 1989; Knight and McCabe, 1997; Lundqvist, 1997; Moller, 2006; Finlayson et al., 2010) across the study area, the data do support a need for 'precursor' sediment that could then be subsequently moulded/deformed.

While it may be desirable to determine a single model of formation for all transverse moraines (Dunlop and Clark, 2006a), this discussion is restricted to Rogen moraine – defined as only topographically-unconfined ribbed moraine spatially associated with palimpsest streamlined terrain, but not deglacial landsystems. This may seem a limiting classification, but still includes widespread Rogen moraine in Canada situated in Saskatchewan, Manitoba, Nunavut, Quebec and Labrador (Cowan, 1968; Dyke and Morris, 1988; Aylsworth and Shilts, 1989; Bouchard, 1989; Clark et al., 2000; Dunlop and Clark, 2006b; Stokes et al., 2006; De Angelis and Kleman, 2008; Dyke, 2008; Stokes et al., 2008) and the Rogen moraine of Ireland (Knight and McCabe, 1997; Clark and Meehan, 2001; Knight, 2010). It may be that shearing and stacking is the preferred mechanism for ribbed moraine meltout associated with deglacial landsystems (Shaw, 1979; Linden et al., 2008; Moller, 2010), as evidence for these features has not-yet been encountered in Rogen moraine as defined here.

4.7.1 Rogen moraine and the transient subglacial bed mosaic

Only the instability-generated hypothesis of Rogen moraine initiation (
Table 4-1) appears to be consistent with our observations in Manitoba. Similar Rogen ridge heights, the parallel conformity of ridge axes within fields, and the spatial patterning of Rogen moraine as fields

suggest Rogen moraine generation, modification and preservation likely occur at a landform-field scale, and not at the scale of individual

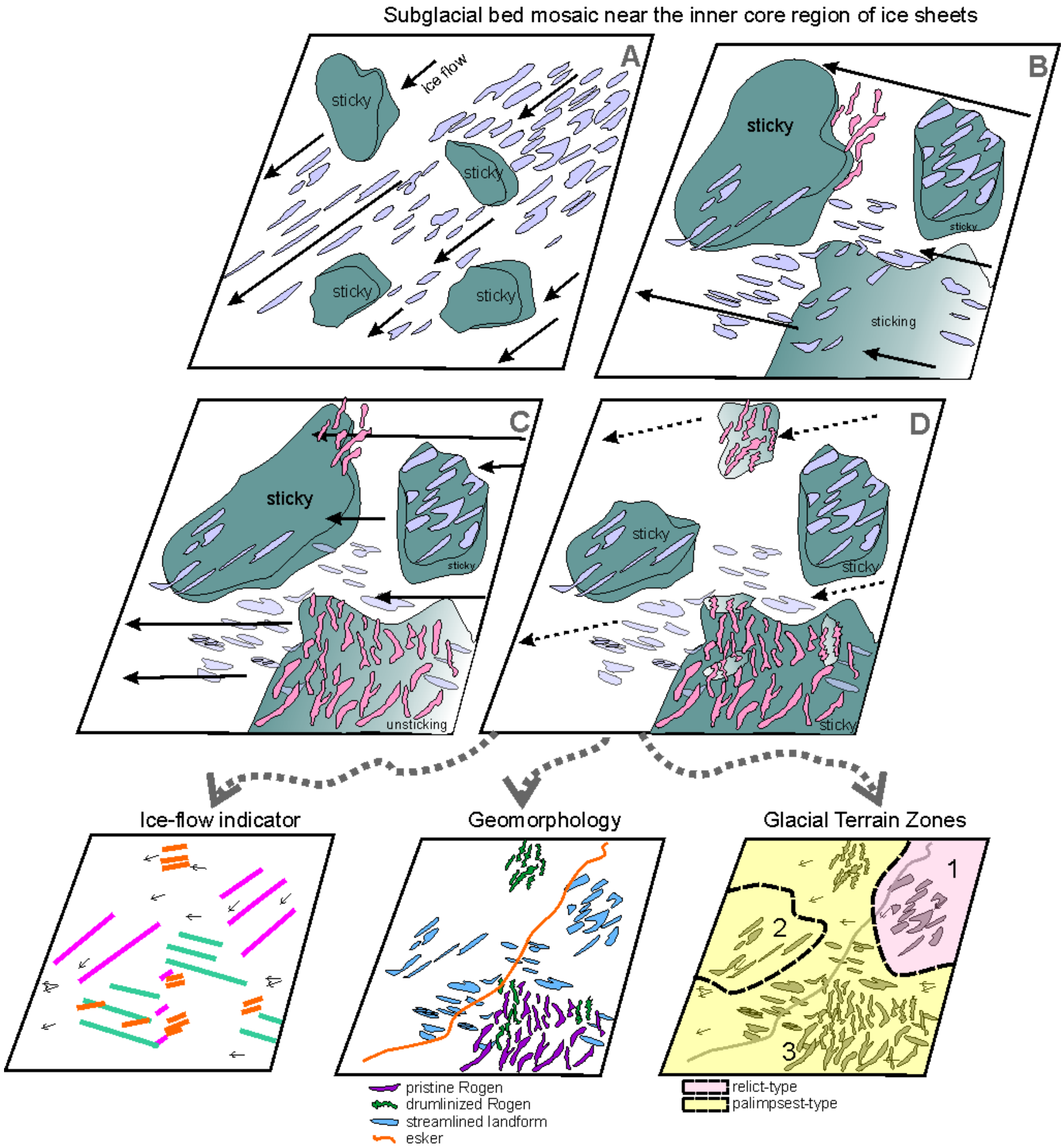


Figure 4-12. Space-time generation of Rogen moraine amongst a mosaic of active and sticky (stable) regions. Four phases of differently-oriented ice flow are depicted, starting with A and moving through time to D. A) Drumlinoid ridge formation occurs within actively reworking spots, amongst small sticky regions as part of the subglacial bed mosaic. B) The size of sticky regions increase, perhaps due to increased shear stress or less subglacial meltwater availability. Drumlinoid ridge formation still occurs within actively reworking spots, and some palimpsest drumlinoid ridges are formed near the boundaries of sticky spots. Rogen moraine may form within a compressive zone as actively reworking spots transition to sticky spots C.) Deflection of ice flow to a different orientation, combined with different regional subglacial conditions may lead to a reorganization of the subglacial bed mosaic; Rogen moraine may form during the 'unsticking' process of a previous sticky region. D.) Weak ice-flow phase with locally changing subglacial conditions, perhaps due to increased subglacial meltwater

availability during deglaciation, could lead to creation of small unsticking zones - within which localized drumlinization of Rogen moraine ridges could occur. Preservation and minor modification of Rogen moraine, streamlined landforms and field-based ice-flow indicators would vary spatially through time, leading to the cumulative ice-flow indicator, geomorphology and Glacial Terrain Zone (GTZ) maps.

landforms. Mimicking the drumlin formation theories proposed by Clark (2010), we suggest Rogen moraine are formed using pre-existing sediment (soft bedrock, till, gravel, etc), as *obstacle moraine* (involving pre-existing dewatered and/or frozen-bed stiff patches) or as *emergent moraine* (instability without large regional changes). The instability theory is unique in that, at this point of development, the theory suggests that instability and amplification of relief could occur through shear deformation as a response to shear strain, which is generated in through both elongation *and* shortening (Chapwanya et al., 2011). This means that generation of Rogen moraine could theoretically occur in both extensional and compression environments.

In keeping with the regional interpretation of the northern Manitoba subglacial landscape as a spatio-temporal mosaic of inheritance vs overprinting (Chapters 2 and 3), and hence relative reworking intensity, we infer paleoglaciologic conditions were also variable in space and time as part of a palimpsest (Knight, 2010) subglacial bed mosaic (Figure 4-12). This inferred mosaic may include spatio-temporal variability not only in 'pre-existing' properties such as substrate roughness, hardness, thickness, and texture, but in paleoglaciologic properties such as basal thermal regime (frozen-bed patches, Kleman, 1992; Kleman and Borgström, 1994; Kleman et al., 1999; Kleman and Glasser, 2007) and effective pressure (Bouchard, 1980; Alley et al., 1989; Benn and Evans, 1996; Boulton et al., 2001a; Eyles, 2006). The resulting subglacial bed would be a mosaic of stable (stiff, sticky regions) and active (moving or reworking) spots (c.f. van der Meer et al., 2003; Piotrowski et al., 2004; Piotrowski et al., 2006). Importantly, as basal ice encounters these transient regions, a mosaic of localized extensive and compressive ice-flow environments may also develop. Instability could be generated through variable shear stresses, which may be formed through localized variations in texture, density and effective pressure (bed-ribbing instability explanation, Hindmarsh, 1999; Dunlop et al., 2008; Chapwanya et al., 2011) – and we suggest that instability would be an inherent part of the subglacial bed mosaic at a landform-field scale.

Rogen moraine generation could then occur amongst this subglacial bed mosaic, perhaps where subglacial dynamics at the ice-bed interface (Figure 4-12B and Figure 4-13A) transition from actively-reworking regimes to sticky regions (Sollid and Sorbel, 1984, 1994; Stokes et al., 2008), or the reverse (ice-stream onset zone, Dyke et al., 1992). Alternatively, Rogen moraine generation could occur *within* sticky regions (Figure 4-12C and Figure 4-13B), perhaps as a region became 'unstuck' and re-mobilized

(Kavanaugh and Clarke, 2001) due to increased dilatancy (Smalley and Unwin, 1968; Denis et al., 2009). 'Unsticking' of large sticky regions could occur as a response to major subglacial changes, such as significant decrease in ice thickness and hence basal shear stress, capture/release of subglacial meltwater into a regional drainage network, or external climatic factors. This would be a form of 'emergent Rogen moraine', whereby slow-flowing subglacially-deforming till would encounter a mosaic of transverse to ice-flow stiff patches that locally inhibited simple down-ice flow of material. Ultimately, this is an expansion of the Rogen moraine subglacial deformation formation hypothesis first proposed by Boulton (1987), with low ice-flow velocity and short sediment-transport distances, but without the need for pre-existing ridges. Questions still exist regarding the predominance of drumlins over Rogen moraine throughout most of the subglacial landscape (Aylsworth and Shilts, 1989; Kleman and Hättestrand, 1999), and the predominance of drumlinized Rogen moraine rather than Rogen moraine overprinting of streamlined landforms, as the model presented in Figure 4-12 would suggest that both could (and should) occur. There must be some limiting factor(s) that we have not yet considered.

4.7.1.1 Rogen moraine preservation

Non-erosive and non-depositional conditions are required for preservation of Rogen moraine ridges during deglaciation (as in Ireland and Scotland, Finlayson et al., 2010; Knight, 2010). Ways to preserve subglacial landforms that were generated further back beneath an ice sheet (including Rogen moraine) include subsequent stagnation or a change to sticky conditions after initial warm-based formation (Figure 4-12) (Stokes et al., 2008). Numerical experiments indeed predict that the areal proportion of warm-based ice decreases after LGM during rapid thinning phases (Tarasov and Peltier, 2004). Because the outer zones of the LIS are likely warm-based following LGM, a reduction of warm-based areal extent thus means that large parts of the core regions become more sticky/sluggish and possibly cold-based. Both mechanisms could occur during thinning of an ice sheet during deglaciation, especially near the outer fringe zones of ice-divides where shear stresses are high (Boulton and Hindmarsh, 1987) and there is a lower availability of subglacial water at the ice-bed interface. Assuming Rogen moraine ridges were generated transverse to ice-flow direction, the glacial history in northeastern Manitoba requires later migration of ice-flow orientation to the SW, WSW (Figure 4-14) and then to ESE, post-generation of the Rogen moraine. This likely required a significant time period, leading us to favour a change to sticky conditions over late deglacial stagnation.

The establishment of the regionally extensive (~700 km wide by at least 500 km long) dendritic esker channel-system in northern Manitoba (Figure 4-1) likely caused spatially variable dewatering of the substrate far back under the ice sheet (Chapter 2). Significant dewatering, over a short time- scale, would have resulted in patchy stiffening of the substrate and slowing of subglacial deformation and/or basal sliding (Boulton et al., 2009; Tylmann et al., 2012). Near the inner core of ice sheets, basal freeze-on (Christoffersen and Tulaczyk, 2003) of remaining subglacial meltwater, combined with ice-sheet

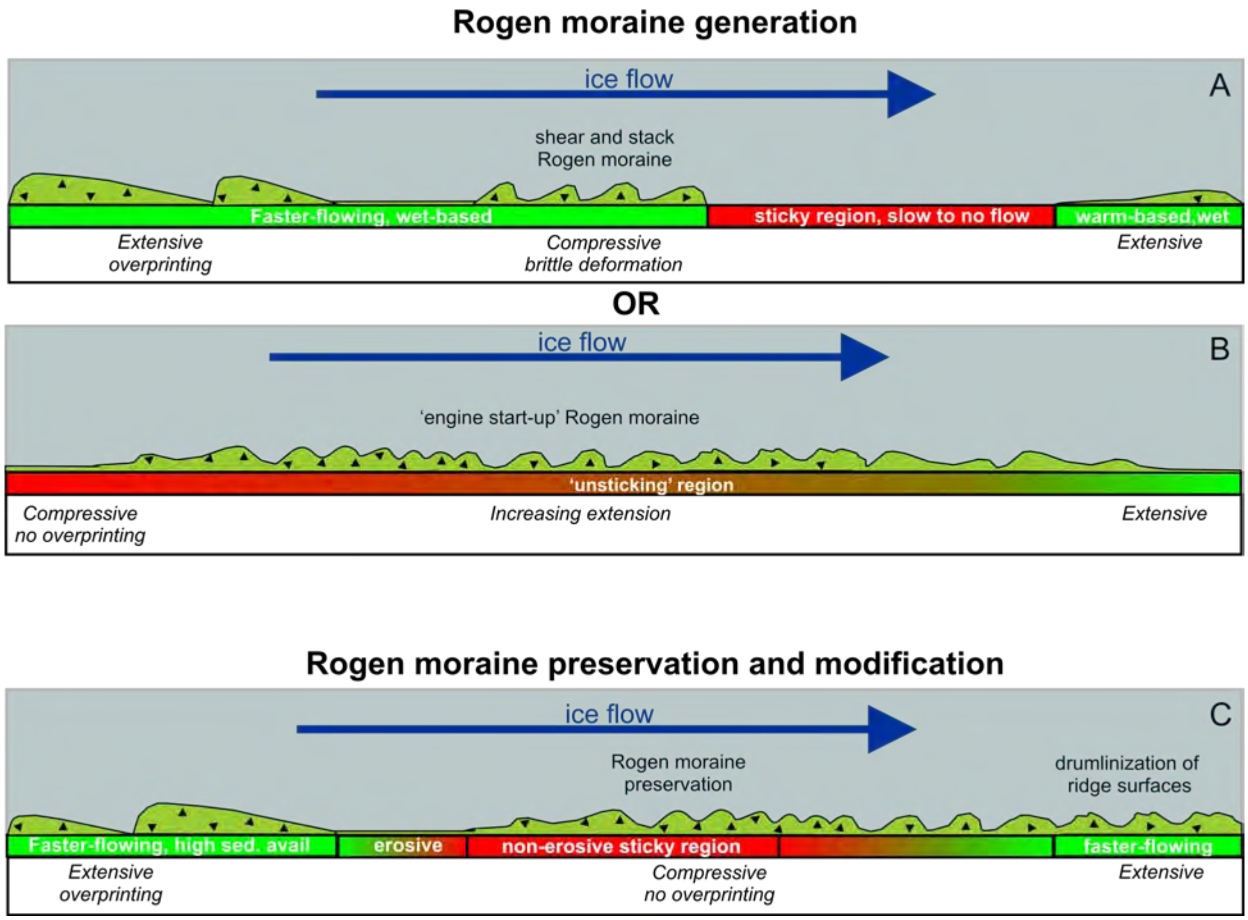


Figure 4-13. Possible Rogen moraine generation (A and B), preservation and modification (C) within a mosaic of localized (A) to regional-scale (B) extensive (active reworking) and compressive (sticky) regions. We envision Rogen moraine could form according to the shear and stack hypothesis at compressive zones generated where an actively-reworking regime transitions to a sticky region (A). This process may be driven by a natural instability mechanism, and could theoretically occur at any compressive zone if the strain regime is high enough – not just near the inner core region of ice sheets. Alternatively, Rogen moraine could be generated within an entire sticky region (B) as the sticky region becomes 'unstuck' and transitions to an actively re-working regime. This regional-scale regime change would likely require larger-scale events to influence subglacial dynamics at the ice-bed interface, such as increased/decreased ice thickness (e.g. build-up and demise of an ice divide), capture/release of subglacial meltwater into a regional drainage network, or external climatic factors. Partial preservation and modification of Rogen moraine (C) would occur during a return to a more localized mosaic of reworking (overprinting) and preservation (inheritance) spots.

thinning and low surface temperatures, may have led to an increase in the abundance of completely frozen patches (Kleman et al., 1999) within the subglacial bed mosaic. As such, we postulate that the conjugate-ridge networks within the relict Caribou River GTZ (Figures 4-1 and 4-3) are evidence of basal frictional heating between sticky, possibly frozen-bed, patches (Rogen moraine) and unstuck wet- based areas of flowing ice (streamlined terrain). Incomplete overprinting of Rogen moraine in this region indicates that the overall regime was still fairly slow and probably cold, suggesting that the conjugate ridges may be evidence for subglacial brittle deformation caused by internal shear at regions of velocity contrast – akin to lateral shear marine moraines (Hindmarsh and Stokes, 2008). Further evidence for localized frozen-bed patches in northern Manitoba include relict GTZ, (Chapter 2), mapped felsenmeer (Dredge and Nixon, 1992) and at least one small patch of regolith (Trommelen, 2011b). Our observation-based interpretation of sticky (possibly frozen) patches within the subglacial bed mosaic (Figure 4-14) is also supported (though over-generalized) by probability modeling of permafrost beneath the LIS at the Last Glacial Maximum (Tarasov and Peltier, 2007; Stokes et al., 2012). Subglacial landscapes in Quebec, Canada (Clärhall and Jansson, 2003), Sweden (Clarhäll and Kleman, 1999) and Scotland (Hall and Glasser, 2003), that exhibit a spatial mosaic of relict, palimpsest and 'young' geomorphic forms may be further evidence for the widespread nature of this landform-field scale subglacial bed mosaic – rather than an widespread regional transition in basal thermal-regime zone as currently interpreted (c.f.Wilch and Hughes, 2000; Clärhall and Jansson, 2003).

Boulder-draping of the Rogen ridge surfaces may occur at the initiation of, or during this long-term preservation stage. As ice continues to deform over and around these sticky (frozen-bed?) patches, a shear zone could develop up-ice of the patch – and propagate englacially just above the patch. Short-travelled debris sourced up-ice of the sticky patches would then accumulate along this zone, and would eventually be deposited through meltout (as postulated in Finland by Sarala and Peuraniemi, 2007). We suggest the boulders are partly buried instead of perched and on the surface like "normal" supraglacial meltout because of the higher concentration of debris accumulated just above the ice-bed interface (as opposed to englacial debris scattered throughout the ice thickness). This is also similar to the debris-rich meltout model of Rogen formation proposed by Bouchard (1989), though we view this as a post-Rogen process. Further excavation of the ridges in northern Manitoba is required to confirm the lack of boulders at depth.

4.7.1.2 SUBSEQUENT DRUMLINIZATION

It may be that variable subglacial and external climatic conditions, during advance and retreat of ice sheets, lead to changes in the configuration and scale of the subglacial bed mosaic. As ice encounters changes in 'pre-existing' properties such as substrate roughness, topography, hardness, thickness, and texture, the mosaic would develop an initial configuration (and scale). Then, we postulate transitions in

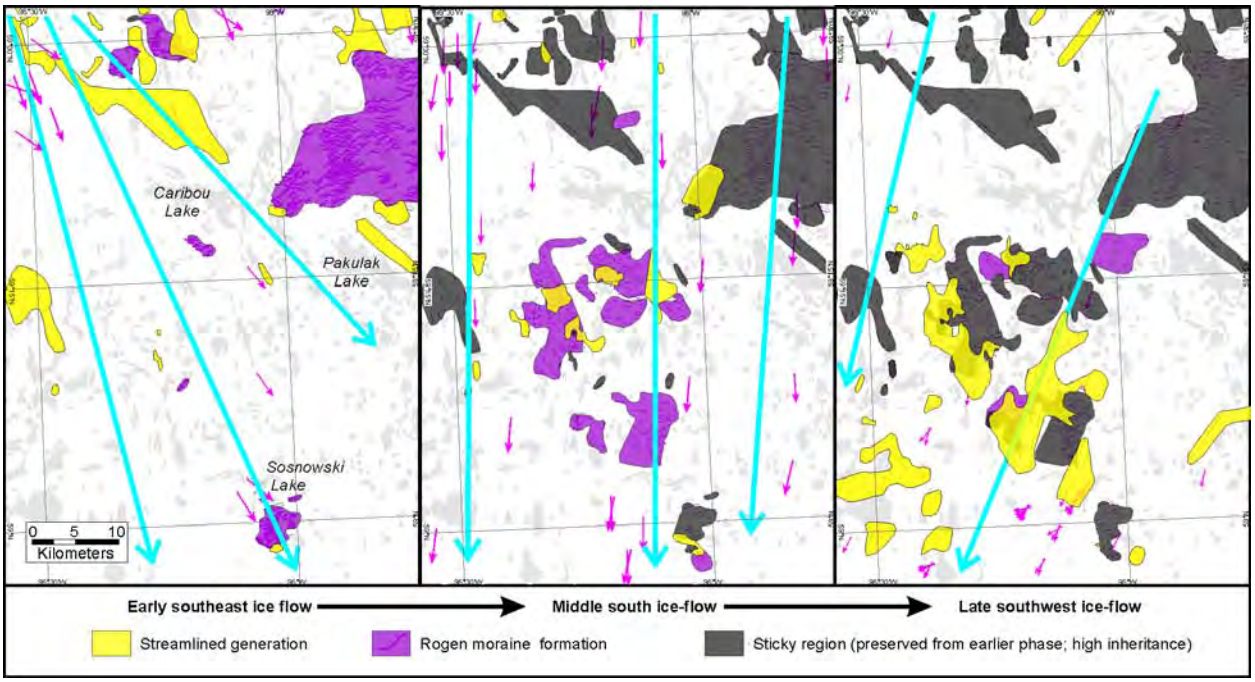


Figure 4-14. Possible subglacial mosaic maps denoting the evolution of spatially variable streamlined landform generation, Rogen moraine formation and preserved or inherited regions indicative of sticky or cold-based patches over time. Small pink arrows are field-based ice-flow indicators. Large blue arrows indicate interpreted regional ice-flow orientation for each time-slice map. ESE to SE ice-flow and generation of Rogen moraine with axes that trend ENE is followed by ice flow to the south, which generated some Rogen moraine with E-W ridge axes and patchy drumlinization of some pre-existing Rogen moraine ridges. Later, ice oriented SSW to WSW flowed across the area, generating a few more Rogen moraine ridges, and forming regionally extensive (but patchy) streamlined terrain.

both the basal thermal regime and the hydrologic regime, as well as distance from an ice divide (c.f. Boulton et al., 2001b), development of ice streams (Bennett, 2003), and climatic amelioration (c.f. Denton et al., 2010), would lead to additional changes in the configuration and scale of the subglacial bed mosaic over time. For example, as subglacial effective water pressure locally changed, perhaps due to regional drainage capture (c.f. Fischer and Clarke, 1997; Boulton et al., 2001a; Kavanaugh and Clarke, 2001), localized stresses within the subglacial mosaic would change correspondently. As such, previously stiff, sticky or frozen non-deforming spots (that were preserving relict/palimpsest landscape, landform, or sediment patches) could become unstuck deforming spots – resulting in superimposition or eradication of existing subglacial landscapes through deposition and/or erosion (Figure 4-12D and Figure 4-14). A modern example of a similar change is the observed transition over time from erosion to drumlin formation beneath an area of the Rutford ice stream in Antarctica (Smith et al., 2007). Regardless of the variables involved, spatio-temporal changes in the subglacial bed mosaic affect basal motion of the ice sheet, which is an important driver of ice-flow velocity, even beneath modern polythermal glaciers (Bingham et al., 2008). Changes in the subglacial bed mosaic may even provide an explanation for glacier speed-up when most other factors have been discounted (c.f. Van der Veen et al., 2011).

In northern Manitoba, empirical evidence indicates pristine Rogen moraine and the associated spectrum of streamlined landforms (drumlinized Rogen moraine to flutes, drumlinoid ridges and drumlins) are associated, but are not interdependent subglacial landforms. Changes in the configuration and scale of the subglacial bed mosaic over time provide a mechanism for patchy fragmented overprinting – that in northern Manitoba could have led to drumlinization of some Rogen moraine ridges and eradication of others (Figure 4-14). Such switches to locally deforming (faster-flowing) ice, as commonly cited to effect a change in landform-type generation (Aario, 1977; Boulton, 1987; Dyke et al., 1992; Hättestrand and Kleman, 1999; Linden et al., 2008; Stokes et al., 2008; Fowler, 2009) are possible – and expected, within the subglacial bed mosaic. The reverse transition to locally non-deforming sticky spots is also expected. The generation of relict and palimpsest GTZ (Chapter 2) is a landscape-scale example of a similar process.

4.8 CONCLUSION

Recent fieldwork in northeast Manitoba, Canada, provided the rare opportunity to study a remote subglacial landscape consisting of transverse ridges and streamlined landforms near the Keewatin Ice Divide of the Laurentide Ice Sheet. The transverse ridges are interpreted as Rogen moraine and defined as ribbed moraine ridges that have a spatial association with streamlined landforms but not deglacial landsystems, and are situated on regionally low-relief terrain without topographic constraints. Ongoing regional paleoglaciological reconstruction allowed assessment of the formation, modification and preservation of these Rogen moraine within the regional spatio-temporal mosaic of subglacial landscape inheritance.

The data we have presented suggests Rogen moraine are a palimpsest subglacial landform, generated primarily from pre-existing sediments correlated to areas of high glacial landscape inheritance near the inner-regions of ice-sheets. In absence of topographic highs (c.f. Bouchard, 1989; Finlayson et al., 2010), or proven pre-existing ridges (c.f. Lundqvist, 1997; Moller, 2006), the impetus for Rogen moraine field generation was most likely spatio-temporal variability within a landscape-scale subglacial bed mosaic. In areas of pre-existing sediment/soft substrate, this transient mosaic of deforming and non-deforming (stiff, sticky) regions may have generated a corresponding transient mosaic of compressive and extensive stresses at the ice-bed interface - triggering instability-driven formation of 'obstacle' or 'emergent' Rogen moraine fields. While the ridge-building mechanism is still unknown, the data herein favour subglacial modification of pre-existing sediment, through instability-generated waves (extensive stresses). Some sticky spots may have remained cold-based or dewatered throughout deglaciation, and allowed for preservation of 'pristine' Rogen ridges. Most sticky spots, however, likely continued to migrate until the ice stagnated, causing weak secondary drumlinization of the ridge surfaces by actively moving ice. Where ice-flow velocity increased substantially, the pre-existing landscape was overprinted and in some areas streamlined terrain was generated.

Landform generation within the subglacial bed mosaic likely occurred over multiple ice-flow phases, with only limited subglacial erosion, transportation or deposition during deglaciation. The idea of landform preservation at patches within a transient subglacial bed mosaic, now allows for a close association between subglacial drumlin and Rogen moraine ridges that may have formed by disconnected and not necessarily coeval or related processes.

Chapter 5 Discussion

Based upon recognition of a highly fragmented, high inheritance subglacial landscape in northeast Manitoba, Canada, this thesis presents a subglacial bed mosaic hypothesis of paleoglaciological evolution for the inner core regions of ice sheets. This hypothesis allows for disjoint subglacial processes to occur within close proximity at the subglacial bed, and allows for expected spatio-temporal variations in substrate properties (texture, heterogeneity, topography, hardness, thickness) and ice-bed interactions (availability of meltwater, basal thermal regime, strain caused by ice thickness). This thesis also suggests that Rogen moraine may have been generated, modified and preserved within this subglacial bed mosaic. The following is a brief review of the strengths and limitations of the methodology used.

5.1 DATASETS USED FOR MAPPING

Mapping of subglacial landforms using remotely-sensed imagery is a common technique in paleoglaciological inversion studies (Boulton and Clark, 1990; Kleman et al., 2006; O Cofaigh and Stokes, 2008). Though using the same general methodology, the end-result mapping can vary greatly depending on the imagery used, the scale chosen, and the biases of each mapper. For example, drastically different \`flowline\` maps of glaciated North America were produced by Kleman et al (2010) and Shaw et al. (2010), the former who used Multi-Spectral Scanner Satellite Imagery (MSS, 60m resolution) with some Landsat Enhanced Thematic Mapper Plus (ETM+, 15m resolution) and rare aerial photographs and the latter who used Landsat 7 ETM+ and Shuttle Radar Topography Mission (SRTM, 90m resolution) elevation data. The Kleman et al (2010) map is more detailed, and is quite similar to the Glacial Map of Canada (Prest et al., 1968) which was produced from aerial photographs. Boulton and Clark (1990) also used MSS imagery, but Kleman et al (2010) were unable to reproduce some of their flowsets. Thus any interpretations using a landform map based on remotely-sensed imagery must acknowledge the limits of the data.

The remotely-sensed subglacial landform map for northern Manitoba (Trommelen and Ross, 2010) was completed using Landsat 7 ETM+ and SRTM, with a few SPOT satellite images (panchromatic 10m resolution). The hillshade model built upon the SRTM DEM was very helpful in establishing the orientation of some flowsets when it was not obvious on the Landsat imagery (NE orientation of Flowset A, rather than SW as mistakenly assigned to the 'Churchill Swarm' by Kleman et al. (2010)).

As of 2012, SPOT satellite imagery is now available for all of Canada through geobase.ca. With the availability of this new resolution of data, along with 0.75 arc second CDEM elevation data (also geobase.ca), this author is currently updating the landform mapping at the Manitoba Geological Survey (Trommelen et al., 2012b). Landform mapping is also compared to geophysical aeromagnetic and electromagnetic data (Residual Total Field), to ensure that bedrock structure is not captured. This newest level of detail helps delineate small fragmented flowsets, as mapped in Figure 2-6 but not on the subglacial landform map (Trommelen and Ross, 2010). While small, their formation must still be accounted for within the regional glacial history, and can lead to significant deviations from mega-geomorphology studies such as Kleman et al. (2010).

As part of this author's work obligations, an area in northwest Manitoba was also mapped (Trommelen, 2011a, b, c). In this area, abundant secondary meltwater has modified the original landscape, which can mask primary subglacial landforms on both aerial photographs and remotely-sensed imagery – and lead to incorrect remotely-sensed mapping. In these areas, fieldwork is necessary to determine the extent of secondary modification, and to enable better recognition of primary features. It should be noted that these meltwater corridors are best identified on aerial photographs and hard to discern from remotely-sensed imagery.

5.2 INADEQUACY OF FLOWSET ASSIGNATION

Flowset mapping (Kleman and Borgstrom, 1996) was the original approach used to reconstruct the ice-flow history in the study area, but given the lack of cross-cutting relationships and fragmented flowsets (going against the close-proximity parameter of Kleman et al., 1994; Greenwood and Clark, 2009b; Stokes et al., 2009; Kleman et al., 2010), this approach did not work well to clarify the relative age of flowsets. In northern Manitoba there is an absence of deglacial dates, parallel conformity to eskers, and cross-cutting relationships. We feel that the absence of these characteristics (i.e. only 'event flowsets') may be common beneath the core regions of paleo-ice sheets, and thus the GTZ method is an important tool in these areas. While it is commonly interpreted that similarly-oriented flowsets that are within close proximity are of the same age, it is also possible that similarly-oriented flowsets that are NOT within close proximity are coeval. This does not mean that all similarly-oriented flowsets were generated at the same time, but it is a reasonable assumption when detailed field-derived relative-age relationships for various indicators suggest similar regional histories.

Because of these difficulties, in this region of suspected high inheritance, I turned to pattern recognition in the landscape. This lead to the eventual development of the glacial terrain zone methodology, borrowing from the terrane concept embraced by bedrock geology. This new tool, combined with detailed field-based ice-flow indicator mapping, was used to help determine timelines, subglacial environment at the time of deposition and subglacial processes over time.

5.3 Clast Dispersal and Clast size fraction

At a detailed scale, clast lithology and arsenic dispersal proved useful lines of evidence to indicate regions of greater compositional inheritance. The general patterns of inherited clasts were present in each grain-size class (2 to 4, 4 to 8 and 8+mm) but were most obvious in the 2-4 mm size fraction – reflecting the low clast-concentrations near the tail-end of dispersal patterns. In Chapter 3 complete regional dispersal patterns were not outlined (given the restrictions of the field program), which makes it difficult to fully compare the strength of inherited vs. overprinted dispersal. This limitation is somewhat mitigated by the dispersal pattern of arsenic, presented in section 4.1.3, which better-defines the transport distances related to the younger SE to SW ice-flow phases. It should be noted that these young dispersal trains are also of short distance, suggesting till production and sediment transport were weak throughout the mid to young ice-flow phases.

5.3.1 Dubawnt Supergroup

Some of the Dubawnt Supergroup rocks were incorrectly described in Chapter 3. It should be noted that all Dubawnt clasts are unmetamorphosed. Secondly, Thelon Formation clasts are buff to light pink quartz sandstones and poorly sorted conglomerates, not red metasandstones as identified herein. Thirdly, Christopher Island Formation rocks include white phenocrysts in addition to phlogopite.

Regardless of these issues, I remain confident that all clasts classified as Dubawnt are indeed derived from the Dubawnt Supergroup. These clasts are quite distinctive from the local lithologies, and are non-calcareous. It is unlikely that they could be Paleozoic red units, as carbonate transport to most of the study area is quite rare. Red shale clasts that were quite friable were interpreted as locally-sourced, and only faceted or rounded clasts were attributed as Dubawnt.

5.3.2 Paleozoic carbonate

The source direction(s) and timing of dispersal of carbonate-bearing clasts in northern Manitoba remains an ongoing subject of study. As part of this authors work obligations, data regarding carbonate-bearing clasts and carbonate within till matrix has been compiled into GIS for the entirety of northern Manitoba. As more detailed fieldwork is completed, I believe that more complex dispersal patterns will continue to emerge. Indeed, there is variability and inherited till composition even within expected areas of carbonate-bearing sediments in north-central Manitoba (Trommelen, 2012). More locally, there is some knowledge of interstratified silty and sandy tills along the North Knife River (Dredge and Nixon, 1992). Unfortunately, detailed compositional analyses have not been completed. In 2012, I visited four sites along this river, and collected samples of a few of these tills. Again these were quick visits, due to project constrains, and the data has not-yet been written up. New samples of surface silty tills were also collected just north of the North Knife River.

5.4 ROGEN MORAINE INVESTIGATIONS

Much work remains to be completed on Rogen moraine in an effort to better-understand their genesis. Because these ridges tend to occur in remote regions of Canada, it is difficult to fully explore their internal characteristics. The small number of clast fabrics completed is another limitation, especially since the results were quite variable. As such, this thesis focused more on the outward characteristics including morphology, relationship to other landforms and spatial patterning.

Instability models of Rogen moraine generation best fit the data presented herein. More work is required to better determine how the instability theory of formation relates to physical processes at the subglacial ice-bed interface – I suggest a possible mechanism worth testing is the instability that an evolving subglacial bed mosaic would generate. A correct model to explain the genesis of Rogen moraine must also explain the predominance of drumlins over Rogen moraine throughout most of the subglacial landscape, and the predominance of drumlinized Rogen moraine rather than rogen moraine overprinting of streamlined landforms.

5.4.1 Internal structure

Due to the lack of natural sections and the impossibility of trenching in such a remote northern region, geophysical techniques were used to profile the internal architecture of Rogen moraine in northeast Manitoba. Ground-penetrating radar (GPR) (good penetration in coarse dry sediment; Neal, 2004; Turesson, 2007; Schrott and Sass, 2008; Van Dam, 2012) and shear-wave seismic equipment (good penetration in fine-grained sediments; Dasios et al., 1999; Turesson, 2007) was taken into the field, because the texture and water content variation within target Rogen moraine ridges was unknown. Several GPR surveys were run on Rogen moraine ridges using a pulseEKKO 100 with 100 Mh antennas with one metre separation and a step size of 0.25 m. Each of the test profiles showed significant signal attenuation, most likely due to the higher clay and moisture content in the surficial layer. As such, GPR surveys were abandoned and our efforts focused on the seismic surveys. It is possible that GPR surveys may achieve better results when the ground is still frozen at the surface.

