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Glacial geomorphology of terrestrial-terminating fast flow lobes/ice stream margins in the southwest Laurentide Ice Sheet

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A B S T R A C T

Glacial geomorphological mapping of southern Alberta, Canada, reveals landform assemblages that are diagnostic of terrestrial-terminating ice streams/fast flowing outlet glaciers with lobate snouts. Spatial variability in features that comprise the landform assemblages reflects changes in (a) palaeo-ice stream activity (switch on/off); and (b) snout basal thermal regimes associated with climate sensitive, steady state flow. Palaeo-ice stream tracks reveal distinct inset sequences of fan-shaped flowsets indicative of receding lobate ice stream margins. Former ice lobe margins are demarcated by (a) major, often glacially overridden transverse moraine ridges, commonly comprising glacitectonically thrust bedrock; and (b) minor, closely spaced recessional push moraines and hummocky moraine arcs. Details of these landform types are well exhibited around the former southern margins of the Central Alberta Ice Stream, where larger scale, more intensive mapping identifies a complex glacial geomorphology comprising minor transverse ridges (MTR types 1–3), hummocky terrain (HT types 1–3), flutings, and meltwater channels/spillways. The MTR type 1 constitute the summit corrugation patterns of glacitectonic thrust moraines or major transverse ridges and have been glacially overrun and moderately streamlined. The MTR type 2 sequences are recessional push moraines similar to those developing at modern active temperate glacier snouts. The MTR type 3 document moraine construction by incremental stagnation because they occur in association with hummocky terrain. The close association of hummocky terrain with push moraine assemblages indicates that they are the products of supraglacial controlled deposition on a polythermal ice sheet margin, where the HT type 3 hummocks represent former ice-walled lake plains. The ice sheet marginal thermal regime switches indicated by the spatially variable landform assemblages in southern Alberta are consistent with palaeoglaciological reconstructions proposed for other ice stream/fast flow lobes of the southern Laurentide Ice Sheet, where alternate cold, polythermal, and temperate marginal conditions associated with climate sensitive, steady state flow sequentially gave way to more dynamic streaming and surging activity.

1. Introduction and rationale

The important role of ice streams in ice sheet dynamics has resulted in them becoming increasingly more prominent as a focus of multidisciplinary research in process glaciology and palaeoglaciology. Ongoing research questions surround the issues of maintenance and regulation of ice flow, temporal and spatial patterns of activation/deactivation, large scale changes in flow regime, and potential linkages/responses to climate. Some insights into these questions are emerging from the studies of former ice sheet beds, but the focus of such research has been largely targeted at marine-terminating ice streams. Details on the marginal activity of terrestrially-terminating ice streams have only recently emerged from the study of the former ice streams of the southern Laurentide Ice Sheet, where retreating ice stream margins clearly constructed lobate assemblages of closely spaced push moraines and associated landforms of a style compatible with seasonally driven ice flow dynamics (Patterson, 1997, 1998; Jennings, 2006; Evans et al., 2008; Ó Cofaigh et al., 2010; Evans et al., 2012). Whereas inset sequences of seasonally deposited push moraines and flutings are traditionally associated with active temperate outlet glaciers exhibiting steady state normal (<10^2 m y^{-1}) to fast (>10^2 m y^{-1}) flow (cf. Clarke, 1987), such as the southern Vatnajokull outlet glaciers in Iceland (e.g., Boulton, 1986; Hart, 1999; Evans and Twigg, 2002), the occurrence of such landform assemblages at the margins of the Laurentide Ice Sheet palaeo-ice streams is more difficult to reconcile with ice sheet scale fast flow dynamics of the magnitude observed in modern Antarctic ice streams. Hence mapping the distribution of palaeo-ice stream and active temperate (steady state normal-fast flow) landsystems has the potential to facilitate spatial and temporal reconstructions of Laurentide Ice Sheet ice streaming activity.

The western plains of southern Alberta, southwest Saskatchewan, and northern Montana contain a wealth of glacial landforms that previously have been widely employed in reconstructions of Laurentide Ice Sheet palaeoglaciology (Stalker, 1956, 1977; Christiansen, 1979; Clayton and Moran, 1982; Clayton et al., 1985; Evans and Campbell, 1992, 1995; Evans et al., 1999; Evans, 2000; Evans et al., 2006, 2008;

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Ó Cofaigh et al., 2010; Evans et al., 2012), while at the same time being central to conceptual developments in glacial geomorphology (e.g., Gravenor and Kupsch, 1959; Stalker, 1960; Clayton and Cherry, 1967; Bik, 1969; Stalker, 1973; Clayton and Moran, 1974; Stalker, 1976; Moran et al., 1980; Kehew and Lord, 1986; Tsui et al., 1989; Beaty, 1990; Alley, 1991; Boulton and Caban, 1995; Evans, 1996; Eyles et al., 1999; Mollard, 2000; Boone and Eyles, 2001; Evans, 2003; Clayton et al., 2008; Evans, 2009). Significant debate has also considered alternative, subglacial megaflood interpretations of the landforms of the region (cf. Rains et al., 1993; Sjogren and Rains, 1995; Shaw et al., 1996; Munro-Stasiuk and Shaw, 1997; Munro-Stasiuk, 1999; Beaney and Hicks, 2000; Beaney and Shaw, 2000; Beaney, 2002; Munro-Stasiuk and Shaw, 2002; Rains et al., 2002; Shaw, 2002; Clarke et al., 2005; Benn and Evans, 2006; Evans et al., 2006; Evans, 2010; Shaw, 2010). The research we present provides a landsystem approach to the interpretation of the glacial geomorphological legacy as it pertains to the Late Wisconsinan advance and retreat of the southwest Laurentide Ice Sheet in the context of the palaeo-ice stream activity demonstrated by Shetsen (1984), Evans et al. (1999, 2006, 2008, 2012), Evans (2000), and Ó Cofaigh et al. (2010). This approach makes the assumption at the outset that subglacially streamlined bedforms and ice-flow transverse landforms are not the product of megafloods, an assumption soundly based in the arguments presented in a number of carefully reasoned ripostes (Clarke et al., 2005; Benn and Evans, 2006; Evans et al., 2006) to the megaflood hypothesis. Also, the landsystem approach has demonstrated that the western plains contain an invaluable record of palaeo-ice stream activity pertaining to the dynamics of terrestrially-terminating systems, wherein spatial and temporal patterns of ice stream operation within an ice sheet are recorded in the regional glacial geomorphology. This forms a contrast to the vertical successions of marine sediments that record the activity of marine-terminating ice streams in offshore depo-centres such as trough-mouth fans (e.g., Vorren and Laberg, 1997; Vorren et al., 1998; Ó Cofaigh et al., 2003; Vorren, 2003; Dowdeswell et al., 2008; Ottesen et al., 2008).

The overall aim of this research is to augment recent developments of the till sedimentology and stratigraphy of the western Laurentide Ice Sheet palaeo-ice streams (Evans et al., 2012) with investigations of the landform signatures of these terrestrially-terminating systems (Fig. 1). Specific objectives include (i) the use of SRTM and Landsat ETM+ imagery and aerial photographs to map the glacial geomorphology of southern Alberta, with particular focus on the impact of the palaeo-ice streams/lobes proposed by Evans et al. (2008); and (ii) the identification of diagnostic landforms or landform assemblages (landsystem model) indicative of terrestrial-terminating ice stream margins and an assessment of their implications for reconstructing palaeo-ice stream dynamics.

2. Study area and previous research

The study area is located in the North America Interior Plains, in the southern part of Alberta, western Canada. It is bordered by the Rocky Mountain Foothills to the west, the Tertiary gravel-topped monadnocks of the Cypress Hills in the southeast, and Milk River Ridge to the south (Fig. 1; Leckie, 2006). Geologically, southern Alberta lies within the Western Canadian Sedimentary Basin, on a northerly dipping anticline known as the Sweet Grass Arch (Westgate, 1968). The Interior Plains in this area are composed of upper Cretaceous and Tertiary sediments, which consist of poorly consolidated clay, silt, and sand (Stalker, 1960; Klassen, 1989; Beaty, 1990). The preglacial and interglacial landscapes were dominated by rivers flowing to the north and northeast and which repeatedly infilled and reincised numerous pre-glacial and interglacial valleys, with sediments ranging in age from late Tertiary/early Quaternary (Empress Group) to Wisconsinan (Stalker, 1968; Evans and Campbell, 1995; Atkinson and Lyster, 2010a,b). The Cypress Hills and Del Bonita Highlands of the Milk River Ridge formed nunataks during Quaternary glaciations (Klassen, 1989).

The glacial geomorphology of southern Alberta was primarily formed during the late Wisconsinan by ice lobes/streams flowing from the Keewatin sector of the Laurentide Ice Sheet, which coalesced with

Fig. 1. Location and palaeo-ice stream maps of the study area: (A) location maps, showing the province of Alberta, Canada, and the study area outlined by two boxes. The larger box covers the area depicted in Fig. 3 and the smaller box the area depicted in Fig. 2. (B) Palaeo-ice stream map superimposed on the SRTM imagery of Alberta and western Saskatchewan, from Ó Cofaigh et al. (2010), with ice stream activity represented as numbered phases. The CAIS and HPIS are part of the phase 1 activity in the western half of the image. (C) Location map of the study area depicted in Fig. 2, showing geographical features and place names.

