Glacial Geomorphology of southern Alberta, Canada

YOUNG, NATHANIEL,JOSEPH,PETER

Glacial Geomorphology of southern Alberta, Canada

Nathaniel Joseph Peter Young

A thesis submitted for the degree of
Master of Science

Department of Geography
University of Durham

April 2009

Declaration of Copyright

I confirm that no part of the material presented in this thesis has previously been submitted by me or any other person for a degree in this or any other university. In all cases, where it is relevant, material from the work of others has been acknowledged.

Nathaniel Joseph Peter Young

Department of Geography
University of Durham
April 2009

Abstract

During deglaciation from the Last Glacial Maximum three terrestrial ice streams within the south western sector of the Laurentide Ice Sheet competed and coalesced in southern Alberta; the High Plains Ice Stream (HPIS), Central Alberta Ice Stream (CAIS) and the east lobe. The ice streams are characterised by smoothed corridors along which lie lineations that identify multiple flow events, transverse ridges of thrust and push origin, esker networks and large sequences of parallel and transverse meltwater channels. The CAIS and HPIS were dynamic and transitory in nature creating a 'time-transgressive' imprint.

The CAIS terminated within southern Alberta creating a wealth of landforms composed of controlled, hummocky, push and thrust block moraines, along with doughnut hummocks, ice walled lake plains, recessional meltwater, tunnel and large spillway channels. The CAIS margin is interpreted to have been polythermal in nature, creating a continuum of landforms that is dominated by active marginal recession.

The methodologies used were placed within an overarching 'scale approach', whereby the research initially focused on a small, regional scale and gradually moved to large scale, local investigations. Firstly, Shuttle Radar Topography Mission (SRTM) and Landsat 7 Enhanced Thematic Mapper (Landsat ETM+) data sets were used to map the regional picture. Then, Aerial Photo Investigation (API), ground truthing and sedimentary analyses were employed to provide a detailed, localised focus into the landform sediment assemblages in southern Alberta.

Contents

Page

List of Figures Page

List of Tables **Page**

Acknowledgements

I would firstly like to thank my parents for all of their support and generosity during the unexpectedly long time it has taken to complete this thesis. Without them this would not have been possible, thank you mum and dad. Similarly, I would like to thank my Grandmother for her remarkable generosity and support during this time.

My thanks go to my auntie for giving me a home in Durham during the writing of the thesis, and for her fantastic company after long hard days in the GIS lab. My thanks go to my uncle in Calgary who very kindly gave me a place to stay during my fieldwork in Alberta and for making my stay much more comfortable and enjoyable than it could have been. A big thank you must go to Matthew Dick for his enjoyable company during our time in Canada, especially during fieldwork.

Special thanks go to Colm Ó Cofaigh for his help, patience and guidance during this project, especially during the last mad month of writing this. Similarly, a big thank you goes to David Evans for his help and patience during this project and for meeting me in the middle of nowhere in southern Alberta. Thanks also to Veronica Crooks for her help throughout the masters and making sure I was still a student at Durham University.

A huge thank you goes to the Foundation for Canadian Studies in the United Kingdom for their financial support towards my field work costs, and Michael Hellyer for his help during the process. My thanks also go to St Cuthbert's Society for their financial support towards fieldwork costs and for 5 fantastic years in Durham.

Chapter 1

Introduction

1.1. Rationale

During the late Wisconsinan Southern Alberta was covered by the southwestern sector of the Laurentide ice sheet (Clayton and Moran, 1982; Dyke and Prest, 1987; Dyke et al., 2002). Small scale mapping by Prest et al. (1968) and more recently Evans et al. (2008) and Ó Cofaigh et al. (in press) has identified that numerous, highly dynamic and transitory ice streams operated in this region during deglaciation from the late Wisconsinan. It is likely that these ice streams exerted strong controls over the dynamics and deglacial style of Laurentide Ice sheet (Bentley, 1987; Clark, 1994a; Bamber et al., 2000; Kaplan et al., 2001; Stokes and Clark, 2001; Bennett, 2003; Hulbe and Fahnestock, 2004). Therefore, in order to assess the ice streams overall impact upon the south western sector of the Laurentide Ice Sheet during this time, it is crucial that we accurately constrain the geomorphic imprint recorded by these ice streams in order to reconstruct palaeo-ice stream dynamics.

Ice streams are fast flowing corridors of ice within an inland ice sheet (cf. Swithinbank, 1954 and Bentley, 1987), with velocities many times greater than that of the surrounding ice, capable of reaching velocities of up to 12 km a^{-1} (Price and Whillans, 2001). They act as the main arteries draining an ice sheet, and so regulate its stability and dynamics. For example, it is estimated that ice streams discharge over 90% of all ice and sediment from the Antarctic Ice Sheet whilst only occupying 13% of the ice sheet perimeter (Morgan et al., 1982; Paterson, 1994). Due to the controls exerted over ice sheet mass balance, ice streams have the ability to significantly affect sea-level, ocean and atmospheric circulation and climate (Alley and Bindschadler, 2001). To this end, ice streams have been suggested as the driving force behind numerous and significant climate perturbations during the late Wisconsinan (Bond and Lotti, 1990; MacAyeal, 1993b).

Due to their importance and wide reaching impacts, ice streams present one of the most important multi-disciplinary research frontiers today. Many key questions still remain, such as, how do they maintain and regulate their flow? How do they activate themselves in time and space? What initiated large scale changes in their flow regimes? Do ice streams directly respond to climate? The answers to these and many other questions are crucial if we are to fully understand ice streams and their impacts. The corollary of which would enable unprecedented accuracy in modelling ice streams and the impacts they may have upon the dynamics of the Antarctic and Greenland ice sheets.

The literature review below considers the key concepts relevant to research on terrestrial palaeo-ice stream margins; in particular, it highlights the specific aspects of the glacial geomorphology of western Canada that will form the focus of this research. The glacial geomorphology of southern Alberta provides an excellent opportunity to add fresh insights into an important avenue of Quaternary science and palaeo-glaciology. This research can be justified by three key areas:

1) The subglacial settings of palaeo-ice streams have been the primary focus of palaeo-ice stream research. This has led to a relative neglect of the marginal settings of terrestrial ice streams. The corollary of this is that we do not fully understand the role of terrestrial ice stream margins on ice stream dynamics and ultimately ice sheet operation as a whole. It is therefore crucial that ongoing research seeks not only to explain subglacial regimes but also the marginal landform suites of ice streams. We can thereby provide a more holistic assessment of all the components of terrestrial palaeo-ice streams, including their behaviour, geomorphic imprints and hence their 'life cycle' (Clark and Stokes, 2003).

   Key Questions for this research:

   - What is the glacial landsystem present in the marginal setting of the ice streams that operated in southern Alberta during the decline from the LGM?
   - What insights can individual landforms and the collective landform suite of southern Alberta provide into terrestrial ice stream marginal dynamics?
   - What control did the marginal area of the southern Alberta ice streams exert over their behaviour and stability?

2) Mapping of the glacial geomorphology of southern and central Alberta (Stalker, 1960; 1977; Prest et al., 1968; Bik, 1969; Westgate, 1968; Shetsen, 1987, 1990; Fulton, 1995, Evans et al., 1999, 2006a, 2008) has enabled a broad identification of flow patterns and landform assemblages. However, the mapping compiled so far lacks the resolution necessary for the detailed assessments of regional palaeo-ice sheet reconstruction. With the advent of new global data sets that contain elevation data, in particular the Shuttle Radar Topography Mission (SRTM), there is an opportunity to produce a detailed map of the glacial geomorphology of southern Alberta with a view to compiling a regional picture of palaeo-ice sheet dynamics.

   Key Questions:

- What is the regional glacial geomorphology of southern Alberta composed of?
- Exactly what landforms comprise the marginal setting of the Alberta ice streams?
- How useful is SRTM data for the mapping of glacial geomorphology on a regional scale?

3) Alberta has been subject to scientific research for over a century due its wealth of readily accessible, complex and varied glacial geomorphology (e.g. Dawson, 1875; Dowling, 1917; Wickenden, 1937; Wyatt et al., 1941; Stalker, 1960, 1977; Westgate, 1968; Bik, 1969; Shaw, 1983, 2002; Shetsen, 1984, 1987, 1990; Evans, 1996, 2000; Kulig, 1996; Evans et al., 1999, Evans et al., 2006a, Evans et al., 2008). More recently the identification of palaeo-ice streams in the province has re-established the more traditional view that subglacial streamlined bedforms and marginal moraine construction were responsible for the geomorphic signature rather than catastrophic subglacial meltwater floods (e.g. Shaw, 1983, 2002; Rains et al., 1993; Munro-Stasiuk, 1999; Munro-Stasiuk and Shaw, 1997, 2002; Beaney and Shaw, 2000). The region continues to be central to the megaflood debate (Benn and Evans, 2006; Evans et al., 2006a; Munro Stasiuk and Sjogren, 2006), and will be examined through detailed studies of glacial geomorphology and sedimentology.

   Key Question:

   - How were the streamlined bedforms, hummocky terrain and marginal moraines in southern Alberta formed?

1.2. Aims and Objectives

The overall aim of this research is to investigate the landform-sediment assemblages in southern Alberta, in order to evaluate and potentially reconstruct ice marginal dynamics of a terrestrial palaeo-ice stream.

In order to fulfil this aim, the following objectives will be addressed:

1. Apply a scale dependant research model that will implement the use of SRTM, Landsat ETM+ and aerial photograph data in order to accurately constrain landform assemblages on a small (provincial) scale and large (township/range) scale, and compile these data in glacial geomorphology maps of southern Alberta, with particular

focus on the ice marginal setting of the proposed Central Alberta Ice Stream (Evans et al., 2008).

2. Critically assess SRTM data with respect to mapping small scale glacial geomorphology.
3. Identify any diagnostic landforms or landform groups that may be indicative of proposed terrestrial ice stream margins and assess their implications for ice stream dynamics.
4. Where possible use site specific stratigraphy to critically assess the proposed constructional mechanisms for ice marginal landforms and sediments in southern Alberta in the context of theoretical models for regional till architecture (Boulton, 1996a, b).

1.3. Ice Streams

During the late 1970s the idea of modern day ice sheets containing fast flowing outlets became widely acknowledged (Rose, 1979). This notion provided the basis for major research projects in Greenland and Antarctica (e.g. Alley et al., 1986; Bindschadler et al., 1987), which yielded insights into the controls on these fast flowing outlets. It was hypothesised that if contemporary ice sheets contained such outlets then the palaeo-ice sheets of the last glaciation would also have had similar areas of fast flow (Denton and Hughes, 1981). It therefore followed that any reconstructions of palaeo-ice sheets must account for such features (e.g. Dyke and Prest, 1987; Boulton and Clark, 1990). It is only within the last decade that the true extent and locations of palaeo-ice streams are starting to be fully understood (e.g. Stokes and Clark, 1999; 2001; Clark and Stokes, 2003; Winsborrow et al., 2004). This has been facilitated by significant advances in remote sensing techniques and availability of continental scale data coverage. It also became apparent during this time that palaeo-ice streams have a unique landsystem that can be identified in the geomorphological record (Dyke and Morris, 1988; Kleman and Borgström, 1996; Stokes and Clark, 1999; 2001; Clark and Stokes, 2003). Such terrestrial observations have also been replicated by marine geophysical investigations of the offshore record from high-latitude glaciated continental shelves (e.g. Canals et al., 2000; Wellner et al., 2001; Ó Cofaigh et al., 2002). The coupling of contemporary and palaeo-ice stream data has also helped to strengthen modelling capability and has provided strong evidence that ice streaming occurred across Europe and North America during the LGM (e.g. Bourgeois et al., 2000; Stokes and Clark, 2001; Clark and Stokes, 2003; Piotrowski et al., 2003; Winsborrow et al., 2004; de Angelis and Kleman, 2005; Dowdeswell et al., 2006; Kleman and Glasser, 2007).

1.4. Ice-stream life cycle

Whilst there is at present no unifying theory that can account for the onset, flow dynamics, and eventual shutdown ('life cycle') of ice streams, a number of different and often contradictory theories exist. These are reviewed below.

1.4.1. Onset

During what stage of ice sheet growth or decay and exactly how ice streams initiate is currently unknown. It seems likely that during ice sheet growth ice streams initiate after ice sheet dynamics and properties reach a critical threshold that is dependent upon several factors: distribution of basal stresses, subglacial water availability, distribution and pressures, geology, ice thickness, topography and location (Bentley, 1987; Anandakrishnan et al., 1998; Bell et al., 1998; Bennet, 2003; Bamber et al., 2006; Peters et al., 2006; Kleman and Glasser, 2007; de Angelis and Kleman, 2008). All these factors are intimately linked through multiple feedback loops and so a change in the state of one would likely impact upon the others.

Onset zones or onsets are located at the heads of ice streams and define the area between slow flow through internal deformation and streaming flow through subglacial deformation or basal sliding (Bindschadler et al., 2001). It is likely that within the onset zone, motion is a product of both internal and sediment deformation, and basal sliding (Whillans et al., 2001; Joughin et al., 2002). Joughin et al. (2002) further suggested that two types of onset zone currently exist: tributary and ice stream onsets. Research into onset zones has primarily focused on the Siple Ice Streams in Antarctica (Hodge and Doppelhammer, 1996; Bell et al., 1998; Bindschadler et al., 2000, 2001; Joughin et al., 2002; Wang et al., 2003; Peters et al., 2006; Bell et al., 2007). Research into palaeo-ice stream onsets has revealed that they are seldom preserved during deglaciation (De Angelis and Kleman, 2008), however, some possible geomorphic indicators have been identified, the most common of which are highly convergent lineations and increasing lineation length has also been regarded as indicative of the down-ice section of onsets (Dyke and Morris, 1988; Stokes and Clark, 2003a; Evans et al., 2004; Ottesen et al., 2005; De Angelis and Kleman, 2008). Similarly, ribbed moraine sequences have been linked to onsets (e.g. Dyke and Morris, 1988; Dyke et al., 1992; De Angelis and Kleman, 2008), along with carbonate plumes, lateral sliding boundaries and transverse scarps (Dyke and Morris, 1988; De Angelis and Kleman, 2008).

Within onset zones certain conditions must prevail for ice streaming to begin. In the absence of significant amounts of subglacial water at pressure melting point, accelerated motion through either sliding (Engelhardt and Kamb, 1998) or subglacial deformation (Alley et al., 1989) would not be possible. Additionally, ice sheets may either be entirely warm based or a combination of cold and warm based, implying that onsets within either type of ice sheet must have different properties. Importantly, changes in thermal bed conditions from cold to warm based have been closely linked to the onset of ice streaming (e.g. Dyke et al., 1992; De Angelis and Kleman, 2008). Such changes in basal thermal regimes have been associated with discontinuities in basal topography (Mckintyre, 1985); ice sheet dynamics (Paterson, 1994; Kleman and Glasser, 2007) and geothermal heat flow (Fahnestock et al., 2001; Vogel and Tulaczyk, 2006). Where warm based conditions prevail, sedimentary basins (Anandakrishnan et al., 1998; Bell et al., 1998; Bamber et al., 2006; Peters et al., 2006), subglacial lakes (Bell et al., 2007), ice property instabilities (Wang et al., 2003), longitudinal stresses (Price et al., 2002) and self organization (Payne and Dongelmans, 1997) have all been invoked to explain the onset of ice streaming.

1.4.2. Ice streaming

Why do ice streams flow at such high velocities? Answering this question is a key element in modelling future behaviour of the Antarctic ice streams and hence the Antarctic ice sheet. At present however, fast flow and ice streaming are *"...still only partly understood and their representation in quantitative models of ice [sheets] is rudimentary*" Tulaczyk (2006, pp.353).

Ice stream flow occurs through either subglacial deformation (Alley et al., 1986), basal sliding (Engelhardt and Kamb, 1998) or a combination of the two (Bennett, 2003; Piotrowski et al., 2004). Alley et al. (1986) first suggested 'the deformable bed model', whereby high velocities were possible through the deformation of highly saturated subglacial sediment. The presence of subglacial water at close to overburden pressure decreases the effective stress of the underlying till column, enabling deformation to occur. Subsequent work by Engelhardt and Kamb (1998) and Kamb (2001) suggested that very high porewater pressures enable decoupling of ice from the bed and the high velocities are then reached through basal sliding with little or no subglacial deformation. Due to the logistical problems in accessing the beds of contemporary ice streams it has been difficult to identify a primary flow mechanism. However, it is likely that during its life cycle an individual ice stream will experience both sliding and deformation, with temporal and spatial variations in the relative dominance of each (Piotowski and Kraus, 1997; Smith, 1997; Bennett, 2003). This helps to answer how ice stream velocities fluctuate on such short time scales. It is clear that the key to understanding ice stream flow and velocity fluctuations is reliant upon our ability to accurately constrain the basal hydrology regimes beneath contemporary ice streams.

Where motion occurs through subglacial sediment deformation it is essential that we understand the appropriate till flow laws. This is for two main reasons: firstly, in order to model ice motion through subglacial sediment deformation it is necessary to understand how till behaves rheologically. Secondly, it is widely accepted that subglacial deformation plays a significant role in the production of till (e.g. Evans et al., 2006b), and so elucidating its overall role in that process will increase the value of Pleistocene tills in glacio-dynamic reconstructions.

The exact flow law of till, the contribution of deformation to till genesis and thickness of the deforming layer is a matter of considerable debate (Ó Cofaigh and Stokes, 2008). Two schools of thought have emerged: The first argues that tills deform in a plastic manner on all scales (Tulaczyk et al., 2000a, b); whilst the second argues that the net accumulation of small scale plastic failures shows that at a large scale till behaves in a viscous manner (Hindmarsh, 1997; Alley, 2000; Fowler, 2002). It is generally agreed that till deforms plastically on a small scale (Kamb, 1991, Hindmarsh, 1997; Bennett, 2003). Tills that operate under a plastic flow law cannot support a stress greater than the yield stress of the till and the rate of deformation thereafter is independent of the stress applied (Kamb, 1991; Iverson et al., 1995, 1998; Hooke et al., 1997; Murray, 1997; Tulaczyk, 2000a; Bennet, 2003). However, Hindmarsh (1997) argues that it is necessary to know the rate of deformation above the yield stress in order to model the *"time dependent behaviour of glaciers"* (p. 1041). Therefore, on a large scale such as ice streams and ice sheets, this model requires that till deforms in a viscous manner, as has previously been assumed by Alley et al. (1987) and Boulton (1996a, b). It is also important that these models are able to explain the production of landforms found along the base of an ice stream. In that sense the viscous flow law is able to account for the production of drumlins and lineations along the bed due to the depth at which deformation occurs (Hindmarsh, 1998a, b). Ó Cofaigh and Stokes (2008) suggest from measurements on ice streams and glaciers that the thickness of deformation in generally less than 1 m. Observations by Engelhardt and Kamb (1998) along the Antarctic ice streams show that deformation depth is variable, which suggests that individual ice streams demonstrate variable mosaics of deformation along their bed (Piotrowski et al., 2004).

Interestingly, Stokes and Clark (2003a, b) have demonstrated that a palaeo-ice stream in the northwestern Canadian Shield flowed over hard bedrock. This was previously assumed to be implausible and has only been hypothesized by Punkari (1995) for the Scandinavian Ice Sheet. Its discovery has wide reaching implications for the deformable bed theory and any modelling of ice sheets that assumed soft-sediment geology (Stokes and Clark, 2003a, b). It also highlights the issue that not all contemporary ice streams can act as analogues for all palaeo-ice streams.

1.4.3. Temporal behaviour and shutdown

The preceding two sections have alluded to the velocity fluctuations inherent within ice streams. Their ability to fluctuate on a variety of timescales has significant impacts upon the mass balance of an ice sheet. Of most importance, is understanding the longer term variations in ice stream behaviour. Rapid fluctuations in velocity across an ice stream are most likely due to the localised variations in basal water pressures (Bennett, 2003). However, Binschadler et al. (2003) identified hourly variations in flow induced by a tidal trigger. This 'stick-slip' behaviour is a common characteristic of fast flowing outlet glaciers and helps to explain cyclical velocity fluctuations, whereby motion is controlled by variations in subglacial water pressure (Boulton et al., 2001), further illustrating the importance of basal hydrological regimes.

Ultimately, ice sheets will respond to climate change, however, long term variations and styles of deglaciation within ice streams have been shown to be independent of external forcing (Jennings, 2006; Tulaczyk, 2006, Ó Cofaigh et al., 2008). Research into the eventual shutdown of contemporary ice streams, in particular Ice Stream C has highlighted several factors including surging, loss of lubricating till, ice-shelf back-stress, ice piracy, water piracy and thermal processes (Anandakrishnan et al., 2001). It seems likely that some of these factors were associated with the decline and eventual shutdown of palaeo-ice streams. Similarly, 'sticky spots' have been strongly associated with ice stream longevity (Anandakrishnan and Alley, 1994), onset zones and ice stream shut-down (Stokes et al., 2007). Sticky spots are defined as *"localised areas of high basal drag surrounded by a well lubricated low strength bed"* (cf. Alley, 1993, in Stokes et al., 2007). They have been further divided into four types: bedrock bumps; till free bedrock areas; areas of strong well drained till; and basal freezing of subglacial sediments and meltwater (Stokes et al., 2007). Whilst there is a lack of data concerning such features it seems logical to assume that if present along ice streams beds they could well account for any spatial and temporal flow variability. It is also worth noting that ice stream shear margins exert a significant control on the flow balance of an ice stream. For example, side drag has been calculated to support between 50 and 100% of the driving force, which may vary along the profile of an ice stream (Echelmeyer et al., 1994; Raymond et al., 2001).

1.5. Palaeo-Ice Streams

Palaeo-ice streams have yielded many new insights into the dynamics of fast glacier flow and have many advantages over their contemporary counterparts, most notably the ease of access to those based in terrestrial environments. The geomorphic record left behind by palaeo-ice streams are the key to identifying their location and extent, life cycle, how they may have initiated, style of deglaciation and dynamic nature. Complex questions remain regarding the flow regimes of terrestrial ice streams; for example, were ablation rates at the lobate terminus high enough to maintain fast flow? If not, are we therefore seeing stop-start or surging signatures in the landform record?

1.5.1. Categories and criteria

Many palaeo-ice stream tracks have been identified within the former Laurentide, Cordilleran, British and Scandinavian Ice Sheets (See Stokes and Clark, 2001, and Winsborrow et al., 2004 for a review). Palaeo-ice streams are categorised into two groups: terrestrial and marine-based ice streams (Fig. 1) (Stokes and Clark, 2001; Clark and Stokes, 2003). Terrestrial ice streams, of which there are no contemporary examples are characterised by a large lobate terminus that likely extended significantly past the regional ice sheet margin (Clark and Stokes, 2003; Jennings, 2006). All contemporary ice streams are marine based, and are characterised by calving directly into the ocean or via an ice shelf (Clark and Stokes, 2003). Ice streams can be further divided into 'pure' and 'topographic' ice streams, the former being unrelated to bedrock topography (Stokes and Clark, 1999).

The life cycle of a terrestrial ice stream is directly recorded in the landform record and can be characterised by two styles of imprint: the 'rubber stamped' or 'smudged' imprint (Fig. 2) (Stokes and Clark, 2001; Clark and Stokes, 2003). The 'rubber stamp' imprint is indicative of a rapid retreat of the ice stream margin, enabling good bedform preservation and hence represents a 'snap shot' of ice stream activity at a point in time (Clark and Stokes, 2003). In contrast the smudged or time-transgressive imprint records an ice stream that continued to operate as it retreated (Clark and Stokes, 2003). This in turn leaves a complex, overprinted and often cross cutting landform signature (Evans et al., 2008). Recently, evidence from the Antarctic continental shelf has identified three different styles of retreat for marine-terminating palaeo-ice streams: fast, episodic and slow (Ó Cofaigh et al., 2008). Similar to the rubber stamped imprint, the rapid retreat style is characterised by preservation of subglacial landforms and a lack of transverse grounding zone wedges (GZWs) or recessional moraines (Ó Cofaigh et al., 2008). Rapid retreat is possible by floatation and calving (Ó Cofaigh et al., 2008). The episodic retreat style is characterised by mega-scale glacial lineations (MSGLs) interrupted by overprinted GZWs (Ó Cofaigh et al., 2008). The slow style is represented by recessional moraines and GZWs (Ó Cofaigh et al., 2008). In order to help identify palaeo-ice streams Stokes and Clark (1999, 2001) proposed a list of geomorphological criteria: *1) Characteristic shape and dimensions; 2) Highly convergent flow patterns; 3) Highly attenuated subglacial*

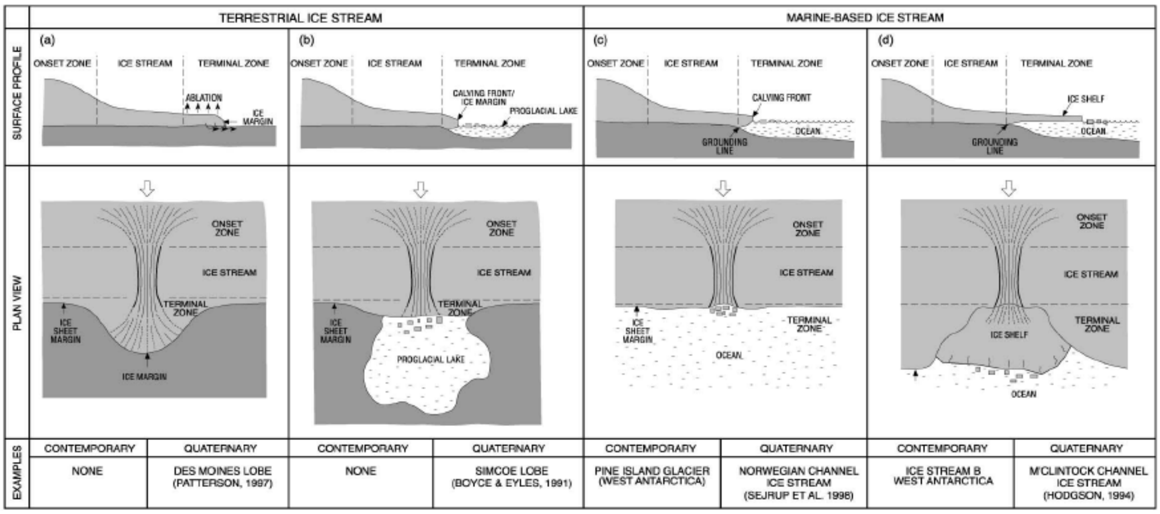


***Figure.1**. The two styles of ice streams. Both show clear flow convergence into a zone of rapid velocity that terminates in either a large lobate margin or calves into open water or an ice shelf. Marine based ice streams are easily able to remove material which renders them more susceptible to rapid velocity fluctuations (from Stokes and Clark, 2001)*

bedforms (length:width > 10: 1); 4) Boothia-type erratic dispersal trains; 5) Abrupt lateral margins (<2km); 6) Ice stream marginal moraines; 7) Glaciotectonic and geotechnical evidence of pervasively deformed till; 8) Submarine accumulation such as till deltas or sediment fans.

1.5.2. Dynamics

Reconstructions of palaeo-ice streams rely on the identification of the above geomorphological criteria because it has been widely demonstrated that such landforms were produced by ice streams (e.g. Clark, 1993, 1994b; Stokes and Clark, 1999; 2001; Clark and Stokes, 2003). However, further complications arise due to the complexity of the landform record, often displaying multiple flowsets that dissect and overprint each other. This problem was first identified by Clark (1993), who discovered cross-cutting lineations that could not have formed during a single flow event. This clearly showed the highly transitory and dynamic nature of ice streams. Subsequently, numerous examples of cross cutting palaeo-ice stream tracts have been identified in the landform record (e.g. de Angelis and Kleman, 2005; Stokes et al., 2006). Recent work by Ó Cofaigh et al. (in press) has highlighted the ability of a terrestrial palaeo-ice stream to reorganise its flow orientation multiple times within a single deglaciation, possibly on time scales as short as 100 years. In addition, the authors suggest that flow reorganisation was controlled by basal thermal conditions, with the decline of one ice stream being the trigger for the onset of others. This concurs with De Angelis and Kleman (2008) that the basal thermal boundary between cold and warm based ice is a key element in ice stream initiation.

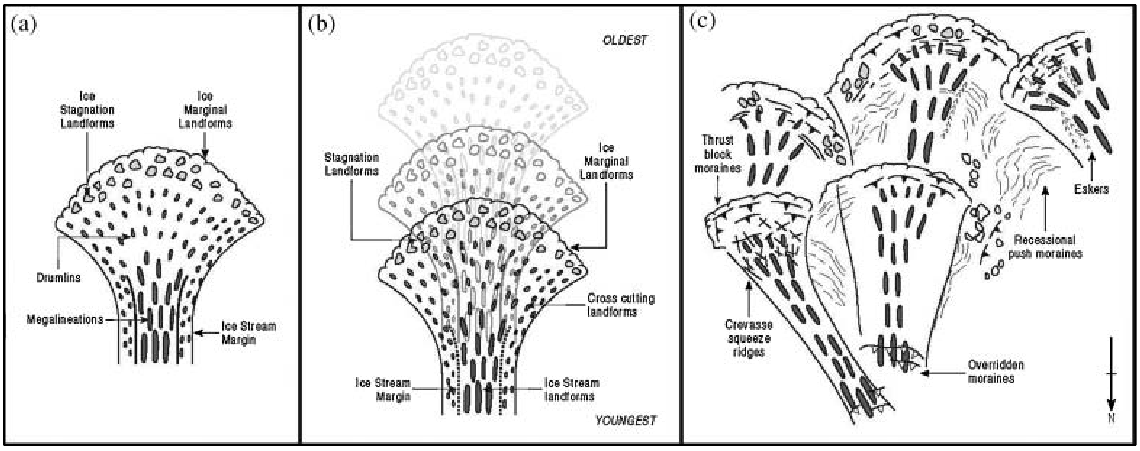


***Figure.2**. a) represents the 'rubber stamped' imprint and b) the 'time transgressive' imprint. c) represents the complex landform record of multiple overprinting flow-sets. In addition to this the transitory nature of ice streams can show flow orientation changes of up to 90° (e.g. Ó Cofaigh et al., in press). Image taken from Evans et al. (2008), adapted from Clark and Stokes (2003).*

1.6. The Geomorphic Imprint

The imprint left by a palaeo-ice stream is distinct and readily identifiable in the landform record. This section will review the types of landforms prominent in the subglacial and marginal settings of a terrestrial palaeo-ice stream, and comment on the various genetic theories.

1.6.1. Mega-scale glacial lineations

Whilst such subglacial landforms have long been recognized (e.g. Fairchild, 1907), the term MSGL was first introduced by Clark (1993). They are generally defined as *"elongate streamlined ridges of sediment produced subglacially and are similar to flutes and drumlins but much larger in all dimensions"* (Clark et al., 2003, p.240). MSGLs generally exist as part of a set of multiple and spatially coherent landforms that together form a flow-set (Clark, 1994b; 1997) Their sizes generally range from 6-100 km in length, widths from 200-1300 m, with spacing of 200 m - 5 km (Clark et al., 2003). For example a 35 km megafluting complex has been identified in southern Alberta (Evans, 1996). Clark and Stokes (2003) suggest that because they form under high strain rates and fast velocity they can be used as an indicator of palaeo-ice stream location. Almost all documented examples of palaeo-ice streams display these landforms with only a few exceptions where ice streaming is reported to have occurred over hard bedrock (Roberts and Long, 2005). An arbitrary elongation ratio (ER) (length:width) ≥10:1 has also been proposed by Stokes and Clark (2002) to be indicative of fast ice flow. This idea has recently been further supported by Briner (2007) with a study of the New York drumlin field and Ottesen et al. (2008). For terrestrially terminating ice streams ERs increase towards the lobate margin and similarly, for marine based ice streams ERs should increase towards the grounding line. Despite the general morphological criteria outlined above, exactly when a drumlin or lineation becomes an MSGL is not well documented (Ottesen et al., 2008), especially since they are possibly part of a landform continuum (Rose, 1987). No arbitrary length or width requirements will be applied for lineations to comprise a flowset, and in agreement with Clark et al's. (2003) criteria, MSGLs are lineations greater than 6 km in length.

Since their existence was demonstrated along the fore-field of contemporary ice streams on the Antarctic Peninsula continental shelf (e.g. Shipp et al., 1999; Canals et al., 2000; Wellner et al., 2001; Ó Cofaigh et al., 2002), such subglacial landforms almost certainly originate from ice streaming. The long held assumption that MSGLs form under fast flow therefore seems to be appropriate. However, there is a currently no consensus as to their genesis. There are multiple theories for drumlin and fluting genesis, they include sediment deformation (Dyson, 1952, Smalley and Unwin, 1968; Boulton, 1976; Smalley and Piotrowski, 1987; Boulton, 1987; Boulton and Hindmarsh, 1987; Clark, 1993; Benn, 1994; Benn and Evans, 1996; Eklund and Hart, 1996), basal lodgement (Fairchild, 1907; Boulton, 1982), depositional meltout (Shaw, 1980), fluvial deposition in subglacial cavaties (Shaw, 1983; Dardis and McCabe, 1983; Shaw and Kvill, 1984; Sharpe, 1987) and remnant erosional forms (Shaw, 1994; Fisher and Spooner, 1994; Shaw, 1996; Munro-Stasiuk and Shaw, 2002; Shaw, 2002). Apart from the fluvial origins the majority of these theories suggest that drumlins and flutings should have a limiting size (Menzies and Rose, 1987). Therefore, whilst tenants of these theories may be applicable at a larger scale, any application to MSGL scale must be carefully considered. Benn and Evans (1998) suggest that MSGLs form either through the erosion of intervening hollows, the accretion of sediment in hills, or some combination of the two. The meltwater hypothesis (Section 1.6.3.) has evolved from a theory that viewed drumlins as a product of fluvial deposition into subglacial cavities, to one that invokes a genesis of remnant erosional landforms (Shaw, 1996, 2002). More recently a groove ploughing theory has been put forward by Tulaczyk et al. (2001) and Clark et al. (2003). This model proposes that keels generated in the base of the ice by bedrock contact, flow over deformable sediment and carve elongate grooves that displace sediment up into intervening ridges. The theory does not require the till to behave with either viscous or plastic rheology and also accepts that groove ploughing does not always occur.

1.6.2. Ribbed 'Rogen' moraines

Ribbed moraines are subglacial transverse ridges found in the interior parts of formerly glaciated areas (Hätterstrand and Klemen, 1999), and have been reported along palaeo-ie stream beds (Dunlop and Clark, 2006; Stokes et al., 2008). A comprehensive review of 33,000 ribbed moraine by Dunlop and Clark (2006) shows a huge range of scale in their morphology: 17-1116 m (average 278 m) wide, 1-64 m (av. 17 m) high, 45-16214 m (av. 688 m) long and wavelengths from 12-5800 m (av. 505 m). Dyke et al. (1992) were the first to recognize their occurrence within palaeo-ice stream tracks on the Prince of Wales Island, Arctic Canada. The link between ribbed moraine and ice streaming has been further identified by Stokes and Clark (2003a), Dunlop and Clark (2006); Finlayson and Bradwell (2008) and Stokes et al. (2008).

It is clear from the current literature that ribbed moraines remain poorly understood. Theories on their genesis can be placed into four groups: i) Shear and stacking of till slabs or debris rich basal ice through compressive flow (Shaw, 1979; Aylsworth and Shilts, 1989; Bouchard, 1989; Finlayson and Bradwell, 2008; Lindén et al., 2008; Stokes et al., 2008); ii) fracturing of frozen sediment sheets at the transition from cold to warm based conditions (Lundqvist, 1969; Dyke et al., 1992; Hätterstrand, 1997; Hätterstrand and Kleman, 1999; Sarla, 2006; Finlayson and Bradwell, 2008); iii) remoulding of a pre-existing landforms such as ice marginal transverse ridges (Möller, 2006), or flutes and drumlins following a ~90° shift in flow direction (Boulton, 1987); iv) cavity infills from subglacial meltwater (Fisher and Shaw, 1992; Shaw, 2002). Finlayson and Bradwell (2008) alluded to ribbed moraines being a polygenetic landform, perhaps the wide range of morphological and spatial characteristics identified by Dunlop and Clark (2006) is evidence of this.

Along palaeo-ice stream tracks ribbed moraine are found in onset zones and along the main trunk of the ice stream. In order to explain their presence along ice streams, it is necessary to invoke ice dynamics on an ice stream and even ice sheet scale. Ribbed moraines found in the onset zones of palaeo-ice streams have been explained by changing thermal conditions and hence the initiation of ice streaming (Dyke et al., 1992; De Angelis and Kleman, 2008). Stokes et al. (2008) used superimposed ribbed moraines along the trunk of the Dubawnt Lake Ice Stream as evidence of ice stream shutdown. The authors proposed that they represent a 'sticky spot' (Section 1.4.3.) and were formed by compressional flow initiated by dewatering, which ultimately caused the ice stream to shut down (Stokes et al., 2007, 2008). Hätterstrand (1997) and Hätterstrand and Kleman (1999) suggest that ribbed moraines demarcate the boundary between cold and warm based ice and so can be used in reconstructing the extent of cold based zones within ice sheets. Lastly, ribbed moraines have been used as evidence to infer that huge meltwater discharges are responsible for a suite of subglacial landforms (Fisher and Shaw, 1992; Shaw, 2002). That ribbed moraines represent a very important landform in relation to ice dynamics is unquestionable and so elucidating their origins is vital.

1.6.3. Meltwater hypothesis

The meltwater hypothesis explains a suite of subglacial landforms through the erosional or depositional action of catastrophic subglacial meltwater discharges beneath the Laurentide ice Sheet. These landforms include drumlins, flutings and MSGLs (Shaw, 1983; Shaw and Kvill, 1984; Shaw et al., 1989; 2000; Shaw, 1996; Munro-Stasiuk and Shaw, 2002; Shaw, 2002), ribbed moraine (Fisher and Shaw, 1992, Shaw, 2002), hummocky terrain (Munro-Stasiuk and Shaw, 1997; Munro-Stasiuk, 1999; Sjogren, 1999; Munro-Stasiuk and Sjogren, 2006), transverse ridges (Beaney and Shaw, 2000) tunnel channels (Sjogren, 1999; Beaney and Hicks, 2000; Beaney, 2002) and Scablands (Sjogren, 1994; Sjogren and Rains, 1995). Like almost all of the formational theories presented here, those that fall under the meltwater hypothesis have both weaknesses and strengths. However, as Shaw (2002) comments it is the assumptions on which these theories are built that hold the most pressing questions: where and how would such a significant amount of water occur and what would cause it to drain so catastrophically over such vast areas of an ice sheet bed (cf. Clarke et al., 2005; Benn and Evans, 2006)?

In reply, numerous scientists have dismissed the meltwater hypothesis based upon its inability to explain simply the field evidence and also on the basis of unsound scientific principles (Benn and Evans, 1998, 2006; Clarke et al, 2005). Benn and Evans (2006) argue that the evidence used to invoke the meltwater hypothesis has been interpreted in such a way as to make it unfalsifiable. Clearly, the meltwater hypothesis has instigated strong debate within the scientific literature, and will hopefully act as a catalyst for elucidating many unanswered palaeoglaciological questions.

1.7. Ice-marginal features

In comparison to subglacial landforms and dynamics, the marginal landforms of terrestrial palaeo-ice streams remain largely unexplored. The palaeo-ice stream landsystem recognises a lobate outer zone of stagnation landforms (hummocks) and ice marginal landforms (Fig. 2) (Clark and Stokes, 2003), however, there is a lack of detailed information on the extent, style and morphology of these landforms and landform suites. This is a reflection of the lack of terrestrially based palaeo-ice streams identified so far. Therefore, an assessment of the landforms and landform suites indicative of marginal processes in terrestrial ice streams is crucial to reconstructing the palaeoglaciology of the Laurentide Ice Sheet. The occurrence of southern Laurentide Ice lobes flowing as ice stream outlets or surging margins has often been suggested (Wright, 1973; Matthews, 1974; Mickelson et al., 1983; Boulton et al., 1985; Clayton et al., 1985; Clayton et al., 1989; Kulig, 1996; Patterson, 1997, 1998; Jennings, 2006). If this is correct then the large body of research on glacial landforms identified in North America may have significant implications for broader conceptual issues concerning terrestrial ice stream marginal dynamics.

The margins of terrestrial palaeo-ice streams in North America and Europe have been compared to those of surging glaciers (Clayton et al., 1985; Patterson, 1997, 1998; Hart, 1999; Colgan et al., 2003; Van der Wateren, 2003; Jennings, 2006; Evans et al., 2008; Ó Cofaigh et al., in press). This has been based on the presence of thrust-block moraines, widespread hummocky terrain and crevasse squeeze ridges which are all components of the surging landsystem (Evans and Rea, 1999, 2003). Alternatively, palaeo-ice stream margins within Alberta have also been shown to show elements of the active temperate landsystem (Evans et al., 1999; Evans, 2003; Evans et al., 2006a; Evans et al., 2008). Specifically, the abundance of low amplitude arcuate moraine ridges within southern Alberta is similar in every respect to moraines produced at contemporary temperate margins, where each ridge is proposed to record a seasonal response to climate (Evans and Twigg, 2002, Evans, 2003). This tendency for terrestrial ice stream margins to reflect both surging and active temperate behaviour highlights the need to fully understand their dynamics and controls.

1.7.1. Transverse ridges

Thrust-block moraines form part of a landform suite produced by glaciotectonics and have been documented at the margins of surging and non-surging glaciers (Sharp, 1985b; Croot, 1988; Evans and England, 1991; Fitzsimmons, 1996; Bennett et al., 1999; Evans and Rea, 1999, 2003; Van der Wateren, 2003). They are thus not necessarily indicative of surging activity (Evans and Rea, 2003). However, in order to form, significant stresses must be applied to proglacial sediments, the magnitude of which is most easily satisfied by glaciers with fast flow regimes, for example at Eyjabakkajökull, Iceland (Evans and Rea, 1999, 2003). At a surging margin they form by an acceleration or surge of the glacier snout into proglacial sediments, which may be *"seasonally frozen, unfrozen or contain discontinuous permafrost*" Evans and Rea (2003, p.260). This leads to failure of the sediments in the form of folding, overturning and thrusting of slabs (Croot, 1988; Jennings, 2006). Importantly, thrust-block moraines have been documented at several locations along the margins of southern Laurentide palaeo-ice streams (Westgate, 1968; Moran et al., 1980; Bluemle and Clayton, 1984; Shetsen, 1987; Tsui et al., 1989; Evans, 1996, 2000; Patterson, 1997, 1998; Jennings, 2006; Evans et al., 2006a; Evans et al., 2008; Ó Cofaigh et al., in press).

Low amplitude transverse ridges have also been documented along the margins of terrestrial ice streams (Westgate, 1968; Evans et al., 1999; Evans, 2000, 2003; Jennings, 2006;

Evans et al., 2008). Within Alberta such ridges are located along the High Plains Ice Stream (HPIS) and at the southern limit of the CAIS (Evans et al., 2008). They have been interpreted as minor push moraines formed by processes similar to those along contemporary active temperate glacier margins (Westgate, 1968; Stalker, 1977; Evans et al., 1999; Evans, 2000, 2003; Evans et al., 2006a; Evans et al., 2008; Evans, 2009) or alternatively as 'controlled moraine' along the HPIS (Gravenor and Kupsch, 1959; Johnson and Clayton, 2003). The presence of controlled moraines implies englacial and supraglacial origins, ice marginal polythermal conditions, and the possible presence of permafrost. They are part of a process-form continuum controlled by basal thermal regimes (Evans, 2009). Controlled moraines were first introduced by Gravenor and Kupsch (1959) but have recently been more accurately defined as *"supraglacially deposited hummocky moraine that posses clear linearity due to the inheritance of the former pattern of debris concentrations in the parent ice*" Evans (2009). Morphologically similar low amplitude ridges along the margins of the Des Moines Lobe have been interpreted as crevasse fill ridges (Jennings, 2006). Significantly, crevasse fill/squeeze ridges are well documented along contemporary surging margins (Clarke et al., 1984; Sharp 1985a, b; Evans and Rea, 1999, 2003; Evans et al., 2007). Similarly, Ó Cofaigh et al. (in press) identified crevasse squeeze ridges along the trunk of 'Ice Stream 1' in Saskatchewan, which they interpreted as evidence for surging activity. Elucidating the controls on the genesis of low amplitude transverse ridges around terrestrial ice stream margins will help unravel the dynamics which operated within these terminal zones.

1.7.2. Hummocky terrain

In agreement with Munro-Stasiuk and Shaw (1997), Munro-Stasiuk and Sjogren (2006) and Evans et al. (2008) the term 'hummocky terrain' will be used instead of 'hummocky moraine', as it is descriptive rather than implying any pre-determined landform genesis. Hummocky terrain is the most common landform type around the reconstructed margins of the southern Laurentide and Fennoscandian ice sheets (Colgan et al., 2003; Johnson and Clayton, 2003). Within these large zones of hummocky terrain there are several morphologically different types of hummock; including simple hummocks, doughnut hummocks and ice walled lake plains. Hummocks, also known as till hummocks, refer to irregular mounds and knobs surrounded by depressions, which are composed primarily of till and occur as part of chaotic zone of similar features. Doughnut hummocks are circular mounds with a central depression and have also been referred to as disintegration rings, doughnuts, circular disintegration ridges, closed ridges, rim ridges, rimmed kettles, humpies, circular moraine, extrusion moraine, Veiki moraines (in Sweden) and Pulji moraines (in Finland; Gravenor and Kupsch, 1959; Parizek, 1969; Aartolahti, 1974; Lagerbäck, 1988; Boutlon and

Caban, 1995; Mollard, 2000; Colgan et al., 2003; Knudsen, 2006). Ice walled lake plains are conspicuous irregular mounds formed from the infilling of ice walled lakes. They have also been referred to as moraine plateaux (Stalker, 1960), moraine lake plateaux (Parizek, 1969), and high kames (Stone and Peper, 1982)

Several alternative theories have been proposed for the formation of all these hummock types, almost all of which consider the landforms to have formed in a 'stagnant glacial regime' (Knudsen, 2006 p.161). Hummocks and doughnut hummocks are thought to have formed by one of three ways: first by collapse and slumping into holes of supraglacial sediments on stagnant ice (Gravenor and Kupsch, 1959; Clayton, 1967; Boulton, 1967, 1972; Parizek, 1969; Clayton and Moran, 1974; Eyles, 1979, 1983; Krüger, 1983; Paul, 1983; Clayton et al., 1985; Sollid and Sørbel, 1988: Johnson et al., 1995; Ham and Attig, 1996; Patterson, 1997, 1998; Mollard, 2000; Johnson and Clayton, 2003; Jennings, 2006; Knudsen, 2006); second is subglacially, by pressing and squeezing of saturated material into holes and crevasses in the ice or by squeezing around disintegrating ice blocks (Gravenor and Kupsch, 1959; Stalker, 1960; Aartolahti, 1974; Eyles et al., 1999; Mollard, 2000; Boone and Eyles, 2001); finally, some researchers suggest formation by subglacial meltwater erosion (Shaw, 1996; Munro-Stasiuk and Shaw, 1997; Munro-Stasiuk and Sjogren, 2006). Doughnut hummocks have also been interpreted to represent proglacial blow-out of over-pressured groundwater (Bluemle, 1993; Boutlon and Caban, 1995; Evans et al, 1999; Evans, 2003, 2009). Hummocks are quite likely a polygenetic landform and so show evidence for both supraglacial and subglacial origins. Ice walled lake plains occur individually and can reach up to a few km in diameter and as much as a few tens of metres above surrounding terrain (Clayton et al., 2008). They have been confidently interpreted as supraglacial ice walled lakes which persisted long into the stagnation cycle due to the permafrost conditions that persisted during deglaciation (Clayton and Cherry, 1967; Parizek, 1969; Attig, 1993; Ham and Attig, 1996; Clayton et al., 2001; Attig et al., 2003, Colgan et al., 2003; Clayton et al., 2008).