5.5 Ice Sheet Reconstruction

Though a significant amount of detailed work has taken place in northeastern Manitoba during the course of this project, a detailed ice-sheet reconstruction was not produced. The area is complex and the remotely-sensed landscape analysis alone is insufficient to lead to a robust reconstruction. The area of detailed fieldwork is already smaller than the area presented in Chapters 2 and 4. Each foray into a new part of the field area revealed new surprises, including different orientations of ice-flow indicators, and unexpected till compositions. Thus while it was initially desired by this author to attempt a detailed reconstruction, it was later felt that a full regional reconstruction was out of the scope of this thesis and would require new fieldwork beyond what was possible to cover during this project. At this point, I suggest that regional ice sheet reconstructions of the study area (Dredge and Cowan, 1989; Dredge and Nixon, 1992) likely do not reflect the correct interaction between the Keewatin Ice Divide, ice flowing southwest/west from Hudson Bay and northwest from the Quebec/Labradorean sector of the LIS. One existing problem is the assignation of ages to flow orientations, owing to a lack of material datable by radiocarbon (Dyke, 2004; Finlayson, 2012). Other dating methodologies are necessary to further reconstruction in the study area, which could include cosmogenic sampling of various deglacial landforms (Stroeven et al., 2011) or U-Th of buried wood from sections in the Hudson Bay Lowlands (Allard et al., 2012).

Chapter 6 Main Contributions

Detailed field study of this inner core region of the LIS has lead to significant gains in understanding of the glacial landscape in northern Manitoba. By re-examining the predictions of both theory-based and observation-based hypotheses, this thesis has led to important new insights into the regional glacial geology as well as into the paleoglaciology and glacial landscape evolution of inner-core regions of ice sheets. The principle 'key findings' of this PhD. thesis are outlined below.

6.1 Specific northern Manitoba findings

Objectives one and two, outlined in Chapter 1 of this thesis, relate to the palaeoglaciology of northeastern Manitoba. As hypothesized, there is more inheritance in the till composition, ice-flow indicator, landform and landscape record than was previously mapped. Glacial Terrain Zone (GTZ) analysis (Chapter 2) combined the spatial pattern and orientation of streamlined landforms with other spatial characteristics of the glacial landscape (eskers, Rogen moraine, field-based ice-flow indicators) to identify one relict-type, three palimpsest-type and one deglacial-type GTZs. Till-clast and arsenic dispersal patterns (Chapters 3, 4) also exhibit amoeboid dispersal, which combined with the ice-flow indicator record indicate the likely presence of a pre-existing till sheet(s) in northern Manitoba. The mapped spatial mosaic of hybrid tills with different levels of inheritance and overprinting further indicated variable spatio-temporal reworking (comminution and dilution) and preservation. Together, this observation-based evidence has lead to the recognition that northeastern Manitoba is a predominately palimpsest and/or relict subglacial landscape – with minimal ice-retreat or ice-marginal elements – as predicted by theory-based hypotheses for inner core regions of ice sheets.

While the regional configuration of the LIS throughout the Late Wisconsin is still uncertain, important findings related to ice-sheet reconstruction in northern Manitoba include:

- Landforms previously thought to relate to late deglaciation (SE-trending during 9 to 8 ka BP stage, Dyke and Prest, 1987b) are now though to be a much older relict from an early ice-flow phase. Till composition data from this area should be collected to confirm the expected increased concentration of Dubawnt erratics in the Caribou River GTZ.
- The younger southwest ice-flow phase, delineated by streamlined landforms, some arsenic and clast dispersal, and field-based ice-flow indicators, originates up-ice from an area near, or within, Hudson Bay. This new field data, together with rare, old northeast-trending ice-flow indicators,

supports the existence (and migration) of an ice saddle in northeast Manitoba, in the Late Wisconsinan, as depicted by Dyke and Prest (1987b) for their 12 ka BP stage and modeled by Tarasov and Peltier (2004). This ice-flow phase is not recognized by Dredge and Nixon (1992), and together with the point above require significant changes to the current deglacial reconstructions. This ice saddle is consistent with evidence of ice sheet draw-down in northern Hudson Bay linked to a major ice stream operating in Hudson Strait at multiple times during the last glaciation (Andrews and MacLean, 2003; Ross et al., 2011).

- Dispersal of clasts and arsenic within the till matrix is heterogeneous within palimpsest-type Glacial Terrain Zones (GTZs), and dispersal patterns do not locally match the expected glacial history gleaned from streamlined-landform flowsets in the field area. Important for drift exploration, is the knowledge that near areas of Rogen moraine inherited amoeboid dispersal patterns may extend in the opposite ice flow direction to that indicated by the overlying streamlined landform orientations.
- There is a patch of carbonate-bearing till and glaciolacustrine sediment in the Great Island area, which is most likely a remnant, inherited outlier of a penultimate till sheet. The elevated concentrations of both Dubawnt Supergroup clasts (transported 300 to 600 km) and carbonate-bearing clasts (transported 100 km) at these sites suggest that dynamic ice-flow phases capable of transporting sediments over long distances occurred prior to the LGM. If correct, the removal of Dubawnt and carbonate-bearing detritus from other regions of the map area implies there was a significant period of erosion prior to formation of the local amoeboid dispersal patterns. Further detailed fieldwork south, southeast and southwest of the study area is required to determine the maximum extent of calcareous sediment and to identify any ice-flow indicators that may best explain the timing of deposition of these sediments in the Great Island area.
- The time of significant till production in the area is not well constrained, but required either ice-marginal warm-based deposition (pre-LGM advance or retreat phases c.f. (Boulton et al., 2001b)?) or significant warm-based conditions and good till production beneath the thick inner core region of the LIS. Based on the inferred predominance of a pre-existing till sheet(s), and the presence of only poorly-developed ice-marginal landsystems related to the last ice sheet retreat phase, the data support deposition of an 'advance-phase till' followed by low net-erosive conditions for most of the glacial cycle – as predicted by Boulton et al. (2001b). However, the term "advance tills" may be incorrect here, as it is unlikely that significant till production could occur within the inner-core regions during ice advances. Ice sheet build-up in these inner regions appears to

involve cold-based ice and viscous flow in initial stages, until the ice gets thick enough to allow melting at its base (Kleman et al., 2010). As such, I suspect significant till production was more likely to occur during phases of maximum ice thickness (OIS 4 and OIS 2). This reduced areal extent of warm-based conditions in-between phases of maximum ice thickness is also modeled by Stokes et al. (2012).

- Landscape evolution is more complex at a local scale (see section 6.4 below). Based on the presence of ice-flow indicators that transition in orientation (Figure 2-6) from east-southeast through to west-southwest, followed by late deglacial east-southeast ice flow, there were at least brief phases of rapid subglacial landscape evolution – in contrast to predictions by Marshall and Clark (2002), Kleman and Glasser (2007) and Tarasov and Peltier (2007). The long-standing view of long duration radial-flow and slow ice-divide migration, with warm-based ice maintained during translocation, may also require revision.

6.2 DEVELOPMENT OF THE GLACIAL TERRAIN ZONE (GTZ) CONCEPT

The GTZ concept was developed during this PhD as a tool for paleoglaciological glacial landscape analyses – to help 'untangle' the complex, and often contradictory, data mapped in the study area. It builds on similar zonal (c.f. Stea and Finck, 2001) and assemblage (c.f. Ross et al., 2009) concepts, by identifying geologically distinct domains that each contain an internally consistent age, origin and assembly history. The guidelines presented herein use the subglacial landscape to work backwards to an eventual understanding of the spatio-temporal evolution of the landscape beneath the core of ice-sheets. GTZ analysis, together with flowset analysis (c.f. Kleman and Borgstrom, 1996), helps to answer questions such as temporal relationships (where cross-cutting relationships are lacking), degree of inheritance and/or overprinting, and continuity of ice-flow regimes within a specific area.

I suspect that similar regions near or within the outer core region of ice sheets may be similarly complex. Indeed, during the production of this thesis, papers from Scotland (Finlayson et al., 2010; Finlayson, 2012) and Ireland (Knight, 2010) hinted at similar previously unrecognized complexity in the sediment, landform and landscape record. As such, this new methodology should be of interest to the international community.

6.3 ROGEN MORAINE INVESTIGATIONS

As per objective three, outlined in Chapter 1 of this thesis, I investigated the characteristics of these transverse-to ice-flow ridges at landscape-scale (mapping and spatial analysis) and landform-scale (internal structure using high-resolution shear wave (S-wave) seismic reflection surveys, sedimentological characteristics, clast-fabric analyses and terrestrial cosmogenic nuclide analyses of till). Based on this work, I suggest that Rogen moraine may be a type of palimpsest subglacial landform, generated and later preserved during evolution of the subglacial bed mosaic as it encounters varying external (climatic, ice-sheet slope, etc) and internal (bed roughness, topography, sediment texture, porosity, effective water pressure, etc) conditions. Rogen moraine genesis is still largely unknown, though I have put forth a mechanism for regional instability that will hopefully be testable in a computer modeling environment and supported by further field evidence. Furthermore, it may aid our understanding of the subglacial landscape to now view each individual landform field as a record of localized subglacial activity that ultimately resulted in a polygenetic discordant subglacial landscape, rather than as 'transitions' between landforms (c.f. Aario, 1977; Boulton, 1987; Rose, 1987; Dunlop and Clark, 2006a).

6.4 THE SUBGLACIAL BED MOSAIC CONCEPT

The recognition that the subglacial landscape is a fragmented mosaic with different assembly histories, and variable levels of inheritance and overprinting, was first inferred by Stea and Finck (2001), Knight (2010) and Finlayson et al. (2010) and has now been confirmed for the southern Keewatin Sector of the LIS. While this zonal or assemblage concept has been accepted for ice stream and inter-ice stream regions as a paleo-ice stream landsystem (Clark and Stokes, 2003; Ross et al., 2009 and others), we now extend this concept to the entire glacial landscape.

It is generally accepted that micro-scale subglacial deformation may occur at the ice-bed interface as a mosaic of deforming and stable spots (Piotrowski et al., 2004; Tylmann et al., 2012). Hence, I propose it is not a huge leap to up-scale this mosaic of deforming and stable (sticky) spots to a landform or even landform-field scale. These processes may occur everywhere, but far back beneath the ice, near the inner regions of ice sheets, these processes are likely 'slowed down' due to decreased effective water pressure, increased shear stress and the postulated presence of basal permafrost. As per Boulton (1987), it is likely that inner core regions of ice sheets are where the various stages of subglacial bed evolution would be more easily preserved. Examples presented herein suggest how this fragmented landscape could have developed, and outline a subglacial bed mosaic model for the evolution of subglacial landscapes.

6.5 NEW EVIDENCE THAT WILL IMPACT ICE SHEET EVOLUTION MODELS

Fieldwork and analyses in northern Manitoba has led to the idea that landform generation, within the subglacial bed mosaic, likely occurred over multiple ice-flow phases in this inner region of the Laurentide Ice Sheet (LIS), with only limited subglacial erosion, transportation or deposition during deglaciation. This is a similar conclusion to that drawn by Knight (2010) for an inner core region of the British-Irish Ice Sheet. Because relict flowsets may be mistaken for landforms that are part of a deglacial GTZ (e.g. within Caribou River GTZ), it is important to verify age-relationships by all possible proxies before including flowsets into ice-sheet evolution models. Similarly, in relict and palimpsest-type GTZ, it should not be assumed that streamlined landforms parallel to a known deglacial ice-flow orientation are of significant strength to completely overprint the underlying glacial sediment composition.

The presence of significant sticky spots (dewatered, or cold-based), necessary to preserve relict and palimpsest terrain, may lend credence to computer models that suggest the Keewatin Sector of the LIS was cold-based (zero ice-flow velocity) during LGM (Marshall et al., 2000; Marshall and Clark, 2002; Tarasov and Peltier, 2007) or mostly cold-based (Stokes et al., 2012). Future work should involve computer modeling of theoretical sticky and stable spots – at landform and landform field scales, to assess the possibility and implications of an evolving subglacial bed mosaic.

References

Aario, R.O., 1977. Associations of fluting, drumlins, hummocks and transverse ridges. Geojournal 1, 65-72.

Aario, R.O. and Peuraniemi, V., 1992. Glacial dispersal of till constituents in morainic landforms of different types. Geomorphology 6, 9-25.

Allard, G., Roy, M., Ghaleb, B., Richard, P.J.H., Larouche, A.C., Veillette, J.J. and Parent, M., 2012. Constraining the age of the last interglacial-glacial transition in the Hudson Bay lowlands (Canada) using U-Th dating of buried wood. Quaternary Geochronology 7, 37-47.

Alley, R.B., Blankenship, D.D., Rooney, S.T. and Bentley, C.R., 1989. Water-pressure coupling of sliding and bed deformation: III. Application to ice stream B, Antarctica. Journal of Glaciology 35, 130-139.

Alysworth, J.M. and Shilts, W.W., 1989. Bedforms of the Keewatin Ice Sheet, Canada. Sedimentary Geology 62, 407-428.

Anderson, J.B. and Oakes Fretwell, L., 2008. Geomorphology of the onset area of a paleo-ice stream, Marguerite Bay, Antarctic Peninsula. Earth Surface Processes and Landforms 33, 503-512.

Anderson, S.D., Böhm, C.O. and Syme, E.C., 2010a. Far North Geomapping Initiative: bedrock geological investigations in the Seal River region, northeastern Manitoba (parts of NTS 54L, M, 64I, P), In: Manitoba Innovation Energy and Mines Manitoba Geological Survey (Ed.), Report of Activities 2010, pp. 6-22.

Anderson, S.D., Böhm, C.O. and Syme, E.C., 2010b. Precambrian geology of the Seal River region, Manitoba (parts of 54L, M, 64I, P). Manitoba Innovation Energy and Mines Manitoba Geological Survey, Preliminary Map 2010-1, 1:175 000.

Anderson, S.D., Böhm, C.O., Syme, E.C., Carlson, A.R. and Murphy, L.A., 2009a. Bedrock geology of the Great Island area, Manitoba (parts of 54L13, 54M4, 64I15, 16, 64P1,2. Manitoba Innovation Energy and Mines Manitoba Geological Survey, Preliminary Map PMAP2009-4, 1:75 000.

Anderson, S.D., Böhm, C.O., Syme, E.C., Carlson, A.R. and Murphy, L.A., 2009b. Far North Geomapping Initiative: geological investigations in the Great Island area, Manitoba (parts of NTS 54L13, 54M4, 64I15, 16, 64P1,2. Manitoba Innovation Energy and Mines Manitoba Geological Survey, 132-147.

Andrews, J.T. and Miller, G.H., 1979. Glacial erosion and ice sheet divides, northeastern Laurentide Ice Sheet, on the basis of the distribution of limestone erratics. Geology 7, 592-596.

Andrews, J.T. and MacLean, B., 2003. Hudson Strait ice streams: a review of stratigraphy, chronology and links with North Atlantic Heinrich events. Boreas 32, 4-17.

Avery, R., 2010. Technical report on the geology of, and results from, the northwest Manitoba project, In: Avery, R.W., P.Geo (Ed.).

Aylsworth, J.M., 1986. Surficial Geology, Nueltin Lake, District of Keewatin, Northwest Territories. Geological Survey of Canada, 1:125 000.

Aylsworth, J.M. and Shilts, W.W., 1989. Bedforms of the Keewatin Ice Sheet, Canada. Sedimentary Geology 62, 407-428.

Aylsworth, J.M., Cunningham, C.M. and Shilts, W.W., 1990. Surficial geology, Edehon Lake, District of Keewatin, Northwest Territories. Geological Survey of Canada, 1:125 000.

Benn, D.I., 1994. Fabric shape and the interpretation of sedimentary fabric data. Journal of Sedimentary Research A64, 910-915.

Benn, D.I., 1995. Fabric signature of subglacial till deformation, Breidamerkurjokull, Iceland. Sedimentology 42, 735-747.

Benn, D.I. and Evans, D.A., 1996. The interpretation and classification of subglacially-deformed materials. Quaternary Science Reviews 15, 23-52.

Benn, D.I. and Evans, D.A., 1998. Glaciers and Glaciation. Edward Arnold, London.

Bennett, M.R., 2003. Ice streams as the arteries of an ice sheet: their mechanics, stability and significance. Earth Science Reviews 61, 309-339.

Bingham, R.G., Hubbard, A.L., Nienow, P.W. and Sharp, M.J., 2008. An investigation into the mechanisms controlling seasonal speedup events at a High Arctic glacier. Journal of Geophysical Research. 113.

Blake, W.J., 1982. Geological Survey of Canada, radiocarbon dates XXII. Canada, G.S.o., 22.

Blundon, P., Bell, T. and Batterson, M.J., 2010. Ice streaming in the Newfoundland Ice Cap: implications for the reconstruction of ice flow and drift prospecting. Survey, N.a.L.D.o.N.R.G., 143-157.

Bouchard, M.A., 1980. Late Quaternary geology of the Temiscamie area, central Quebec, Canada. Unpublished Ph.D. thesis, McGill University.

Bouchard, M.A., 1989. Subglacial landforms and deposits in central and northern Quebec, Canada, with emphasis on Rogen moraines. Sedimentary Geology, 62, 293-308.

Boulton, G.S., 1987. A theory of drumlin formation by subglacial sediment deformation., In: Menzies, J., Rose, J. (Eds.), Drumlin Symposium, Balkema, Rotterdam,The Netherlands, pp. 25-80.

Boulton, G.S., 1996. Theory of glacial erosion, transport and deposition as a consequence of subglacial sediment deformation. Journal of Glaciology 42, 43-62.

Boulton, G.S. and Hindmarsh, R.C.A., 1987. Sediment deformation beneath glaciers: rheology and geological consequences. Journal of Geophysical Research. 92, 9059-9082.

Boulton, G.S. and Clark, C.D., 1990. A highly mobile Laurentide ice sheet revealed by satellite images of glacial lineations. Nature 346, 813-817.

Boulton, G.S., Dobbie, K.E. and Zatsepin, S., 2001a. Sediment deformation beneath glaciers and its coupling to the hydraulic system. Quaternary International 86, 3-28.

Boulton, G.S., Dongelmans, P., Punkari, M. and Broadgate, M., 2001b. Palaeoglaciology of an ice sheet through a glacial cycle: the European ice sheet through the Weichselian. Quaternary Science Reviews 20, 591-625.

Boulton, G.S., Lunn, R., Vidstrand, P. and Zatsepin, S., 2007. Subglacial drainage by groundwater-channel coupling, and the origin of esker systems: part II - theory and simulation of a modern system. Quaternary Science Reviews 26, 1091-1105.

Boulton, G.S., Hagdorn, M., Maillot, P.B. and Zatsepin, S., 2009. Drainage beneath ice sheets: groundwater-channel coupling, and the origin of esker systems from former ice sheets. Quaternary Science Reviews 28, 621-638.

Bradwell, T., Stoker, M. and Krabbendam, M., 2008. Megagrooves and streamlined bedrock in NW Scotland: The role of ice streams in landscape evolution. Geomorphology 97, 135-156.

Briner, J.P., Miller, G.H., Thompson Davis, P. and Finkel, R.C., 2006. Cosmogenic radionuclides from fjord landscapes support differential erosion by overriding ice sheets. Geological Society of America Bulletin 118, 406-420.

Campbell, J.E., 2001. Phelps Lake project: highlights of the Quaternary investigations in the Bonokoski Lake area (NTS 64M NW). Saskatchewan Geological Survey, *in* Summary of Investigations 2001, Volume 2, Miscellaneous Report 2001-4.2, 19-27.

Campbell, J.E., 2002. Phelps Lake project: highlights of the Quaternary investigations in the Keseechewun Lake area (NTS 64M-9,-10,-15 and -16). Saskatchewan Geological Survey Sask. Industry Resources, Summary of Investigations 2002, Volume 2, Miscellaneous Report 2002-4.2, 16.

Campbell, J.E., Trommelen, M.S., McCurdy, M.W., Böhm, C.O. and Ross, M., 2012. Till composition and ice-flow indicator data, Great Island-Caribou Lake area (parts of NTS 54L, 54M, 64I, and 64P), northeast Manitoba. Geological Survey of Canada, Open File 6967, Manitoba Geological Survey Open File 2011-4, 26 pages, 21 CD-ROM.

Carr, S.J. and Rose, J., 2003. TIll fabric patterns and significance: particle response to subglacial stress. Quaternary Science Reviews 22, 1415-1426.

Chapwanya, M., Clark, C.D. and Fowler, A.C., 2011. Numerical computations of a theoretical model of ribbed moraine formation. Earth Surface Processes and Landforms 36, 1101-1112.

Charbonneau, R. and David, P.P., 1993. Glacial dispersal of rock debris in central Gaspesie, Quebec, Canada. Canadian Journal of Earth Sciences 30, 1697-1707.

Christoffersen, P. and Tulaczyk, S., 2003. Signature of palaeo-ice-stream stagnation: till consolidation induced by basal freeze-on. Boreas 32, 114-129.

Clarhäll, A. and Kleman, J., 1999. Distribution and glaciological implications of relict surfaces on the Ultevis plateau, northwestern Sweden. Annals of Glaciology 28, 202-208.

Clärhall, A. and Jansson, K.N., 2003. Time perspectives on glacial landscape formation - glacial flow chronology at Lac aux Goelands, northeastern Quebec, Canada. Journal of Quaternary Science 18, 441-452.

Clark, C.D., 1993. Mega-scale glacial lineations and cross-cutting ice-flow landforms. Earth Surface Processes and Landforms 18, 1-29.

Clark, C.D., 1999. Glaciodynamic context of subglaical bedform generation and preservation. Annals of Glaciology 28, 23-32.

Clark, C.D., 2010. Emergent drumlins and their clones: from till dilatancy to flow instabilities. Journal of Glaciology 51, 1011-1025.

Clark, C.D. and Meehan, R.T., 2001. Subglacial bedform geomorphology of the Irish Ice Sheet reveals major configuration changes during growth and decay. Journal of Quaternary Science 16, 483-496.

Clark, C.D. and Stokes, C.R., 2003. Paleo-ice stream landsystem, In: Evans, D.A. (Ed.), Glacial Landsystems. Hodder Arnold, London, pp. 205-227.

Clark, C.D., Knight, J. and Gray, J.T., 2000. Geomorphological reconstruction of the Labrador Sector of the Laurentide Ice Sheet. Quaternary Science Reviews 19, 1343-1366.

Clark, P.U., Dyke, A.S., Shakun, J.D., Carlson, A.E., Clark, J., Wohlfarth, B., Mitrovica, J.X., Hostetler, S.W. and McCabe, A.M., 2009. The last glacial maximum. Science 325, 710-714.

Clark, C.D., Hughes, A.L.C., Greenwood, S.L., Jordan, C. and Sejrup, H.P., 2012. Pattern and timing of retreat of the last British-Irish Ice Sheet. Quaternary Science Reviews 44, 112-146.

Clarke, G.K.C., 2005. Subglacial Processes. Annual Review of Earth and Planetary Sciences 33, 247-276.

Colgan, P.M., Mickelson, D.M. and Cutler, P.M., 2003. Ice-marginal terrestrial landsystems: southern Laurentide Ice Sheet margin, In: Evans, D.A. (Ed.), Glacial Landsystems. Hodder Arnold, London, pp. 111-142.

Cowan, W.R., 1968. Ribbed moraine: till-fabric analysis and origin. Canadian Journal of Earth Sciences 5, 1145-1159.

Cunningham, C.M. and Shilts, W.W., 1977. Surficial geology of the Baker Lake area, District of Keewatin. Geological Survey of Canada, 311-314.

Dasios, A., McCann, C., Astin, T.R., MCann, D.N. and Fenning, P., 1999. Seismic imaging of the shallow subsurface: Shear-wave case histories. Geophysical Prospecting 47, 565-591.

De Angelis, H., 2007. Glacial geomorphology of the east-central Canadian Arctic. Journal of Maps, 323-341.

De Angelis, H. and Kleman, J., 2005. Palaeo-ice streams in the northern Keewatin sector of the Laurentide ice sheet. Annals of Glaciology 42, 135-144.

De Angelis, H. and Kleman, J., 2008. Palaeo-ice-stream onsets: examples from the north-eastern Laurentide Ice Sheet. Earth Surface Processes and Landforms 33, 560-572.

Denis, M., Buoncristiani, J. and Guiraud, M., 2009. Fluid-pressure controlled soft-bed deformation sequence beneath the surging Breioamerkurjokull (Iceland, Little Ice Age). Sedimentary Geology, 221, 71-86.

Denton, G., Anderson, R.F., Toggweiler, J.R., Edwards, R.L., Schaefer, J.M. and Putnam, A.E., 2010. The last glacial termination. Science 328, 1652-1656.

Denton, G.H. and Hughes, T.J., 1981. The last great ice sheets. John Wiley and Sons, New York.

DiLabio, R.N.W. and Kaszycki, C.A., Till geochemistry of the Brochet area, NTS 64F, Manitoba. Geological Survey of Canada, 37.

DiLabio, R.N.W., Kaszycki, C.A., Way Nee, V.J. and Nielsen, E., 1986. Surficial geology, Granville Lake, Manitoba. Geological Survey of Canada, 1:125 000.

Dredge, L.A., 1982. Relict ice-scour marks and late phases of Lake Agassiz in northermost Manitoba. Canadian Journal of Earth Sciences 19, 1079-1087.

Dredge, L.A., 1983. Character and development of northern Lake Agassiz and its relation to Keewatin and Hudsonian ice regimes, In: Teller, J.T., Clayton, L. (Eds.), Glacial Lake Agassiz. Geological Association of Canada, pp. 117-131.

Dredge, L.A., 1988. Drift carbonate on the Canadian Shield. II: Carbonate dispersal and ice-flow patterns in northern Manitoba. Canadian Journal of Earth Sciences 25, 783-787.

Dredge, L.A. and Nixon, F.M., 1981a. Surficial geology, Churchill, Manitoba. Geological Survey of Canada, 1:250 000.

Dredge, L.A. and Nixon, F.M., 1981b. Surficial geology, Nejanilini Lake, Manitoba. Geological Survey of Canada, 1:250 000.

Dredge, L.A. and Nixon, F.M., 1981c. Surficial geology, Cape Churchill, Manitoba. Geological Survey of Canada, 1:250 000.

Dredge, L.A. and Nixon, F.M., 1982a. Surficial geology, Shethanei Lake, Manitoba. Geological Survey of Canada, Preliminary Map 6-1980, 1:250 000.

Dredge, L.A. and Nixon, F.M., 1982b. Surficial geology, Caribou River, Manitoba. Geological Survey of Canada, Preliminary Map 5-1980, 1:250 000.

Dredge, L.A. and Nixon, F.M., 1986. Surficial geology, northeastern Manitoba. Geological Survey of Canada, Map 1617A, 1:500 000.

Dredge, L.A. and Thorleifson, L.H., 1987. The Middle Wisconsinan history of the Laurentide Ice Sheet. Geographie physique et Quaternaire 41, 215-235.

Dredge, L.A. and Cowan, W.R., 1989. Quaternary geology of the southwestern Canadian Shield, In: Fulton, R.J. (Ed.), Quaternary geology of Canada and Greenland. Geological Survey of Canada, Geology of Canada, no 1, pp. 214-248.

Dredge, L.A. and Nixon, F.M., 1992. Glacial and environmental geology of northeastern Manitoba. Geological Survey of Canada, Memoir, 80.

Dredge, L.A. and Pehrsson, S.J., 2006. Geochemistry and physical properties of till in northernmost Manitoba (NTS 54E,F,K,L,M; 64I,J,K,N,O,P). Geological Survey of Canada, Open File, 134.

Dredge, L.A. and McMartin, I., 2007. Geochemical reanalysis of archived till samples from northernmost Manitoba. Geological Survey of Canada, 98.

Dredge, L.A. and McMartin, I., 2011. Glacial stratigraphy of northern and central Manitoba. Geological Survey of Canada, Bulletin, 27p.

Dredge, L.A., Nixon, F.M. and Richardson, R.J.H., 1982a. Surficial geology, Kasmere Lake, Manitoba. Geological Survey of Canada, Preliminary Map, Map 19-1981, 1:250 000.

Dredge, L.A., Nixon, F.M. and Richardson, R.J.H., 1982b. Surficial geology, Tadoule Lake, Manitoba. Geological Survey of Canada, Preliminary Map, Map 17-1981, 1:250 000.

Dredge, L.A., Nixon, F.M. and Richardson, R.J.H., 1985. Surficial geology, northwestern Manitoba. Geological Survey of Canada, Map 1608A, 1:500 000.

Dredge, L.A., Nixon, F.M. and Richardson, R.J.H., 1986. Quaternary geology and geomorphology of northwestern Manitoba. Geological Survey of Canada, Memoir, 418, 38.

Dredge, L.A., Morgan, A.V. and Nielsen, E., 1990. Sangamon and pre-Sangamon interglaciations in the Hudson Bay Lowlands of Manitoba. Geographie physique et Quaternaire 44, 319-336.

Dredge, L.A., McMartin, I. and Pyne, M., 2007. Surface materials and landforms, northernmost Manitoba. Geological Survey of Canada, Open File, 24.

Dreimanis, A. and Vagners, U.J., 1971. Bimodal distribution of rock and mineral fragments in basal tills:, In: Goldthwait, R.P. (Ed.), Till, a Symposium. Ohio State University Press, pp. 237-250.

Drummond, R.N., 1965. The glacial geomorphology of the Cambrian Lake area, Labrador-Ungava. Unpublished Ph.D. thesis, McGill University.

Dunlop, P., 2004. The characteristics of ribbed moraine and assessment of theories for their genesis. Unpublished Ph.D. thesis, Department of Geography, The University of Sheffield. 363pp.

Dunlop, P. and Clark, C.D., 2006a. The morphological characteristics of ribbed moraine. Quaternary Science Reviews 25, 1668-1691.

Dunlop, P. and Clark, C.D., 2006b. Distribution of ribbed moraine in the Lac Naococane region, central Quebec, Canada. Journal of Maps, 59-70.

Dunlop, P., Clark, C.D. and Hindmarsh, R.C.A., 2008. Bed ribbing instability explanation: testing a numerical model of ribbed moraine formation arising from coupled flow of ice and subglacial sediment. Journal of Geophysical Research. 113, 15.

Dyke, A.S., 2004. An outline of North American deglaciation with emphasis on central and northern Canada, In: Ehlers, J., Gibbard, P.L. (Eds.), Quaternary Glaciations - Extent and Chronology, Part II. Elsevier B.V., North America, pp. 373-424.

Dyke, A.S., 2008. The Steensby Inlet Ice Stream in the context of the deglaciation of Northern Baffin Island, Eastern Arctic Canada. Earth Surface Processes and Landforms 33, 573-592.

Dyke, A.S. and Prest, V.K., 1987a. Late Wisconsinan and Holocene history of the Laurentide Ice Sheet. Geographie physique et Quaternaire 41, 237-263.

Dyke, A.S. and Prest, V.K., 1987b. Late Wisconsinan and Holocene Retreat of the Laurentide Ice Sheet. Geological Survey of Canada, 1:5 000 000.

Dyke, A.S. and Morris, T.F., 1988. Drumlin fields, dispersal trains, and ice streams in Arctic Canada. Canadian Geographer 32, 86-90.

Dyke, A.S. and Dredge, L.A., 1989. Quaternary geology of the northwestern Canadian Shield, In: Fulton, R.J. (Ed.), Chapter 3 of Quaternary Geology of Canada and Greenland. Geological Survey of Canada, Geology of Canada, no. 1, Ottawa, Ontario, pp. 189-214.

Dyke, A.S. and Evans, D.J.A., 2003. Ice-marginal terrestrial landsystems: northern Laurentide and Innuitian ice sheet margins., In: Evans, D.J.A. (Ed.), Glacial Landsystems. Arnold, London, pp. 143-165.

Dyke, A.S., Dredge, L.A. and Vincent, J.S., 1982. Configuration and dynamics of the Laurentide Ice Sheet during the Late Wisconsinan Maximum. Geographie physique et Quaternaire 36, 5-14.

Dyke, A.S., Moore, A.L. and Robertson, L., 2003. Deglaciation of North America. Geological Survey of Canada, 1:30 000 000.

Dyke, A.S., Morris, T.F., Green, E.C. and England, J.H., 1992. Quaternary geology of Prince of Wales Island, Arctic Canada. Canada, G.S.o.

Dyke, A.S., Andrews, J.T., Clark, P.U., England, J.H., Miller, G.H., Shaw, J. and Veillette, J.J., 2002. The Laurentide and Innuitian ice sheets during the Last Glacial Maximum. Quaternary Science Reviews 21, 9-31.

Evans, D.A., 2003a. Introduction to glacial landsystems, In: Evans, D.A. (Ed.), Glacial Landsystems. Hodder Arnold, London, pp. 1-11.

Evans, D.A., 2003b. Glacial Landsystems. Hodder Arnold, London, p. 532.

Evans, D.A., 2009. Controlled moraines: origins, characteristics and palaeoglaciological implications. Quaternary Science Reviews 28, 183-208.

Evans, D.A., Hiemstra, J.F. and O Cofaigh, C., 2007. An assessment of clast macrofabrics in glacigenic sediments based on A/B plane data. Geografiska Annaler Series a-Physical Geography 89, 103-120.

Evans, D.J.A., Clark, C.D. and Rea, B.R., 2008. Landforms and sediment imprints of fast glacier flow in the southwest Laurentide Ice Sheet. Journal of Quaternary Science 23, 249-272.

Evans, D.J.A., PhillipS, E.R., Hiemstra, J.F. and Auton, C.A., 2006. Subglacial till: Formation, sedimentary characteristics and classification. Earth Science Reviews 78, 115-176.

Eyles, N., 2006. The role of meltwater in glacial processes. Sedimentary Geology 190, 257-268.

Fabel, D., Stroeven, A., Harbor, J., Kleman, J., Elmore, D. and Fink, D., 2002. Landscape preservation under Fennoscandian ice sheets determined from in situ produced ^{10}Be and ^{26}Al. Earth and Planetary Science Letters 201, 397-406.

Fabel, D., Fink, D., Fredin, O., Harbor, J., Land, M. and Stroeven, A., 2006. Exposure ages from relict lateral moraines overridden by the Fennoscandian ice sheet. Quaternary Research 65, 136-146.

Finlayson, A.G., 2012. Ice dynamics and sediment movement: last glacial cycle, Clyde basin, Scotland. Journal of Glaciology 58, 487-500.

Finlayson, A.G. and Bradwell, T., 2008. Morphological characteristics, formation and glaciological significance of Rogen moraine in northern Scotland. Geomorphology 101, 607-617.

Finlayson, A.G., Merrit, J., Browne, M., Merrit, J., McMillan, A. and Whitbread, K., 2010. Ice sheet advance, dynamics, and decay configurations: evidence from west central Scotland. Quaternary Science Reviews 29, 969-988.

Fischer, U.H. and Clarke, G.K.C., 1997. Stick-slip sliding behaviour at the base of a glacier. Annals of Glaciology 24, 390-396.

Fisher, T.G. and Shaw, J., 1992. A depositional model for Rogen moraine, with examples from Avalon Peninsula, Newfoundland. Canadian Journal of Earth Sciences 29, 669-686.

Fortin, R., Coyle, M., Carson, J.M. and Kiss, F., 2009. Airborne geophysical surveys of the Great Island and Seal River area, Manitoba.

Fourriere, A., Claudin, P. and Andreotti, B., 2010. Bedforms in a turbulent stream: formation of ripples by primary linear instability and of dunes by nonlinear pattern coarsening. Journal of Fluid Mechanics 649, 287-328.

Fowler, A.C., 2009. Instability modelling of drumlin formation incorporating lee-side cavity growth. Proceedings of the Royal Society of London, Series A: mathematical, physical and engineering sciences 465, 2681-2702.

Geobase, 2005-2010. GeoBase orthoimage 2005–2010: Manitoba datasets; Canada, N.R., http://www.geobase.ca/geobase/en/find.do?produit=imr, January–June

Glasser, N.F. and Bennett, M.R., 2004. Glacial erosional landforms: origins and significance for palaeoglaciology. Progress in Physical Geography 28, 43-75.