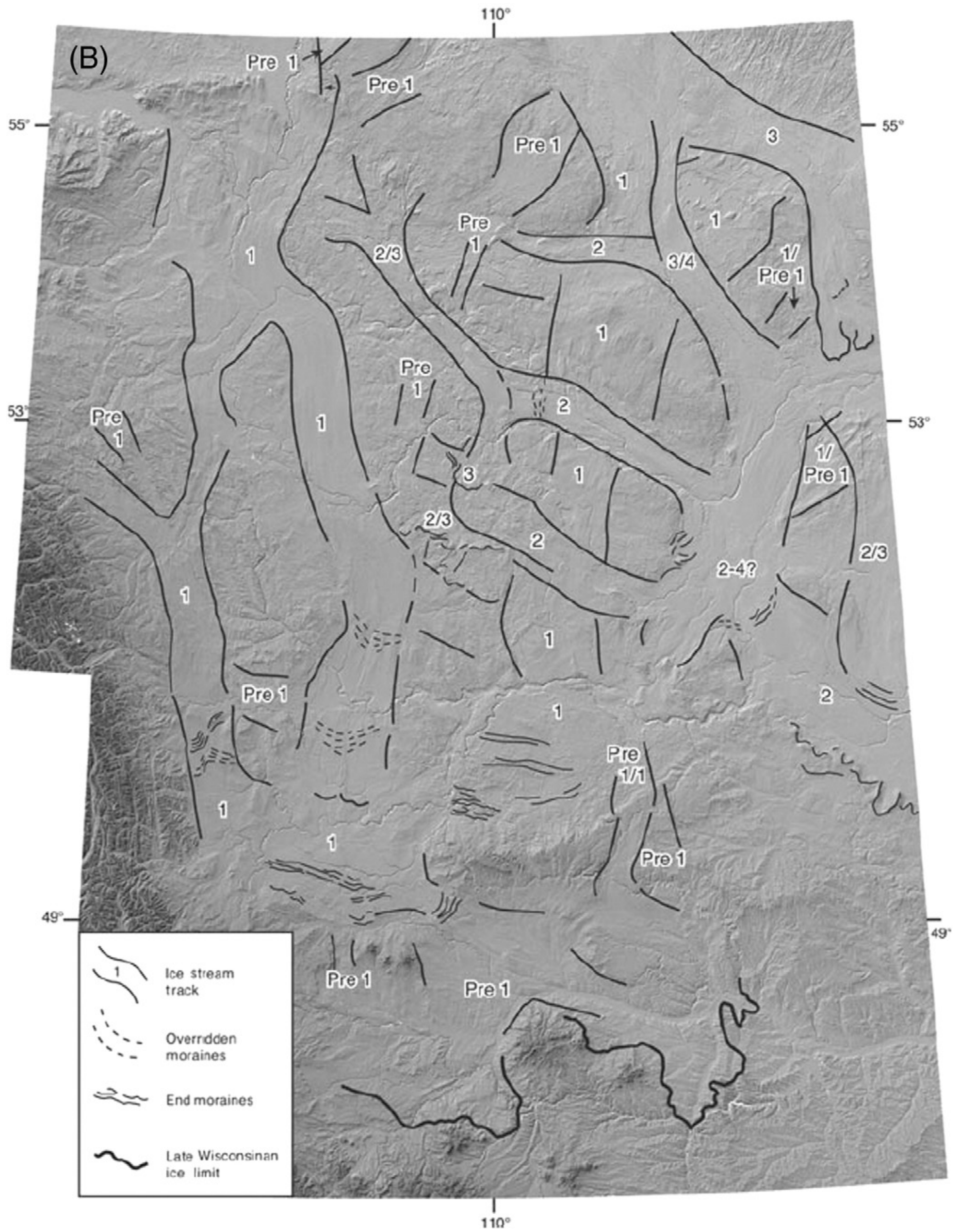


Fig. 1 (*continued*).

the Cordilleran Ice Sheet over the high plains to form a southerly flowing suture zone marked by the Foothills Erratics Train (Stalker, 1956; Jackson et al., 1997; Rains et al., 1999; Dyke et al., 2002; Jackson and Little, 2004). At its maximum during the late Wisconsinan, the 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 ka BP and had retreated to the 'Lethbridge moraine' by c. 12.3 ka BP, after which it receded rapidly to the north (Stalker, 1977; Clayton and Moran, 1982; Dyke and Prest, 1987; Dyke, 2004).

Mapping of the glacial geomorphology of southern and central Alberta (Stalker, 1960; Prest et al., 1968; Westgate, 1968; Stalker, 1977; Shetsen, 1987, 1990; Fulton, 1995; Evans et al., 1999, 2006, 2008) has enabled a broad identification of ice flow patterns and ice-marginal landform assemblages. Three prominent fast-flowing ice lobes appear to have operated within the region and were identified as the 'east', 'central' and 'west lobes' by Shetsen (1984) and Evans (2000). Recently, Evans et al. (2008) suggested that the west and central lobes be referred to as the High Plains Ice Stream (HPIS) and Central Alberta Ice Stream (CAIS), respectively, owing to their connection to corridors of highly streamlined terrain that are interpreted as the imprint of trunk zones of fast ice flow (Fig. 1B). The CAIS has also been referred to as the 'Lethbridge lobe' by Eyles et al. (1999), who highlighted that its margins were defined by the McGregor, Lethbridge, and Suffield moraine belts. These moraine belts comprise landforms of various glacigenic origins, including thrust moraines (Westgate, 1968; Stalker, 1973, 1976; Tsui et al., 1989; Evans, 1996, 2000; Evans and Rea, 2003; Evans et al., 2008), 'hummocky terrain' (cf. Gravenor and Kupsch, 1959; Stalker, 1960, 1977; Shetsen, 1984, 1987, 1990; Clark et al., 1996; Munro-Stasiuk and Shaw, 1997; Evans et al., 1999; Eyles et al., 1999; Boone and Eyles, 2001; Burgess et al., 2003; Evans, 2003; Johnson and Clayton, 2003; Evans et al., 2006; Munro-Stasiuk and Sjogren,

Fig. 1 (*continued*).

2006), and recessional push moraines and/or controlled moraine (Evans et al., 1999; Evans, 2003; Johnson and Clayton, 2003; Evans et al., 2006, 2008). Glacially overridden and streamlined moraines also appear in the trunk zones of the fast glacier flow tracks (Evans et al., 2008), although their origins and ages remain to be elucidated. Localized case studies of large-scale moraine mapping by Evans et al. (1999, 2006, 2008) have identified a spatial variability that potentially reflects changing thermal regimes at the sheet margin in addition to surging activity during later stages of recession, similar to the trends identified by Colgan et al. (2003) in the northern USA.

During deglaciation of the region, numerous proglacial lakes developed in front of the receding lobate ice stream margins, resulting in the incision of numerous spillways (Christiansen, 1979; Evans, 2000). These spillways have been either cut through preexisting preglacial valley fills or have created new flood tracks through the soft Cretaceous bedrock (Evans and Campbell, 1995). As meltwaters decanted generally eastwards they appear to have penetrated beneath the ice sheet margin in some places to produce subglacial meltwater channels (Sjogren and Rains, 1995). This pattern of drainage was most likely enhanced by the northeasterly dip of the glacioisostatically depressed land surface beneath the receding ice sheet.

A complex stratigraphy of pre-Quaternary and Quaternary glacial and interglacial deposits exists in the study region (Stalker, 1963, 1968, 1969, 1983; Stalker and Wyder, 1983; Evans and Campbell, 1992, 1995; Evans, 2000). Of significance to this study are the extensive outcrops of glacigenic sediment relating to the last glaciation, which have been employed in palaeoglaciological reconstructions of ice streams and ice sheet marginal recession patterns by Evans (2000), Evans et al. (2006, 2008, 2012), and Ó Cofaigh et al. (2010; Fig. 1B). These studies have highlighted the marginal thickening of subglacial traction tills in association with individual ice streams/lobes, thereby verifying theoretical models of subglacial deforming layers (e.g., Boulton, 1996a,b) beneath ice sheets.

The findings of the research outlined above are assimilated in this study with new observations and data on the glacigenic landforms of the region in order to assess the regional imprint of ice stream marginal sedimentation. Local variations in the patterns of landform assemblages in turn facilitate a better understanding of ice stream dynamics during the deglaciation of the southwest Laurentide Ice Sheet.

3. Methods

Glacial geomorphological mapping was undertaken by using three different aerial image sources, including the 2000 Shuttle Radar Topography Mission (SRTM 2-arc second data), Landsat 7 Enhanced Thematic Mapper Plus (Landsat ETM+) and aerial photograph mosaics flown and compiled by the Alberta Department of Lands and Forest in the 1950s. The SRTM data have been used to create digital elevation models (DEMs) of the Alberta landscape.

Global Mapper™ produced a smoothed, rendered pseudo-colour image of the SRTM data that could be manipulated to accentuate features, produce three-dimensional images and change sun illumination angles. Following the procedures of Smith and Clark (2005), multiple illumination angles were also used during mapping. An alternative method was employed to compare, verify, and supplement the SRTM mapping. This involved the use of ENVI 4.3 software to open the SRTM data in a grey-scale format; nearest neighbour sampling was used to correct for missing sample points and was automatically applied to the same missing data points when opening the images in Global Mapper. Additional geomorphological mapping was conducted through interpretation of the high resolution Landsat ETM+ panchromatic band (band 8: 0.52–0.90 μm) images.

The SRTM and Landsat ETM+ mapping is at a scale appropriate to the identification of regional scale landform patterns, including subglacial bedforms and cross-cutting lineations (Clark, 1999). Once identified and mapped, lineations were divided into flowsets using characteristics such as parallel conformity, length, and morphology; they were then simplified by drawing flowlines aligned with and parallel to the lineation direction, following the procedures outlined by Clark (1997, 1999), so that each flowset represents a collection of glacial features formed during the same flow phase and under the same conditions. 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. Such quantitative analyses of subglacial bedforms have been widely demonstrated to be critical in the reconstruction of palaeo-ice streams and their dynamics (e.g., Stokes and Clark, 2003; Roberts and Long, 2005; Stokes et al., 2006; Storrar and Stokes, 2007).

A series of ten, 1:63,360 (1 in. to 1 mile) aerial photograph mosaics captured in 1951 by the Alberta Department of Lands and Forest were utilized for large-scale investigations of the geomorphology of the southern Alberta palaeo-ice streams. Landforms were mapped according to their morphometric characteristics prior to interpretation, although genetic terms were later used to identify features on the maps. Linear depositional features, ice flow-parallel (flutings, eskers) and ice flow-transverse (major and minor ridges and moraines) were mapped as single lines representing their summit crests. In areas of 'hummocky terrain' (sensu Benn and Evans, 2010), the complexity and density of individual hummocks rendered the mapping of every mound inappropriate; and hence the hummocky terrain is represented by black shading of the interhummock depressions. This approach effectively illustrates the relative degrees of linear versus chaotic patterns.

4. Results of geomorphological mapping

4.1. Regional palaeo-ice stream geomorphology: small scale mapping case studies of the HPIS and CAIS tracks

The glacial geomorphology of southern Alberta is dominated by the imprints of two fast ice flow or palaeo-ice stream tracks (the HPIS and CAIS of Evans et al., 2008), which appear as corridors of smoothed topography bordered by lobate marginal landforms and inter-lobate/inter-stream hummocky terrain. Also, in the eastern part of the province, the subglacial bedforms and marginal moraines of Ó Cofaigh et al.'s (2010) 'Ice Stream 1' ('east lobe' of Shetsen, 1984; Evans, 2000) terminate on the north slopes of the Cypress Hills. Previous work on regional mapping in Alberta by Evans et al. (2008) identified the fast flow tracks and various ice-flow transverse ridges, some of which were difficult to interpret because of the low resolution of the DEMs available at the time. Here we report on the comprehensive and systematic mapping and quantification of landforms in the HPIS and CAIS tracks (Figs. 1 and 2) based on higher resolution SRTM data and the further development of that mapped by Ó Cofaigh et al. (2010; Fig. 1B).