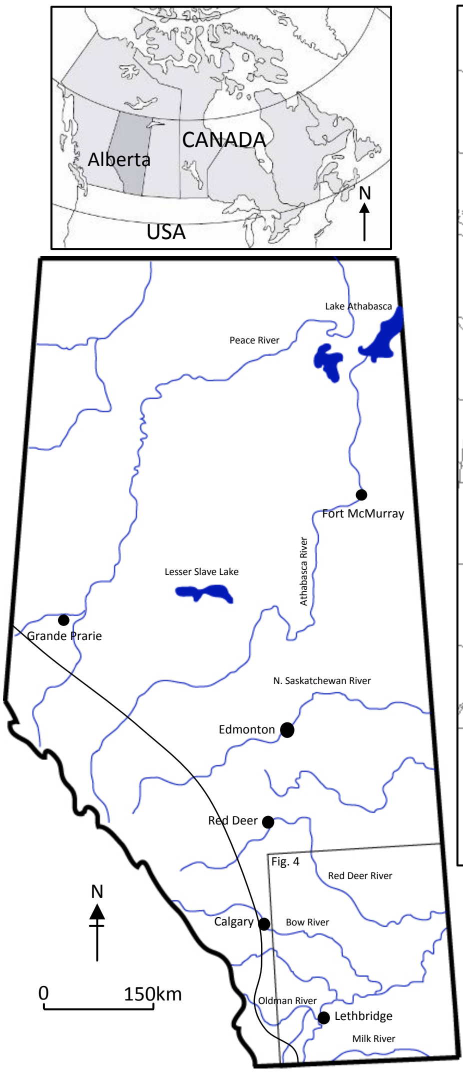
Hummocky terrain along contemporary ice margins is also generally believed to occur through meltout of supraglacial debris, and is associated with ice stagnation and surging margins (Johnson, 1972; Wright, 1980; Sharp, 1985b; Evans and Rea, 1999, 2003; Evans et al., 2007). Alternatively, some studies have suggested that hummocky terrain may be a product of active ice margins (Dyke and Savelle, 2000; Lukas, 2005, 2007) rather than extensive regional stagnation.

The widespread presence of hummocky terrain around the southern Laurentide margins has also been used as evidence of surging (Clayton et al., 1985; Colgan et al., 2003). Some of the hummocky terrain landforms described above are yet to be confidently identified at contemporary margins, and so care must be taken when using form analogy to assess the origins of Pleistocene hummocky terrain (Johnson and Calyton, 2003). Colgan et al. (2003) believe that the sequences of landforms at and behind the margins of the southern Laurentide lobes can be distinguished into three main landsystem types: Low relief till plains and low relief end moraines (Landsystem A); drumlins and high relief hummocky end moraines (Landsystem B); and low relief aligned hummocks and ice thrust masses (Landsystem C). Each landsystem is a record of differing marginal regimes and, depending on whether or not the southern Laurentide ice lobes represent palaeo-ice streams (Patterson, 1997, 1998; Jennings, 2006), may represent terrestrial ice stream margins. Interestingly, Landsystem C was interpreted by Colgan et al. (2003) to record fast flow or glacier surging, resulting in widespread stagnation of ice. It is also worth noting that the aligned hummocks show a strong resemblance to the controlled moraine described by Evans (2009).

A recurring theme throughout southern Laurentide and hummocky terrain research is the presence and impact of permafrost on ice marginal dynamics and landform genesis (Moran et al., 1980; Attig and Clayton, 1986; Attig et al., 1989; Johnson, 1990; Mooers, 1990; Colgan, 1996; Cutler et al., 1998; Cutler et al., 1999; Cutler et al., 2000; Clayton et al., 2001; Colgan et al., 2003; Mickelson and Colgan, 2003; Bauder et al., 2005). Its presence along the former southern Laurentide ice sheet margins is widely recognised and therefore has particular relevance to landform development around terrestrial ice stream margins. Specifically, the potentially thin nature of the southern Laurentide ice lobes rendered them sensitive to any internal changes, in particular basal thermal regimes. Therefore, the presence of permafrost, in the form of ice wedge polygons and ice wedge casts represents a marginal frozen bed which in turn would have affected ice marginal dynamics and hence landform construction (Bauder et al., 2005).

1.8. Alberta

Alberta is situated in the western prairies of Canada (Fig.3) and is bordered by British Columbia to the west, Saskatchewan to the east and Northwest Territories to the North. The study area is located in southern Alberta ranging from 110°-114°W and 49°-52°N. The province contains a wealth of glacial geomorphology that was primarily formed during the late Wisconsinan by ice lobes/streams that flowed from the north east, from the Keewatin sector of the Laurentide Ice Sheet (Clague, 1989). Coalasence of the Laurentide and Cordilleran Ice Sheets occurred during the late Wisconsinan and was located along the high plains (Dyke et al., 2002; Jackson et al., 2007) forcing ice flow in a south, south-easterly direction, and is recorded by the Foothills Erratics Train (Stalker, 1956; Jackson et al., 1997; Rains et al., 1999; Jackson et al., 2007). At its maximum during the late Wisconsinan extent ice flowed through Alberta and into northern Montana (Colton et al., 1961; Westgate, 1968; Colton and Fullerton, 1986; Dyke and Prest, 1987; Fulton, 1995; Kulig, 1996; Dyke et al., 2002; Fullerton et al., 2004a, b; Davies et al., 2006). Ice sheet reconstructions suggest that deglaciation from Montana started c.14 cal. ka BP (all dates are referred to in calendar years), with the central lobe occupying the Lethbridge moraine until c.12.3 ka BP, after which it receded rapidly to the north (Clayton and Moran, 1982; Dyke and Prest, 1987). Mapping of the glacial geomorphology of southern and central Alberta (Stalker, 1960; 1977; Prest et al., 1968; Westgate, 1968; Shetsen, 1987, 1990; and Fulton, 1995, Evans et al., 1999, 2006a, 2008) has enabled a broad identification of flow patterns and landform assemblages. Three prominent ice streams seem to have operated within the province and were identified as the east, central and west lobes (Shetsen, 1984; Evans, 2000). Recently, Evans et al. (2008) suggested that the west and central flow paths be referred to as the High Plains Ice Stream (HPIS) and Central Alberta Ice Stream (CAIS) respectively. The CAIS has also been referred to as the Lethbridge lobe (Eyles et al., 1999) and



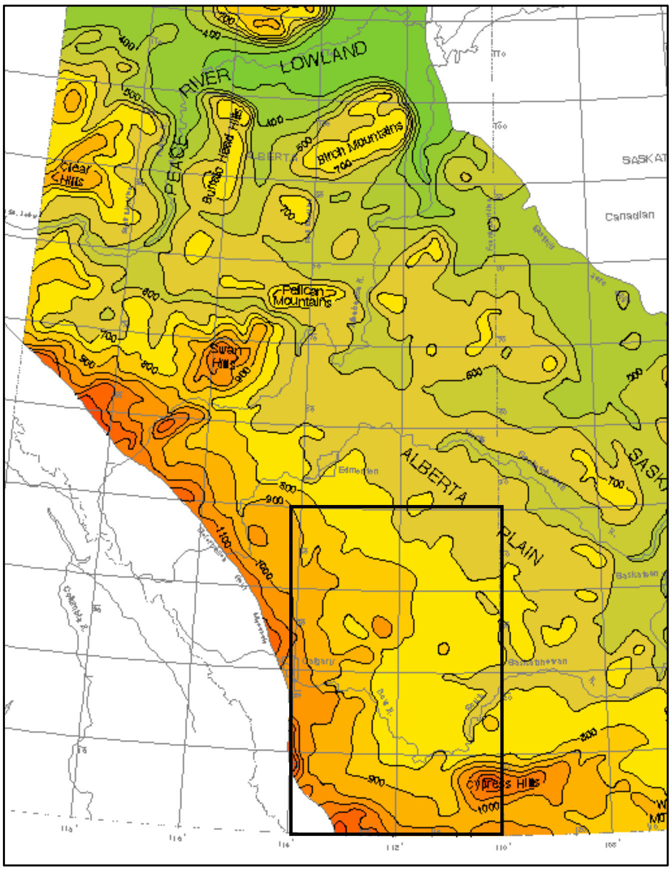


Figure 3: Location and bedrock topography maps of the study area. The location maps show the province of Alberta, Canada and the box highlights the study area (Fig. 4). The topography map is taken from The Geological Atlas of the Western Canadian Sedimentary Basin (Alberta Energy and Utilities Board/Alberta Geological Survey, 1994). The map highlights the regional north, north east dipping slope. The line on the Alberta location map identifies the western limit of the bedrock topography data.

is bordered by the McGregor, Lethbridge and Suffield moraines. During deglaciation of the province, numerous proglacial lakes formed in depressions left by the CAIS, many of which decanted forming spillways (Evans, 2000). Numerous examples of glaciotectonic disturbance have been reported (Westgate, 1968; Stalker, 1973, 1976; Tsui et al., 1989; Evans, 1996, 2000; Evans and Rea, 2003; Evans et al., 2008; Ó Cofaigh et al., in press), some of which have been interpreted as evidence of surging activity (Evans and Rea, 2003). Southern Alberta also has extensive hummocky terrain (Gravenor and Kupsch, 1959; Stalker, 1960; 1977; Shetsen, 1984, 1987, 1990; Clark et al., 1996; Munro-Stasiuk and Shaw, 1997; Evans et al., 1999; Munro-Stasiuk and Sjogren, 2006).

1.8.1. Alberta Geology

The study area is bordered by the Rocky Mountain Foothills to the west, the Cypress Hills to the east, and the Sweet Grass Hills and Milk River Ridge to the south. The whole of the study area lies within North America interior plains, within which lies the western Canadian Sedimentary Basin. The southern Alberta plains lie on a northerly dipping anticline (Fig. 3) known as the Sweet Grass Arch (Westgate, 1968). The preglacial landscape was dominated by rivers flowing in large valleys that drained to the north and northeast. The pre-glacial valleys were numerous and located throughout southern Alberta. They have been infilled with sediments ranging from late Tertiary/Early Quaternary (Empress Group) to Wisconsinan time

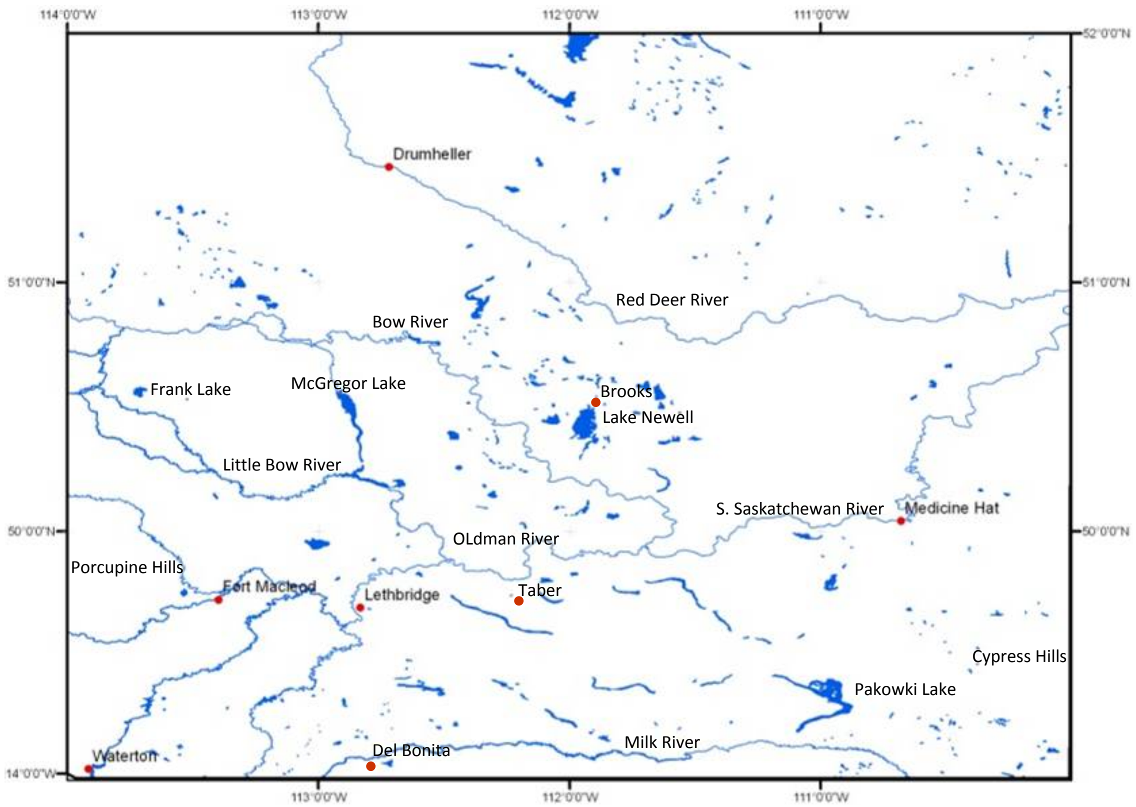


Figure 4: Location map of the study area in southern Alberta showing the main towns, rivers and lakes.

(Evans and Campbell, 1995). The interior plains of southern and central Alberta are composed of Upper Cretaceous and Tertiary sediments, which consist of poorly consolidated clay, silt and sand originating from the underlying bedrock (Stalker, 1960; Klassen, 1989). Importantly, significant quantities of bentonite are common throughout the study area, creating very poor drainage conditions (Klassen, 1989; Beaty, 1990), and exerting a strong control on till rheology when wet (Stalker, 1960). The Cypress Hills and Del Bonita Highlands remained as nunataks during the Quaternary (Klassen, 1989).

1.9. Summary

As discussed throughout this chapter ice streams are a very important and highly dynamic characteristic of both contemporary and palaeo-ice sheets. However, the extent to which terrestrial ice streams impact upon ice sheet dynamics is not fully understood. Moreover, the lack of any contemporary examples hinders our ability to fully elucidate and reconstruct their behaviour. As such, this research aims to accurately constrain the landform record of southern Alberta where terrestrial ice streams are reported to have existed during the late Wisconsinan. In particular, the focus of this research is to document the landform sediment assemblages at the margin of a terrestrial ice stream with a view to reconstructing palaeo-ice stream dynamics and elucidating any controls an ice stream margin may have upon its longevity and dynamics.

Chapter 2

Methods

This chapter reviews the methods used to achieve the aims of this research and is subdivided into three sections. The first section reviews the scale dependent model which has been implemented to give this research a logical and holistic structure. The second section explains the three mapping techniques adopted, which include SRTM, Landsat ETM+ and Aerial Photo mapping. The third section describes the sedimentary analyses used and their benefits in helping understand landform genesis.

2.1. Scale approach

A 'scale approach' is used to provide a framework for investigating glacial geomorphology on multiple scales. This notion acts as a structure to the research and requires that the mapping methodologies implemented will start at a small regional scale and move up to large local scale mapping, increasing the scale of focus in the smallest possible incremental steps. This will encompass the greatest quantity of possible data and provide a logical undercurrent that will produce the best possible results. Therefore, using this model as a framework this research will map the regional glacial geomorphology of southern Alberta with SRTM data and then introduce Landsat ETM+ which is a more accurate data set to verify and add to mapping, followed by aerial photo investigation. The model's primary function is to provide a structure that will enable accurate data results from data sets that vary in scale.

2.2. Mapping

Three different data sources will be used within this research in order to produce precise regional and local scale glacial geomorphology maps of Southern Alberta: Shuttle Radar Topography Mission (SRTM), Landsat 7 Enhanced Thematic Mapper Plus (Landsat ETM+) and Aerial Photograph Investigation (API).

2.2.1. SRTM:

SRTM data have been employed to create digital elevation models (DEMs) of the landform record. DEMs are raster files that display topographic data through tonal changes, as each pixel represents an averaged vertical height for the area it covers. Through recent advances in the availability of national and global data sets, the ease with which they can now be created and manipulated, their higher resolution and coverage render them one of the most useful tools for glacial geomorphology investigations (Smith and Clark, 2005).

The SRTM was flown in February 2000, obtaining elevation data on a near global scale (60°N - 56°S) and generating the most complete high-resolution topographic database on earth (Ramirez, 2006). The SRTM used interferometric synthetic aperture radar (InSAR) to record topographic data; this technique uses dual radar antennas situated in separate locations that take images simultaneously (Rabus et al., 2003). The mission produced three data sets: 1 arc-second, 30 m resolution data set for the US only (SRTM1); 3 arc-seconds, 90 m resolution (SRTM3) and 30 arc-seconds, 1 km resolution (SRTM30) data set with near global coverage (NASA, 2001). All three data sets have an approximate swath width of 225 km and the data is provided in an '.hgt' file format, which is a 16-bit signed integer data in a simple binary raster (NASA, 2001). All elevation data is in metres and referenced to the World Geodetic System/ Earth Geopotential Model (WGS84/EGM96). The mission objectives required absolute and vertical accuracy of ±16 m for 90% of the data and ±6 m respectively, and horizontal accuracy of ±20 m (Smith et al., 2006). Both absolute and vertical accuracy requirements were met, with the horizontal accuracy exceeding the specification by almost an order of magnitude (Rabus et al., 2003).

Whilst other available topographic data sets exists (ASTER, CDED), SRTM3 was chosen, as it was considered the best available data set for regional mapping of southern Alberta and its use in glacial geomorphology requires further testing and appraisal. ASTER (Advanced Spaceborne Thermal Emission and Reflection Radiometer) provides data with higher or similar resolution (15-90 m) to SRTM; however, the data is only available on demand and requires payment. Due to the financial constraints of this research and methodological approach discussed below, ASTER data has been dismissed as a mapping technique. Canadian Digital Elevation Data (CDED) provides topographic information with a resolution varying between 0.75-3 arc seconds and is referenced to the National Topographic Data Base (NTDB) (Fig.5).

Only within the last five years has SRTM data been available for use in glacial geomorphology, and as yet it has rarely been employed. Several authors, Glasser and Jansson, (2005); Bolch et al., (2005) and Heyman et al., (2008) have used SRTM data for mapping geomorphology; however, each author used it in conjunction with another data set such as Landsat 7 ETM+ and ASTER. This suggests that by itself SRTM data is inadequate as a sole mapping technique, most likely due to the resolution. However, recently Ó Cofaigh et al. (in press) used solely SRTM data to map ice streams in Saskatchewan and Alberta, yielding excellent results. It is therefore important that as part of this research the use of SRTM data for mapping glacial geomorphology be critiqued with specific focus on landform identification. Smith et al. (2006) have suggested that spaceborne sensors such as SRTM are not sensitive enough to map detailed morphology. Similarly, Falorni et al. (2005) have commented on a link between high topography and vertical accuracy errors within SRTM data sets. The above evidence suggests that SRTM imagery will provide a good regional scale picture, yet where landforms exist at scales smaller than or approaching pixel resolution it is likely that these landforms will not be visible. This would provide a generalized, rather than a comprehensive illustrative map of the glacial geomorphology.

2.2.1.1. Method 1

It was necessary to convert the SRTM file format '.hgt', as it was only recognized by a few GIS packages. Two GIS packages were available to convert the files: Global Mapper™ and ENVI 4.3™. Global Mapper™ produced a smoothed, rendered pseudo-colour image (Fig.6) of the SRTM data that could be manipulated to accentuate features, produce 3D images and

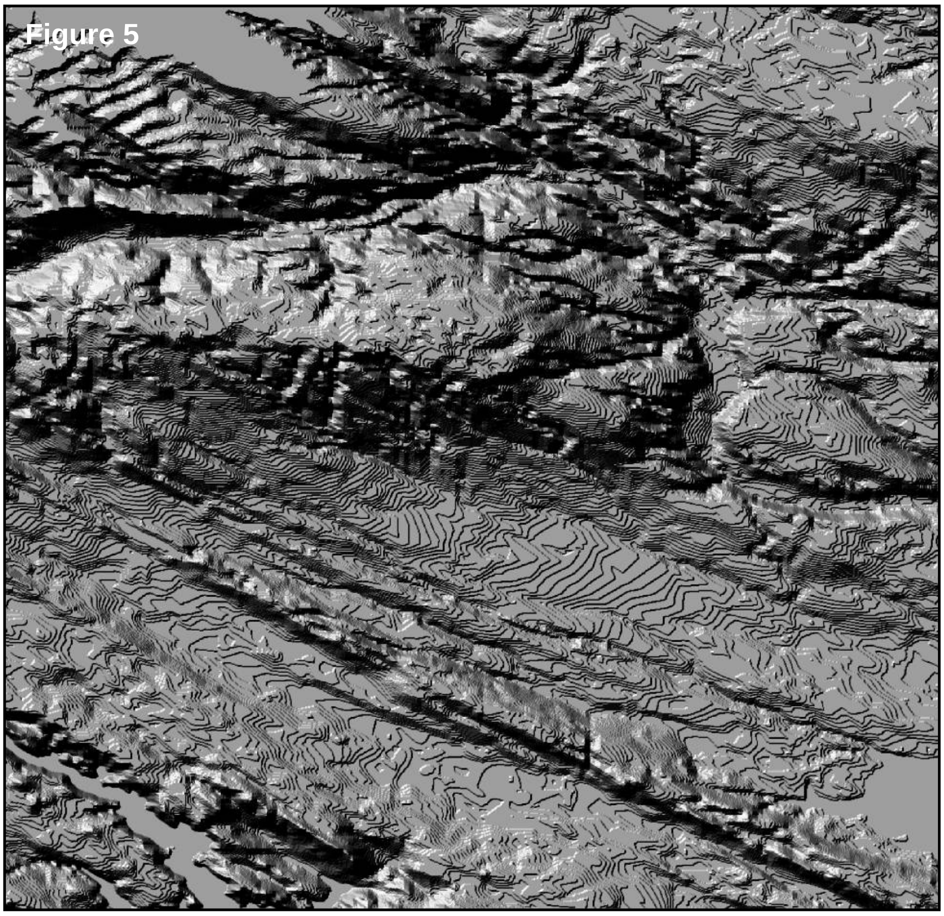


***Figure 5**: DEM created from CDED. The DEM is situated SE of Pakowki lake (Fig. 4), which is represented by the single grey tone in the NW corner of the image. The NTDB contour data easily visible throughout the image distracts from the landforms and so hinders the mapping and interpretation process.*

change sun illumination angles. By vertically stretching the elevation data, it is possible to more easily identify landforms within the data set, as they become more defined. Vertical stretching exaggerates the height of data above a set height, which is normally the lowest point in the data set. For example, a moraine ridge that is 10 m high could become 20 m high if the data set was stretched by a factor of two. However, this method can also start to misrepresent features due to morphology changes, and so one must be careful not to use this method as a sole means of mapping but rather to help interpret and distinguish landforms. In agreement with Smith and Clark (2005) multiple illumination angles were used to help map the landforms. Illuminating DEMs from an orientation at right angles to the main lineation direction provides excellent results (Smith and Clark, 2005). However, a single azimuth illumination, normal to lineation direction, highlights those features but weakens those of different orientation, creating an azimuth bias (Onorati et al., 1992; Smith and Clark, 2005). With respect to southern Alberta, where there are myriad landforms all of varying orientation and scale, it was necessary to implement illumination angles orthogonal to each other, and when mapping flowsets parallel and normal to the main lineation direction (Smith and Clark, 2005). Lastly, 3D images were created (Fig.7) to help verify mapped data, as they enabled

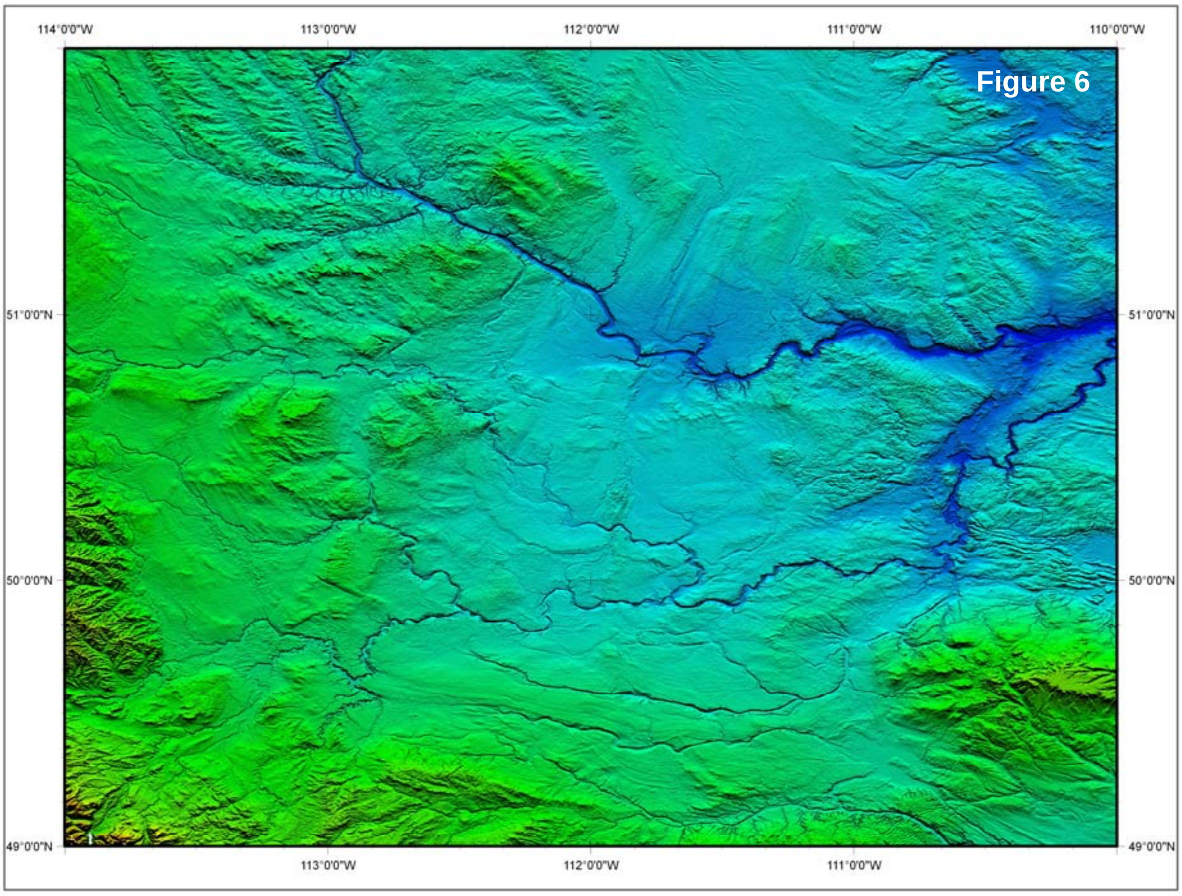


Figure 6: *Exported GeoTIFF of the fieldsite from Global Mapper. The pseudo-colour DEM is automatically created when opening the file; blue represents the lowest ground and darker green to brown the highest ground. Rendition is for illustration purposes only and no scale is given.*

multiple viewing angles of the same image.

The Global Mapper^{TM} interface did not provide the ability to easily map the glacial geomorphology and so these manipulations were completed in Global Mapper and then exported as a GeoTIFF. GeoTIFFs, however, provide only georeferenced raster imagery with no topographic information, hence the need to perform the DEM manipulations discussed above prior to their creation. These images were then opened in Erdas Imagine 9.0, a GIS package that enabled easy mapping of the glacial landforms. In order to map these features, vector layers were created and placed on top of the exported GeoTIFFs. These vector layers essentially act like tracing paper, allowing multiple layers to be used simultaneously, providing a very efficient method of mapping glacial geomorphology on a regional scale.

2.2.1.2. Method 2

An alternative method was employed to compare and verify the SRTM mapping and to add any additional landforms to the overall geomorphology map. ENVI 4.3 software was used to open the SRTM data, in a grey scale format (Fig. 8), and where data points were missing nearest neighbour sampling was used to correct for this. The nearest neighbour sampling was automatically applied to the same missing data points when opening the images in Global Mapper. The files were then exported from ENVI as Bitmap Graphic files '.img', which are simply raster files that can carry both georeferenced and topographic information. This option was not available when exporting out of Global Mapper. These images were opened in

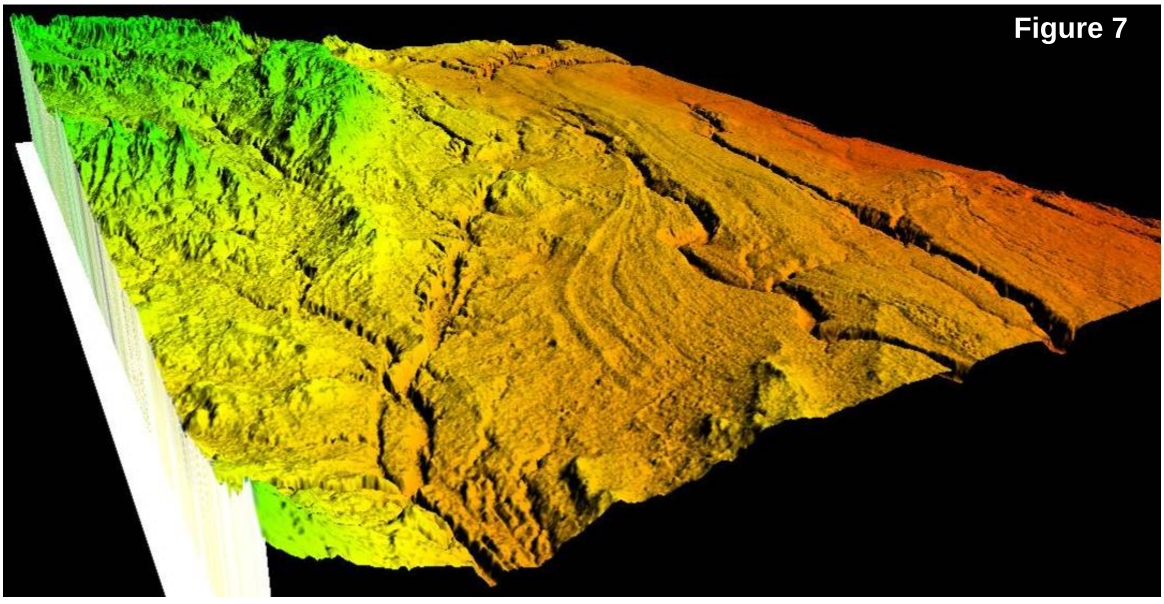


Figure 7*: 3D image created in Global Mapper. The 360° view provides a new perspective on landforms, but also highlights the differences in relief. This image was taken looking WNW from SE Alberta; the highland topography on the left is the Milk River Ridge. Rendition is for illustration purposes only and no scale is given*

Erdas Imagine 9.0 as relief shaded DEMs (Fig. 9). The DEMs were manipulated in exactly the same manner as above, with sun illumination changes, vertical exaggeration and 3D profiling. In similar fashion to the above method, vector files were overlaid on the DEMs to map the landforms. The results were then compared against the mapping performed from the GeoTIFFs. Each individual vector file represented a landform type and so the vector files from each method were compared to identify any discrepancies between the two methods.

The initial glacial geomorphology map was created through the above two methods, both having been subject to the same manipulation techniques, in order to create the best possible representation of the landform record in southern Alberta. Distinct landform assemblages will be apparent regardless of visual manipulations performed, yet depending on the glaciodynamic conditions responsible for their formation it is often the more subtle landforms that are of particular importance when interpreting landform assemblages (Smith and Clark, 2005). In order to fully elucidate the palaeo-ice dynamics operating in southern Alberta it is vital that landforms are identified at as many scales as possible and so the SRTM mapping will be used in conjunction with Landsat ETM+ and API. The finished images were rendered and exported from ArcMap™ Version 9.0.

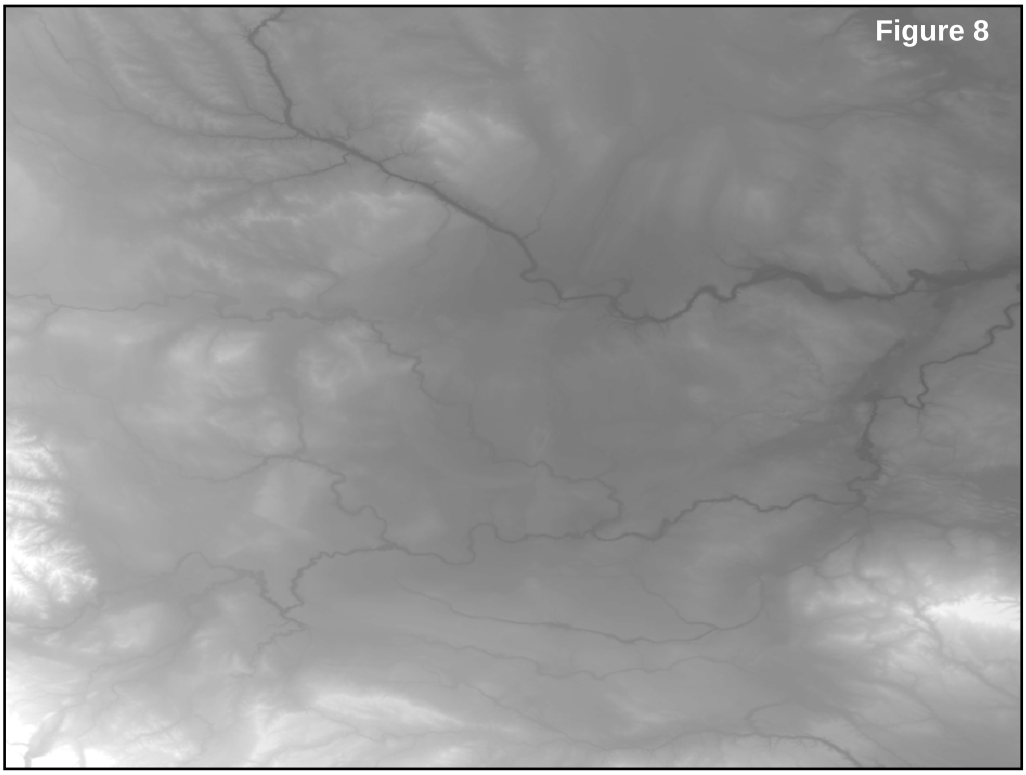


***Figure 8**: Grey scale image exported from ENVI 4.3. The '.img' file format carries topographic information, with white representing higher ground and dark grey the lowest ground. From this, relief shaded profiles were created. Rendition is for illustration purposes only and no scale is given.*

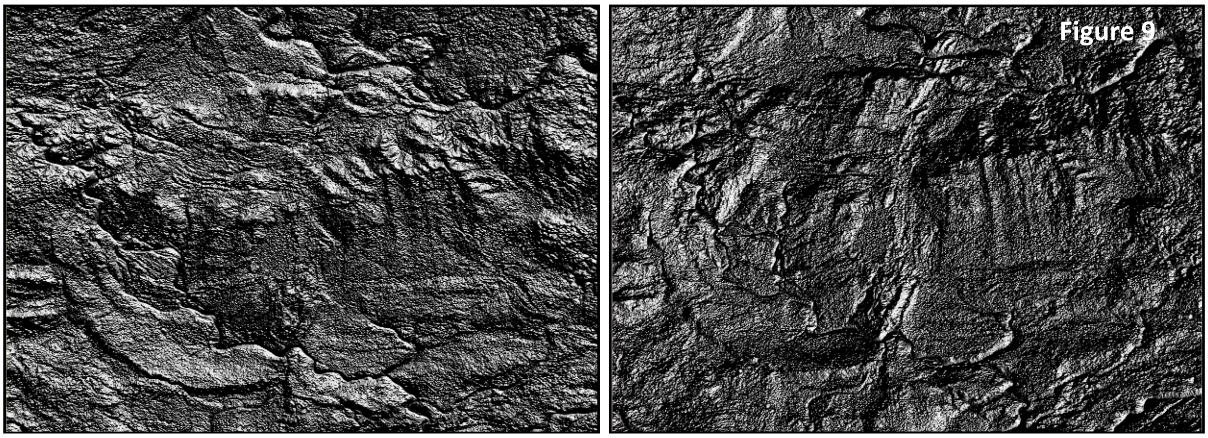


***Figure 9**: Relief shaded profiles illuminated perpendicular to each other. Note the distinct visual change in the imagery from a simple DEM manipulation. Rendition is for illustration purposes only and no scale is given.*

2.2.2. LANDSAT ETM+

Landsat ETM+ consists of 8 spectral bands. Bands 1-5 and 7 have a 30 m resolution and allow colour composites to be created, enabling different band combinations to aid image interpretation. Band 6 (thermal infared) has a resolution of 60 m. Band 8 (panchromatic) has a resolution of 15 m (USGS, 2006). Landsat ETM+ is a sensor that was carried onboard the Landsat 7 satellite. Each scene size is approximately 170 km north-south by 183 km east-west (USGS, 2006).

Additional geomorphological mapping was conducted through interpretation of the panchromatic band (band 8: 0.52-0.90 µm) due to its high resolution. A total of 13 scenes were mosaiced together providing full coverage of the field site (Table.1). These were downloaded from the GeoBase website (http://www.geobase.ca/geobase/en/index.html) overseen by the Canadian Council on Geomatics (CCOG). All images were in GeoTIFF format, and have been georeferenced with the North American datum of 1983 (NAD83) corresponding to the Universal Transverse Mercator (UTM) projection; UTM Zone 12 for Alberta.

Table.1

GeoBase ID	Type	Path	Row	Acquisition Date
043024, *Edition 1.02*	ETM+	043	024	2000/10/06
042024, Edition 1.01	ETM+	042	024	2002/08/18
042025, Edition 1.02	ETM+	042	025	2000/08/28
042026, Edition 1.01	ETM+	042	026	2001/08/15

041024, Edition 1.00	ETM+	041	024	1999/08/03
041025, Edition 1.00	ETM+	041	025	2001/07/07
041026, Edition 1.01	ETM+	041	026	2001/07/07
040024, Edition 1.00	ETM+	040	024	1999/07/27
040025, Edition 1.00	ETM+	040	025	1999/07/27
040026, Edition 1.01	ETM+	040	026	2000/07/29
039024, Edition 1.00	ETM+	039	024	2000/08/23
039025, Edition 1.00	ETM+	039	025	2000/07/22
039026, Edition 1.01	ETM+	039	026	2001/08/10

The same method was adopted for mapping the Landsat ETM+ images. The images were opened in Erdas Imagine 9.0 and overlaid with the same vector layers that were used to map the DEMs. This allowed first order verification of the SRTM interpretations and also additional features to be mapped with ease. Whilst the Landsat ETM+ and SRTM data sets are georeferenced to different co-ordinate systems no problems were encountered when overlaying vector files on the different raster formats.

The colour composite images created by alternating band combinations provided no advantage over the higher resolution panchromatic band. This is most likely due to the 30 m resolution and the extensive agriculture and subsequent modification of subtle landforms, requiring higher resolution imagery to identify such features. The 15 m resolution of the panchromatic band enabled more detailed mapping of smaller features but the lack of any topographic information hindered the identification of the larger, regional scale landforms. Band 8 was significantly more useful when identifying smaller lineations and any subtle differences within sections of hummocky terrain than SRTM.

Geographical Information Systems (GIS) provide a unique tool with which to assess and hence understand the relationships between individual glacial landforms, in particular specific landform assemblages and the insights into ice dynamics that can be gained from that (Napieralski et al., 2007). Clear advantages exist through using remote sensing for mapping glacial geomorphology, notably the ease and speed of mapping, the wide range of scales on which mapping can be performed, the availability of data, and the ability to integrate multi source data (Smith and Clark, 2005). Additionally, and more specific to this research it provides an excellent way to view large sets of landform assemblages, helping unravel any landform signatures on ice sheet scales (Clark, 1997). Landsat ETM+ has been used for many glacial geomorphology investigations (i.e. Glasser and Jansson, 2005, Heyman et al., 2008). In particular, it has been instrumental in palaeo-ice stream investigations (Stokes and Clark, 2003a; De Angelis and Klemen, 2005, 2008; Stokes et al., 2006). An assortment of other remotely sensed data have been used for investigating glacial geomorphology, such as Landsat Multi-Spectral Scanner (MSS) (Boulton and Clark, 1990; Clark, 1994; Stokes and Clark, 2003a), Landsat Thematic Mapper (TM) (Clark and Stokes, 2001; Stokes, 2002), SPOT High Resolution Visible (HRV) (Smith et al., 2000) and ERS Synthetic Aperture Radar (SAR) (Clark et al., 2000; Clark and Stokes, 2001). However, the higher resolution provided by the ETM+ data was considered the best possible data source for smaller scale landform mapping and independent verification on the accuracy of the SRTM data.

The combination of both these remote sensing methods should provide a comprehensive overview of the glacial geomorphology in southern Alberta, helping identify any individual or key suites of landforms that require further investigation using API, field checking and sedimentary analysis. On a regional scale the mapping will identify whether there are any cross-cutting lineations, different flow sets and the overall landform signature, aiding the reconstruction of palaeo-ice stream flow patterns and relative deglaciation pattern. Flowsets will be reconstructed in accordance with the criteria proposed by Clark (1999). Identifying flowsets provides an excellent method by which to understand any changes in ice stream flow, configuration and dynamics. Once identified, flowsets were mapped by drawing flowlines orientated parallel to the lineation direction. Where possible, quantitative analyses examined average lineation length, orientation, elongation ratios (ER) and average distance between lineations, in order to identify any similarities or differences between flowsets. Quantitative analyses into lineation morphology and pattern has been demonstrated to be useful in helping reconstruct palaeo-ice streams and their dynamics (Stokes and Clark, 2003a; Roberts and Long, 2005; Stokes et al., 2006; Storrar and Stokes, 2007).

Landforms were mapped using their morphometric characteristics and then interpreted. Whilst, a prior knowledge of the field site, landform morphology and type is essential before mapping any glacial geomorphology, it is vital that interpretations do not place any bias on the mapping process and the researcher stays as objective as possible (Hubbard and Glasser, 2005). Secondary data sources (Shetsen, 1987) were also used to help confirm or reject any contentious mapped landforms.

2.2.3. Aerial photo Investigation (API)

Aerial photos were used for large scale investigations into the marginal landform record of the CAIS (49°00’00”N, 49°42’06.44”N; -110°00’00”W, -113°14’47.21W). The remote sensing methods were inadequate for detailed investigation of the CAIS margin and so API and ground truthing were employed. Aerial photos provided by far the best available data source for investigating the marginal landforms of the CAIS because of the excellent resolution. A series of 10, 1:63,360 (1 inch to one mile) aerial photo mosaics captured in 1951 by the Alberta Department of Land and Forests were mapped by hand to create a detailed glacial geomorphology map. The marginal area of the CAIS was chosen because of the quality of the landform record.

The landforms were mapped using the same method as above, firstly based on their morphometric characteristics and then interpreted to determine whether they were of glacial origin. Moraines and thrust ridges were represented by a single line along their crest; lineations were similarly represented by a single line orientated along their main axis. Where possible, individual hummocks were outlined, however, due to the complexity and sporadic orientation of some sections of the hummocky terrain this was not possible. In these sections the hummocky terrain has been represented by shading the inter hummock depressions. Any hummocks that showed strong linearity were represented by a single line along that orientation. Distinct differences existed between types of hummocky terrain but these were not illustrated on the map due to scale; however these differences are comprehensively covered in the next chapter.

Once completed the maps were scanned into digital form and made into a TIF file. Individual images were mosaiced in Adobe Photoshop®, creating a map of the marginal setting of the CAIS. Specific landforms and suites of landforms were identified through API and further investigated through ground truthing and where possible sedimentary analysis.

2.3. Sedimentary Analysis

Key landforms and suites of landforms identified as having particular importance in unraveling the complex glacial signatures and assemblages were identified from the API. Where possible, stratigraphic logs, cross sections and clast analyses were used to assess the sedimentary makeup of the landforms. Stratigraphic logs and cross sections are an essential technique in attempting to reconstruct glacial dynamics (Evans et al., 2007). In particular, vertical and lateral variations in sedimentary structures and architecture, bed contacts, sorting and texture were examined at sections (Evans and Ó Cofaigh, 2003). Large sequences often show distinct lithostratigraphic units relating to advances and retreat of the margin, occasionally allowing relative dating of the stratigraphy (Westgate, 1968). Clast macrofabrics were taken (imbrication of the A/B plane; Evans and Benn, 2004) where possible to provide information on the directions of ice movement, allow differentiations between styles of deposition and different sedimentary facies, and to provide indications of the relative strain signature in a sediment (Hubbard and Glasser, 2005). A minimum target of 50 clasts was chosen when taking clast data in order to make the results statistically significant (Bridgland, 1986). However, this was not always possible at all sites. The stratigraphic and sedimentological data will allow constructional mechanisms to be accurately identified, helping to assess the different theories of landform construction under palaeo-ice stream margins. Stratigraphic logs and cross sections are used as the primary data sources for any sedimentary analysis and when available were supported by clast macrofabrics.

2.4. Summary

The sequence of methods adopted here have deliberately followed a scale approach that employs a small to large scale structure in order to provide a holistic and logical basis for this scientific research. This in turn will help provide comprehensive data collection of the landform record, which will enable a thorough assessment of the glacial dynamics that acted in their formation. These specific methods have been employed to create:

- A regional glacial geomorphology map of southern Alberta.
- Palaeo-flow set map, indicating flow orientation and dynamism of ice streaming within southern Alberta.
- A detailed glacial geomorphology map of the marginal setting of the CAIS from aerial photo investigation.
- Sedimentary logs, cross-sections and stereonets used to help reconstruct ice marginal dynamics at the time of landform genesis.

Chapter 3

Results

This chapter is divided into three sections: regional and ice marginal glacial geomorphology, and glacial sedimentology. The first section presents the results obtained from SRTM and landsat ETM+ mapping; Section 2, glacial geomorphology of the CAIS marginal area and Section 3, the glacial sedimentology of the CAIS marginal area. Whilst the marginal setting of the CAIS is the primary focus of this research, by first assessing the regional geomorphology it is possible to place the results within a local and regional context.

3.1. Regional results

The field area covered approximately 99,000 km^2 and demonstrates a wealth of landforms (Fig.10). The glacial geomorphology is primarily dominated by two palaeo-ice streams tracks, the HPIS and CAIS (Evans et al., 2008), and inter-lobate (inter stream) hummocky terrain. Additionally, along the eastern margin of Alberta, Efs_1, also known as 'Ice Stream 1' (Ó Cofaigh et al., In Press) or the east lobe (Shetsen, 1984; Evans, 2000), terminates to the north of the Cypress Hills (Fig. 10). The ice stream corridors are identified by smoothed topography, multiple flow sets (Fig. 12), MSGLs, and extensive esker and metltwater networks. In addition, both the HPIS and CAIS display narrower trunks of fast flow and lobate margins, in accordance with the palaeo-ice stream geomorphological criteria set out by Stokes and Clark (1999, 2001). Figure 10 distinguishes between different types of transverse ridge sets but the reasons for these interpretations are discussed in the Chapter 4.

3.1.1. High Plains Ice Stream

The HPIS track extends into the study area in a south, south-easterly direction, turning to the east and south-east as highlighted by Figure 12. The HPIS is approximately 250 km in length (within the study area) and its width varies from around 50 km along the main trunk to 85 km across the lobate terminus. The HPIS is bounded by hummocky terrain to the east and the foothills of the Rocky Mountains to the west. The HPIS was located roughly 250 m higher than the CAIS (Fig. 13b) and traversed across a west to east dipping slope (Fig. 13b, c). The main landforms are lineations situated primarily to the north of the study area, with eskers and esker networks flowing parallel to lineation direction. Reconstructing lineations has identified at least seven (Hfs_1-7) different flow sets along the HPIS trunk, and to the west five flow sets have been identified as being sourced from the Rocky Mountains (Cfs_1-5). Anastamosing meltwater channels are juxtaposed with a large sequence of transverse ridges that lie along the western edge of the ice stream. As mentioned above the HPIS splays out into a lobate form along its southern margin, terminating at the western edge of the CAIS. The marginal deposits of the HPIS are composed of hummocky terrain bands west of Del Bonita and the McGregor Moraine (Evans et al., 2006a).