Goodfellow, B.W., Stroeven, A.P., Hättestrand, C., Kleman, J. and Jansson, K.N., 2008. Deciphering a non-glacial/glacial landscape mosaic in the northern Swedish mountains. Geomorphology 93, 213-232.

Grant, A.C. and Sanford, B.V., 1988. Bedrock geology mapping and basin studies in Hudson Bay region, In: Geological Survey of Canada (Ed.), Current research part B, pp. 287-296.

Greenwood, S.L. and Clark, C.D., 2009a. Reconstructing the last Irish Ice Sheet 2: a geomorphologically-driven model of ice sheet growth, retreat and dynamics. Quaternary Science Reviews 28, 3101-3123.

Greenwood, S.L. and Clark, C.D., 2009b. Reconstructing the last Irish Ice Sheet 1: changing flow geometries and ice flow dynamics deciphered from the glacial landform record. Quaternary Science Reviews 28, 3085-3100.

Hall, A.M. and Glasser, N.F., 2003. Reconstructing the basal thermal regime of an ice stream in a landscape of selective linear erosion: Glen Avon, Cairngorm Mountains, Scotland. Boreas 32, 191-207.

Hall, A.M. and Migon, P., 2010. The first stages of erosion by ice sheets: Evidence from central Europe. Geomorphology 123, 349-363.

Harbor, J., Stroeven, A., Fabel, D., Clärhall, A., Kleman, J., Yingkui, L., Elmore, D. and Fink, D., 2006. Cosmogenic nuclide evidence for minimal erosion across two subglacial sliding boundaries of the late glacial Fennoscandian ice sheet. Geomorphology 75, 90-99.

Hart, J.K. and Smith, B., 1997. Subglacial deformation associated with fast ice flow, from the Columbia Glacier, Alaska. Sedimentary Geology 111, 177-197.

Hättestrand, C., 1997. Ribbed moraines in Sweden - distribution pattern and palaeoglaciological implications. Sedimentary Geology, 111, 41-56.

Hättestrand, C. and Kleman, J., 1999. Ribbed moraine formation. Quaternary Science Reviews 18, 43-61.

Hättestrand, C. and Stroeven, A., 2002. A relict landscape in the centre of Fennoscandian glaciation: Geomorphological evidence of minimal Quaternary glacial erosion. Geomorphology 44, 127-143.

Hättestrand, C., Goodwillie, D. and Kleman, J., 1999. Size distribution of two cross-cutting drumlin systems in northern Sweden: a measure of selective erosion and formation time length. Annals of Glaciology 28, 146-152.

Henderson, E.P., 1959. A glaical study of central Quebec-Labrador. Geological Survey of Canada.

Henderson, P.J., 1989. Data report - Description and composition of cores and grab samples, Hudson 87-028, Hudson Bay. Geological Survey of Canada, Open File, 157.

Henderson, P.J. and McMartin, I., 2008. Surficial geology, Flin Flon, Manitoba-Saskatchewan. Geological Survey of Canada, 1:50 000.

Hicock, S.R. and Lian, O.B., 1999. Cordilleran Ice Sheet lobal interactions and glaciotectonic superposition through stadial maxima along a mountain front in southwestern British Columbia, Canada. Boreas 28, 531-542.

Hildes, D.H.D., Clarke, G.K.C., Flowers, G.E. and Marshall, S.J., 2004. Subglacial erosion and englacial sediment transport modelled for North American ice sheets. Quaternary Science Reviews 23, 409-430.

Hindmarsh, R.C.A., 1998. The stability of a viscous till sheet coupled with ice flow, considered at wavelengths less than the ice thickness. Journal of Glaciology 44, 285-292.

Hindmarsh, R.C.A., 1999. Coupled till-ice dynamics and the seeding of drumlins and bedrock forms. Annals of Glaciology 28, 221-230.

Hindmarsh, R.C.A. and Stokes, C.R., 2008. Formation mechanisms for ice-stream lateral shear margin moraines. Earth Surface Processes and Landforms 33, 610-626.

Jansson, K.N., 2005. Map of the glacial geomorphology of north-central Quebec-Labrador, Canada. Journal of Maps, 46-55.

Jansson, K.N., Kleman, J. and Marchant, D.R., 2002. The succession of ice-flow patterns in north-central Quebec-Labrador, Canada. Quaternary Science Reviews 21, 503-523.

Johnson, M.D., Schomacker, A., Benediktsson, I.O., Geiger, A.J., Ferguson, A. and Ingolfsson, O., 2010. Activce drumlin field revealed at the margin of Mulajokull, Iceland: A surge-type glacier. Geology 38, 943-946.

Johnston, W.G.Q. and Schreiner, B.T., 2011. Distribution of distinctive glacial erratics in western Canada, Joint Annual Meeting of Geological Association of Canada, the Mineralogical Association of Canada, the Society of Economic Geologists and the Society for Geology Applied to Mineral Deposits., Ottawa, Ontario.

Kaszycki, C.A., 1989. Surficial Geology and Till Composition, Northwestern Manitoba. Geological Survey of Canada, Open File 50.

Kaszycki, C.A. and Way Nee, V.J., 1989a. Surficial geology, Big Sand Lake, Manitoba (64G). Geological Survey of Canada, 1:125 000.

Kaszycki, C.A. and Way Nee, V.J., 1989b. Surficial geology, Brochet, Manitoba. Geological Survey of Canada, 1:125 000.

Kaszycki, C.A. and Way Nee, V.J., 1989c. Surficial geology, Kississing, Manitoba (63N). Canada, G.S.o., 1:125 000.

Kaszycki, C.A. and Way Nee, V.J., 1989d. Surficial geology, Uhlman Lake, Manitoba (64B). Geological Survey of Canada, 1:125 000.

Kaszycki, C.A. and Way Nee, V.J., 1990a. Surficial geology, Brochet, Manitoba. Geological Survey of Canada, 1:250 000.

Kaszycki, C.A. and Way Nee, V.J., 1990b. Surficial geology, Big Sand Lake, Manitoba. Geological Survey of Canada, 1:250 000.

Kaszycki, C.A. and Way Nee, V.J., 1990c. Surficial geology, Uhlman Lake, Manitoba. Geological Survey of Canada, 1:250 000.

Kaszycki, C.A. and Way Nee, V.J., 1990d. Surficial geology, Granville Lake, Manitoba. Canada, G.S.o., 1:250 000.

Kaszycki, C.A. and Way Nee, V.J., 1990e. Surficial geology, Kississing Lake, Manitoba. Geological Survey of Canada, 1:250 000.

Kaszycki, C.A., Dredge, L.A. and Groom, H., 2008. Surficial geology and glacial history, Lynn Lake - Leaf Rapids area, Manitoba. Geological Survey of Canada, Open File, 105.

Kavanaugh, J.L. and Clarke, G.K.C., 2001. Abrupt glacier motion and reorganization of basal shear stress following the establishment of a connected drainage system. Journal of Glaciology 47, 472-480.

King, E.C., Hindmarsh, R.C.A. and Stokes, C.R., 2009. Formation of mega-scale glacial lineations observed beneath a West Antarctic ice stream. Nature Geoscience 2, 585-588.

Kjaer, K.H. and Kruger, J., 1998. Does clast size influence fabric strength? Journal of Sedimentary Research 68, 746-749.

Kjaer, K.H., Houmark-Nielsen, M. and Richardt, N., 2003. Ice-flow patterns and dispersal of erratics at the southwestern margin of the late Scandinavian Ice Sheet: signature of palaeo-ice streams. Boreas 32, 130-148.

Klassen, R.A., 1995. Drift composition and glacial dispersal trains, Baker Lake area, District of Keewatin, Northwest Territories. Geological Survey of Canada, 68.

Klassen, R.A., 1997. Glacial history and ice flow dynamics applied to drift prospecting and geochemical exploration, In: Gubins, A.G. (Ed.), Proceedings of Exploration 97: Fourth Decennial International Conference on Mineral Exploration, pp. 221-232.

Klassen, R.W., 1983. Lake Agassiz and the late glacial history of northern Manitoba., In: Teller, J.T., Clayton, L. (Eds.), Glacial Lake Agassiz. Geological Association of Canada, pp. 97-115.

Klassen, R.W., 1986. Surficial geology of north-central Manitoba. Geological Survey of Canada, 57.

Klassen, R.W. and Netterville, J.A., 1980. Surficial geology, Kettle Rapids, East of Principal Meridian, Manitoba (54D). Geological Survey of Canada, 1:250 000.

Klassen, R.W. and Netterville, J.A., 1985. Surficial geology, north-central Manitoba. Geological Survey of Canada, 1:500 000.

Kleman, J., 1992. The palimpsest glacial landscape in northwestern Sweden. Geografiska Annaler Series a-Physical Geography 74, 305-325.

Kleman, J. and Borgström, I., 1994. Glacial land forms indicative of a partly frozen bed. Journal of Glaciology 40.

Kleman, J. and Borgstrom, I., 1996. Reconstruction of palaeo-ice sheets: the use of geomorphological data. Earth Surface Processes and Landforms 21, 893-909.

Kleman, J. and Stroeven, A., 1997. Preglacial surface remnants and Quaternary glacial regimes in northwest Sweden. Geomorphology 19, 35-54.

Kleman, J. and Hättestrand, C., 1999. Frozen-bed Fennoscandinavian and Laurentide ice sheets during the Last Glacial Maximum. Nature 402.

Kleman, J. and Glasser, N.F., 2007. The subglacial thermal organization (STO) of ice sheets. Quaternary Science Reviews 26, 585-597.

Kleman, J., Borgström, I. and Hättestrand, C., 1994. Evidence for a relict glacial landscape in Quebec-Labrador. Palaeogeography, Palaeoclimatology, Palaeoecology 111, 217-228.

Kleman, J., Hättestrand, C. and Clärhall, A., 1999. Zooming in on frozen-bed patches: Scale-dependent controls on Fennoscandian ice sheet basal thermal zonation. Annals of Glaciology 28, 189-194.

Kleman, J., Fastook, J. and Stroeven, A., 2002. Geologically and geomorphologically constrained numerical model of Laurentide Ice Sheet inception and build-up. Quaternary International 95-96, 87-98.

Kleman, J., Stroeven, A. and Lundqvist, J., 2008. Patterns of Quaternary ice sheet erosion and deposition in Fennoscandia and a theoretical framework for explanation. Geomorphology 97, 73-90.

Kleman, J., Hättestrand, C., Borgström, I. and Stroeven, A., 1997. Fennoscandian palaeoglaciology reconstructed using a glacial geological inversion model. Journal of Glaciology 43, 283-299.

Kleman, J., Hättestrand, C., Stroeven, A., Jansson, K.N., De Angelis, H. and Borgström, I., 2006. Reconstruction of paleo-ice sheets - inversion of their glacial geomorphological record, In: Knight, P. (Ed.), Glaciology and Earth's Changing Environment. Blackwell, pp. 192-198.

Kleman, J., Jansson, K.N., De Angelis, H., Stroeven, A., Hättestrand, C., Alm, G. and Glasser, N.F., 2010. North American Ice Sheet build-up during the last glacial cycle, 115-21 kyr. Quaternary Science Reviews 29, 2036-2051.

Knight, J., 2010. Basin-scale patterns of subglacial sediment mobility: Implications for glaciological inversion modelling. Sedimentary Geology, 232, 145-160.

Knight, J. and McCabe, A.M., 1997. Identification and significance of ice-flow-transverse subglacial ridges (Rogen moraines) in northern central Ireland. Journal of Quaternary Science 12, 519-524.

Knight, J., McCarron, S.G. and McCabe, A.M., 1999. Landform modification by paleo-ice streams in east-central Ireland. Annals of Glaciology 28, 161-167.

Lenton, P.G. and Kaszycki, C.A., 2005. Till geochemistry in northwestern Manitoba (NTS 63N, 64B, 64F and 64G and parts of 63K, 63O, 64A and 64C). Survey, M.I.E.D.a.M.M.G., 1 CD-ROM.

Lian, O.B. and Hicock, S.R., 2000. Thermal conditions beneath parts of the last Cordilleran Ice Sheet near its centre as inferred from subglacial till, associated sediments, and bedrock. Quaternary International 68-71, 147-162.

Lian, O.B. and Hicock, S.R., 2010. Insight into the character of palaeo-ice-flow in the upland regions of mountain valleys during the last major advance (Vashon Stade) of the Cordilleran Ice Sheet, southwest British Columbia. Boreas 39, 171-186.

Lian, O.B., Hicock, S.R. and Dreimanis, A., 2008. Laurentide and Cordilleran fast ice flow: some sedimentological evidence from Wisconsinan subglacial till and its substrate. Boreas 32, 102-113.

Linden, M., Moller, P. and Adrielsson, L., 2008. Ribbed moraine formed by subglacial folding, thrust stacking and lee-side cavity infill. Boreas 37, 102-131.

Livingstone, S.J., Cofaigh, C.O. and Evans, D.J.A., 2008. Glacial geomorphology of the central sector of the last British-Irish Ice Sheet. Journal of Maps Sp. Iss. 2, 358-377.

Lundqvist, J., 1969. Problems of the so-called Rogen moraine. Swedish Geological Survey, 32.

Lundqvist, J., 1989. Rogen (ribbed) moraine - identification and possible origin. Sedimentary Geology 62, 281-292.

Lundqvist, J., 1997. Rogen moraine - an example of two-step formation of glacial landscapes. Sedimentary Geology 111, 27-40.

Manitoba Energy and Mines, 1980. Churchill, NTS 54L. Manitoba Department of Energy and Mines, M.R.D., Bedrock Geology Compilation Map, 1:250 000.

Marich, A., Batterson, M.J. and Bell, T., 2005. The morphology and sedimentological analyses of rogen moraines, central Avalon Peninsula, Newfoundland. Newfoundland and Labrador Department of Natural Resources, 1-14.

Mark, D.M., 1973. Analysis of axial orientation data, including till fabrics. Geological Society of America Bulletin 84, 1369-1374.

Marshall, S.J. and Clark, P.U., 2002. Basal temperature evolution of North American ice sheets and implications for the 100-kyr cycle. Geophysical Research Letters 29, 769.

Marshall, S.J., Tarasov, L., Clarke, G.K.C. and Peltier, W.R., 2000. Glaciological reconstruction of the Laurentide Ice Sheet: physical processes and modelling challenges. Canadian Journal of Earth Sciences 37, 769-793.

Matile, G.L.D., 2006. Surficial geology of the Kasmere-Putahow lakes area, northwestern Manitoba (parts of NTS 64N10, 11, 14, 15). Manitoba Science Technology Energy and Mines. Manitoba Geological Survey, 1:50 000.

McMartin, I. and Henderson, P., 2004. Evidence from Keewatin (central Nunavut) for paleo-ice divide migration. Geographie physique et Quaternaire 58, 163-186.

McMartin, I., Campbell, J.E., Dredge, L.A. and Robertson, L., 2010. A digital compilation of ice-flow indicators for central Manitoba and Saskatchewan: datasets, digital scalable maps and 1:500 000 scale generalized map. Geological Survey of Canada.

McMartin, I., Dredge, L.A., Grunksy, E. and Pehrsson, S.J., in press. Till geochemistry in west-central Manitoba: interpretation of provenance and mineralization based on glacial history knowledge and principal component analysis. Economic Geology xx, xx.

Meriano, M. and Eyles, N., 2009. Quantitative assesment of the hydraulic role of subglaciofluvial interbeds in promoting deposition of deformation till (Northern Till, Ontario). Quaternary Science Reviews 28, 608-620.

Miller, G.H., Briner, J.P., Lifton, N.A. and Finkel, R.C., 2006. Limited ice-sheet erosion and complex in situ cosmogenic Be-10, Al-26, and C-14 on Baffin Island, Arctic Canada. Quaternary Geochronology 1, 74-85.

Moller, P., 2006. Rogen moraine: an example of glacial reshaping of pre-existing landforms. Quaternary Science Reviews 25, 362-389.

Moller, P., 2010. Melt-out till and ribbed moraine formation, a case study from south Sweden. Sedimentary Geology 232, 161-180.

Murray, T., 1997. Assessing the paradigm shift: deformable glacier beds. Quaternary Science Reviews 16, 995-1016.

Neal, A., 2004. Ground-penetrating radar and its use in sedimentology: principles, problems and progress. Earth Science Reviews 66, 261-330.

Nielsen, E., 1987. Till geochemistry of the Seal River area, east of Great Island, Manitoba. Manitoba Energy and Mines Geological Services, 28.

Nielsen, E., 2001. Quaternary stratigraphy, till provenance and kimberlite indicator mineral surveys along the lower Hayes River. Manitoba Industry Trade and Mines. Manitoba Geological Survey, 121-125.

Nielsen, E., 2002a. Quaternary stratigraphy and ice-flow history along the lower Nelson, Hayes, Gods and Pennycutaway rivers and implications for diamond exploration in northeastern Manitoba, In: Manitoba Industry Trade and Mines. Manitoba Geological Survey (Ed.), Report of Activities 2002, pp. 209-215.

Nielsen, E., 2002b. Pebble lithology of sections in the Hudson Bay Lowland, Manitoba, Canada, Unpublished Raw Data. Manitoba Geological Survey.

Nielsen, E. and Dredge, L.A., 1982. Quaternary stratigraphy and geomorphology of a part of the lower Nelson River, In: Geological Association of Canada Mineralogical Association of Canada (Ed.), p. 56.

Nielsen, E., Morgan, A.V., Morgan, A., Mott, R.J., Rutter, N.W. and Causse, C., 1986. Stratigraphy, paleoecology and glacial history of the Gillam area, Manitoba. Canadian Journal of Earth Sciences 23, 1641-1661.

Nixon, F.M., Dredge, L.A. and Richardson, R.J.H., 1982. Surficial geology, Munroe Lake, Manitoba. Geological Survey of Canada, Preliminary Map 20-1981, 1:250 000.

O Cofaigh, C. and Stokes, C.R., 2008. Reconstructing ice-sheet dynamics from subglacial sediments and landforms: introduction and overview Earth Surface Processes and Landforms 33, 495-502.

O Cofaigh, C., Evans, D.A. and Smith, I.R., 2010. Large-scale reorganization and sedimentation of terrestrial ice streams during late Wisconsinan Laurentide Ice Sheet deglaciation. Geological Association of America Bulletin 122, 743-756.

O Cofaigh, C., Pudsey, C.J., Dowdeswell, J.A. and Morris, P., 2002. Evolution of subglacial bedforms along a paleo-ice stream, Antarctic Peninsula continental shelf. Geophysical Research Letters 29, 1199.

Parent, M., Paradis, S.J. and Boisvert, E., 1995. Ice-flow patterns and glacial transport in the eastern Hudson Bay region: implications for the late Quaternary dynamics of the Laurentide Ice Sheet. Canadian Journal of Earth Sciences 32, 2057-2070.

Parent, M., Paradis, S.J. and Doiron, A., 1996. Palimpsest glacial dispersal trains and their significance for drift prospecting. Journal of Geochemical Exploration 56, 123-140.

Patterson, C.J., 1998. Laurentide glacial landscapes: The role of ice streams. Geology 26, 643-646.

Paul, D., Hanmer, S., Tella, S., Peterson, T.D. and LeCheminant, A.N., 2002. Compilation, bedrock geology of part of the western Churchill Provice, Nunavut - Northwest Territories. Geological Survey of Canada, 1:1000000.

Peltier, W.R., 2004. Global glacial isostasy and the surface of the ice-age earth: the ICE-5G (VM2) Model and GRACE. Annual Review of Earth and Planetary Sciences 32, 111-149.

Peterson, T.D., 2006. Geology of the Dubawnt Lake area, Nunavut-Northwest Territories. Geological Survey of Canada, 56.

Philips, E., Everest, J. and Diaz-Doce, D., 2010. Bedrock controls on subglacial landform distribution and geomorphological processes: Evidence from the Late Devensian Irish Sea Ice Stream. Sedimentary Geology 232, 98-118.

Piotrowski, J.A., Larsen, N.K. and Junge, F.W., 2004. Reflections on soft subglacial beds as a mosaic of deforming and stable spots. Quaternary Science Reviews 23, 993-1000.

Piotrowski, J.A., Larsen, N.K., Menzies, J. and Wysota, W., 2006. Formation of subglacial till under transient bed conditions: deposition, deformation, and basal decoupling under a Weichselian ice sheet lobe, central Poland. Sedimentology 53, 83-106.

Plouffe, A., Bednarski, J.M., Huscroft, C., Anderson, J.B. and McCuaig, S.J., 2011a. Late Wisconsinan glacial history in the Bonaparte Lake map area, south-central British Columbia: implications for glacial transport and mineral exploration. Canadian Journal of Earth Sciences 48, 1091-1111.

Plouffe, A., Anderson, R.G., Gruenwald, W., Davis, W.J., Bednarski, J.M. and Paulen, R.C., 2011b. An integrated approach to boulder tracing: an example from south central British Columbia, Geohydro 2011, Quebec City.

Prest, V.K., Grant, D.R. and Rampton, V.N., 1968. Glacial map of Canada. Geological Survey of Canada, 1: 5000000.

Prest, V.K., Donaldson, J.A. and Mooers, H.D., 2000. The Omar story: The role of Omars in assessing glacial history of west-central North America. Geographie physique et Quaternaire 54, 257-270.

Punkari, M., 1997. Subglacial processes of the Scandinavian ice Sheet in Fennoscandia inferred from flow-parallel features and lithostratigraphy. Sedimentary Geology 111, 263-283.

Raunholm, S., Sejrup, H.P. and Larsen, E., 2003. Lateglacial landform associations at Jaeren (SW Norway) and their glaci-dynamic implications. Boreas 32, 462-475.

Refsnider, K.A. and Miller, G.H., 2010. Reorganization of ice sheet flow patterns in Arctic Canada and the mid-Pleistocene transition. Geophysical Research Letters 37, 5.

Richardson, R.J.H., Dredge, L.A. and Nixon, F.M., 1982. Surficial geology, Whiskey Jack Lake, Manitoba. Geological Survey of Canada, Preliminary Map, Map 18-1981, 1:250 000.

Rose, J., 1987. Drumlins as part of a glacier bedform continuum., In: Menzies, J., Rose, J. (Eds.), Drumlin Symposium. A.A. Balkema, Rotterdam, pp. 103-116.

Ross, M., Lajeunesse, P. and Kosar, K., 2011. The subglacial record of northern Hudson Bay: insights into the Hudson Strait Ice Stream catchment. Boreas 40, 73-91.

Ross, M., Campbell, J.E., Parent, M. and Adams, R.S., 2009. Paleo-ice streams and the subglacial landscape mosaic of the North American mid-continental prairies. Boreas 38, 421-439.

Roy, M., 1998. Pleistocene stratigraphy of the lower Nelson River area: implications for the evolution of the Hudson Bay lowland of Manitoba, Canada. Unpublished M.Sc thesis, Université du Québec à Montreal, Montreal, Quebec, 220 p.

Sarala, P., 2006. Ribbed moraine stratigraphy and formation in southern Finnish Lapland. Journal of Quaternary Science 21, 387-398.

Sarala, P. and Peuraniemi, V., 2007. Exploration using till geochemistry and heavy minerals in the ribbed moraine area of southern Finnish Lapland. Geochemistry, Exploration, Environment, Analysis 7, 195-205.

Schoof, C., 2007. Pressure-dependent viscosity and interfacial instability in coupled ice-sediment flow. Journal of Fluid Mechanics 570, 227-252.

Schrott, L. and Sass, O., 2008. Application of field geophysics in geomorphology: advances and limitations exemplified by case studies. Geomorphology 93, 55-73.

Seaman, A.A., Stea, R.R. and Allard, S., 2011. The Appalachian Glacier Complex in the canadian maritime provinces during the Illinoian and Wisconsinan, Geohydro 2011, Quebec City, QC, Canada, p. 8.

Shaw, J., 1979. Genesis of the Sveg tills and Rogen moraines of central Sweden: a model of basal melt out. Boreas 8, 409-426.

Shaw, J., 2002. The meltwater hypothesis for subglacial bedforms. Quaternary International 90, 5-22.

Shaw, J., Sharpe, D.R. and Harris, J., 2010. A flowline map of glaciated Canada based on remote sensing data. Canadian Journal of Earth Sciences 47, 89-101.

Shilts, W.W., 1980. Flow patterns in the central North American ice sheet. Nature 286, 213-218.

Shilts, W.W., 1993. Geological Survey of Canada's contribution to understanding the composition of glacial sediments. Canadian Journal of Earth Sciences 30, 333-353.

Shilts, W.W., Cunningham, C.M. and Kaszycki, C.A., 1979. Keewatin Ice Sheet - Re-evaluation of the traditional concept of the Laurentide Ice Sheet. Geology 7, 537-541.

Sladen, W.E., 2011. Permafrost. Geological Survey of Canada, Open File 6724,

Smalley, I.J. and Unwin, D.J., 1968. The formation and shape of drumlins and their distribution and orientation in drumlin fields. Journal of Glaciology 7, 377-390.

Smith, A.M., Murray, T., Nicholls, K.W., Makison, K., Adalgeirsdottir, G., Behar, A.E. and Vaughan, D.G., 2007. Rapid erosion, drumlin formation, and changing hydrology beneath an Antarctic ice stream. Geology 35, 127-130.

Smith, J.S., 2006. Northeast Wollaston Lake Project: Quaternary investigations of the Cochrane River (NTS map sheets 64L-10,-11,-14 and -15) and Charcoal Lake (NTS map sheets 64L-9 and -16) areas. Saskatchewan Geological Survey; Saskatchewan Industry Resources, *in* Summary of Investigations 2006, Volume 2, Miscellaneous Report 2006-4.2, Ppaer A-6, 15.

Smith, J.S. and Kaczowka, A., 2007. Northeast Wollaston Lake Project: Quaternary investigations in the Wellbelove Bay - Ross Channel - Rabbabou Bay area, northeast Wollaston Lake, Saskatchewan (Parts of NTS 64L/06, 07, 10 and 11). Saskatchewan Geological Survey; Saskatchewan Ministry of Energy and Resources, *in* Summary of Investigations 2007, Volume 2, Miscellaneous Report 2007-4.2, Paper A-4, 24.

Smith, M.J. and Knight, J., 2011. Palaeoglaciology of the last Irish ice sheet reconstructed from striae evidence. Quaternary Science Reviews 30, 147-160.

Sollid, J.L. and Sorbel, L., 1984. Distribution and genesis of moraines in central Norway. Striae 20, 63-67.

Sollid, J.L. and Sorbel, L., 1994. Distribution of glacial landforms in southern Norway in relation to the thermal regime of the last continental ice sheet. Geografiska Annaler Series a-Physical Geography 76, 25-35.

Staiger, J.W., Gosse, J.C., Little, E.C., Utting, D.J., Finkel, R.C., Johnson, J.V. and Fastook, J., 2006. Glacial erosion and sediment dispersion from detrital cosmogenic nuclide analyses of till. Quaternary Geochronology 1, 29-42.

Stea, R.R., 1994. Relict and palimpsest glacial landforms in Nova Scotia, Canada., In: Warren, W.P., Groots, D.C. (Eds.), Proceedings of the Commission on the Formation and Deformation of Glacial Deposits. Balkema, Dublin, Ireland, pp. 141-148.

Stea, R.R. and Brown, Y., 1989. Variation in drumlin orientation, form and stratigraphy relating to successive ice flows in southern and central Nova Scotia, In: Menzies, J., Rose, J. (Eds.), Subglacial Bedforms - Drumlins, Rogen Moraine and Associated Subglacial Bedforms. Sedimentary Geology, pp. 223-240.

Stea, R.R. and Pe-Pier, G., 1999. Using whole rock geochemistry to locate the source of igneous erratics from drumlins on the Atlantic coast of Nova Scotia. Boreas 28, 308-325.

Stea, R.R. and Finck, P.W., 2001. An evolutionary model of glacial dispersal and till genesis in Maritime Canada, In: McClenaghan, M.B., Bobrowsky, P.T., Hall, G.E.M., Cook, S.J. (Eds.), Drift Exploration in Glaciated Terrain. The Geological Society of London, pp. 237-265.

Stea, R.R., Turner, R.G., Finck, P.W. and Graves, R.M., 1989. Glacial dispersal in Nova Scotia: a zonal concept, In: Dilabio, R.N.W., Coker, W.B. (Eds.), Drift Prospecting. Geological Survey of Canada, pp. 155-169.

Stokes, C.R. and Clark, C.D., 2003a. The Dubawnt Lake palaeo-ice stream: evidence for dynamic ice sheet behaviour on the Canadian Shield and insights regarding the controls on ice-stream location and vigour. Boreas 32, 263-279.

Stokes, C.R. and Clark, C.D., 2003b. Laurentide ice streaming on the Canadian Shield: A conflict with the soft-bedded ice stream paradigm? Geology 31, 347-350.

Stokes, C.R. and Tarasov, L., 2010. Ice streaming in the Laurentide Ice Sheet: A first comparision between data-calibrated numerical model output and geological evidence. Geophysical Research Letters 37, 5.

Stokes, C.R., Clark, C.D. and Storrar, R., 2009. Major changes in ice stream dynamics during deglaciation of the north-western margin of the Laurentide Ice Sheet. Quaternary Science Reviews 28, 721-728.

Stokes, C.R., Tarasov, L. and Dyke, A.S., 2012. Dynamics of the North American Ice Sheet Complex during its inception and build-up to the Last Glacial Maximum. Quaternary Science Reviews 50, 86-104.

Stokes, C.R., Clark, C.D., Lian, O.B. and Tulaczyk, S., 2006. Geomorphological map of ribbed moraine on the Dubawnt Lake Ice Stream bed: a signature of ice stream shut-down? Journal of Maps, 1-9.

Stokes, C.R., Lian, O.B., Tulaczyk, S. and Clark, C.D., 2008. Superimposition of ribbed moraines on a palaeo-ice-stream bed: implications for ice stream dynamics and shutdown. Earth Surface Processes and Landforms 33, 593-609.

Stroeven, A., Fabel, D., Hättestrand, C. and Harbor, J., 2002. A relict landscape in the centre of Fennoscandian glaciation: cosmogenic radionuclide evidence of tors preserved through multiple glacial cycles. Geomorphology 44, 145-154.

Stroeven, A., Fabel, D., Harbor, J., Fink, D., Caffee, M. and Dahlgren, T., 2011. Importance of sampling across an assemblage of glacial landforms for interpreting cosmogenic ages of deglaciation. Quaternary Research 76, 148-156.

Stroeven, A.P., Harbor, J., Fabel, D., Kleman, J., Hättestrand, C., Elmore, D., Fink, D. and Fredin, O., 2006. Slow, patchy landscape evolution in northern Sweden despite repeated ice-sheet glaciation. Tectonics, climate and Landscape Evolution, Geological Society of America Special Papers 398, 387-396.

Sugden, D.E., 1977. Reconstruction of the morphology, dynamics, and thermal characteristics of the Laurentide ice sheet at its maximum. Arctic and Alpine Research 9, 21-47.

Sugden, D.E., 1978. Glacial erosion by the Laurentide Ice Sheet. Journal of Glaciology 20, 367-391.

Sutinen, R., Jakonen, M., Piekkari, M., Haavikko, P., Narhi, P. and Middleton, M., 2010. Electrical-sedimentary anisotrophy of Rogen moraine, Lake Rogen area, Sweden. Sedimentary Geology 232, 181-189.

Taberlet, N., Morris, S.W. and McElwaine, J.N., 2007. Washboard Road: The Dynamics of Granular Rippes Formed by Rolling Wheels. Physical Review Letters 99, 4.

Tarasov, L. and Peltier, W.R., 2004. A geophysically constrained large ensemble analysis of the deglacial history of the North American ice-sheet complex. Quaternary Science Reviews 23, 359-388.

Tarasov, L. and Peltier, W.R., 2007. Coevolution of continental ice cover and permafrost extent over the last glacial-interglacial cycle in North America. Journal of Geophysical Research-Earth Surface 112.

Tella, S., Paul, D., Berman, R.G., Davis, W.J., Peterson, T.D., Pehrsson, S.J. and Kerswill, J.A., 2007. Bedrock geology compliation and regional synthesis of parts of the Hearne and Rae domains, western Churchill Province, Nunavut-Manitoba. Geological Survey of Canada, Open File 5441, 1:550 000.

Teller, J.T. and Leverington, D.W., 2004. Glacial Lake Agassiz: a 5000 yr history of change and its relationship to the d18O record of Greenland. Geological Society of America Bulletin 116, 729-742.

Thorleifson, L.H., 1996. Review of Lake Agassiz history, In: Teller, J.T., Thorleifson, L.H., Matile, G.L.D., Brisbin, W.C. (Eds.), Sedimentology, geomorphology and history of the central Lake Agassiz Basin. Geological Association of Canada pp. 55-84.

Thorleifson, L.H., Wyatt, P.H., Shilts, W.W. and Nielsen, E., 1992. Hudson Bay Lowland Quaternary Stratigraphy: evidence for early Wisconsinan glaciation centred in Quebec, In: Clark, P.U., Lea, P.D. (Eds.), The Last Interglacial-Glacial Transition in North America. Geological Society of America, pp. 207-220.

Tremblay, T., Paulen, R.C. and Lamothe, M., 2011. The role of geomorphological and sedimentological settings in the interpretation of heavy minerals and geochemistry in till, northern Melville Peninsula, Nunavut, Geohydro 2011, Quebec City, QC, Canada.

Trommelen, M.S., 2011a. Surficial Geology, Synder Lake, northwestern Manitoba (parts of NTS 64N5). Manitoba Innovation Energy and Mines Manitoba Geological Survey, Preliminary Map PMAP2011-4, 1:50 000

Trommelen, M.S., 2011b. Far North Geomapping Initiative: Quaternary geology of the Snyder-Grevstad lakes area, far northwestern Manitoba (parts of NTS 64N5), In: Manitoba Innovation Energy and Mines Manitoba Geological Survey (Ed.), Report of Activities 2011, pp. 18-28.

Trommelen, M.S., 2011c. Field-based ice-flow indicator data, Snyder-Grevstad lakes area, northwestern Manitoba (parts of NTS 64N5). Manitoba Innovation Energy and Mines Manitoba Geological Survey, Data Repository Item, Microsoft Excel® file.

Trommelen, M.S., 2012. Quaternary geology of the Knee Lake area, northeastern Manitoba (NTS 53L14, 15, 53M1, 2). Manitoba Innovation Energy and Mines, M.G.S., Report of Activities 2012, GS17, 178-188.

Trommelen, M.S. and Ross, M., 2009. Manitoba Far North Geomapping Initiative: field reconnaissance of surficial sediments, glacial landforms and ice-flow indicators, Great Island and Kellas Lake areas, Manitoba (NTS 54L, 64I, P). Manitoba Innovation Energy and Mines; Manitoba Geological Survey, 148-153.

Trommelen, M.S. and Ross, M., 2010. Subglacial landforms in northern Manitoba, Canada, based on remote sensing data. Journal of Maps 2010, 618-638.

Trommelen, M.S. and Ross, M., 2011a. Far North Geomapping Initiative: palimpsest bedrock macroforms and other complex ice-flow indicators near Churchill, northern Manitoba (part of NTS 54L16), In: Manitoba Innovation Energy and Mines Manitoba Geological Survey (Ed.), Report of Activities 2011, pp. 29-35.

Trommelen, M.S. and Ross, M., 2011b. Field-based ice-flow indicator data, Churchill area, northeastern Manitoba (part of NTS 54L16). Manitoba Innovation Energy and Mines Manitoba Geological Survey, Data Repository Item microsoft excel file.

Trommelen, M.S. and Campbell, J.E., 2012a. Surficial geology, Great Island - Seal River area, Manitoba. Geological Survey of Canada; Manitoba Innovation Energy and Mines Manitoba Geological Survey, Canadian Geoscience Map CGM-42, Geoscience Map MAP2011-3, 1:50 000.

Trommelen, M.S. and Campbell, J.E., 2012b. Surficial geology, Gordon River area, Manitoba. Geological Survey of Canada; Manitoba Innovation Energy and Mines Manitoba Geological Survey, Canadian Geoscience Map CGM-40, Geoscience Map MAP2011-1, 1:50 000.

Trommelen, M.S. and Campbell, J.E., 2012c. Surficial geology, Sosnowski Lake area, Manitoba. Geological Survey of Canada; Manitoba Innovation Energy and Mines Manitoba Geological Survey, Canadian Geoscience Map CGM-43; Geoscientific Map MAP2011-4, 1:50 000.