The study area contains the 250-km-long HPIS (Evans et al., 2008), which varies in width from 50 km along the main trunk to 85 km across the lobate terminus, and approximately 320 km of the total length of the CAIS, over which distance its width increases from 97 to 160 km at its lobate margin (Figs. 2 and 3). A total of 714 flow-parallel lineations were identified along the CAIS and HPIS and together comprise seven individual flowsets (defined using the criteria proposed by Clark, 1997, 1999), although large areas of the smoothed corridors that demarcate the fast flow tracks do not contain terrain sufficiently fluted to enable confident flowset mapping (Fig. 3). The main landforms in the HPIS trunk include at least five (Hfs_1–5) different flowsets (Fig. 3), four of which (Hfs_2–5) record marginal splaying or lobate flow within the HPIS toward the McGregor moraine belt. One flowset (CAfs_1) was identified along the CAIS trunk and one (CAfs_2) in its southeast corner (Fig. 3), each flowset relating to different phases of ice stream flow.

Flowset Hfs_4 contained the largest number of lineations (260), although all flowsets tended to display strong spatial coherency; and CAfs_1 contained the longest lineation (35 km; cf. Evans, 1996). Because of the resolution of SRTM imagery, elongation ratios (ERs) could not be accurately determined; however, most lineations have ERs of >10:1. The smallest lineations were found in Hfs_1 (see Table 1 for flowset data).

Flowsets display distinct relationships with ice flow transverse ridges and/or hummocky terrain arcs (Evans et al., 2008). Extensive sequences of transverse ridges exist throughout the study area, not only in marginal settings as sharp-crested features but also along the HPIS and CAIS flow corridors as smoothed or streamlined features (Figs. 2, 4–8). These ridges are loosely classified below as minor or major features according to their relative sizes.

Transverse ridges and hummocky terrain associated with the HPIS reveal a clear pattern of ice-marginal advance and recession. For example, flowsets Hfs_4 and 5 terminate in zones of hummocky terrain and/or minor transverse ridges, demarcating lobate ice marginal positions that are compatible with the flowsets that terminate on their proximal sides (cf. Evans et al., 1999, 2006, 2008). The landform assemblage TR_1a occupies 100 km of the western half of the HPIS track and includes an extensive sequence of low amplitude (3–6 m high), inset, and arcuate minor transverse ridges (cf. Evans et al., 1999; Evans, 2003; Johnson and Clayton, 2003). These minor ridges appear to be draped over, or superimposed on, two major ridges (TR_1(N) and

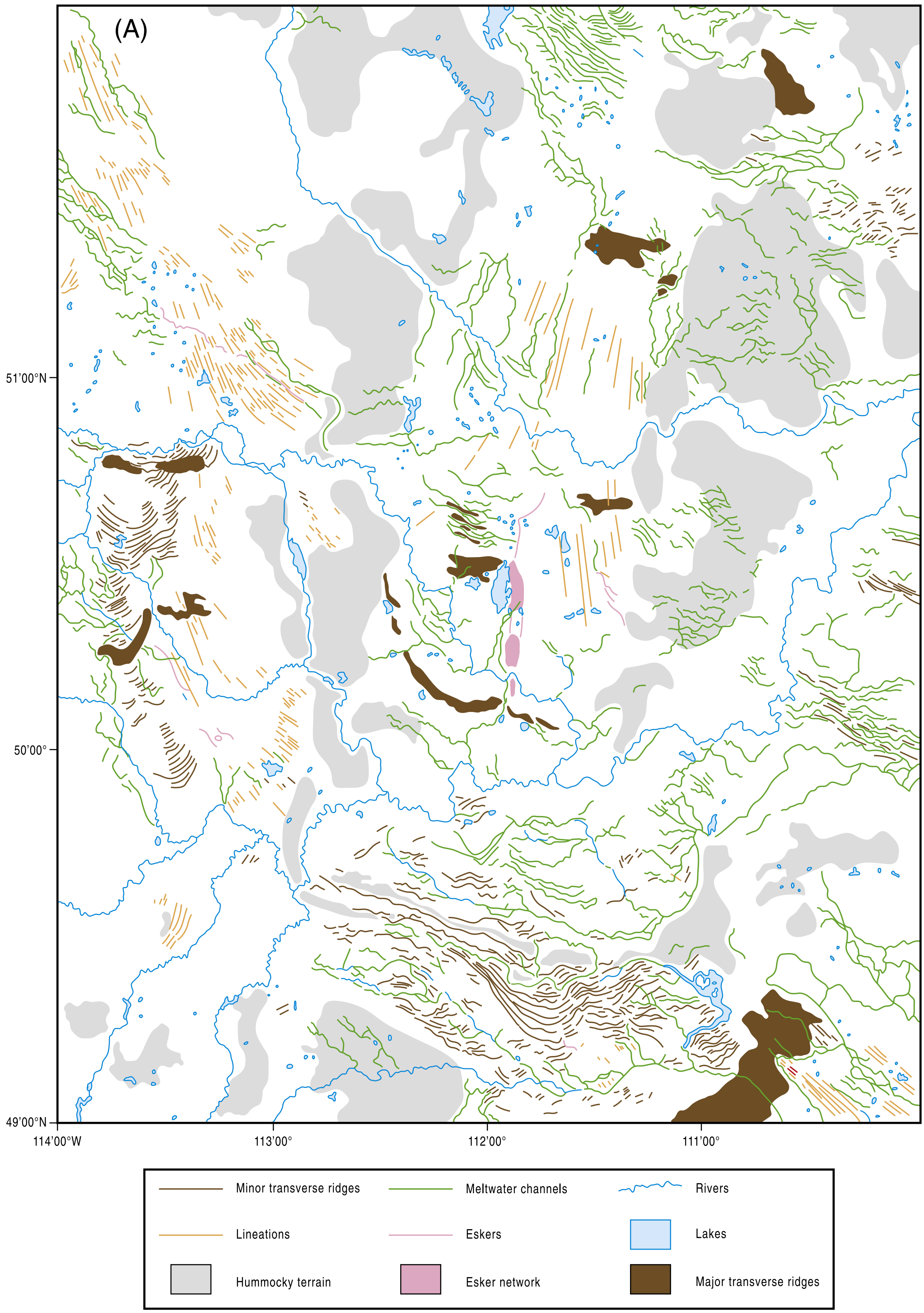


Fig. 2. Glacial geomorphology map of southern Alberta based upon the mapping of SRTM imagery undertaken in this study: (A) map of landforms with genetic classifications; (B) map of landforms annotated with place names and the locations of figures used in this paper, the transverse ridge sets, and topographic cross profiles A–E (see Fig. 7).

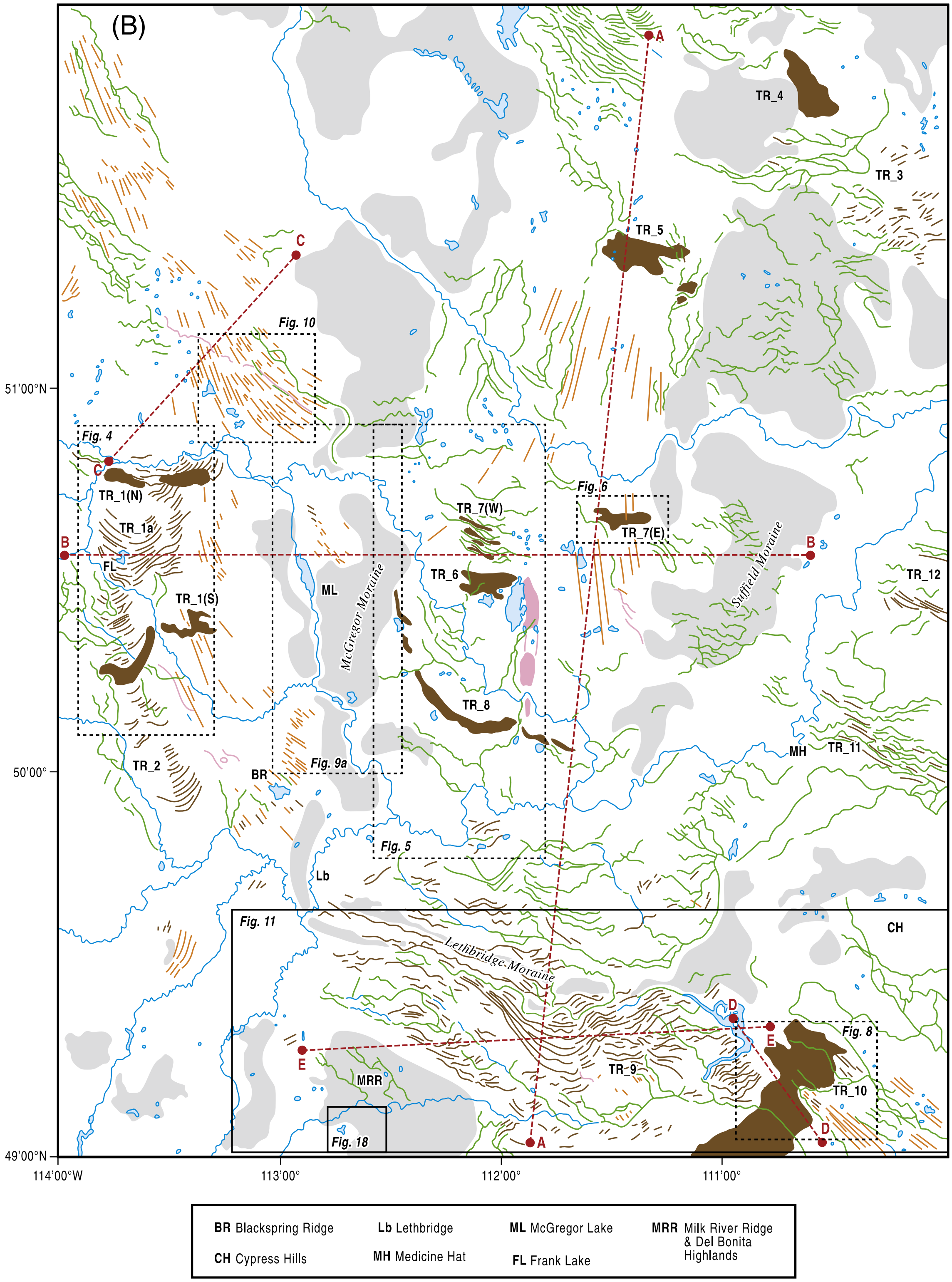


Fig. 2 (*continued*).