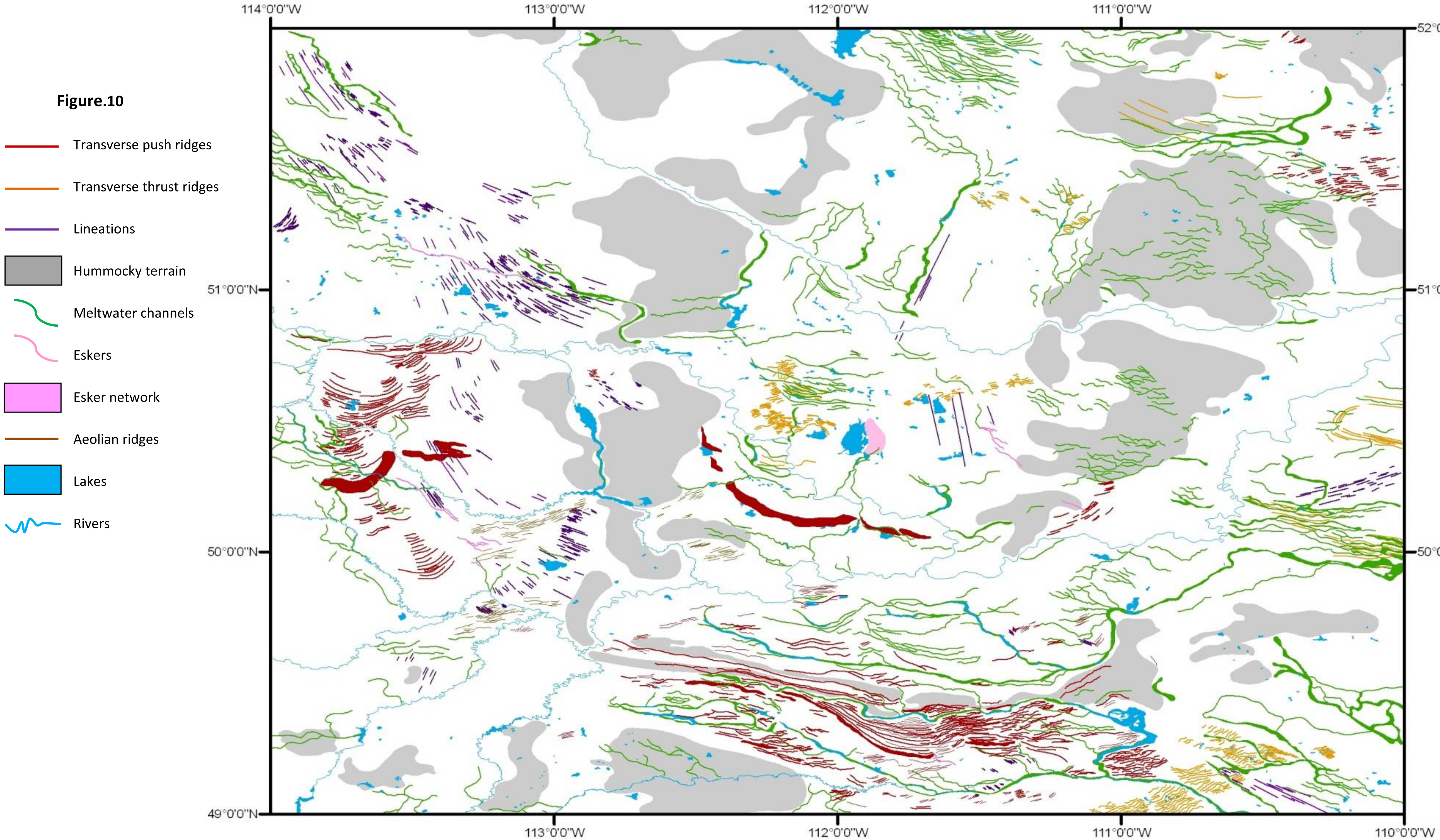


Figure 10*: Glacial geomorphology map of southern Alberta.*

Figure 11

BR - Blackspring Ridge

CH - Cypress Hills

Lb - Lethbridge

MH - Medicine Hat

ML - McGregor Lake

FL - Frank Lake

MRR - Milk River Ridge

BI - Bow Island

WI - Wolf Island

CC - Cherry Coulée

Crevasse-squeeze ridges, Set 1

Crevasse-squeeze ridges, Set 2

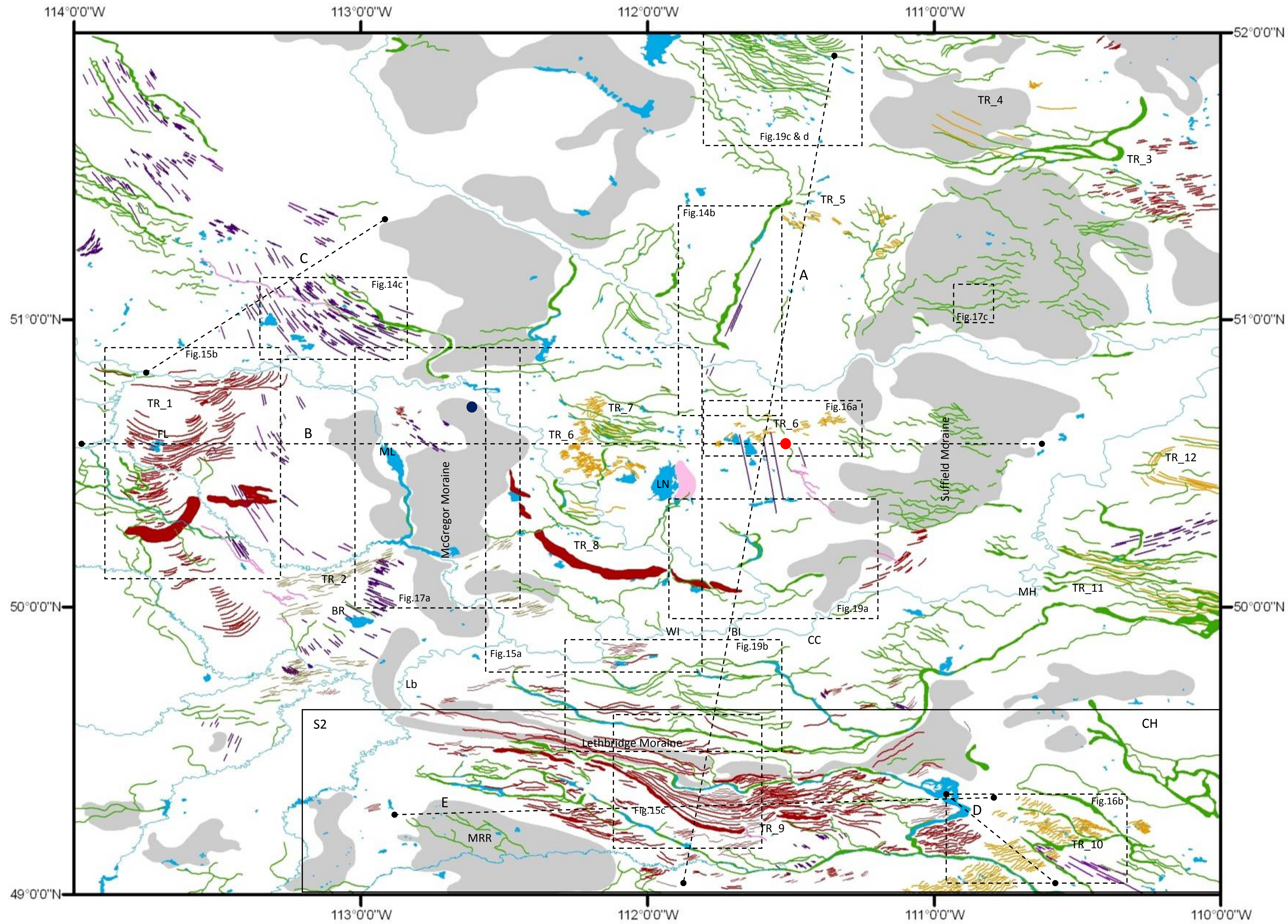


***Figure 11**: Locations of Figures 13-19 and transverse ridge sets within the field area.*

Figure 12

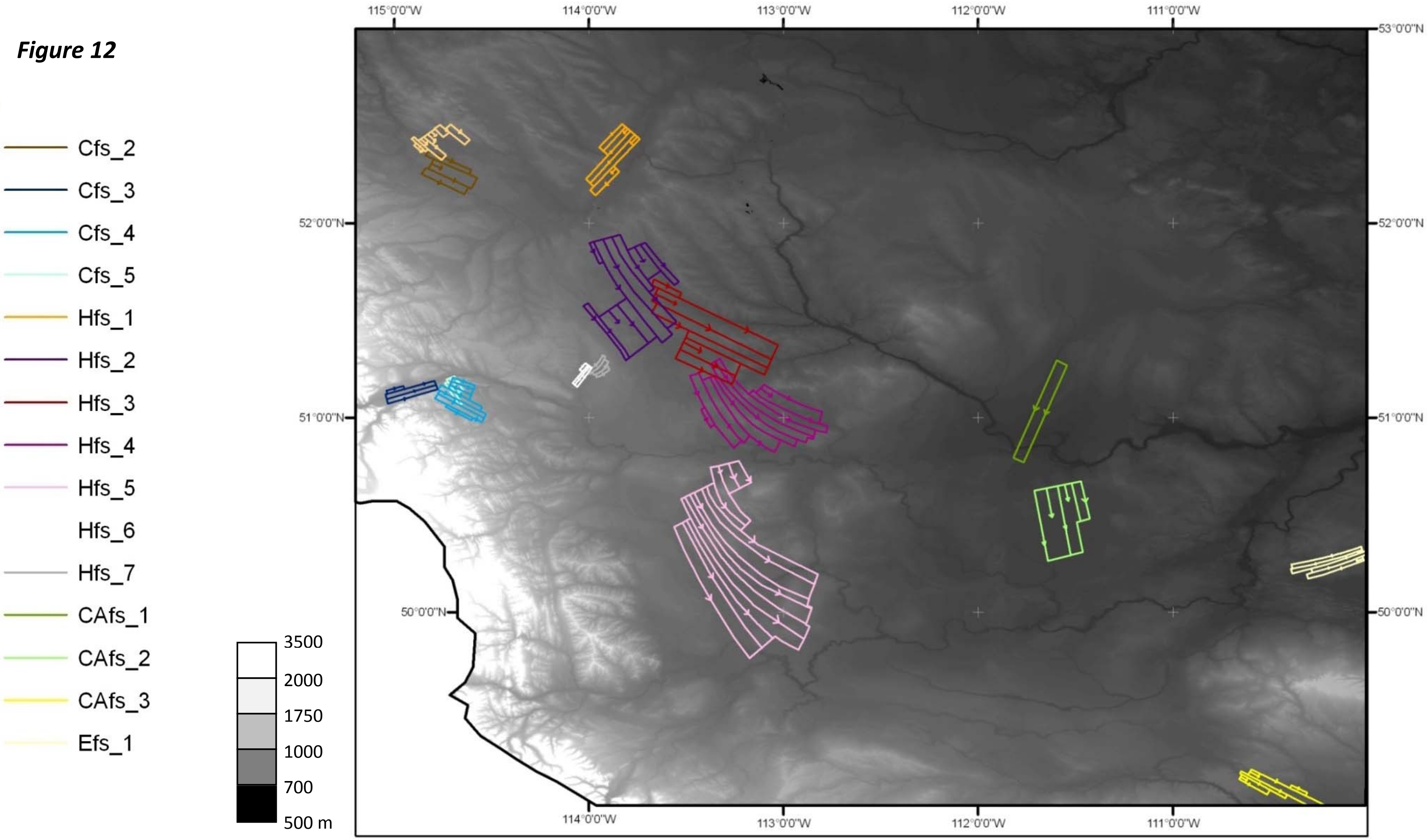


Figure 12*: Flow-sets reconstructed from glacial lineations. Lineations were grouped into flow sets based primarily on their orientation but also their proximity and location (Clark, 1999). Cfs indicates a Cordilleran origin; Hfs are found along the High Plains Ice Stream; CAfs along the Central Alberta Ice Stream and Efs along the east lobe.*

3.1.2. Central Alberta Ice Stream

The CAIS track extends through the field area in a southerly direction, orientated perpendicular to Hfs_3 and 4. The CAIS is roughly 320 km in length (within the study area) and becomes progressively wider down ice: 97 km across northern trunk, widening to 160 km across the margin as the terminus spreads out into a large lobate margin (Fig. 10). The CAIS advanced against an inclined proglacial slope (Fig. 13A), in particular around the margin. The ice stream is easily identifiable as it is bound to the west, east and partly to the south by hummocky terrain; also known as the McGregor, Suffield and Lethbridge moraines respectively (Fig. 11) (Shetsen, 1984; Evans, 2000). Additionally, mapping by Evans et al. (2008) identified a lateral moraine along the western margin of the CAIS. Two flow sets were identified along the CAIS (CAfs_1 and 2), and one in the south east corner CAfs_3, recording at least two different phases of flow. The margin is composed of a series of sub-parallel transverse ridges, contiguous to hummocky terrain sections that in plan form run parallel to the ridges. Meltwater channels and couleés present in this area also appear lobate in form. Moving northwards the meltwater channels continue to trend parallel to the marginal landforms. In addition, a large sequence of anastamosing meltwater channels occurs at the northern most tip of the CAIS. They trend in an east, south east direction following the regional slope. Transverse ridges are found primarily in the CAIS marginal area but are also found along the ice stream trunk, and in the north east corner and east of the study area. Eskers and esker networks occur along the centre and eastern edge of the ice stream trunk.

3.1.3. Flow-sets and lineations

Overall 653 lineations were identified along the CAIS and HPIS and combined they produced 16 individual flow-sets (Fig. 12). Hfs_4 contained the largest amount of lineations (130) showing strong spatial coherency, and CAfs_1 contained the largest lineation (35km long). Due to the resolution of SRTM imagery no elongation ratios could be taken, however, it is highly likely that some if not all lineations had an ER of greater than 10:1. The smallest examples were found in Hfs_1 and 7 and Cfs_3, and the largest in CAfs_1 and 2 (See Table 2 for flow set data). Hfs_4, 5 and CAfs_2 terminate in hummocky terrain zones, whereas Efs_1 marginal zone is composed of large transverse ridges. CAfs_1 lies between two sets of multiple transverse ridges and 70 km up ice of a large perpendicular lobate transverse ridge. In similar fashion CAfs_3 is located on the down ice side of a sequence of transverse ridges and the Blackspring Ridge lineations occur on the down ice side of an escarpment.

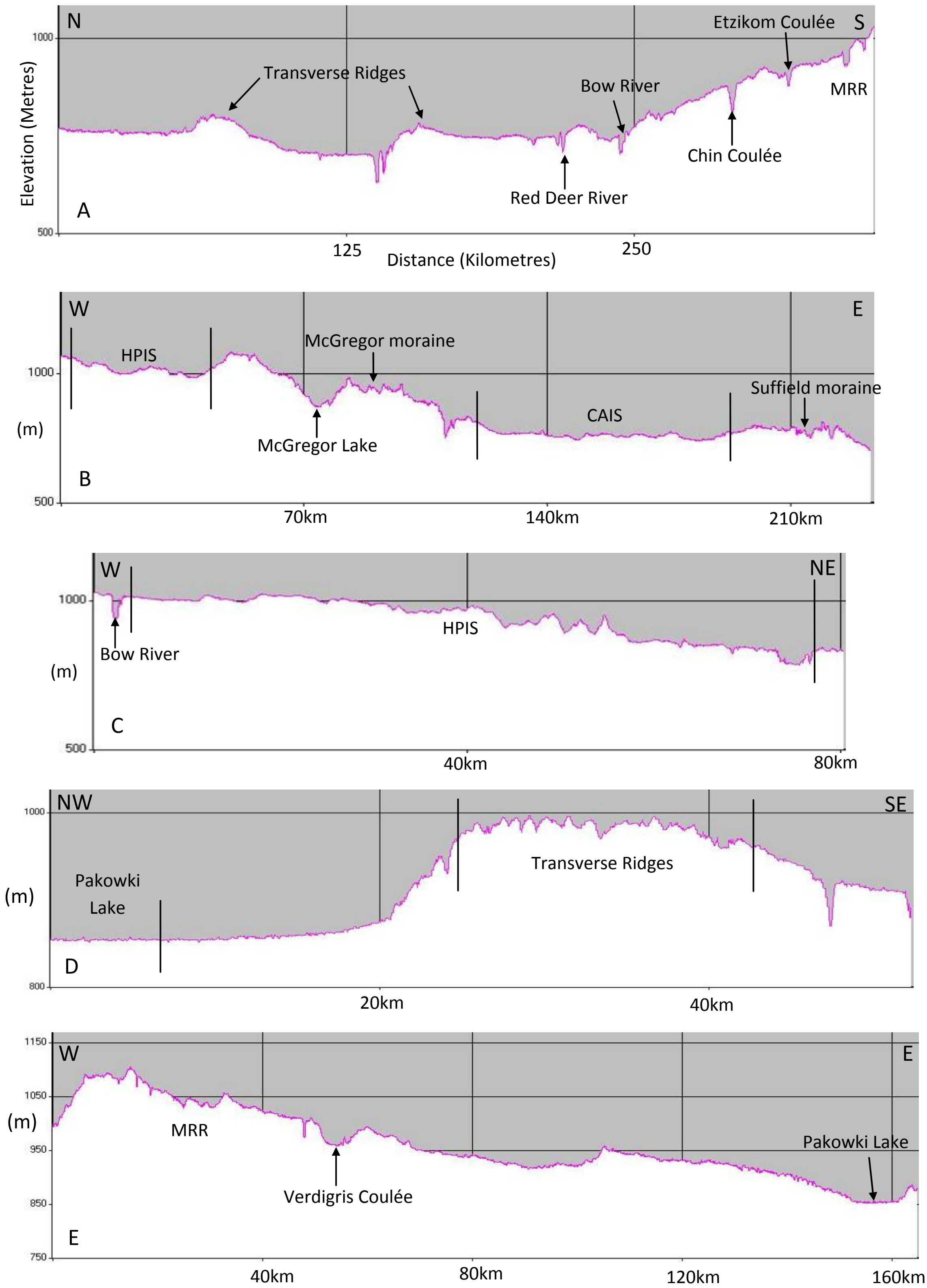


Figure 13: Profiles taken from SRTM data (See Figure 11 for location). A) long profile of the CAIS. B) is a cross section of the study area. C) shows the slope which the HPIS traversed across. D) shows the profile from Pakowki Lake across the transverse ridges located on the preglacial drainage divide in the SE corner of the study area. E) is a cross section of the CAIS marginal area.

Flow-set	No. Lineam-ents	Mean Length (km)	Mean Direction (°)	Flow-set Area (km^2)	Mean Transverse Wavelength (km)
Cfs_1	11	2.60	128	289.53	1.64
Cfs_2	8	3.60	114	276.09	1.62
Cfs_3	32	1.21	74	133.18	0.22
Cfs_4	55	2.32	117	373.85	0.31
Cfs_5	13	3.37	146	73.74	0.37
Hfs_1	61	1.56	224	501.16	0.24
Hfs_2	81	3.42	141	2162.29	0.44
Hfs_3	66	2.34	119	1631.27	0.46
Hfs_4	130	3.58	127	2075.77	0.27
Hfs_5	99	3.52	129	4473.44	0.52
Hfs_6	11	2.01	222	42.53	0.33
Hfs_7	32	1.99	227	83.72	0.19
CAfs_1	6	10.04	203	262.27	0.76
CAfs_2	4	20.89	168	577.92	4.17
CAfs_3	20	4.17	118	424.56	0.52
Efs_1	28	2.93	252	251.56	0.31

Table 2: *Data showing the specific characteristics of the flow-sets, which in turn act as a device to help differentiate between particular flow sets.*

3.1.4. Transverse ridges

Large sequences of transverse ridges exist throughout the study area, not only in marginal settings but along the HPIS and CAIS flow corridors. The ridges demonstrate a wide range of distinctly different morphological characteristics.

A large sequence of transverse ridges stretches approximately 100 km along the western half of the HPIS (TR_1). The ridges are low amplitude, generally only a few metres in height, but can reach up to around 6 m. Within the sequence are two large ridges, superimposed on which are four ridges roughly 10-15 m in height (Fig. 15b). The two ridges are dissected by a branch of the Bow River, and the eastern ridge is overprinted by lineations from Hfs_5 (Fig. 15b). Cross-cutting the southern section of Hfs_5 and along the western edge of the CAIS is a sequence of densely spaced thin, sharp crested transverse ridges (TR_2). They range from hundreds of metres to a few km in length and reach up to around 25 m in height.

The north eastern corner of the study area is composed of a sequence of inset ridges that are bordered by hummocky terrain to the east, west and south (TR_3). The sequence of extends for 30 km with crest wavelength varying between 500 to 1000 m. The ridges range

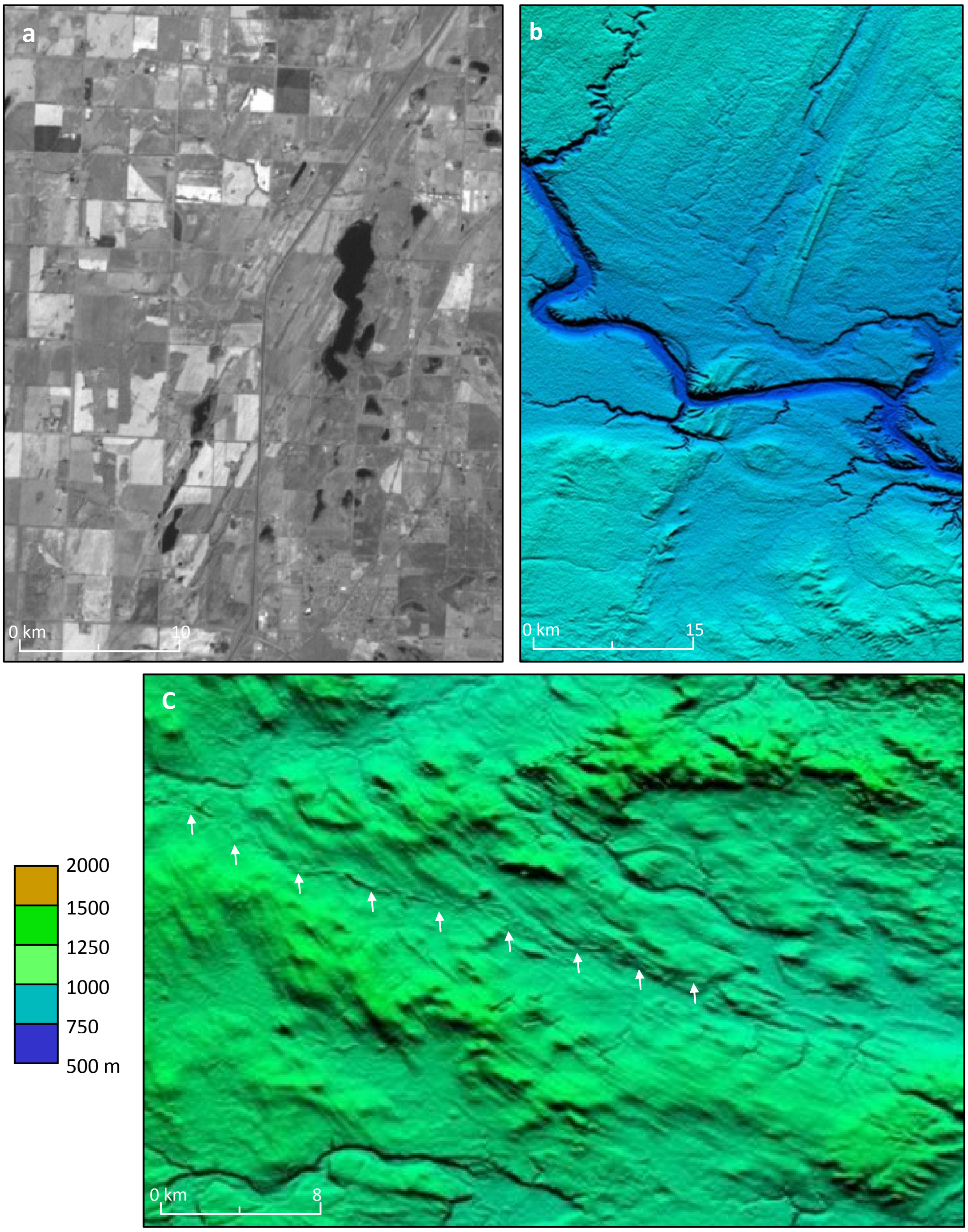


Figure 14: a) Landsat ETM+ image of Hfs_1; b) CAfs_1 from SRTM data in GeoTIFF format; c) Hfs_4 from SRTM data in GeoTIFF format. c) demonstrates the high level of spatial coherency of a single flow set and the white arrows highlight the large esker that runs through the flow set.

from roughly 10-20 m in height and are generally a few km in length. Situated to the north west of TR_3 are several large ridges that dominate an area of hummocky terrain (TR_4). The ridge crests run for 10 km and stand up to 20 m above the surrounding terrain. To the north of CAfs_1 lies a sequence of ridges orientated at 45° to lineation direction (TR_5). The sequence is composed of three parallel subsets of ridges spaced 4.5 km either side of the central set. All the ridges are similar in morphology, demonstrating a distinct crenulate plan view. The sequence is composed of 40 ridges, ranging from 1-4 km in length. The subsets rise up to 30 m above the surrounding land on which each ridge crest is about 5 m high. 20 km to the south of CAfs_1 lie two sets of ridges (Fig. 15a & 16a): a large series of ridges that are identical in morphology to those situated just to the north of CAfs_1 (TR_6); and a small set of larger ridges that have been accentuated by meltwater (TR_7). TR_6 is composed of over one hundred ridges that run east to west across the length of the CAIS. These ridge crests demonstrate a strong orientation from north east to south west and lie perpendicular to CAfs_2. From west to east the ridges and their crest wavelengths become gradually larger, ranging from a couple of metres up to 5 m in height. The ridges within the second set are characterised by longer, smoother crests of 8 km in length and are up to 5 m in height. Meltwater channels run in and behind them accentuating their overall size and the ridges are orientated perpendicular to CAfs_1 and a large lobate ridge 20 km to the south (TR_8). The large lobate ridge is situated across the majority of the CAIS between the Bow and Oldman Rivers (51°N and 112°W). The eastern edge of the ridge trends into an area of hummocky terrain (10 km east) that is orientated parallel to the lobe. The ridge is dissected by meltwater channels at several locations but overall extends for about 70 km. The ridge is asymmetric, with a steeper proximal slope and its height gradually increases from west to east (20 – 30 m).

At the southern margin of the CAIS lies a large sequence of transverse ridges (TR_9), north of which lies the Lethbridge Moraine (Fig. 11) (Stalker, 1977; Evans; 2000). This sequence comprises part of the CAIS ice marginal landscape and provides the focus for Section 3.2. The sequence is approximately 45 km in depth, 150 km in length and composed of numerous long, low amplitude ridges that are aligned en echelon to each other and are lobate in shape. The largest ridges extend for the length of the sequence and are no more than a few meters in height. Also within this sequence are numerous meltwater channels and coulées that run parallel to the ridges. 15 km to the south east of Pakowki Lake lies a sequence of 170 sub-parallel ridges (TR_10), just up-ice of, and perpendicular to CAfs_3 (Fig. 16b). The ridges lie on the preglacial drainage divide that was located between the Cypress and Sweet Grass Hills (Westgate, 1968) and is about 150 m higher than the Pakowki Lake depression (Fig.13D). The ridges reach 20 m in height to the west CAfs_3 and descend to roughly 5 m to the east. Ridge wavelength also decreases from west to east, varying from 1 km to 250 m. Two further sets of transverse ridges are located to the north and south of Efs_1. The southern set (TR_11) is orientated at 45° to, and located at the terminus of Efs_1. The sequence is composed of large parallel ridges up to 30 m in height, which are accentuated by major meltwater channels that surround them. The TR_12 features are predominantly parallel to those of TR_11 but also show a 90° change in orientation at the western edge. The sequence is located at the south west corner of 'Ice Stream 1' identified by Ó Cofaigh et al. (in press). The ridges are predominantly 5 m in height but two larger ridges located up-ice are almost 30 m in height.

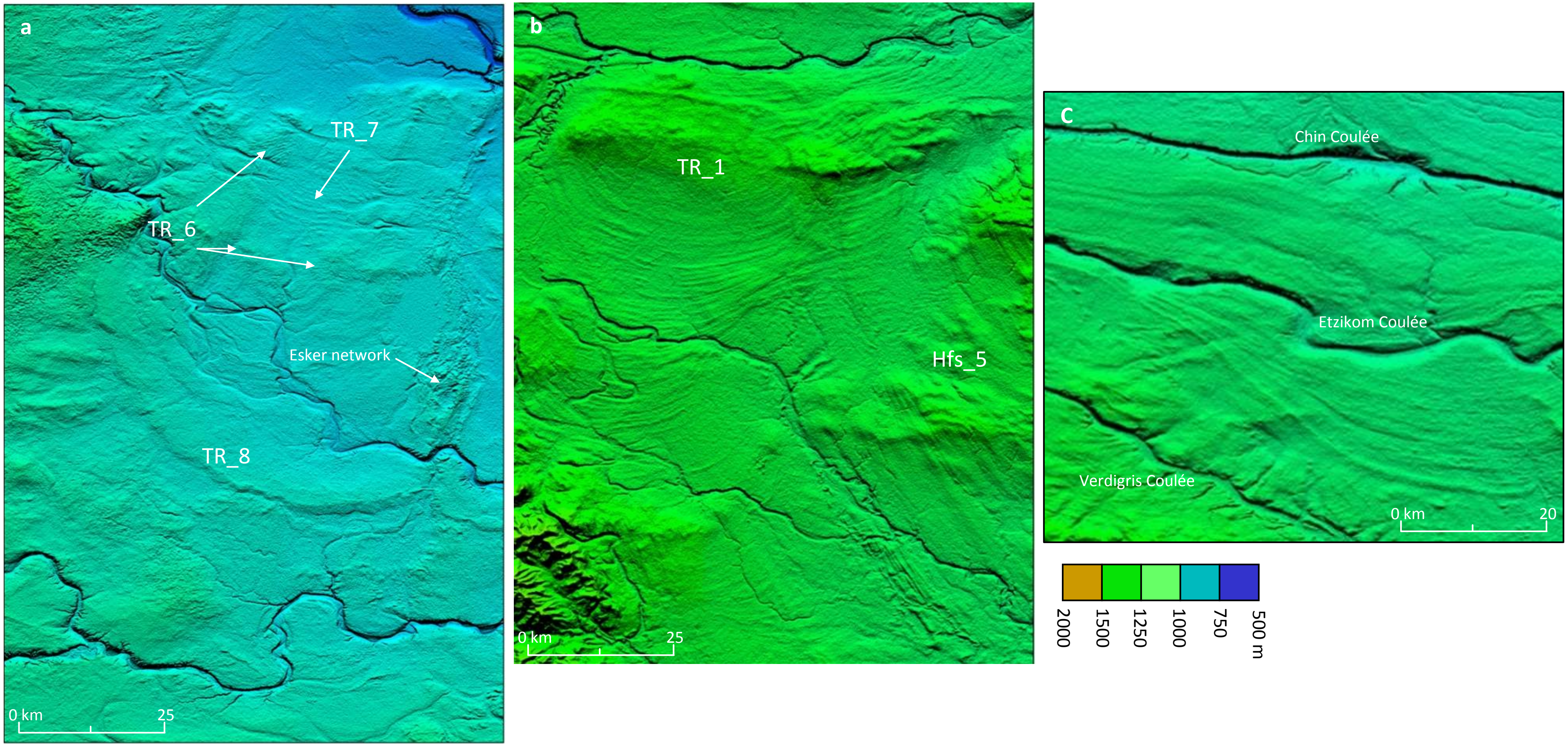


Figure 15*: Examples taken from SRTM data of transverse ridges situated in southern Alberta. a) large lobate ridge situated along the CAIS. The Bow River flows through the centre of the image and the Oldman River along the bottom. Also in a) are TR_6, TR_ and TR_8, and an esker network situated to the right centre of the image. b) displays the long set of ridges situated along the HPIS trunk (TR_1). Note the streamlined features that make up Hfs_5 to the right of the image. c) shows the sequence of long, low amplitude ridges that form the lobate plan pattern of the CAIS margin (TR_9). Also visible are the large Chin, Etzikom and Verdigris Coulées that run parallel to the ridges.*

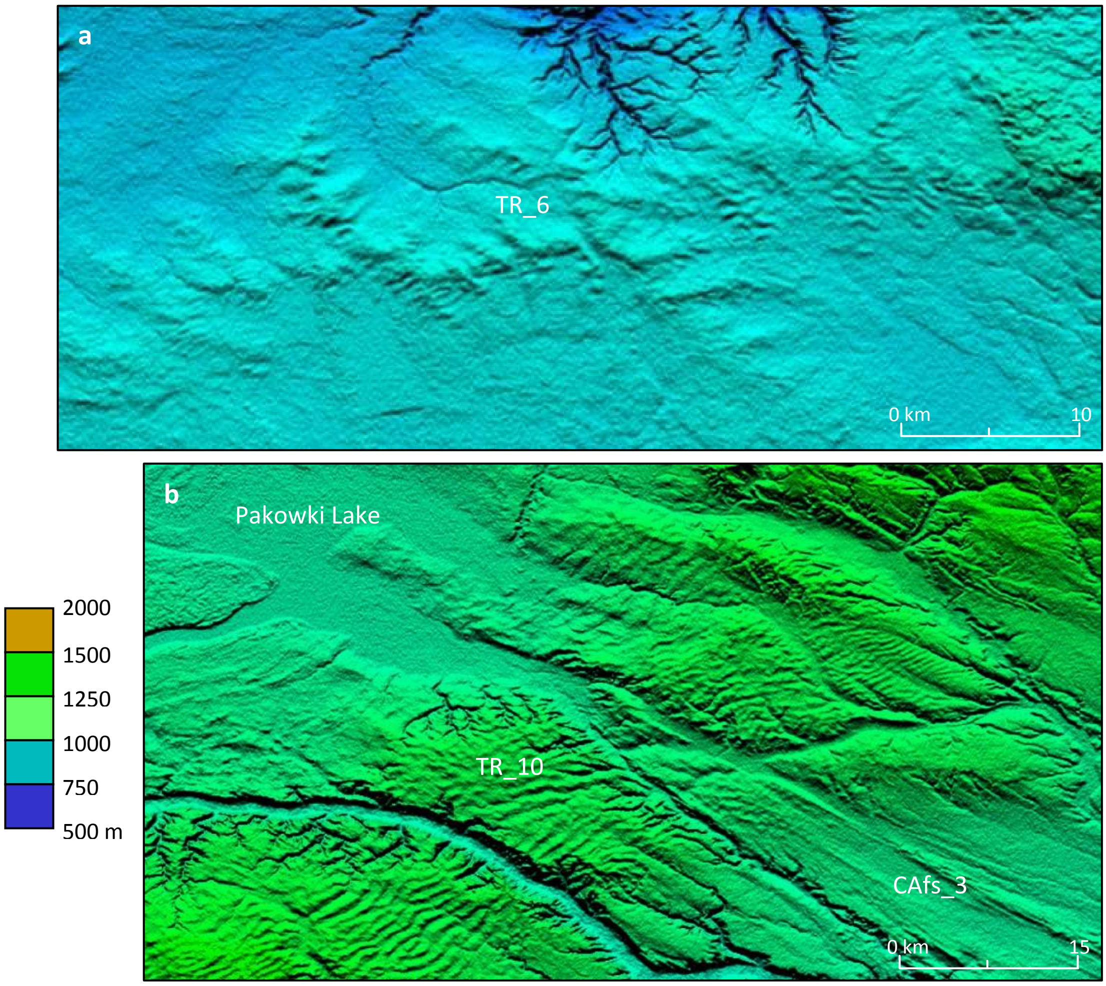


Figure 16: *a) western section of transverse ridges that cross the entire CAIS (TR_6, Fig. 11). b) shows the sequence of ridges in the south eastern corner of Alberta (TR_10). Note the lineations situated just down ice of the ridges (CAfs_3) and the* smooth flat topography in the north west corner represents Pakowki Lake.

3.1.5. Hummocky terrain

Hummocky terrain comprises a large proportion of the study area and is found primarily in between ice stream corridors but also along the margin of the CAIS (Fig. 10 & Fig.17). The areas identified demarcate the limits of smoothed topography characteristic of the ice stream corridors. The resolution of the SRTM and to an extent the Landsat ETM+ data is insufficient to make significant insights into the morphology of the hummocky terrain. Rather, it has been possible to identify general areas of hummocky topography. The hummocky terrain areas mapped show strong similarity in the distribution of hummocky terrain to that of Prest et al. (1968); Shetsen (1987, 1990); Clark et al. (1996) and Evans (2000).

Significantly, a hummocky terrain band (Lethbridge Moraine) runs continuously from the edge of the Blackspring Ridge across the CAIS marginal area up to and around the Cypress Hills. In plan form the band demonstrates a strong lobate pattern that runs parallel to the transverse ridges but internally consists of chaotic hummocks. Additionally, within the areas of

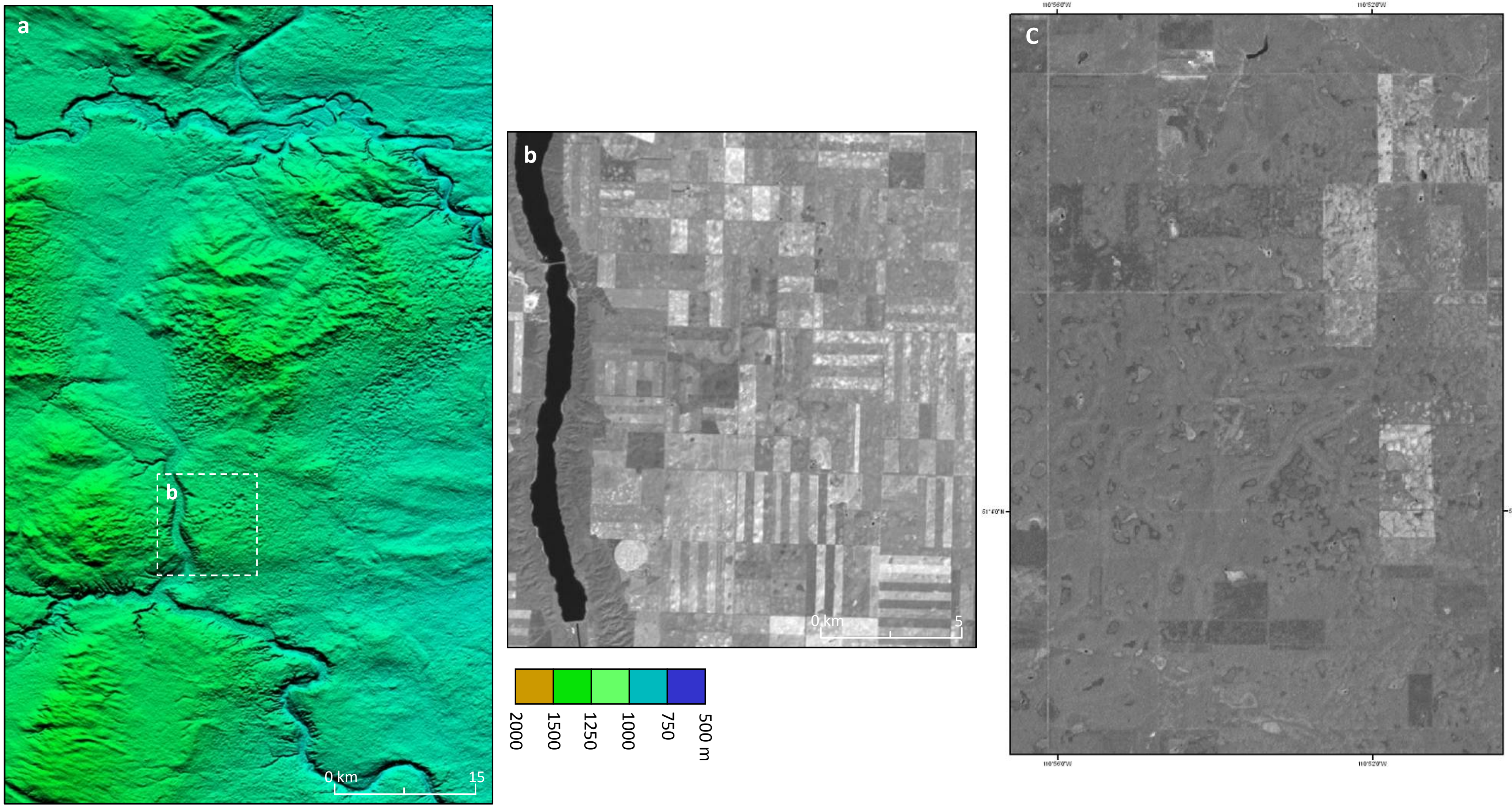


Figure 17: *Examples of hummocky terrain. a) hummocky terrain around McGregor Lake, which is shown by the flat, smooth area in the left centre. b) larger scale image of the hummocky terrain at the south east of McGregor Lake. The image is taken from Landsat ETM+ data. c) chaotic pattern of ridge like structures that lies within hummocky terrain in the north east of the study area (Fig. 11). The ridges are significantly larger than the surrounding hummocks and are the best developed example of this type of feature identified within the field area.*

hummocky terrain, several examples of conspicuous chaotic ridge forms were present. The best developed example was located at 110°54’0”W, 51°4’0”N (Fig. 17c). The ridges have no distinct crest and are symmetrical in shape; they are significantly larger than the surrounding hummocks with individual ridges reaching up to 4 km in length.

3.1.6. Crevasse squeeze ridges

Crevasse squeeze ridges have been located at two locations within the field site (Fig. 11 & 18). Set 1 are located 15 km to the north east of McGregor Lake at 112°37’0”W, 50°40’0”N, within the McGregor moraine (Fig. 18a). The sequence is 1.5 km long and the ridges are thin, sharp crested and extend for up to 750 m in length. The second set was identified with Landsat ETM+ data and is located at 111°28’29”W, 50°32’38”N. The ridges are located within CAfs_2, approximately 10 km down ice of the large transverse ridge sequence (TR_6). The sequence stretches for just under 6 km, with ridge length varying between 500 m and 1500 m. Morphologically they are very similar to set 1, and are characterised by very thin, low amplitude, sharp crested ridges.

3.1.7. Esker networks

Eskers are prominent throughout the study area, but resolution constraints meant that only the largest features were identifiable. It was also common to find that large eskers in close proximity to each other (Figs. 13a, b) were part of a series of smaller features that could not be confidently distinguished. The largest esker identified was situated along Hfs_4 (Fig. 14c), reaching 45 km in length. The esker runs at 25° offset to the lineations and follows the natural slope. Further to the south a sequence of prominent eskers is situated along the centre

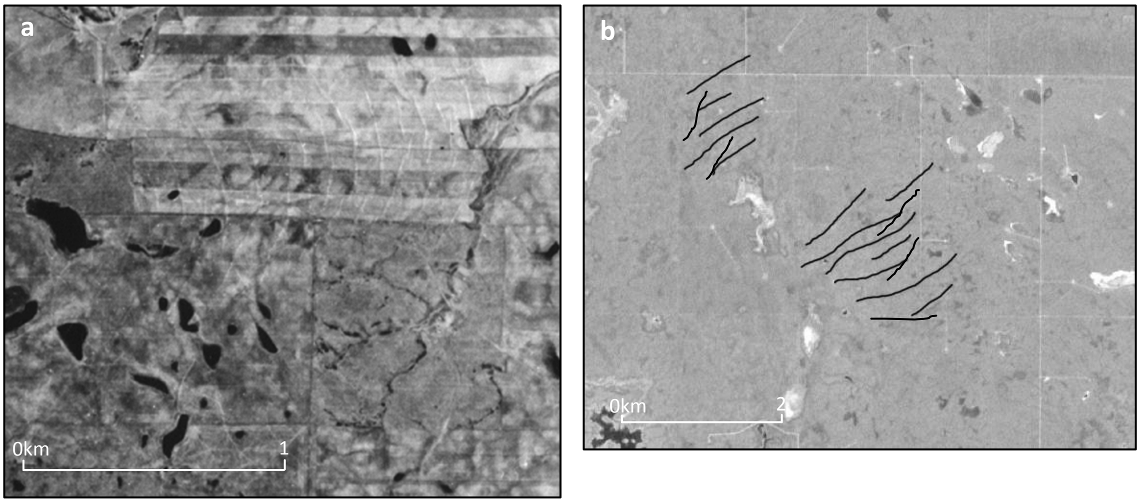


Figure 18: *The two sets of crevasse squeeze ridges found. a) arcuate cross cutting ridges that trend north to south and were identified in the north west corner of aerial photo grid 82I10 (Fig. 11). b) ridges located just north west corner of CAfs_2. The ridges have been highlighted as they are barely identifiable in the image.*

of the HPIS corridor and the western edge of Hfs_5 (Fig. 13b). They form a 40 km network of eskers that run parallel to lineation direction. Additional, eskers were identified along the centre and eastern half of the CAIS. A network of eskers was located along the eastern edge of Lake Newell. The network initiates 20 km south of CAfs_1 and terminates just south of Lake Newell. 30 km to the east, more prominent esker ridges were located at the edge of CAfs_2. The eskers run for up to 15 km at roughly 20° offset to lineation direction and are likely surrounded by numerous smaller eskers

3.1.8. Meltwater channels

Extensive networks and individual meltwater channels occur throughout the study area (Fig. 19). The mapped meltwater channels are predominantly from SRTM and Landsat ETM+ data but were also cross referenced against Shetsens' (1987) glacial geology map when resolution constraints hindered identification. Two large sequences of channels are located along the western edge of the study area with some originating from the Rocky Mountain Foothills (Fig. 10). They follow the natural slope, cross cutting and in parts running parallel to the HPIS. To the north of the CAIS a large sequence of spillway channels known as the Coronation Spondin scablands (Sjogren and Rains, 1995) dominate the landscape (Fig. 19c & d). Within the study area the sequence runs for up to 40 km, with channels reaching up to 20 m in depth. Unlike the HPIS, meltwater channels along the CAIS are orientated perpendicular and parallel to flow direction (Fig. 10 & 19b). Towards the margin of the CAIS there are extensive sequences of channels that run en echelon with transverse ridges. In particular, conspicuous large coulées dominate the marginal landforms, extending for over 100 km and reaching up to 60 m in depth. Meltwater channels are also common throughout the hummocky terrain sections, often occurring as a series of interconnected channels.

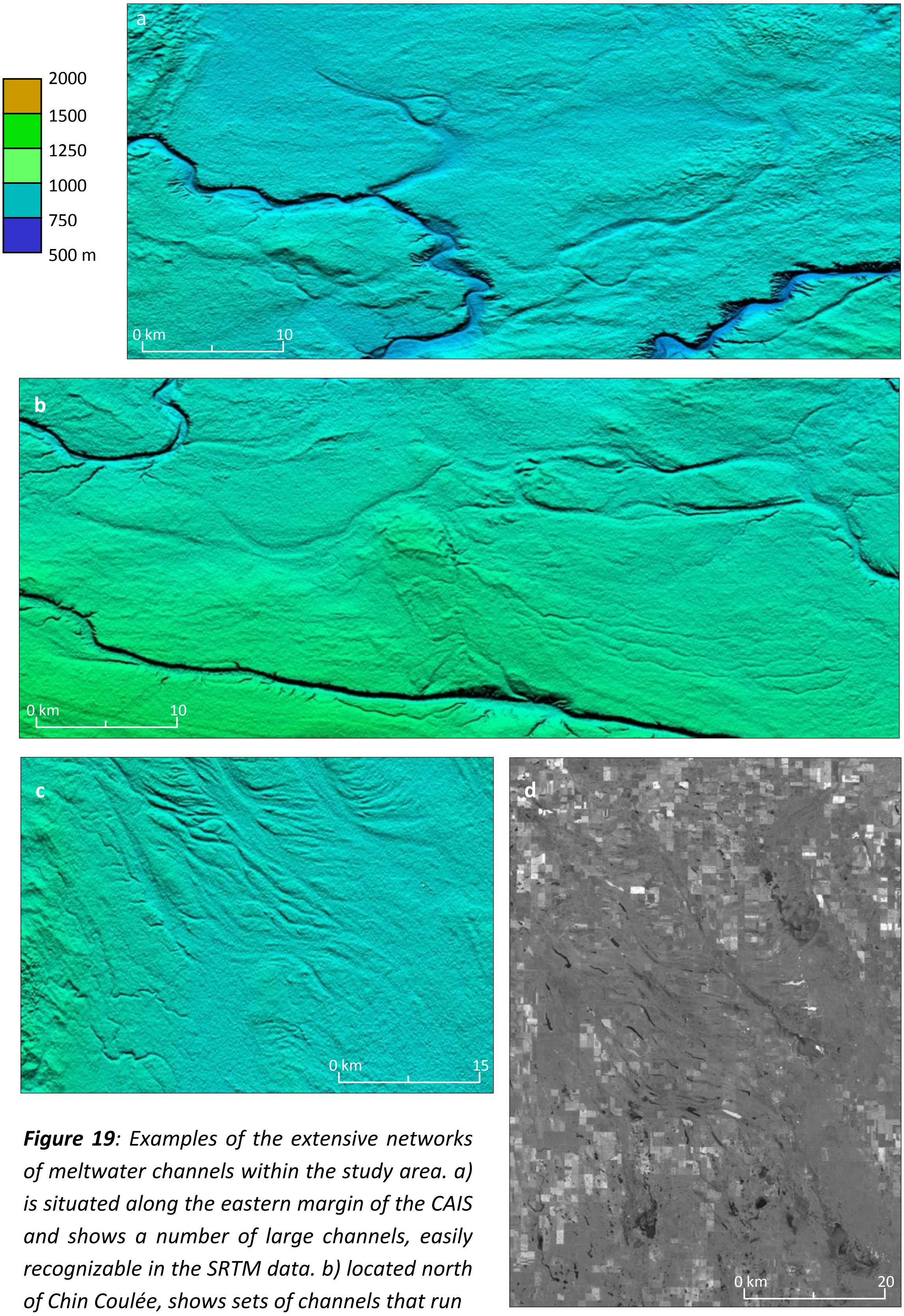


***Figure 19**: Examples of the extensive networks of meltwater channels within the study area. a) is situated along the eastern margin of the CAIS and shows a number of large channels, easily recognizable in the SRTM data. b) located north of Chin Coulée, shows sets of channels that run parallel to the transverse ridges to the south. c) and d) show the Coronation-Spondin channels (Sjogren and Rains, 1995) that cross-cut the CAIS corridor. They are excellent examples of subglacial channels formed by large discharges of water stored beneath the ice (Evans et al., 2008).*

3.2. Ice stream marginal landscape

Section 3.2 presents the results obtained from mapping of the CAIS margin. It is immediately obvious from the landforms mapped that the glacial history of this area is distinctly more varied and complex than that identified through the SRTM data (Fig. 20, insert at the back of this document). The glacial geomorphology is composed of a complex mixture of transverse ridges, hummocky terrain, flutings and meltwater channels juxtaposed with each other (Fig. 20). This area is characterised by a west to east dipping slope that is accentuated by the Milk River River and Pakokwi Lake depression, and a north to south inclined slope (Fig. 12A & 12E).