Trommelen, M.S. and Campbell, J.E., 2012d. Surficial geology, Stubner Lake area, Manitoba. Geological Survey of Canada; Manitoba Innovation Energy and Mines Manitoba Geological Survey, Canadian Geoscience Map CGM-41, Geoscience Map MAP2011-2, 1:50 000.

Trommelen, M.S., Ross, M. and Campbell, J.E., 2010. Far North Geomapping Initiative: Quaternary geology of the Great Island-Kellas Lake area, northern Manitoba (parts of NTS 54L, M, 64I, P). Manitoba Innovation Energy and Mines Manitoba Geological Survey, *in* Report of Activities 2010, GS-3, 36-49.

Trommelen, M.S., Ross, M. and Campbell, J.E., 2012a. Glacial Terrain Zone analysis of a fragmented paleoglaciological record, southeast Keewatin sector of the Laurentide Ice Sheet. Quaternary Science Reviews 40, 1-20.

Trommelen, M.S., Keller, G.R. and Lenton, B.K., 2012b. Quaternary geology of Manitoba: digital compilation of point and line data, with updating of the dataset using remotely-sensed (SPOT) imagery. Manitoba Innovation Energy and Mines, M.G.S., Report of Activities 2012, GS-18, 189-193.

Trommelen, M.S., Ross, M. and Campbell, J.E., accepted. Inherited clast dispersal patterns: implications for paleoglaciology of the southeast Keewatin Sector of the Laurentide Ice Sheet. Boreas xx, xx.

Turesson, A., 2007. Comparative analysis of the multi-offset ground-penetrating radar and shear-wave reflection methods. Journal of Environmental and Engineering Geophysics 12, 163-171.

Turner, R.G. and Stea, R.R., 1987. The Garden of Eden dispersal train, In: Nova Scotia Department of Mines and Energy; Mines and Minerals Branch (Ed.), Report of Activities 1986, pp. 165-169.

Tylmann, K., Piotrowski, J.A. and Wysota, W., 2012. The ice/bed interface mosaic: deforming spots intervening with stable areas under the fringe of the Scandinavian Ice Sheet at Samplawa, Poland. Boreas.

Tyrrell, J.B., 1898a. Report on the Doobount, Kazan, and Ferguson Rivers, and the northwest coast of Hudsons Bay; and on two overland routes from Hudson Bay to Lake Winnipeg. Canada, Geological Survey of Canada, 1-218.

Tyrrell, J.B., 1898b. The glaciation of north central Canada. Journal of Geology 6, 147-160.

United States Geological Survey, 2002. Shuttle Radar Topography Mission, digital topographic data; United States Geological Survey, http://dds.cr.usgs.gov/srtm/,

Van Dam, R.L., 2012. Landform characterization using geophysics - recent advances, applications and emerging tools. Geomorphology 137, 57-73.

van der Meer, J.J.M., Menzies, J. and Rose, J., 2003. Subglacial till: the deforming glacier bed. Quaternary Science Reviews 22, 1659-1685.

Van der Veen, C.J., Plummer, J.C. and Stearns, L.A., 2011. Controls on the recent speed-up of Jakobshavn Isbræ, West Greenland. Journal of Glaciology 57, 770-782.

Van Landeghem, K.J.J., Wheeler, A.J. and Mitchell, N.C., 2009. Seafloor evidence for palaeo-ice streaming and calving of the grounded Irish Sea Ice Stream: implications for the interpretation of its final deglaciation phase. Boreas 38, 119-131.

Veillette, J.J. and Roy, M., 1995. The spectacular cross-striated outcrops of James Bay, Quebec, In: Geological Survey of Canada (Ed.), Current Research 1995-C, pp. 243-248.

Vincent, J.S. and Prest, V.K., 1987. The Early Wisconsinan history of the Laurentide Ice Sheet. Geographie physique et Quaternaire 41, 199-213.

Waller, R.I., van Dijk, T.A.G.P. and Knudsen, O., 2008. Subglacial bedforms and conditions associated with the 1991 surge of Skeioararjokull, Iceland. Boreas 37, 179-194.

Wilch, E. and Hughes, T.J., 2000. Calculating basal thermal zones beneath the Antarctic ice sheet. Journal of Glaciology 46, 297-310.

Winsborrow, M., Clark, C.D. and Stokes, C.R., 2004. Ice streams of the Laurentide Ice Sheet. Geographie physique et Quaternaire 58, 269-280.

Appendix

SUBGLACIAL LANDFORMS IN NORTHERN MANITOBA, CANADA, BASED ON REMOTE SENSING DATA (JOURNAL OF MAPS, V2010, P. 618-638)

SURFICIAL GEOLOGY MAPS FOR THE STUDY AREA

Surficial geology, Gordon River area, Manitoba (GSC CGM-40, MGSMAP2011-1)

Surficial geology, Stubner Lake area, Manitoba (GSC CGM-41, MGS MAP2011-2)

Surficial geology, Great Island – Seal River area, Manitoba (GSC CGM-42, MGS MAP2011-3)

Surficial geology, Sosnowski Lake area, Manitoba (GSC CGM-43, MGS MAP2011-4)

*This paper is published as Trommelen and Ross, Journal of Maps, 2010, pages 618-638

Subglacial landforms in northern Manitoba, Canada, based on remote sensing data

Michelle Trommelen and Martin Ross

Department of Earth and Environmental Sciences, University of Waterloo, Waterloo, Ontario, Canada, N2L 3G1; michelle.trommelen@gmail.com.

Abstract

This paper presents a new subglacial landform map for northern Manitoba (58°-60°N). The region was formerly covered by the Laurentide Ice Sheet and is located at the southern margin of the Late Wisconsinan (~25-10 Ka BP) deglacial Keewatin Ice Divide just west of Hudson Bay. Mapping was focused on determining the location and orientation of streamlined landforms (drumlins and megaflutes), Rogen moraines, and eskers for the 109 366 km^2 region. Based on the theory that landforms such as drumlins and Rogen moraines all result from subglacial ice flow processes, this map forms the basis for reconstruction of ice flow sets that indicate past glacial phases in northern Manitoba. It shows that the geomorphologic record, and hence ice flow history, is more complex than previously reported. There are several successive generations of ice flow indicators superimposed on top of each other, sometimes at abrupt (90°) angles.

Special attention is paid to the location and ridge crest orientation of Rogen moraines in northern Manitoba. Hence the map also provides insights into the characteristics of Rogen moraines in northern Manitoba, which are critical for investigating formative processes. 11,007 individual landforms were mapped using Landsat 7 Enhanced Thematic Mapper Plus (ETM+) satellite imagery in combination with a Shuttle Radar Topography Mission (SRTM) digital elevation model and several SPOT 4 satellite images. The results are presented as a printable map at 1:1 125 000 scale.

Introduction

Northern Manitoba (58°-60°N) is situated at the southern margin of the migrating Late Wisconsinan deglacial Keewatin Ice Divide (Dredge et al. 1986; Aylsworth and Shilts 1989; Dyke and Dredge 1989; Dredge and Nixon 1992; McMartin and Henderson 2004). The map associated with this text shows the detailed location and orientation of streamlined landforms, eskers, and Rogen moraines in northern Manitoba. It builds on previous glacial landform maps by Prest et al. (1968), Dredge et al. (1986), Aylsworth and Shilts (1989), and Dredge and Nixon (1992), and provides detailed landform information in a mapping gap between Nunavut (De Angelis 2007) and central Manitoba (McMartin et al. 2010) (Figure 1). Ice flow studies in this area have been limited, and the landform record as well as ice flow directional data are highly simplified even on existing regional-scale maps (Dredge and Nixon 1981a; b; 1982a; b; c; Dredge et al. 1982a; b; Nixon et al. 1982; Richardson et al. 1982). The availability of Landsat 7 Enhanced Thematic Mapper Plus (ETM+) satellite imagery has allowed for detailed mapping of overlapping features which often cannot be displayed on regional maps. The main purpose of this map is to provide a more detailed, georeferenced, and updated view of the regional glacial landscape of northern Manitoba. This map is well suited for identifying individual ice flow phases (i.e. flow-sets: see Clark 1999) and regional deglaciation features (esker patterns), both of which are essential prerequisites for a detailed reconstruction of the regional glacial history.

The secondary objective of this study is to provide detailed information about the geomorphological characteristics of Rogen (ribbed) moraine ridges, their spatial distribution, and the relationship between Rogen moraine fields and other subglacial landforms in northern Manitoba. It is generally agreed upon that these ridges formed subglacially and transverse to regional ice flow under the Laurentide, Fennoscandian, and British-Irish ice sheets (Cowan 1968; Lundqvist 1969; Shaw 1979; Aylsworth and Shilts 1989; Bouchard 1989; Lundqvist 1989; Fisher and Shaw 1992; Hattestrand 1997; Knight and McCabe 1997; Lundqvist 1997; Hattestrand and Kleman 1999; Clark et al. 2000; Clark and Meehan 2001; Marich et al. 2005; Dunlop and Clark 2006; Moller 2006; Sarala 2006; Finlayson and Bradwell 2008; Linden et al. 2008; Stokes et al. 2008; Moller 2009). However, the interpretation of Rogen (ribbed) moraine ridges is contentious and some authors suggest that all ribbed moraines are formed by a similar mechanism (Hattestrand 1997; Hattestrand and Kleman 1999; Dunlop et al. 2008), while others suggest that various types may have been formed by somewhat different processes or impetuses (c.f. Moller 2006; Linden et al. 2008; Moller 2009). Emphasis was placed on the ridge crest orientation and the presence/absence of drumlinization, as the orientation relative to known ice flow direction and the spatial landscape assemblages are necessary inputs for determining a model to explain the formation of Rogen moraines. The map presented herein thus provides geomorphological observations about Rogen moraines in northern Manitoba and can be used as part of a larger effort to test the various existing models of formation based on the glacial theory. Particularly, knowledge about the relationship between Rogen moraines and other geomorphic landforms in northern Manitoba will strengthen paleoglaciologic reconstructions and may improve development and interpretation of associated Rogen moraine formation models; which will contribute to an overall understanding of ice sheet dynamics near the Keewatin ice divide.

Lastly, drift exploration in Canada is currently carried out by surface sampling along grid lines, without consideration of glacial landforms. Till sampling programs generally assume that the proportion of local detritus increases with depth and is highest just above the bedrock surface (c.f. Klassen 1997; McClenaghan et al. 1997; Stanley 2009). As such, surface sampling typically allows for regional-scale sampling and interpretation. However, several models to explain the formation of Rogen and/or ribbed moraine suggest instead that the moraine ridges may be draped by local detritus and that sediment at depth is further-travelled. Hence, in order to properly interpret till compositional data and make sound decisions based on drift prospecting results in Rogen moraine terrain, a thorough understanding of paleoglaciologic conditions and formation of these landforms is required. While it is premature and out of the scope of this paper to draw conclusions about these questions, we briefly discuss some of the potential processes and implications of key geomorphological features of the map.

Physiography and Bedrock Geology

Northern Manitoba (58°-60°N) is bounded by Hudson Bay in the east, and extends west to the Saskatchewan border, rising in elevation to about 500m a.s.l. (Figure 1).

Figure 1. Physiography of the study area, northern Manitoba, illustrating a rise in topography from east to west. The boundary between the Precambrian basement rocks and the overlying Paleozoic Hudson carbonate platform is located in the southeast corner of the map. Numbers denote the location of the remaining figures.

Northern Manitoba has a gently undulating landscape consisting of glacial landforms, till plains, bedrock outcrops, and abundant small lakes. Streamlined landforms, Rogen moraine ridges, and eskers provide local relief of 2-30 m. The eastern and southernmost parts of the area, which were covered by the Tyrrell Sea and Glacial Lake Agassiz, respectively, form extensive relatively flat areas draped by organic deposits (peat bogs and fens) that often mask glacial landforms (Dredge et al. 1986; Dredge and Nixon 1992).

Most of the study area is underlain by granitoid and granodiorite rocks of the Archean-Paleoproterozoic Hearne domain, Churchill Province, which is a part of the Canadian Shield (Manitoba Geological Survey 2006). Paleoproterozoic metasedimentary and metavolcanic rocks also occur throughout the area. In the southeast, the Precambrian basement is overlain by the Paleozoic Hudson carbonate platform. The boundary between the basement and the carbonate platform can be seen in the southeast corner of Figure 1.

Glacial History

Northern Manitoba has been repeatedly glaciated during the Quaternary, resulting in the formation of a palimpsest landscape (Dredge et al. 1986; Dredge et al. 1990; Dredge and Nixon 1992; Nielson 2001; 2002; Nielson and Fedikow 2002; McMartin et al. 2010). In the Late Wisconsinan deglaciation, the migrating Keewatin Ice Divide was situated just north of the study area and may have extended into Manitoba (Dredge et al. 1986; Aylsworth and Shilts 1989; Dyke and Dredge 1989; Dredge and Nixon 1992; McMartin and Henderson 2004). The Keewatin Ice divide was likely present near the study area in the Early Wisconsinan as well (Boulton and Clark 1990).

Glacial ice of northward (Keewatin) provenance covered most of northern Manitoba (Dredge et al. 1986; Dredge and Nixon 1992; Dredge et al. 2007). Contrastingly, in the southeast portion of the study area (predominately southeast of the dashed line in Figure 1), glacial ice was flowing from an ice divide located to the east (Quebec/Labrador sector of the ice sheet) (Boulton and Clark 1990; Dredge and Nixon 1992).

During deglaciation, Glacial Lake Agassiz abutted the retreating ice margin in the southern portion of the study area, resulting in the deposition of silts, clays, and sands. Incursion of the Tyrrell Sea from Hudson Bay into northeast Manitoba occurred once Glacial Lake Agassiz had drained (Dyke et al. 1982; Dredge et al. 1986; Dyke and Dredge 1989; Dredge and Nixon 1992; Dyke 2004).

Methods

The main map is based on digital mapping of glacial landforms (streamlined landforms, Rogen moraine, and eskers) using the panchromatic band of Landsat Enhanced Thematic Mapper Plus (ETM+) satellite imagery (15 m resolution) and several SPOT 4 satellite images (10 m resolution; Great Island area), at a scale of 1:30 000 in ArcMap (Figure 2). Larger landforms (> 7 km) were also mapped at 1:80 000, using a Shuttle Radar Topography Mission (SRTM) digital elevation model (90 m resolution; Figure 2) to avoid any scale-bias in detection. Mapping of individual landform crests allows for more detailed spatial analysis of the landforms in comparison to previous work (Prest et al. 1968; Dredge and Nixon 1981c; b; a; 1982c; b; a; Dredge et al. 1982a; b; Nixon et al. 1982; Richardson et al. 1982; Aylsworth and Shilts 1989; Shaw et al. 2010), and provides a better basis for reconstruction of the glacial history. Final map presentation is at 1:1 125 000 scale.

Figure 2. Imagery used to create the main map and examples of digitized landforms. A. Arcmap hillshade of the Shuttle Radar Topography Mission (SRTM) digital elevation model (90 m resolution). B. SPOT 4 satellite imagery (10 m resolution). C. Landsat Enhanced Thematic Mapper Plus (ETM+) satellite imagery (15 m resolution).

Previous maps and reports provided useful cross-references for streamlined landform orientations (Prest et al. 1968; Dredge et al. 1986; Dredge and Nixon 1992; Fulton 1995; Matile 2005; 2006). Care was taken to ensure that bedrock structures were not mapped as glacial landforms by referring to various published bedrock maps (Manitoba Industry 2000; 2001; 2002; Manitoba Industry 2005; Anderson et al. 2009). Landform length is estimated using the digitized landforms based on the ETM+ satellite imagery. All field photographs and preliminary sedimentology descriptions are from northeast Manitoba (Great Island – Kellas Lake) visited in 2009 and 2010 (Trommelen and Ross 2009; Trommelen et al. in progress).

Map Description

This map, based on remote sensing data, identifies the regional subglacial landforms present in northern Manitoba. It does not attempt to encompass every single landform, due to the limited resolution of the data, but is meant to provide a regional overview of subglacial landforms. Aerial-photograph based mapping is required to provide further local-scale details.

Streamlined landforms

Streamlined landforms form parallel to the direction of ice flow. The mapped landforms include various types of drumlins and other more elongated ridges here referred to as megaflutings, as well as crag and tail landforms (a tadpole-shaped landform developed by glacial erosion of rocks of unequal resistance). A total of 5136 streamlined landforms were identified, symbolized as a line on the map which extends along the a-axis of each individual landform. These landforms are abundant in northern Manitoba, ranging from subdued low-lying swales in the northeast (Figure 3A), to drumlins in the northwest (Figure 3B), and to well-developed megaflutes in the central north (Figure 3C). Digitized streamlined landforms in northern Manitoba consist of individual ridge segments 0.2-9.8 km long, with a mean segment length of 1.3 km (n=7395).

Figure 3. Streamlined landforms as seen from panchromatic landsat imagery in northern Manitoba (see Figure 1 for locations) A. Subdued low-lying streamlined landforms in north-central Manitoba (NTS 64P). Note the cross-cutting relationship with older landforms (dashed). B. Drumlins south of Kasmere Lake (NTS 64N). Note the esker in the bottom right, aligned sub-parallel to ice flow. C. Streamlined landforms just north of Quinn Lake, likely formed during the Quinn Lake re-advance into Glacial Lake Agassiz around 8300 BP (Dredge et al. 1986).

Most streamlined landforms are of low relief in the northeast and higher relief in the northwest (Dredge et al. 1986; Dredge and Nixon 1992) with elongation ratios (length:width; not shown on map; n=200) that range from 1:1 to 3:1. Several differently-oriented sets of large crag and tail landforms exist, four to nine kilometers in length with elongation ratios of 3:1 to 4:1 (n=10). In the northwest, and at Shethani Lake (UTM 567450, 6518940), there are numerous drumlins and megaflutes with lengths that range from one to four kilometers and have elongation ratios of 4:1 to 7.5:1 (n=100).

Rogen Moraine

Rogen moraines are characterized by “anastomosing to curved ridges and intervening troughs, all lying transverse to former ice-flow direction” (Lundqvist 1969; 1989), and exhibit a gradual up-and/or down-ice flow-direction transition to drumlins (Lundqvist 1969; Boulton 1987; Bouchard 1989; Hattestrand and Kleman 1999) and/or a non-transitional lateral shift to streamlined terrain (c.f. Alysworth and Shilts 1989). Rogen moraines in northern Manitoba consist of individual ridge segments 0.1-9.3 km long, with a typical segment length of 0.5-3.2 km (mean=1.8 km, n=2870*)*. The ridge segments (2870 mapped) are found throughout northern Manitoba, almost exclusively north of 59° (see map). These Rogen moraine ridges are classified as jagged, anastomosing, broad arcuate, downstream curving, and barchan shaped (Figure 4), following Dunlop and Clark's (2006) terminology.

Figure 4. Rogen moraine types in northern Manitoba as seen from panchromatic landsat imagery. Classifications are modified from Dunlop and Clark (2006) A. Pristine Rogen moraine, jagged type. B. Pristine Rogen moraine, anastamosing. C. Pristine Rogen moraine, anastamosing type transitioning to drumlinized rogen moraine, broad arcuate. D. Drumlinized Rogen moraine, downstream curving. E. Drumlinized Rogen moraine, broad arcuate. F. Drumlinized Rogen moraine, barchan shaped.

Pristine Rogen Moraine

Rogen moraines that are not significantly drumlinized, but still part of a ribbed moraine – streamlined terrain landscape assemblage, are herein referred to as pristine Rogen moraine ridges (Figures 4A,B,C and 5) (1260 mapped segments, 44% of total).

Figure 5. Rogen moraine north of Sosnowski Lake, northeast Manitoba (See Figure 1 for locations). A. Anastamosing Rogen moraine with shallow lakes between ridges. B. Down-ice side of a ridge, person for scale. C. Undulating surface of a ridge, illustrating the high surface boulder content.

Drumlinized Rogen Moraine

Ribbed moraine ridges that have been overridden by actively flowing ice, resulting in streamlining of their surfaces (Figures 4 D,E,F and 6) are herein termed "drumlinized Rogen moraine" (1610 mapped segments, 56% of total). The degree and size of streamlining/modification is and often transitional, ranging from complete drumlinization of the surface (Figure 6A) to minor modification resulting in superimposition of small drumlins (Figure 6 B).

Figure 6. Drumlinized Rogen moraine north of Sosnowski Lake in northern Manitoba. (see Figure 1 for locations) A. Transverse ridge with a modified, planed surface. Ice flow was from top left to bottom right. Looking closely, it is possible to see the steep down-ice ‘cliffs’ of the ridges. B. Minor drumlinized surface of several Rogen ridges; ice flow was towards the helicopter tail. C. Undulating surface of a ridge, illustrating the lower surface boulder content.

Eskers

Eskers are prevalent throughout northern Manitoba, and 742 individual segments have been mapped (Figure 7). Eskers are elongated and more or less sinuous sand and gravel ridges that are a record of the dendritic interconnected drainage network that existed underneath the Laurentide Ice Sheet over hard subglacial substrates (c.f. Boulton et al. 2009). Most of the mapped eskers are large ‘main’ ridges, which are often fed by smaller tributary eskers. Esker segments mapped range in length from 0.2 to 98.0 km, with an average length of 4.8 km (n=742). The total ‘connected’ main ridges are often 150-200 km in length. Smaller eskers are present in northern Manitoba, but have not been identified on this map as they are not visible at the resolution of the Landsat imagery. Likewise, some eskers occur within tunnel channels or meltwater corridors, but that level of detail is beyond the scope of this map.

Figure 7. Several esker ridges in northeast Manitoba; see Figure 1 for locations. Figure B is a wave-washed esker, washed by the postglacial Tyrrell Sea.

Discussion

Streamlined landforms in northern Manitoba, formed parallel to paleoice-flow, show approximately ten different spatial orientations. The deviations in landform orientation provides supporting evidence that the study area was affected by the migrating Keewatin Ice Divide, which continued to cause fluctuations in ice flow direction. Only a few glacial landforms are mapped in portions of the southern and eastern study area, because these areas are masked by sediments deposited by Glacial Lake Agassiz and the Tyrrell Sea (Dredge et al. 1986; Dredge and Nixon 1992). Generally, elongation ratios of >10:1 are thought to represent landforms created by fast-flowing ice (c.f. Hart and Smith 1997; O Cofaigh et al. 2002; Stokes and Clark 2002; Bradwell et al. 2008; Andreassen and Winsborrow 2009). Thus the lower the elongation ratio, the slower or more sluggish the ice flow. Using these definitions, most of the streamlined landforms in northern Manitoba were formed during slow or sluggish ice flow conditions, as would be expected near the Keewatin Ice Divide.

Implications for Rogen moraine formation and ice flow dynamics

This is the first regional study of Rogen moraine ridges that highlights not only the orientation and shape of the ridge crests, but the spatial landscape assemblage between pristine Rogen moraines, drumlizined rogen moraines and streamlined terrain. If, as numerous papers suggest, Rogen moraine ridges are formed by slow, sluggish ice (Aario 1977; Boulton 1987; Dyke et al. 1992; Hattestrand and Kleman 1999; Stokes et al. 2006; Linden et al. 2008; Stokes et al. 2008; Fowler 2009), then the spatially-variable modification by later ice flow suggests a change in the regional ice dynamics to faster-flowing ice at some point in time. Currently, we are uncertain whether the pristine Rogen moraines are preserved more or less randomly or whether the pristine Rogen moraine fields are akin to local sticky spots of stiffened well-drained till or cold-based ice. The map shows that there are up-ice, down-ice, and abrupt lateral transitions between pristine and drumlinized Rogen moraine ridges, as well as streamlined landforms (Figure 8). There is not a single observation of Rogen moraine overprinted on eskers. Additionally, Rogen moraine ridge orientation is often oblique to esker ridge orientation, rather than transverse as would be expected of a coeval landform. This fact, combined with the variable overprinting by streamlined landforms, suggests that the Rogen moraines are not ice-marginal deglacial deposits, and but rather were formed subglacially further back under the ice sheet. A lack of Rogen moraine ridges exists in the central region, most likely due to extensive drumlinization during an enhanced flow of a lobe into Glacial Lake Agassiz (c.f. Dredge et al. 1986). The large, widespread esker network in northern Manitoba suggests that deglaciation was warm-based, and that the ice sheet was thin and non-accumulating.

Figure 8. Patchy glacial landscape in northern Manitoba (see Figure 1 for locations). Here there are localized patches of pristine Rogen moraine (orange), streamlined landforms (yellow), and drumlinized Rogen moraine (white). The boundaries between these fields range from abrupt to gradational. Eskers variably drape or cut all landforms as they were formed during the deglacial phase. The lower two images demonstrate the spatial relationship between Rogen moraine fields and streamlined landforms. In figure B there is an abrupt lateral transition between swaths of Rogen moraine and streamlined terrain. The streamlined landforms clearly overprint Rogen moraines which is more obvious near the edges of the swaths. Figure C illustrates a down-ice transition from streamlined landforms to Rogen moraines that vary in the degree of overprint/modification.

West of Sand Lake (UTM 623264, 6629099) and between Nueltin and Nejanilini lakes (see map, UTM 506600, 6612425), there are large crag and tail landforms oriented to the NE and ESE, respectively. These orientations deviate significantly from all other streamlined landform orientations in northern Manitoba (SE to SW). Comparison with published ice flow data from the surrounding regions suggests that the crag and tail landforms may be related to an Early-Mid Wisconsinan glaciation where ice flowed east into Hudson Bay (Boulton and Clark 1990; McMartin and Henderson 2004). The map also shows that most Rogen moraine ridges are oriented parallel to the nearby larger and older crag and tail landforms, and perpendicular to the smaller and younger streamlined landforms that overprint their surfaces.

Conclusions

Subglacial landform mapping in northern Manitoba reflects a significantly more complex glacial record than recognized in earlier publications. The palimpsest landform record suggests at least seven different ice flow phases may have occurred, some of which may be pre-Late Wisconsinan. While all areas of Manitoba appear to have been affected by multi-directional ice flow, not all ice flow directions are preserved in the landform record in any one area. Further fieldwork, including the collection of striation and other ice flow indicator data, is required to determine the spatial extent of each ice flow phase. Fieldwork should also help ascertain whether there is a correlation between the orientation of older (pre-Late Wisconsinan?) crag and tail landforms and Rogen moraine ridges. Interpretations based on the landform map and additional fieldwork will have implications for ice flow history, Keewatin Ice Divide migration, ice flow dynamics (sticky spots) and Rogen moraine formation. Additionally, these interpretations will enable better analysis of till compositional data collected during drift prospecting programs in Rogen moraine terrain, in northern Manitoba and in other similar regions of Canada.

Software

All imagery was manipulated in ESRI ArcMap. Glacial landforms are stored as line shapefiles. Map preparation and final production was accomplished using Corel Draw.

Acknowledgements

This map is part of a PhD project at the University of Waterloo. M. Trommelen is supported by the Geological Survey of Canada's (GSC) Research Affiliate Program. The authors are grateful for the logistical support provided by the Manitoba Geological Survey and the Geological Survey of Canada (GSC) through the Geomapping for Energy and Minerals Program (GEM) Manitoba Far North project. Computer and software infrastructure was funded by the Canada Foundation for Innovation (CFI). Discussion about the glacial history and Quaternary geology of northern Manitoba and surrounding regions was kindly provided by L. Dredge, I. McMartin, and J. Campbell of the GSC. This manuscript has benefited from internal reviews by L. Dredge and J. Campbell. D. Berg, P. Dunlop and D.S. Sharpe are thanked for their critical and thought-provoking external reviews, which have greatly enhanced this manuscript.

Map Design

A hillshade derived from the SRTM imagery was used as the background, to provide the reader with an indication of the topography and landform record in northern Manitoba. As such, lakes are grey, not blue, to allow for better color differentiation from the background.

References

Aario, R.O. 1977: Associations of fluting, drumlins, hummocks and transverse ridges; Geojournal, v. 1: p. 65-72. doi:10.1007/BF00195540

Alysworth, J.M. and Shilts, W.W. 1989: Bedforms of the Keewatin Ice Sheet, Canada; Sedimentary Geology, v. 62: p. 407-428.

Anderson, S.D., Bohm, C.O., Syme, E.C., Carlson, A.R. and Murphy, L.A. 2009: Bedrock geology of the Great Island area, Manitoba (parts of 54L13, 54M4, 64I15, 16, 64P1,2, Manitoba Innovation Energy and Mines Manitoba Geological Survey. Preliminary Map PMAP2009-4. 1:75 000.

Andreassen, K. and Winsborrow, M. 2009: Signature of ice streaming in Bjornoyrenna, Polar North Atlantic, through the Pleistocene and implications for ice-stream dynamics; Annals of Glaciology, v. 50,(52): p. 17-26. doi:10.3189/172756409789624238

Aylsworth, J.M. and Shilts, W.W. 1989: Bedforms of the Keewatin Ice Sheet, Canada; Sedimentary Geology, v. 62: p. 407-428. doi:10.1016/0037-0738(89)90129-2

Bouchard, M.A. 1989: Subglacial landforms and deposits in central and northern Quebec, Canada, with emphasis on Rogen moraines; Sedimentary Geology, v. 62: p. 293-308. doi:10.1016/0037-0738(89)90120-6

Boulton, G. S. 1987: A theory of drumlin formation by subglacial sediment deformation. *In* Drumlin Symposium. *Edited by* J Menzies and J. Rose. Balkema, Rotterdam,The Netherlands25-80

Boulton, G. S. and Clark, C. D. 1990: A highly mobile Laurentide ice sheet revealed by satellite images of glacial lineations; Nature, v. 346: p. 813-817. doi:10.1038/346813a0

Boulton, G. S., Hagdorn, M., Maillot, P.B. and Zatsepin, S. 2009: Drainage beneath ice sheets: groundwater-channel coupling, and the origin of esker systems from former ice sheets; Quaternary Science Reviews, v. 28: p. 621-638. doi:10.1016/j.quascirev.2008.05.009

Bradwell, T., Stoker, M. and Krabbendam, M. 2008: Megagrooves and streamlined bedrock in NW Scotland: The role of ice streams in landscape evolution; Geomorphology, v. 97: p. 135-156. doi:10.1016/j.geomorph.2007.02.040

Clark, C. D. 1999: Glaciodynamic context of subglaical bedform generation and preservation.; Annals of Glaciology, v. 28: p. 23-32.

Clark, C. D., Knight, J and Gray, J.T. 2000: Geomorphological reconstruction of the Labrador Sector of the Laurentide Ice Sheet; Quaternary Science Reviews, v. 19: p. 1343-1366. doi:10.1016/S0277-3791(99)00098-0

Clark, C. D. and Meehan, R.T. 2001: Subglacial bedform geomorphology of the Irish Ice Sheet reveals major configuration changes during growth and decay; Journal of Quaternary Science, v. 16,(5): p. 483-496. doi:10.1002/jqs.627

Cowan, W.R. 1968: Ribbed moraine: till-fabric analysis and origin; Canadian Journal of Earth Sciences, v. 5: p. 1145-1159.

De Angelis, H. 2007: Glacial geomorphology of the east-central Canadian Arctic; Journal of Maps, v.: p. 323-341. doi.org/10.4113/jom.2007.90

Dredge, L. A., McMartin, I. and Pyne, M. 2007: Surface materials and landforms, northernmost Manitoba; *in* Open File 5435, Geological Survey of Canada; p. 24, 1 sheet (1:500 000).

Dredge, L. A., Morgan, A.V. and Nielson, E. 1990: Sangamon and pre-Sangamon interglaciations in the Hudson Bay Lowlands of Manitoba; Geographie Physique et Quaternaire, v. 44,(3): p. 319-336.

Dredge, L. A. and Nixon, F.M. 1981a: Surficial geology, Cape Churchill, Manitoba, Geological Survey of Canada. Preliminary Map 3-1980. 1:250 000.

Dredge, L. A. and Nixon, F.M. 1981b: Surficial geology, Churchill, Manitoba, Geological Survey of Canada. Preliminary Map 4-1980. 1:250 000.

Dredge, L. A. and Nixon, F.M. 1981c: Surficial geology, Nejanilini Lake, Manitoba, Geological Survey of Canada. Preliminary Map 7-1980. 1:250 000.

Dredge, L. A. and Nixon, F.M. 1982a: Surficial geology, Caribou River, Manitoba, Geological Survey of Canada. Preliminary Map 5-1980. 1:250 000.

Dredge, L. A. and Nixon, F.M. 1982b: Surficial geology, Shethanei Lake, Manitoba, Geological Survey of Canada. Preliminary Map 6-1980. 1:250 000.

Dredge, L. A. and Nixon, F.M. 1982c: Surficial Geology, York Factory, East of Second Meridian-east, Manitoba, Geological Survey of Canada. Preliminary Map 2-1980. 1:250 000.

Dredge, L. A. and Nixon, F.M. 1992: Glacial and environmental geology of northeastern Manitoba; *in* Memoir 432, Geological Survey of Canada; p. 80.

Dredge, L. A., Nixon, F.M. and Richardson, R.J.H. 1982a: Surficial geology, Kasmere Lake, Manitoba, Geological Survey of Canada. Prelimary Map, Map 19-1981. 1:250 000.

Dredge, L. A., Nixon, F.M. and Richardson, R.J.H. 1982b: Surficial geology, Tadoule Lake, Manitoba, Geological Survey of Canada. Preliminary Map, Map 17-1981. 1:250 000.

Dredge, L. A., Nixon, F.M. and Richardson, R.J.H. 1986: Quaternary geology and geomorphology of northwestern Manitoba; *in* Memoir 418, Geological Survey of Canada; p. 38.

Dunlop, P. and Clark, C. D. 2006: The morphological characteristics of ribbed moraine; Quaternary Science Reviews, v. 25: p. 1668-1691. doi:10.1016/j.quascirev.2006.01.002

Dunlop, P., Clark, C. D. and Hindmarsh, R.C.A. 2008: Bed ribbing instability explanation: testing a numerical model of ribbed moraine formation arising from coupled flow of ice and subglacial sediment; Journal of Geophysical Research., v. 113: p. 15. doi:10.1029/2007JF000954

Dyke, A. S. 2004: An outline of North American deglaciation with emphasis on central and northern Canada. *In* Quaternary Glaciations - Extent and Chronology, Part II. *Edited by* J. Ehlers and P.L. Gibbard. Elsevier B.V., North America. Development in Quaternary Science Series. 373-424.doi:10.1016/S1571-0866(04)80209-4

Dyke, A. S. and Dredge, L. A. 1989: Quaternary geology of the northwestern Canadian Shield. *In* Chapter 3 of Quaternary Geology of Canada and Greenland. *Edited by* R. J. Fulton. Geological Survey of Canada, Geology of Canada, no. 1, Ottawa, Ontario189-214

Dyke, A. S., Dredge, L. A. and Vincent, J. S. 1982: Configuration and dynamics of the Laurentide Ice Sheet during the Late Wisconsinan Maximum; Geographie physique et Quaternaire, v. 36,(1-2): p. 5-14.

Dyke, A. S., Morris, T.F., Green, E.C. and England, J. H. 1992: Quaternary geology of Prince of Wales Island, Arctic Canada; *in* Memoir 433, Geological Survey of Canada; p.

Finlayson, A.G. and Bradwell, T. 2008: Morphological characteristics, formation and glaciological significance of Rogen moraine in northern Scotland.; Geomorphology, v. 101: p. 607-617. doi:10.1016/j.geomorph.2008.02.013

Fisher, T. G. and Shaw, J. 1992: A depositional model for Rogen moraine, with examples from Avalon Peninsula, Newfoundland.; Canadian Journal of Earth Sciences, v. 29: p. 669-686.

Fowler, A.C. 2009: Instability modelling of drumlin formation incorporating lee-side cavity growth; Proceedings of the Royal Society of London, Series A: mathematical, physical and engineering sciences, v. 465,(2109): p. 2681-2702. doi:10.1098/rspa.2008.0490

Fulton, R. J., compiler,. 1995: Surficial Materials of Canada, Map 1880A. 1:5,000,000.