TR_1(S); Fig. 4). The summits of the two major ridges each comprise up to five component subridges 10–15 m high and are overprinted by flutings, the most prominent relating to flowset Hfs_5 (Fig. 4), which continues in a southeasterly direction to cover the area known as Blackspring Ridge (Fig. 2; Munro-Stasiuk and Shaw, 2002). A further extensive series of inset arcuate minor ridges (TR_2) lies immediately south of the southernmost major ridge and, together with the TR_1a sequence, has previously been interpreted by Evans et al. (1999) and Evans (2003) as a recessional push moraine sequence that has been deposited over the Hfs_5 flutings and the larger transverse ridges TR_1(N) and TR_1(S).

On the CAIS footprint, CAfs_1 terminates north of the largest major transverse ridge in the study area (TR_8; Fig. 5), which displays a bi-lobate front and is 70 km long and crosses most of the CAIS between the Bow and Oldman Rivers, with its eastern edge connecting to an area of hummocky terrain. The ridge is weakly asymmetric, with a steeper

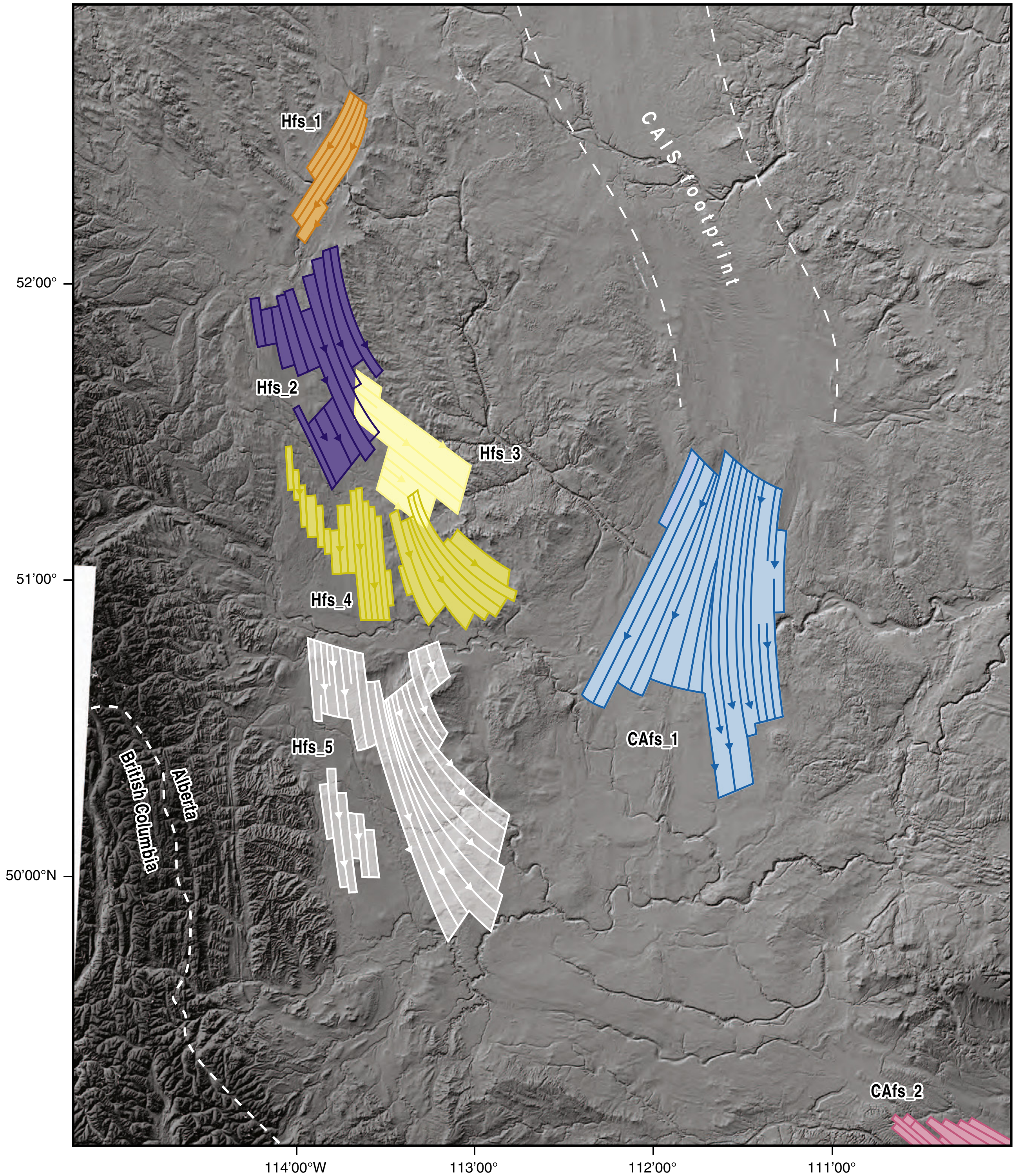


Fig. 3. Flowsets reconstructed from glacial lineations. Lineations were grouped into flowsets based primarily on their orientation and their proximity and location (Clark, 1999). Hfs_1–5 relate to the High Plains Ice Stream and CAfs_1 and 2 relate to the Central Alberta Ice Stream.

distal slope and its height gradually increases from west to east from 20 to 30 m. The centre of flowset CAfs_1 is connected to TR_8 via an esker network (Evans, 1996, 2000) that joins the ridge at its re-entrant or inflexion point (Figs. 2 and 5). Two sets of major transverse ridges also occur in the area located between major ridge TR_8 and the southern end of flowset CAfs_1 (Figs. 5 and 6). Assemblage TR_7(E) comprises broad, shallow, parallel ridges superimposed with numerous discontinuous, narrow, and sharp ridges that give the landform a corrugated appearance (Fig. 6). These have previously been interpreted as glacitectonic thrust ridges by Evans and Campbell (1992) and Evans (2000) based upon field exposures displaying deformed Cretaceous bedrock overlain by till. Assemblage TR_7(W) includes only the narrow, sharp ridges, which appear to continue southward into those in TR_6 but occupy proglacial/spillway flood tracks previously mapped by Evans (1991, 2000) and therefore have most likely been accentuated by fluvial erosion.

Table 1
Data showing the specific characteristics of the flowsets, which in turn act as a device to help differentiate between particular flowsets.

Flowset	Number of lineations	Mean length (km)	Mean direction (°)	Flowset area (km^2)
Hfs_1	81	1.56	224	702
Hfs_2	110	3.42	141	3162
Hfs_3	66	2.34	119	1631
Hfs_4	260	3.58	170	4150
Hfs_5	147	3.52	160	5964
CAfs_1	30	10.0	182	6154
CAfs_2	20	4.17	118	849

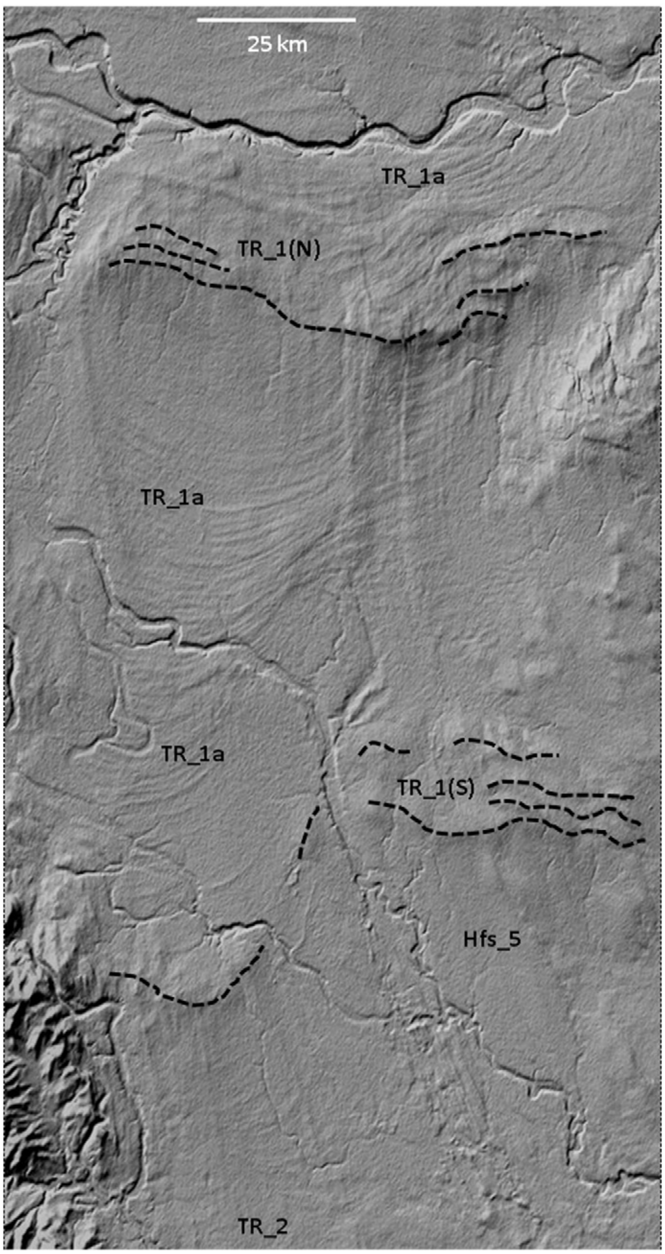


Fig. 4. SRTM data of transverse ridges situated along the HPIS trunk. TR_1a and TR_2 are minor transverse ridges and TR_1(N) and TR_1(S) are major transverse ridges whose significant ridge crests are outlined by broken lines. Note that the minor ridges drape the major ridges. The major ridges are also superimposed by the streamlined features (flutings) that make up Hfs_5, particularly to the right and bottom of the image. An esker network also appears in the bottom right corner.