3.2.1. Transverse Ridges

TR_9 and 10 comprise part of the transverse ridge sequences in this area (Fig. 11). The large transverse ridge bands in Figure 10 are actually composed of numerous smaller transverse ridges and hummocks contiguous to each other. The majority of transverse ridges are located to the south and south-east of the Lethbridge Moraine and Etzikom Coulée (Fig. 20). In particular, the most extensive sequences lie directly south of Crow Indian Lake, Verdigris Coulee and south east of Pakowki Lake (Fig. 20). Within the CAIS marginal setting three types of transverse ridge sets were identified through the aerial photo mapping (Fig. 20).

Type 1: Within the CAIS marginal area these ridges are only located in the south east corner (Fig. 10 & 20). The ridges were large enough to be identified in the regional mapping from the SRTM data and their morphology is described in Section 3.1.4. However, large scale mapping has identified that the sequence TR_10 is actually composed of three sets of ridges. The main set (Set A) are large sub-parallel ridges that lie up ice and perpendicular to CAfs_3 and are characterised by long wavelengths between crests, and intervening hollows that are filled with numerous small lakes (Fig. 20 & 22). Aerial photos also revealed that the ridges, in particular those either side of the Milk River are overprinted flutings. The second set (Set B) lies parallel to Set A but is located along CAfs_3 and is composed of very subtle, discontinuous, densely spaced ridges that share some resemblance to the transverse ridges of Type 2. The ridges reach up to 1 km long and are no greater than 2 m high. The third set of ridges (Set C) are located down ice of Set A and just north of Set B (Fig. 20 & 22). They show a clear resemblance to the smaller ridges within Set A, with similar smooth crests and water filled depressions but are orientated north to south, roughly 45° to lineament direction. Their length varies between 1-3 km.

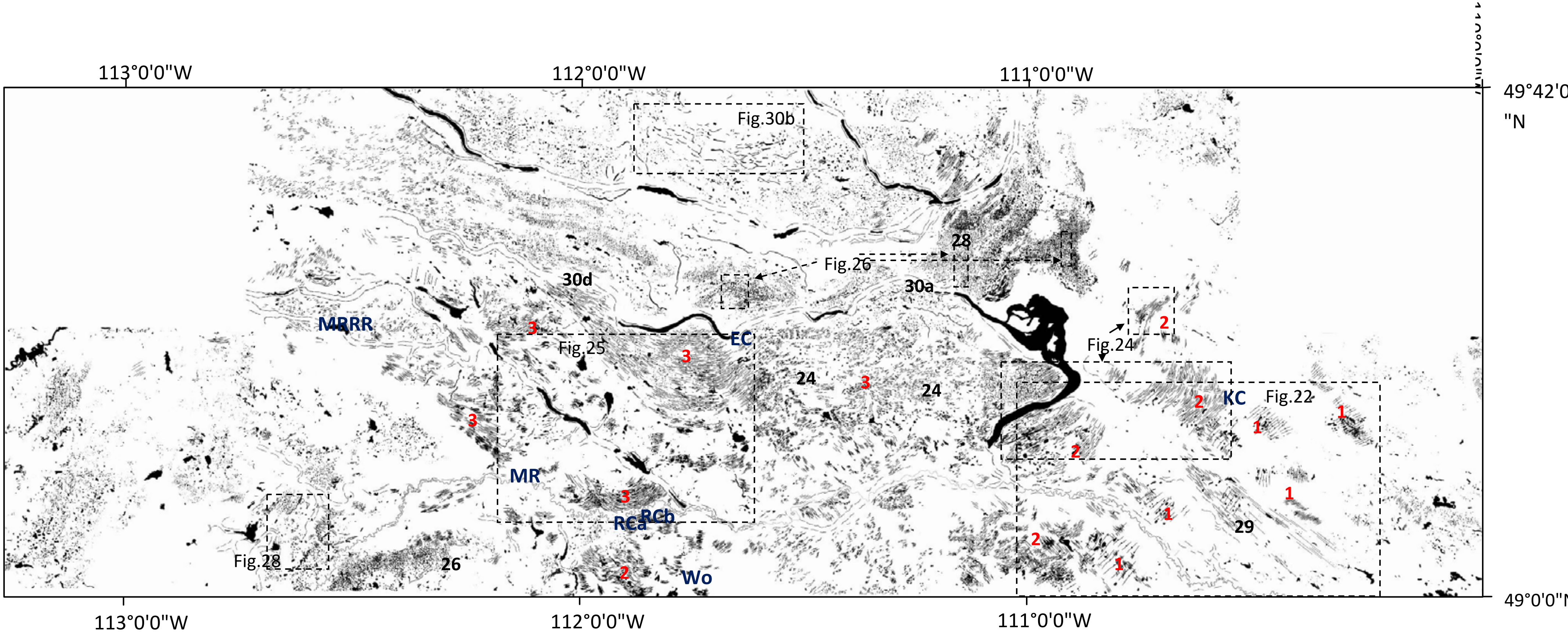


Figure 21*: Glacial geomorphology map of the CAIS marginal area. The location of Figures 13E, 18-36 are highlighted and transverse ridge sets are located in red.*

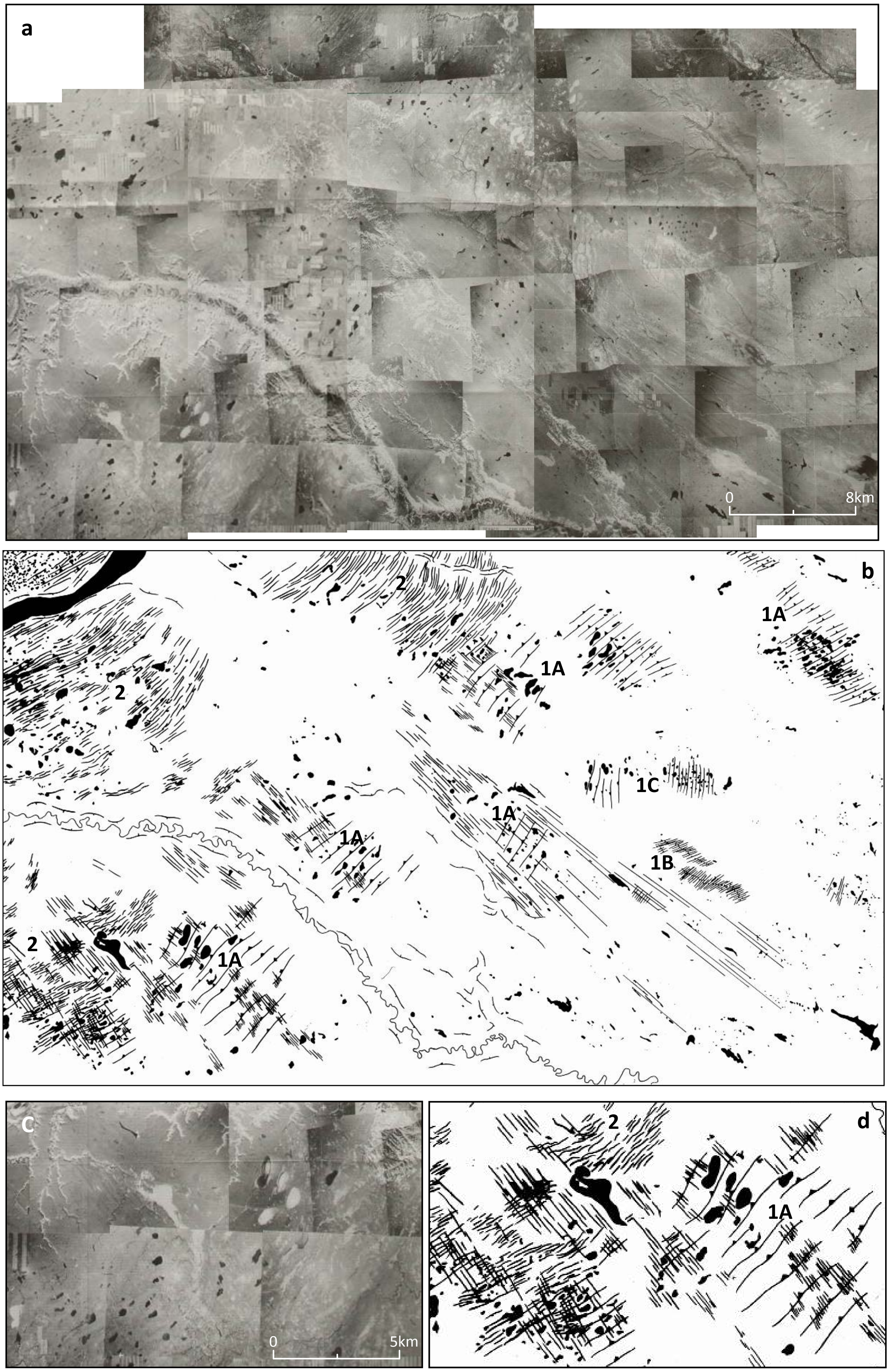


Figure 22: *Transverse ridge sets type 1 and 2 located in the SE corner of the marginal area. Individual ridge types are identified in b) and d). The map identifies lineations overprinting the ridges particularly prominent in c) and d).*

Type 2: Are found primarily along the flat terrain (Fig. 12D) between Pakowki Lake and Type 1 ridges but also south of the Milk River (Fig. 20 & 22). Type 2 ridges are characterised by conspicuous ridge sets that reach up to 5 m in height and in general have continuous ridge crests (Fig. 23). The ridges located along the south east margin of Pakowki Lake extend for up to 15 km, but in general the ridges range from 1-5 km. The type 2 ridges situated south of Milk River (Fig. 21) are more subtle and smaller than those to the south east of Pakowki Lake. Additionally, individual type 2 ridges are found throughout the study area.

Type 3: The most common type of transverse ridge are located to the west of Pakowki Lake, and are most extensive just south of Etzikom Coulee and Verdigris Coulee (Fig. 20, 21 & 23). Type 3 ridges are characterised by discontinuous transverse ridges that lie parallel and contiguous to individual hummocks, which in plan view demonstrate clear linearity (Fig. 23). Individual ridges are more subtle than type 2 with smoothed crests, and heights generally no greater than 3 m for contiguous hummocks as well. Additionally, individual hummocks are also located in between ridges and linear hummock bands, where as type 2 have no features in between crests (Fig. 23). The variety in individual hummock types will be discussed further in Section 3.2.2. Type 3 sequences show clear lobate form as highlighted in both the regional and large scale geomorphology maps (Fig. 10 & 20), and are located on the inclined slope of the ice stream marginal area (Fig. 12A). Also common are numerous small water filled depressions located in between and next to ridges. Type 3 ridges also demonstrate subtle overprinting creating cross-cutting ridge sets (Fig. 25a).

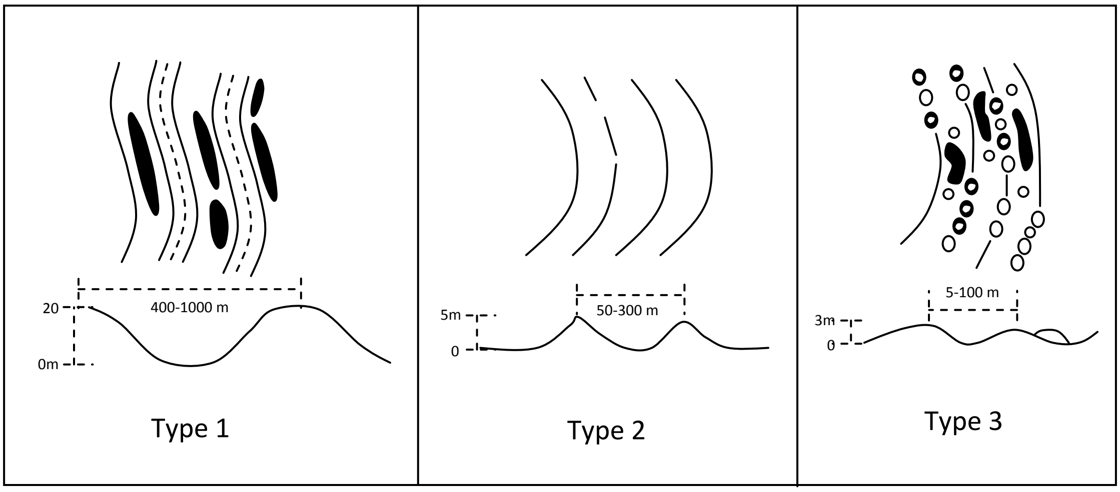
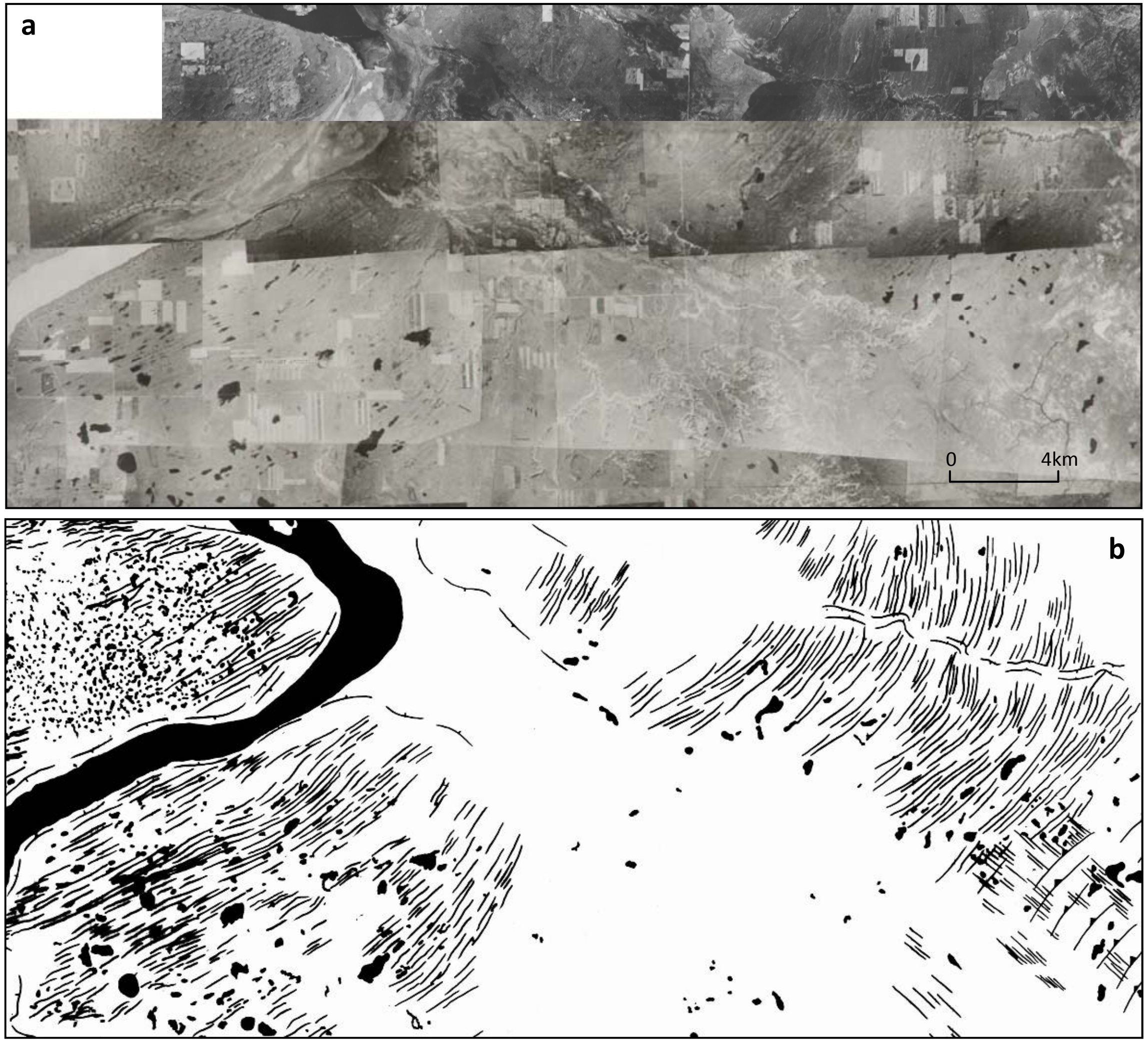


Figure 23: *Morphological characteristics of transverse ridge sets within the CAIS marginal zone. Type 1 are symmetrical in form, and have smoothed tops and water filled depression (the dotted line represents the crest of the ridge). Type 2 have sharper crests and vary in wavelength. Type 3 are composed of numerous hummocks and ridges strongly orientated with water filled depressions and occasional hummocks between ridges.*



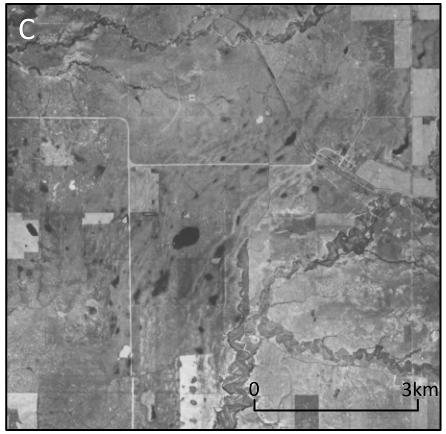


Figure 24*: Aerial photo images of type 2 transverse ridges, located to the east of Pakowki Lake (Fig. 21). Top left corner of a) type 3 ridges blending into hummocky terrain type 1. c) type 2 ridges located 5km to the north of image a (Fig. 20 & 21). d) and e) parallel, smooth crested and discontinuous nature of type 3 transverse ridges.*

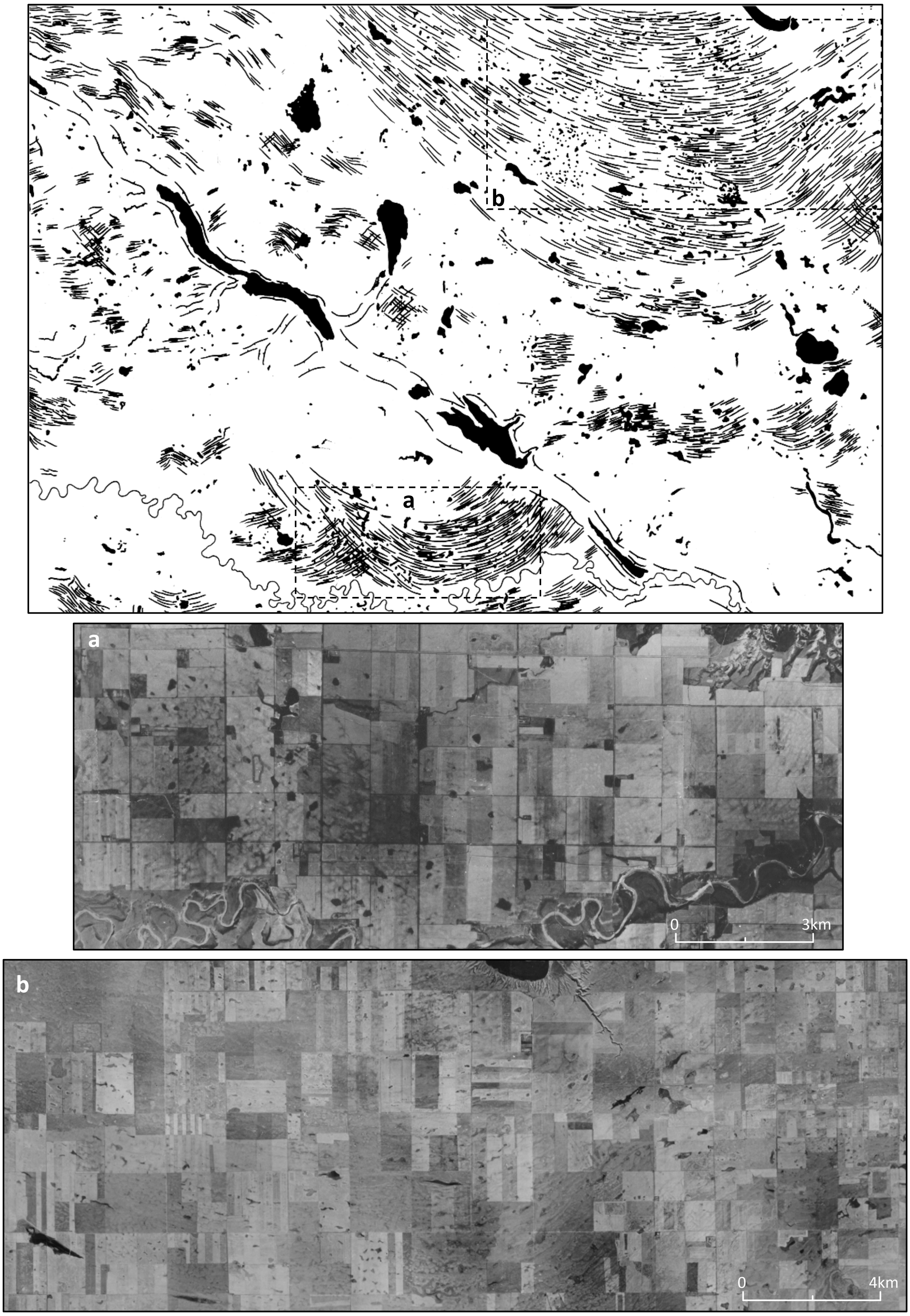


Figure 25: *Type 3 transverse ridges located in the central portion of the CAIS marginal zone (Fig. 20 & 21). Individual hummocks and transverse ridges lie contiguous to each other, leaving clear linearity in the landform record. a) located between Verdigris Coulée and the Milk River. b) is located just south of Crow Indian Lake which lies within Etzikom Coulée.*

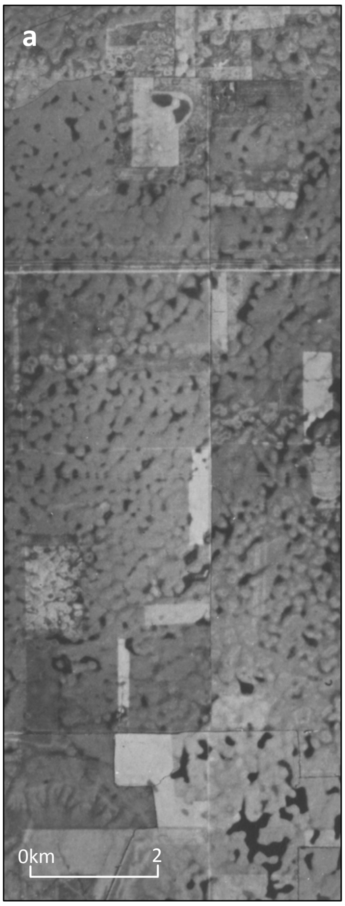
3.2.2 Hummocky terrain

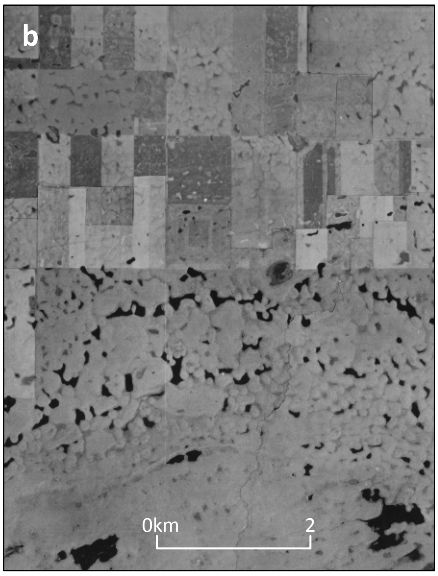
Hummocky terrain is the most common landform within the CAIS marginal zone, demonstrating a wealth of different hummock types. Internally, hummocky assemblages are chaotic and demonstrate little to no linearity. However, at a large scale they exhibit an overall lobate form that runs parallel to the sequences of transverse ridges (Fig. 10 & 20). North of Etzikom Coulee several long thin hummocky terrain bands run parallel to transverse ridges and meltwater channels. The largest extends for 60 km from west of 112°0'0"W, between Etzikom and Chin Coulée eastwards to the north of Pakowki Lake (Fig. 20). This hummocky terrain forms part of the Lethbridge Moraine which extends from Lethbridge across the marginal area and north around the Cypress Hills (Fig. 10) (Westgate, 1968; Bik, 1969). Hummocky terrain is also found in the south west corner of the study area, where it wraps around the Del Bonita Highlands and along the MRR. The linear hummocks within type 3 transverse ridge sets are composed of type 1 and 2 hummocks with no type 3 hummocks identified. Close inspection of these bands reveals three different types of hummock.

Type 1: The majority of the hummocky terrain consists of densely spaced, low relief hummocks that show little or no orientation (Fig. 26 & 27). The hummocks vary significantly in size from 1-5 m in height and generally between 5-50 m in diameter (Fig. 26). Their morphology varies from individual circular and oval shaped hummocks to interconnected larger hummocks with less rounded tops. Type 1 and type 2 hummocks lie randomly juxtaposed with each other and make up 99% of the hummocky terrain bands. Numerous small ponds fill the depressions between the hummocks.

Type 2: These hummocks generally occur randomly assorted with type 1 but occasionally areas within hummocky terrain bands are composed solely of type 2 (Fig. 21 & 26c). They are characterised by circular mounds, with a cylindrical often water filled hollow at their centre. (Fig. 27). This creates a ring or doughnut shape that is noticeably different in morphology to type 1 hummocks. Conspicuous ridges have also been identified within the hummock bands (Fig. 28). These ridges weave through the hummocks, showing no singular orientation and occasionally make up the rims of type 2 hummocks.

Type 3: Are the largest features within the hummocky terrain bands, reaching up to 20 m in height and 1 km in width (Fig. 27 & 29). They have a roughly cylindrical to oval plan form and are up to twice as high as the surrounding hummocky terrain. Some have large rims defining their shape, but all have a flat centre that lies below the outer edge. They are the least common feature of the three but easily the most conspicuous. Type 3 hummocks are best developed and primarily located in the south west corner of the study area around the Del Bonita Highlands (Fig. 28).





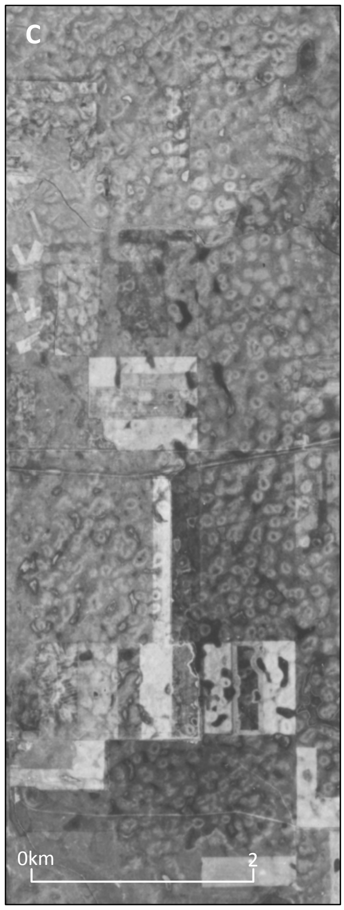


Figure 26: *Examples of type 1 and 2 hummocks. a) and b) show predominantly type 1, c) contains mainly type 2. a) and c) are located north of Pakowki Lake and b) just north of Crow Indian Lake (Fig. 21).*

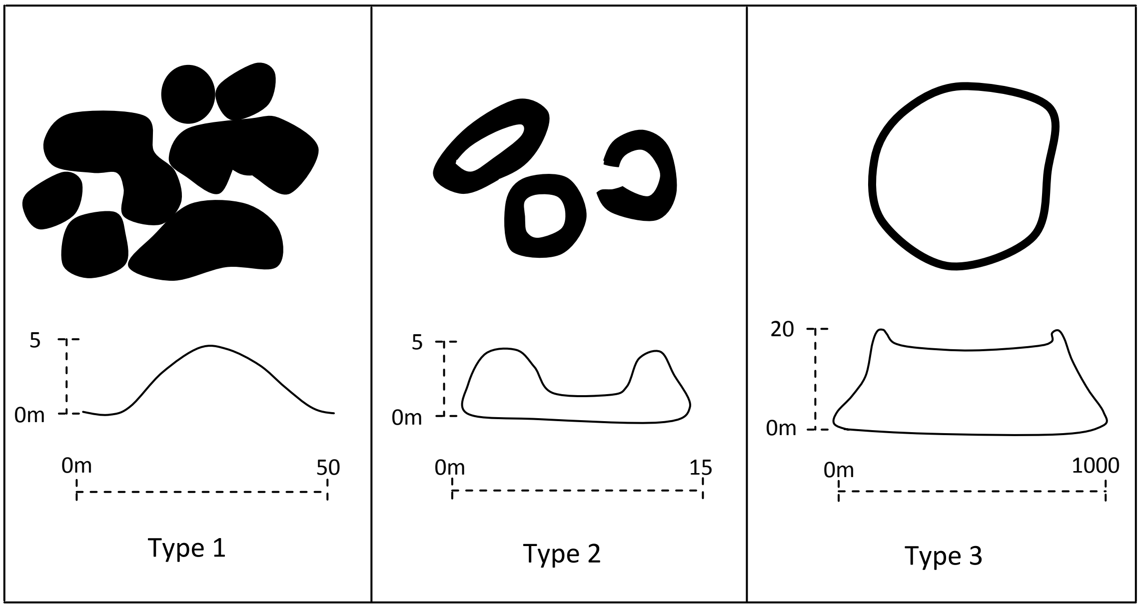


Figure 27: *Morphological characteristics of hummocks within the Lethbridge Moraine sequence. The measurements used here represent the larger sizes of each type.*

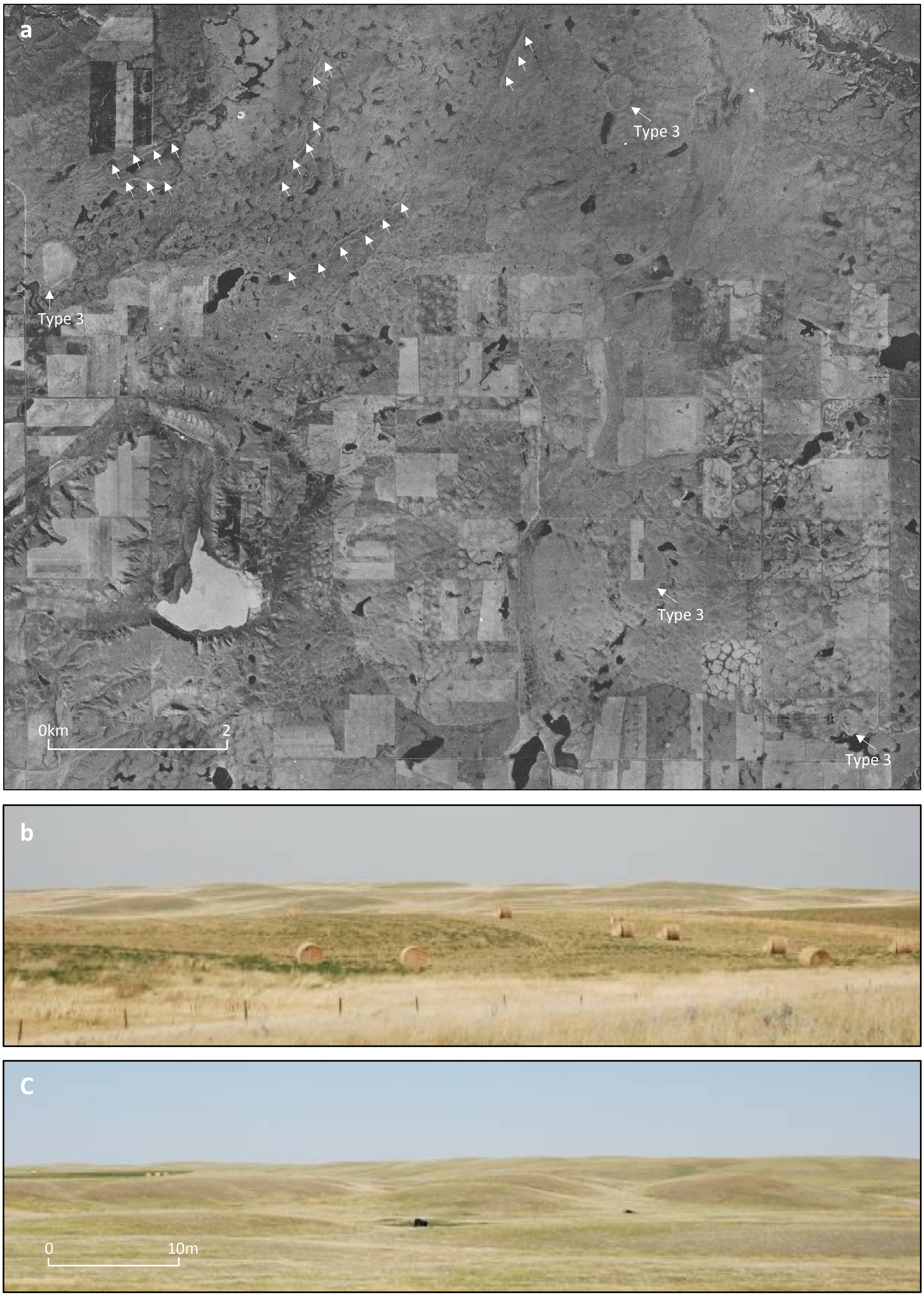


Figure 28: *Hummocky terrain examples. a) located just east of Del Bonita and shows the juxtaposition of all 3 types of hummocks. Also within the image are the ridges (highlighted by the white arrows) that run through some hummocky terrain bands. Note that here they run between type 2 hummocks and make up their rims in parts. b) and c) show examples of type 1 hummocks. The hay bales in b) are roughly 1.5m high.*

3.2.3. Lineations

Lineations are located predominantly along the eastern portion of the Milk River and south and south east of Pakowki Lake, but also north of Tyrrell Lake (Fig. 20 & 21). Lineations range from 1-9 km in length with an average of 2 km. Lineations located north and south of the Milk River clearly overprint transverse ridges (type 1, Fig. 20 & 22) at right angles and are only a few feet in height, which makes them very difficult to recognise during fieldwork (Westgate, 1968). CAfs_3 initiates between two sets of type 1 transverse ridges and is easily recognisable as the lineations are notably bigger in all morphological aspects than the other lineations mapped (Fig.29). Within the flow-set lineations reach up to 9 km in length and 6 m height. The flow set is approximately 30 km long and 5 km wide and shows an increase in lineation length along its profile. There are at least double the amount of lineaments mapped by from the aerial photographs compared to the SRTM data. Measured elongation ratios range from 12:1 up to 85:1 along the CAfs_3.

Figure 29: *Photo mosaic of a lineation within CAfs_3. The feature initiated on the right of the photo and continued for over 2 km. The stakes in the foreground are a metre high.*

3.2.4. Glaciofluvial landforms

Four major spillways extend across the study area: Forty Mile Coulée, Chin Coulée, Etzikom Coulée and Verdigris Coulée. The spillways run parallel to the transverse ridges and conform to the lobate plan form found throughout the marginal landform record (Fig. 20). They trend across the majority of the Lethbridge Moraine sequence as dominant features reaching up to 500 m in width and 60 m in height (Fig. 30). An extensive network of smaller channels situated north of Chin Coulée (Fig. 20 & 30) run predominantly parallel but also perpendicular to the spillway. The shallow channels reach up to 10 km in length and 200 m in width (Fig. 30). Larger channels are found to the north of Crow Indian Lake dissecting the hummocky terrain band at right angles. These channels reach up to 20 km in length and 100 m in width. Additional meltwater channels are found sporadically throughout the study area. Only a few eskers were identified and are located chiefly in the north east corner of the study area (Fig. 20).

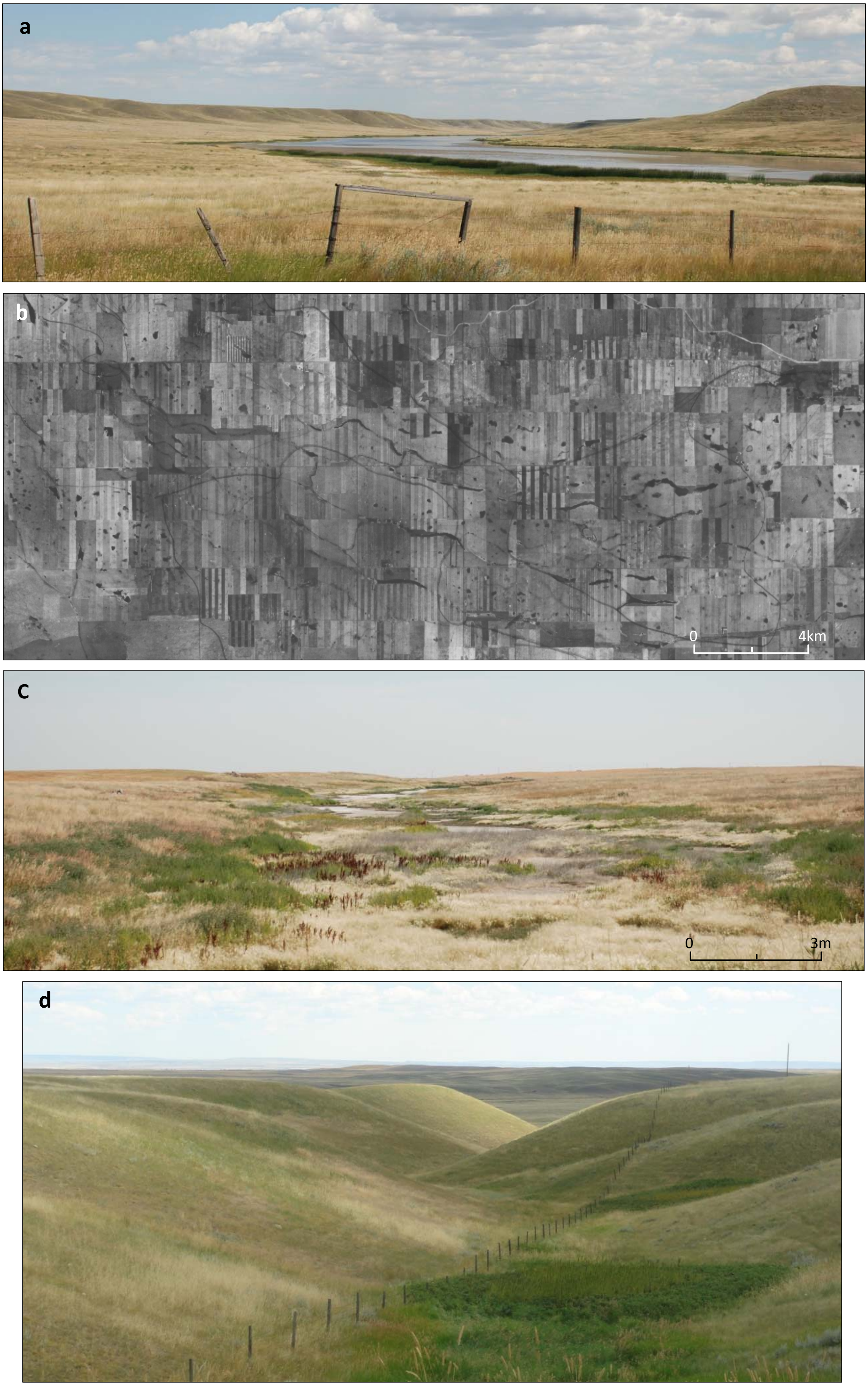


Figure 30*: a) looks west to east along Etzikom Coulée, approx. 750m in width. b) is an aerial photograph of the network of channels to the north Chin Coulée. c) small example of the network of shallow channels in b). d) shows a meltwater channel that flows into Etzikom Coulée, the channel 20m in width (Fig. 21).*

3.3. Sedimentary Results

Inherent problems lie with finding and accessing sedimentary exposures when conducting research within ancient glaciated terrain. Therefore, all sedimentary data found during field work is presented and supplemented with unpublished examples (D.J.A. Evans, Unpublished data) some of which lie just north of the mapped CAIS marginal area.

3.3.1. Etzikom Coluée (EC): Etzikom coulée section (111°39'0"W; 49°22'0"N) lies in a gravel pit situated within hummocks at the eastern end of Crow Indian Lake, on the northern side of the coulée (Fig. 31). A clast-supported boulder gravel (Bms/Gms) with clasts greater than 30 cm lies at the base of the sequence. This bouldery unit is overlain by a clast-supported gravel (Gm), which grades into a matrix supported gravel (Gms). The clasts are notably smaller (≤10 cm diameter) than the base layer. There is no grading in clast size, and clasts are sub-rounded (SR) to rounded (R). Clasts within the gravel are also no greater than 10 cm and generally sub-angular (SA) to sub-rounded (SR). Above this lies a 15 cm thick dark brown massive silty sand layer (Sm). A well consolidated matrix supported, sandy diamicton (Dmm) lies above and possesses a sharp, erosional boundary. The diamict contains clasts of varying size ranging from 1-15 cm in diameter and some outsized clasts up to 30 cm, all of which are SA and SR. Internally, the diamict was composed of a predominantly sandy texture. Clast fabric data collected from the upper gravel unit (EC1) (Fig. 31) shows a strong bimodal cluster (S_1 = 0.752) with clasts dipping at shallow angles towards the north and south. The lower gravel unit also has a strong bimodal cluster (S_1 = 0.807) with a low dipping angle.

3.3.2. Milk River (MR): This section (112°05'0"W; 49°09'0"N) is located on the distal side of a sequence of type 3 transverse ridges and does not seem to cut through any distinguishable landform (Fig. 21 & 32). At the base of the Milk River section are three separate units of massive sand/silts (Fm/Sm) which lie conformably on top of each other. A sharp contact separates the massive sands/silts with the lower matrix-supported gravel (Gms), with clasts size (≤10 cm in diameter) and generally SR to R. A further 45 cm of massive silts/sands lie above and this is sharply overlain by 15 cm of matrix-supported gravel (Gms) with clast size (≤8 cm in diameter). The above matrix-supported gravel (Gms) widens and then shrinks from north to south, forming a lense shape with SR to R clasts (2-15 cm in diameter). A 25 cm thick matrix supported gravel (Gms) is situated above, differentiated by a decrease in clast size (≤10 cm). Above lies 35 cm of well consolidated sandy silt which contains small clasts (≤2 cm) dispersed throughout. The section is topped by a 20 cm and 90 cm matrix-supported gravel (Gms) units. Both have a significantly lower clast content compared to the lower Gms units and occasional

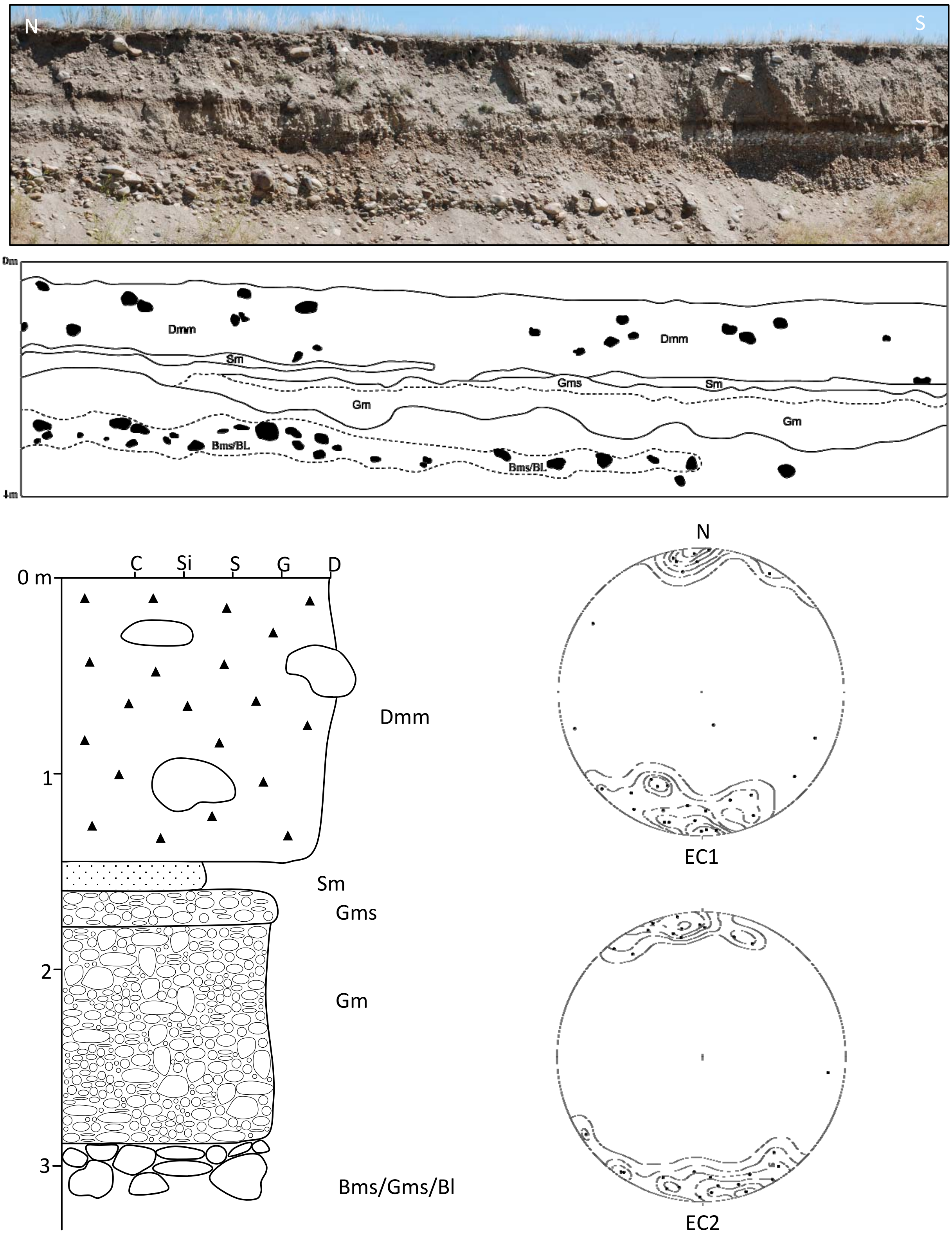


Figure 31: *Cross section and log taken from a gravel pit within Etzikom Coluée (See Fig. 21 for location).*

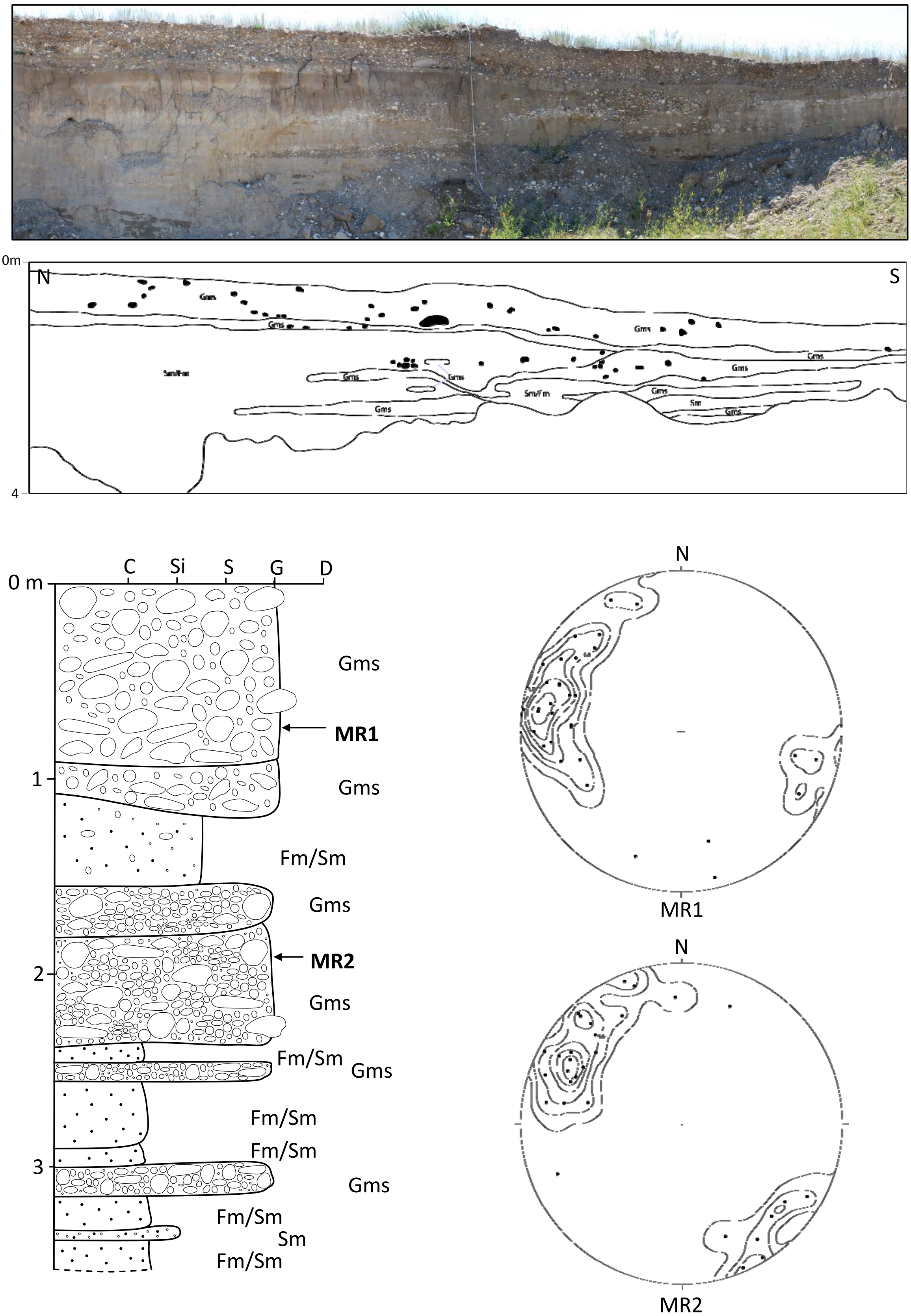


Figure 32*: Cross-section and log taken from a gravel pit 5km west of the town Milk River (Fig. 21). Tape measure in the photo is 4m in length.*

outsized clasts (30 cm in diameter). The lower Gms thins from north to south in the section, has a lighter brown colour, clast size (2-10 cm in diameter) and SR. The uppermost Gms unit is 90 cm thick and composed of SR to R clasts (1-15 cm in diameter). Clast fabric data from the upper Gms unit (MR1) shows a strong bimodal cluster (S_1 = 0.745) with a shallow dipping angle, orientated WNW to ESE (Fig. 32). The lower Gms unit (MR2) also demonstrated a fairly strong bimodal cluster (S_1 = 0.776) and shallow dipping angles with a shift in orientation to NW to SE.