Hart, J. K. and Smith, B. 1997: Subglacial deformation associated with fast ice flow, from the Columbia Glacier, Alaska; Sedimentary Geology, v. 111: p. 177-197. doi:10.1016/S0037-0738(97)00014-6

Hattestrand, C. 1997: Ribbed moraines in Sweden - distribution pattern and palaeoglaciological implications; Sedimentary Geology, v. 111: p. 41-56. doi:10.1016/S0037-0738(97)00005-5

Hattestrand, C. and Kleman, J. 1999: Ribbed moraine formation; Quaternary Science Reviews, v. 18: p. 43-61. doi:10.1016/S0277-3791(97)00094-2

Klassen, R. A. 1997. Glacial history and ice flow dynamics applied to drift prospecting and geochemical exploration. *In* Proceedings of Exploration 97: Fourth Decennial International Conference on Mineral Exploration. 221-232.

Knight, J and McCabe, A.M. 1997: Identification and significance of ice-flow-transverse subglacial ridges (Rogen moraines) in northern central Ireland; Journal of Quaternary Science, v. 12,(6): p. 519-524. doi:10.1002/(SICI)1099-1417(199711/12)12:6<519::AID-JQS313>3.0.CO;2-Q

Linden, M., Moller, P. and Adrielsson, L. 2008: Ribbed moraine formed by subglacial folding, thrust stacking and lee-side cavity infill; Boreas, v. 37: p. 102-131. doi:10.1111/j.1502-3885.2007.00002.x

Lundqvist, J. 1969: Problems of the so-called Rogen moraine.; *in* Series C, 648, Swedish Geological Survey; p. 32.

Lundqvist, J. 1989: Rogen (ribbed) moraine - identification and possible origin; Sedimentary Geology, v. 62: p. 281-292. doi:10.1016/0037-0738(89)90119-X

Lundqvist, J. 1997: Rogen moraine - an example of two-step formation of glacial landscapes; Sedimentary Geology, v. 111: p. 27-40. doi:10.1016/S0037-0738(97)00004-3

Manitoba Geological Survey. 2006: Geology of Manitoba, 1979 Based on Manitoba Geological Services, Geological map of Manitoba. Map 79-2. 1:1 000 000.

Manitoba Industry, Economic Development and Mines, Manitoba Geological Survey. 2005: Caribou River, NTS 54M, Economic Development and Mines Manitoba Industry, Manitoba Geological Survey. Bedrock Geology Compilation Map 54M. 1:250 000.

Manitoba Industry, Trade and Mines, Geological Survey. 2000: Kasmere Lake, NTS 64N, Trade and Mines Manitoba Industry, Geological Survey. Bedrock Geology Compilation Map 64N. 1:250 000.

Manitoba Industry, Trade and Mines, Manitoba Geological Survey. 2001: Munroe Lake, NTS 64O, Trade and Mines Manitoba Industry, Geological Survey'. Bedrock Geology Compilation Map 64O. 1:250 000.

Manitoba Industry, Trade and Mines, Manitoba Geological Survey. 2002: Nejanilini Lake, NTS 64P, Trade and Mines Manitoba Industry, Manitoba Geological Survey'. 1:250 000.

Marich, A., Batterson, M. J. and Bell, T. 2005: The morphology and sedimentological analyses of rogen moraines, central Avalon Peninsula, Newfoundland; *in* Current Research 2005, Report 05-1, Newfoundland and Labrador Department of Natural Resources; p. 1-14.

Matile, G.L.D. 2005: Surficial geology of Nejanilini Lake, Manitoba (parts of NTS 64P5, 12 and 13), Manitoba Science Manitoba Geological Survey, Technology, Energy and Mines. Preliminary Map PMAP2005-4. 1:35 000.

Matile, G.L.D. 2006: Surficial geology of the Kasmere-Putahow lakes area, northwestern Manitoba (parts of NTS 64N10, 11, 14, 15), Manitoba Science Manitoba Geological Survey, Technology, Energy and Mines. Preliminary Map PMAP2006-6. 1:50 000.

McClenaghan, M. B., Thorleifson, L. H. and DiLabio, R. N. W. 1997. Till geochemistry and indicator mineral methods in mineral exploration. *In* Proceedings of Exploration 97: Fourth Decennial International Conference on Mineral Exploration. 233-248.

McMartin, I., Campbell, J E, Dredge, L. A. and Robertson, L. 2010: A digital compilation of ice-flow indicators for central Manitoba and Saskatchewan: datasets, digital scalable maps and 1:500 000 scale generalized map; *in* Open File 6405, Geological Survey of Canada; p.

McMartin, I. and Henderson, P. 2004: Evidence from Keewatin (Central Nunavut) for Paleo-Ice Divide Migration; Geographie physique et Quaternaire, v. 58,(2-3): p. 163-186.

Moller, P. 2006: Rogen moraine: an example of glacial reshaping of pre-existing landforms; Quaternary Science Reviews, v. 25: p. 362-389. doi:10.1016/j.quascirev.2005.01.011

Moller, P. 2009: Melt-out till and ribbed moraine formation, a case study from south Sweden; Sedimentary Geology, v. in press: p. 20 p. doi:10.1016/j.sedgeo.2009.11.003

Nielson, E. 2001: Quaternary stratigraphy, till provenance and kimberlite indicator mineral surveys along the lower Hayes River; *in* GS-18, Trade and Mines Manitoba Industry, Manitoba Geological Survey; p. 121-125.

Nielson, E. 2002: Quaternary stratigraphy and ice-flow history along the lower Nelson, Hayes, Gods and Pennycutaway rivers and implications for diamond exploration in northeastern Manitoba; Trade and Mines Manitoba Industry, Manitoba Geological Survey; p. 209-215.

Nielson, E. and Fedikow, M.A.F. 2002: Kimberlite indicator-mineral surveys, lower Hayes River; *in* GP2002-1, Trade and Mines Manitoba Industry, Geological Survey; p.

Nixon, F.M., Dredge, L. A. and Richardson, R.J.H. 1982: Surficial geology, Munroe Lake, Manitoba, Geological Survey of Canada. Preliminary Map 20-1981. 1:250 000.

O Cofaigh, C., Pudsey, C.J., Dowdeswell, J. A. and Morris, P. 2002: Evolution of subglacial bedforms along a paleo-ice stream, Antarctic Peninsula continental shelf; Geophysical Research Letters, v. 29,(8): p. 1199. doi:10.1029/2001GL014488

Prest, V. K., Grant, D.R. and Rampton, V. N. 1968: Glacial map of Canada, Geological Survey of Canada. A Series Map 1253A. 1: 5000000.

Richardson, R.J.H., Dredge, L. A. and Nixon, F.M. 1982: Surficial geology, Whiskey Jack Lake, Manitoba, Geological Survey of Canada. Preliminary Map, Map 18-1981. 1:250 000.

Sarala, P. 2006: Ribbed moraine stratigraphy and formation in southern Finnish Lapland; Journal of Quaternary Science, v. 21,(4): p. 387-398. doi:10.1002/jqs.995

Shaw, J. 1979: Genesis of the Sveg tills and Rogen moraines of central Sweden: a model of basal melt out; Boreas, v. 8: p. 409-426. doi:10.1111/j.1502-3885.1979.tb00437.x

Shaw, J., Sharpe, D.R. and Harris, J. 2010: A flowline map of glaciated Canada based on remote sensing data; Canadian Journal of Earth Sciences, v. 47,(1): p. 89-101. doi:10.1139/E09-068

Stanley, C. R. 2009: Geochemical, mineralogical, and lithological dispersal models in glacial till: physical process constraints and application in mineral exploration. *In* Application of till and stream sediment heavy mineral and geochemical methods to mineral exploration in western and northern Canada. *Edited by* R. C. Paulen and I. McMartin. Geological Association of Canada. GAC Short Course Notes 18.

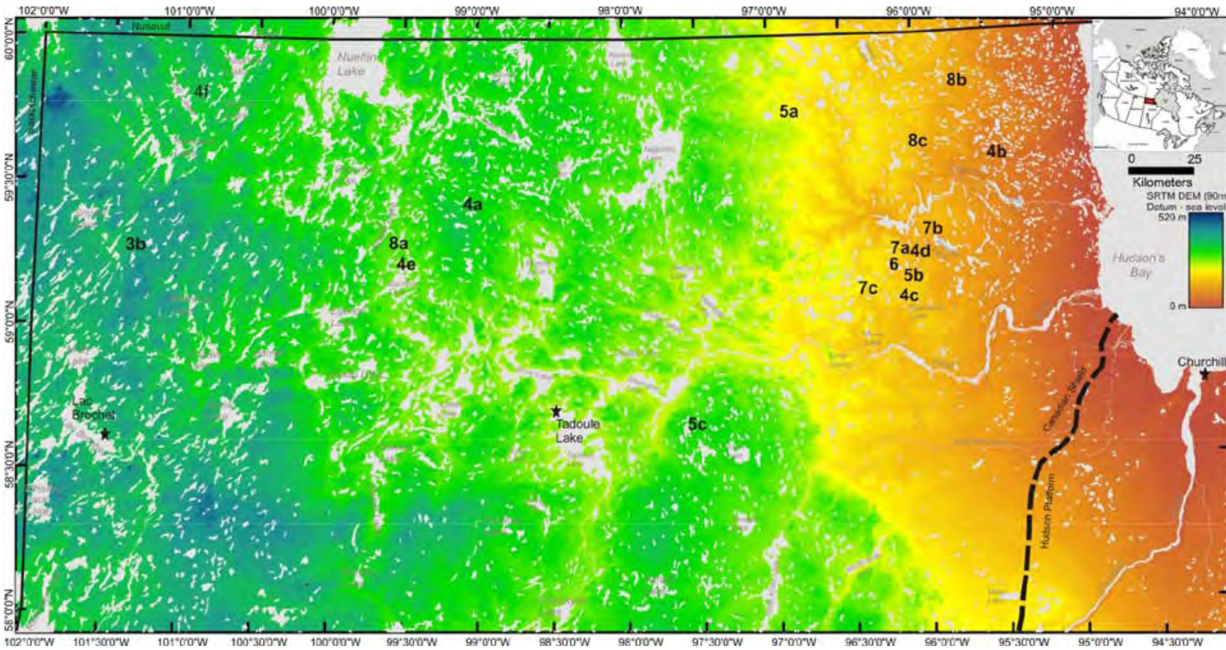
Stokes, C.R. and Clark, C. D. 2002: Are long subglacial bedforms indicative of fast ice flow?; Boreas, v. 31,(3): p. 239-249. doi:10.1080/030094802760260355

Stokes, C.R., Clark, C. D., Lian, O. B. and Tulaczyk, S. 2006: Geomorphological map of ribbed moraine on the Dubawnt Lake Ice Stream bed: a signature of ice stream shut-down?; Journal of Maps, v.: p. 1-9. doi:10.4113/jom.2006.43

Stokes, C.R., Lian, O. B., Tulaczyk, S. and Clark, C. D. 2008: Superimposition of ribbed moraines on a palaeo-ice-stream bed: implications for ice stream dynamics and shutdown; Earth Surface Processes and Landforms, v. 33,(4): p. 593-609. 10.1002/esp.1671

Trommelen, M. S. and Ross, M. 2009: Manitoba Far North Geomapping Initiative: field reconnaissance of surficial sediments, glacial landforms and ice-flow indicators, Great Island and Kellas Lake areas, Manitoba (NTS 54L, 64I, P); *in* Report of Activities 2009, GS-14, Manitoba Innovation Energy and Mines; Manitoba Geological Survey; p. 148-153.

Trommelen, M. S., Ross, M and Campbell, J E. in progress: Far North Geomapping Initiative: Quaternary geology of the Great Island - Kellas Lake region, northern Manitoba (parts of NTS 54L, 54M, 64I, 64P); *in* Report of Activities 2010, GS-X17, Manitoba Innovation Energy and Mines Manitoba Geological Survey; p. xx-xx.



Nunavut
Nueltin Lake
Lac Brochet
Tadoule Lake
Churchill
Hudson's Bay
Canadian Shield
Hudson Platform
Kilometers
SRTM DEM (90m)
Datum - sea level
520 m
0 m
4f
3b
4a
8a
4e
5a
8b
8c
4b
7b
7a
4d
6
5b
7c
4c
5c

Figure 1

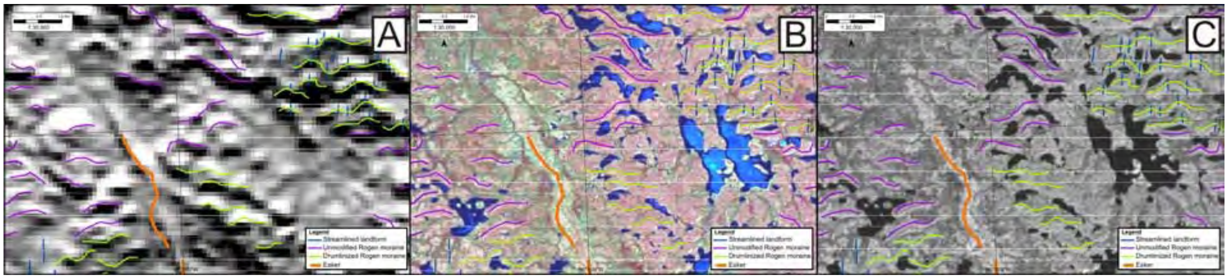

A
Legend
Streamlined landform
Unmodified Rogen moraine
Drumlinized Rogen moraine
Esker
B
Legend
Streamlined landform
Unmodified Rogen moraine
Drumlinized Rogen moraine
Esker
C
Legend
Streamlined landform
Unmodified Rogen moraine
Drumlinized Rogen moraine
Esker

Figure 2

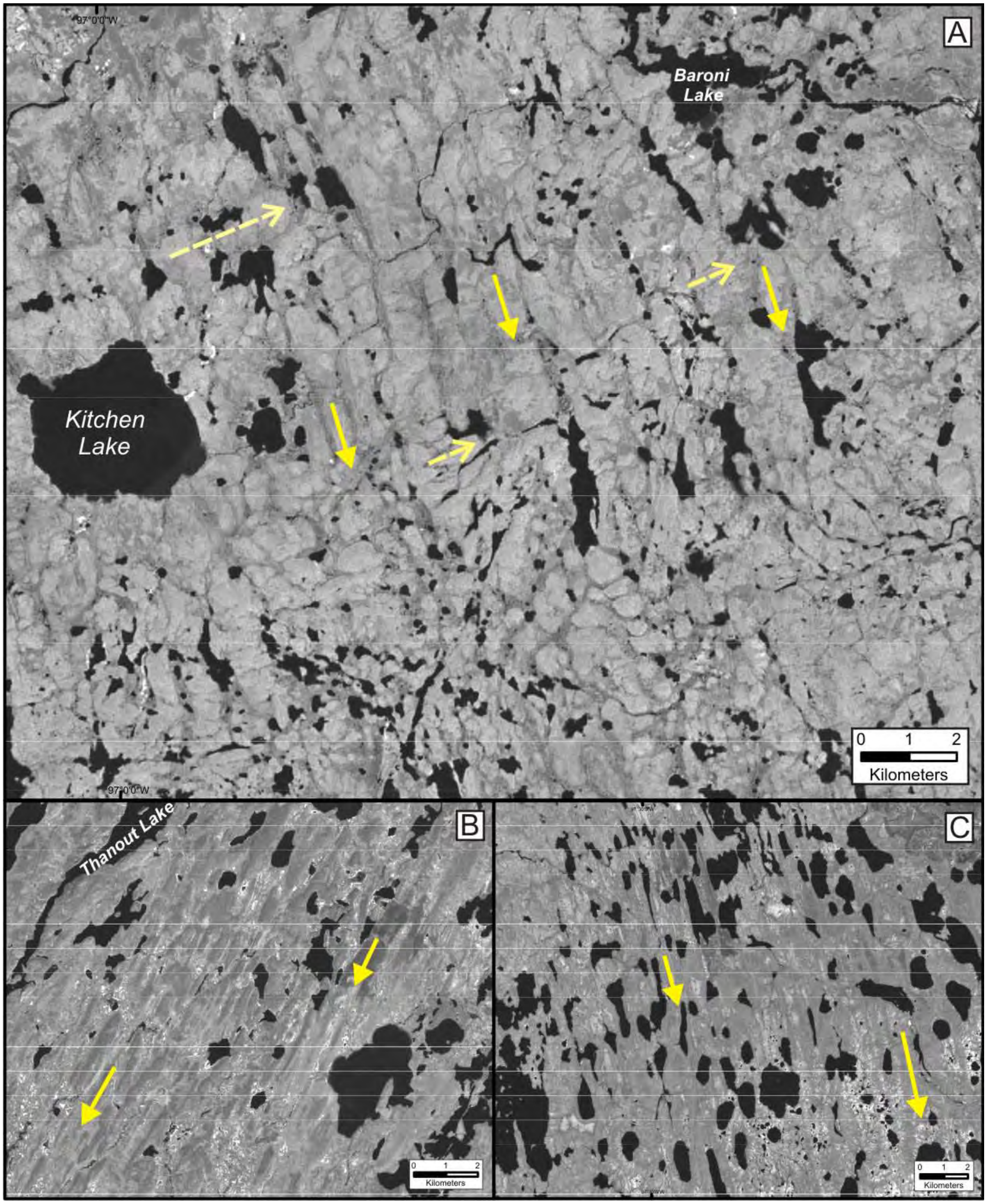


Figure 3

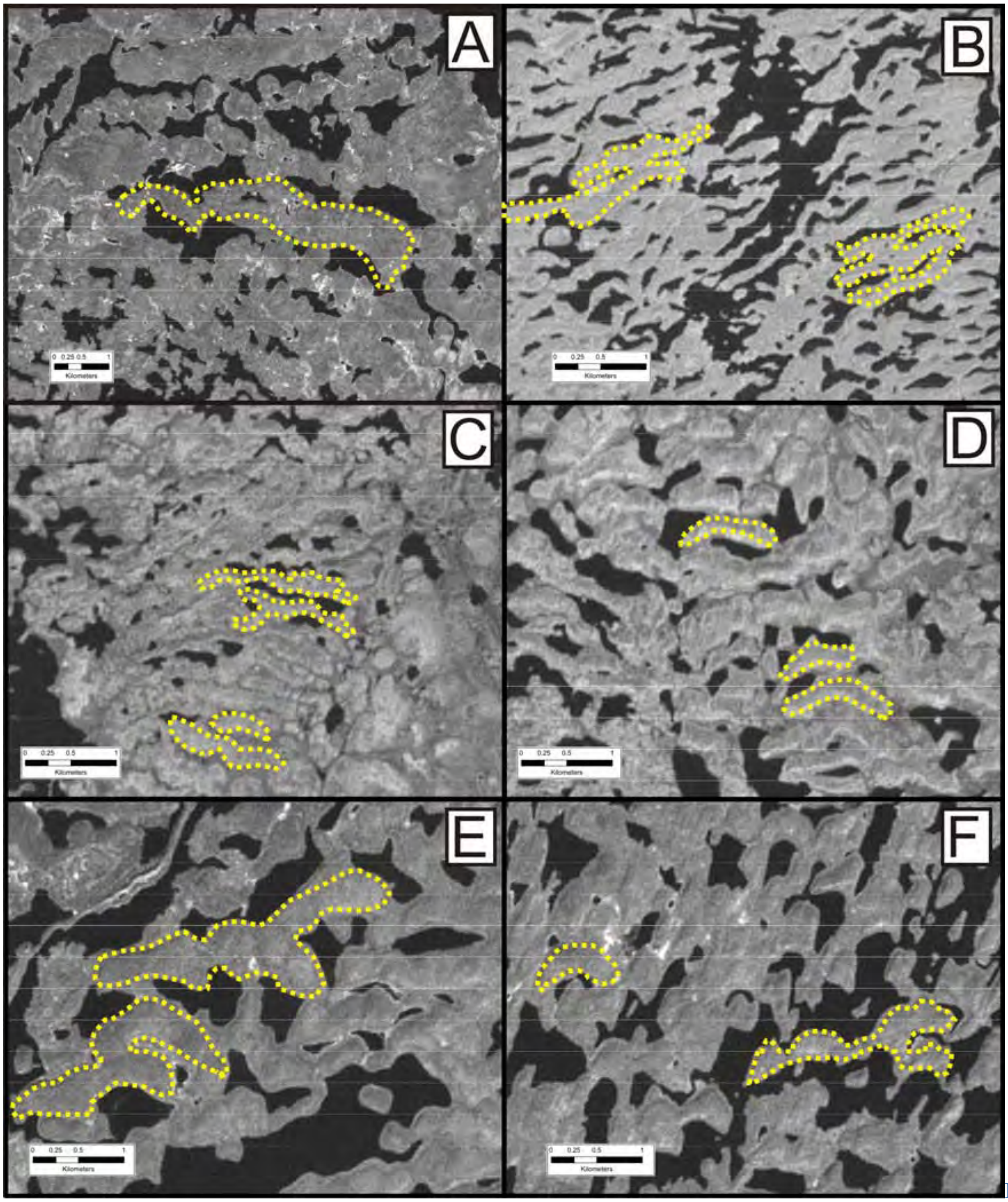


Figure 4

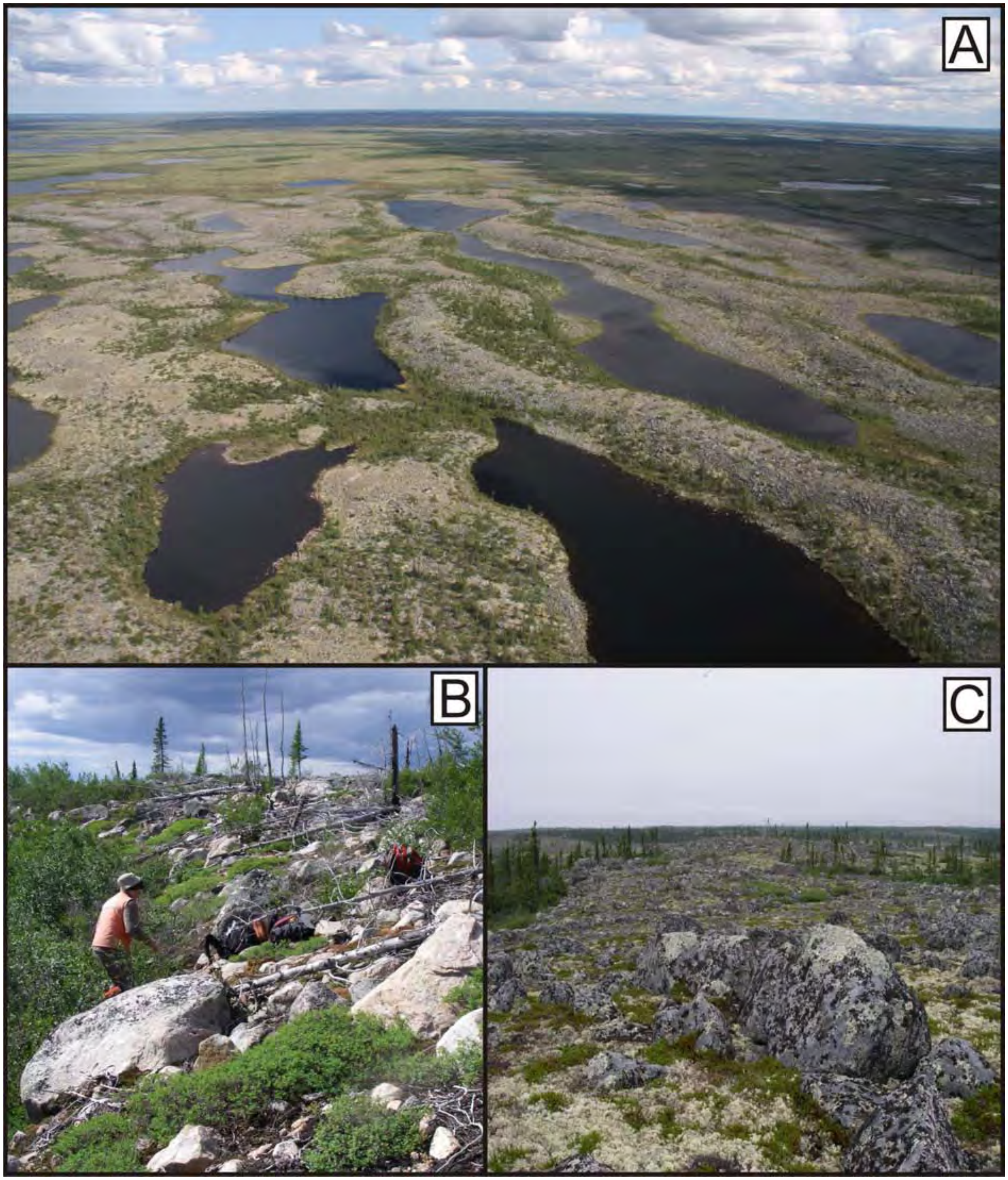


Figure 5

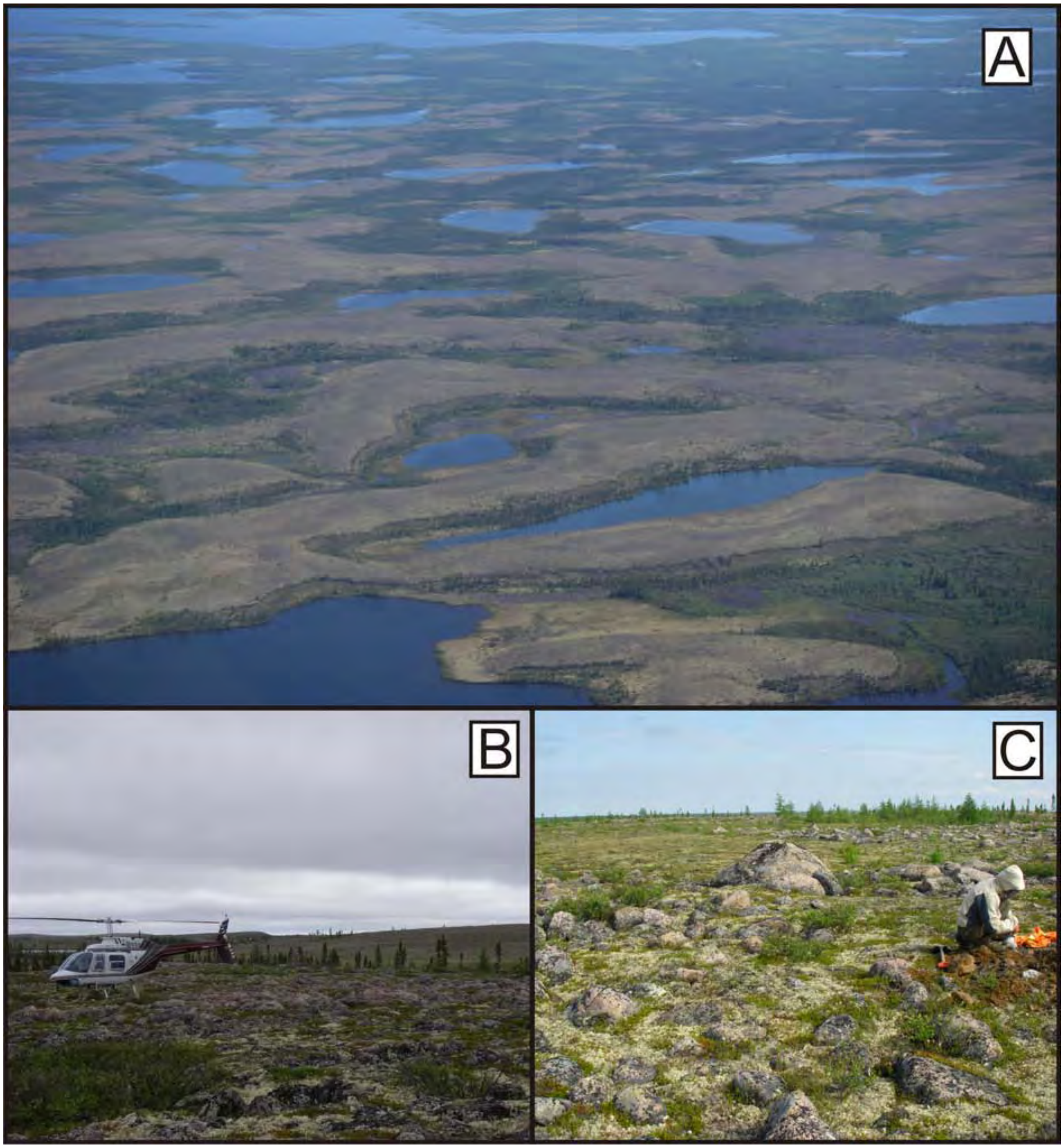


Figure 6

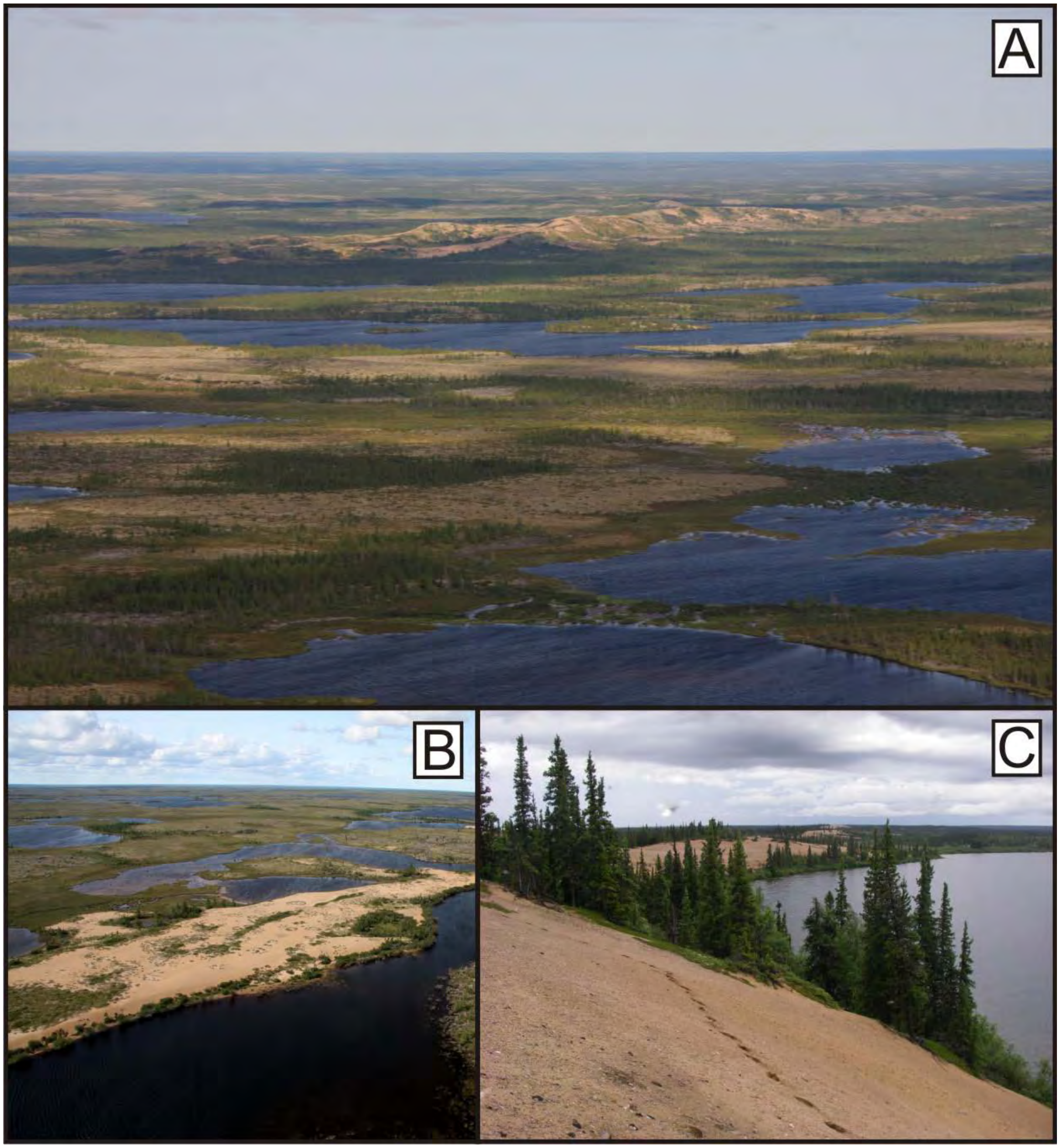
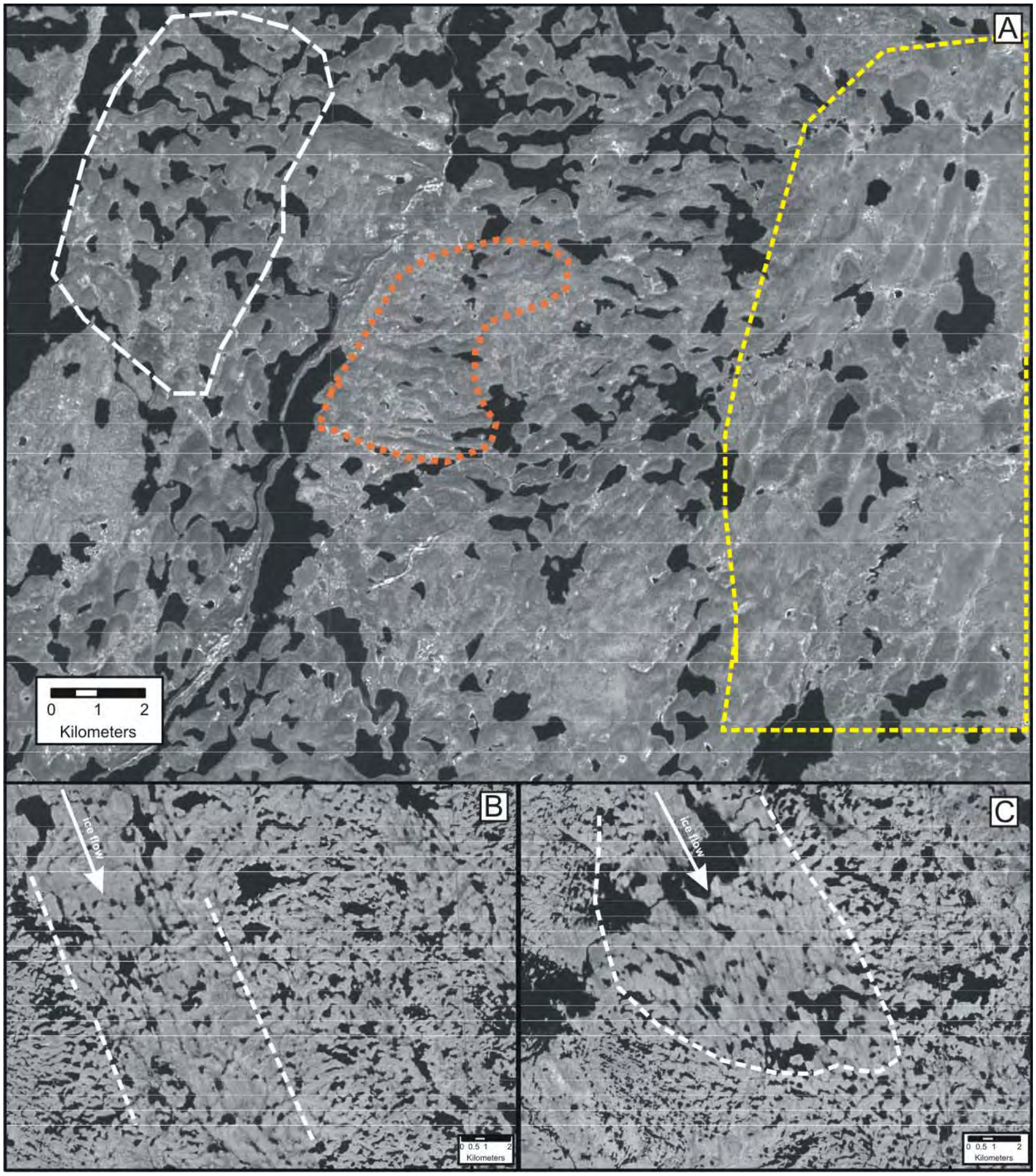
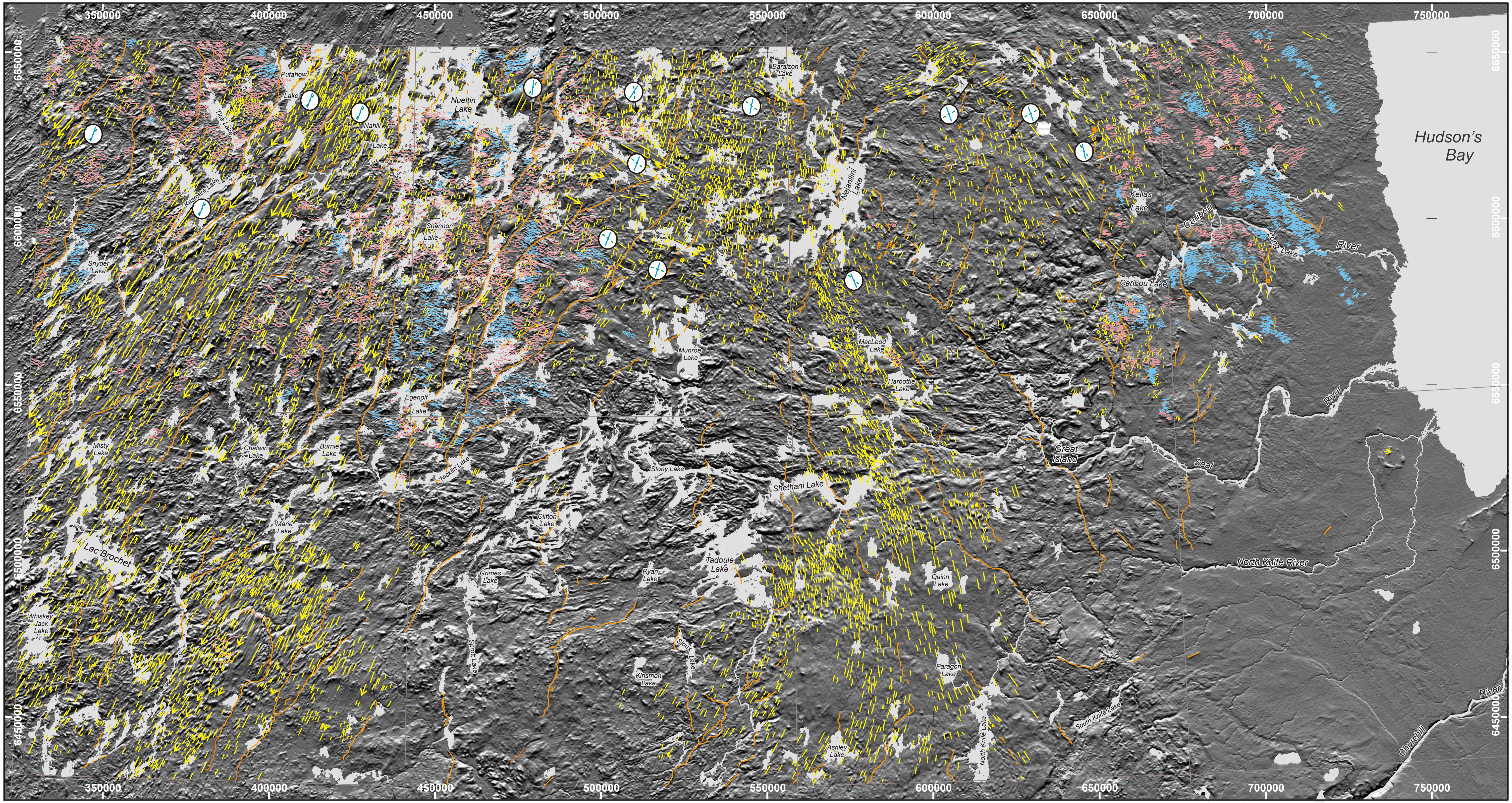
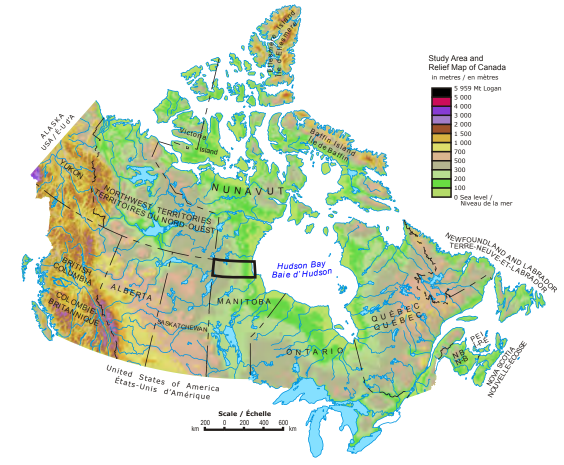
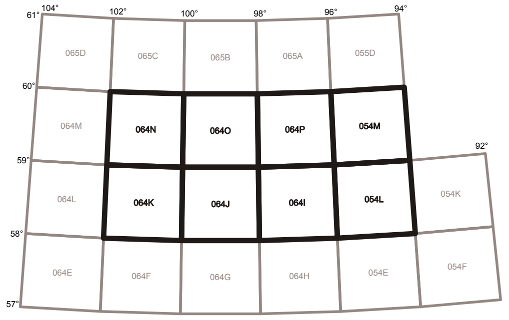


Figure 7


A
0 1 2
Kilometers
B
Ice flow
0 0.5 1 2
Kilometers
C
Ice flow
0 0.5 1 2
Kilometers
Figure 8







Index to the Canadian National Topographic System

Recommended Citation:
Trommelen, M. and Ross, M. (2010)
Subglacial landforms in northern Manitoba, Canada, based on remote sensing data
Journal of Maps.