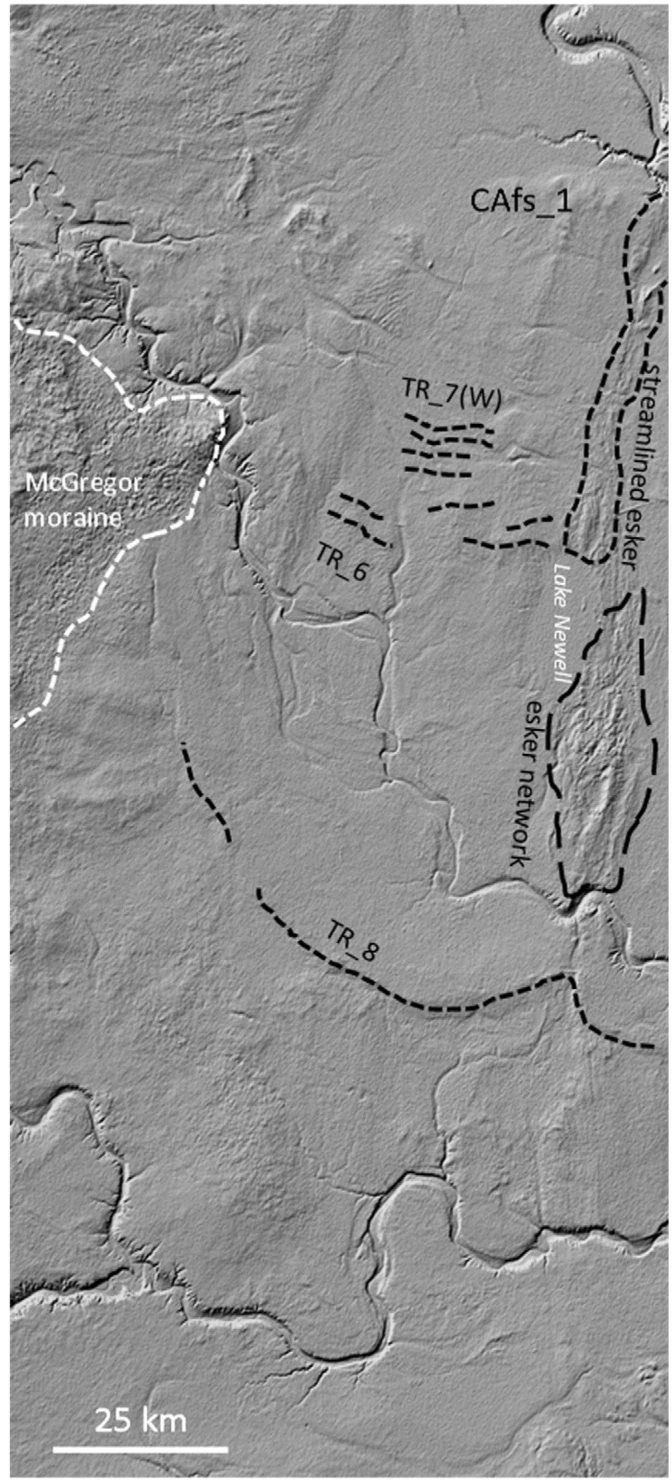


Fig. 5. SRTM data of major transverse ridges, particularly the large lobate ridge TR_8, situated along the CAIS. The Bow River flows through the centre of the image and the Oldman River along the bottom. Also shown are TR_6, TR_7(W), and an esker network situated to the right centre of the image, the northern part of which has been streamlined as part of CAfs_1.

Further north in the CAIS footprint, CAfs_1 apparently starts immediately downflow of a streamlined, major transverse ridge complex (TR_5; cf. Evans, 1996; Evans et al., 2008), comprising three parallel subsets of ridges rising up to 30 m above the surrounding terrain (Fig. 7A). In detail, the sequence is composed of 40 ridges, ranging from 1 to 4 km in length. Other transverse ridges in this area, but outside the CAIS footprint, include a cluster of minor ridges (TR_3),

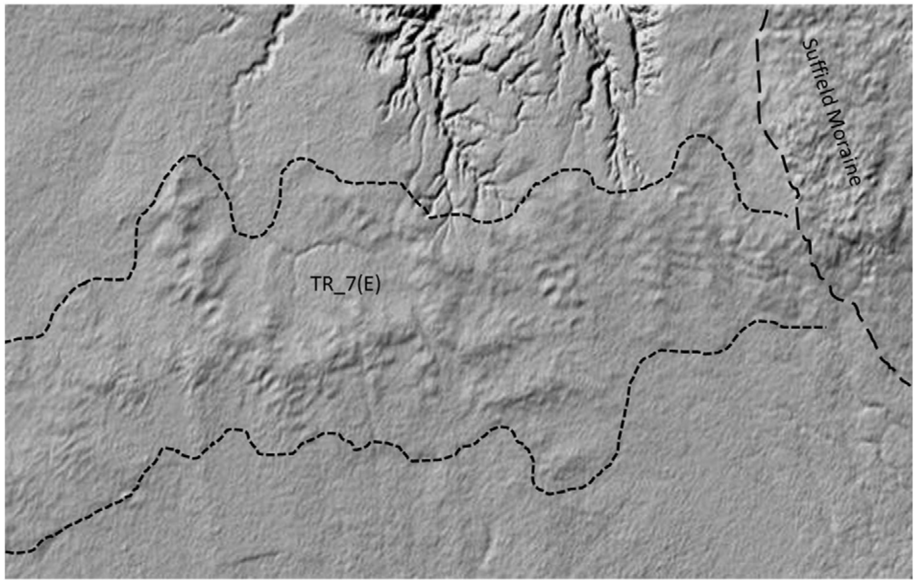


Fig. 6. SRTM data of the TR_7(E) major ridge (outlined by short dash lines) located immediately south of the Dinosaur Provincial Park badlands (visible at top of image) and showing the details of the discontinuous ridged crestline that gives the landform a corrugated appearance. The hummocky terrain of the Suffield moraine is visible at the top right of the image (outlined by long dash line).

30 km long and 10–20 m high, with crest wavelengths of 500–1000 m, and bordered by hummocky terrain to the east, west, and south. Individual ridges within the sequence are only a few kilometres in length. To the northwest of TR_3 are several large ridges set within and dominating an area of hummocky terrain (TR_4). The ridge crests are 10 km long and stand up to 20 m above the surrounding hummocks. These large transverse ridge complexes are strongly asymmetric with steeper, north-facing, proximal slopes.

In the extreme south of the study area, on the preglacial drainage divide that was located between the Cypress and Sweet Grass hills (Westgate, 1968) and 150 m above the Pakowki Lake depression (Fig. 7D), flowset CAfs_2 is located on the summit and downice side of major ridge assemblage TR_10. The summit of TR_10 comprises a series of prominent and closely spaced, sharp crested transverse ridges (Fig. 8) that vary in height from 20 to 5 m and in wavelength from 1 km to 250 m and give the landform a corrugated appearance. Further details of the smaller transverse summit ridges on TR_10 and their relationships with minor transverse ridges and flutings in the area are presented in the next section based upon aerial photograph mapping.

Ridge complex TR_10 is separated from TR_8, located 130 km to the north, by a wide zone of minor transverse ridges, including the 'Lethbridge moraine' of Stalker (1977; Fig. 2), which has been developed on the northern slopes of Milk River Ridge and in the Milk River drainage basin. Immediately south of the Lethbridge moraine lies a 45-km-wide and 150-km-long arc of low amplitude, minor transverse ridges (TR_9; Fig. 2B), associated with numerous ridge-parallel meltwater channels and coulees (Fig. 7E). This landform assemblage has been mapped at greater detail using aerial photographs and is reviewed in the next section as a landsystem indicative of lobate terrestrial ice stream margins.

Two further sets of minor transverse ridges (TR_11 and TR_12) are located at the southwest corner of Ó Cofaigh et al.'s (2010) Ice Stream 1. These landforms record the incursion of an 'east lobe' onto the northern slopes of the Cypress Hills and against the east side of the Suffield moraine (Fig. 2).

Hummocky terrain covers a large proportion of the study area and defines the margins of palaeo-ice stream/lobe tracks (cf. Evans, 2000; Evans et al., 2008). It occurs primarily between the smoothed, fast ice flow corridors (Fig. 7B) and along the southern margin of the CAIS (Figs. 2 and 9). The SRTM and Landsat ETM+ imagery reveals a pattern of hummocky terrain that is similar to that depicted by Prest et al. (1968), Shetsen (1987, 1990), Clark et al. (1996), and Evans (2000). Detailed mapping of the landforms that occur within the hummocky terrain belts, particularly in the McGregor moraine (Fig. 9) has previously revealed that they comprise areas of linear to chaotic hummock chains interspersed with minor ridges, interpreted by Evans (2000, 2009) and Evans et al. (2006) as a landform imprint of glacier margins that alternated between polythermal and temperate basal conditions during recession. Significantly in this respect, hummocky terrain bands (part of Stalker's, 1977, Lethbridge moraine) run continuously from the edge of Blackspring Ridge across the CAIS marginal area up to and around the Cypress Hills. In planform the hummocky terrain bands demonstrate a strong lobate pattern and run parallel to intervening belts of transverse ridges, even though they internally consist of chaotic hummocks. The SRTM data reveal that the hummocky terrain and associated minor ridges are superimposed on larger physiographic features (Fig. 9A), which are likely representative of remnant uplands in the preglacial land surface (Fig. 1B; cf. Leckie, 2006). The regional distribution and prominence of this hummocky terrain or moraine belt have previously been employed to demarcate the limit of the Lethbridge lobe (Eyles et al., 1999) or CAIS (Evans et al., 2008, 2012) and that this was superimposed on the HPIS footprint. The details of the hummocky terrain and associated minor ridges are presented at larger scale in the next section through a case study of the CAIS ice-marginal landsystem.

Eskers are found on the small-scale imagery throughout the study area as narrow winding ridges, but resolution constraints allowed the identification of only the largest features. The largest esker identified in this study was 45 km long and situated along Hfs_4 (Fig. 10). Further south, a sequence of prominent eskers is situated along the centre of the HPIS corridor, particularly in association with Hfs_5 (Figs. 2 and 4), forming a 40-km-long network running parallel to lineation direction. Another prominent network of eskers is located along the eastern edge of Lake Newell and emerges 20 km south of CAfs_1 and terminates just south of Lake Newell at the inflexion point of the dual-lobate ridge TR_8 (see above; Fig. 5; cf. Evans, 1996, 2000). Additional eskers were identified along the centre and eastern half of the CAIS.

4.2. Ice stream/lobe marginal landsystem: large-scale mapping case study of the CAIS

Although ice flow transverse ridges have been identified at a regional scale, as described above (Figs. 2, 4–8), landform mapping from aerial photographs in combination with the SRTM data (Fig. 11) reveals a complex glacial geomorphology at larger and more localized scales, comprising minor transverse ridges, hummocky terrain, flutings, and meltwater channels/spillways. These features have been developed on a land surface characterized by Tertiary gravel-capped monadnocks (e.g., Del Bonita highlands/Milk River Ridge, Cypress Hills) and substantial depressions related to long term drainage networks (e.g., Pakowki Lake depression). Previous research has investigated the nature and origins of minor transverse ridges at the margins of the HPIS and CAIS in the McGregor moraine belt, concluding that spatial variability in morphology (controlled moraine to push moraine) likely reflects changes in the basal thermal regime of the ice sheet margin during recession (Evans et al., 2006; Evans, 2009). In order to test this hypothesis, the minor transverse ridge assemblages that demarcate the receding lobate margins of the CAIS are now analysed in detail.