3.3.3. Writing on Stone (WoS): This site is a gravel pit located 5 km south west of Writing on Stone Provincial Park and lies within poorly developed type 2 ridges (Fig. 21 & 33) (111°44’0”W; 49°02’0”N). The 2.5 m high section consists of a 1 m thick lower unit of parallel laminated sand and silt (Sl/Fl). Above lies 15 cm of matrix-supported gravel (Gms), dipping at an angle of 20° from east to west, clast size (5 mm-10 cm in diameter) and SA to SR. Above lies a cross laminated sand/silt unit that widens from left to right (15-75 cm) and consists of cross laminations that dip at 10° in the same direction. The unit is similar to the lowermost unit (Sl/Fl) but also contains small clasts (1-2 cm in diameter) located randomly throughout the unit and a greater sand content. A distinct boundary marks the change from laminated sands and silts to the above matrix supported gravel (Gms). The layer extends diagonally across section from west to east and with clasts size (1-5 cm in diameter) and occasional outsized clasts (15 cm). A further four matrix supported gravel (Gms) units lie above this layer that are differentiated by clear changes in clast size but have no clear contact boundaries. The immediate matrix supported gravel (Gms) also thins diagonally from west to east (30-10 cm) and is comprised of SA to SR clasts that vary from 3-4 cm up to 20 cm in diameter. The above matrix supported gravel is 30 cm in width and characterised by a distinct decline in clast size (5mm to 5cm in diameter), and are also SA to SR in form. The penultimate matrix supported gravel (Gms) has larger clasts (5-20 cm in diameter) of similar shape and dips at a low angle from west to east. The uppermost matrix-supported gravel (Gms) is 20 cm in width and composed of clasts (1-6 cm in diameter) and SA to SR.

3.3.4. Ketchem Creek (KC): This section is located at the eastern extent of a sequence of type 2 transverse ridges within Ketchem Creek, 20 km east of Pakowki Lake (Fig. 21, 110°36’0”W; 49°17’0”N). The section is composed of a single, well consolidated, matrix-supported diamict (Dmm), with a sand/silt texture, clasts dispersed throughout and occasional outsized boulders reaching up to 30 cm in diameter (Fig. 33). Clast size varied between 2-10 cm in diameter and were SR to R in form.

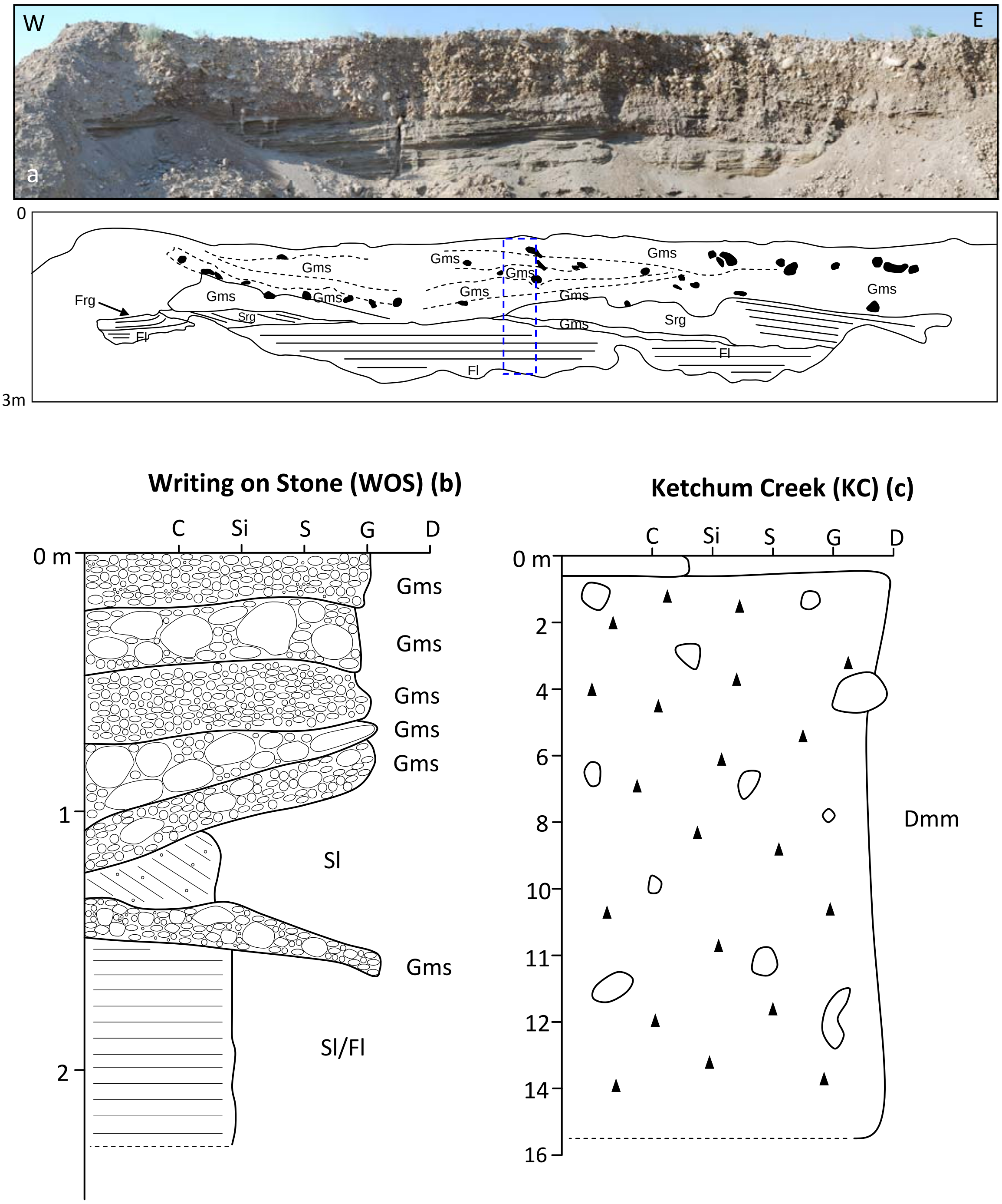


Figure 33. *a) exposure located south west of Writing on Stone Provincial Park (Fig. 21). Section log (b) is highlighted by boxed area. Section log of Ketchem Creek (c) is located 20 km east of Pakowki Lake (Fig. 21).*

3.3.5. Milk River Ridge Reservoir (MRRR): This site is situated on the north shore of the Milk River Ridge Reservoir is located within type 2 transverse ridges (Fig. 21, 112°35’0”W; 49°22’0”N). The section is comprised of a 2 m thick massive matrix-supported diamict (Dmm) that contains sand lenses; topped by a 20 cm band of massive clay/silt (Fig. 34)

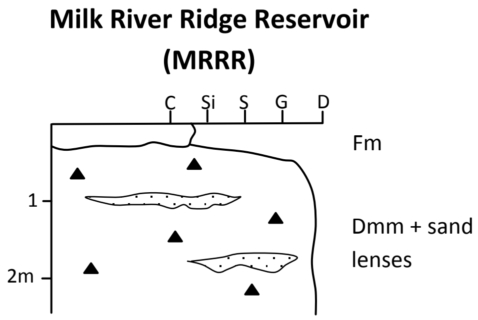


Figure 34: *Milk River Ridge Reservoir (MRRR) data provided by D.J.A. Evans (Unpublished data).*

3.3.6. Red Creek (RCa): This site is located 8 km south of Verdigris Coulée, just east of where the Red Creek meets the Milk River (Fig. 21, 111°53’0”W; 49°6’0”N). The exposure is a river cutting through a large hummock (40 m in width) positioned within a band of type 3 transverse ridges (Fig. 35). Unfortunately it could not be accessed for investigation and so the following description is from detailed photo investigation only. The exposure is composed of two diamict units highlighted by a sharp boundary that marks a change from light brown to grey/brown colour and a sharp boundary that marks the change. The lower unit (Dmm) reaches up to 2.5 m high and has at least two horizontal structures that run parallel to this boundary. Outsized clasts are visible in both units reaching up to 20 cm in diameter. The upper diamict (Dmm) varies between 1-1.5 m high and contains slumping structures. On top of this diamict lies a gravel lag. This section is capped by a unit of massive light coloured fines (Fm) that is likely representative of an alluvium layer common throughout the area (Westgate, 1968).

3.3.7. Red Creek (RCb): This exposure is located 500 m to the east of RCa (111°52’0”W; 49°6’0”N) and extends for 250 m from north to south and is located within type 3 transverse ridges with a low amplitude hummock capping the section (Fig. 36). The section is composed of 3 different diamicts differentiated by sharp boundaries that mark a distinct change in colour. The lowermost diamict is massive (Dmm), contains outsized clasts (size unknown) and extends throughout the whole sequence. The middle diamict is matrix-supported and stratified (Dms), through which the underlying diamict has penetrated forming a diapir structure. Either side of the intrusion the diamict has folded and deformed into a circular structure. This unit also contains a lense type feature, identified by its distinct shape and colour. To the north (Fig. 33b) the third diamict seems to be massive matrix supported diamict (Dmm) and is

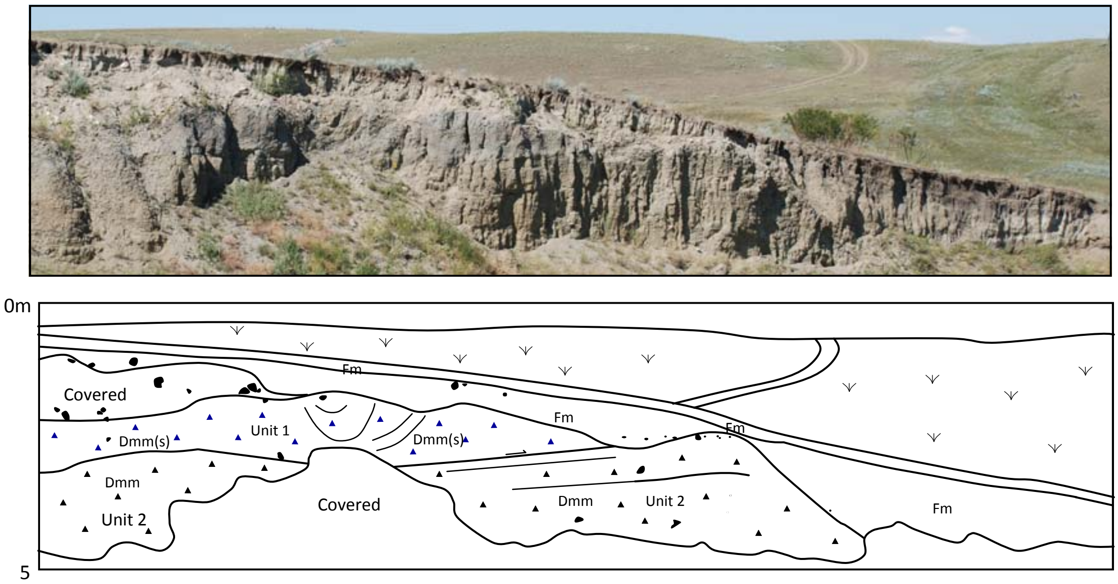


Figure 35: *Red Creek (RCa) section is located within a hummock at a river cutting situated along Road 151 South. RCb is located to the right of this image.*

distinguished by a change from the lighter brown lowermost diamict to a distinct grey colour. At the northern end of the exposure (Fig. 33b) there is an intrusive feature that is similar to a clastic dyke (Evans and Benn, 2004). Above the diamict units lies a 2 m thick, massive unit (Fm), most probably composed of fines (Westgate, 1968).The section is capped by a 1.5 m hummock form that is horizontally laminated (Sl/Fl).

3.3.8. Bow Island (BI): This site is located 10 km north of CAIS marginal area (111°43'0"W; 49°55'0"N), where the Oldman River meets the Bow River (Fig. 11). The Bow Island data (Fig. 37) is from D.J.A. Evans (Unpublished data), modified from Stalker and Berendregt (1988). The sequence is underlain by 20 m of bedrock from the Foremost Formation, which has undergone significant glaciotectonic folding and thrusting. Above this lies a gravel/boulder lag (Gm/Glg/Bl) and then roughly 6m of predominantly clays that show graded cross laminations. The sequence is capped by a further clast supported gravel layer (Gm) and a fines unit (Fm). The grain size chart shows that the top units are composed of around 90% clay and the silt content of the Foremost Formation increases with depth.

3.3.9. Pendant d'Oreille (PO): This site is located on the southern extent of Pakowki Lake (111°03'0"W; 49°11'0"N) within the type 2 transverse ridge sequence (Fig. 38, D.J.A. Evans, Unpublished data). The base of the section is composed of a matrix supported diamict with occasional laminations (Dmm/Dml). Contiguous to this are trough cross bedded (St), planar cross bedded (Sp) and ripple cross laminated (Sr) lenses that have varied sand/silt content confirmed by the grain size chart (Fig. 38). The diamict units are partially separated by a two ripple cross laminated (Sr) lenses. The uppermost 2.5 m of the exposure is composed of

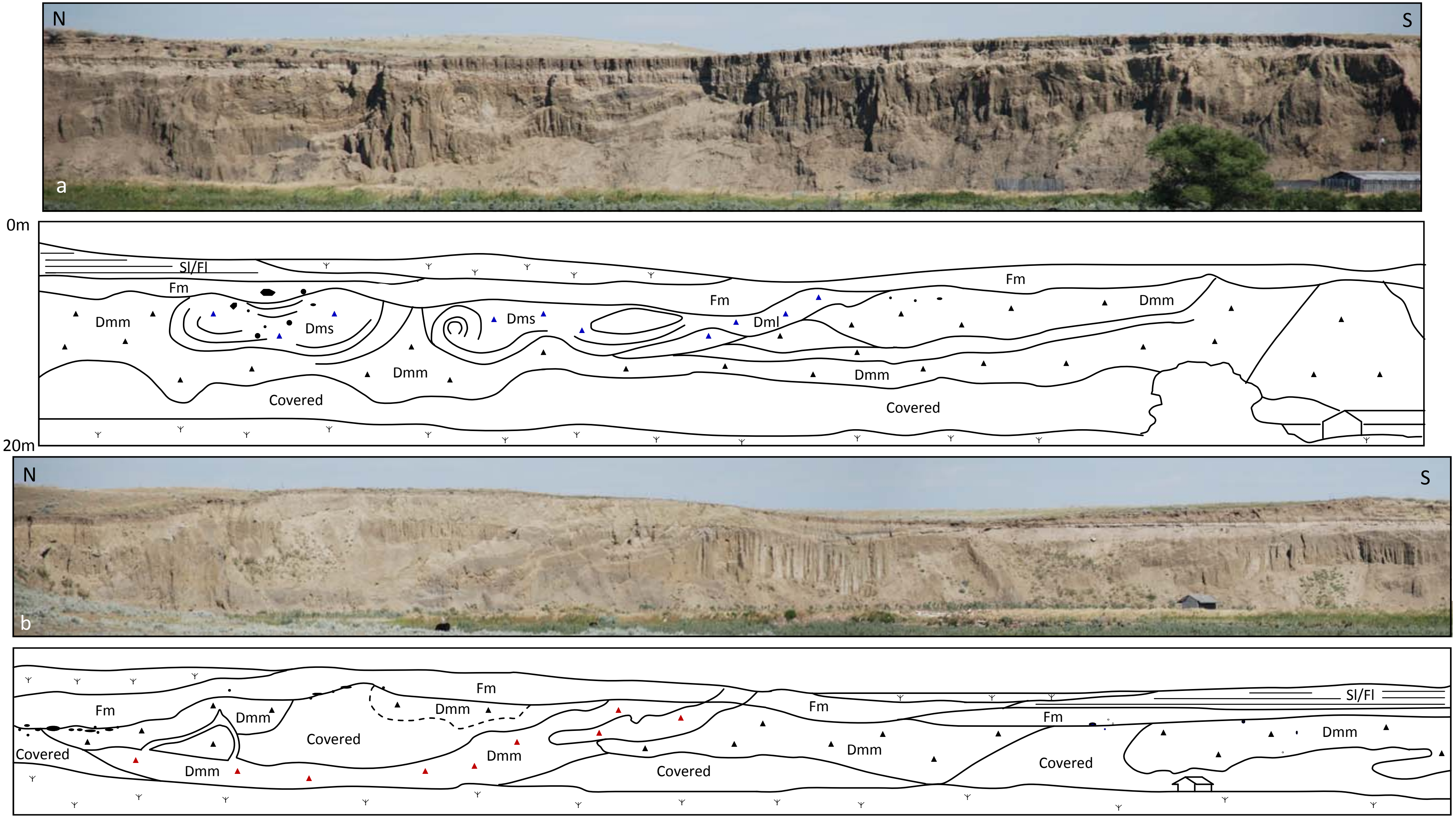


Figure 36: *Red Creek Section b (See Fig. 21 for location). The images combine to produce the whole section running from right to left. The section demonstrates multiple diamict units and a clear diapir structure in a).*

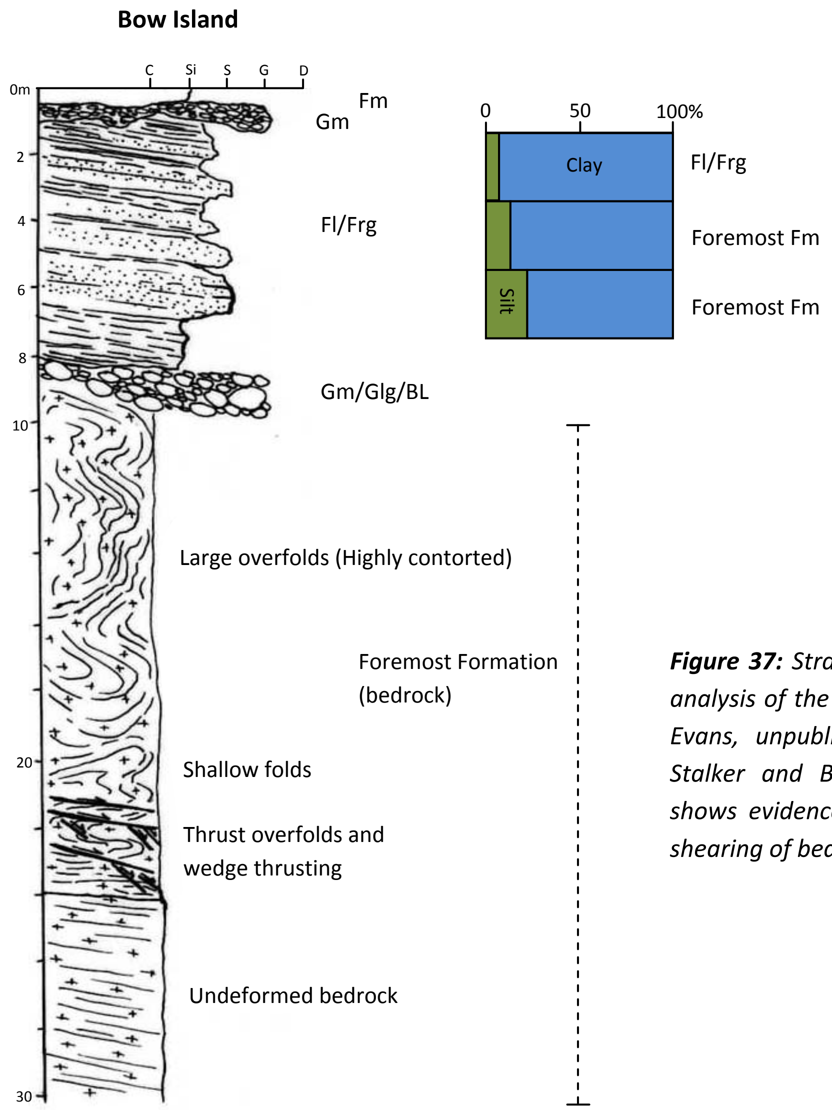


Figure 37: *Stratigraphic log and grain size analysis of the Bow Island exposure (D.J.A. Evans, unpublished data, modified from Stalker and Berendregt, 1988). The log shows evidence of significant folding and shearing of bedrock.*

of a matrix supported diamict that is partially laminated and contains sheared sand lenses throughout (Dml (s)).

3.3.10. Cherry Coulée (CC): This site is located 3km north west of Bow Island town, where Cherry Coulée joins the Saskatchewan River (Fig. 11) (111°28’0”W; 49°53’0”N). The data (Fig. 39a) is from D.J.A. Evans (Unpublished data). The section is underlain by almost 10 m of graded and climbing ripple cross laminations (Frg) composed of almost completely of clay. Above lies 19 m of matrix supported diamict (Dmm) that is composed of three different units. The lowermost dimaict contains a deformed bedrock raft, silt wisps and attenuated lenses. This grades into matrix supported, laminated diamict (Dml) with larger (5-50 cm) attenuated silt lenses and bedrock rafts. The middle unit is a matrix supported diamict (Dmm) that varies in thickness (2-4 m), and is predominatly composed of clay (75%). The section is topped by 8-10 m of matrix supported diamict with attenuated sand and silt lenses throughout (Dms). The upper diamict is the only unit to have any sand content with around 65%. The other units contain only silt and clay, with the silt content gradually decreasing with depth.

3.3.11. Wolf Island (WI): This site is located 10 km north of Purple Springs along the Oldman River (Fig. 11, 111°53'0"W; 49°55'0"N). The Wolf Island data (Fig. 39b) is provided by D.J.A. Evans (Unpublished data), modified from Stalker and Barendregt (1988). At the base of the section is 3 m of bedrock; on top of which lies a 2 m massive granule unit (GRm), a 1 m trough/planar cross bedded sand unit (Sp/St) and an eastwards thickening silt unit all of which comprise the Empress Group. Above lies 8 m matrix supported diamict (Dmm) which shows laminations, sand and silt wisps at its base (Dml) and several attenuated bedrock rafts in the top 4 m of the unit. A megablock raft composed of bedrock and clast supported gravel (Gm, Empress Group) is situated above the Dmm unit. A further 1.5 m of matrix supported diamict (Dmm (s)) lies above, with sheared sand lenses and a zone of rounded (Fig. 39b). The Upper 10 m is predominantly composed of fine silt laminations with dropstones (Fl (d)) but also numerous gravel lenses (Gm)and an occasional palimpsest/bedload lag (Glg). The grain size data shows that clay content in the lower diamict increases with depth.

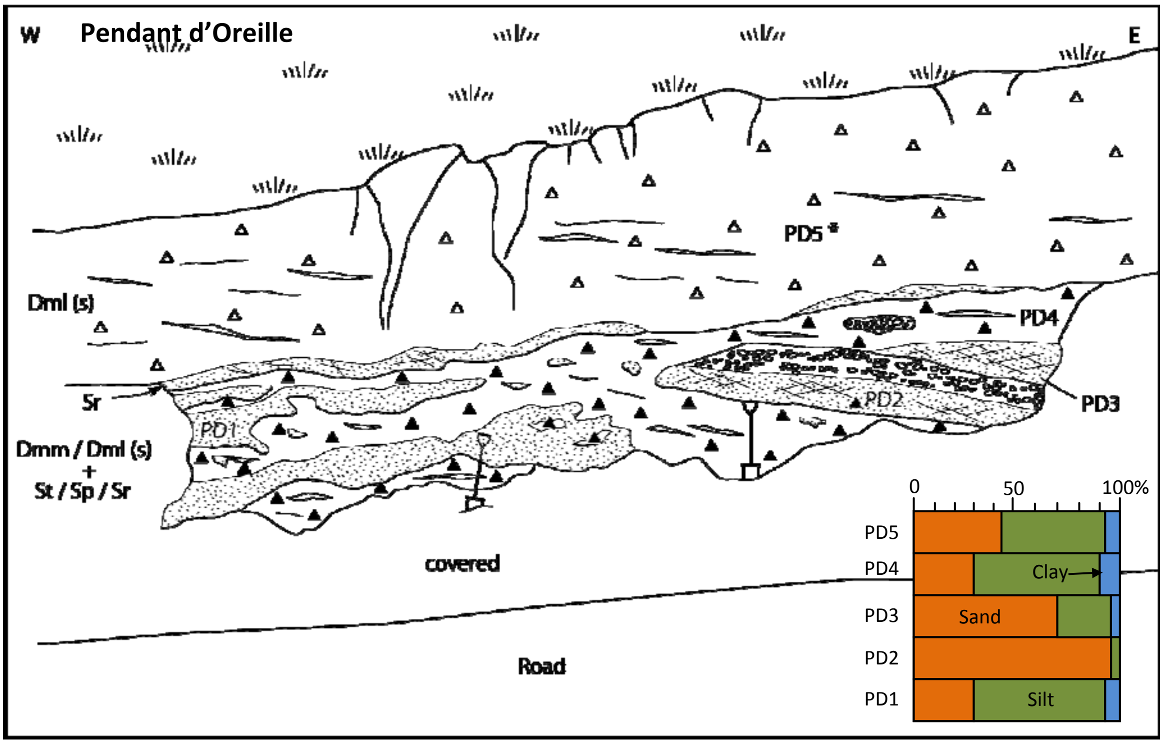


Figure 38: *a) cross section located south east of Pakowki Lake and type 2 transverse ridges (Fig. 21) (D.J.A. Evans, unpublished data).The chart represents grain size content and the spades are 1m in height.*

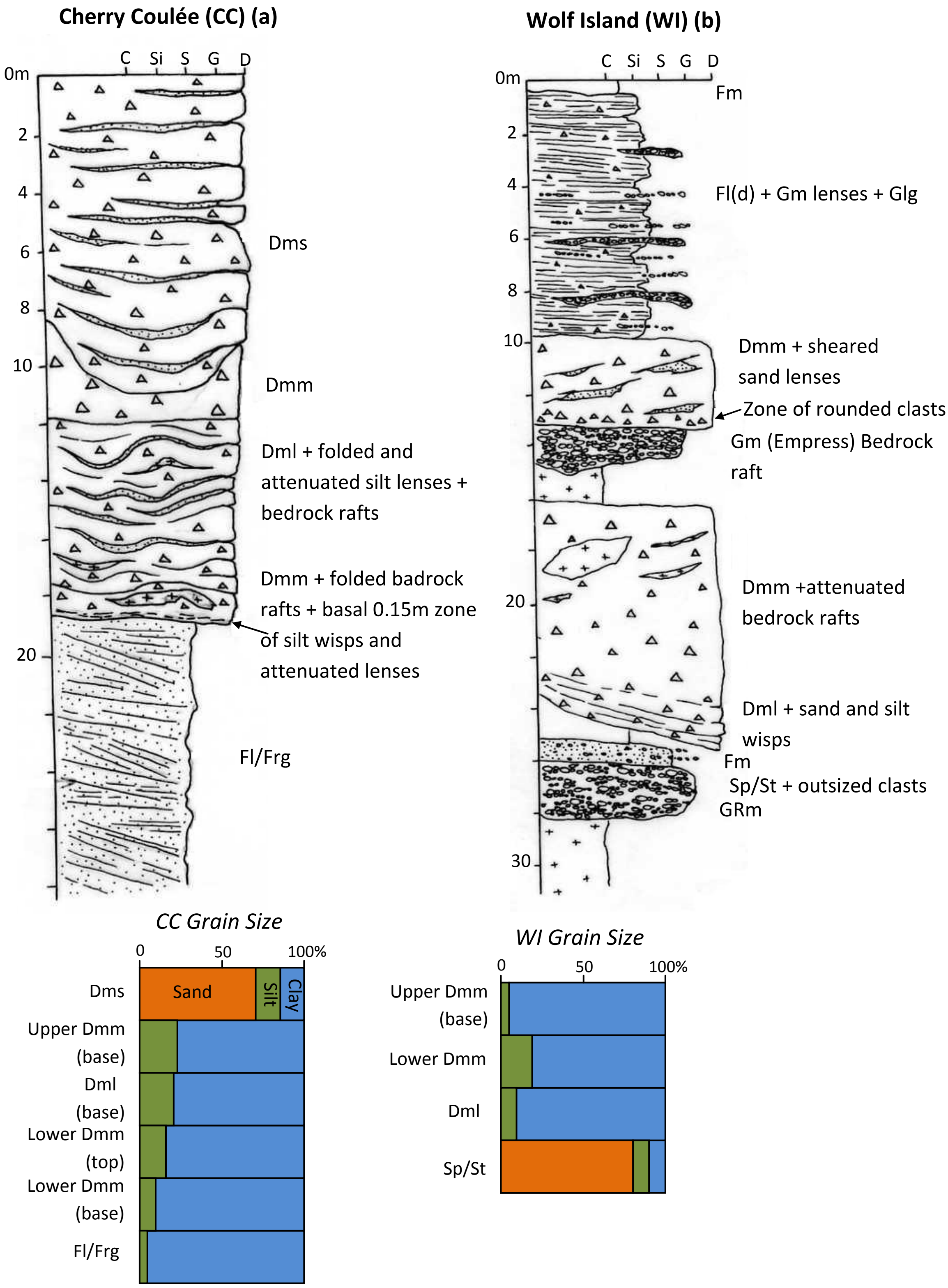


Figure 39: *Section logs taken from sites just north of the Lethbridge moraine sequence (Fig. 11). The two sections lie within 35km of each other along the Oldman and Saskatchewan Rivers. Data provided by D.J.A. Evans (Unpublished data). Both profiles demonstrate significant glaciotectonic activity.*

Chapter 4

Interpretations

4.1. Glacial Sedimentology

4.1.1. Till production

Previous investigations of the CAIS marginal area show that it is characterised by thick till sequences composed of multiple till units (Westgate, 1968; Stalker, 1969; Fullerton and Colton, 1986; Kulig, 1996; Evans et al., 2008). This is also illustrated in the Ketchem Creek (Fig. 33c), Red Creek a and b (Fig. 35 & 36), Pendant d'Oreille (Fig. 38) and Cherry Creek (Fig. 39) sections presented in this thesis. In particular, exposures along the Milk River show thick sequences of Quaternary sediments up to 25m in places (Westgate, 1968). In contrast, the streamlined corridor of the CAIS demonstrates a very thin sediment cover (Evans, 2000). Within southern Alberta lie till wedges which thicken down ice and multiple stacked till sequences formed through multiple re-advances of the CAIS (Westgate, 1968; Kulig, 1996; Evans et al., 2008). This evidence is consistent with subglacial deformation theory (Alley, 1991; Boulton, 1996a, b), which suggests that it was a major control on both the production and distribution of sediments in southern Alberta. Importantly, Evans and Hiemstra (2005) and Evans et al. (2006b) demonstrate that the overall effect of subglacial processes is ice marginal thickening through advected material. Therefore, it is likely that several subglacial processes, such as lodgement, deformation and clast and ice keel ploughing (Evans and Hiemstra, 2005) acted to create the CAIS sedimentary profile of which subglacial deformation was a major contributor.

The upper diamict in the Pendant d'Oreille section (Fig. 38) demonstrates attenuated rafts of stratified sediment. It is possible that these features were derived from the underlying diamict and sand interbeds through deformation and incorporation during ice advance (Evans, 2006b). Additionally, the exposure is located just outside of the type 2 transverse ridge sequence (Fig. 21) and most likely relates to the advance which formed the surrounding lineations, which are interpreted to have formed during a fast flow event (Section 4.6). This evidence is consistent with a subglacial traction till, emplaced during an advance through the south eastern corner of Alberta (Evans et al. 2006b). Importantly, preserved rafts in sedimentary exposures within southern Alberta have been interpreted as evidence for short transport distances in the traction layer, which in turn suggests that basal sliding was the primary flow mechanism and surrounding lineations were formed through ploughing by ice keels (Evans et al., 2008).

4.1.2. Glaciotectonics and sedimentary structrues

The Bow Island (Fig. 37), Cherry Coulée and Wolf Island (Fig. 39a & b) section logs, and previously published work by Westgate (1968), Stalker (1973; 1976), Evans (2000), Aber and

Ber (2007), Evans et al., (2008) demonstrates that glaciotectonic activity was widespread throughout the CAIS marginal area. The presence of folded and sheared lenses within diamicts, bedrock rafts, highly contorted bedrock and sediments are all well documented evidence for glaciotectonism (Aber et al., 1989; Evans and Ó Cofaigh, 2003; Evans et al., 2006b; Aber and Ber, 2007). This evidence suggests that both ice marginal and subglacial glaciotectonism occurred throughout the CAIS marginal area (Aber and Ber, 2007). Furthermore, evidence presented by Evans et al. (2008) of glaciotectonic features on the down ice side of preglacial valleys suggests the valley fills and bedrock cliffs were an important source for sediment production.

The inclined proglacial slope within the CAIS marginal area (Fig. 13A) and high porewater pressures created through poor drainage and ice loading is likely to have produced a setting that is highly conducive to glaciotectonics (Aber et al., 1989). Poor drainage may have occurred due to the high montmorillonite content within the local geology (Klassen, 1989) and patchy permafrost (Bednarski, Personal communication). This in turn will have increased the likelihood of proglcial sediment failure, especially within rapidly advancing margins (Aber et al., 1989; Aber and Ber, 2007). Evans et al. (2006a) demonstrated that low amplitude ridges and linear hummocks around Travers Reservoir had glaciotectonic origins; importantly, these landforms closely relate to those at the CAIS margin and so may have similar origins. However, without extensive sections through landforms it is difficult to ascertain their exact origins.

The intrusive structure situated within the Red Creek b (Fig. 36) section is interpreted as a diapir that originated from below the folded unit. As the diapir was forced upwards it caused the above material to fold into an almost concentric form. It is possible that the diapir formed as a result of density contrasts within poorly drained saturated sediments, with gravity acting to propel the lower-density material upwards (Aber and Ber, 2007). Whilst no other documented examples exist within the study area, similar density driven structures have been documented by Rijsdijk (2001), Hiemstra et al. (2005) and Aber and Ber (2007). Boulton and Caban (1995) comment on similar structures in relation to hydrodynamic blowout or 'extrusion moraine' and that the structures are a product of glacier loading and changes in effective pressure. However, the feature does not record any surface expression and so the surrounding hummocks are most likely a product of other mechanisms. Furthermore, due to relatively small loading stresses imparted by a thin margin, it is likely that the diapiric structure is not of any glaciotectonic origin. The origin of the clastic dyke is unknown but it is possible that it formed through similar processes.

4.2. Transverse Ridges

SRTM, Landsat ETM+ and aerial photograph mapping identified four types of transverse ridge sets: thrust, push, push and linear hummocks, and crevasse squeeze ridges. Each will now be reviewed in turn.

4.2.1. Thrust ridges (Type 1)

4.2.1.1. CAIS marginal area

All three type 1 ridge sets have been overrun by a re-advancing ice margin. The largest ridges (Set A, Fig. 21, 22 & 23; also referred to as TR_10 in Figure 11) are overprinted with lineations and their tops have been smoothed by the advancing ice. The ridges are composed of deformed bedrock (Beaney and Shaw, 2000), which is confidently interpreted as evidence for proglacial thrusting (Aber et al., 1989; Evans et al., 2008). Previous mapping of the transverse ridges has also interpreted them as ice thrust ridges (Westgate, 1968, Shetsen, 1987; Evans et al., 2008). Alternatively, Beaney and Shaw (2000) interpreted them as either a polygenetic landform, formed by thrusting and fluvial erosion or solely by fluvial erosion. Their morphology shares distinct resemblance to ice thrust ridges described by Bluemle and Clayton (1984) in North Dakota and Tsui et al. (1989) in central Alberta (Beaney and Shaw, 2000). Based on this evidence they are confidently interpreted as overrun ice thrust ridges, composed almost entirely of bedrock. In agreement with Beaney and Shaw (2000) their location along the preglacial divide suggests that topography was significant in their formation. The margin will certainly have experienced significant compressive forces as the margin met the higher topography (Fig. 13D), producing a setting highly conducive to glaciotectonism (Bluemle and Clayton, 1984; Aber et al., 1989). Tsui et al. (1989) comment on the importance of the impounding of proglacial water bodies by inclined slopes as a crucial factor in ice thrust ridge production. Whilst the implications of this were ridges located in topographic troughs, the same settings are interpreted to have helped form these ridges. Specifically, high porewater pressures created by swelling montmorillonite contained within bentonite in the surficial geology and proglacial ponding against the inclined slope created poor drainage conditions favourable to glaciotectonism (Aber et al., 1989). Moreover, the combination of high porewater pressures, weak bedrock and compressive marginal flow will have enhanced the likelihood of sediment failure and glaciotectonic deformation (Aber et al., 1989).

The origins of type 1 (Set B) transverse ridges are unknown (Fig. 22). They have been heavily modified by re-advance of the margin and are barely distinguishable in the landform record. Consequently, further investigation is required to elucidate their origins but they could also be thrust ridges produced by fast flow and high porewater pressures common in this area. Their orientation parallel to Set A suggests they may formed during the same advance. Type 1

(Set C) transverse ridges are very similar in form to Set A but have been significantly modified into more subtle and smoothed ridges. Based on their similar morphology and position on the preglacial divide they are also interpreted as overridden thrust ridges. Their different orientation and more subtle form indicates that they most likely formed during a previous flow episode through this area (Wesgate, 1968; Kulig, 1996). The sequence of ridges and overprinted lineations within the south east corner therefore records at least three advances of the margin through this area (Westgate, 1968; Kulig, 1996).

Thrust ridges are common within contemporary surging forelands (Sharp, 1985b; Croot, 1988; Bennett et al., 1999; Evans and Rea, 1999; 2003) and have been documented at the margins of terrestrial palaeo-ice streams (Patterson, 1997; 1998; Colgan et al., 2003; Van der Wateren, 2003; Jennings, 2006). However, they have also been documented along non-surging, sub-polar margins where permafrost is present (Evans and England, 1991) and are not necessarily indicative of fast flow (Evans and Rea, 2003; Van der Wateren, 2003). The overprinting and surrounding lineations all have elongation ratios greater than 10:1, which has been demonstrated as evidence for fast flow (Stokes and Clark, 2002). Therefore, evidence for multiple re-advances of the CAIS margin, thrust ridges and lineations juxtaposed with each other strongly suggests that several fast flow events occurred within this area. Moreover, similar glaciotectonic deformation evidence to that presented in this thesis (Figs. 37, 39a and b) has been suggested as evidence of a surging margin (Boulton et al., 1985; Clayton et al., 1985; Fisher et al., 1985; Clark, 1994b; Evans and Rea, 1999; 2003; Colgan et al., 2003; Evans and Ó Cofaigh, 2003) and ice streaming (Colgan et al., 2003; Van der Wateren, 2003; Jennings, 2006). This implies that fast flow was not constrained to the south east corner of the CAIS marginal area.

4.2.1.2. Southern Alberta

The large arcuate ridges (TR_1b, Fig. 10 & 11) dissected by the Little Bow River have been streamlined and overprinted by lineations, documenting a more southerly advance of the HPIS. The arcuate nature of the ridges implies they are ice marginal features and so likely record an earlier advance of the HPIS to this location. The large ridge forms lie on a rise in bedrock topography positioned 30-60m above the surrounding land (Geiger, 1967) and are significantly different in morphology to the surrounding recessional ridges (Fig. 10). Their size, multiple crests and location on a rise in bedrock topography are compatible with glaciotectonic origins (Bluemle and Clayton, 1984; Aber et al., 1989; Aber and Ber, 2007). Ice thrust masses have also been located just up ice of this location around the Frank Lake area (Evans et al., 1999). As mentioned above the geology of southern Alberta contains bentonite a type of clay that swells when wet (Klassen, 1989), rendering the underlying sediments almost impermeable

(Beaty, 1990). The poor drainage combined with ice loading would produce high porewater pressures in ice marginal and subglacial settings which in turn are strongly associated with glaciotectonic activity (Aber et al., 1989). These conditions will have been common throughout southern Alberta and likely significantly affected ice dynamics and landform production. This circumstantial evidence suggests that these ridges are highly likely to be of glaciotectonic origin.

The north east corner of the study area is composed of Transverse ridge sets 3 and 4 (TR_3 & 4) (Fig. 11). Previous mapping in the area of TR_3 by Kjearsgaard (1976) and Shetsen (1987) identified significantly fewer transverse and ice thrust ridges than mapped in Figure 10. The large and sharp crested nature of ridges is different from the other recessional transverse ridge sets (TR_1 & 9) and it is possible that they have glaciotectonic origins (Moran et al., 1980; Evans et al., 2008). Further research is necessary in order to ascertain their origins. Transverse ridge set 4 (TR_4) match the mapped landforms by Kjearsgaard (1976), Shetsen (1987) and Evans et al. (2008) and so in agreement with these authors are interpreted as ice thrust ridges. Furthermore, part of the sequence directly corresponds with the location of the Neutral Hills, a well known composite ridge (Aber and Ber, 2007).

Transverse ridge sets 5 and 6 (TR_5 & 6) (Fig. 11) share an almost identical affinity in location and morphology and so are interpreted to have the same genetic origins. Both sets of ridges are found on top of bedrock highs (Fig. 13A) in similar fashion to TR_10, the corollary is that their formation was influenced by topographical controls in similar style to TR_10 and TR_1b (Bluemle and Clayton, 1984; Aber et al., 1989). They show multiple ridges in similar style to the Neutral Hills and other documented composite, thrust ridges (Aber and Ber, 2007). Glaciotectonized bedrock has been documented within TR_6 (Evans and Campbell, 1992; Evans, 1996) and the overall arcuate plan form of both sets of ridges supports an ice marginal origin. Based on this evidence both sets of ridges are confidently interpreted as ice thrust ridges formed by compressive marginal flow caused by higher topography (Evans, 1996, 2000; Evans et al., 2008). A thin till cover situated on top of the ridges suggests that they are actually cupola hills (Benn and Evans, 1998; Evans, 2000). The till cover also implies they were overridden during a re-advance of the CAIS (Evans, 2000). The timing of their formation and subsequent re-advance is unknown. Transverse ridge set 7 (TR_7) are located within TR_6 and so it seems likely they share similar origins. They are also interpreted as thrust ridges with differences in morphology associated with meltwater channels that run between the ridge crests, accentuating and modifying them.

Transverse ridges sets 11 and 12 (TR_11 & 12; Fig. 11) are interpreted as a single sequence of ridges as opposed to two separate sets, formed at the margin of the east lobe ['Ice Stream 1'] (Ó Cofaigh et al., in press). Extensive sections through the ridges show that they have been glaciotectonically thrust and stacked (Ó Cofaigh et al., in press). Based on this evidence and their size, they are interpreted as ice thrust ridges.

4.2.2. Push ridges (Type 2)

Type 2 transverse ridge sequences (Fig. 23) are primarily located east and south east of Pakowki Lake and south of the Milk River (Fig. 20 & 21). Their morphology and inset nature shares a close resemblance to active temperate margins in Iceland, for example at Breiðamerkurjökull and Fjallsjökull (Evans and Twigg, 2002; Evans, 2003). The whole sequence of transverse ridges within the CAIS marginal area has been interpreted by Evans et al. (1999), Evans (2003) and Evans et al. (2008) to represent recessional push ridges. However, there are differences between type 2 and 3 transverse ridge sets and so that interpretation is not wholly accepted. Type 2 ridges are interpreted as push ridges based on their individual and collective resemblance to contemporary examples. This therefore implies that the CAIS margin must have been warm based during this time, responding to seasonal climate variations and hence is representative of an active temperate margin (Evans and Twigg, 2002; Evans, 2003). Alternatively, Westgate (1968) interpreted type 2 ridges as washboard moraine, linear disintegration ridges and ridged end moraine. Further investigation into their sedimentary architecture is needed to confidently confirm and reject these interpretations.

The exact constructional mechanisms of these ridges are unknown but they are best developed along the flat terrain of the Pakowki Lake depression (Fig. 13D, E & 20). As previously mentioned poor drainage conditions and highly saturated sediments will have prevailed in the marginal setting of the CAIS and so it seems likely that Price's (1970) squeeze model for moraine construction operated in this setting. However, also likely are other constructional models, such as Sharp (1984), Matthews et al. (1995), Krüger (1996) and Evans and Heimstra (2005).

The large lobate ridge (TR_8) has previously not been identified. The ridge is hereafter named the Vauxhall Ridge after the nearest town and is almost certainly ice marginal based on its lobate plan form. Furthermore, it lies down ice and perpendicular to CAfs_1, which suggests that it records the re-advance limit of the CAIS during this flow event. The ridge also aligns with hummocky terrain and transverse ridges to the east, which are interpreted to have formed contemporaneously. Its geomorphic expression provides few indicators as to its genetic origins, and so further investigation is required.

4.2.3. Push ridges and linear hummocks (Type 3)

Type 3 ridge sequences are composed of linear hummock tracks and discontinuous ridges which lie contiguous to each other (Figs. 20, 21, 23-25). The small scale mapping (Fig.

10) shows a clear recessional pattern of the CAIS and HPIS margin, similar to contemporary active temperate glaciers (Evans et al., 1999; Evans and Twigg, 2002; Evans, 2003, 2009). The large scale mapping (Fig. 20) shows that the landform record is significantly more complex but still records an active recession of the margin. Importantly, type 2 and type 3 ridges formed during the same advance and retreat episode (Westgate, 1968; Kulig, 1996) and so lie on a continuum of landforms. Morphologically type 3 ridges are similar to type 2 (Fig. 23), however, any explanation for type 3 sequences must account for the production of discontinuous transverse ridges contiguous to linear hummocks.