Legend

Streamlined landform (drumlin, flute; direction unknown)
Streamlined landform (crag and tail, roche moutnee; direction known)
Pristine rogen moraine
Drumlinized rogen moraine
Esker (direction unknown)
Lake
River

Cross-cutting Relationships
younger (blue, solid)
older (green, dashed)

Subglacial landforms in northern Manitoba, Canada, based on remote sensing data

M. Trommelen and M. Ross
Department of Earth and Environmental Sciences
University of Waterloo, Waterloo, Ontario, Canada, N2L 3G1

Scale 1:1 125 000
0 10 20 30 40 50 Km

Universal Transverse Mercator Grid (NAD 1983, UTM Zone 14)

Glacial landforms were mapped at various scales from Landsat 7 Enhanced Thematic Mapper Plus (ETM+) satellite imagery, in combination with a Shuttle Radar Topography Mission (SRTM) digital elevation model (background) and SPOT imagery in the Great Island area. This map is part of a PhD project at the University of Waterloo which receives funding from the Geological Survey of Canada's (GSC) Research Affiliate Program, through the Geomapping for Energy and Minerals Program (GEM), Manitoba Far North project.

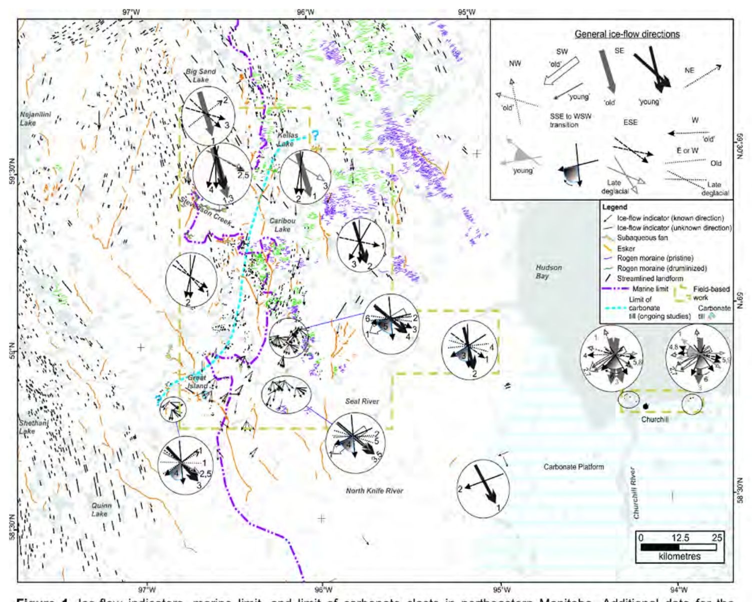


Figure 1. Ice-flow indicators, marine limit, and limit of carbonate clasts in northeastern Manitoba. Additional data for the study and field areas. Subglacial landforms (eskers, Rogen moraine, streamlined landforms, and subaqueous glaciofluvial fans) and field-based ice-flow indicators (striae, grooves, cresentic fractures, gouges, roche moutonnée) are shown. Large circles highlight the region-wide relationship between indicators discovered during field mapping and the relative timing of variously oriented ice-flow phases at each site. Data from the Churchill area is published in Trommelen and Ross (2011), and Trommelen (2011). The general ice-flow directions box provides a summary of ice-flow orientation for the entire region. The maximum marine inundation limit of the Tyrrell Sea (dashed purple) is around 180 m a.s.l. The northwestern limit of carbonate clast dispersal (dashed blue) from the Carbonate Platform to the east also extends across the area, compiled from detailed till pebble counts (Campbell et al., 2012).

REFERENCES

Boulton, G.S. and Clark, C.D., 1990. A highly mobile Laurentide ice sheet revealed by satellite images of glacial lineations; Nature, v. 346, p. 813–817.

Campbell, J.E., Trommelen, M.S., McCurdy, M.W., Böhm, C.O., and Ross, M., 2012. Till composition and ice-flow indicator data, Great Island–Caribou Lake area (parts of NTS 54L, 54M, 64-I and 64P), northeast Manitoba; Geological Survey of Canada, Open File 6967, 1 CD-ROM; Manitoba Innovation, Energy and Mines, Manitoba Geological Survey, Open File OF2011-4.

Dredge, L.A. and Nixon, F.M., 1981. Surficial geology, Nejanilini Lake, Manitoba; Geological Survey of Canada, Preliminary Map 7-1980, scale 1:250 000.

Dredge, L.A. and Nixon, F.M., 1982. Surficial geology, Caribou River, Manitoba; Geological Survey of Canada, Preliminary Map 5-1980, scale 1:250 000.

Dredge, L.A. and Nixon, F.M., 1992. Glacial and environmental geology of northeastern Manitoba; Geological Survey of Canada, Memoir 432, 80 p.

Dredge, L.A. and Thorleifson, L.H., 1987. The Middle Wisconsinan history of the Laurentide Ice Sheet; Geographie physique et Quaternaire, v. 41, no. 2, p. 215–235.

Dredge, L.A., Nixon, F.M., and Richardson, R.J.H., 1986. Quaternary geology and geomorphology of northwestern Manitoba; Geological Survey of Canada, Memoir 418, 38 p.

Dredge, L.A., Morgan, A.V., and Nielson, E., 1990. Sangamon and pre-Sangamon interglaciations in the Hudson Bay Lowlands of Manitoba; Geographie physique et Quaternaire, v. 44, no. 3, p. 319–336.

Kaszycki, C.A., Dredge, L.A., and Groom, H., 2008. Surficial geology and glacial history, Lynn Lake–Leaf Rapids area, Manitoba; Geological Survey of Canada, Open File 5873, 1 CD-ROM. doi:10.4049/225935

Klassen, R.W., 1986. Surficial geology of north-central Manitoba; Geological Survey of Canada, Memoir 419, 57 p.

Trommelen, M.S., 2011. Field-based ice-flow indicator data, Churchill area, northeastern Manitoba (part of NTS 54L16); Manitoba Innovation, Energy and Mines, Manitoba Geological Survey, Data Repository Item DRI2011001, Microsoft® Excel® file.

Trommelen, M.S. and Ross, M., 2009. Manitoba Far North Geomapping Initiative: field reconnaissance of surficial sediments, glacial landforms and ice-flow indicators, Great Island and Kellas Lake areas, Manitoba (NTS 54L, 64I, P); *in* Report of Activities 2009; Manitoba Innovation, Energy and Mines, Manitoba Geological Survey, p. 148–153.

Trommelen, M.S. and Ross, M., 2011. Far North Geomapping Initiative: palimpsest bedrock macroforms and other complex ice-flow indicators near Churchill, northern Manitoba (part of NTS 54L16); *in* Report of Activities 2011; Manitoba Innovation, Energy and Mines, Manitoba Geological Survey, p. 29–35.

Trommelen, M.S., Ross, M., and Campbell, J.E., 2010. Far North Geomapping Initiative: Quaternary geology of the Great Island–Kellas Lake area, northern Manitoba (parts of NTS 54L, M, 64I, P); *in* Report of Activities 2010; Manitoba Innovation, Energy and Mines, Manitoba Geological Survey, p. 36–49.

ACKNOWLEDGMENTS

This map was completed as part of a Ph.D. degree at the University of Waterloo, supervised by M. Ross, funding for which was provided by the Geological Survey of Canada's (GSC) Research Affiliate Program. Field logistics and helicopter support were provided by the Manitoba Geological Survey. This mapping exercise also formed part of the GSC Geomapping for Energy and Minerals Program (GEM) Northeast Manitoba project - Natural Resources Canada. Field assistance by B. Clarke, T. Stainton, and D. Vessey is greatly appreciated. The authors thank D. Kerr (GSC) and G. Matile (Manitoba Geological Survey) for critical review of the map.

Abstract

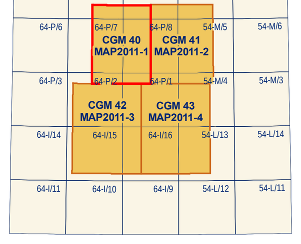
Northeast Manitoba is mantled by glacial and postglacial sediments, with scarce bedrock outcrops. Past ice-flow reconstructions in northern Manitoba suggest that the region has been covered at least twice by ice from the Keewatin Sector, and at least three times by ice from the Labradorean Sector (Dredge et al., 1986; Klassen, 1986; Dredge and Thorleifson, 1987; Boulton and Clark, 1990; Dredge et al., 1990; Dredge and Nixon, 1992; Kaszycki et al., 2008). This map builds on previous 1:250 000 surficial mapping completed in the 1980s (Dredge and Nixon, 1981, 1982).

The northern part of the study area is characterized by extensive swaths of bouldery drumlinized and pristine (nondrumlinized) Rogen moraine ridges alternating with swaths of streamlined terrain. The remaining area is characterized by bedrock topography draped by a mix of till blankets and till veneers. Long, large eskers are present throughout the area, at roughly 18 km intervals. Where the eskers are located below approximately 200 m a.s.l., they have been partially eroded by lacustrine and/or marine waters. Below 150 m a.s.l., the eskers exist as washed, low-lying sand and gravel blankets rather than ridges. A mix of organic blankets and marine sediment is present in the eastern portion of the study area, predominantly below 150 m a.s.l. The study area has, in part, been wave washed by either or both the postglacial Tyrrell Sea and glacial Lake Agassiz or other smaller, disconnected glacial lakes. The marine limit in the study area is around 180 m a.s.l. Field data were obtained by helicopter-assisted ground truthing in 2009 and 2010. Further description of map units, with photos, can be found in Trommelen and Ross (2009) and Trommelen et al. (2010). Field data, including relative age of ice-flow indicators, are available in Campbell et al. (2012).

Résumé

Le nord-est du Manitoba est recouvert d'une nappe de sédiments glaciaires et postglaciaires à travers de laquelle percent de rares affleurements rocheux. Dans le nord du Manitoba, les reconstitutions des anciens écoulements glaciaires donnent à penser que la région a été couverte au moins deux fois par des glaces du Secteur du Keewatin et au moins trois fois par des glaces du Secteur du Labrador (Dredge et al., 1986; Klassen, 1986; Dredge et Thorleifson, 1987; Boulton et Clark, 1990; Dredge et al., 1990; Dredge et Nixon, 1992; Kaszycki et al., 2008). La présente carte s'appuie sur des travaux de cartographie des matériaux superficiels à l'échelle 1/250 000 réalisés dans les années 1980 (Dredge et Nixon, 1981, 1982).

La partie nord de la région à l'étude est caractérisée par de grandes bandes de terrain occupées par des crêtes de moraine de Rogen à blocs, modelées en drumlins ou conservant leur forme d'origine, qui alternent avec des bandes de terrain aux formes fuselées. Le reste de la région est caractérisé par un relief défini par la surface du socle rocheux et la couverture moulante de nappes et de plaques de till. De longs et larges eskers sont présents dans toute la région, à des intervalles d'environ 18 km. Lorsque les eskers sont situés à une altitude inférieure à environ 200 m ASL, ils ont été en partie érodés par l'action d'eaux lacustres ou marines. Au-dessous d'une altitude de 150 m ASL, les eskers ne subsistent que sous la forme de nappes de sable et gravier délavés à faible relief, plutôt que de crêtes. Un mélange de nappes de dépôts organiques et de sédiments marins sont présents dans la partie est de la zone à l'étude, surtout à des altitudes inférieures à 150 m ASL. La région à l'étude a été délavée par l'action de vagues de la Mer de Tyrrell postglaciaire, du Lac glaciaire Agassiz ou d'autres d'autres petits lacs glaciaires discontinus. Dans la région à l'étude, la limite marine est située à une altitude d'environ 180 m ASL. Les données de terrain ont été obtenues lors de vérifications des réalités de terrain menées à l'aide d'un hélicoptère en 2009 et 2010. Des descriptions plus détaillées des unités cartographiques, avec photos, peuvent être consultées dans Trommelen et Ross (2009) et Trommelen et al. (2010). Les données de terrain, dont celles concernant les âges relatifs des indicateurs d'écoulement glaciaire, sont fournies dans Campbell et al. (2012).



National Topographic System reference and index to adjoining published maps

Cover illustration:
Bouldery till plain, somewhat washed by proglacial meltwater. Photograph by M.S. Trommelen, Manitoba Geological Survey.

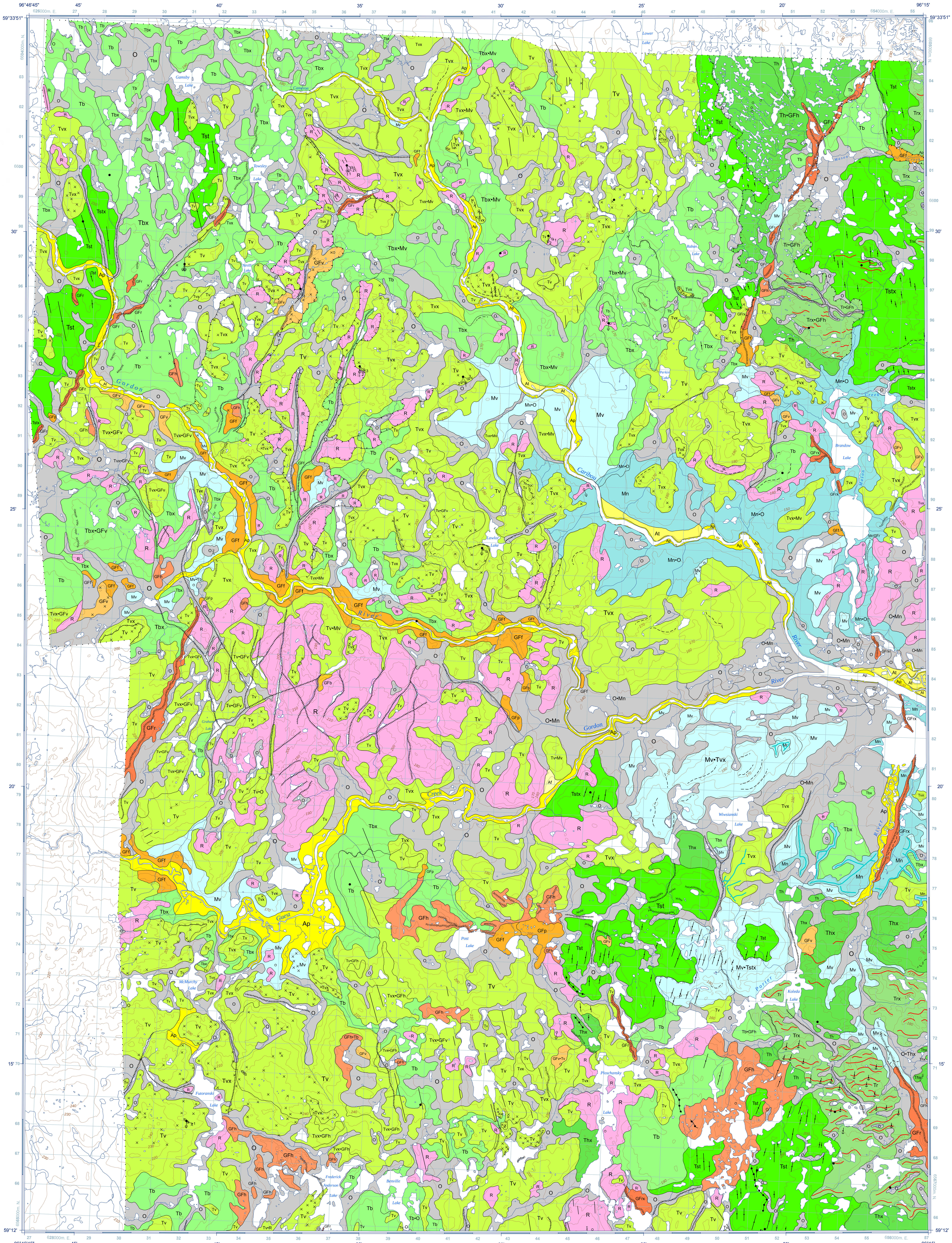
Printed map:
Catalogue No. M183-1/40-2011E
ISBN 978-0-660-20097-2

Digital map:
Catalogue No. M183-1/40-2011E-PDF
ISBN 978-1-100-19081-5

doi:10.4095/288956

Natural Resources Canada   Ressources naturelles Canada

GEOLOGICAL SURVEY OF CANADA
CANADIAN GEOSCIENCE MAP 40
MANITOBA GEOLOGICAL SURVEY
GEOSCIENTIFIC MAP MAP2011-1
SURFICIAL GEOLOGY
GORDON RIVER
Manitoba
1:50 000



This legend is common to CGM 40, CGM 41, CGM 42, and CGM 43.
Coloured legend blocks indicate map units that appear on this map.
Not all symbols shown in the legend appear on this map.

QUATERNARY

SURFICIAL DEPOSITS

HOLOCENE

NONGLACIAL ENVIRONMENTS

O — **ORGANIC DEPOSITS:** Undifferentiated peat and muck; 1 m to greater than 5 m thick; formed by the accumulation of plant material in various stages of decomposition; generally occur as flat, wet terrain (swamps and bogs) over poorly drained substrates. Fibric fens are present along some water channels. Thickness varies from thin organic veneers (20–40 cm) overlying till and boulder fields to organic plains greater than 3 m thick. Thick organic deposits typically overlie fine-grained glaciolacustrine and marine sediments. Permafrost is commonly present underlying and/or within thick organic deposits, as seen by the prevalent raised bogs with ice-wedge polygons. Small, unmapped deposits commonly occur in most terrain units. Peat mantles most geological units.

ALLUVIAL DEPOSITS: Sorted sand, silt, and clay with minor gravel and organic detritus; commonly stratified; deposited along and/or within all modern rivers and streams.

Ap — **Floodplain sediments:** sorted sand, silt, clay, minor gravel, and organic detritus greater than 1 m thick; forming active floodplains close to river and stream level; includes terraces too small to show at this map scale.

At — **Fluvial terraces:** inactive terraces above modern floodplain; greater than 2 m thick; consisting of gravel, sand, and overbank silt and organic detritus on the Seal and Caribou rivers. Annual spring ice-push continues to build up sediment along the side of these terraces.

Af — **Fan-delta sediments:** poorly sorted sand and organic detritus deposited at the western side of Caribou Lake.

L — **LACUSTRINE DEPOSITS:** Undifferentiated; massive to stratified, sorted sand, silt, clay, and minor organic detritus deposited adjacent and/or within modern ponds and lakes.

MARINE SEDIMENTS: Poor to well sorted sand and silt with 0–20% pebbles, cobbles, and occasional boulders (ice rafted and lags), deposited in the postglacial Tyrrell Sea. Clasts are typically subrounded to subangular, occasionally striated and/or faceted and/or bullet-shaped, derived from the reworking of till. The marine limit is between 165 m a.s.l. and 180 m a.s.l., defined by washing limits on eskers and till plains and by the elevations of sand blankets and beaches. The exact elevation is uncertain, owing to the likelihood that glacial Lake Agassiz was coeval to the Tyrrell Sea during deglaciation. Near the marine limit, glaciomarine sediment also occurs. These sand and silt deposits locally include pockets of debris-flow sediments, till, and/or minor dropstones, deposited from suspension and iceberg rafting.

Mv — **Marine veneer:** discontinuous sand less than 1–2 m thick that drape the existing topography; overlies wave-washed till between 180 m a.s.l. and 140 m a.s.l.; below 140 m a.s.l. present as sandy patches overlying bedrock outcrops where all till has been removed.

Mb — **Marine blanket:** flat to gently undulating plain of fine sand, silt, and clay greater than 2 m thick; often overlain by a layer of organic material (less than 1 m thick); sparsely fossiliferous; offshore sediment.

Mn — **Nearshore sediments:** poor to well sorted, sand, silt, and clay; occur as veneers and blankets of sediment overlying till and/or bedrock; commonly less than 2 m thick. Predominantly derived from reworking of till and/or glaciofluvial deposits.

Mr — **Littoral sediments:** poor to well sorted, stratified sand with 5–20% pebbles and cobbles; typically 1–2 m thick. Beach ridges, consisting of sand and cobbles derived from the underlying till are present at elevations of 155–170 m a.s.l. More common are linear patches of pebbly sand with occasional spits, derived from esker and crevasse ridges. The latter typically contain a higher percentage of exotic lithologies. Where esker and crevasse ridges occur below marine limit, wave washing has commonly reduced the ridges down to a common height of 0.25–1 m and redistributed the sand creating veneers and blankets of light orange, granitic pebbly sand. Low-lying regions or depressions often have an organic veneer overlying the sand and silt.

LATE WISCONSINAN

PROGLACIAL AND GLACIAL ENVIRONMENTS

GLACIOLACUSTRINE DEPOSITS: moderate to well sorted clay, silt, and very fine to fine sand; massive to bedded; moderately dense; deposited in glacial Lake Agassiz or other small glacial lakes along the margin of the retreating Laurentide Ice Sheet. Usually overlain by less than 0.5 m thick organic deposits in lowlands with flat topography. Some littoral sand may be marine in origin, given that the Tyrrell Sea incursion occurred in the same area, and the genesis is uncertain. Sand encountered above 180 m has been assigned as glaciolacustrine, whereas sand below 180 m is considered marine. Sediment is derived from the Archean and Paleoproterozoic rocks in the area, and predominately consists of feldspar and quartz. In the west part of Great Island, carbonate, red mudstone, and black mudstone clasts were found in calcareous silty clay between 200–260 m a.s.l. This material is quite similar to calcareous glaciolacustrine pelite found to the west of the map area (Dredge et al., 1986), and was likely derived from Hudsonian and/or Labradorean drift deposited into glacial Lake Agassiz.

GLd — **Ice-contact deltaic sediments:** well to moderately stratified sand and gravel, forming a deltaic deposit where a meltwater channel entered a glacial lake during regression and lowering of lake levels. Surface is kettled and the landform has a steep front.

GLv — **Glaciolacustrine veneer:** discontinuous cover less than 1–2 m thick; underlying topography is discernible. Interspersed with small till or glaciofluvial deposits.

GLb — **Glaciolacustrine blanket:** continuous cover greater than 2 m thick; forming flat to undulating topography that locally obscures underlying geomorphology.

GLACIOFLUVIAL DEPOSITS: light orange, pebbly sand with occasional (2%) cobbles and boulders at surface deposited behind, at, or in front of the ice margin by flowing glacial meltwater. The sand is often well sorted and massive, though occasional bedding is present in some esker ridges. Where the suffix "x" has been added to the terrain unit label (i.e. GFrx) it indicates the sediments have had significant surface reworking by glacial Lake Agassiz and/or the Tyrrell Sea.

GFv — **Glaciofluvial veneer:** discontinuous sand and gravel cover, less than 1–2 m thick; underlying topography is discernible.

GFb — **Glaciofluvial blanket:** continuous sand and gravel cover greater than 2 m thick, forming flat to undulating topography that locally obscures underlying units and associated geomorphic patterns. Occasional thinner patches of sediment may occur. Unit GFbx indicates significant surface reworking by glacial Lake Agassiz and/or the Tyrrell Sea.

GFp — **Subaerial outwash sediments:** massive to stratified sand to pebbly sand with occasional (~5%) cobbles and boulders, deposited in a subaerial environment at or in front of the ice margin by glacial meltwater. Sediments are greater than 2 m thick and may drape the underlying topography like a blanket, or where thicker, mask underlying topography completely. The surface may be kettled; unit includes fan sediments deposited at the ice margin at the portal of an englacial or subglacial meltwater channel.

GFt — **Terraced sediments:** inactive terraces above modern floodplain; deposited during glacial meltwater flow in meltwater channels. The terrace along the Seal River contains about 10–20% carbonate clasts, in addition to the local shield-derived lithologies.

GFf — **Subaqueous outwash sediments:** massive to stratified sand to pebbly sand, occasionally rippled and/or crossbedded; interbedded with gravel and diamictic units of variable thickness; rare (~5%) cobbles and boulders present; sediments deposited into a shallow, subaqueous glaciolacustrine or marine environment (Tyrrell Sea), at or near the retreating ice front by meltwater turbidity currents.

GFh — **Ice-contact glaciofluvial sediments:** undifferentiated deposits; poorly sorted sand and gravel with minor diamicton; deposited by glacial meltwater in direct contact with the glacier; 1 m to greater than 20 m thick; forming gently undulating to hummocky topography related to melting of underlying ice. Features include kettles, kames, and ridges.

GFr — **Eskers and esker systems:** stratified sand and gravel with minor diamicton, deposited by meltwater flow within tunnels beneath or within the glacier; present as large (3–10 m high), long (10–25 km), regularly spaced (10–18 km) esker segments, with smaller (1–5 m high) and shorter esker ridges found between the large ridges. Esker segments consist of kame and kettle topography up to 20 m high. Eskers and crevasse-fill ridges well below marine limit have been extensively wave washed which has created resultant 'ridges' 0.25–2 m high, and a blanket of pebbly sand near the 'ridge' location, and are mapped as unit Mn.

GLACIAL DEPOSITS: unsorted to poorly sorted diamicton (till) with a sandy-silt to silty-sand matrix, deposited in subglacial or ice-marginal environments. May locally contain blocks of pre-existing sediments and/or stratified drift. Tills consist mainly of granitic material in regions overlying granitic bedrock, and consist of a more variable lithology in supracrustal bedrock regions. The till has been emplaced by ice flowing from the Keewatin Sector, within the Laurentide Ice Sheet. Where the suffix "x" has been added to the terrain unit label (e.g. Tvx, Thx, Tbx, Tstx) it indicates that the sediments have had significant surface reworking by meltwater and/or the Tyrrell Sea.

Tv — **Till veneer:** discontinuous till cover less than 1–2 m thick; underlying topography is discernible. Surface may be washed in the vicinity of meltwater channels and where marine sediments are present.

Tb — **Till blanket:** continuous till cover greater than 2 m thick, forming flat to undulating topography that locally obscures underlying units and associated geomorphic patterns. Occasional thinner patches of till may occur. Surface may be washed in the vicinity of meltwater channels and where marine sediments are present.

Tst — **Streamlined till:** till greater than 2 m thick, moulded beneath the glacier into linear ridges and/or furrows parallel to ice flow; drumlins, drumlinoid ridges, and flutings. Ridges are typically 0.1–3 km long and only 1–3 m high.

Th — **Hummocky till:** supraglacial meltout (ablation) tills deposited by melting of stagnant ice; loose, texturally variable, sandy to gravelly matrix, some sorting; angular to subangular clasts; locally includes poorly sorted sand and gravel; gently undulating to hummocky topography.

Tr — **Rogen moraine:** anastamosing to curved ridges and intervening troughs, all lying transverse to former ice-flow direction. The Rogen ridges may exhibit gradual up- and down-ice flow-direction transition to drumlinoid ridges and flutings and/or a nontransitional lateral shift to streamlined terrain. Ridges are typically 0.1–3.0 km long, with a typical segment length of 760 m (n=507). There are both 'pristine' ridges and 'drumlinized' ridges, the latter of which have been overridden by actively flowing ice, resulting in streamlining of their surfaces (*see* attribute table for delineation). The degree and size of streamlining is often transitional between minor modification to complete drumlinization.

hu — Weakly calcareous, carbonate-bearing till with a clayey-silt matrix encountered within the subsurface in the southwest portion of Great Island. This till was likely deposited by west- or northwest-flowing ice from the Labradorean Sector (Hudsonian Ice) of the Laurentide Ice Sheet.

PRE-QUATERNARY

BEDROCK

R — **Precambrian rocks:** metasedimentary, metavolcanic rocks and associated intrusive rocks; may be overlain by a thin, discontinuous veneer of till in upland, unwashed areas, and/or a thin discontinuous veneer of sand and/or pebbly sand below marine limit or within meltwater corridors that rarely exceeds 1 m thick.

NOTE: In areas where the surficial cover forms a complex pattern, the area is coloured according to the dominant unit and labelled in descending order of cover (e.g. O•Tr). Where underlying stratigraphic units are known, areas are coloured according to the overlying unit and labelled in the following manner: O/Tr.

Geological contact (defined, inferred)
Rogen moraine crest: *see* attribute table (Leg_Label field) for delineation between pristine and drumlinized ridges
Limit of mapping
Esker, direction known
Major moraine ridge
Crevasse ridge
Drumlinoid ridge or fluting
Drumlin
Streamlined bedrock, direction unknown
Minor meltwater channel, direction unknown
Major meltwater corridor
Raised beach, wave-cut notch
Limit of submergence, marine and/or glaciomarine (wave-cut benches, washing limits)
Limit of submergence, glaciolacustrine (wave-cut benches, washing limits)
Small outcrop
Field site with sample
Field site without sample
Striation, direction known, numbers indicate relative age (1 - oldest) (*see* attribute table (Rel_age field) for relative ages)
Striation, direction unknown
Roche moutonnée
Kame
Kettle

Recommended citation:
Trommelen, M.S. and Campbell, J.E., 2012. Surficial geology, Gordon River, Manitoba; Geological Survey of Canada, Canadian Geoscience Map 40; Manitoba Innovation, Energy and Mines, Manitoba Geological Survey, Geoscientific Map MAP2011-1, scale 1:50 000. doi:10.4095/288956

GSC CANADIAN GEOSCIENCE MAP 40 • MGS GEOSCIENTIFIC MAP MAP2011-1

Canadian Geoscience Maps
Cartes géoscientifiques du Canada

Authors: M.S. Trommelen[1] and J.E. Campbell[2]
[1]Manitoba Geological Survey, 360-1395 Ellice Avenue, Winnipeg, Manitoba R3G 3P2
[2]Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario K1A 0E8

Aerial photograph interpretation (1:50 000 scale) and geology by M.S. Trommelen (2009 and 2010 field seasons)

Digital compilation by L. Robertson, 2010–2011

Cartography by D. Viner

Scientific editing by E. Inglis

Map projection Univeral Transverse Mercator, zone 14. North American Datum 1983

Base map at the scale of 1:50 000 from Natural Resources Canada, with modifications.

SURFICIAL GEOLOGY
GORDON RIVER
Manitoba
1:50 000

1 0 1 2 3 4 km

Elevations in metres above mean sea level.

Magnetic declination 2012, 0°37' E decreasing 9.0' annually.

The Geological Survey of Canada and Manitoba Geological Survey welcome corrections or additional information from users.
This map was produced from processes that conform to the Scientific and Technical Publishing Services Quality Management System, registered to the ISO 9001:2008 standard.

This publication, including digital data, can be downloaded free of charge from GeoPub (http://geopub.nrcan.gc.ca/). It is also available from the Geological Survey of Canada Bookstore (http://gsc.nrcan.gc.ca/bookstore).

This publication can also be downloaded in PDF, free of charge, from the Manitoba government web site at http://manitoba.ca/minerals.

GEOLOGICAL SURVEY OF CANADA
CANADIAN GEOSCIENCE MAP 40
MANITOBA GEOLOGICAL SURVEY
GEOSCIENTIFIC MAP MAP2011-1
SURFICIAL GEOLOGY
GORDON RIVER
Manitoba

Canada   Manitoba

Four trim marks around perimeter of map sheet. Trim map sheet first, then fold at folding marks.

Cover and additional panels are 17cm wide when folded.