4.2.1. Transverse ridges of the CAIS southern margins

Transverse ridges are aligned obliquely to former ice flow and are in places contiguous with bands of hummocky terrain, forming large arcuate bands and thereby allowing the regional lobate pattern of ice stream marginal deposition to be mapped. At larger scales the transverse ridges display significant variability in form and thereby inform a higher resolution palaeoglaciology. The majority of transverse ridges are located to the south and southeast of the Lethbridge moraine and Etzikom Coulee and the most extensive sequences lie directly south of Crow Indian Lake (in Etzicom Coulee), Verdigris Coulee, and southeast of Pakowki Lake (Figs. 2B and 11), where they document the early recessional phases of the CAIS margin. Within the CAIS marginal setting, three types of minor transverse ridge sets are identified and classified as MTR types 1–3 (Figs. 12–15); MTR type 1 relate to the series of prominent and closely spaced, sharp-crested transverse ridges that comprise the corrugated summit of TR_10 (Fig. 8). Additionally, three types of hummocky terrain form are recognized and classified as HT types 1–3 (Figs. 16 and 17).

The MTR type 1 have largely symmetrical cross profiles and consistent wavelengths (Figs. 8 and 12), occur only in the southeast corner of the CAIS margin on the TR_10 ridge complex (Figs. 2, 8 and 11), and are large enough to be identified in the regional mapping using the SRTM data (Fig. 8). Because of its ripple-like or corrugated appearance in planform, the TR_10 ridge complex has been interpreted by Beaney and Shaw (2000) as an erosional surface scoured by subglacial megaflood waters. Our large-scale mapping reveals that the complex ridge TR_10 comprises closely spaced, sharp-crested ridges that are remarkably parallel over most of their lengths despite significant plan form crenulations (Fig. 8). The intervening hollows are commonly filled with numerous, elongate small lakes (Figs. 11 and 13), and the aerial photographs reveal that the ridges are more widely overprinted by flutings than is apparent from the SRTM image (compare Figs. 8 and 13). These MTR type 1 pass northwesterly into MTR type 2 at around the location where the regional topographic slope drops abruptly into the Pakowki Lake depression (Fig. 7). Discontinuous and faint flutings, which are widespread in association with the MTR type 2, continue across the boundary between the MTR types 1 and 2; their long axis alignments are parallel to those of CAfs_2 flutings but they are less prominent, even though they are contiguous. These fluting characteristics indicate that TR_10 has been overridden by glacier ice, likely during the production of CAfs_2.

The MTR type 2 are characterized by low relief and sharp-crested ridges with largely asymmetrical cross profiles and variable wavelengths; ridges often locally overlap or overprint each other and possess crenulate or sawtooth planforms (Fig. 12; Evans, 2003). They dominate primarily flat terrain of the Milk River drainage basin (Fig. 11). They are characterised by conspicuous ridge sets up to 5 m high and with generally continuous crests (Fig. 14). The ridges located along the southeast margin of Pakowki Lake extend for up to 15 km, but in general the ridges range from 1to 5 km long. The ridges situated south of the Milk River (Fig. 11) are more subtle and smaller than those to the southeast of Pakowki Lake.

The MTR type 3 are characterized by discontinuous, low relief, and sharp-crested ridges that are aligned parallel and contiguous with chains of hummocks to form continuous lines when viewed over large areas. Between the high points, strongly orientated depressions often filled with ponds and occasionally containing isolated hummocks accentuate the overall linearity (Figs. 12 and 15). They are the most common ridge type located to the west of Pakowki Lake and are most extensive just south of Etzikom Coulee and Verdigris Coulee (Fig. 11). Individual ridges and associated hummocks are more subtle than MTR type 2, with smoothed crests and heights generally no greater than 3 m. They also show clear lobate form on both the regional and large-scale geomorphology maps (Figs. 2 and 11). Like MTR type 2, the type 3 ridges also demonstrate subtle overlapping or overprinting (Fig. 15A).

4.2.2. Hummocky terrain of the CAIS southern margins

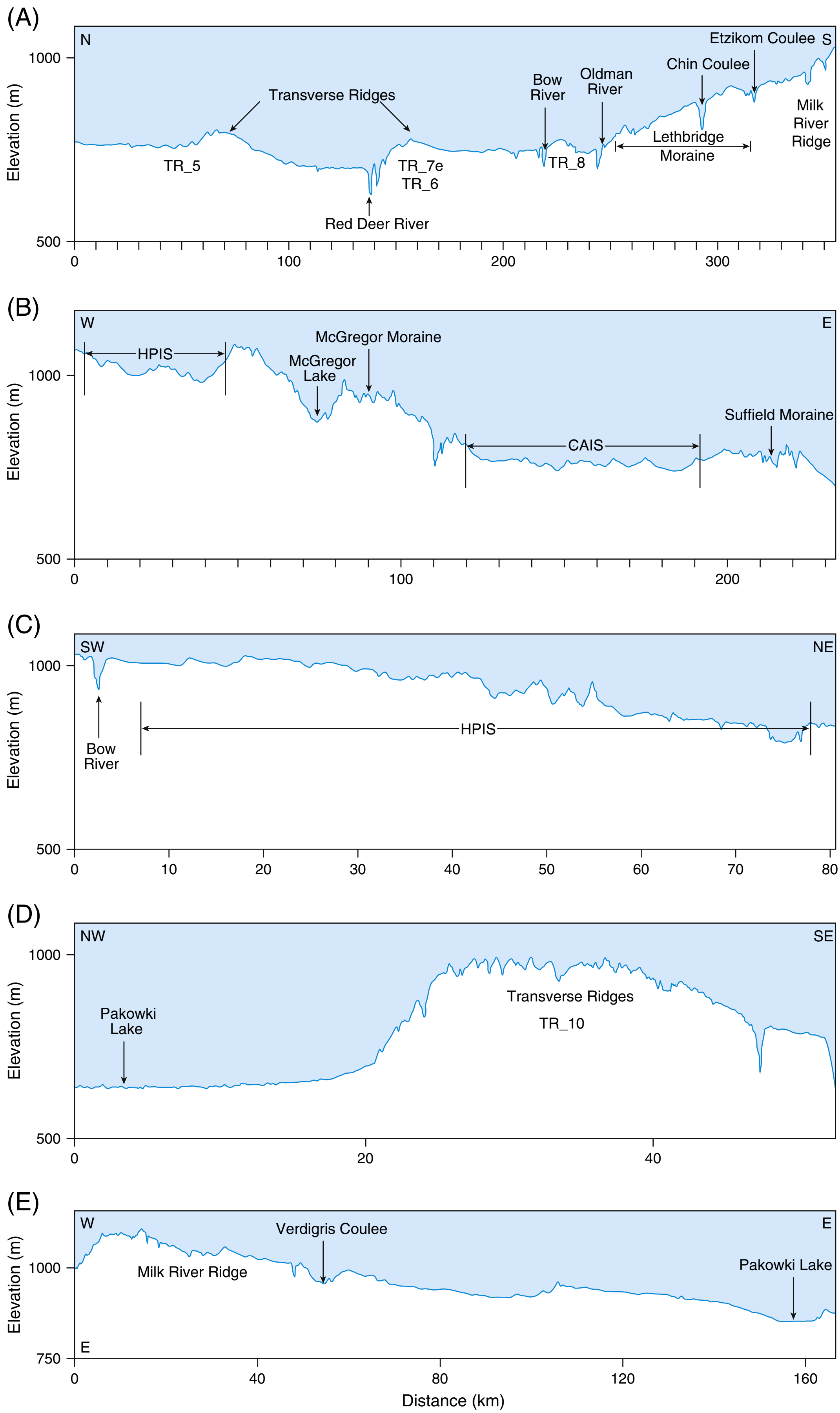
Hummocky terrain is the most common landform within the CAIS marginal zone and contains a wide range of hummock types (Figs. 16–18). At large scales, hummock assemblages are chaotic and demonstrate little to no linearity, but when viewed at smaller scales they exhibit curvilinear or lobate patterns aligned parallel to sequences of transverse ridges (Figs. 2 and 11). 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 the area between Etzikom and Chin Coulees to north of Pakowki Lake (Fig. 11). This hummocky terrain forms part of the Lethbridge moraine which extends from Lethbridge to the north slopes of the Cypress Hills (Fig. 2; Westgate, 1968; Bik, 1969; Stalker, 1977). Hummocky terrain also occurs in the southwest corner of the study area, where it wraps around the Del Bonita highlands and along the Milk River Ridge. Close inspection of these hummocky terrain bands reveals three different types of hummock (HT types 1–3; Fig. 17).

The HT type 1 hummocks form the majority of the hummocky terrain and consist of densely spaced, low relief hummocks with little or no orientation but when viewed over larger areas can appear to be crudely aligned (Figs. 16 and 18). The hummocks vary significantly in size, up to 5 m in height and generally <30 m in diameter (Fig. 17). Their morphology varies from individual circular and oval shaped hummocks to interconnected larger hummocks with less rounded tops. The HT 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.

The HT type 2 hummocks are generally randomly juxtaposed with HT type 1 and form occasional larger zones within other hummocky terrain bands (Fig. 16C). They are characterised by circular mounds with a cylindrical, often water-filled hollow at their centre (Fig. 17). This creates a ring or *doughnut* shape that is noticeably different in morphology to HT type 1 hummocks. Conspicuous ridges also occur within the larger zones of HT type 2 hummocks (Fig. 18). These ridges weave through the hummocks, showing no singular orientation, and occasionally make up parts of the rims of hummocks.

The HT type 3 hummocks are the largest of the hummock types, being up to 20 m high and 1 km wide (Fig. 17). They have a roughly cylindrical to oval planform and are up to twice as high as the surrounding hummocky terrain. Some have large rims and all have a flat surface. They are the least common of the three hummock types but the most conspicuous. The HT type 3 hummocks are best developed and

Fig. 7. Topographic profiles taken from SRTM data (see Fig. 2 for location) across the study area: (A) long profile of the bed of the CAIS; (B) transverse profile across the beds of the HPIS and CAIS and the McGregor and Suffield moraine belts; (C) transverse profile across the terrain traversed by the HPIS; (D) ice flow parallel profile from Pakowki Lake across the transverse ridges located on the preglacial drainage divide in southeastern Alberta; (E) transverse profile across the terrain covered by the CAIS marginal landforms.