The best analogues to type 3 ridges are located around Frank Lake (Fig. 11; Gravenor and Kupsch, 1959; Evans et al., 1999; Evans, 2003, 2009; Johnson and Clayton, 2003) and along the margins of the southern Laurentide lobes, for example North and South Dakota (Colgan et al., 2003). The landform assemblage of the Frank Lake area has been interpreted as controlled moraine (Gravenor and Kupsch, 1959; Johnson and Clayton, 2003), or push moraines and proglacial blow-out features (Evans et al., 1999; Evans, 2003, 2009). The controlled moraine explanation proposes that englacial structures in the parent ice control the landform imprint formed, leaving linear hummocks and superimposed ring forms (Gravenor and Kupsch, 1959; Johnson and Clayton, 2003). The aligned hummock tracts of landsystem C (Colgan et al., 2003) share some resemblance to those within southern Alberta and are interpreted to represent ice marginal stagnation following a surge or fast flow event. Johnson and Clayton (2003) also comment on similar landform assemblages formed from supraglacial sources associated with stagnant margins, however, they also suggest that up ice of the margin ice will still have been active and so the final landform record is a product of re-advances incorporating the debris rich stagnant margins.

The individual hummocks within type 3 ridge sets vary between type 1 and type 2 hummocks, which are interpreted as having formed supraglacially (Section 4.3). In order for such landforms to form to the extent found within the CAIS, large debris concentrations would have been necessary in englacial and supraglacial conditions. This will have occurred primarily through englacial thrusting of debris rich ice due to compressive flow accentuated by the inclined proglacial slope (Fig. 13A) (Boulton, 1967, 1970; Ham and Attig, 1996; Hambrey et al., 1997, 1999; Bennett et al., 1998; Glasser and Hambrey, 2003; Swift et al., 2006; Roberts et al., 2009). The reincorporation and stacking of stagnant debris charged ice may have further thickened the debris charged margin if the previous advance of the CAIS had resulted in ice marginal stagnation (Eyles, 1983; Sharp, 1985b; Ham and Attig, 1996, Patterson, 1997; Jennings, 2006; Clayton et al., 2008). For type 3 ridge sequences to be of controlled moraine origin (Gravenor and Kupsch, 1959; Johnson and Clayton, 2003) they would need to have exceptional preservation potential that is not seen at contemporary margins (Evans, 2009;

Roberts et al., 2009). Furthermore, if they were controlled moraines then this would suggest that the CAIS margin was cold based (Evans, 2009), which is not reflected in the landform record as shown by both large and small scale mapping (Fig. 10 & 20).

The overall imprint highlighted in the glacial geomorphology maps (Fig. 10 & 20) is confidently interpreted as active marginal recession of the HPIS and CAIS following advance episodes (Evans et al., 1999; Evans and Twigg, 2002; Evans, 2003). Thus, the individual ridges within type 3 ridge sets are consistent with active temperate margins (Evans et al., 1999; Evans and Twigg, 2002; Evans, 2003) and are also interpreted as push moraines formed during seasonal re-advance. Therefore, the margins of both ice streams did not stagnate after this re-advance episode and remained warm based during recession. During the advance episode it seems possible that the englacial structures formed through ice marginal compression against the inclined proglacial slope which acted as a linear control on hummock emplacement (Evans, 2009). However the resulting linear hummocks contained minimal ice cores due to the active nature of the margin and as such linearity was preserved. Furthermore, the linear nature of the landform imprint will have been accentuated by seasonal re-advance of the margin during which time the transverse ridges were formed (Evans, 2003). If the margin remained warm based then is seems likely that some hummocks will have been amalgamated and destroyed during seasonal re-advance and push ridge production. However, it is also possible that winter freeze on (Evans and Heimstra, 2005) helped develop englacial structures and preserve hummocks (Evans, 2009).

4.2.4. Crevasse squeeze ridges

Crevasse squeeze ridges have been located within hummocky terrain and along the trunk of the CAIS (Figs. 11 & 18) and are strong evidence for surging activity (Sharp, 1985a, b; Evans and Rea, 1999; 2003; Evans et al., 2007). The orientation of the ridges in the McGregor moraine sequence suggests they formed by ice flowing west to east and so are tentatively interpreted to be part of Hfs_4. The corollary is that Hfs_4 represents a surge of the HPIS and the surrounding hummocks formed through ice stagnation and downwasting of supraglacial debris (Sharp, 1985a; Evans and Rea, 1999; 2003). The juxtaposition of crevasse squeeze ridges and CAfs_2 are also interpreted as evidence for ice stream surging. The suite of landforms is directly compatible with the landforms of contemporary surging forelands and therefore represents a surging landsystem (Evans and Rea, 1999; 2003). The association between ice streaming and surging will be considered in the following chapter.

4.3. Hummocky terrain

Three types of hummocks were identified within the CAIS marginal area. Types 1 and 2 comprise over 99% of the hummocks and were found in both type 3 ridge sequences and the numerous hummock bands (Figs. 20, 25 & 26). Type 3 hummocks were occasionally found within hummocky terrain bands (Figs. 20 & 28) but no examples were found in type 3 ridge sequences. The presence of all three types of hummock juxtaposed with each other suggests that they all formed though similar origins.

4.3.1. Hummocks (Type 1)

4.3.1.1. CAIS marginal area

Type 1 hummocks represent the largest proportion of hummocky terrain within the CAIS marginal area. Concentrations of type 1 hummocks occur around the Del Bonita highlands and in the lobate bands of hummocks north of Etzikom Coluée (Fig. 20) also known as the Lethbridge moraine (Stalker, 1977). Previous work in Alberta (Gravenor and Kupsch, 1959; Stalker, 1960; Bik, 1969) has identified that a significant proportion of type 1 hummocks are composed of till. Johnson and Clayton et al. (2003) argue that due to the identical morphology of till hummocks to those composed of collapsed lake sediment and outwash, it is likely they are also of supraglacial origins. The Red Creek (b) exposure (Fig. 36) shows that a low amplitude hummock capping the section is composed of horizontal laminations. These laminations are consistent with glaciolacustrine settings (Shaw, 1977; Fitzsimons, 1992; Campbell et al., 2001) and so are interpreted to represent collapsed lake sediments formed in a supraglacial position and let down during marginal retreat (Johnson and Clayton, 2003). As mentioned above (Section 4.2.3) several processes are likely to have operated at the CAIS margin, producing large concentrations of englacial and supraglacial debris. In addition to these it also likely that basal freeze (Boulton, 1972, Evans, 2009) and supercooling (Alley et al., 1998; 1999; Evenson et al., 1999; Evans, 2009) further enhanced the debris load at the margin.

Based on the resemblance to southern Laurentide and Fennoscandian ice sheet margins, the presence next to ice walled lake plains and the likelihood of a debris laden margin, type 1 hummocks are interpreted as supraglacial landforms (Gravenor and Kupsch, 1959; Clayton, 1967; Boulton, 1967, 1972; Parizek, 1969; Clayton and Moran, 1974; Eyles, 1979, 1983; Krüger, 1983; Paul, 1983; Clayton et al., 1985; Sollid and Sørbel, 1988: Johnson et al., 1995; Ham and Attig, 1996; Patterson, 1997, 1998; Mollard, 2000; Johnson and Clayton, 2003; Jennings, 2006; Knudsen, 2006). However, their presence next to active recessional landforms (Fig. 20 & 25) suggests that type 1 hummocks are not restricted to stagnant ice margins. Differential melting and debris quantities will have controlled the formation of the hummocks, creating individual, irregular shaped hummocks of varying sizes. The general low amplitude character of type 1 hummocks could likely be a product of the poorly drained, clay nature of the surficial sediments and debris supply (Clayton, 1967; Boulton, 1972; Johnson and Clayton, 2003). It is also likely that some subglacial pressing occurred at the margin of the CAIS (Stalker, 1960; Eyles et al., 1999; Boone and Eyles, 2001) due to the heavily saturated sediments that will have been common (Klassen, 1989). In agreement with Mollard (2000) and Boone and Eyles (2001) it is likely that hummocky terrain is a polygenetic landform of both subglacial and supraglacial origins, with supraglacial processes dominating within the CAIS marginal area.

The presence of internally chaotic hummock bands, such as the Lethbridge moraine and those around Del Bonita are interpreted to represent ice marginal stagnation and a cold based margin. The overall lobate form shares a strong resemblance to the controlled moraine described by Evans (2009) and hummock assemblages along the southern Laurentide margins (e.g. Landsystem C, Colgan et al., 2003; Johnson and Clayton, 2003). This in turn suggests that the CAIS margin was cold based and the lack of any linearity is a product of ice cores during emplacement (Evans, 2009). Therefore, the overall landform imprint of the CAIS marginal area is one of a polythermal margin. Numerous studies along the southern and south western Laurentide lobes have reconstructed significant permafrost concentrations at the margins of the lobes (e.g. Clayton et al., 2001, Bauder et al., 2005) and so it seems highly probable that such conditions will have prevailed at the margin of CAIS at some point during deglaciation. No surficial periglacial evidence was identified within the CAIS marginal area but several generations of ice wedge casts around Del Bonita (Bednarski, personal communication) and around the Cypress Hills (Westgate, 1968) suggests that permafrost was discontinuous throughout the CAIS marginal area. Furthermore, the presence of well developed hummocky terrain at these locations supports a cold based margin and stagnation origin (Clayton et al., 2008). It is quite possible that permafrost development was hindered by the presence of proglacial lakes common in southern Alberta (Klassen, 1989; Cutler, 2006).

4.3.1.2. Southern Alberta

The hummocky terrain throughout southern Alberta is extensive and well developed, and as such it has been the subject of numerous investigations (Stalker, 1960, Munro-Stasiuk and Shaw, 1997; Eyles et al., 1999; Boone and Eyles, 2001; Evans et al., 2006a). Correlation between Figure 10 and previous mapping (Shetsen, 1984, 1987; Clark et al., 1996; Evans et al., 1999) shows that hummocky terrain is accurately located and that SRTM data is sufficient for overall identification of the landform type. Due to its position between corridors of fast flow it has also been termed ‘interlobate’ terrain (Evans et al., 2008); however, the more generic term ‘hummocky terrain’ is preferred here. The change from smoothed topography to hummocky terrain along the CAIS is interpreted as a change in subglacial regimes, and hence demarcates the flow paths of the ice streams (Dyke and Morris, 1988; Patterson, 1998; Evans et al., 2008; Ó Cofaigh et al., in press).

Hfs_4 and 5 (Fig. 12) flow into hummocky terrain known as the 'McGregor moraine', which is interpreted as the marginal zone of the HPIS and so the term 'interlobate' is misleading. A detailed investigation into the exact composition of the hummocky terrain surrounding McGregor Lake by Evans et al. (2006a) reveals that the hummocky terrain is composed of inset recessional push ridges, with flutings terminating at the ridges. This landsystem directly corresponds to contemporary active temperate margins in Iceland (Evans et al., 1999; Evans and Twigg, 2002; Evans, 2003; Evans et al., 2006a; Evans et al., 2008) and the recessional ridges along the HPIS trunk (Fig. 10). Reconstructed ice margins show that they were formed by ice flowing in from the north-west (Evans et al., 2006a), and so most likely represent the marginal area of Hfs_5. This implies that Hfs_5 encompasses a significantly larger area than reconstructed in Figure 12. Glaciotectonic evidence identified along the north shore of the Travers Reservoir, demonstrated that some linear hummocks and low amplitude ridges are in fact thrust block moraines (Evans et al., 2006a). This suggests that some linear hummocks and low amplitude ridges common throughout the CAIS margin area may have glaciotectonic origins. Furthermore, these landforms were formed by ice flow from the north east, which suggests that the CAIS advanced into this area after the HPIS had receded. It is clear that the HPIS and CAIS competed around this area during deglaciation (Fig. 40), creating a composite suite of landforms. This point is further illustrated by the presence of crevasse squeeze ridges within the same hummocky terrain band (Figs. 11 & 18).

The juxtaposition of chaotic ridge forms next to hummocks and within hummocky terrain bands in southern Alberta (Fig. 17c) implies that they also have supraglacial origins. They may represent crevasse fills and are present because of the debris charged margins that operated in this area. Alternatively, their morphology shares some resemblance to eskers and could represent esker networks that formed whilst ice stagnated and downwasted in these areas. Further research is required to fully understand these landforms.

The above examples clearly show that hummocky terrain in southern Alberta is more complex than illustrated in SRTM data sets. However, in order to produce regional glacial geomorphology maps from data sources such as SRTM these areas should continue to be recognized as generalised hummocky terrain areas until they have been investigated at a larger scale. Large scale mapping and sedimentary investigations are required to fully elucidate the landforms and their origins within hummocky terrain.

4.3.2. Doughnut Hummocks (Type 2)

Type 2 hummocks are interpreted as doughnut hummocks or ring forms that are common to many deglaciated forelands, for example southern Laurentide lobes and Europe (Gravenor and Kupsch, 1959; Parizek, 1969; Aartolahti, 1974; Lagerbäck, 1988; Boutlon and Caban, 1995; Mollard, 2000; Colgan et al., 2003; Knudsen, 2006). Evidence presented by Johnson and Clayton (2003) demonstrates that doughnut hummocks within Alberta and across North America are predominantly composed of clayey till, which they suggest is an important characteristic in their formation. Whilst several alternative theories have been proposed for their formation, all of which consider the landforms to have formed in a 'stagnant glacial regime' (Knudsen, 2006 p.161), they remain poorly understood (Evans, 2009). Their presence contiguous to push ridges contradicts this assumption and clearly demonstrates that they are not solely found at stagnant margins. Based on the interpretations for hummocks (type 1) and ice walled lake plains (type 3) it seems that doughnut hummocks must also have originated from supraglacial positions. They most likely formed via differential melting regimes of thick supraglacial debris, whereby debris flow into sink holes and irregularities in the surface formed multiple low relief landforms (Gravenor, 1955; Clayton, 1967; Mollard, 2000; Knudsen, 2006). Possible alternative origins are proglacial blow-out of over pressured fluids, previously documented in poorly drained sediments (Bluemle, 1993; Boutlon and Caban, 1995; Evans et al, 1999; Evans, 2003, 2009), or subglacial pressing of saturated sediments (Gravenor and Kupsch, 1959; Stalker, 1960; Aartolahti, 1974; Eyles et al., 1999; Mollard, 2000; Boone and Eyles, 2001).

The ridges identified within the hummock bands around Del Bonita (Fig. 28) are interpreted as different features to those described above (Section 4.3.1.2. Fig. 17c) because they are significantly smaller, occur individually and comprise part of hummock rims. They may be the product of sink holes that have become interconnected and so the rims of doughnut hummocks connect to form long ridges. Alternatively, it is also possible that whilst ice stagnated and remained ice cored for significant amounts of time, supraglacial streams developed along sink holes and hollows in the ice cored surface, eventually creating larger channels. After the ice cores melted and debris was let down, debris along these channels were left amongst the hummocks, impart comprising their ridges. Further study of their composition is required in order to accurately elucidate their origins.

4.3.3. Ice walled Lake Plains (Type 3)

Type 3 hummocks show an exact morphological resemblance to ice walled lake plains identified along the margins of the southern Laurentide lobes, in Minnesota, North Dakota, Wisconsin, Michigan, southern New England (Colgan et al., 2003; Clayton et al., 2008) and throughout Europe (Strehl, 1998; Knudsen et al., 2006). Strong evidence presented by Clayton et al. (2008) demonstrates that ice walled lake plains cannot be of subglacial origin based on molluscs present within ice walled lake deposits. Their presence therefore is strongly associated with supraglacial origins, the corollary of which strongly suggests that the surrounding hummocky terrain is also of supraglacial origin (Johnson and Clayton, 2003; Clayton et al., 2008). In agreement with the other hummocky and controlled moraine interpretations, the ice walled lake plains are also supraglacially derived landforms. Their dominant size is explained by continued development after ice recession due to a thick debris cover insulating the buried ice walls (Attig, 1993; Clayton et al., 2001; Attig et al., 2003; Clayton et al., 2008), hence why they are not located within the active recessional imprint of the CAIS marginal area The close association between ice walled lake plain development and permafrost (Attig, 1993; Clayton et al., 2001; Attig et al., 2003) is also evident within the CAIS marginal area. The largest ice walled lake plains are located around the Del Bonita Highlands, located almost exactly where permafrost features have been identified (Bednarski, personal communication). This evidence suggests that permafrost directly impacts upon the development of ice walled lake plains (Attig, 1993; Ham and Attig, 1996; Clayton et al., 2001; Attig et al., 2003; Clayton et al., 2008)

4.4. Glaciofluvial and Glaciolacustrine evidence

4.4.1. CAIS marginal area

Numerous gravel pits located south of Etzikom Coulée demonstrate evidence that is consistent with glaciofluvial outwash, in particular sandur deposits (Benn and Evans, 1998). Importantly, the gravel pits lie within the active recessional zone of the margin, which suggests that glaciofluvial activity was extensive within and around the margin during this time (Benn and Evans, 1998; Evans, 2003).

The lowermost boulder lag in the Etzikom Coulée section (Fig. 31) is a common attribute to the outer spillway zone model of Kehew and Lord (1986, 1987), however, the presence on the inside of the inner channel suggests that flow was either unable to carry the boulders or that they were laid down during meltwater discharge decline (Keyhew and Clayton, 1983; Keyhew and Lord, 1986, 1987). The strong north to south clast fabric orientation of the gravels within the Etzikom Coulée section suggests they were emplaced by southwards glaciofluvial flow (Evans and Benn, 2004). This contrast with the east to west orientation of the coulée implies that the gravels were deposited after spillway production. In addition the gravel pit is located within a set of hummocks located inside the coulée and so the overlying diamict must also have formed after spillway production. This evidence is consistent with a re-advance of the CAIS margin into the coulée.

The gravel units situated within the Milk River section (Fig. 32) are similarly interpreted as evidence of glaciofluvial activity with clast size changes reflecting fluctuations in discharge velocities (Benn and Evans, 1998). The clast fabric data (Fig. 32) suggests that proglacial outwash followed the natural slope within the CAIS margin area (Fig. 13E). The juxtaposition of gravel units that grade laterally and massive sand/silt units are consistent with outwash deposits most likely deposited at an active margin (Rust and Romanelli, 1975; Benn and Evans, 1998). Lateral variations in outwash deposits document alternating discharge velocities, sediment load and proximity of the ice margin (Rust and Romanelli, 1975; Benn and Evans, 1998). Furthermore, the massive sand/silt unit may be representative of subaqueous outwash formed by the rapid deposition of high sediment loads (Rust and Romanelli, 1975).

The lowermost unit exposed at the Writing on Stone section (Fig. 33) is horizontally laminated which is consistent with formation in a glaciolacustrine setting (Shaw, 1975; 1977; Fitzsimons, 1992), and is interpreted as evidence that a proglacial lake was located in this area. Based on clast size data the upper gravel units represent differing discharge velocities most likely a product of seasonal discharge (Benn and Evans, 1998).

The Etzikom, Chin, Forty Mile and Verdigris Coulées situated within the CAIS marginal area document the decanting of proglacial lakes common in southern Alberta (Stalker, 1962; Evans, 2000; Evans et al., 2008). The remarkable coherence with other meltwater channels and transverse ridges implies that the decanting water was controlled by the natural northwest to southeast slope (Figs. 3, 13A & E). The channels located north of the Lethbridge moraine and Chin Coulée drained eastwards around the Cypress Hills. The anomalous position of Forty Mile Coulée is explained by the pre-glacial Winnifred divide which forced east flowing meltwater to the south, south east (Westgate, 1968). Meltwater channels that formed south of Etzikom Coulée drained towards Pakowki Lake and into northern Montana.

The transverse channels are interpreted as ice marginal meltwater channels, which document the recession of the margin and that meltwater was forced from west to east by the proglacial slope. Channels orientated north to south such as those that dissect the hummocky terrain north of Etzikom Coulée (Fig. 20) flowed up the proglacial slope. This could only have occurred under significant pressure and so these channels are interpreted to have formed subglacially (Hooke and Jennings, 2006). Their dimensions and characteristics directly compare to tunnel valleys along the southern Laurentide lobes (Mooers, 1989; Clayton et al., 1999; Cutler, 2002; Hooke and Jennings, 2006). Furthermore, they are located within the Lethbridge moraine which is interpreted to represent a frozen margin, a key characteristic in tunnel valley formation (Hooke and Jennings, 2006). Therefore it seems quite possible that these channels represent tunnel valleys that formed under significant pressure whilst marginal drainage was impeded by discontinuous permafrost and a frozen margin (Hooke and Jennings, 2006).

4.4.2. Southern Alberta

Individual eskers and esker networks have been identified by Shetsen (1987, 1990) and Evans et al. (2008). As previously mentioned, only the largest esker forms were identifiable and so where individual eskers are in close proximity to each they are interpreted to represent an esker network. The two esker networks positioned along the central portion of the HPIS and CAIS document that subglacial water was concentrated along these points (Evans et al., 2008). Similarly, the distribution of eskers along (Fig. 10, CAfs_2) the flat terrain of the eastern flank of the CAIS also shows that subglacial water was concentrated in this area. Where eskers cross cut MSGLs (Fig. 10 & 14c) they are interpreted to have formed after lineation production most likely during ice marginal recession (Stokes and Clark, 2003a; Stokes et al., 2008). Similar evidence has been interpreted by Stokes and Clark (2003a) to suggest that retreat direction may not have been parallel to flow orientation.

Meltwater channels are extensive across the study area, forming both anabranched and en-echelon sequences along the ice stream pathways (Fig. 10 & 19). The large sequence of anabranching meltwater channels that cross cut the CAIS in the north of the study area (Fig. 10 & 19) were traditionally interpreted as proglacial spillway channels (Gravenor and Bayrock, 1955; Christiansen, 1979) but have more recently been unequivocally interpreted as subglacial in origin (Sjorgren and Rains, 1995; Evans et al., 2008). The large meltwater channel that runs parallel to CAfs_1 (Fig. 10 & 14b) is connected to the southern arm of this sequence of channels, which suggests that it formed during the same discharge episode and is also subglacial in origin. Meltwater channels situated at the margin of Efs_1 are also interpreted as spillways based on previous research (Evans, 2000; Ó Cofaigh et al., in press).

The origin of the channels along the western margin of the HPIS is unknown; however, it is possible that they are lateral channels that formed between higher topography of the Porcupine Hills to the west and HPIS to the east. Some of these channels have previously been interpreted as tunnel channels and evidence of megafloods in southern Alberta (Rains et al., 2002). Their size and form are similar to the spillway channels along the CAIS marginal area and so are consistent with large meltwater discharge landforms (Evans, 2000; Rains et al., 2002). Similarly, spillway channels (Milk River, Lost River, Canal Creek and Sage Creek) in the south east corner of the CAIS marginal area have been interpreted as tunnel valleys (Beaney, 2002). The channels are located down ice of the preglacial divide and evidence presented shares a resemblance to documented tunnel valley morphologies (Ó Cofaigh, 1996). Alternatively, reconstructed glacial Lake Pakowki (Westgate, 1968) is situated just up-ice of where the channels initiate, which strongly suggests they actually have a proglacial origin, formed by the decanting of glacial Lake Pakowki (Wesgate, 1968). Evidence of large discharges of meltwater in the landform record such as tunnel valleys (e.g. Attig et al., 1989; Clayton et al., 1999; Johnson, 1999; Cutler et al., 2002; Beaney, 2002; Rains et al., 2002) and spillways (Evans, 2000; Evans et al., 2008) are easily explained by localised drainage events. There seems to be no reason to invoke provincial scale megafloods for the production of these landforms (Beaney, 2002; Rains et al., 2002), especially when they can be more easily explained by other mechanisms.

It is also worth commenting on the extensive networks of channels that exist within hummocky terrain bands, in particular the Suffield moraine. Their overall shape resembles rivers within a catchment area and it seems likely that they formed as meltwater flowed under gravity rather than hydraulic gradients. The corollary is that they formed as ice stagnated in this area of hummocky terrain.

4.5. Lineations and MSGLs

4.5.1. Southern Alberta

In agreement with Evans et al. (2008) the smoothed corridors are interpreted as ice streams produced by ice flowing at higher velocities than the surrounding ice (Swithinbank, 1954; Bentley, 1987). Furthermore, comparison of glacial geomorphology mapped in Figure 10 against Stokes and Clark's (1999, 2001) geomorphological criteria for ice streaming shows a very strong correlation (Table 3).

The smoothed corridors demonstrate landforms synonymous with fast flow (Stokes and Clark, 2001) which in turn are surrounded and delineated by a change in the *"smoothness*" (Ó Cofaigh et al., in press, p. 5) to hummocky terrain associated with slow moving, cold based ice and stagnation (Dyke and Morris, 1988, Stokes and Clark, 2002, Evans et al., 2008; Ó Cofaigh et al., in press). The MSGLs and smoothed topography directly compare to previously identified palaeo ice streams (Patterson 1997, 1998; Stokes and Clark, 1999, 2001; Clark and Stokes, 2003; Jennings, 2006) and to the forelands of contemporary ice streams on the Antarctic Shelf (Shipp et al., 1999; Canals et al., 2000; Wellner et al., 2001; Ó Cofaigh et al., 2002). The onset zones of both the HPIS and CAIS are unknown and mapping by Prest et al. (1968) and Evans et al. (2008) do not identify any clear convergent flow patterns. Till pebble lithology data (Shetsen, 1984) also suggests that both the HPIS and CAIS demonstrate Boothia type dispersal. The MSGLs show that the HPIS was transitory in nature, but any shifts in flow orientation were relatively small. There are only a few MSGLs situated along the CAIS within the study area, but their differing orientations identify at least two separate flow events.

Further up ice of the study area large MSGL sets are found along the CAIS, for example the Athabasca Fluting Field (Shaw et al., 2000). Extensive glaciotectonic evidence has been demonstrated (Fig. 37, 39a & b) along the CAIS and HPIS and their marginal areas (Evans et al., 2008). Based on the reconstructed flow sets and landforms it seems clear that both the HPIS and CAIS represent a 'time-trangressive' ice stream (Clark and Stokes, 2003).

Based on the field area cross profile in Figure 13B and topographic maps (Geiger, 1967) the CAIS and the east lobe (O' Cofaigh et al., in press) are interpreted as 'pure' ice streams and the HPIS a 'topographic' Ice stream (Clark and Stokes, 2003). In particular, along its southern section (south of 51°N) the HPIS is continuously bound by the Rocky Mountain Foothills to the west and in parts to the east. The HPIS traversed across the natural slope in the High Plains, with orientations moving towards the slope direction along Hfs_2-5 (Fig. 12). It is quite possible that the natural slope will have exerted some control over flow direction. In agreement with Dyke et al. (2002) it is assumed that the Cordilleran and Laurentide ice did coalesce during the LGM. It is therefore likely that the Cordilleran ice exerted strong controls over the orientation of the HPIS, forcing it in a south easterly direction. This is highlighted by the different orientations of Hfs_1 and Cfs_1 and 2 (Fig. 12) which are interpreted to have sourced from the Laurentide and Cordilleran ice sheets respectively. Additionally, the 90° shift of the HPIS between Hfs_1 and 2 (Fig. 12) is positioned approximately where the Foothills Erratics train is located, which has been used to mark the location of ice sheet coalescence (Stalker, 1956; Jackson et al., 1997; Rains et al., 1999). The multiple flow-sets along the HPIS therefore document numerous small scale flow re-organisations and re-advances during deglaciation. Whether Hfs_6 and 7 (Fig. 12) document a radical shift in the orientation of the HPIS is unknown. However, it seems more likely that the western portion of the HPIS flowed around this higher topography as its marginal area spread out into a lobate form. Hfs_5 (Fig. 12) is composed of numerous lineations that on a small scale demonstrate strong spatial coherency. However, large scale mapping compiled by Evans et al. (2006a) identify cross cutting lineations which must have been formed during more than one flow event. This evidence therefore suggests that Hfs_5 represents more than one flow event and highlights

Ice Stream Geomorphological Criteria (Stokes and Clark, 1999, 2001)	CAIS	HPIS
1. Characteristic shape and Dimensions	YES	YES
2. Highly convergent flow patterns	Unknown	NO
3. Highly attenuated bedforms	YES	YES
4. Boothia type erratic dispersal train	YES	YES
5. Abrupt lateral margins	YES	NO
6. Ice stream marginal moraines	YES	YES
7. Glaciotectonic and geotechnical evidence of pervasively deformed till	YES	YES
8. Submarine till delta or sediment fan	NA	NA

Table 3: *CAIS and HPIS compared with ice stream geomorphology criteria proposed by Stokes and Clark (1999, 2001).*

the need for, and accuracy of, large scale mapping. As mentioned above the HPIS terminated at the McGregor moraine sequence but it also spread out in the south west corner of the study area highlighted by lineations, hummocky terrain and meltwater channels (Fig. 10). The coalescence of the HPIS and CAIS is reconstructed to have been located around the McGregor moraine, Lethbridge and Travers Reservoir.

Cfs_1 and 2 are composed of lineations varying from 1.3 - 4.8km and seem to have elongation ratios greater than 10:1, which would suggest that they formed under a fast flow regime. Interestingly, there are no documented ice streams that sourced from the south eastern portion of the Cordilleran ice sheet. Whether this is representative of an ice stream sourcing from the Rocky Mountains is unknown. Cfs_3 is known as the Morley Flats drumlin field (Fisher and Spooner, 1994) and is located along the Bow Valley. Cfs_4 overprints Cfs_5 (Fig. 12) and so is interpreted to have occurred after Cfs_5. All of these flow sets are interpreted to have formed by Cordilleran ice flowing out of the Bow Valley but do not record evidence of ice streaming based on the geomorphological criteria proposed by Stokes and Clark (1999, 2001).

Few lineations were identified along the CAIS but it is clear from those mapped that at least two advance phases are recorded in the landform record. This is based on the 45° shift in orientation between CAfs_1 and 2 (Fig. 12). Allied with this, mapping by Sjogren and Rains (1995) and Evans et al. (2008) identifies significantly more glacial lineations along the CAIS, and also seem to record two separate flow events. CAfs_1 (Fig. 10) lies up ice and perpendicular to the large lobate ridge (TR_8); due to their juxtaposition with each other it seems likely that TR_8 could represent the maximum position of this re-advance. Based on this evidence it is suggested that CAfs_2 represents an earlier flow episode of the CAIS and CAfs_1 represents ice streaming sometime after this. Furthermore, Evans (1996, 2000) interpreted the large lineation (35km long) within CAfs_1 (Figs. 10 & 12) as a streamlined esker, which suggests that CAfs_1 formed after esker formation, during a later flow event.

Efs_1 (Fig. 12) is situated within the south west marginal zone of the east lobe. The investigation by Ó Cofaigh et al. (in press) did not acknowledge this set of lineations and so it is tentatively proposed that Efs_1 represents fast flow of this portion of ice stream margin. It is likely that the flow-set was formed as the east lobe moved into the study area forming contemporaneously with the large transverse ridges (TR_11; Figs. 10 & 12).

4.5.2. CAIS marginal area

All lineations within the CAIS marginal area are located to the south and south east (Fig. 20) and are interpreted as MSGLs if they are greater than 6km in length (Clark et al., 2003). Where lineations overprint type 1 and type 2 transverse ridges to the south and south east of Lake Pakowki (Fig. 20) they represent a re-advance of the margin, most probably the Altawan advance of Kulig (1996) also known as the Wild Horse advance (Westgate, 1968). This advance is interpreted to have occurred between the Cypress Hills and 112°W, approximately 15km east of Del Bonita. The lineations run parallel to CAfs_3 and so based on their strong parallel coherency are interpreted to represent the same flow event. Lineation length gradually increases from north west to south east, trending into a several MSGLs within CAfs_3 (Fig. 20). All measured ERs within the CAIS marginal area are greater than the 10:1 threshold proposed by Stokes and Clark (2002), which implies they formed under fast flowing ice. The locations of CAfs_1, 2 and 3 (Fig. 12) on the down ice side of bedrock highs and at locations where the proglacial slope is away from the margin (Fig. 13A & D) suggests that topography may have been a controlling factor in their production. Similar lineation occurrences on the down ice side of higher topography are found within Hfs_5 on the Blackspring Ridge (Fig. 10) (Munro-stasiuk and Shaw, 2002) and the Athabasca fluting field in central Alberta (Shaw et al., 2000). This point was also touched on by Westgate (1968) where he mentioned the largest lineations only occur where the regional slope was away from the margin. If this were a factor in lineation and MSGL production then it would explain why there are so few lineations along the CAIS as the proglacial slope is predominantly inclined against the margin (Fig. 13A). This evidence is consistent with the groove ploughing theory for lineation production (Clark et al., 2003) whereby keels produced by bedrock bumps, in this case bedrock highs (Fig. 13A & D) carve grooves in the bed and deform sediments into the intervening ridges.

4.6. Deglacial style of the CAIS

The regional glacial geomorphology primarily records the deglacial dynamics of the western margin of the south west Laurentide Ice Sheet. From this evidence it is clear that during the Late Wisconsinan advance of the Laurentide Ice Sheet, ice flowed south through Alberta, then through the south eastern corner of Alberta to its maximum extent in northern Montana (Clayton and Moran, 1982; Dyke and Prest, 1987; Kulig, 1996; Dyke et al., 2002). During this time three ice streams competed and coalesced within southern Alberta. Within the study area the High Plains Ice Stream flowed in a south, south easterly direction; the CAIS flowed mainly in a south, south westerly direction with part of the ice stream flowing south easterly into Montana and the east lobe flowed in a south westerly direction from neighbouring Saskatchewan. During deglaciation from the Late Wisconsinan ice receded through southern Alberta depositing a wealth of ice marginal landforms.

The glacial geomorphology in southern Alberta is interpreted to have formed by ice streams and hence direct ice contact with the underlying substrate. Furthermore, the evidence presented here for ice marginal features is consistent with a wealth of literature that documents direct glacial action on the landscape at contemporary margins (e.g. Evans, 2003). Therefore, the interpretation that the smoothed corridors and their landforms reflect subglacial megaflood pathways (e.g. Rains et al., 1993; Shaw et al., 1996; Shaw, 2002) is rejected. Numerous flaws exist concerning the meltwater hypothesis, namely there is no attempt to group the landforms into a landsystem, rather the landforms are interpreted individually. Therefore, the hypothesis is unable to explain for example, why the megaflood would produce transverse ridges that lie down ice and perpendicular to MSGLs. Why are they lobate in form? How would the megaflood form three different types of hummocks, some of which lie contiguous to push ridges? In addition, the images produced to highlight the flow paths of these megafloods (Rains et al., 1993) show the flood direction opposite to the reconstructed ice flow direction in this research and by Ó Cofaigh et al. (in press) of the east lobe. The south westerly flow direction of the east lobe from Saskatchewan is indisputable and illustrates a major flaw with these reconstructions. Simply, the theory is not compatible with both the geomorphic and sedimentary evidence (Benn and Evans, 2006)

Based on the geomorphology mapped here and previous work on deglacial timelines of the southwest Laurentide Ice Sheet (Westgate, 1968; Clayton and Moran, 1982; Dyke and Prest, 1987; Kulig, 1996) a relative retreat map has been produced (Fig. 40). By around 14ka BP recession from the southern limit in Montana had reached into Alberta (Westgate, 1968; Clayton and Moran, 1982; Dyke and Prest, 1987; Kulig, 1996) and the CAIS margin occupied the Lethbridge moraine until approximately 12.3 ka BP (Clayton and Moran, 1982; Dyke and Prest, 1987). Stalker (1977) showed the Lethbridge moraine to be the classical Wisconsin ice limit; however, this limit is interpreted to coincide with Etzikom drift limit of Westgate (1968). In agreement with Liverman et al. (1989); Young et al. (1989, 1993); Burns et al. (1993) and Kulig (1996) the Late Wisconsinan was the most extensive to affect the study area.

Reconstructions in the area (Shetsen, 1987; Evans, 2000) have identified significant glaciolacustrine silts and clays throughout southern Alberta, for example in the Pakowki Lake depression, formed by glacial Lake Pakowki during ice margin recession (Westgate, 1968; Kulig, 1996). It is argued that a deep proglacial lake will likely have also been located here during the advance of the margin during LGM, contained between the preglacial divide and the ice margin. It is postulated that the presence of a proglacial lake destabilised the margin leading to a rapid advance over the preglacial divide (Dredge and Cowan, 1989; Evans and Ó Cofaigh, 2003; Stokes and Clark, 2004).

Westgate (1968) identified 5 distinct morphostratigraphic units each of which represents a re-advance limit in the south east Alberta (110° - 112°W; 49° - 50°N) based on petrography and morphology: Elkwater drift; Wild Horse drift; Pakowki drift; Etzikom drift and

Oldman drift. The Elkwater drift relates to the upper ice limit on the Cypress Hills. The Wild Horse drift extends into northern Montana where it terminates at a large 15-20m transverse ridge sequence and is interpreted to represent the final advance of the CAIS margin into Montana sometime around 14 ka BP. The Pakokwi drift (Fig. 40) is marked by the outer extent of the push moraines to the south east of Lake Pakowki and runs along the northern tip of the Milk River and north around the Cypress Hills (Wesgate, 1968; Bik, 1969; Kulig, 1996). Therefore, all landforms to the south of this point were formed during an earlier advance, most likely the Altawan advance (15ka BP; Kulig, 1996). The Pakowki advance (Fig. 40) not recognized in Christiansen's (1979) and Dyke and Prest's (1987) deglacial sequences, most likely occurred between 14-13.5ka BP (Kulig, 1996) and relates to Clayton and Moran's (1982) Stage F - H. The Etzikom drift limit is interpreted as the Lethbridge moraine limit of Stalker (1977) and is marked in Figure 20 by the broad band of hummocky terrain just north of Etzikom Coulée. This advance episode maintained its position along the Lethbridge moraine until around 12.3ka BP (Stage I, Clayton and Moran, 1982; Dyke and Prest, 1987; Kulig, 1996). The Oldman drift limit (Fig. 40) is found outside of the mapped CAIS marginal area, just south of the Oldman River. Importantly, the correlation between the thrust ridges by Travers Reservoir (Evans et al., 2006a) and the Oldman limit suggests that they were formed during this re-advance episode. The corollary is that the HPIS had already receded further to the north. This re-advance (Stage J – L, Clayton and Moran, 1982) most likely occurred just after 12ka BP. Based on the regional geomorphology map (Fig. 10) it is suggested that a further re-advance occurred (Vauxhall advance), the limit of which is marked by the Vauxhall Ridge and must have occurred sometime after 12ka BP. Evans (2000) suggests that the CAIS margin had receded to the north of the study area by 12ka BP. Based on the Vauxhall advance evidence the CAIS must have receded later than that proposed by Evans (2000). Importantly, Dyke and Prest (1987) have their margin to north of the study area by this time, and so this suggests that the CAIS may have remained within southern Alberta for longer than previously thought. The Vauxhall ridge is interpreted to mark the final re-advance of the CAIS after which time it receded rapidly (Evans, 2000). The exact timing of the HPIS and east lobe retreat is unclear; however, it seems likely that the HPIS had receded somewhere north of Bow River by 12ka BP. Kulig (1996) documented significantly different deglacial styles between the CAIS and the east lobe, with the east lobe being characterised by a stable margin. The difference in deglacial styles is possibly a product of differing subglacial regimes and dispersal centres.

The presence of at least three till units within CAIS marginal area and overprinted landforms highlights the complex nature of terrestrial ice stream marginal recession styles and their landform imprint. The morphological record of the CAIS margin is therefore a palimpsest of at least three re-advance episodes. This seeming cyclical behaviour of the CAIS is characteristic of contemporary surging margins (e.g. Sharp, 1985b; Raymond, 1987; Evans and Rea, 1999, 2003) and has been interpreted as evidence of surging activity in the Laurentide Ice Sheet (e.g. Clayton et al., 1985; Kulig, 1996). However, it is not known whether the re-advances were forced by increased ice streaming driven by subglacial mechanisms or external forcing. It is clear that subsequent re-advance limits were positioned further north each time, most likely due to overall ice sheet recession, rather than subglacial conditions.

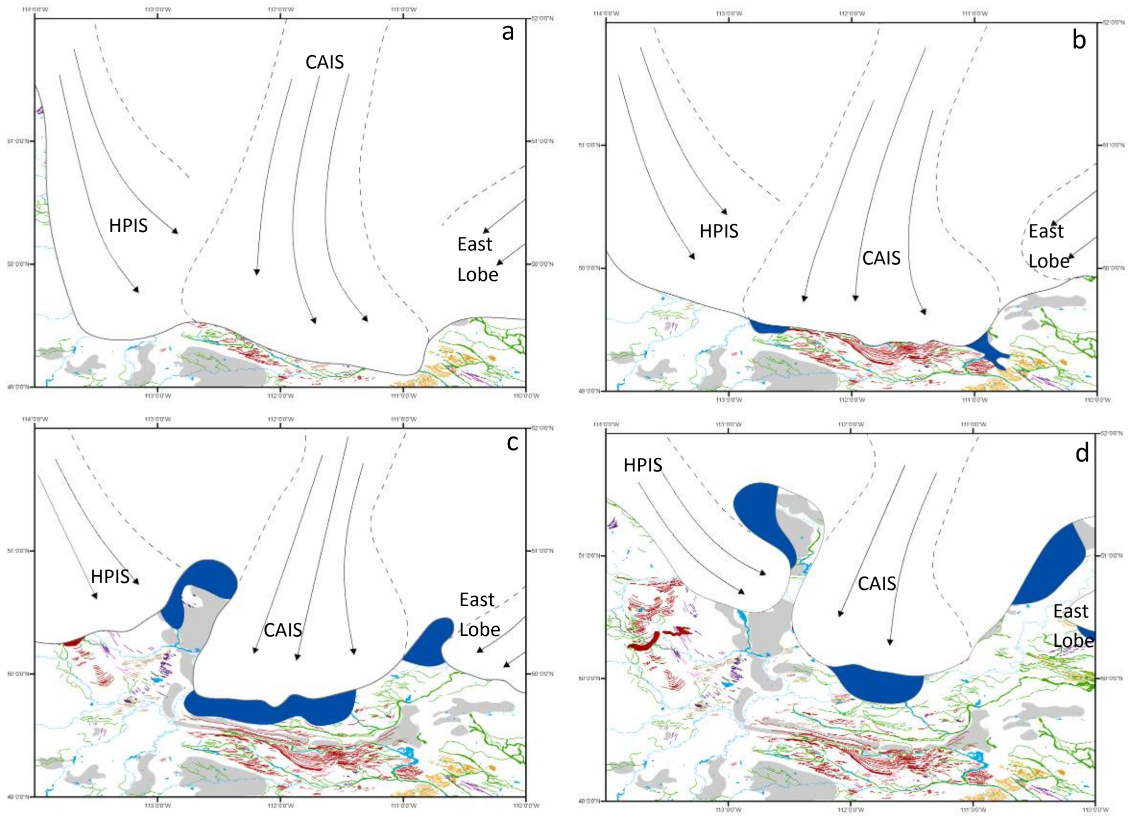


Figure 40: *Reconstructed deglacial limits in southern Alberta. a) Pakowki advance limit around 14-13.5ka BP (Westgate, 1968; Clayton and Moran, 1982; Dyke and Prest, 1987; Kulig, 1996). b) Etzikom limit located along the Lethbridge moraine maintained its position until 12.3ka BP (Clayton and Moran, 1982; Dyke and Prest, 1987; Kulig, 1996). c) Oldman limit around 12ka BP (Westgate, 1968; Clayton and Moran, 1982). d) Vauxhall limit possibly around 11.7ka BP (Clayton and Moran, 1982). The reconstructed position of the HPIS is based solely on geomorphology and so limits are quite speculative. The proglacial lakes are reconstructed from Westgate (1968), Shetsen (1987) and Evans (2000).*

Chapter 5

Discussion

5.1. Methods and Data Sources

5.1.1. Scale Approach (SA)

When reconstructing ice dynamics such as ice streaming and ice marginal settings it is crucial that we appreciate the variety of scales on which landforms are formed (Smith et al. 2006); for example MSGLs are more readily identifiable using small scale visualisation methods such as SRTM, but differentiating between types of hummocks requires data sets with much higher resolution. It therefore follows that mapping here has required the integration of data sets at multiple scales and that *"...successful landform identification is scale dependent*" (Smith et al., 2006, p.163). In order to fully incorporate this idea and hence gain the best possible objective results a 'Scale Approach' has been implemented as a framework to guide and structure this research. Its application has benefited the study in numerous ways; most importantly it enabled a holistic overview that highlighted the complex and fascinating nature of the glacial dynamics that operated in southern Alberta. In addition, it allowed independent verification of mapping techniques through multiple data sets, which in turn improved the accuracy of the mapping. The SA adopted within this research has proved to be a successful methodology with which to approach mapping at local and regional scales. Furthermore, its success within this research suggests that it would be appropriate for current research in glacial geomorphology due to the wide availability of data sets at numerous scales.

5.1.2. Mapping and Sedimentology

SRTM data alone provides an excellent source of 'reconnaissance' style mapping, identifying regional trends, landform suites and the larger individual landforms. The elevation data contained within the data sets, the ease of visualisation manipulations and their availability render SRTM data an excellent data resource for small scale mapping (Rabus et al., 2003). Without the use of SRTM data, many landforms would not have been identified within Landsat ETM+ data sets, for example the Vauxhall Ridge. Recent studies using solely SRTM data demonstrate the benefits of using SRTM (Ó Cofaigh et al., in press); however, problems arise when attempting to produce detailed maps due to the resolution constraints. This is well illustrated by the lack of any detail provided by SRTM data for the hummocky terrain areas within southern Alberta, which are a key component of the landsystem. This is almost completely resolved when SRTM data is used in conjunction with other data sets as commonly seen within contemporary studies (Glasser and Jansson, 2005; Bolch et al., 2005; Heyman et al., 2008).

Manipulations that are performed on the data must recognise and appreciate any bias they introduce (Smith and Clark, 2005); for example, though illuminating lineations perpendicular to their orientation such features are significantly accentuated within the landform record (Smith and Clark, 2005). This research has attempted to minimise such bias through using multiple illumination angles (Smith and Clark, 2005) and Landsat ETM+ as an independent verification tool (Glasser and Jansson, 2005). The Landsat ETM+ provided good comparison with SRTM mapping and also identified landforms that were at scales below the resolution of the SRTM data. However, low relief and subtle landforms were difficult or impossible to recognise and were only identified in SRTM data. Therefore, when used in conjunction they provide greater coverage and enable first order verification of each other.

Recent advances in mapping techniques now enable more accurate mapping than ever; for example, NEXTMap data produces DEMs with a spatial resolution of 5 m, yielding unprecedented results for the BRITICE project (Clark et al., 2004; Evans et al., 2005). However, data coverage of this quality and scale is rare and so aerial photographs still provide the most accurate data sets outside of specialist sources. The results gained from the aerial photo mapping are comprehensive, accurate and enabled significant insights to be gained from the landform record. This illustrates well that whilst mapping techniques will continue to improve, good quality aerial photographs will continue to provide one of the best data sources available for mapping glacial geomorphology.

The lack of sedimentary data relating to landform types within the CAIS marginal area has weakened the ability to accurately constrain their origins. Moreover, it weakens the ability to create process form models (Clayton and Moran, 1974) that can be applied to other landform assemblages. Unfortunately, good exposures through landforms within the ancient landform record are rare, and an inherent fact from which this research suffers.

5.2. Terrestrial Ice Stream Margin

In order to accurately model and interpret ambiguous evidence within the landform record it is crucial that we accurately constrain not only the individual landform type but the surrounding landform suite. Our ability to accurately model landsystems vastly increases our accuracy in identifying and interpreting ancient landscapes (Evans, 2003). However, care must be taken when applying modern analogues as certain landform types and assemblages are not found within contemporary glacier forelands, for example ice walled lake plains and doughnut hummocks (Johnson and Clayton, 2003). Moreover, the lack of any contemporary terrestrial ice streams hinders our ability to assess the extent to which ice marginal dynamics can affect ice stream flow (Jennings, 2006). It is worth reiterating at this point that the southern Laurentide margin contained several ice streams the margins of which have been interpreted to be represented by lobes. It follows that their landform assemblages are of significant importance with respect to this research (Patterson, 1997, 1998; Winsborrow et al., 2004; Jennings, 2006).