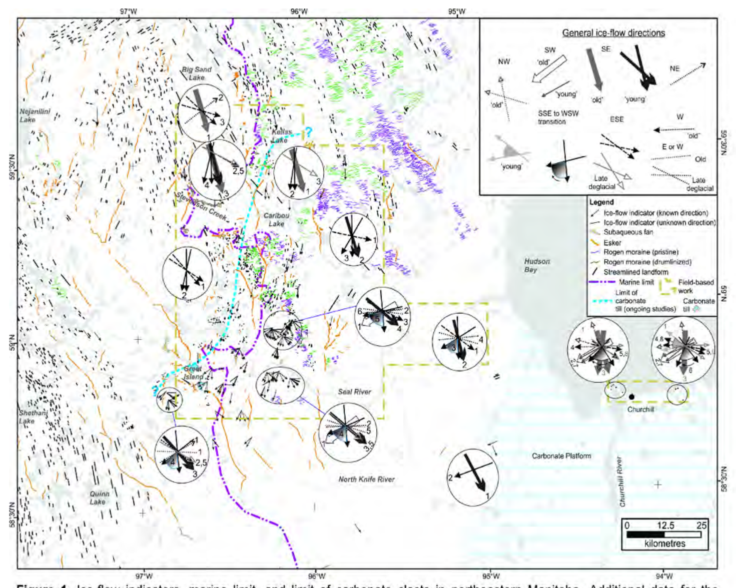


Figure 1. Ice-flow indicators, marine limit, and limit of carbonate clasts in northeastern Manitoba. Additional data for the study and field areas. Subglacial landforms (eskers, Rogen moraine, streamlined landforms, and subaqueous glaciofluvial fans) and field-based ice-flow indicators (striae, grooves, cresentic fractures, gouges, roche moutonnée) are shown. Large circles highlight the region-wide relationship between indicators discovered during field mapping and the relative timing of variously oriented ice-flow phases at each site. Data from the Churchill area is published in Trommelen and Ross (2011), and Trommelen (2011). The general ice-flow directions box provides a summary of ice-flow orientation for the entire region. The maximum marine inundation limit of the Tyrrell Sea (dashed purple) is around 180 m a.s.l. The northwestern limit of carbonate clast dispersal (dashed blue) from the Carbonate Platform to the east also extends across the area, compiled from detailed till pebble counts (Campbell et al., 2012).

REFERENCES

Boulton, G.S. and Clark, C.D., 1990. A highly mobile Laurentide ice sheet revealed by satellite images of glacial lineations; Nature, v. 346, p. 813–817.

Campbell, J.E., Trommelen, M.S., McCurdy, M.W., Böhm, C.O., and Ross, M., 2012. Till composition and ice-flow indicator data, Great Island–Caribou Lake area (parts of NTS 54L, 54M, 64-I and 64P), northeast Manitoba; Geological Survey of Canada, Open File 6967, 1 CD-ROM; Manitoba Innovation, Energy and Mines, Manitoba Geological Survey, Open File OF2011-4.

Dredge, L.A. and Nixon, F.M., 1981. Surficial geology, Nejanilini Lake, Manitoba; Geological Survey of Canada, Preliminary Map 7-1980, scale 1:250 000.

Dredge, L.A. and Nixon, F.M., 1982. Surficial geology, Caribou River, Manitoba; Geological Survey of Canada, Preliminary Map 5-1980, scale 1:250 000.

Dredge, L.A. and Nixon, F.M., 1992. Glacial and environmental geology of northeastern Manitoba; Geological Survey of Canada, Memoir 432, 80 p.

Dredge, L.A. and Thorleifson, L.H., 1987. The Middle Wisconsinan history of the Laurentide Ice Sheet; Geographie physique et Quaternaire, v. 41, no. 2, p. 215–235.

Dredge, L.A., Nixon, F.M., and Richardson, R.J.H., 1986. Quaternary geology and geomorphology of northwestern Manitoba; Geological Survey of Canada, Memoir 418, 38 p.

Dredge, L.A., Morgan, A.V., and Nielson, E., 1990. Sangamon and pre-Sangamon interglaciations in the Hudson Bay Lowlands of Manitoba; Geographie physique et Quaternaire, v. 44, no. 3, p. 319–336.

Kaszycki, C.A., Dredge, L.A., and Groom, H., 2008. Surficial geology and glacial history, Lynn Lake–Leaf Rapids area, Manitoba; Geological Survey of Canada, Open File 5873, 1 CD-ROM. doi:10.4049/225935

Klassen, R.W., 1986. Surficial geology of north-central Manitoba; Geological Survey of Canada, Memoir 419, 57 p.

Trommelen, M.S., 2011. Field-based ice-flow indicator data, Churchill area, northeastern Manitoba (part of NTS 54L16); Manitoba Innovation, Energy and Mines, Manitoba Geological Survey, Data Repository Item DRI2011001, Microsoft® Excel® file.

Trommelen, M.S. and Ross, M., 2009. Manitoba Far North Geomapping Initiative: field reconnaissance of surficial sediments, glacial landforms and ice-flow indicators, Great Island and Kellas Lake areas, Manitoba (NTS 54L, 64I, P); in Report of Activities 2009; Manitoba Innovation, Energy and Mines, Manitoba Geological Survey, p. 148–153.

Trommelen, M.S. and Ross, M., 2011. Far North Geomapping Initiative: palimpsest bedrock macroforms and other complex ice-flow indicators near Churchill, northern Manitoba (part of NTS 54L16); in Report of Activities 2011; Manitoba Innovation, Energy and Mines, Manitoba Geological Survey, p. 29–35.

Trommelen, M.S., Ross, M., and Campbell, J.E., 2010. Far North Geomapping Initiative: Quaternary geology of the Great Island–Kellas Lake area, northern Manitoba (parts of NTS 54L, M, 64I, P); in Report of Activities 2010; Manitoba Innovation, Energy and Mines, Manitoba Geological Survey, p. 36–49.

ACKNOWLEDGMENTS

This map was completed as part of a Ph.D. degree at the University of Waterloo, supervised by M. Ross, funding for which was provided by the Geological Survey of Canada's (GSC) Research Affiliate Program. Field logistics and helicopter support were provided by the Manitoba Geological Survey. This mapping exercise also formed part of the GSC Geomapping for Energy and Minerals Program (GEM) Northeast Manitoba project - Natural Resources Canada. Field assistance by B. Clarke, T. Stainton, and D. Vessey is greatly appreciated. The authors thank D. Kerr (GSC) and G. Matile (Manitoba Geological Survey) for critical review of the map.

Abstract

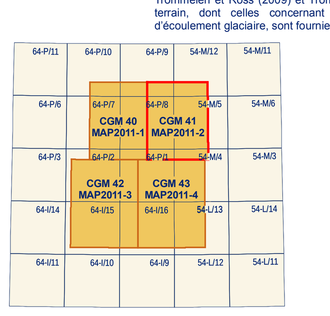
Northeast Manitoba is mantled by glacial and postglacial sediments, with scarce bedrock outcrops. Past ice-flow reconstructions in northern Manitoba suggest that the region has been covered at least twice by ice from the Keewatin Sector, and at least three times by ice from the Labradorean Sector (Dredge et al., 1986; Klassen, 1986; Dredge and Thorleifson, 1987; Boulton and Clark, 1990; Dredge et al., 1990; Dredge and Nixon, 1992; Kaszycki et al., 2008). This map builds on previous 1:250 000 surficial mapping completed in the 1980s (Dredge and Nixon, 1981, 1982).

The northern part of the study area is characterized by extensive swaths of bouldery drumlinized and pristine (nondrumlinized) Rogen moraine ridges alternating with swaths of streamlined terrain. The remaining area is characterized by bedrock topography draped by a mix of till blankets and till veneers. Long, large eskers are present throughout the area, at roughly 18 km intervals. Where the eskers are located below approximately 200 m a.s.l., they have been partially eroded by lacustrine and/or marine waters. Below 150 m a.s.l., the eskers exist as washed, low-lying sand and gravel blankets rather than ridges. A mix of organic blankets and marine sediment is present in the eastern portion of the study area, predominantly below 150 m a.s.l. The study area has, in part, been wave washed by either or both the postglacial Tyrrell Sea and glacial Lake Agassiz or other smaller, disconnected glacial lakes. The marine limit in the study area is around 180 m a.s.l. Field data were obtained by helicopter-assisted ground truthing in 2009 and 2010. Further description of map units, with photos, can be found in Trommelen and Ross (2009) and Trommelen et al., 2010. Field data, including relative age of ice-flow indicators, are available in Campbell et al. (2012).

Résumé

Le nord-est du Manitoba est recouvert d'une nappe de sédiments glaciaires et postglaciaires à travers de laquelle percent de rares affleurements rocheux. Dans le nord du Manitoba, les reconstitutions des anciens écoulements glaciaires donnent à penser que la région a été couverte au moins deux fois par des glaces du Secteur du Keewatin et au moins trois fois par des glaces du Secteur du Labrador (Dredge et al., 1986; Klassen, 1986; Dredge et Thorleifson, 1987; Boulton et Clark, 1990; Dredge et al., 1990; Dredge et Nixon, 1992; Kaszycki et al., 2008). La présente carte s'appuie sur des travaux de cartographie des matériaux superficiels à l'échelle 1/250 000 réalisés dans les années 1980 (Dredge et Nixon, 1981, 1982).

La partie nord de la région à l'étude est caractérisée par de grandes bandes de terrain occupées par des crêtes de moraine de Rogen à blocs, modelées en drumlins ou conservant leur forme d'origine, qui alternent avec des bandes de terrain aux formes fuselées. Le reste de la région est caractérisé par un relief défini par la surface du socle rocheux et la couverture moulante de nappes et de plaques de till. De longs et larges eskers sont présents dans toute la région, à des intervalles d'environ 18 km. Lorsque les eskers sont situés à une altitude inférieure à environ 200 m ASL, ils ont été en partie érodés par l'action d'eaux lacustres ou marines. Au-dessous d'une altitude de 150 m ASL, les eskers ne subsistent que sous la forme de nappes de sable et gravier délavés à faible relief, plutôt que de crêtes. Un mélange de nappes de dépôts organiques et de sédiments marins sont présents dans la partie est de la zone à l'étude, surtout à des altitudes inférieures à 150 m ASL. La région à l'étude a été délavée par l'action de vagues de la Mer de Tyrrell postglaciaire, du Lac glaciaire Agassiz ou d'autres d'autres petits lacs glaciaires discontinus. Dans la région à l'étude, la limite marine est située à une altitude d'environ 180 m ASL. Les données de terrain ont été obtenues lors de vérifications des réalités de terrain menées à l'aide d'un hélicoptère en 2009 et 2010. Des descriptions plus détaillées des unités cartographiques, avec photos, peuvent être consultées dans Trommelen et Ross (2009) et Trommelen et al., 2010. Les données de terrain, dont celles concernant les âges relatifs des indicateurs d'écoulement glaciaire, sont fournies dans Campbell et al. (2012).



National Topographic System reference and index to adjoining published maps

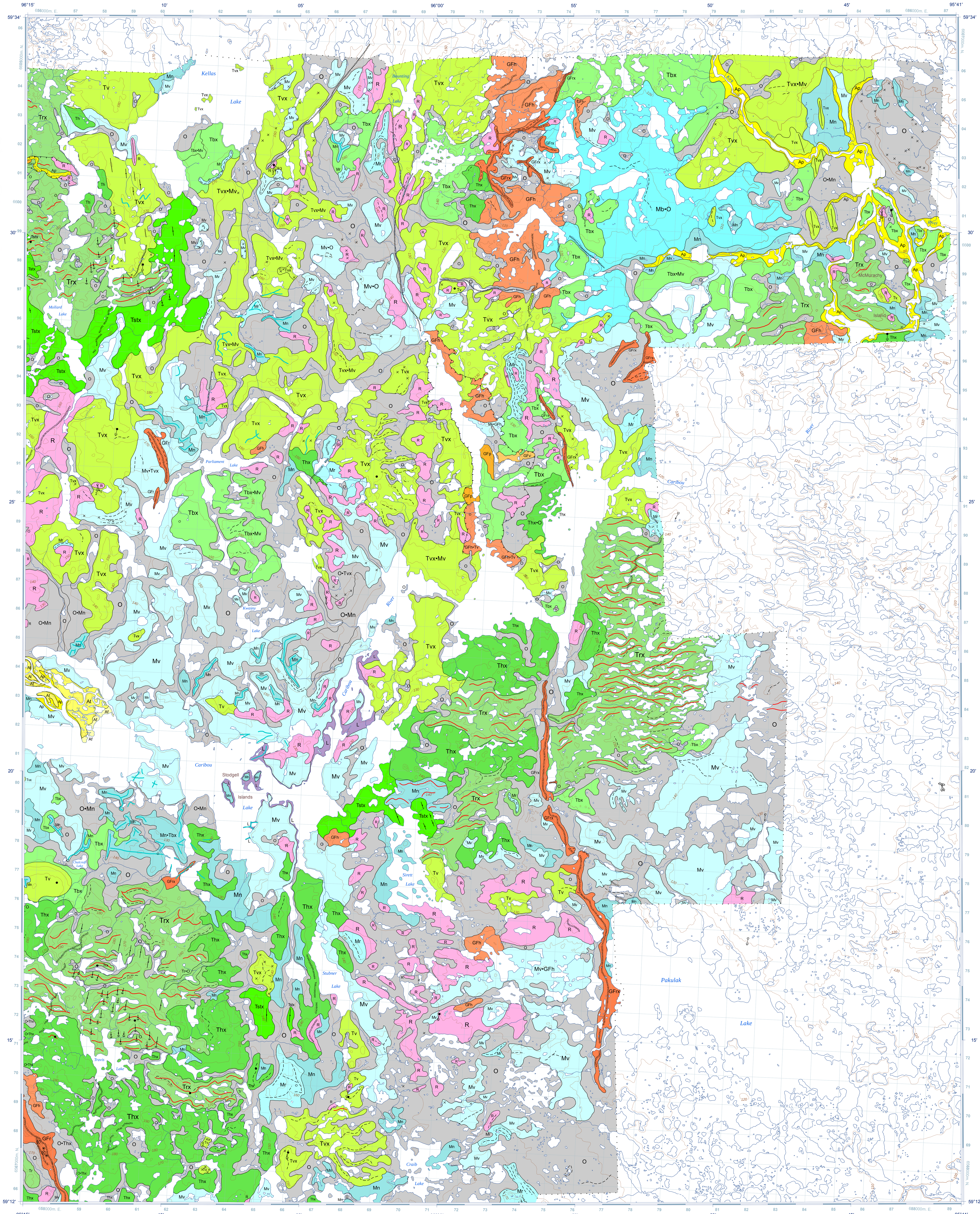
Cover illustration:
Hummocky glaciofluvial deposits east of Kellas Lake. Photograph by M.S. Trommelen, Manitoba Geological Survey.

Printed map:
Catalogue No. M183-1/41-2011E
ISBN 978-0-660-20099-6
Digital map:
Catalogue No. M183-1/41-2011E-PDF
ISBN 978-1-100-19082-2

doi:10.4095/288957

Natural Resources Canada   Ressources naturelles Canada

GEOLOGICAL SURVEY OF CANADA
CANADIAN GEOSCIENCE MAP 41
MANITOBA GEOLOGICAL SURVEY
GEOSCIENTIFIC MAP MAP2011-2
SURFICIAL GEOLOGY
STUBNER LAKE
Manitoba
1:50 000



This legend is common to CGM 40, CGM 41, CGM 42, and CGM 43.
Coloured legend blocks indicate map units that appear on this map.
Not all symbols shown in the legend appear on this map.

QUATERNARY

SURFICIAL DEPOSITS

HOLOCENE

NONGLACIAL ENVIRONMENTS

O — **ORGANIC DEPOSITS:** Undifferentiated peat and muck; 1 m to greater than 5 m thick; formed by the accumulation of plant material in various stages of decomposition; generally occur as flat, wet terrain (swamps and bogs) over poorly drained substrates. Fibric fens are present along some water channels. Thickness varies from thin organic veneers (20–40 cm) overlying till and boulder fields to organic plains greater than 3 m thick. Thick organic deposits typically overlie fine-grained glaciolacustrine and marine sediments. Permafrost is commonly present underlying and/or within thick organic deposits, as seen by the prevalent raised bogs with ice-wedge polygons. Small, unmapped deposits commonly occur in most terrain units. Peat mantles most geological units.

ALLUVIAL DEPOSITS: Sorted sand, silt, and clay with minor gravel and organic detritus; commonly stratified; deposited along and/or within all modern rivers and streams.

Ap — **Floodplain sediments:** sorted sand, silt, clay, minor gravel, and organic detritus greater than 1 m thick; forming active floodplains close to river and stream level; includes terraces too small to show at this map scale.

At — **Fluvial terraces:** inactive terraces above modern floodplain; greater than 2 m thick; consisting of gravel, sand, and overbank silt and organic detritus on the Seal and Caribou rivers. Annual spring ice-push continues to build up sediment along the side of these terraces.

Af — **Fan-delta sediments:** poorly sorted sand and organic detritus deposited at the western side of Caribou Lake.

L — **LACUSTRINE DEPOSITS:** Undifferentiated; massive to stratified, sorted sand, silt, clay, and minor organic detritus deposited adjacent and/or within modern ponds and lakes.

MARINE SEDIMENTS: Poor to well sorted sand and silt with 0–20% pebbles, cobbles, and occasional boulders (ice rafted and lags), deposited in the postglacial Tyrrell Sea. Clasts are typically subrounded to subangular, occasionally striated and/or faceted and/or bullet-shaped, derived from the reworking of till. The marine limit is between 165 m a.s.l. and 180 m a.s.l., defined by washing limits on eskers and till plains and by the elevations of sand blankets and beaches. The exact elevation is uncertain, owing to the likelihood that glacial Lake Agassiz was coeval to the Tyrrell Sea during deglaciation. Near the marine limit, glaciomarine sediment also occurs. These sand and silt deposits locally include pockets of debris-flow sediments, till, and/or minor dropstones, deposited from suspension and iceberg rafting.

Mv — **Marine veneer:** discontinuous sand less than 1–2 m thick that drape the existing topography; overlies wave-washed till between 180 m a.s.l. and 140 m a.s.l.; below 140 m a.s.l. present as sandy patches overlying bedrock outcrops where all till has been removed.

Mb — **Marine blanket:** flat to gently undulating plain of fine sand, silt, and clay greater than 2 m thick; often overlain by a layer of organic material (less than 1 m thick); sparsely fossiliferous; offshore sediment.

Mn — **Nearshore sediments:** poor to well sorted, sand, silt, and clay; occur as veneers and blankets of sediment overlying till and/or bedrock; commonly less than 2 m thick. Predominantly derived from reworking of till and/or glaciofluvial deposits.

Mr — **Littoral sediments:** poor to well sorted, stratified sand with 5–20% pebbles and cobbles; typically 1–2 m thick. Beach ridges, consisting of sand and cobbles derived from the underlying till are present at elevations of 155–170 m a.s.l. More common are linear patches of pebbly sand with occasional spits, derived from esker and crevasse ridges. The latter typically contain a higher percentage of exotic lithologies. Where esker and crevasse ridges occur below marine limit, wave washing has commonly reduced the ridges down to a common height of 0.25–1 m and redistributed the sand creating veneers and blankets of light orange, granitic pebbly sand. Low-lying regions or depressions often have an organic veneer overlying the sand and silt.

LATE WISCONSINAN

PROGLACIAL AND GLACIAL ENVIRONMENTS

GLACIOLACUSTRINE DEPOSITS: moderate to well sorted clay, silt, and very fine to fine sand; massive to bedded; moderately dense; deposited in glacial Lake Agassiz or other small glacial lakes along the margin of the retreating Laurentide Ice Sheet. Usually overlain by less than 0.5 m thick organic deposits in lowlands with flat topography. Some littoral sand may be marine in origin, given that the Tyrrell Sea incursion occurred in the same area, and the genesis is uncertain. Sand encountered above180 m has been assigned as glaciolacustrine, whereas sand below 180 m is considered marine. Sediment is derived from the Archean and Paleoproterozoic rocks in the area, and predominately consists of feldspar and quartz. In the west part of Great Island, carbonate, red mudstone, and black mudstone clasts were found in calcareous silty clay between 200–260 m a.s.l. This material is quite similar to calcareous glacio-lacustrine pelite found to the west of the map area (Dredge et al., 1986), and was likely derived from Hudsonian and/or Labradorean drift deposited into glacial Lake Agassiz.

GLd — **Ice-contact deltaic sediments:** well to moderately stratified sand and gravel, forming a deltaic deposit where a meltwater channel entered a glacial lake during regression and lowering of lake levels. Surface is kettled and the landform has a steep front.

GLv — **Glaciolacustrine veneer:** discontinuous cover less than 1–2 m thick; underlying topography is discernible. Interspersed with small till or glaciofluvial deposits.

GLb — **Glaciolacustrine blanket:** continuous cover greater than 2 m thick; forming flat to undulating topography that locally obscures underlying geomorphology.

GLACIOFLUVIAL DEPOSITS: light orange, pebbly sand with occasional (2%) cobbles and boulders at surface deposited behind, at, or in front of the ice margin by flowing glacial meltwater. The sand is often well sorted and massive, though occasional bedding is present in some esker ridges. Where the suffix "x" has been added to the terrain unit label (i.e. GFrx) it indicates the sediments have had significant surface reworking by glacial Lake Agassiz and/or the Tyrrell Sea.

GFv — **Glaciofluvial veneer:** discontinuous sand and gravel cover, less than 1–2 m thick; underlying topography is discernible.

GFb — **Glaciofluvial blanket:** continuous sand and gravel cover greater than 2 m thick, forming flat to undulating topography that locally obscures underlying units and associated geomorphic patterns. Occasional thinner patches of sediment may occur. Unit GFbx indicates significant surface reworking by glacial Lake Agassiz and/or the Tyrrell Sea.

GFp — **Subaerial outwash sediments:** massive to stratified sand to pebbly sand with occasional (~5%) cobbles and boulders, deposited in a subaerial environment at or in front of the ice margin by glacial meltwater. Sediments are greater than 2 m thick and may drape the underlying topography like a blanket, or where thicker, mask underlying topography completely. The surface may be kettled; unit includes fan sediments deposited at the ice margin at the portal of an englacial or subglacial meltwater channel.

GFt — **Terraced sediments:** inactive terraces above modern floodplain; deposited during glacial meltwater flow in meltwater channels. The terrace along the Seal River contains about 10–20% carbonate clasts, in addition to the local shield-derived lithologies.

GFf — **Subaqueous outwash sediments:** massive to stratified sand to pebbly sand, occasionally rippled and/or crossbedded; interbedded with gravel and diamictic units of variable thickness; rare (~5%) cobbles and boulders present; sediments deposited into a shallow, subaqueous glaciolacustrine or marine environment (Tyrrell Sea), at or near the retreating ice front by meltwater turbidity currents.

GFh — **Ice-contact glaciofluvial sediments:** undifferentiated deposits; poorly sorted sand and gravel with minor diamicton; deposited by glacial meltwater in direct contact with the glacier; 1 m to greater than 20 m thick; forming gently undulating to hummocky topography related to melting of underlying ice. Features include kettles, kames, and ridges.

GFr — **Eskers and esker systems:** stratified sand and gravel with minor diamicton, deposited by meltwater flow within tunnels beneath or within the glacier; present as large (3–10 m high), long (10–25 km), regularly spaced (10–18 km) esker segments, with smaller (1–5 m high) and shorter esker ridges found between the large ridges. Esker segments consist of kame and kettle topography up to 20 m high. Eskers and crevasse-fill ridges well below marine limit have been extensively wave washed which has created resultant 'ridges' 0.25–2 m high, and a blanket of pebbly sand near the 'ridge' location, and are mapped as unit Mn.

GLACIAL DEPOSITS: unsorted to poorly sorted diamicton (till) with a sandy-silt to silty-sand matrix, deposited in subglacial or ice-marginal environments. May locally contain blocks of pre-existing sediments and/or stratified drift. Tills consist mainly of granitic material in regions overlying granitic bedrock, and consist of a more variable lithology in supracrustal bedrock regions. The till has been emplaced by ice flowing from the Keewatin Sector, within the Laurentide Ice Sheet. Where the suffix "x" has been added to the terrain unit label (e.g. Tvx, Thx, Tbx, Tstx) it indicates that the sediments have had significant surface reworking by meltwater and/or the Tyrrell Sea.

Tv — **Till veneer:** discontinuous till cover less than 1–2 m thick; underlying topography is discernible. Surface may be washed in the vicinity of meltwater channels and where marine sediments are present.

Tb — **Till blanket:** continuous till cover greater than 2 m thick, forming flat to undulating topography that locally obscures underlying units and associated geomorphic patterns. Occasional thinner patches of till may occur. Surface may be washed in the vicinity of meltwater channels and where marine sediments are present.

Tst — **Streamlined till:** till greater than 2 m thick, moulded beneath the glacier into linear ridges and/or furrows parallel to ice flow; drumlins, drumlinoid ridges, and flutings. Ridges are typically 0.1–3 km long and only 1–3 m high.

Th — **Hummocky till:** supraglacial meltout (ablation) tills deposited by melting of stagnant ice; loose, texturally variable, sandy to gravelly matrix, some sorting; angular to subangular clasts; locally includes poorly sorted sand and gravel; gently undulating to hummocky topography.

Tr — **Rogen moraine:** anastamosing to curved ridges and intervening troughs, all lying transverse to former ice-flow direction. The Rogen ridges may exhibit gradual up- and down-ice flow-direction transition to drumlinoid ridges and flutings and/or a nontransitional lateral shift to streamlined terrain. Ridges are typically 0.1–3.0 km long, with a typical segment length of 760 m (n=507). There are both 'pristine' ridges and 'drumlinized' ridges, the latter of which have been overridden by actively flowing ice, resulting in streamlining of their surfaces (*see* attribute table for delineation). The degree and size of streamlining is often transitional between minor modification to complete drumlinization.

hu — Weakly calcareous, carbonate-bearing till with a clayey-silt matrix encountered within the subsurface in the southwest portion of Great Island. This till was likely deposited by west- or northwest-flowing ice from the Labradorean Sector (Hudsonian Ice) of the Laurentide Ice Sheet.

PRE-QUATERNARY

BEDROCK

R — **Precambrian rocks:** metasedimentary, metavolcanic rocks and associated intrusive rocks; may be overlain by a thin, discontinuous veneer of till in upland, unwashed areas, and/or a thin discontinuous veneer of sand and/or pebbly sand below marine limit or within meltwater corridors that rarely exceeds 1 m thick.

NOTE: In areas where the surficial cover forms a complex pattern, the area is coloured according to the dominant unit and labelled in descending order of cover (e.g. O•Tr). Where underlying stratigraphic units are known, areas are coloured according to the overlying unit and labelled in the following manner: O/Tr.

Geological contact (defined, inferred)
Rogen moraine crest: *see* attribute table (Leg_Label field) for delineation between pristine and drumlinized ridges
Limit of mapping
Esker, direction known
Major moraine ridge
Crevasse ridge
Drumlinoid ridge or fluting
Drumlin
Streamlined bedrock, direction unknown
Minor meltwater channel, direction unknown
Major meltwater corridor
Raised beach, wave-cut notch
Limit of submergence, marine and/or glaciomarine (wave-cut benches, washing limits)
Limit of submergence, glaciolacustrine (wave-cut benches, washing limits)
Small outcrop
Field site with sample
Field site without sample
Striation, direction known, numbers indicate relative age (1 - oldest) (*see* attribute table (Rel_age field) for relative ages)
Striation, direction unknown
Roche moutonnée
Kame
Kettle

Recommended citation:
Trommelen, M.S. and Campbell, J.E., 2012. Surficial geology, Stubner Lake, Manitoba; Geological Survey of Canada, Canadian Geoscience Map 41; Manitoba Innovation, Energy and Mines, Manitoba Geological Survey, Geoscientific Map MAP2011-2, scale 1:50 000. doi:10.4095/288957

GSC CANADIAN GEOSCIENCE MAP 41 • MGS GEOSCIENTIFIC MAP MAP2011-2

Canadian Geoscience Maps
Cartes géoscientifiques du Canada

Canada   Manitoba

Authors: M.S. Trommelen[1] and J.E. Campbell[2]
[1]Manitoba Geological Survey, 360-1395 Ellice Avenue, Winnipeg, Manitoba R3G 3P2
[2]Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario K1A 0E8

Aerial photograph interpretation (1:50 000 scale) and geology by M.S. Trommelen (2009 and 2010 field seasons)

Digital compilation by L. Robertson, 2010–2011

Cartography by D. Viner

Scientific editing by E. Inglis

Map projection Univeral Transverse Mercator, zone 14. North American Datum 1983

Base map at the scale of 1:50 000 from Natural Resources Canada, with modifications.

SURFICIAL GEOLOGY
STUBNER LAKE
Manitoba
1:50 000

1 0 1 2 3 4 km

Elevations in metres above mean sea level.

Magnetic declination 2012, 0°5' W increasing 8.0' annually

The Geological Survey of Canada and Manitoba Geological Survey welcomes corrections or additional information from users.
This map was produced from processes that conform to the Scientific and Technical Publishing Services Quality Management System, registered to the ISO 9001:2008 standard.

This publication, including digital data, can be downloaded free of charge from GeoPub (http://geopub.nrcan.gc.ca/). It is also available from the Geological Survey of Canada Bookstore (http://gsc.nrcan.gc.ca/bookstore).

This publication can also be downloaded in PDF, free of charge, from the Manitoba government web site at http://manitoba.ca/minerals.

GEOLOGICAL SURVEY OF CANADA
CANADIAN GEOSCIENCE MAP 41
MANITOBA GEOLOGICAL SURVEY
GEOSCIENTIFIC MAP MAP2011-2
SURFICIAL GEOLOGY
STUBNER LAKE
Manitoba

Four trim marks around perimeter of map sheet. Trim map sheet first, then fold at folding marks.

Cover and additional panels are 17cm wide when folded.

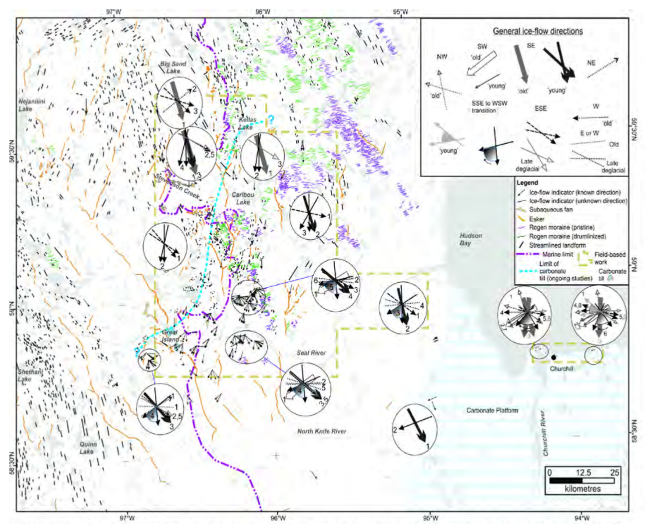
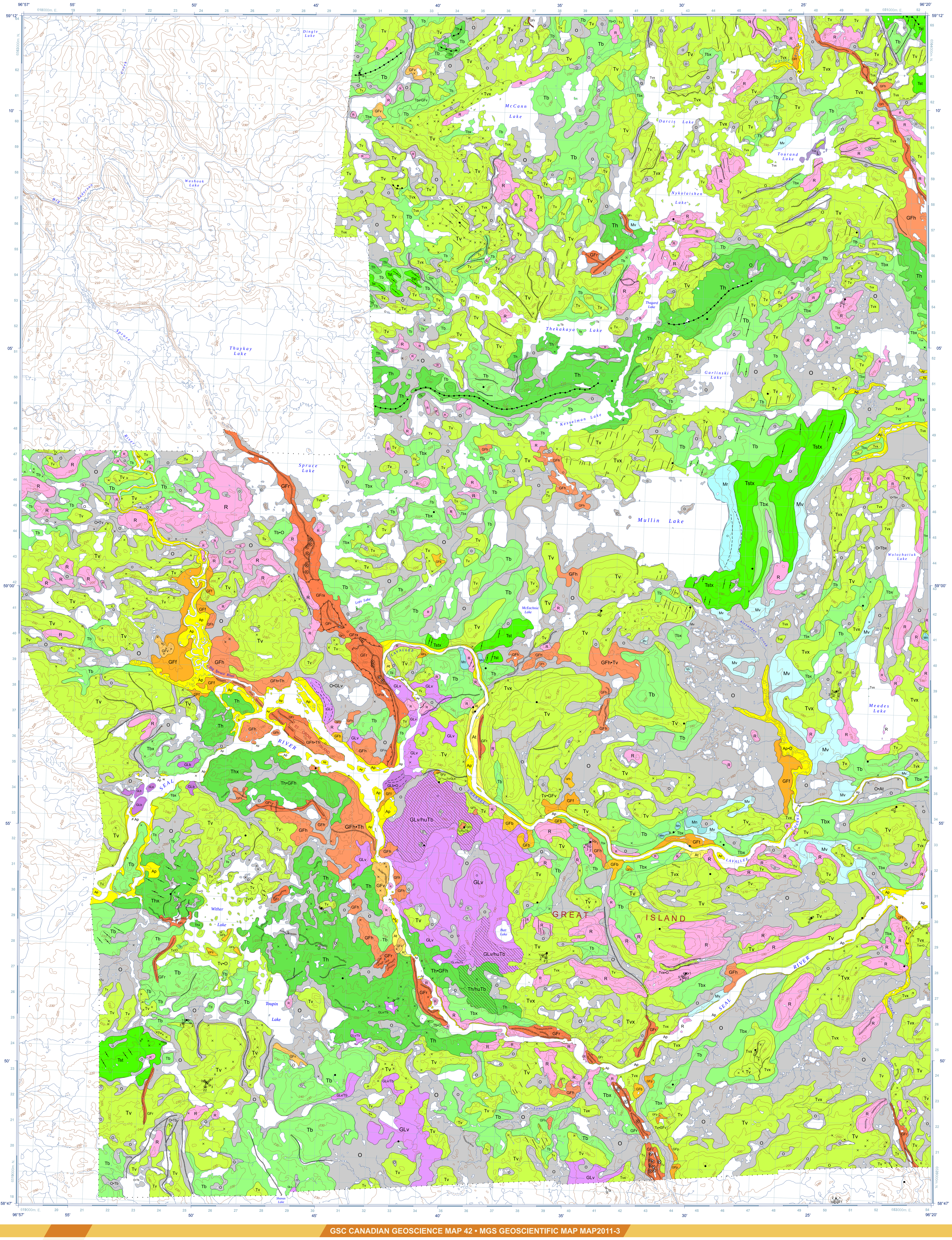


Figure 1. Ice-flow indicators, marine limit, and limit of carbonate clasts in northeastern Manitoba.



This legend is common to CGM 40, CGM 41, CGM 42, and CGM 43.
Coloured legend blocks indicate map units that appear on this map.
Not all symbols shown in the legend appear on this map.

QUATERNARY

SURFICIAL DEPOSITS

HOLOCENE

NONGLACIAL ENVIRONMENTS

O — **ORGANIC DEPOSITS:** Undifferentiated peat and muck; 1 m to greater than 5 m thick; formed by the accumulation of plant material in various stages of decomposition; generally occur as flat, wet terrain (swamps and bogs) over poorly drained substrates. Fibric fens are present along some water channels. Thickness varies from thin organic veneers (20–40 cm) overlying till and boulder fields to organic plains greater than 3 m thick. Thick organic deposits typically overlie fine-grained glaciolacustrine and marine sediments. Permafrost is commonly present underlying and/or within thick organic deposits, as seen by the prevalent raised bogs with ice-wedge polygons. Small, unmapped deposits commonly occur in most terrain units. Peat mantles most geological units.

ALLUVIAL DEPOSITS: Sorted sand, silt, and clay with minor gravel and organic detritus; commonly stratified; deposited along and/or within all modern rivers and streams.

Ap — **Floodplain sediments:** sorted sand, silt, clay, minor gravel, and organic detritus greater than 1 m thick; forming active floodplains close to river and stream level; includes terraces too small to show at this map scale.

At — **Fluvial terraces:** inactive terraces above modern floodplain; greater than 2 m thick; consisting of gravel, sand, and overbank silt and organic detritus on the Seal and Caribou rivers. Annual spring ice-push continues to build up sediment along the side of these terraces.

Af — **Fan-delta sediments:** poorly sorted sand and organic detritus deposited at the western side of Caribou Lake.