(A)
N
S
Elevation (m)
1000
500
0
100
200
300
Transverse Ridges
TR_5
TR_7e
TR_6
Red Deer River
Bow River
TR_8
Oldman River
Lethbridge Moraine
Chin Coulee
Etzikom Coulee
Milk River Ridge
(B)
W
E
1000
500
0
100
200
HPIS
McGregor Moraine
McGregor Lake
CAIS
Suffield Moraine
(C)
SW
NE
1000
500
0
10
20
30
40
50
60
70
80
Bow River
HPIS
(D)
NW
SE
1000
500
0
20
40
Pakowki Lake
Transverse Ridges
TR_10
(E)
W
E
1000
750
0
40
80
120
160
Milk River Ridge
Verdigris Coulee
Pakowki Lake
E
Distance (km)

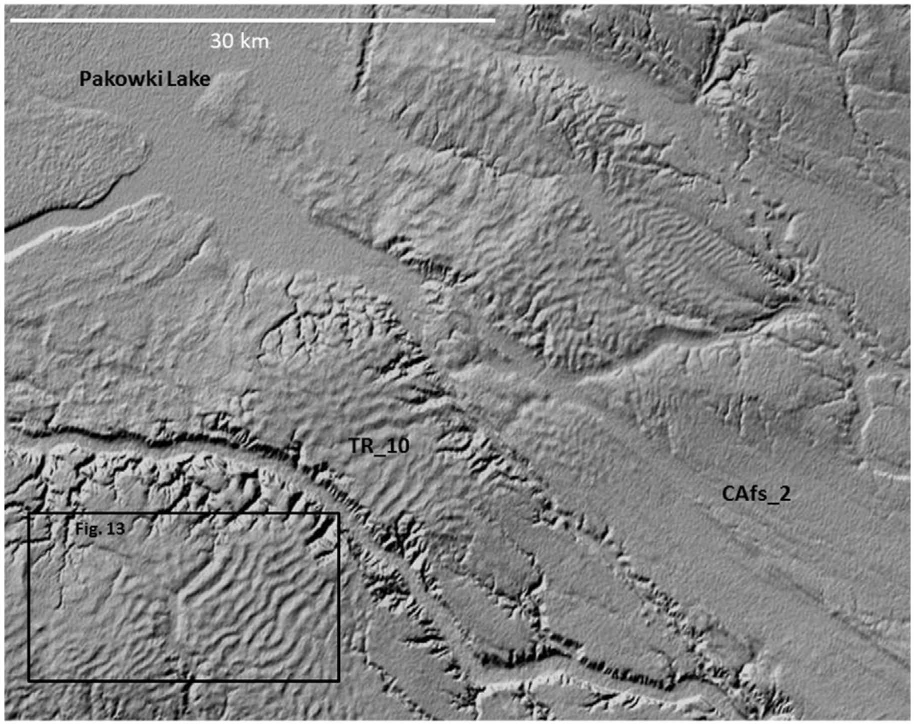


Fig. 8. SRTM data of the sequence of MTR type 1 ridges in the southeastern corner of Alberta (TR_10). Note the lineations situated just downice of the ridges (CAfs_2) and the smooth flat topography in the northwest corner representing Pakowki Lake. Although the ridges appear nonfluted in this image, they are overprinted by faint flutings, as illustrated by Fig. 13 (see box for location).

primarily located in the southwest corner of the study area around the Del Bonita highlands (Fig. 18).

4.2.3. Flutings of the CAIS southern margins

Flutings near the margin of the CAIS are located predominantly along the eastern portion of the Milk River and south and southeast of Pakowki Lake, and north of Tyrrell Lake (Fig. 11). They range from 1 to 9 km in length with an average of 2 km. Flutings located north and south of the Milk River clearly overprint MTR type 1 (Figs. 11 and 13) at right angles and are <2 m in amplitude, making them difficult to recognize on the ground (Westgate, 1968). The flutings that constitute flowset CAfs_2 are notably larger than any other lineations in the CAIS marginal zone, individual forms being up to 9 km long and 6 m high and the whole flowset covering an area 30 km long and 5 km wide. As a result, the aerial photographs reveal at least double the amount of flutings compared to the SRTM data. This scale of resolution allows further assessment of fluting dimensions, including elongation ratios, which range from 12:1 up to 85:1 along the CAfs_2 with fluting length increasing in a downflow direction.

4.2.4. Spillways at the CAIS southern margins

Four major spillways extend across the study area including Forty Mile Coulee, Chin Coulee, Etzikom Coulee and Verdigris Coulee and lie parallel to the transverse ridges, conforming to the lobate planform displayed by the ice-marginal landform record (Fig. 11). They extend across the majority of the Lethbridge moraine sequence as dominant features, reaching up to 500 m wide and 60 m deep (Fig. 19). An extensive network of smaller channels situated north of Chin Coulee (Figs. 11 and 19) lies predominantly parallel and perpendicular to the spillway. These shallow channels are up to 10 km long and 200 m wide (Fig. 19). Longer channels up to 20 km long and 100 m wide are found to the north of Crow Indian Lake in Etzicom Coulee, dissecting the hummocky terrain band at right angles. Only a few eskers were identified and are located chiefly in the northeast corner of the area mapped in Fig. 11.

5. Interpretations of geomorphology mapping

5.1. Smoothed corridors, lineations and flutings

Smoothed corridors of terrain on the plains of western Canada have been previously interpreted as palaeo-ice stream tracks (Evans et al., 2008; Ó Cofaigh et al., 2010) based in part upon the geomorphological criteria proposed by Stokes and Clark (1999, 2001; Table 2). The corridors contain megascale glacial lineations (MSGL) or flutings and are delineated by a change in smoothed topography, created by fast ice flow, to hummocky terrain associated with slow moving, cold based ice and stagnation (Dyke and Morris, 1988; Stokes and Clark, 2002; Evans et al., 2008; Evans, 2009; Ó Cofaigh et al., 2010). Similarly, we compare the lineations and smoothed topography of southern Alberta 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), and thereby substantiate propositions for the former occurrence of the HPIS and CAIS in the southwest Laurentide Ice Sheet. The onset zones of the HPIS and CAIS are unknown, and mapping by Prest et al. (1968) and Evans et al. (2008) does not identify any clear convergent flow patterns. However, we identify divergent flow patterns in the flowsets Hfs_4 and Hfs_5 and CAfs_1. Additionally, pebble lithology data (Shetsen, 1984) demonstrate a Boothia type (Dyke and Morris, 1988) dispersal by the HPIS and CAIS. Topographic cross profiles (Fig. 7B) and bedrock topographic maps (Geiger, 1967; Atkinson and Lyster, 2010b) reveal that the CAIS is a pure ice stream and that the HPIS is a predominantly topographic ice stream (Clark and Stokes, 2003).

Few flowsets were identified along the CAIS track, but a lack of obvious cross-cutting patterns hampers any identification of changing flow directions. However, the orientation of flowset CAfs_1 appears to relate to lobate ice flow toward the bi-lobate ridge TR_8 (Figs. 3 and 5), indicating that TR_8 could represent the maximum position of a readvance during which flowset CAfs_1 terminated at a lobate ice margin. Transverse ridge sets TR_6 and TR_7 appear to represent later readvances by the CAIS lobe that terminated north of TR_8. This would explain the streamlining of a major esker network by CAfs_1 to the north of TR_6 and TR_7 and its preservation in a nonstreamlined state to the south (Evans, 1996, 2000). The nonstreamlined section of the esker network documents the development of a significant subglacial/englacial drainage pathway at the junction of two ice flow units in the CAIS; this is especially apparent where the esker is coincident with the re-entrant within the dual lobate TR_8 ridge (Fig. 5; see Section 5.2 below).

In the marginal zone of the CAIS in south and southeast Alberta (Fig. 11), MSGL and smaller flutings overprint MTR types 1 and 2, specifically to the south and southeast of Lake Pakowki. The minor flutings in the area run parallel to flowset CAfs_2 and thus, based on their strong parallel coherence, are interpreted to represent the same flow event.

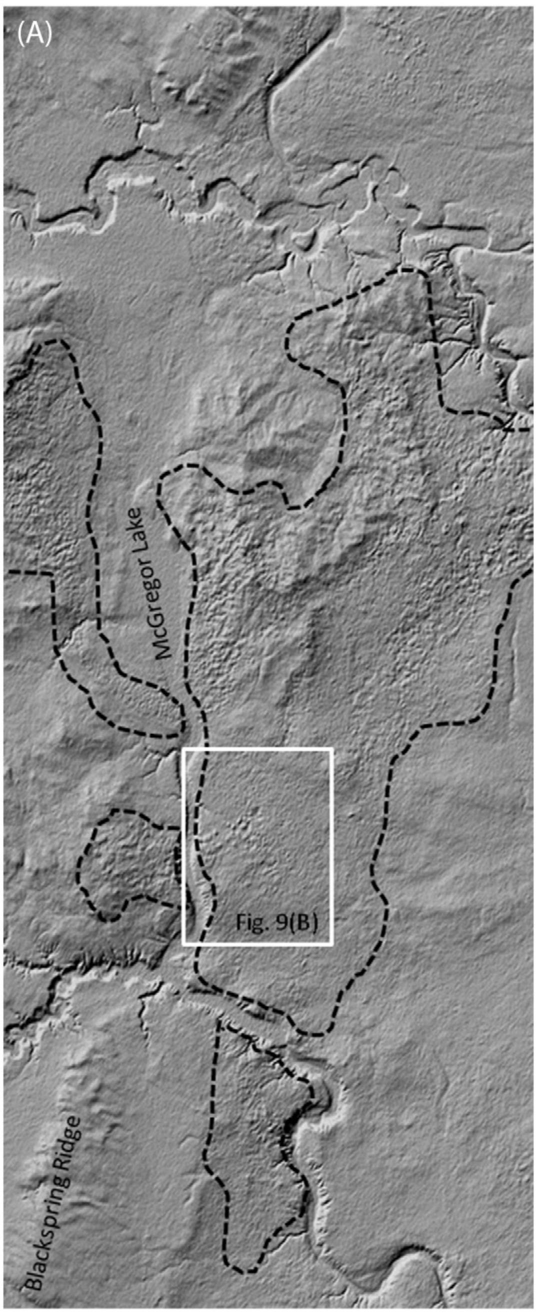


Fig. 9. Example of hummocky terrain in the McGregor moraine: (A) Landsat ETM+ image of the moraine assemblage (areas outlined by broken line), with the Little Bow and Bow Rivers at the bottom and top of the image, respectively; (B) larger scale aerial photograph image of the hummocky terrain to the southeast of McGregor Lake, located by the box in (A).