The conceptual terrestrial ice stream margin landsystem (Fig. 41) is based on the interpretations of the CAIS marginal area, with the landform types and assemblages sharing a strong resemblance to other southern Laurentide ice stream lobes. The extent to which the proglacial slope has affected landform development has been discussed in detail; however, the exact extent to which it impacted on the overall landsystem remains unclear. It seems highly likely that it exerted some control on the landsystem, with ice dynamics acting as the primary control. The model does not attempt to act as the single template for ice stream margins due to the wide range of landform assemblages found in such settings. Interestingly, the ice marginal landsystem includes elements of multiple landystems: active temperate (Evans and Twigg, 2002; Evans 2003), surging (Evans and Rea, 1999, 2003; Evans et al., 2007), landsystem C of the southern Laurentide (Colgan et al., 2003) and supraglacial (Johnson and Clayton, 2003). This implies that terrestrial ice stream margins undergo a variety of different dynamics and regimes over short spatial and temporal scales; for example Cutler et al. (2001) demonstrate different geomorphic and sedimentary imprints of three neighbouring southern Laurentide lobes. The ability of terrestrial ice stream margins to create a wide variety of landforms reflects the different settings and controls exerted on them, in particular bedrock topography and geology (Colgan et al., 2003). This in turn exerts strong controls on subglacial thermal and hydrology regimes and hence ice streaming (Anandakrishnan et al., 1998; Bell et al., 1998; Bennett, 2003; Bamber et al., 2006).

The active marginal signal of both the HPIS and CAIS (Evans et al., 2008) is not common to other terrestrial ice stream margins (Patterson, 1997, 1998; Colgan et al., 2003; Johnson and Calyton, 2003; van de Wateren, 2003; Jennings, 2006; Ó Cofaigh et al., in press). This characteristic suggests that the recession of the CAIS and HPIS may have been synchronous, which is in strong contrast to the asynchronous behaviour of the southern Laurentide Lobes (Jennings, 2006). However, the CAIS marginal area also records evidence for a cold based margin that may have persisted for some time. This evidence correlates with the other southern Laurentide ice stream margins described by Patterson (1997, 1998), Colgan et al. (2003), Johnson and Calyton (2003) and Jennings (2006). It seems that CAIS marginal area and other terrestrial ice stream margins demonstrate a continuum of landforms (Fig. 41) that are common to a variety of documented landsystems, which are primarily controlled by the basal thermal regimes operating at the margin. Therefore, at one end of this continuum there are active, warm based landforms, such as recessional push moraines and at the other cold based landforms such as controlled moraine.

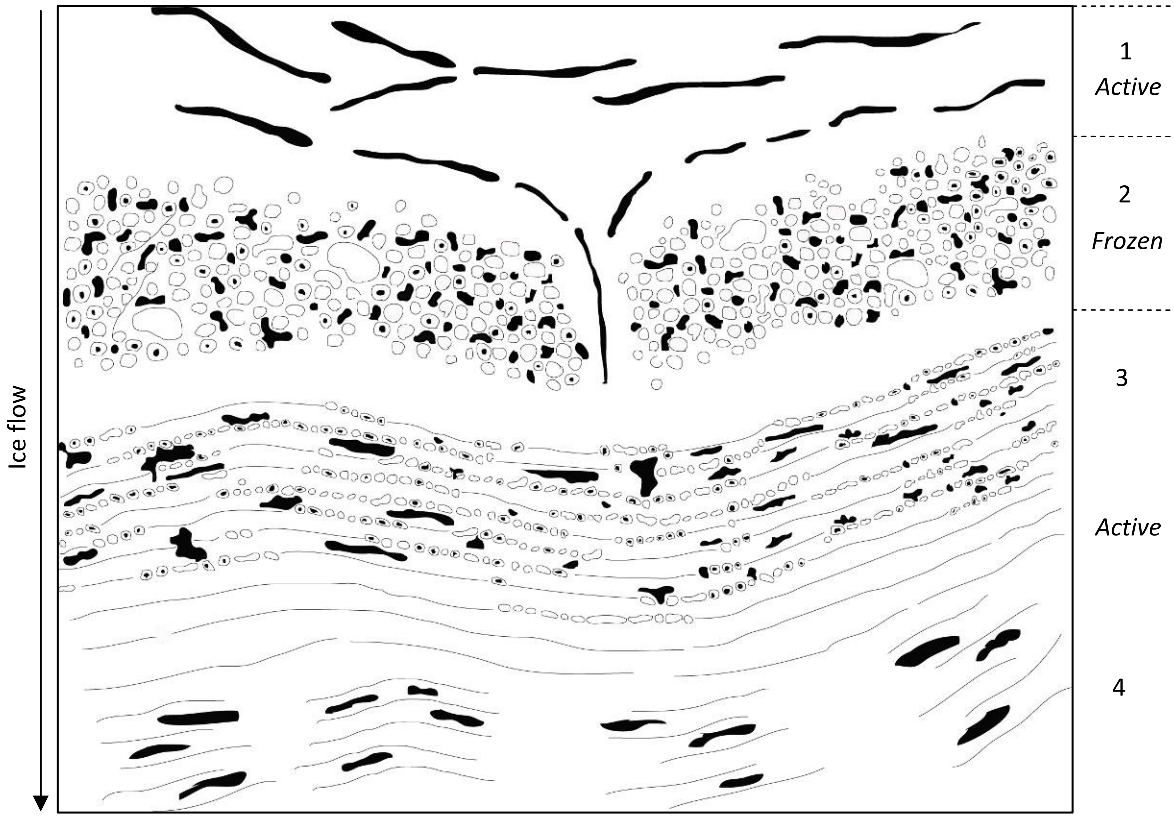


Figure.41: *A continuum of landforms created by terrestrial ice stream margins based primarily on the CAIS margin. Ice flow is from north to south. The ability of an individual margin to create such a wide variety of landforms and landform assemblages documents its susceptibility to a wide variety of subglacial, marginal and localised conditions.1) Recessional meltwater channels that run parallel to the reconstructed ice margin in much the same manner as the transverse ridges. 2) Hummocky terrain bands that demonstrate a large scale lobate form are interpreted to document a cold based ice margin that persisted for enough time to form extensive hummocky terrain and for ice walled lake plains to form. The largest hummocks represent ice walled lake plains. Tunnel channels are also a likely feature in such settings. 3) Active recessional imprint comprised of push ridges that blend into contiguous linear hummock bands. The hummocks are interpeted to originate from supraglacial positions and are aligned due to structures within the ice margin formed by thrusting and compression. This signal is accentuated by a winter re-advance producing the push ridges. 4) Thrust ridges are common at ice stream margins with high porewater pressures, weak bedrock, topographic obstacles and permafrost.*

It is important that a distinction be made between evidence for ice streaming and for surging in terrestrial marginal settings because they share similar landform types yet are associated with different ice dynamics (Jennings, 2006). Furthermore, differentiating between the styles of re-advance found in southern Alberta will help to confirm ice marginal dynamics and landform origins. Clark and Stokes (2003) comment on the problem of discriminating between surging and ice streaming suggesting that the term 'surging' be restricted to glaciers alone whereas 'ice stream' refers to fast flow within an ice sheet. However, if a terrestrial ice stream were to experience uncontrolled accelerated flow beyond the usual fast flow regime then this surely would constitute a surge. Problems arise with actually recognising such an occurrence in contemporary ice streams as flow velocities fluctuate on various timescales

(Bindscahdler et al., 2003) and so difficulties are found in modelling and identifying such an event (Radok et al., 1987; Anandakrishnan et al., 2001). The periodic and transitory nature of ice streams is well documented (Clark and Stokes, 2003) and so re-advances of the margin are here referred to as fast flow events; surging should only be used when the landsystem formed by a palaeo-ice stream directly reflects that of a surging glacier and not based on individual landform types.

The southern and south western Laurentide ice stream margins share a close affinity to contemporary surging glaciers (Evans and Rea, 1999, 2003; Jennings, 2006), in particular, large quantities of hummocky terrain and thrust ridges (Boulton et al., 1985; Clayton et al., 1985; Fisher et al., 1985; Clark, 1994b; Kulig, 1996; Marshall et al., 1996; Cutler et al., 2001; Colgan et al., 2003; Ó Cofaigh et al., in press). Thrust ridges such as those along the margin of the CAIS, east lobe and Superior lobe (Westgate, 1968; Shetsen, 1987; Jennings, 2006; Ó Cofaigh et al., in press) were most likely formed by a rapidly advancing margin (Sharp, 1985b; Evans and Rea, 2003) and so could represent either accelerated or fast flowing ice. Importantly, Jennings (2006) suggests that the presence of thrust ridges at the margin of the Superior Lobe are evidence that high velocities occurred all the way to the margin and so their presence is not necessarily evidence of a surging ice stream but rather fast flow. Reconstructions of permafrost along the southern Laurentide lobes suggest it was extensive (Clayton et al., 2001, Bauder et al., 2005) and its presence may have impacted upon thrust ridge formation in these areas (Aber et al., 1989; Jennings, 2006). Alternatively, discontinuous permafrost in southern Alberta (Westagte, 1968; Bednarski, Personal communication) implies that thrust ridge formation occurred through different mechanisms, and so their production is likely to occur through a variety of mechanisms at ice stream margins (Aber et al., 1989). They have also been documented in sub-polar environments where permafrost is present (e.g. Evans and England, 1991) and are not necessarily a product of fast flow (Evans and Rea, 2003). Hummocky terrain is common to almost all terrestrial ice stream margins along the southern and south western Laurentide ice sheet, in the form of well developed, aligned, chaotic and collectively lobate hummocks (Johnson et al., 1995; Ham and Attig, 1996; Patterson 1997; 1998; Colgan et al., 2003; Johnson and Clayton, 2003, Jennings, 2006; Clayton et al., 2008). Its presence is generally interpreted as evidence for ice marginal stagnation of debris rich ice (e.g. Colgan et al., 2003; Johnson and Clayton, 2003), which is easily explained at both ice stream and surging margins. Therefore, thrust ridges and hummocky terrain comprise a significant component of terrestrial ice stream margins and should not act as evidence of a surging ice stream.

Ó Cofaigh et al. (in press) suggest that the east lobe may have surged to its extent around Medicine Hat based on stagnation evidence and crevasse squeeze ridges overprinting

MSGLs. In a similar fashion poorly developed crevasse squeeze ridges are juxtaposed with MSGLs along the CAIS and lie up ice of hummocky terrain and glaciotectonic evidence (Evans et al. 2008). This strong similarity with the surging landsystem (Evans and Rea, 1999, 2003; Evans et al., 2007) suggests that a re-advance episode of the CAIS may have been a surge. In particular, crevasse squeeze ridges cannot form without longitudinal extentsion of the ice (Evans and Rea, 2003) which in turn suggests that the CAIS did not maintain its flow.

5.3. Ice stream dynamics

In order to try and understand the wider implications for the results and interpretations presented in the last two chapters, the idea of the ice stream 'life cycle' (Clark and Stokes, 2003) will be revisited with each element being considered in turn.

5.3.1. Onset and Location

The location of the onset zone of the HPIS and CAIS is unknown. Mapping by Evans et al. (2008) shows that the HPIS and CAIS likely originated from the same ice stream that flowed south through Alberta and split into two individual ice streams at around 54°N. Reconnaissance on Google Earth™ and mapping by Prest et al. (1968) and Evans et al. (2008) traces MSGLs northwards to at least Athabasca Lake, yet no obvious onset zone is visible. The HPIS is interpreted to be a product of the CAIS and so does not demonstrate any convergent flow or an onset zone. The CAIS is most likely to have initiated somewhere in northern Alberta based on the mapping by Prest et al. (1968) and Evans et al. (2008). It is also quite probable that the onset zone was not preserved during deglaciation (De Angelis and Kleman, 2008). Further research is required to resolve the location of the onset of the CAIS.

Early research recognised that if contemporary ice sheets contained ice streams then palaeo-ice sheets most likely also had similar features (Denton and Hughes, 1981). The location of ice streams within the Laurentide Ice Sheet has been strongly associated with the availability of a deforming bed (e.g. Alley, 1991; Clark, 1992, 1994a; Marshall et al., 1996; Patterson, 1998) and partly topographic controls (Marshall et al., 1996; Kaplan et al., 2001). Evidence from contemporary ice streams also suggests that geological controls are a key factor in ice stream location (Anandakrishnan et al., 1998; Bell et al., 1998). However, more recently evidence presented by Stokes and Clark (2003a, b) suggests that deformable beds are not essential for ice streaming to occur. Importantly, the CAIS and HPIS are located over the Western Canadian Sedimentary Basin, which contains substantial areas of soft bed geology (Mathews, 1974; Boulton and Jones, 1979; Klassen, 1989). Numerous other fast flow pathways in the south western corner of the Laurentide Ice Sheet are also located over the Canadian

Sedimentary Basin (Prest et al., 1968; Evans et al., 2008; Ó Cofaigh et al., in press). Therefore, whilst ice streaming may occur in a variety of geological settings, the location of ice streaming in the south west corner of the Laurentide Ice Sheet was primarily controlled by the availability of a deformable bed.

5.3.2. Ice streaming: temporal and spatial behaviour

The CAIS, HPIS and east lobe flowed over a deformable bed which is composed of Cretaceous and Tertiary sediments, consisting of poorly consolidated clay, sand and silt (Stalker, 1960; Klassen, 1989). The Cretaceous beds in particular are weak due to a high bentonite content, which is reflected by the quantity of thrust features within southern Alberta. Combined with this, poor drainage conditions caused by swelling clays will have almost certainly created elevated porewater pressure. Furthermore, Beaty (1990) comments on localised differences in drainage conditions but where bentonic clays were present the underlying geology became almost impermeable. Therefore, weak and poorly drained sediments will have created low shear stresses and high porewater pressures that made ice streaming and fast flow in southern Alberta very likely (Stalker, 1973; Clayton et al., 1985; Fisher et al., 1985; Klassen, 1989; Clark, 1994a; Evans et al., 2008). The corollary is that the geologic setting can exert strong controls on the location and flow dynamics of ice streams (Anandakrishnan et al., 1998; Bell et al., 1998; Bamber et al., 2006).

Bedrock highs will likely have acted as some resistance to flow (e.g. Alley, 1993; Joughin et al., 2001; Price et al., 2002; Stokes et al., 2007) and caused localised compression, highlighted by the presence of thrust ridges at these locations. It is also suggested that the inclined proglacial slope along the CAIS (Fig. 13A) will have impacted on ice marginal dynamics through acting as a restraint to ice marginal advance, causing significant marginal compressive flow and enabling proglacial lakes to form (Klassen, 1989). This is illustrated by numerous exposures that show glaciotectonic activity, well developed controlled moraine, large spillway channels that record the decanting of proglacial lakes and significant distribution of glaciolacustrine sediments (Shetsen, 1987; Evans, 2000; Evans et al., 2008).

Unfortunately, it is difficult to ascertain whether fast ice motion occurred through deformation or sliding. Numerous till units and up ice thickening till wedges within southern Alberta (Westagte, 1968; Evans and Campbell, 1992; Evans et al., 2008) are consistent with the theory of subglacial deformation (Alley, 1991; Boulton, 1996a, b). However, work by Piotrowski and Kraus (1997), Piotrowski et al. (2001; 2004) suggests that subglacial deformation is not widespread but concentrated to localised spots along the bed. Furthermore, Evans et al. (2008) argue that the presence of large subglacial channels and thin tills overlying stratified sediments along the CAIS trunk provides evidence to suggest deformation was subordinate to sliding. No one process can be confidently outlined as a primary flow mechanism, rather, the evidence demonstrates that the CAIS flowed through some combination of deformation and sliding.

5.3.3. Shutdown

Results presented here and mapping by Prest et al. (1968) and Evans et al. (2008) highlights multiple flow sets throughout Alberta, recording dynamic and transitory ice streams that re-organised themselves on numerous occasions during deglaciation from the LGM. In agreement with Ó Cofaigh et al. (in press) it is quite possible that the eventual shutdown of the CAIS and HPIS during deglaciation subsequently caused them to re-organise their flow regime north of the study area and shift in direction (Prest et al., 1968; Evans et al., 2008). Within southern Alberta there is evidence of five re-advance episodes of the CAL between c. 15-12ka BP in the form of multiple till units and varying ice marginal landform assemblages (Wesgate, 1968; Clayton and Moran, 1982; Kulig, 1996). Each re-advance limit of the CAIS is located further north each time, which suggests that overall climate warming during this period impacted upon ice stream extent (Dyke et al., 2003). The overall landform imprint of the CAIS margin is a palimpsest of multiple re-advance episodes, which record primarily active recession but also cold based dynamics. In order to understand the overall deglacial dynamics of the CAIS two key elements are addressed: Why did ice streaming stop? What caused the CAIS to periodically re-advance?

5.3.3.1. Why did the CAIS shutdown?

That the presence of an ice stream draws down the surface profile and elevation of its drainage basin and hence a portion of the ice sheet is widely acknowledged (e.g. Bentley, 1987; Clark, 1994a; Bamber et al., 2000; Kaplan et al., 2001; Stokes and Clark, 2001; Bennett, 2003; Hulbe and Fahnestock, 2004) and that ice streams flowing through and into Alberta also had low driving stresses (Bentley, 1987; Bennett, 2003). It follows that a terrestrial ice stream margin will be susceptible to changes in basal thermal regimes and that this particular characteristic makes ice streams prone to abrupt changes in flow regimes and direction (Hulbe and Fahnestock, 2004).

Active marginal recession is recorded by the CAIS margin, which shows that the margin of the CAIS was warm based (Hart, 1999; Evans and Twigg, 2002; Evans, 2003). The progressively shorter re-advance episodes suggest that the CAIS recorded both large and small scale climate impacts. It follows that the re-advance limit was likely not controlled by ice marginal dynamics as suggested by Jennings (2006) but processes acting up ice of the margin. This may have been due to a lack of ice flux in the catchment area from which the ice stream sourced (Johannesson, et al., 1989) which in turn meant it was unable to continue flowing. Possible reasons for the decline in ice flux could be due to deglaciation and ice sheet re-organisation from the LGM (Dyke et al., 2003), or, possibly shifting ice sheds in the south western sector of the Laurentide Ice Sheet due to changes in ice sheet configuration and geometry caused by ice stream draw down and climate (Bamber et al., 2000; Stokes and Clark, 2001; Dyke et al., 2003). The extent and response of the drainage basin to changing ice sheet configurations (Dyke et al., 2003) from which the CAIS sourced is unknown. Their asynchronous nature means that other drainage basins are likely to have responded in a different manner depending on their individual characteristics (Johannesson et al., 1989; Ó Cofaigh et al., 2008); which in turn suggests that ice streams will record a variety of deglacial styles, one control of which is the size of its drainage basin (Ó Cofaigh et al., 2008). Joughin et al. (1999) document shared source regions for ice streams in West Antarctica, which suggests that the amount of ice being used by an individual ice stream can change over time. Similarly, Joughin et al. (1999) and Anandakrishnan et al. (2001) suggest that ice piracy may account for the shutdown of ice stream C. This concept may be applicable in Alberta, as the presence of neighbouring ice streams such as the HPIS and the Saskatchewan ice streams (Ò Cofaigh et al., In Press) may have sourced greater ice flux during deglaciation causing the CAIS to shut down.

Whilst ice stream dynamics may change through external or marginal controls it is ultimately the conditions at the ice sheet bed that actively shut the ice stream down (Anandakrishnan et al., 2001). The availability of subglacial water at high pressure is crucial for ice streaming and so any changes will affect ice stream dynamics (Parizek et al., 2002; Vogel et al., 2003; Bennett, 2003). Joughin and Tulaczyk (2002) and Jennings (2006) suggest that any changes in flow, advance or decline are controlled by mechanisms at the ice stream bed and not by reduced ice availability through climate change. In particular, Jennings (2006) associates drainage of the lobe with its decline and eventual halt. The presence of large spillways in SE Alberta represents rapid drainage of proglcaial lakes (Evans et al., 2006a), during decanting they will also have drained the margin causing the lobe to slow or stop (Jennings, 2006). This rapid drainage may have been sufficient to cause the margin to freeze to the bed as is reflected by the Lethbridge moraine. Furthermore, the advance of the CAIS may have been sufficient to cause a thermal change in bed conditions due to thin ice being closer to the bed, leading to its decline (MacAyeal, 1993a, b; Payne, 1995).

The large continuous lobate bands of hummocky terrain that demarcate the Lethbridge moraine limit are interpreted as controlled moraine (Section 4.2.3) and as such are evidence for a frozen margin (Evans, 2009). This evidence is also consistent with that of the southern Laurentide lobes, where frozen margins were common features (Attig et al., 1989; Ham and Attig, 1996; Clayton et al., 2001; Colgan et al., 2003), creating broad, lobate hummock bands (Patterson, 1997; 1998; Colgan et al., 2003; Johnson and Clayton, 2003; Jennings, 2006). It is possible that freeze on of the margin occurred after ice stagnation (Colgan et al., 2003) and so is either a product of processes further up ice of the margin, or, it could have caused the CAIS to slow and shutdown. Freeze on would most likely have occurred through an increase in the vertical temperature gradient, vertical and horizontal advection of cold ice and a reduction in porewater pressure due to the thinning and rapid advance of the margin (Alley et al., 1997; Tulaczyk et al., 2000a; Christoffersen and Tulaczyk, 2003; Hulbe and Fahnestock, 2004). Furthermore, thermal processes have also been suggested for the decline of ice stream C in Antarctica (Anandakrishnan et al., 2001; Bennett, 2003).

5.3.3.2. What caused the CAIS to re-advance?

The periodic re-advances of the margin show that the CAIS did not shutdown completely, rather, conditions periodically enabled the ice stream to readvance. Such readvances were most likely driven by subglacial and marginal dynamics (de Angelis and Kleman, 2005; Jennings, 2006), the most important of which will have been the availability of subglacial water at pressure (Parizek et al., 2002; Vogel et al., 2003; Bennett, 2003). What drove the re-advance episodes is poorly understood, however, it is possible that the presence of proglacial lakes facilitated ice stream advance (Dredge and Cowan, 1989; Evans, 2000; Evans and Ó Cofaigh, 2003; Stokes and Clark, 2004). This is supported by the widespread presence of glaciolacustrine sediments north of Chin Coulée (Shetsen, 1987; Pawlowicz and Fenton, 1995; Evans, 2000).

Proglacial lakes significantly impact on ice marginal dynamics in particular basal heat balance and subglacial drainage (Cutler, 2006). Importantly, work by Stokes and Clark (2004) suggests that the presence of a large proglacial lake initiated the Dubawnt Lake ice stream. It seems possible that during recession of the margin the development of proglacial lakes, enhanced by glacioisostatic depression (Evans, 2000) and discontinuous permafrost blocked drainage routes and caused porewater pressure to rise (Cutler, 2006). This in turn may have raised porewater pressure up ice and destabilised the margin. The water body therefore acted as a *"trigger" (Stokes and Clark, 2004, p.170)* to ice marginal advance.

Chapter 6

Conclusion

SRTM data sets confirm that the glacial geomorphology of southern Alberta is dominated by imprints of ice streams which sourced from the Keewatin dome of the Laurentide Ice Sheet. SRTM data provides an invaluable tool with which to map small (ice stream) scale glacial geomorphology but is best supplemented with other, higher resolution data. Based on previous research (Evans et al., 2006a; Evans et al., 2008) and the geomorphological criteria set out by Stokes and Clark (2001) both the HPIS and CAIS are confidently interpreted as terrestrial palaeo-ice streams. Mapping by Prest et al. (1968), Evans et al. (2008) and Ó Cofaigh et al. (in press) show that ice streams continued to operate within Alberta and neighbouring Saskatchewan during the recession of the south western Laurentide ice sheet from the Late Wisconsinan. Further north individual ice streams likely had short life cycles documented by numerous 'rubber-stamped' (Clark and Stokes, 2003) ice stream pathways in this area and their shutdown likely led to the onset of a new ice stream (Ó Cofaigh et al., in press). The location of ice streams seem to have been primarily controlled by the widespread availability of soft bed geology in the form of the Canadian Sedimentary Basin (Klassen, 1989). Furthermore, the general south to east flow orientation of ice streaming in this area is a product of a regional slope from west to east and the coalescence of the Laurentide and Cordilleran Ice Sheets deflecting flow.

The HPIS and CAIS were dynamic and transitory in nature creating a smudged imprint that is highlighted by multiple flow-set reconstructions and typical of time-transgressive ice streams (Clark and Stokes, 2003). Ice stream dynamics were directly affected by the geology and topography over which the ice streams flowed (Anandakrishnan et al., 1998; Bell et al., 1998; Bamber et al., 2006), with bedrock highs likely acting as a trigger for thrust ridge and lineation production (Aber et al., 1989; Tulaczyk et al., 2001; Clark et al., 2003). Evidence left by the CAIS in the form of till wedges that thicken down ice and large subglacial meltwater channels suggests that the ice stream flowed by some combination of sliding and deformation (Evans et al., 2008). The CAIS recorded several re-advance episodes during overall ice sheet recession which is thought to be a product of external climate forcing affecting ice stream drainage basins and fluctuating subglacial and marginal dynamics.

Reconstructed ice marginal dynamics show that conditions prevailing at the margin can have significant affects on the health of an ice stream (Jennings, 2006). In particular, where frozen toe zones exist it is quite possible that the margin will cause ice streaming to slow and even shutdown (Christoffersen and Tulaczyk, 2003). The CAIS marginal area is composed of controlled, hummocky, push and thrust block moraines, along with doughnut hummocks, ice walled lake plains, recessional meltwater, tunnel and large spillway channels. The CAIS margin seems to have been polythermal in nature, with ice thickness acting as a strong control on the basal thermal regime of the margin. The alternating thermal nature of the margin created a continuum of landforms that is dominated by active marginal recession between 15-12 ka BP (Westgate, 1968; Clayton and Moran, 1982; Kulig, 1996; Dyke et al., 2003). Similarly, the HPIS also documents active recession of its margin.

Lobate hummock bands document a frozen outer margin and are representative of controlled moraine (Evans, 2009). The presence of well developed ice walled lake plains within hummocky terrain and controlled moraine bands indicates that buried ice persisted long after deglaciation due to insulation from thick supraglacial debris (Clayton et al., 2008). Furthermore, the best developed ice walled lake plains are located where permafrost features are present, suggesting that its presence helped in their development (Ham and Attig, 1996; Clayton et al., 2001; Attig et al., 2003; Clayton et al., 2008).

In comparison to other terrestrial ice stream margins (Patterson, 1997; 1998; Jennings, 2006, Ó Cofaigh et al., in press) it seems that hummocky terrain bands and large thrust ridges are the most common landform types found in such setting. However, no other documented terrestrial ice stream margins demonstrate active recession in the form push ridges. It therefore seems that active recession lies at one extreme of the terrestrial ice stream margin landform continuum, with frozen outer limits and stagnation found at the other.

6.1. Limitations and recommended further research

This research has suffered from a lack detailed sedimentary data concerning constructional processes within the CAIS marginal area, and so has relied heavily on the geomorphology. Whilst this is not necessarily a weakness the veracity of the findings would be strengthened by additional sedimentary evidence. A recommendation for future research would be to further explore the field area with a sole view to documenting sedimentary exposures, in order to provide greater insight into the glacial history, dynamics and hence reconstructing terrestrial ice stream marginal behaviour. Depending upon the location of the exposures in relation to landforms this further research could also target landform constructional processes, specifically hummocky terrain and transverse ridges. Moreover, it could then be used to critically assess the findings detailed here and other theories for landform genesis.

This research has started to develop a detailed model for terrestrial ice stream marginal landsystems. The model now requires stringent testing and comparison with other such landsystems, and so ongoing research should aim to accurately constrain landform sediment assemblages in other terrestrial ice stream marginal settings. This would build the data and are knowledge of such setting and enable testing and development of the model. Furthermore, further research in this area is crucial to understanding and answering what controls and impacts an ice stream margin exerts on its longevity and dynamics.

Interestingly, the proglacial slope of the CAIS was inclined against the margin. This undoubtedly will have affected the dynamics that operated at the margin. One suggestion for further research is to focus on the nature of the topography at terrestrial ice stream margins and compare the types on landforms and landsystem created there. This would aim to further elucidate the impact such topographical differences could have had on marginal dynamics. Helpfully, this would be relatively easy to investigate.

Whilst significantly more difficult than the other ideas proposed here, it is important that further research considers the wider impacts of terrestrial ice streams on the dynamics and stability of ice sheets. Continuing research should attempt to model the potential impacts of terrestrial ice streams such as the CAIS/HPIS on drainage basin size, location and ability to change and in doing so overall ice sheet morphology and stability. This would go some way to understanding the overall role terrestrial ice streams and potentially their margins played in the stability, morphology and eventual decline of the last great ice sheets.

As shown here SRTM data is very useful for small scale, regional mapping. Its use in glacial geomorphology is still relatively new, and so it requires further testing using a variety of examples on different scales of focus. A recommendation for further research is that SRTM data be quantitatively critiqued with respect to glacial geomorphology. This could perhaps be done in similar fashion to Smith and Clark (2005), using a variety of data sets, looking at several landform assemblages and directly comparing the results.

References:

Aartolahti, T. 1974. Ring ridge hummocky moraines in northern Finland. *Fennia* 134: 1–21.

Aber, J.S., Ber, A. 2007. *Glaciotectonism*. Developments in Quaternary Science 6. Elseveir, London.

Aber, J.S., Croot, D.G., Fenton, M.M. 1989. *Glaciotectonic landforms and structures*. Kluwer Academic, Boston.

Alley, R.B. 1991. Deforming-bed origin for southern Laurentide till sheets? *Journal of Glaciology* 37: 67-76.

Alley, R.B. 1993. In search of ice-stream sticky spots. *Journal of Glaciology* 39 (133): 447–454.

Alley, R.B. 2000. Continuity comes first: recent progress in understanding subglacial deformation. In, Maltman, A.J., Hubbard, M.J., Hambrey, M.J. (Eds.) Deformation of Glacial Material. *Geological Society Special Publication* 176: 171-180.

Alley, R.B., Bindschadler, R. (eds.) 2001. *The West Antarctic Ice Sheet: Behaviour and Environment*. Antarctic Research Series 77: 283-294. American Geophysical Union, Washington DC

Alley, R.B., Blankenship, D.D., Bentley, C.R., Rooney, S.T. 1986. Deformation of till beneath Ice Stream B, West Antarctica. *Nature* 322: 57-59

Alley, R.B., Blankenship, D.D., Bentley, C.R., Rooney, S.T. 1987. Till beneath Ice Stream B, 4. A coupled ice–till flow model. *Journal of Geophysical Research* 92: 8921– 8929.

Alley, R.B., Blankenship, D.D., Rooney, S.T., 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.

Alley, R.B., Cuffey, K.M., Evenson, E.B., Strasser, J.C., Lawson, D.E. & Larson, G.J. 1997. How glaciers entrain and transport basal sediment: physical constraints. *Quaternary Science Reviews 16:* 1017–1038.

Alley, R.B., Lawson, D.E., Evenson, E.B., Strasser, J.C., Larson, G.J., 1998. Glaciohydraulic supercooling: a freeze-on mechanism to create stratified debris-rich basal ice: II. Theory. *Journal of Glaciology* 44: 562–568.

Alley, R.B., Strasser, J.C., Lawson, D.E., Evenson, E.B., Larson, G.J. 1999. Glaciological and geological implications of basal-ice accretion in overdeepenings. In, Mickelson, D.M., Attig, J.W. (eds.) *Glacial Processes Past and Present*. Geological Society of America Special Paper, 337: 1–9.

Anandakrishnan, S., Alley, R.B. 1994. Ice Stream C, Antarctica, sticky spots detected by microearthquake monitoring. *Annals of Glaciology* 20: 183–186.

Anandakrishnan, S., Alley, R.B., Jacobel, R.W., Conway, H. 2001. The flow regime of Ice Stream C and hypotheses concerning its recent stagnation. In, Alley, R.B., Bindschadler, R. (Eds.) *The West Antarctic Ice Sheet: Behaviour and Environment*. Antarctic Research Series 77: 283-294. American Geophysical Union, Washington DC.

Anandakrishnan, S., Blankenship, D.D., Alley, R.B., Stoffa, P.L. 1998. Influence of subglacial geology on the position of a West Antarctic ice stream from seismic observations. *Nature* 394(6688): 62-66.

Attig, J.W. 1993. Pleistocene geology of Taylor County, Wisconsin. *Wisconsin Geological and Natural History Survey Bulletin* 90: 25.

Attig, J. W. & Clayton, L. 1986b. Late Wisconsin continuous permafrost in northern Wisconsin ended about 13,000 years ago. *Geological Society of America Abstracts with Programs 18*: 278.

Attig, J.W., Clayton, Lee, Johnson, M.D., Patterson, C.J., Ham, N.R., Syverson, K.M. 2003. Ice-walled-lake plains in the mid-continent - what they tell us about late glacial ice marginal processes and environments. *Geological Society of America Program and Abstracts 35*: 61.

Attig, J.W., Michelson, D.M., Clayton, L. 1989. Late Wisconsin landform distribution and glacier bed conditions in Wisconsin. *Sedimentary Geology* 62: 399-405.

Atwater, B.F. 1986. Pleistocene glacial-lake deposits of the Sampoli River valley, northeastern Washington. *United States Geological survey, Bulletin*, 1661: 39.

Ayslworth, J.M., Shilts, W.W. 1989. Bedforms of the Keewatin Ice Sheet, Canada. *Sedimentary Geology* 62: 407-428.
Bamber, J.L., Ferraccioli, F., Joughin, I., Shepherd, T., Rippin, D.M., Siegert, M.J., Vaughan, D.G. 2006. East Antarctic ice stream tributary underlain by a major sedimentary basin. *Geology* 34(1): 33-36.
Bamber, J.L., Vaughan, D.G., Joughin, I. 2000. Widespread complex flow in the interior of the Antarctic Ice Sheet. *Science* 287: 1248-1250.
Bauder, A., Mickelson., D.M., Marshall, S.J. 2005. Numerical modeling investigations of the subglacial conditions of the southern Laurentide ice sheet. Annals of Glaciology 40: 219-224.
Beaney, C.L., 2002. Tunnel channels in southeast Alberta, Canada: evidence for catastrophic channelized drainage. *Quaternary International* 90: 67-74.
Beaney, C.L., Hicks, F.E., 2000. Hydraulic modelling of subglacial tunnel channels, south-east Alberta, Canada. *Hydrological Processes* 14: 2545-2557.
Beaney, C.L., Shaw, J. 2000. The subglacial geomorphology of southeast Alberta: evidence for subglacial meltwater erosion. *Canadian Journal of Earth Sciences* 37: 51–61
Beaty, C.B. 1990. Milk River in Southern Alberta: A classic underfit stream. *Canadian Geographer,* 34(2): 171-174.
Bell, R.E., Blankenship, D.D., Finn, C.A., Morse, D.L., Scambos, T.A., Brozena, J.M., Hodge, S.M. 1998. Influence of subglacial geology on the onset of a West Antarctic ice stream from aerogeophysical observations. *Nature* 394(6688): 58–62.
Bell, R.E., Studinger, M., Shuman, C.A., Fahnestock, M.A., Joughin, I. 2007. Large subglacial lakes in East Antarctica at the onset of fast flowing ice streams. *Nature* 445(7130): 904–907.
Benn, D.I. 1994. Fluted moraine formation and till genesis below a temperate glacier: Slettmarkbreen, Jotunheimen, Norway. *Sedimentology* 41: 279-292.
Benn, D.I., Evans, D.J.A. 1996. The interpretation and classification of subglacially-deformed material. *Quaternary Science Reviews* 15: 23-52.
Benn, D.I., Evans, D.J.A. 1998. *Glaciers & Glaciation*. Arnold, London, p.431-437.
Benn, D.I., Evans, D.J.A. 2006. Subglacial megafloods: outrageous hypothesis or just outrageous? In, Knight, P.G. (ed.), *Glacier Science and Environmental Change.* Blackwell: Oxford; 42–46.
Bennett, M.R. 2003. Ice streams as the arteries of an ice sheet: their mechanics, stability and significance. *Earth Science Reviews* 61: 309–339.
Bennett, M.R., Hambrey, M.J., Huddart, D., Glasser, N.F., Crawford, K.R. 1999. The landform and sediment assemblages produced by a tidewater glacier surge in Kongsfjorden, Svalbard. *Quaternary Science Reviews* 18: 1213-1246.
Bentley, C.R. 1987. Antarctic ice streams: a review. *Journal of Geophysical Research* 92: 8843–8858.
Bik, M. J. J. 1969. The origin and age of the prairie mounds of southern Alberta, Canada. *Biuletyn Peryglacjalny* 19: 85-130.
Bindschadler, R.A., Bamber, J., Anandakrishnan, S. 2001. Onset of streaming flow in the Siple Coast Region, West Antarctica. In, Alley, R.B., Bindschadler, R.A. (Eds) *The West Antarctic Ice Sheet: Behaviour and Environment. American* Geophysical Union, Antarctic Research Series 77: 123-136.
Bindschadler, R., Chen, X., Vornberger, P. 2000. The onset area of Ice Stream D, West Antarctica. *Journal of Glaciology* 46(152): 95-101.
Bindschadler, R.A., King, M.A., Alley, R.B., Anandakrishnan, S., Padman, L. 2003. Tidally Controlled Stick-Slip Discharge of a West Antarctic Ice Stream. *Science* 22: 1087-1089.
Bindschadler, R.A., Stephenson, S.N., MacAyeal, D.R., Shabataie, S. 1987. Ice dynamics at the Mouth of Ice Stream B, Antarctica. *Journal of Geophysical Research* 92: 8885-8894.
Bluemle, J.P. 1993. Hydrodynamic blowouts in North Dakota. In, Aber, J.S. (ed.) *Glaciotectonics and Mapping Glacial Deposits*. Canadian Plains Research Centre, University of Regina, p. 259-266.

Bluemle, J.P., Clayton, L. 1984. Large-scale glacial thrusting and related processes in North Dakota. *Boreas* 13: 279-299.
Bolch, T., Kamp, U., Olsenholler, J. 2005. Using ASTER and SRTM DEMs for studying geomorphology and glaciations in high mountain areas. In, Oluić, M. (ed.) *New Strategies for European Remote Sensing,* Millipress, Rotterdam: p119-127.
Bond, G.C., Lotti, R. 1990. Iceberg discharges into the North Atlantic on millennial time scales during the last glaciations. *Science* 267: 1005-1009.
Boone, S.J., Eyles, N. 2001. Geotechnical model for great plains hummocky moraine formed by till deformation below stagnant ice. *Geomorphology* 38: 109-124.
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. 1967. The development of a complex supraglacial moraine at the margin of Sørbreen, Ny Friesland, Vestspitsbergen. *Journal of Glaciology* 6: 717-735.
Boulton, G.S. 1970. On the origin and transport of englacial debris in Svalbard glaciers. *Journal of* Glaciology 9: 213–229.
Boulton, G.S. 1972. Modern arctic glaciers as depositional models for former ice sheets. *Journal of the Geological Society of London* 128: 361-393.
Boulton, G.S. 1976. The origin of glacially-fluted surfaces: observations and theory. *Journal of Glaciology* 17: 287-309.
Boulton, G.S. 1982. Subglacial processes and the development of glacial bedforms. In Davidson-Arnott, R., Nickling, W., Fahey, B.D. (Eds) *Research in Glacial, Glacio-fluvial, and Glacio-lacustrine systems.* Geo Books, Norwich, 1-31.
Boulton, G.S. 1987. A theory of drumlin formation by subglacial sediment. In Menzies, J., Rose, J., (eds.), *Drumlin symposium.* Rotterdam: Balkema.
Boulton, G.S. 1996a. The origin of till sequences by subglacial sediment deformation beneath mid-latitude ice sheets. *Annals of Glaciology* 22: 75–84.
Boulton, G.S. 1996b. Theory of glacier erosion, transport and deposition as a consequence of subglacial sediment deformation. *Journal of Glaciology* 42: 43–62.
Boulton, G.S., Caban, P.E. 1995. Groundwater flow beneath ice sheets: Part II – its impact on glacier tectonic structures and moraine formation. *Quaternary Science Reviews* 14: 563–587.
Boulton, G.S., Clark, C.D. 1990. A highly mobile Laurentide Ice Sheet revealed by satellite images of glacial lineations. *Nature* 346: 813–817.
Boulton, G.S., Hindmarsh, R.C.A. 1987. Sediment deformation beneath glaciers: rheology and geological consequences. *Journal of Geophysical Research* 92: 9059-9082.
Boulton, G.S., Jones, A.S. 1979. Stability of temperate ice sheets resting on beds of deformable sediments. *Journal of Glaciology* 24: 29-43.
Boulton, G.S., Dobbie, K.E., Zatsepin, S. 2001. Sediment deformation beneath glaciers and its coupling to the subglacial hydraulic system. *Quaternary International* 86: 3-28.
Boulton, G.S., Smith, G.D., Jones, A.S., Newsome, J. 1985. Glacial geology and glaciology of the last mid-latitude ice sheets. *Journal of the Geological Society of London* 142: 447-474.
Bourgeois, O., Dauteuil, O., Vliet-Lanoë, B.V. 2000. Geothermal control on flow patterns in the Last Glacial Maximum ice sheet of Iceland. *Earth Surface Processes and Landforms* 25 (1): 59-76.
Bridgland, D.R. 1986. *Clast lithology analysis*. Quaternary Research Association, Cambridge, p.207.
Briner, J.P. 2007. Supporting evidence from the New York drumlin field that elongate subglacial bedforms indicate fast ice flow. *Boreas* 37: 143-147.
Burns, J.W., Young, R.R., Arnold, L.D. 1993. *Don't look fossil gift horses in the mouth, they, too, have tongues of ice*. Geological Association of Canada Mineralogical Association of Canada Program with Abstracts, Joint Annual Meeting, Edmonton, pp.A14.
Campbell, J.E., Fenton, M.M., Pawlowicz, J.G. 2001. Surficial geology of the Wandering River area, Alberta (NTS 83P/SE). *Alberta Geological Society*, Map 242.

Canals, M., Urgeles, R., Calafat. 2000. Deep sea floor evidence of past ice streams off the Antarctic Peninsula. *Geology* 28 (1): 31-34.
Christiansen, E.A. 1979. The Wisconsinan deglaciation of southern Saskatchewan and adjacent areas. *Canadian Journal of Earth Sciences*, 16: 913-938.
Christofferson, P., Tulaczyk, S. 2003. Signature of palaeo-ice stream stagnation: till consolidation induced by basal freeze-on. *Boreas* 32: 114–129.
Clague, J.J. 1989. Relationship of the Cordilleran and Laurentide glaciers. In, Fulton, R.J. (ed.) *Quaternary Geology of Canada and Greenland*, Geological Survey of Canada, Geology of Canada no. 1, pp. 42-43.
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. 1994b. Large scale ice-moulded landforms and their glaciological significance. *Sedimentary Geology* 91: 253–268.
Clark, C.D. 1997. Reconstructing the evolutionary dynamics of former ice sheets using multi-temporal evidence, remote sensing and GIS. *Quaternary Science Reviews* 24: 1533-1541.
Clark, C.D. 1999. Glaciodynamic context of subglacial bedform generation and preservation. *Annals of Glaciology* 28: 23–32.
Clark, C.D., Stokes, C.R. 2001. Extent and basal characteristics of the M'Clintock Channel Ice Stream. *Quaternary International* 86: 81–101.
Clark, C.D., Stokes, C.R. 2003. Palaeo-ice stream landsystem. In, Evans, D.J.A. (ed.) *Glacial Landsystems.* Arnold, London, p.204–227.
Clark, C.D., Evans, D.J.A., Khatwa, A., Bradwell, T., Jordan, C., Marsh, S.H., Mitchell, W.A., Bateman, M.D. 2004. Map and GIS database of glacial landforms and features related to the last British Ice Sheet. *Boreas* 33: 359–375.
Clark, C.D., Knight, J.K., Gray, J.T. 2000. Geomorphological reconstruction of the Labrador Sector of the Laurentide Ice Sheet. *Quaternary Science Reviews* 19: 1343–1366.
Clark, C.D., Tulaczyk, S.M., Stokes, C.R., Canals, M. 2003. A groove-ploughing theory for the production of mega scale glacial lineations, and implications for ice-stream mechanics. *Journal of Glaciology* 49(165): 240–256.
Clark, P.U. 1992. Surface form of the southern Laurentide Ice Sheet and its implications to ice-sheet dynamics. *Geological Society of America Bulletin* 104: 595-605.
Clark, P.U. 1994a. Unstable behaviour of the Laurentide Ice Sheet over deforming sediment and its implications for climate change. *Quaternary Research* 41: 19-25.
Clark, P.U., Licciardi, J.M., MacAyeal, D.R., Jenson, J.W. 1996. Numerical reconstruction of a soft-bedded Laurentide Ice Sheet during the last glacial maximum. *Geology* 24: 679–682.
Clarke, G.K.C., Collins, S.G., Thompson, D.E. 1984. Flow, thermal structure, and subglacial conditions of a surge type glacier. *Canadian Journal of Earth Science* 21(2): 232-240.
Clarke, G.K.C., Leverington, D.W., Teller, J.W., Dyke, A.S., Marshall, S.J. 2005. Fresh arguments against the Shaw megaflood hypothesis. A reply to comments by David Sharpe on ''Paleohydraulics of the last outburst flood from glacial Lake Agassiz and the 8200 BP cold event''. *Quaternary Science Reviews* 24: 1533-1541.
Clayton, L. 1967. Stagnant-glacier features of the Missouri Coteau in North Dakota. *North Dakota Geological Survey,* Miscellaneous Series 30: 25-46.
Clayton, L., Cherry, J.A. 1967. Pleistocene superglacial and ice walled lakes of west-central North America. *North Dakota Geological Survey Miscellaneous Series* 30: 47–52.
Clayton, L., Moran, S.R. 1974. A glacial process-form model. In Coates, D.R. (ed.) *Glacial Geomorphology*. State University of New York, Binghamton, 89-119.
Clayton, L. and Moran, S.R. 1982. Chronology of Late Wisconsinan glaciations in middle North America. *Quaternary Science Reviews*, v. 1, p. 55-82.
Clayton, L., Attig. J.W., Michelson, D.M. 1989. Evidence against pervasively deforming bed material beneath rapidly moving lobes of the southern Laurentide Ice Sheet. *Sedimentary Geology* 62: 203-208.
Clayton, L., Attig. J.W., Michelson, D.M. 1999. Tunnel channels in Wisconsin. In: Michelson, D.M.,

Attig, J.W. (Eds.), *Glaciers Past and Present*. Geological Society of America Special Paper 337, pp. 69-82.
Clayton, L., Attig. J.W., Michelson, D.M. 2001. Effects of late Pleistocene permafrost on the landscape of Wisconsin, USA. *Boreas* 30: 173-188.
Clayton, L., Attig, J.W., Ham, N.R., Johnson, M.D., Jennings, C.E., Syverson, K.M. 2008. Ice-walled-lake plains: Implications for the origin of hummocky glacial topography in middle North America. *Geomorphology* 97: 237–248.
Clayton, L., Teller, J.T., Attig, J.W. 1985. Surging of the southwestern part of the Laurentide Ice Sheet. *Boreas* 14: 235-241.
Croot, D.G. 1988. Glaciotectonics and surging glaciers, a correlation based on Vestspitsbergen, Svalbard, Norway. In, Croot, D.G. (ed.) *Glaciotectonics: forms and processes*. Balkema, Rotterdam, p49-61.
Colgan, P. M. 1996. *The Green Bay and Des Moines Lobes of the Laurentide Ice Sheet: Evidence for Stable and Unstable Glacier Dynamics, 18000 to 12 000 BP*. 293 pp. Ph.D. dissertation, University of Wisconsin, Madison
Colgan, P.M., Mickelson, D.M., Cutler, P.M. 2003. Ice marginal terrestrial landsystems: southern Laurentide ice sheet margin. In, Evans, D.J.A. (ed.) *Glacial Landsystems*, Arnold, London: p111-142.
Colton, R.B., Lemke, R.W., Lindvall, R.M. 1961. Glacial Map of Montana East of the Rocky Mountains. *U.S. Geological Survey Miscellaneous Investigations Series Map I-327*, scale 1:500 000, 1 sheet.
Colton, R.B., Fullerton, D.S. 1986. Proglacial lakes along the Laurentide Ice Sheet margin in Montana. *Geological Society of America Abstracts with Programs* 18 (5), 347.
Cutler, P.M. 2006. Modelling Impact of Glacier-permafrost Interaction on Subglacial Water Flow. In, Knight, P.G. (ed.) *Glacier Science and Environmental Change*. Blackwell, Oxford, p67-68.
Cutler, P.M., Colgan, P.M., Mickelson, D.M. 2002. Sedimentologic evidence for outburst floods from the Laurentide Ice Sheet margin in Wisconsin, USA: implications for tunnel-channel formation. *Quaternary International* 90: 23-40.
Cutler, P. M., MacAyeal, D. R., Mickelson, D. M., Colgan, P.M. 1998. A numerical investigation of ice-lobe–permafrost interactions around the southern Laurentide Ice Sheet. *Geological Society of America Abstracts with Programs 30*, 112.
Cutler, P.M., MacAyeal, D.R., Mickelson, D.M., Parizek, B.R., Colgan, P.M. 2000. A numerical investigation of ice-lobe–permafrost interaction around the southern Laurentide Ice Sheet. *Journal of Glaciology* 46: 311–325.
Cutler, P.M., Mickelson, D.M., Colgan, P.M., MacAyeal, D.R., Parizek, B.R. 2001. Influence of the Great Lakes on the dynamics of the southern Laurentide ice sheet: numerical experiments. *Geology* 29: 1039-1042.
Cutler, P. M., Mickelson, D. M., MacAyeal, D. R., Colgan, P.M. 1999. Permafrost around the southern Laurentide Ice Sheet: its influence on ice dynamics, landform genesis and water drainage. *Geological Society of America Abstracts with Programs 31*A, 204.
Dardis, G.F., McCabe, A.M. 1983. Facies of subglacial channel sedimentation in late Pleistocene drumlins, Northern Ireland. *Boreas* 12: 263-278.
Davies, N.K., Locke, W.W., Pierce, K.L., Finkel, R.C. 2006. Glacial Lake Musselshell: Late Wisconsin slackwater on the Laurentide ice margin in central Montana, USA. *Geomorphology* 75: 330-345.
Dawson, G.M. 1875. *Report on the geology and resources of the region in the vicinity of the Forty-Ninth parallel from Lake of the woods to the Rocky Mountains*; British North American Boundary Commission, Montreal, p.379.
De Anglais, H., 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., Kleman, J. 2008. Palaeo-ice stream onsets: examples from the north-eastern Laurentide Ice Sheet. *Earth Surface Processes and Landforms* 33: 560-572.
Denton, G.H., Hughes, T.J. 1981. *The Last Great Ice Sheets*. Wiley, New York.