L — **LACUSTRINE DEPOSITS:** Undifferentiated; massive to stratified, sorted sand, silt, clay, and minor organic detritus deposited adjacent and/or within modern ponds and lakes.

MARINE SEDIMENTS: Poor to well sorted sand and silt with 0–20% pebbles, cobbles, and occasional boulders (ice rafted and lags), deposited in the postglacial Tyrrell Sea. Clasts are typically subrounded to subangular, occasionally striated and/or faceted and/or bullet-shaped, derived from the reworking of till. The marine limit is between 165 m a.s.l. and 180 m a.s.l., defined by washing limits on eskers and till plains and by the elevations of sand blankets and beaches. The exact elevation is uncertain, owing to the likelihood that glacial Lake Agassiz was coeval to the Tyrrell

Mv — **Marine veneer**

Mb — **Marine blanket**

Mn — **Nearshore sediments**

Mr — **Littoral sediments**

LATE WISCONSINAN

PROGLACIAL AND GLACIAL ENVIRONMENTS

GLACIOLACUSTRINE DEPOSITS

GLd — **Ice-contact deltaic sediments**

GLv — **Glaciolacustrine veneer**

GLb — **Glaciolacustrine blanket**

GLACIOFLUVIAL DEPOSITS

GFv — **Glaciofluvial veneer**

GFb — **Glaciofluvial blanket:** … associated geomorphic patterns. Occasional thinner patches of sediment may occur. Unit GFbx indicates significant surface reworking by glacial Lake Agassiz and/or the Tyrrell Sea.

GFp — **Subaerial outwash sediments:** massive to stratified sand to pebbly sand with occasional (~5%) cobbles and boulders, deposited in a subaerial environment at or in front of the ice margin by glacial meltwater. Sediments are greater than 2 m thick and may drape the underlying topography like a blanket, or where thicker, mask underlying topography completely. The surface may be kettled; unit includes fan sediments deposited at the ice margin at the portal of an englacial or subglacial meltwater channel.

GFt — **Terraced sediments:** inactive terraces above modern floodplain; deposited during glacial meltwater flow in meltwater channels. The terrace along the Seal River contains about 10–20% carbonate clasts, in addition to the local shield-derived lithologies.

GFf — **Subaqueous outwash sediments:** massive to stratified sand to pebbly sand, occasionally rippled and/or crossbedded; interbedded with gravel and diamictic units of variable thickness; rare (~5%) cobbles and boulders present; sediments deposited into a shallow, subaqueous glaciolacustrine or marine environment (Tyrrell Sea), at or near the retreating ice front by meltwater turbidity currents.

GFh — **Ice-contact glaciofluvial sediments:** undifferentiated deposits; poorly sorted sand and gravel with minor diamicton; deposited by glacial meltwater in direct contact with the glacier; 1 m to greater than 20 m thick; forming gently undulating to hummocky topography related to melting of underlying ice. Features include kettles, kames, and ridges.

GFr — **Eskers and esker systems:** stratified sand and gravel with minor diamicton, deposited by meltwater flow within tunnels beneath or within the glacier; present as large (3–10 m high), long (10–25 km), regularly spaced (10–18 km) esker segments, with smaller (1–5 m high) and shorter esker ridges found between the large ridges. Esker segments consist of kame and kettle topography up to 20 m high. Eskers and crevasse-fill ridges well below marine limit have been extensively wave washed which has created resultant 'ridges' 0.25–2 m high, and a blanket of pebbly sand near the 'ridge' location, and are mapped as unit Mn.

GLACIAL DEPOSITS: unsorted to poorly sorted diamicton (till) with a sandy-silt to silty-sand matrix, deposited in subglacial or ice-marginal environments. May locally contain blocks of pre-existing sediments and/or stratified drift. Tills consist mainly of granitic material in regions overlying granitic bedrock, and consist of a more variable lithology in supracrustal bedrock regions. The till has been emplaced by ice flowing from the Keewatin Sector, within the Laurentide Ice Sheet. Where the suffix "x" has been added to the terrain unit label (e.g. Tvx, Thx, Tbx, Tstx) it indicates that the sediments have had significant surface reworking by meltwater and/or the Tyrrell Sea.

Tv — **Till veneer:** discontinuous till cover less than 1–2 m thick; underlying topography is discernible. Surface may be washed in the vicinity of meltwater channels and where marine sediments are present.

Tb — **Till blanket:** continuous till cover greater than 2 m thick, forming flat to undulating topography that locally obscures underlying units and associated geomorphic patterns. Occasional thinner patches of till may occur. Surface may be washed in the vicinity of meltwater channels and where marine sediments are present.

Tst — **Streamlined till:** till greater than 2 m thick, moulded beneath the glacier into linear ridges and/or furrows parallel to ice flow; drumlins, drumlinoid ridges, and flutings. Ridges are typically 0.1–3 km long and only 1–3 m high.

Th — **Hummocky till:** supraglacial meltout (ablation) tills deposited by melting of stagnant ice; loose, texturally variable, sandy to gravelly matrix, some sorting; angular to subangular clasts; locally includes poorly sorted sand and gravel; gently undulating to hummocky topography.

Tr — **Rogen moraine:** anastamosing to curved ridges and intervening troughs, all lying transverse to former ice-flow direction. The Rogen ridges may exhibit gradual up- and down-ice flow-direction transition to drumlinoid ridges and flutings and/or a nontransitional lateral shift to streamlined terrain. Ridges are typically 0.1–3.0 km long, with a typical segment length of 760 m (n=507). There are both 'pristine' ridges and 'drumlinized' ridges, the latter of which have been overridden by actively flowing ice, resulting in streamlining of their surfaces (*see* attribute table for delineation). The degree and size of streamlining is often transitional between minor modification to complete drumlinization.

hu — Weakly calcareous, carbonate-bearing till with a clayey-silt matrix encountered within the subsurface in the southwest portion of Great Island. This till was likely deposited by west- or northwest-flowing ice from the Labradorean Sector (Hudsonian Ice) of the Laurentide Ice Sheet.

PRE-QUATERNARY

BEDROCK

R — **Precambrian rocks:** metasedimentary, metavolcanic rocks and associated intrusive rocks; may be overlain by a thin, discontinuous veneer of till in upland, unwashed areas, and/or a thin discontinuous veneer of sand and/or pebbly sand below marine limit or within meltwater corridors that rarely exceeds 1 m thick.

NOTE: In areas where the surficial cover forms a complex pattern, the area is coloured according to the dominant unit and labelled in descending order of cover (e.g. O•Tr). Where underlying stratigraphic units are known, areas are coloured according to the overlying unit and labelled in the following manner: O/Tr.

- Geological contact (defined, inferred)
- Rogen moraine crest: *see* attribute table (Leg_Label field) for delineation between pristine and drumlinized ridges
- Limit of mapping
- Esker, direction known
- Major moraine ridge
- Crevasse ridge
- Drumlinoid ridge or fluting
- Drumlin
- Streamlined bedrock, direction unknown
- Minor meltwater channel, direction unknown
- Major meltwater corridor
- Raised beach, wave-cut notch
- Limit of submergence, marine and/or glaciomarine (wave-cut benches, washing limits)
- Limit of submergence, glaciolacustrine (wave-cut benches, washing limits)
- Small outcrop
- Field site with sample
- Field site without sample
- Striation, direction known, numbers indicate relative age (1 - oldest) (*see* attribute table (Rel_age field) for relative ages)
- Striation, direction unknown
- Roche moutonnée
- Kame
- Kettle

Recommended citation:
Trommelen, M.S. and Campbell, J.E., 2012. Surficial geology, Great Island–Seal River, Manitoba; Geological Survey of Canada, Canadian Geoscience Map 42; Manitoba Innovation, Energy and Mines, Manitoba Geological Survey, Geoscientific Map MAP2011-3, scale 1:50 000. doi:10.4095/288955

REFERENCES

ACKNOWLEDGMENTS

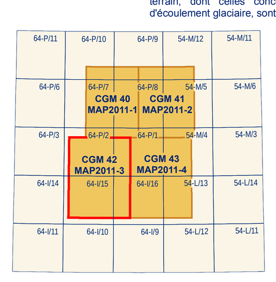
… Manitoba project - Natural Resources Canada. Field assistance by B. Clarke, T. Stainton, and D. Vessey is greatly appreciated. The authors thank D. Kerr (GSC) and G. Matile (Manitoba Geological Survey) for critical review of the map.

Abstract

Northeast Manitoba is mantled by glacial and postglacial sediments, with scarce bedrock outcrops. Past ice-flow reconstructions in northern Manitoba suggest that the region has been covered at least twice by ice from the Keewatin Sector, and at least three times by ice from the Labradorean Sector (Dredge et al., 1986; Klassen, 1986; Dredge and Thorleifson, 1987; Boulton and Clark, 1990; Dredge et al., 1990; Dredge and Nixon, 1992; Kaszycki et al., 2008). This map builds on previous 1:250 000 surficial mapping completed in the 1980s (Dredge and Nixon, 1981, 1982).
The northern part of the study area is characterized by extensive swaths of bouldery drumlinized and pristine (nondrumlinized) Rogen moraine ridges alternating with swaths of streamlined terrain. The remaining area is characterized by bedrock topography draped by a mix of till blankets and till veneers. Long, large eskers are present throughout the area, at roughly 18 km intervals. Where the eskers are located below approximately 200 m a.s.l., they have been partially eroded by lacustrine and/or marine waters. Below 150 m a.s.l., the eskers exist as washed, low-lying sand and gravel blankets rather than ridges. A mix of organic blankets and marine sediment is present in the eastern portion of the study area, predominantly below 150 m a.s.l. The study area has, in part, been wave washed by either or both the postglacial Tyrrell Sea and glacial Lake Agassiz or other smaller, disconnected glacial lakes. The marine limit in the study area is around 180 m a.s.l. Field data were obtained by helicopter-assisted ground truthing in 2009 and 2010. Further description of map units, with photos, can be found in Trommelen and Ross (2009) and Trommelen et al. (2010). Field data, including relative age of ice-flow indicators, are available in Campbell et al. (2012).

Résumé

Le nord-est du Manitoba est recouvert d'une nappe de sédiments glaciaires et postglaciaires à travers de laquelle percent de rares affleurements rocheux. Dans le nord du Manitoba, les reconstitutions des anciens écoulements glaciaires donnent à penser que la région a été couverte au moins deux fois par des glaces du Secteur du Keewatin et au moins trois fois par des glaces du Secteur du Labrador (Dredge et al., 1986; Klassen, 1986; Dredge et Thorleifson, 1987; Boulton et Clark, 1990; Dredge et al., 1990; Dredge et Nixon, 1992; Kaszycki et al., 2008). La présente carte s'appuie sur des travaux de cartographie des matériaux superficiels à l'échelle 1/250 000 réalisés dans les années 1980 (Dredge et Nixon, 1981, 1982).
La partie nord de la région à l'étude est caractérisée par de grandes bandes de terrain occupées par des crêtes de moraine de Rogen à blocs, modelées en drumlins ou conservant leur forme d'origine, qui alternent avec des bandes de terrain aux formes fuselées. Le reste de la région est caractérisé par un relief défini par la surface du socle rocheux et la couverture moulante de nappes et de plaques de till. De longs et larges eskers sont présents dans toute la région, à des intervalles d'environ 18 km. Lorsque les eskers sont situés à une altitude inférieure à environ 200 m ASL, ils ont été en partie érodés par l'action d'eaux lacustres ou marines. Au-dessous d'une altitude de 150 m ASL, les eskers ne subsistent que sous la forme de nappes de sable et gravier délavés à faible relief, plutôt que de crêtes. Un mélange de nappes de dépôts organiques et de sédiments marins sont présents dans la partie est de la zone à l'étude, surtout à des altitudes inférieures à 150 m ASL. La région à l'étude a été délavée par l'action de vagues de la Mer de Tyrrell postglaciaire, du Lac glaciaire Agassiz ou d'autres d'autres petits lacs glaciaires discontinus. Dans la région à l'étude, la limite marine est située à une altitude d'environ 180 m ASL. Les données de terrain ont été obtenues lors de vérifications des réalités de terrain menées à l'aide d'un hélicoptère en 2009 et 2010. Des descriptions plus détaillées des unités cartographiques, avec photos, peuvent être consultées dans Trommelen et Ross (2009) et Trommelen et al. (2010). Les données de terrain, dont celles concernant les âges relatifs des indicateurs d'écoulement glaciaire, sont fournies dans Campbell et al. (2012).



National Topographic System reference and index to adjoining published maps

Cover illustration: Undulating till blanket with esker in background and organics in lows. Photograph by M.S. Trommelen. Manitoba Geological Survey.

Natural Resources Canada / Ressources naturelles Canada

GEOLOGICAL SURVEY OF CANADA
CANADIAN GEOSCIENCE MAP 42
MANITOBA GEOLOGICAL SURVEY
GEOSCIENTIFIC MAP MAP2011-3
SURFICIAL GEOLOGY
GREAT ISLAND–SEAL RIVER
Manitoba
1:50 000

Canadian Geoscience Maps
Cartes géoscientifiques du Canada

Authors: M.S. Trommelen[1] and J.E. Campbell[2]
[1]Manitoba Geological Survey, 360-1395 Ellice Avenue, Winnipeg, Manitoba R3G 3P2
[2]Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario K1A 0E8

Aerial photograph interpretation (1:50 000 scale) and geology by M.S. Trommelen (2009 and 2010 field seasons)

Digital compilation by L. Robertson, 2010–2011

Cartography by M. Méthot

Scientific editing by E. Inglis

Map projection Univeral Transverse Mercator, zone 14. North American Datum 1983

Base map at the scale of 1:50 000 from Natural Resources Canada, with modifications.

SURFICIAL GEOLOGY
GREAT ISLAND–SEAL RIVER
Manitoba
1:50 000

Elevations in metres above mean sea level.

Magnetic declination 2012, 0°56' E decreasing 9.0' annually.

The Geological Survey of Canada and the Manitoba Geological Survey welcomes corrections or additional information from users.
This map was produced from processes that conforms to the Scientific and Technical Publishing Services Quality Management System, registered to the ISO 9001:2008 standard.

This publication, including digital data, can be downloaded free of charge from GeoPub (http://geopub.nrcan.gc.ca/). It is also available from the Geological Survey of Canada Bookstore (http://gsc.nrcan.gc.ca/bookstore).

This publication can also be downloaded in PDF, free of charge, from the Manitoba government web site at http://manitoba.ca/minerals.

GEOLOGICAL SURVEY OF CANADA
CANADIAN GEOSCIENCE MAP 42
MANITOBA GEOLOGICAL SURVEY
GEOSCIENTIFIC MAP MAP2011-3
SURFICIAL GEOLOGY
GREAT ISLAND–SEAL RIVER
Manitoba

Four trim marks around perimeter of map sheet. Trim map sheet first, then fold at folding marks.

Cover and additional panels are 17cm wide when folded.

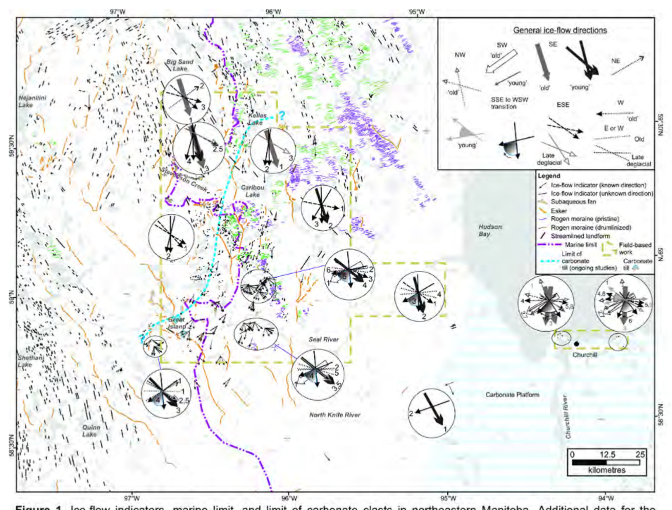
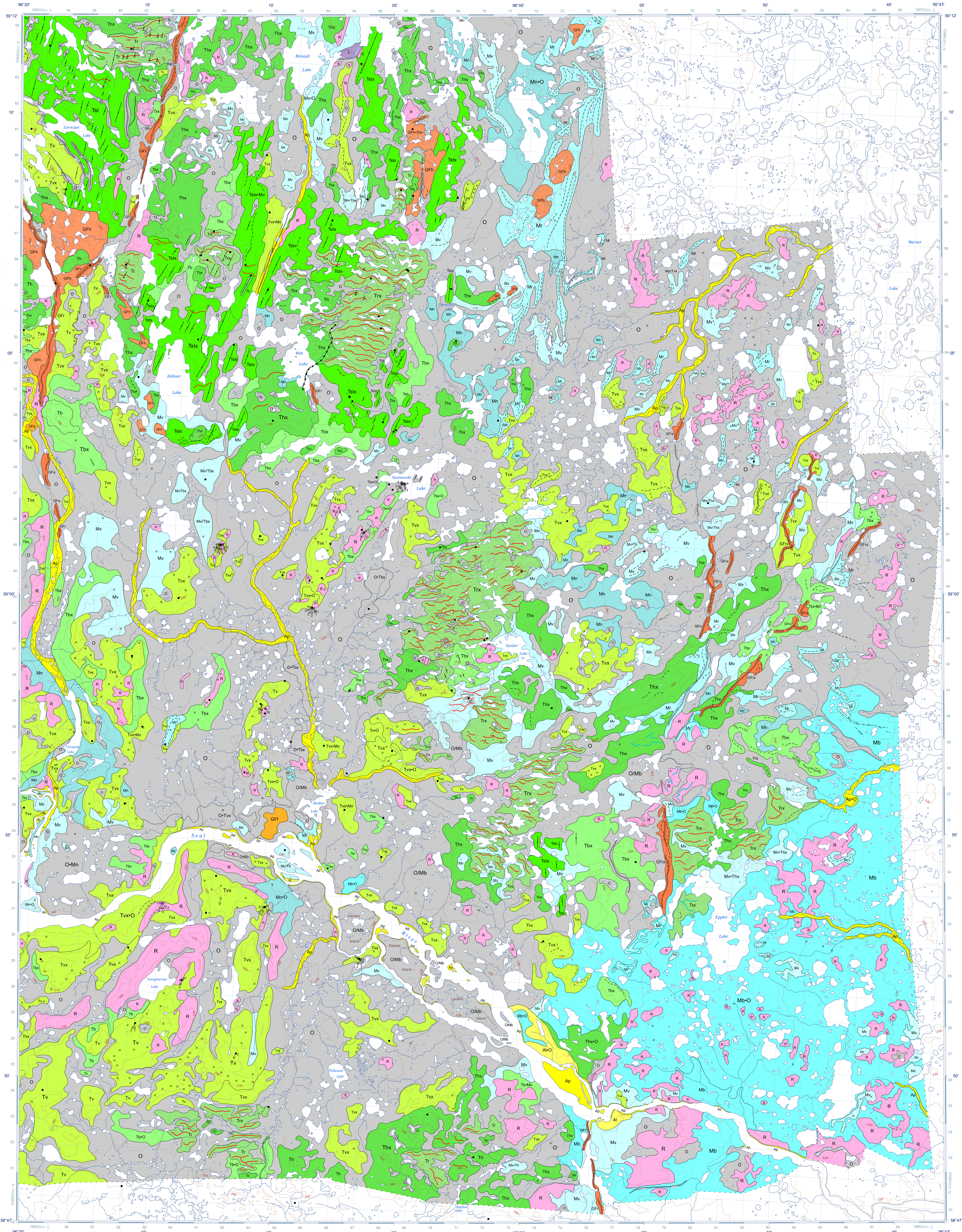


Figure 1. Ice-flow indicators, marine limit, and limit of carbonate clasts in northeastern Manitoba.



REFERENCES

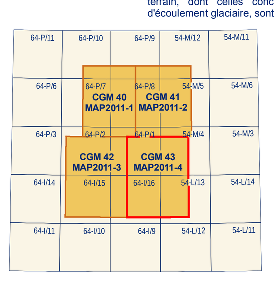
Boulton, G.S. and Clark, C.D., 1990. A highly mobile Laurentide ice sheet revealed by satellite images of glacial lineations; Nature, v. 346, p. 813–817.

ACKNOWLEDGMENTS

This map was completed as part of a Ph.D. degree at the University of Waterloo, supervised by M. Ross, funding for which was provided by the Geological Survey of Canada's (GSC) Research Affiliate Program. Field logistics and helicopter support were provided by the Manitoba Geological Survey. This mapping exercise also formed part of the GSC Geomapping for Energy and Minerals Program (GEM) Northeast Manitoba project - Natural Resources Canada. Field assistance by B. Clarke, T. Bastien, and C. Vassey is greatly appreciated. The authors thank D. Kerr (GSC) and G. Matile (Manitoba Geological Survey) for critical review of the map.

Abstract

Résumé



National Topographic System reference and index to adjoining published maps

Cover illustration:
Streamlined terrain foreground, Rogen moraine background, just north of Sosnowski Lake. Photograph by M.S. Trommelen, Manitoba Geological Survey.

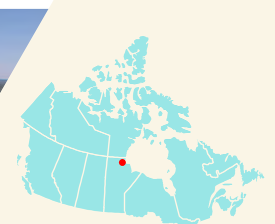
Printed map:
Catalogue No. M183-1/43-2011E
ISBN 978-0-660-20103-0
Digital map:
Catalogue No. M183-1/43-2011E-PDF
ISBN 978-1-100-19084-6

doi:10.4095/288958

Natural Resources Canada | Ressources naturelles Canada

GEOLOGICAL SURVEY OF CANADA
CANADIAN GEOSCIENCE MAP 43
MANITOBA GEOLOGICAL SURVEY
GEOSCIENTIFIC MAP MAP2011-4

SURFICIAL GEOLOGY
SOSNOWSKI LAKE
Manitoba
1:50 000



Canadian Geoscience Maps
Cartes géoscientifiques du Canada

Four trim marks around perimeter of map sheet. Trim map sheet first, then fold at folding marks.

Cover and additional panels are 17 cm wide when folded.

This legend is common to CGM 40, CGM 41, CGM 42, and CGM 43.
Coloured legend blocks indicate map units that appear on this map.
Not all symbols shown in the legend appear on this map.

QUATERNARY

SURFICIAL DEPOSITS

HOLOCENE

NONGLACIAL ENVIRONMENTS

O — **ORGANIC DEPOSITS:** Undifferentiated peat and muck; 1 m to greater than 5 m thick; formed by the accumulation of plant material in various stages of decomposition; generally occur as flat, wet terrain (swamps and bogs) over poorly drained substrates. Fibric fens are present along some water channels. Thickness varies from thin organic veneers (20–40 cm) overlying till and boulder fields to organic plains greater than 3 m thick. Thick organic deposits typically overlie fine-grained glaciolacustrine and marine sediments. Permafrost is commonly present underlying and/or within thick organic deposits, as seen by the prevalent raised bogs with ice-wedge polygons. Small, unmapped deposits commonly occur in most terrain units. Peat mantles most geological units.

ALLUVIAL DEPOSITS: Sorted sand, silt, and clay with minor gravel and organic detritus; commonly stratified; deposited along and/or within all modern rivers and streams.

Ap — **Floodplain sediments:** sorted sand, silt, clay, minor gravel, and organic detritus greater than 1 m thick; forming active floodplains close to river and stream level; includes terraces too small to show at this map scale.

At — **Fluvial terraces:** inactive terraces above modern floodplain; greater than 2 m thick; consisting of gravel, sand, and overbank silt and organic detritus on the Seal and Caribou rivers. Annual spring ice-push continues to build up sediment along the side of these terraces.

Af — **Fan-delta sediments:** poorly sorted sand and organic detritus deposited at the western side of Caribou Lake.

L — **LACUSTRINE DEPOSITS:** Undifferentiated; massive to stratified, sorted sand, silt, clay, and minor organic detritus deposited adjacent and/or within modern ponds and lakes.

MARINE SEDIMENTS: Poor to well sorted sand and silt with 0–20% pebbles, cobbles, and occasional boulders (ice rafted and lags), deposited in the postglacial Tyrrell Sea. Clasts are typically subrounded to subangular, occasionally striated and/or faceted and/or bullet-shaped, derived from the reworking of till. The marine limit is between 165 m a.s.l. and 180 m a.s.l., defined by washing limits on eskers and till plains and by the elevations of sand blankets and beaches. The exact elevation is uncertain, owing to the likelihood that glacial Lake Agassiz was coeval to the Tyrrell Sea during deglaciation. Near the marine limit, glaciomarine sediment also occurs. These sand and silt deposits locally include pockets of debris-flow sediments, till, and/or minor dropstones, deposited from suspension and iceberg rafting.

Mv — **Marine veneer:** discontinuous sand less than 1–2 m thick that drape the existing topography; overlies wave-washed till between 180 m a.s.l. and 140 m a.s.l.; below 140 m a.s.l. present as sandy patches overlying bedrock outcrops where all till has been removed.

Mb — **Marine blanket:** flat to gently undulating plain of fine sand, silt, and clay greater than 2 m thick; often overlain by a layer of organic material (less than 1 m thick); sparsely fossiliferous; offshore sediment.

Mn — **Nearshore sediments:** poor to well sorted, sand, silt, and clay; occur as veneers and blankets of sediment overlying till and/or bedrock; commonly less than 2 m thick. Predominantly derived from reworking of till and/or glaciofluvial deposits.

Mr — **Littoral sediments:** poor to well sorted, stratified sand with 5–20% pebbles and cobbles; typically 1–2 m thick. Beach ridges, consisting of sand and cobbles derived from the underlying till are present at elevations of 155–170 m a.s.l. More common are linear patches of pebbly sand with occasional spits, derived from esker and crevasse ridges. The latter typically contain a higher percentage of exotic lithologies. Where esker and crevasse ridges occur below marine limit, wave washing has commonly reduced the ridges down to a common height of 0.25–1 m and redistributed the sand creating veneers and blankets of light orange, granitic pebbly sand. Low-lying regions or depressions often have an organic veneer overlying the sand and silt.

LATE WISCONSINAN

PROGLACIAL AND GLACIAL ENVIRONMENTS

GLACIOLACUSTRINE DEPOSITS: moderate to well sorted clay, silt, and very fine to fine sand; massive to bedded; moderately dense; deposited in glacial Lake Agassiz or other small glacial lakes along the margin of the retreating Laurentide Ice Sheet. Usually overlain by less than 0.5 m thick organic deposits in lowlands with flat topography. Some littoral sand may be marine in origin, given that the Tyrrell Sea incursion occurred in the same area, and the genesis is uncertain. Sand encountered above 180 m has been assigned as glaciolacustrine, whereas sand below 180 m is considered marine. Sediment is derived from the Archean and Paleoproterozoic rocks in the area, and predominately consists of feldspar and quartz. In the west part of Great Island, carbonate, red mudstone, and black mudstone clasts were found in calcareous silty clay between 200–260 m a.s.l. This material is quite similar to calcareous glacio-lacustrine pelite found to the west of the map area (Dredge et al., 1986), and was likely derived from Hudsonian and/or Labradorean drift deposited into glacial Lake Agassiz.

GLd — **Ice-contact deltaic sediments:** well to moderately stratified sand and gravel, forming a deltaic deposit where a meltwater channel entered a glacial lake during regression and lowering of lake levels. Surface is kettled and the landform has a steep front.

GLv — **Glaciolacustrine veneer:** discontinuous cover less than 1–2 m thick; underlying topography is discernible. Interspersed with small till or glaciofluvial deposits.

GLb — **Glaciolacustrine blanket:** continuous cover greater than 2 m thick; forming flat to undulating topography that locally obscures underlying geomorphology.

GLACIOFLUVIAL DEPOSITS: light orange, pebbly sand with occasional (2%) cobbles and boulders at surface deposited behind, at, or in front of the ice margin by flowing glacial meltwater. The sand is often well sorted and massive, though occasional bedding is present in some esker ridges. Where the suffix "x" has been added to the terrain unit label (i.e. GFrx) it indicates the sediments have had significant surface reworking by glacial Lake Agassiz and/or the Tyrrell Sea.

GFv — **Glaciofluvial veneer:** discontinuous sand and gravel cover, less than 1–2 m thick; underlying topography is discernible.

GFb — **Glaciofluvial blanket:** continuous sand and gravel cover greater than 2 m thick, forming flat to undulating topography that locally obscures underlying units and associated geomorphic patterns. Occasional thinner patches of sediment may occur. Unit GFbx indicates significant surface reworking by glacial Lake Agassiz and/or the Tyrrell Sea.

GFp — **Subaerial outwash sediments:** massive to stratified sand to pebbly sand with occasional (~5%) cobbles and boulders, deposited in a subaerial environment at or in front of the ice margin by glacial meltwater. Sediments are greater than 2 m thick and may drape the underlying topography like a blanket, or where thicker, mask underlying topography completely. The surface may be kettled; unit includes fan sediments deposited at the ice margin at the portal of an englacial or subglacial meltwater channel.

GFt — **Terraced sediments:** inactive terraces above modern floodplain; deposited during glacial meltwater flow in meltwater channels. The terrace along the Seal River contains about 10–20% carbonate clasts, in addition to the local shield-derived lithologies.

GFf — **Subaqueous outwash sediments:** massive to stratified sand to pebbly sand, occasionally rippled and/or crossbedded; interbedded with gravel and diamictic units of variable thickness; rare (~5%) cobbles and boulders present; sediments deposited into a shallow, subaqueous glaciolacustrine or marine environment (Tyrrell Sea), at or near the retreating ice front by meltwater turbidity currents.

GFh — **Ice-contact glaciofluvial sediments:** undifferentiated deposits; poorly sorted sand and gravel with minor diamicton; deposited by glacial meltwater in direct contact with the glacier; 1 m to greater than 20 m thick; forming gently undulating to hummocky topography related to melting of underlying ice. Features include kettles, kames, and ridges.

GFr — **Eskers and esker systems:** stratified sand and gravel with minor diamicton, deposited by meltwater flow within tunnels beneath or within the glacier; present as large (3–10 m high), long (10–25 km), regularly spaced (10–18 km) esker segments, with smaller (1–5 m high) and shorter esker ridges found between the large ridges. Esker segments consist of kame and kettle topography up to 20 m high. Eskers and crevasse-fill ridges well below marine limit have been extensively wave washed which has created resultant 'ridges' 0.25–2 m high, and a blanket of pebbly sand near the 'ridge' location, and are mapped as unit Mn.

GLACIAL DEPOSITS: unsorted to poorly sorted diamicton (till) with a sandy-silt to silty-sand matrix, deposited in subglacial or ice-marginal environments. May locally contain blocks of pre-existing sediments and/or stratified drift. Tills consist mainly of granitic material in regions overlying granitic bedrock, and consist of a more variable lithology in supracrustal bedrock regions. The till has been emplaced by ice flowing from the Keewatin Sector, within the Laurentide Ice Sheet. Where the suffix "x" has been added to the terrain unit label (e.g. Tvx, Thx, Tbx, Tstx) it indicates that the sediments have had significant surface reworking by meltwater and/or the Tyrrell Sea.

Tv — **Till veneer:** discontinuous till cover less than 1–2 m thick; underlying topography is discernible. Surface may be washed in the vicinity of meltwater channels and where marine sediments are present.

Tb — **Till blanket:** continuous till cover greater than 2 m thick, forming flat to undulating topography that locally obscures underlying units and associated geomorphic patterns. Occasional thinner patches of till may occur. Surface may be washed in the vicinity of meltwater channels and where marine sediments are present.

Tst — **Streamlined till:** till greater than 2 m thick, moulded beneath the glacier into linear ridges and/or furrows parallel to ice flow; drumlins, drumlinoid ridges, and flutings. Ridges are typically 0.1–3 km long and only 1–3 m high.

Th — **Hummocky till:** supraglacial meltout (ablation) tills deposited by melting of stagnant ice; loose, texturally variable, sandy to gravelly matrix, some sorting; angular to subangular clasts; locally includes poorly sorted sand and gravel; gently undulating to hummocky topography.

Tr — **Rogen moraine:** anastamosing to curved ridges and intervening troughs, all lying transverse to former ice-flow direction. The Rogen ridges may exhibit gradual up- and down-ice flow-direction transition to drumlinoid ridges and flutings and/or a nontransitional lateral shift to streamlined terrain. Ridges are typically 0.1–3.0 km long, with a typical segment length of 760 m (n=507). There are both 'pristine' ridges and 'drumlinized' ridges, the latter of which have been overridden by actively flowing ice, resulting in streamlining of their surfaces (*see* attribute table for delineation). The degree and size of streamlining is often transitional between minor modification to complete drumlinization.

hu — Weakly calcareous, carbonate-bearing till with a clayey-silt matrix encountered within the subsurface in the southwest portion of Great Island. This till was likely deposited by west- or northwest-flowing ice from the Labradorean Sector (Hudsonian Ice) of the Laurentide Ice Sheet.

PRE-QUATERNARY

BEDROCK

R — **Precambrian rocks:** metasedimentary, metavolcanic rocks and associated intrusive rocks; may be overlain by a thin, discontinuous veneer of till in upland, unwashed areas, and/or a thin discontinuous veneer of sand and/or pebbly sand below marine limit or within meltwater corridors that rarely exceeds 1 m thick.

NOTE: In areas where the surficial cover forms a complex pattern, the area is coloured according to the dominant unit and labelled in descending order of cover (e.g. O•Tr). Where underlying stratigraphic units are known, areas are coloured according to the overlying unit and labelled in the following manner: O/Tr.

Geological contact (defined, inferred)
Rogen moraine crest: *see* attribute table (Leg_Label field) for delineation between pristine and drumlinized ridges
Limit of mapping
Esker, direction known
Major moraine ridge
Crevasse ridge
Drumlinoid ridge or fluting
Drumlin
Streamlined bedrock, direction unknown
Minor meltwater channel, direction unknown
Major meltwater corridor
Raised beach, wave-cut notch
Limit of submergence, marine and/or glaciomarine (wave-cut benches, washing limits)
Limit of submergence, glaciolacustrine (wave-cut benches, washing limits)
Small outcrop
Field site with sample
Field site without sample
Striation, direction known, numbers indicate relative age (1 - oldest) (*see* attribute table (Rel_age field) for relative ages)
Striation, direction unknown
Roche moutonnée
Kame
Kettle

Recommended citation:
Trommelen, M.S. and Campbell, J.E., 2012. Surficial geology, Sosnowski Lake, Manitoba; Geological Survey of Canada, Canadian Geoscience Map 43; Manitoba Innovation, Energy and Mines, Manitoba Geological Survey, Geoscientific Map MAP2011-4, scale 1:50 000. doi:10.4095/288958

GSC CANADIAN GEOSCIENCE MAP 43 • MGS GEOSCIENTIFIC MAP MAP2011-4

Authors: M.S. Trommelen[1] and J.E. Campbell[2]
[1]Manitoba Geological Survey, 360-1395 Ellice Avenue, Winnipeg, Manitoba R3G 3P2
[2]Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario K1A 0E8

Aerial photograph interpretation (1:50 000 scale) and geology by M.S. Trommelen (2009 and 2010 field seasons)

Digital compilation by L. Robertson, 2010–2011

Cartography by D. Viner

Scientific editing by E. Inglis

Map projection Univeral Transverse Mercator, zone 14. North American Datum 1983

Base map at the scale of 1:50 000 from Natural Resources Canada, with modifications.

SURFICIAL GEOLOGY
SOSNOWSKI LAKE
Manitoba
1:50 000

1 0 1 2 3 4 km

Elevations in metres above mean sea level

Magnetic declination 2012, 0°10' E decreasing 8.0' annually

The Geological Survey of Canada and Manitoba Geological Survey welcome corrections or additional information from users.
This map was produced from processes that conform to the Scientific and Technical Publishing Services Quality Management System, registered to the ISO 9001:2008 standard.

This publication, including digital data, can be downloaded free of charge from GeoPub (http://geopub.nrcan.gc.ca/). It is also available from the Geological Survey of Canada Bookstore (http://gsc.nrcan.gc.ca/bookstore).

This publication can also be downloaded in PDF, free of charge, from the Manitoba government web site at http://manitoba.ca/minerals.

GEOLOGICAL SURVEY OF CANADA
CANADIAN GEOSCIENCE MAP 43
MANITOBA GEOLOGICAL SURVEY
GEOSCIENTIFIC MAP MAP2011-4
SURFICIAL GEOLOGY
SOSNOWSKI LAKE
Manitoba