Dowdeswell, J.A., Ottesen, D., Rise, L. 2006. Flow switching and large scale deposition by ice streams draining former ice sheets. *Geology* 34 (4): 313-316.

Dowling, D.B. 1917. The southern plains of Alberta. *Geological Survey of Canada* 93: 200.

Dredge, A.S., Cowan, W.R. 1989. Quaternary geology of the southwestern Canadian shield. In, Fulton, *Quaternary Geology of Canada and Greenland*. Geological Survey of Canada, Geology of Canada no. 1, pp. 214-235.

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

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

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

Dyke, A.S., Savelle, J.M. 2000. Major end moraines of Younger Dryas age on Wollaston Peninsula, Victoria Island, Canadian arctic: implications for palaeoclimate and for formation of hummocky moraine. *Canadian Journal of Earth Sciences* 37: 601–619.

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

Dyke, A.S., Moore, A., Robertson, L. 2003. Degaciation of North America: *Geological Survey of Canada*, Technical Report Open File 1574.

Dyke, A.S. Morris, T.F., Green, D.E.C., England, J., 1992. Quaternary geology of Prince of Wales Island, Arctic Canada. *Geological Survey of Canada Memoir* 433.

Dyson, J.L. 1952. Ice-ridged moraines and their relation to glaciers. *American Journal of Science* 250: 204-212.

Echelmeyer, K.A., Harrison, W.D., Larsen, C., Mitchell, J.E. 1994. The role of the margins in the dynamics of an active ice stream. *Journal of Glaciology* 40 (136): 527–538.

Engelhardt, H., Kamb, B. 1998. Basal sliding of Ice Stream B, West Antarctica. *Journal of Glaciology* 44: 223-230.

Ekland, A., Hart, J.K. 1996. Glaciotectonic deformation within a flute from the Isfallsglaciären, Sweden. *Journal of Quaternary Science* 11: 299-310.

Evans, D.J.A. 1996. A possible origin for a megafluting complex on the southern Alberta prairies, Canada. *Zeitschrift für Geomorphologie* Suppl. Bd. 106: 125-148.

Evans, D.J.A. 2000. Quaternary geology and geomorphology of the Dinosaur Provincial Park area and surrounding plains, Alberta, Canada: the identification of former glacial lobes, drainage diversions and meltwater flood tracks. *Quaternary Science Reviews* 19: 931–958.

Evans, D.J.A. 2003. Ice-marginal terrestrial landsystems: active temperate glacier margins. In, Evans, D.J.A. (ed.), *Glacial Landsystems*. Arnold, London: p. 12–43.

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

Evans, D.J.A., Benn, D.I. 2004. *A practical guide to the study of glacial sediments*, Arnold, London.

Evans, D.J.A., Campbell, I.A. 1992. Glacial and postglacial stratigraphy of Dinosaur Provincial Park and surrounding plains, southern Alberta, Canada. *Quaternary Science Reviews* 11: 535–555.

Evans, D.J.A., Campbell, I.A. 1995. Quaternary stratigraphy of the buried valleys of the lower Red Deer River, Alberta, Canada. *Journal of Quaternary Science* 10: 123–148.

Evans, D.J.A., England, J. 1991. Canadian Landform Examples 19: high arctic thrust block moraines. *Canadian Geographer* 35: 93-97.

Evans, D.J.A., Hiemstra, J.F. 2005. Till deposition by glacier submarginal, incremental thickening. *Earth Surface Processes and Landforms* 30: 1633–1662.

Evans, D.J.A., Ó Cofaigh, C. 2003. Depositional evidence for marginal oscillations of the Irish Sea ice stream in southeast Ireland during the last glaciations. *Boreas* 32: 76-101.

Evans, D.J.A., Rea, B.R. 1999. Geomorphology and sedimentology of surging glaciers: a landsystems approach. *Annals of Glaciology* 28: 75-82.

Evans, D.J.A., Rea, B.R. 2003. Surging glacier landsystem. In *Glacial Landsystems*, Evans, D.J.A. (ed.). Arnold: London; 259–288.

Evans, D.J.A., Twigg, D.R. 2002. The active temperate glacial landsystem: a model based on Breiðamerkurjökull, Iceland. *Quaternary Science Reviews* 21: 2143-2177.

Evans, D.J.A., Clark, C.D., Mitchell, W.A. 2005. The last British Ice Sheet: a review of the evidence utilized in the compilation of the Glacial Map of Britain. *Earth Science Reviews* 70: 253-312.

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

Evans, D.J.A., Lemmen, D.S., Rea, B.R. 1999. Glacial landsystems of the southwest Laurentide Ice Sheet: modern Icelandic analogues. *Journal of Quaternary Science* 14: 673–691.

Evans, D.J.A., Rea, B.R., Hiemstra, J.F., Ó Cofaigh, C. 2006a. A critical assessment of subglacial mega-floods: a case study of glacial sediments and landforms in south-central Alberta, Canada. *Quaternary Science Reviews* 25: 1638-1667.

Evans, J., Dowdeswell, J., Ó Cofaigh, C. 2004. Late Quaternary submarine bedforms and ice-sheet flow in Gerlache Strait and on the adjacent continental shelf, Antarctic Peninsula. *Journal of Quaternary Science* 19(4): 397-407.

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

Evans, D.J.A., Twigg, D.R., Rea, B.R., Shand, M. 2007. *Surficial geology and geomorphology of the Brúarjökull surging glacier landsystem*. 1:25000 Map/Poster. University of Durham.

Evenson, E.B., Lawson, D.E., Strasser, J.C., Larson, G.J., Alley, R.B., Ensminger, S.L., Stevenson, W.E. 1999. Field evidence for the recognition of glaciohydrologic supercooling. In, Mickelson, D.M., Attig, J.W. (eds.) *Glacial Processes Past and Present*. Geological Society of America Special Paper, 337: 23–35.

Eyles, N. 1979. Facies of supraglacial sedimentation on Icelandic and alpine temperate glaciers. *Canadian Journal of Earth Scientists* 16: 1341-1361.

Eyles, N. 1983. Modern Icelandic glaciers as depositional models for 'hummocky moraine' in the Scottish Highlands. In, Evenson, E.B. (ed.) *Tills and related deposits*. Balkema, Rotterdam, p. 47-59.

Eyles N., Boyce, J.I., Barendregt, R.W. 1999. Hummocky moraine: sedimentary record of stagnant Laurentide Ice Sheet lobes resting on soft beds. *Sedimentary Geology* 123:163–174.

Fahnestock, M., Abdalati, W., Joughin, I., Brozena, J., Gogineni, P. 2001. High geothermal heat flow, basal melt, and the origin of rapid ice flow in central Greenland. *Science* 294(5550): 2338-2342.

Fairchild, H.L. 1907. Drumlin structure and origin. *Geological Society of America* 17: 702-706

Falorni, G., Teles, V., Vivoni, E.R., Bras, R.L., Amaratunga, K.S. 2005. Analysis and characterization of the vertical accuracy of digital elevation models from the Shuttle Radar Topography Mission. *Journal of Geophysical Research* 110, F02005, DOI: 10.1029/2003JF000113, 2005

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

Fisher, D.A., Reeh, N., Langley, K. 1985. Objective reconstructions of the Late Wisconsinan Ice Sheet. *Géographie Physique et Quaternaire* 39: 229-238.

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

Fisher, T.G., Spooner, I. 1994. Subglacial meltwater origin and subaerial meltwater modification of drumlins near Morley, Alberta, Canada. *Sedimentary Geology* 91: 285–298.

Fitzsimons, S.J. 1992. Sedimentology and depositional model for glaciolacustrine deposits in an ice dammed tributary valley, western Tasmania, Australia. *Sedimentology* 39: 393-410.

Fitzsimons, S.J. 1996. Formation of thrust block moraines at the margins of dry-based glaciers, south Victoria Land, Antarctica. *Annals of Glaciology* 22: 68-74.

Fowler, A.C. 2002. Correspondence. Rheology of subglacial till. *Journal of Glaciology* 48: 631-632.

Fullerton, D.S., Colton, R.B. 1986. *Stratigraphy and correlation of the glacial deposits of the Montana plains*. In, Sibrava, V., Bowen, D.Q., G.M. Richmond, G.M. (eds.) Quaternary Glaciations in the Northern Hemisphere, Pergamon Press, Oxford, Quaternary Science Reviews, Special Volume 5, p. 69-82.

Fullerton, D.S., Bush, C.A., Colton, R.B., Straub, A.W. 2004a. Map showing spatial and temporal relations of mountain and continental glaciations on the northern plains, primarily in northern Montana and northwestern North Dakota. *U.S. Geological Survey Geologic Investigations Series Map I-2843*, scale 1: 1,000,000, 1 sheet with accompanying text.

Fullerton, D.S., Colton, R.B., Bush. C.A. 2004b. Limits of mountain and continental glaciations east of the Continental Divide in northern Montana and north-western North Dakota, USA. In: Ehlers, J. and Gibbard, P.L. (eds.), *Quaternary Glaciations – Extent and Chronology Part II: North America*, Elsevier, London.

Fulton, R.J. 1995. Surficial Materials of Canada. Map 1880A, *Geological Survey of Canada*: Ottawa.

Geiger, R.W. 1967. Bedrock topography of the Gleichen map area, Alberta. Research Council of Alberta, Report 67-2, 14pp.

Glasser, N.F., Hambrey, M.J. 2003. Ice-marginal terrestrial landsystems: Svalbard polythermal glaciers. In, Evans, D.J.A. (ed.) *Glacial Landsystems*. Arnold, London, pp. 65–88.

Glasser, N.F., Jansson, K.N. 2005. Fast-flowing outlet glaciers of the Last Glacial Maximum Patagonian Icefield. *Quaternary Research* 63: 206-211.

Gravenor, C.P. 1955. The origin and significance of prairie mounds. *American Journal of Science* 253: 475-483.

Gravenor, C.P., Bayrock, L.A. 1955. Glacial geology of the Coronation district, Alberta. *Research Council of Alberta, Preliminary Report* 55-1.

Gravenor, C.P., Kupsch, W.O. 1959. Ice disintegration features in western Canada. *Journal of Geology* 67: 48–64.

Ham, N.R., Attig, J.W. 1996. Ice wastage and landscape evolution along the southern margin of the Laurentide Ice Sheet, north-central Wisconsin. *Boreas* 25: 171-186.

Hambrey, M.J., Huddart, D., Bennett, M.R., Glasser, N.F. 1997. Genesis of 'hummocky moraine' by thrusting in glacier ice: evidence from Svalbard and Britain. *Journal of the Geological Society of* London 154: 623–632.

Hambrey, M.J., Bennett, M.R., Dowdeswell, J.A., Glasser, N.F., Huddart, D. 1999. Debris entrainment and transport in polythermal valley glaciers, Svalbard. *Journal of Glaciology* 45: 69–86.

Hart, J.K. 1999. Identifying fast ice flow from landform assemblages in the geological record: a discussion. *Annals of Glaciology* 28: 59-67.

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

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

Heyman, J., Hätterstrand, C.H., Stroeven, A.P. 2008. Glacial geomorphology of the Bayan Har sector of the NE Tibetan Plateau. J*ournal of Maps*, 2008: 42-62

Hiemstra, J.F., Rijsdijk, K.F., Evans, D.J.A., van der Meer, J.J.M. 2005. Integrated Micro- and Macro-scale Sediment Analyses of Irish Sea Diamicts from Abermawr and Traeth y Mwnt, Wales, U.K. *Boreas* 34(1): 61-74.

Hindmarsh, R.C.A. 1997. Deforming beds: viscous and plastic scales of deformation. *Quaternary Science Reviews* 16: 1039-1056.

Hindmarsh, R.C.A. 1998a. The stability of a viscous till sheet coupled with ice flow, considered at wavelengths less than the ice thickness. *Journal of Glaciology* 44(147): 285-292.

Hindmarsh, R.C.A. 1998b. Drumlinization and drumlin-forming instabilities: viscous till mechanisms. *Journal of Glaciology* 44(147): 293-314.

Hodge, S.M., Doppelhammer, S.K. 1996. Satellite imagery of the onset of streaming flow of ice streams C and D, West Antarctica. *Journal of Geophysical Research* 101(3): 6669–6677.

Hooke, R.LeB., Jennings, C.E. 2006. On the formation of the tunnel valleys of the southern Laurentide ice sheet. *Quaternary Science Reviews* 25: 1364-1372.

Hooke, R.L., Hanson, B., Iverson, N.R., Jansson, P., Fischer, U.H. 1997. Rheology of till beneath Storglaciaren, Sweden. *Journal of Glaciology* 38: 348-358.

Hubbard, B., Glasser, N. 2005. *Field Techniques in glaciology and glacial geomorphology*. Wiley, Chichester, p217-262.

Hulbe, C.L., Fahnestock, M.A. 2004. West Antarctic ice-stream discharges variability: mechanism, controls and pattern of grounding-line retreat. *Journal of Glaciology* 50 (171): 471–484.

Iverson, N.R., Hanson, B., Hooke, R., Le, B., Jansson, P. 1995. Flow mechanism of glaciers on soft beds. *Science* 267: 80-81.

Iverson, N.R., Hooyer, T.S., Baker, R.W. 1998. Ring-shear studies of till deformation: coulomb-plastic behaviour and distributed strain in glacier beds. *Journal of Glaciology* 44: 634-642.

Jackson, L.E., Phillips, F.M., Shimamura, K, Little, E.C. 1997. Cosmogenic ^{36}Cl dating of the Foothills erratics train, Alberta, Canada. *Geology* 25: 195–198.

Jennings, C.E. 2006. Terrestrial ice streams: a view from the lobe. *Geomorphology* 75: 100–124.

Johannesson, T., Raymond, C.F., Waddington, E. 1989. Time-scale for adjustment of glaciers to changes in mass balance. *Journal of Glaciology* 35: 355–369.

Johnson, M.D. 1999. Spooner Hills, northwest Wisconsin: high-relief hills carved by subglacial meltwater of the Superior Lobe. In, Michelson, D.M., Attig, J.W. (eds.) *Glaciers Past and Present*. Geological Society of America Special Paper 337, pp. 83–92.

Johnson, M.D., Clayton, L. 2003. Supraglacial landsystems in lowland terrain. In, Evans, D.J.A. (ed.) *Glacial Landsystems.* Arnold, London, p228-251.

Johnson, M.D., Mickelson, D.M., Clayton, L., Attig, J.W. 1995. Composition and genesis of glacial hummocks, western Wisconsin. *Boreas* 24: 97-116.

Johnson, P.G. 1972. The morphological effects of surges of the Donjek Glacier, St Elias Mountains, Yukon Territory, Canada. *Journal of Glaciology* 11: 227-234.

Johnson, W. H. 1990: Ice-wedge casts and relict patterned ground in central Illinois and their environmental significance. *Quaternary Research 33*: 51–72.

Joughin, I., Tulaczyk, S. 2002. Positive mass balance of the Ross Ice Streams, Antarctica. *Science* 295 (5554): 476–480.

Joughin, I., Fahnestock, M., MacAyeal, D., Bamber, J.L., Gogineni, P. 2001. Observation and analysis of ice flow in the largest Greenland ice stream. *Journal of Geophysical Research* 106 (D24) 34: 21–34.

Joughin, I., Gray, J.T., Bindschadler, R.A., Price, S.F., Morse, D.L., Hulbe, C.L., Mattar, K., Werner, C. 1999. Tributaries of West Antarctic ice streams revealed by RADARSAT interferometry. *Science* 286 (5438): 283–286.

Joughin, I., Tulaczyk, S., Bindschadler, R., Price, S.F. 2002. Changes in West Antarctic ice stream velocities: observation and analysis. *Journal of Geophysical Research* 107(11): 10.1029/2001JB001029, 2002.

Lindén, M., Möller, P., Adrielsson, L. 2008. Ribbed moraine formed by subglacial folding, thrust stacking and lee-side cavity fill. *Boreas* 37: 102–131.

Liverman. D.G.E., Catto, N.R. and Rutter, N.W. 1989. Laurentide glaciations in west-central Alberta – a single (late Wisconsinan) event. *Canadian Journal of Earth Sciences*, 26: 266-274.

Lukas, S. 2005. A test of the englacial thrusting hypothesis of 'hummocky' moraine formation: case studies from the northwest Highlands, Scotland. *Boreas* 34: 287-307.

Lukas, S. 2007. ‘A test of the englacial thrusting hypothesis of “hummocky” moraine formation: case studies from the northwest Highlands, Scotland’: Reply to comments. *Boreas* 36: 108-113.

Lundqvist, J. 1969. Problems of the so-called Rogen moraine. *Sverige Geologiske Undersökning*, Series C, 648.

Kamb, B. 1991. Rheological nonlinearity and flow instability in the deforming-bed mechanisms of ice stream motion. *Journal of Geophysical Research* 96B: 16585-16595.

Kamb, B. 2001. Basal zone of the West Antarctic ice streams and its role in lubrication of their rapid motion. In *The West Antarctic Ice Sheet: Behaviour and Environment*, Alley, R.B. and Bindschadler, R. (eds). Antarctic Research Series 77, American Geophysical Union: Washington, DC; 157–199.

Kaplan, M.R., Miller, G.H., Steig, E.J. 2001. Low-gradient outlet glaciers (ice streams?) drained the Laurentide ice sheet. *Geology* 29: 343-346.

Kehew, A.E., Clayton, L. 1983. Late Wisconsinan floods and development of the Souris-Pembina spillway system in Saskatchewan, North Dakota and Manitoba. In, Teller, J.T., Clayton, L. (eds.) *Glacial Lake Agassiz*. Geological Association of Canada Special Paper 26: 187-209.

Kehew, A.E., Lord, M.L. 1986. Origin and large scale erosional features of glacial lake spillways in the northern Great Plains. *Geological Society of America Bulletin* 97: 162-177.

Kehew, A.E., Lord, M.L. 1987. Glacial lake outbursts along the mid-continent margins of the Laurentide ice-sheet. In, Mayer, L., Nash, D. (eds.), Catastrophic Flooding. Allen and Unwin, Boston, 95-120.

Kjaersgaard, A.A. 1976. Reconnaissance soil survey of the Oyen map sheet – 72M (Preliminary Report). *Alberta Institute of Pedology*, ReportS-76-36, 85pp.

Klassen, R.W., 1989. Quaternary geology of the southern Canadian Interior Plains. In, R.J. Fulton, R.J. (ed.), *Quaternary Geology of Canada and Greenland*, Chapter 2. Geological Survey of Canada, Geology of Canada, 1 (also Geological Society of America, the Geology of North America, Vol. K-I).

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

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

Knudsen, C.G., Larsen, E., Sejrup, H.P., Stalsberg, K. 2006. Hummocky moraine landscape on Jæren, SW Norway—implications for glacier dynamics during the last deglaciation. *Geomorphology* 77: 153 -168.

Krüger, J. 1983. Glacial geomorphology and deposits in Denmark, In. Ehlers, J. (ed.) *Glacial deposits in north-west Europe*. Balkema, Rotterdam, p.181-191.

Krüger, J. 1996. Moraine ridges formed from subglacial frozen-on sediment slabs and their differentiation from push moraines. *Boreas* 25: 57–63.

Kulig, J.J. 1996. The glaciations of the Cypress Hills of Alberta and Saskatchewanand its regional implications. *Quaternary International* 32: 53-77.

Lagerbäck, R. 1988. The Veiki moraines in northern Sweden widespread evidence of an Early Weichselian deglaciation. *Boreas* 17: 469–486.

MacAyeal, D.R. 1993a. A low-order model of growth/purge oscillations of the Laurentide Ice Sheet. *Paleooceanography* 8: 767-773.

MacAyeal, D.R. 1993b. Binge/purge oscillations of the Laurentide Ice Sheet as a cause of the North Atlantic’s Heinrich events. *Palaeoceanography* 8: 775–784.

Marshall, S.J., Clarke, G.K.C., Dyke, A.S., Fisher, D.A. 1996. Geologic and topographic controls on fast flow in the Laurentide and Cordilleran Ice Sheets. *Journal of Geophysical Research* 101(B8): 17827-17839.

Mathews, W.H. 1974. Surface profiles of the Laurentide Ice Sheet in its marginal areas. *Journal of*
Glaciology 13: 37-43.

Matthews, J.A., McCarroll, D., Shakesby, R.A. 1995. Contemporary terminal-moraine ridge formation at a temperate glacier: Styggedalsbreen, Jotunheimen, southern Norway. *Boreas* 24: 129–139.

McIntyre, N.F. 1985. The dynamics of outlet glaciers. *Journal of Glaciology* 31(108): 99 –107.

Menzies, J. 1984. *Drumlins: A Bibliography*. Geo Books, Norwich.

Menzies, J., Rose, J. 1987. *Drumlins: trends and perspectives*. *Episodes* 10: 29-31.

Mickelson, D.M., Clayton, L., Fullerton, D.S., Borns, H.W. 1983. The late Wisconsin glacial record of the Laurentide ice sheet in the United States. In, Wright, H.E., Jr. (ed.), *Late Quaternary environments of the United States*. Vol.1, University of Minnesota Press, Minneapolis, p.3-37.

Mickelson, D.M., Colgan, P.M. 2003. The southern Laurentide ice sheet in the United States. *In,* Gillespie, A., Porter, S.B., Atwater, B.F. (*eds). Advances in research in Quaternary geology of the United States*. New York, Elsevier, 1–16.

Mollard, J.D. 2000. Ice-shaped ring-forms in western Canada: their airphoto expressions and manifold polygenetic origins. *Quaternary International* 68: 187–198.

Möller, P. 2006. Rogen moraine: an example of glacial re-shaping of pre-existing landforms. *Quaternary Science Reviews* 25: 362-389.

Mooers, H.D. 1990. Ice-marginal thrusting of drift and bedrock:thermal regime, subglacial aquifers, and glacial surges. *Canadian Journal of Earth Sciences* 27: 849–862.

Mooers, H.D. 1989. On the formation of tunnel valleys of the Superior Lobe, central Minnesota. *Quaternary Research* 32: 24–35.

Moran, S.R., Clayton, L., Hooke, R.LeB., Fenton, M.M., Andrashak, L.D. 1980. Glacier-bed landforms of the Prairie region of North America. *Journal of Glaciology* 25: 457-476.

Morgan, V.I., Jacka, T.H., Akerman, G.J., Clarke, A.L. 1982. Outlet glacier and mass-balance studies in Enderby, Kemp and Mac. Robertson Lands, Antarctica. *Annals of Glaciology* 3, 204-210.

Munro-stasiuk, M.J. 1999. Evidence for water storage and drainage at the base of the Laurentide ice-sheet, south-central Alberta, Canada. *Annals of Glaciology* 28, 175-180.

Munro-Stasiuk, M.J., Shaw, J. 1997. Erosional origin of hummocky terrain in south-central Alberta, Canada. *Geology* 25: 1027-1030.

Munro-Stasiuk, M.J., Shaw, J. 2002. The Blackspring Ridge flute field, south-central Alberta, Canada: evidence for subglacial sheetflow erosion. *Quaternary International* 90: 75-86.

Munro-Stasiuk, M.J., Sjogren, D. 2006. The erosional origin of hummocky terrain, Alberta, Canada. In, Knight PG (ed.), *Glacier Science and Environmental Change*. Blackwell, Oxford, 33-36.

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

Napieralski, J., Harbor, J., Li, Y. 2007. Glacial geomorphology and geographic information systems. *Earth Science Reviews* 85: 1-22.

NASA. 2001. SRTM Topography. *ftp://e0srp01u.ecs.nasa.gov/srtm/version2/Documentation/SRTM_Topo.pdf* -

Ó Cofaigh, C. 1996. Tunnel valley genesis. *Progress in Physical Geography* 20: 1-19.

Ó Cofaigh, C., Stokes, C.R., 2008. Reconstructing ice-sheet dynamics from subglacial sediments and landforms: introduction and overview. *Earth Surface Processes and Landforms* 22: 495-502.

Ó Cofaigh, C., Dowdeswell, J.A., Evans, J., Larter, R.D. 2008. Geological constraints on Antarctic palaeo-ice-stream retreat. *Earth Surface Processes and Landforms* 33: 513-525.

Ó Cofaigh, C., Evans, D.J.A., Smith, R. In Press. Large-scale reorganistaion and sedimentation of terrestrial ice-streams during a single glacial cycle. *Geological Society of America Bulletin*.

Ó Cofaigh, C., Pudsey, C.J., Dowdeswell, J.A., Morris, P. 2002. Evolution of subglacial bedforms along a palaeo-ice stream, Antarctic Peninsula continental shelf. *Geophysical Research Letters* 29(8): 10.1029/2001GL014488.

Onorati, G., Poscolieri, M., Ventura, R., Chiarini, V., Crucilla, U. 1992. The Digital Elevation Model of Italy for Geomorphology and Structural Geology. *CATENA* 19: 147-178.

Ottesen, D., Dowdeswell, J.A., Rise, L. 2005. Submarine landforms and the reconstruction of fast-flowing ice streams within a large Quaternary ice sheet: the 2500-km-long Norwegian-Svalbard margin (57°– 80°N). *Geological Society of America Bulletin* 117(7/8): 1033-1050.

Ottesen, D., Stokes, C.R., Rise, L., Olsen, L. 2008. Ice-sheet dynamics and ice streaming along coastal parts of northern Norway. *Quaternary Science Reviews* 27: 922-940.

Parizek, R.R. 1969. Glacial ice-contact ridges and rings. *Geological Society of America Special Paper* 123: 49–102.

Parizek, B.R., Alley, R.B., Anandakrishnan, S., Conway, H. 2002. Sub-catchment melt and long-term stability of ice stream D, West Antarctica. *Geophysical Research Letters* 29 (8): 55.1–55.4.

Paterson, W.S.B. 1994. *The Physics of Glaciers*. Pergamon–Elsevier, Oxford.

Patterson, C.J. 1997. Southern Laurentide ice lobes were created by ice streams: Des Moines Lobe in Minnesota, USA. *Sedimentary Geology* 111: 249–261.

Patterson, C.J. 1998. Laurentide glacial landscapes: the role of ice streams. *Geology* 26: 643–646.

Paul, M.A. 1983. The supraglacial landsystem. In, Eyles, N. (ed.) *Glacial geology*. Pergamon, Oxford, p.71-90.

Pawlowicz, J.G., Fenton, M.M. 1995. Bedrock Topography of Alberta. *Alberta Geological Survey*, Map 1:2,000,000.

Payne, A.J. 1995. Limit cycles in the basal thermal regime of ice sheets. *Journal of Geophysical* Research B100: 4249-4263.

Payne, A.J., Dongelmans, P.W. 1997. Self-organisation in the thermomechanical flow of ice sheets. *Journal of Geophysical Research* 102: 12219–12234.

Peters, L.E., Anandakrishnan, S., Alley, R.B., Winberry, J.P., Voigt, D.E. 2006. Subglacial sediments as a control on the onset and location of two Siple Coast ice streams, West Antarctica. *Journal of Geophysical Research* 111: 10.1029/2005JB003766.

Piotrowski, J.A., Kraus, A. 1997. Response of sediment to ice sheet loading in northwestern Germany: effective stresses and glacier bed stability. *Journal of Glaciology* 43: 495–502.

Piotrowski, J.A., Jørgensen, F., Kristiansen, S., Larsen, N.K., Rattas, M., Windelberg, S., Wysota, W. 2003. *Land-based ice streams of the southern Scandinavian Ice Sheet: movement mechanisms and ice/bed interactions*. XVI INQUA Congress, Reno, USA; Abstracts, p. 124.

Piotrowski, J.A., Larsen, N.K., 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., Mickelson, D.M., Tulaczyk, S., Krzyszkowski, D., Junge, F.W. 2001. Were deforming beds beneath past ice sheets really widespread? *Quaternary International* 86: 139–150.

Prest, V.K., Grant, D.R., Rampton, V.N. 1968. Glacial Map of Canada. Map 1253A (1:5000000), *Geological Survey of Canada*: Ottawa.

Price, R.J. 1970. Moraines at Fjallsjökull, Iceland. *Arctic and Alpine Research* 2: 27–42.

Price, S.F., I.M. Whillans. 2001. Crevasse patterns at the onset to Ice Stream B, West Antarctica. *Journal of Glaciology* 47(156): 29-36

Price, S.F., Bindschadler, R.A., Hulbe, C.L., Blankenship, D.D. 2002. Force balance along an inland tributary and onset to Ice Stream D, West Antarctica. *Journal of Glaciology* 48(160): 20-30.

Punkari, M. 1995. Function of the ice streams in the Scandinavian ice sheet: Analyses of glacialgeological data from southwestern Finland: *Royal Society of Edinburgh Transactions, Earth Sciences* 85: 283-302.

Rabus, B., Eineder, M., Roth, A., Bamler, R. 2003. The shuttle radar topography mission – a new class of digital elevation models acquired by spaceborne radar. *Journal of Photogrammetry and Remote Sensing* 57: 241-262.

Radok, U., Jenssen, D., McInnes, B. 1987. On the surging potential of polar ice streams. Antartic surges- aclear and present danger? *U.S. Department of Energy Technical Report*, Washington DC: DOE/ER/60197-H1

Rains, R.B., Shaw, J., Skoye, K.R., Sjogren, D.B., Kvill, D.R. 1993. Late Wisconsin subglacial megaflood paths in Alberta. *Geology* 21: 323–326.

Rains, R.B., Kvill, D., Shaw, J. 1999. Evidence and some implications of coalescent Cordilleran and Laurentide glacier systems in western Alberta. In A World of Real Places: Essays in Honour of William C. Wonders, Smith PJ (ed.). University of Alberta: Edmonton, p. 147–161.

Rains, R.B., Shaw, J., Sjogren, D.B., Munro-Stasiuk, M.J., Robert, R., Young, R.R., Thompson, R.T. 2002. Subglacial tunnel channels, Porcupine Hills, southwest Alberta, Canada. *Quaternary International 90*: 57-65.

Ramirez, E. 2006. *Shuttle Radar Topography Mission: The mission to map the world.* Project Formulation & Management Applications (PFMA) Group. Jet propulsion laboratory. http://www2.jpl.nasa.gov/srtm/

Raymond, C.F. 1987. How do glaciers surge? A review. *Journal of Geophysical Research* 92: 9121-9134.

Raymond, C.F., Echelmeyer, K.A., Whillans, I.M., Doake, C.S.M. 2001. Ice stream shear margins. In: Alley, R.B., Bindschadler, R.A. (Eds.) *The West Antarctic Ice Sheet: Behaviour and Environment*. Antarctic Research Series 77: 137–155. American Geophysical Union, Washington, D.C.

Rijsdijk, K.F. 2001. Density-driven deformation structures in glacigenic consolidated diamicts: Examples from Traeth Y Mwnt, Cardiganshire, Wales, UK. *Journal of Sedimentary Research* 71: 122–135.

Roberts, D.H., Long, A.J. 2005. Streamlined bedrock terrain and fast ice flow, Jakobshavns Isbrae, West Greenland: implications for ice stream and ice sheet dynamics. *Boreas* 34: 25-42.

Roberts, D.H., Yde, J.C., Knudsen, N.T., Long, A.J., Lloyd, J.M. 2009. Ice marginal dynamics during surge activity, Kuannersuit Glacier, Disko Island, West Greenland. *Quaternary Science Reviews* 28: 209-222.

Rose, K.E. 1979. Characteristics of ice flow in Marie Byrd Land, Antarctica. *Journal of Glaciology* 24: 63-74.

Rose, J. 1987. Drumlins as part of a glacier bedform continuum. In, Menzies, J., Rose, J. (eds), Drumlin symposium, p. 103-116. Balkema, Rotterdam.

Rust, B.R., Romanelli, R. 1975. Late Quaternary Subaqueous Outwash Deposits near Ottawa, Canada. In, Jopling, A.V., McDonald, B.C., (eds) *Glaciofluvial and Glaciolacustrine sedimentation*. Special Publication Society of Economic Palaontologists and Minerologists, Tulsa, 23: 177-193.

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

Sharp, M.J. 1984. Annual moraine ridges at Skalafellsjökull, southeast Iceland. *Journal of Glaciology* 30: 82–93

Sharp, M.J. 1985a. 'Crevasse-fill' ridges – a landform type characteristic of surging glaciers? *Geografiska Annaler* 67A: 213-220.

Sharp, M.J. 1985b. Sedimentation and stratigraphy at Eyjabakkajökull – an Iceland surging glacier. *Quaternary Research* 24: 268-284.

Sharpe, D.R. 1987. The stratified nature of drumlins from Victoria Island and Southern Ontario, Canada. In, Menzies, J., Rose, J. (Eds.) *Drumlin Symposium*. Balkema, Rotterdam, 185–214.

Shaw, J. 1975. Sedimentary successions in ice marginal lakes. In, Jopling, A.V., McDonald, B.C., (eds) *Glaciofluvial and Glaciolacustrine sedimentation*. Special Publication Society of Economic Palaontologists and Minerologists, Tulsa, 23: 281-363.

Shaw, J. 1977. Sedimentation in an alpine lake during deglaciation: Okanagan Valley, British Columbia, Canada. Geografiska Annaler 59: 221-240.

Shaw, J. 1979. Genesis of the Sveg till and Rogen moraines of central Sweden: a model of basal Melt out. *Boreas* 8, 409–426.

Shaw, J. 1980. Drumlins and large scale flutings related to glacier folds. *Arctic and Alpine Research* 12: 287-298.

Shaw, J. 1983. Drumlin formation related to inverted meltwater erosional marks. *Journal of Glaciology* 29: 461–479.
Shaw, J., 1994. Hairpin erosional marks, horseshoe vortices and subglacial erosion. *Sedimentary Geology* 91: 269–283.
Shaw, J. 1996. A meltwater model for Laurentide subglacial landscapes. In: McCann, S.B., Ford, D.C. (eds.), *Geomorphology sans frontiere*. Wiley, Chichester, pp. 182–226.
Shaw, J. 2002. The meltwater hypothesis for subglacial bedforms. *Quaternary International* 90: 5-22.
Shaw, J., Kvill, D. 1984. A glaciofluvial origin for drumlins of the Livingstone Lake area, Saskatchewan. *Canadian Journal of Earth Sciences* 12: 1426-1440.
Shaw, J., Kvill, D., Rains, R.B. 1989. Drumlins and catastrophic subglacial floods. *Sedimentary Geology* 62: 177–202.
Shaw, J., Rains, R.B., Eyton, J.R., Weissling L. 1996. Laurentide subglacial outburst floods: landform evidence from digital elevation models. Canadian *Journal of Earth Sciences* 33: 1154–1168.
Shaw, J., Faragini, D., Kvill, D.R., Rains, R.B. 2000. The Athabasca fluting field, Alberta, Canada: implications for the formation of large scale fluting (erosional lineations). *Quaternary Science Reviews* 19: 959–980.
Shetsen I. 1984. Application of till pebble lithology to the differentiation of glacial lobes in southern Alberta. *Canadian Journal of Earth Sciences* 21: 920–933.
Shetsen, I. 1987. *Quaternary Geology, Southern Alberta*. ARC map (1:500000), Alberta Research Council: Edmonton, Canada.
Shetsen, I. 1990. *Quaternary Geology, Central Alberta*. ARC map (1:500000), Alberta Research Council: Edmonton, Canada.
Shipp, S.S., Anderson, J.B., Domack, E.W. 1999. Late Pleistocene–Holocene retreat of the West Antarctic Ice-Sheet system in the Ross Sea: Part 1 – Geophysical results. *Geological Society of America Bulletin* 111: 1486-1516.
Sjogren, D.B. 1994. *Geomorphology of the Coronation-Spondin scabland*. M.Sc. Thesis, University of Alberta, Canada.
Sjogren, D.B., Rains, R.B. 1995. Glaciofluvial erosional morphology and sediments of the Coronation-Spondin scabland, east-central Alberta. *Canadian Journal of Earth Sciences* 32: 565–578.
Sjogren, D.B. 1999. *Formation of the Viking Moraine, east-central Alberta: geomorphic and sedimentary evidence*. Ph.D. Thesis, University of Alberta, Canada, 219pp.
Smalley, I.J., Unwin, D.J. 1968. The formation and shape of drumlins and their distribution and orientation in drumlin fields. *Journal of Glaciology* 7: 377-390.
Smalley, I.J., Piotrowski, J.A. 1987. Critical Strength/stress ratios at the ice-bed interface in the drumlin forming process: from 'dilitancy' to 'cross-over'. In, Menzies, J., Rose, J. (Eds) *Drumlin symposium*. Balkema, Rotterdam, 81-86.
Smith, A.M. 1997. Basal conditions on Rutford Ice Stream, West Antarctica, from seismic observations. *Journal of Geophysical Research* 102 (B1): 543-552.
Smith L.C., Alsdorf D.E., Magilligan F.J., Gomez B., Mertes L.A.K., Smith N.D., Garvin J.B. 2000. Estimation of erosion, deposition, and net volume tric change caused by the 1996 Skeidararsandur jökulhlaup, Iceland, from synthetic aperture radar interferometry. *Water Resources Research* 36: 1583–1594.
Smith, M. J., Clark, C.D. 2005. Methods for the visualization of digital elevation methods for landform mapping. *Earth Surface Processes and Landforms* 30: 885-900.
Smith, M.J., Rose, J., Booth, S. 2006. Geomorphological mapping of glacial landforms from remotely sensed data: An evaluation of the principal data sources and an assessment of the quality. *Geomorphology* 76: 148-165.
Sollid, J.L., Sørbel, L. 1988. Influence of temperature conditions in formation of end moraines in Fennoscandia and Svalbard. *Boreas* 17: 553–558.
Stalker, A. M. 1956. The Erratics Train; Foothills of Alberta. *Geological Survey of Canada, Bulletin* 37: 28pp.

Stalker, A. M. 1960. Ice-pressed drift forms and associated deposits in Alberta. *Geological Survey of Canada Bulletin 57*: 38pp.
Stalker, A. M. 1962. *Surficial Geology, Lethbridge, Alberta*. Geological Survey of Canada, Map 41, 1:253,440.
Stalker, A. M. 1969. Quaternary stratigraphy in southern Alberta. Report II – sections near Medicine Hat. *Geological Survey of Canada,* Paper 69-26, 28pp.
Stalker, A. M. 1973. The large interdrift bedrock blocks of the Canadian Prairies. *Geological Survey Canada*, Paper 75-1A: 421-422
Stalker, A. M. 1976. Megablocks, or the enormous erratic of the Albertan Prairies. *Geological Survey Canada,* Paper 76-1C, p. 185-188.
Stalker, A. M. 1977. The probable extent of the Classical Wisconsin ice in southern and central Alberta. *Canadian Journal of Earth Sciences*, v. 14, p.2614-2619.
Stalker A. M., Barendregt, R. 1988. *Megablocks east of Lethbridge*. Field excursion Guide. University of Lethbridge.
Stokes, C.R., Clark, C.D. 1999. Geomorphological criteria for identifying Pleistocene ice streams. *Annals of Glaciology* 28: 67–74.
Stokes, C.R., Clark, C.D. 2001. Palaeo-ice streams. *Quaternary Science Reviews* 20: 1437–1457.
Stokes, C.R., Clark, C.D. 2002. Are long bedforms indicative of fast ice flow? *Boreas* 31: 239–249.
Stokes, C.R., 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., 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., Clark, C.D. 2004. Evolution of late glacial ice-marginal lakes on the northwestern Canadian Shield and their influence on the location of the Dubawnt Lake palaeo-ice stream. *Palaeogeography, Palaeoclimatology, Palaeoecology* 215: 155-171.
Stokes, C.R., Clark, C.D. 2006. What can the 'footprint' of a palaeo-ice stream tell us? Interpreting the bed of the Dubawnt Lake Ice Stream, northern Keewatin, Canada. In *Glacier Science and Environmental Change*, Knight, P.G. (ed.). Blackwell: Oxford; 208–209.
Stokes, C.R., Clark, C.D., Winsborrow, M.C.M. 2006. Subglacial bedform evidence for a major palaeo-ice stream and its retreat phases in the Amundsen Gulf, Canadian Arctic Archipelago. *Journal of Quaternary Science* 21(4): 399-412.
Stokes, C.R., Clark, C.D., Olav, B., Lian, O.B., Tulaczyk, S. 2007. Ice stream sticky spots: a review of their identification and influence beneath contemporary and palaeo-ice streams. *Earth Science Reviews* 81: 217–249.
Stokes, C.R., Lian, O.B., Tulaczyk, S., Clark, C.D. 2008. Superimposition of ribbed moraines on a palaeo-ice stream: implications for ice stream shutdown. *Earth Surface Processes and Landforms* 33: 593-609.
Stone, B.D., Peper, J.D. 1982. Topographic control of the deglaciation of eastern Massachusettes: Ice lobation and moraine incursion. In, Larsen, G.J., Stone, B.D. (eds.), *Late Wisconsinan glaciations of New England.* Kendell Hunt, Dubuque IA, p.145-166.
Storrar, R., Stokes, C.R. 2007. A Glacial Geomorphological Map of Victoria Island, Canadian Arctic. *Journal of Maps* 2007: 191-210.
Strehl, E. 1998. Glazilimnische Kames in Schleswig-Holstein. *Eiszeitalter u. Gegenwart* 48: 19-22.
Swift, D.A., Evans, D.J.A., Fallick, A.E. 2006. Transverse englacial debris-rich ice bands at Kviarjökull, southeast Iceland. *Quaternary Science Reviews* 25: 1708–1718.
Swithinbank, C.W.M. 1954. Ice Streams. *Polar Record* 7: 185-186.
Tsui, P.C., Cruden, D.M., Thomson, S. 1989. Ice thrust terrains and glaciotectonic settings in central Alberta. *Canadian Journal of Earth Sciences* 26: 1308–1318.
Tulaczyk, S.M. 2006. Fast glacier flow and ice streaming. In *Glacier Science and Environmental Change*, Knight, P.G. (ed.). Blackwell: Oxford; 353–359.

Tulaczyk, S.M., Kamb, B., Engelhardt, H. 2000a. Basal mechanics of Ice Stream B: I. Till mechanics. *Journal of Geophysical Research* 105: 463-481.
Tulaczyk, S.M., Kamb, B., Engelhardt, H. 2000b. Basal Mechanics of Ice Stream B: II. Plastic-undrained-bed model. *Journal of Geophysical Research* 105: 483-494.
Tulaczyk, S.M, Scherer, R.P., Clark, C.D. 2001. A ploughing model for the origin of weak tills beneath ice streams: a qualitative treatment. *Quaternary International* 86: 59–70.
United States Geological Survey. 2006. Landsat Thematic Mapper Plus (ETM+). http://edc.usgs.gov/products/satellite/landsat7.php
Van der Wateren, F.M. 2003. Ice-marginal terrestrial landsystems: southern Scandinavian ice sheet margin. In, Evans, D.J.A. (ed.) *Glacial Landsystems*. Arnold, London, p. 166-202.
Vernon, P.D. 1962. Tills of the Lethbridge area, Alberta; their stratigraphy, fabric, and composition. Unpublished M.S. thesis, University of Calgary, Calgary, Alberta, 181pp.
Vogel, S.W., Tulaczyk, S. 2006. Ice-dynamical constraints on the existence and impact of subglacial volcanism on West Antarctic ice sheet stability. *Geophysical Research Letters* 33: L23502. DOI: 10.1029/2006GL027345
Vogel, S.W., Tulaczyk, S., Joughin, I.R. 2003. Distribution of basal melting and freezing beneath tributaries of Ice Stream C: implication for the Holocene decay of the West Antarctic Ice Sheet. *Annals of Glaciology* 36: 273–282.
Wang, W., Zwally, H.J., Hulbe, C.L., Siegert, M.J., Joughin, I. 2003. Anisotropic ice flow leading to the onset of Ice Stream D, West Antarctica: numerical modelling based on the observations from Byrd Station borehole. *Annals of Glaciology* 37: 397– 403.
Wellner, J.S., Lowe, A.L., Shipp, S.S., Anderson, J.B. 2001. Distribution of glacial geomorphic features on the Antarctic continental shelf and correlation with substrate. *Journal of Glaciology* 47: 397-411.
Westgate, J.A. 1968. Surficial geology of the Foremost – Cypress Hills area, Alberta. *Research Council of Alberta*, Bulletin 22.
Whillans, I.M., Bentley, C.R., van der Veen, C.J. 2001. Ice streams B and C. In, Alley, R.B., Bindschadler, R.A. (eds) *The West Antarctic Ice Sheet: Behavior and Environment*. American Geophysical Union, Antarctic Research Series 77: 257-281.
Wickenden, R.T.D. 1937. Age relations of glacial deposits in southeastern Alberta and southwestern Saskatchewan. *Proc. Royal Society of Canada* 3 (31): 113.
Winsborrow, M.C.M., Clark, C.D., Stokes, C.R. 2004. Ice Streams of the Laurentide Ice Sheet. *Géographie physique et Quaternaire* 58(2-3): 269-280.
Wright, H.E., Jr. 1973. Tunnel valleys, glacier surges, and subglacial hydrology of the Superior lobe, Minnesota. In, Black, R.F., Goldthwait, R.P., Williams, H.B. (eds.) *The Wisconsinan Stage*. Geological Society of America Memoir 135: 251-276.
Wright, H.E., Jr. 1980. Surge moraines of the Klutlan Glacier, Yukon Territory, Canada and application to the Late Glacial in Minnesote. *Quaternary Research* 14: 2-17.
Wyatt, F.A., Newton, J.D., Bowser, W.E., Odynsky, W. 1941. Soil survey of Milk River sheet. University of Alberta, College of Agriculture 14: 76.
Young, R.R., Burns, J.A., Rains, R.B. 1989. A re-evaluation of the Empress Formation gravels and sands in the Beverly Valley, central Alberta. In, CANQUA, *Late Glacial and Postglacial Processes and Environments in Montane and Adjacent Areas*: Programs and Abstracts, pp. 50.
Young, R.R., Burns, J.A., Rains, R.B., Smith, D.G. 1993. *The Late Wisconsinan Laurentide Ice Sheet and its relationship to a theoretical "ice free" corridor*. Program with Abstracts, Geological Association of Canada, Joint Annual Meeting, Edmonton, Alberta, pp. A-113.
Young, R.R., Burns, J.A., Smith, D.G., Arnold, L.D. and Rains, R.B. 1994. A single, late Wisconsin, Laurentide glaciations, Edmonton area and southwestern Alberta. *Geology* 22: 683-686.