Lineation length gradually increases from northwest to southeast, trending into several MSGLs within CAfs_2 (Fig. 11). All measured elongation ratios within the CAIS marginal area are greater than the 10:1 minimum threshold proposed by Stokes and Clark (2002) for fast flowing ice.

The locations of CAfs_1 and 2 (Fig. 3), on the downice side of bedrock highs that appear to have been glacitectonically thrust and stacked (see Section 5.3 below) and at locations where the proglacial slope dips downice (Fig. 7A and D), suggest that topography may have been a controlling factor in their production. Similar lineation occurrences on the downice sides of higher topography are found within Hfs_5 on Blackspring Ridge (Fig. 2; Munro-Stasiuk and Shaw, 2002) and on the Athabasca fluting field in central Alberta (Shaw et al., 2000), an observation also made by Westgate (1968) who further highlights the occurrence of the largest flutings in such settings. If this is a significant factor in lineation and MSGL production, it would explain why so few lineations occur along the CAIS where the regional slope predominantly dips upice (Fig. 7A). This evidence is consistent with the groove ploughing theory for lineation production (Clark et al., 2003) whereby ice keels produced by flow over bedrock bumps carve grooves in the bed and deform sediments into intervening ridges or flutings. The surface form of the northern end of the megafluting complex at the centre of CAfs_1 is instructive in this respect in that it appears as a flat-topped ridge with grooves in its summit (Evans, 1996, 2000).

5.2. Transverse ridges

A variety of large transverse ridges were initially identified on DEMs by Evans et al. (2008) who interpreted them as either overridden or readvance moraines based upon their morphology and some localized exposures, the latter indicating a glacitectonized bedrock origin. The higher resolution SRTM data used in this study facilitate a more detailed assessment of these forms.

The streamlining and lineation overprinting of the two major arcuate ridges within the TR_1 sequence (TR_1(N) and TR_1(S); Figs. 2 and 4) document the southerly advance of the HPIS over the site after major ridge construction. The arcuate or lobate nature of the ridges indicates that they were constructed as ice marginal features and so likely record an earlier advance of the HPIS to this location. The two major ridges (TR_1(N) and TR_1(S)) occur at a location where the bedrock topography rises 30–60 m above the surrounding terrain (Geiger, 1967; Atkinson and Lyster, 2010b) and are significantly different in morphology to the minor ridges (TR_1a) that lie over, between, and south of them (Figs. 2 and 4). Their size, multiple crests, and location on a bedrock rise are compatible with glacitectonic origins, similar to numerous other examples in southern Alberta, where the Cretaceous bedrock is highly susceptible to glacitectonic disruption (Bluemle and Clayton, 1984; Aber et al., 1989; Tsui et al., 1989; Aber and Ber, 2007).

In the northeast, ridge sets TR_3 and 4 (Fig. 2) are part of the northern extension of the Suffield moraine. Ice thrusting was proposed by Kjaersgaard (1976), Shetsen (1987), and Evans et al. (2008) for ridge set TR_4. Previous mapping in the area of TR_3 by Kjaersgaard (1976) and Shetsen (1987) identified significantly fewer transverse ridges but did propose an ice thrust origin for parts of those landforms. Glacitectonic origins are also most likely for TR_5 and 6 (Fig. 2) because they occur on bedrock highs (Fig. 7A) and hence are influenced by topographical controls (Bluemle and Clayton, 1984; Aber et al., 1989; Tsui et al., 1989); comprise closely spaced, parallel, and predominantly linear multiple ridge crests; and internally contain glacitectonized bedrock (Evans and Campbell, 1992; Evans, 1996; Evans et al., 2008). The overall arcuate planforms of TR_5, TR_6, and TR_7 also support an ice-marginal formation by compressive ice marginal flow (cf. Evans, 1996, 2000; Evans et al., 2008). A thin till cover situated on top of the ridges suggests that they are actually cupola hills (Aber et al., 1989; Evans, 2000; Benn and Evans, 2010) produced by the overridding CAIS margin (Evans, 2000). Ridge set TR_7(W) is a locally fluvially modified part of the TR_6 and TR_7 sequence and so it is most likely that they share similar origins.

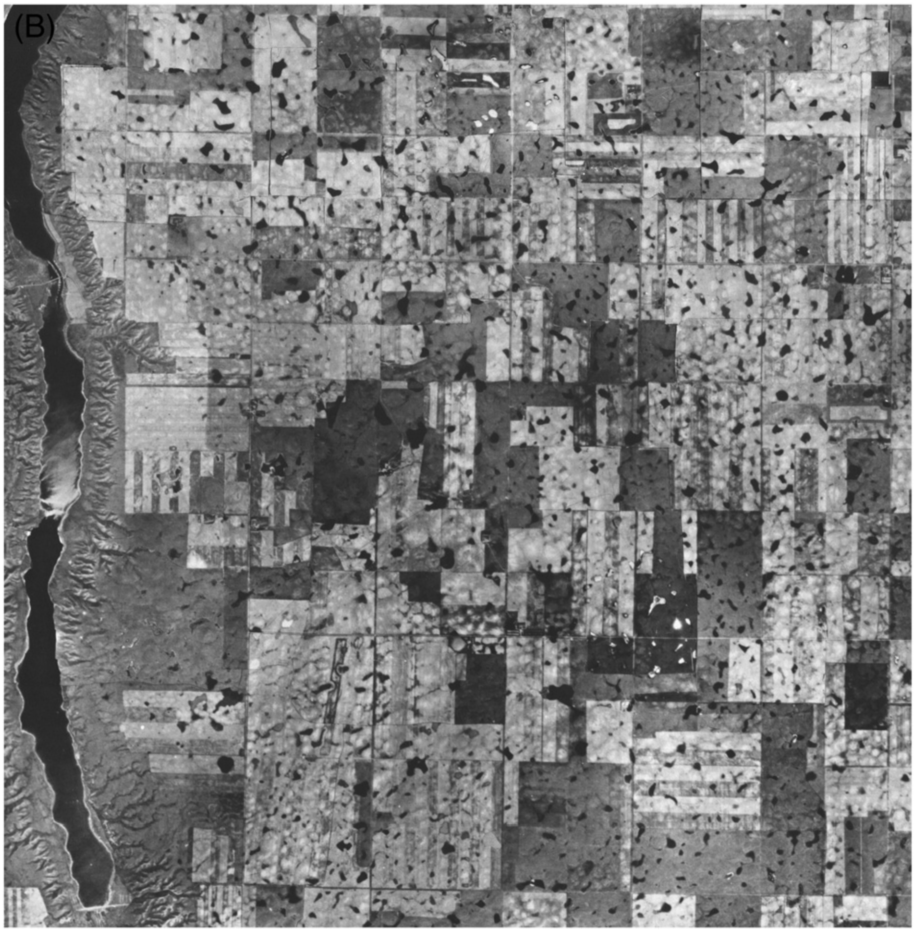


Fig. 9 (*continued*).

The large bi-lobate ridge (TR_8) has previously not been identified and is hereafter named the Vauxhall Ridge after the nearest town. It is almost certainly ice marginal, based on its dual-lobate planform and lies downice of CAfs_1 and the subglacially streamlined Lake Newell esker network (Fig. 5; Evans, 1996), which suggests that it records the readvance limit of the CAIS. The ridge also continues into hummocky terrain and transverse ridges to the east, which are therefore interpreted to have formed contemporaneously. The geomorphic expression of the Vauxhall Ridge provides few indicators as to its precise genetic origins, and so further investigation of subsurface structure is required. Nonetheless, its steeper distal slope is typical of the cross profiles of thrust block moraines or composite ridges (Aber et al., 1989).

Ridge sets TR_11 and 12 (Fig. 2) are interpreted as a single sequence of ridges formed at the margin of the east lobe or Ice Stream 1 of Ó Cofaigh et al. (2010). Extensive sections through the ridges show that they have been glacitectonically thrust and stacked (Ó Cofaigh et al., 2010), indicating an ice thrust and/or submarginal incremental thickening origin (cf. Evans et al., 2012).

Glacitectonic origins are proposed for some of the minor transverse ridges mapped at larger scales in the CAIS margin case study. Specifically, MTR type 1 of the CAIS marginal landsystem (TR_10; Fig. 2) likely originated through glacitectonic thrusting and have been overrun by a readvancing ice margin, as indicated by their overprinting with perpendicular lineations. The ridges are composed of deformed bedrock (Beaney and Shaw, 2000), an observation used to support a proglacial thrusting origin by Westgate (1968), Shetsen (1987), and Evans et al. (2008). Their location along the preglacial drainage divide suggests that topography was significant in their formation; glacier flow would

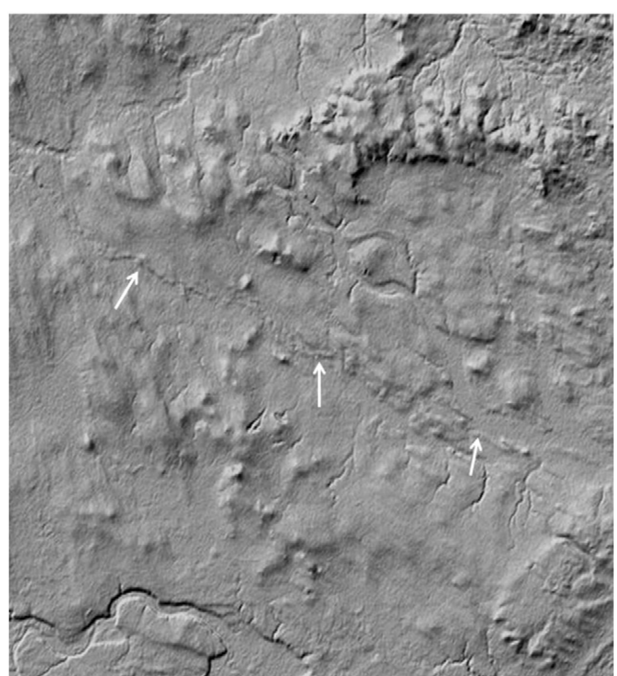


Fig. 10. SRTM data showing flowset Hfs_4 and demonstrating the high level of spatial coherency as well as a large esker indicated by white arrows.

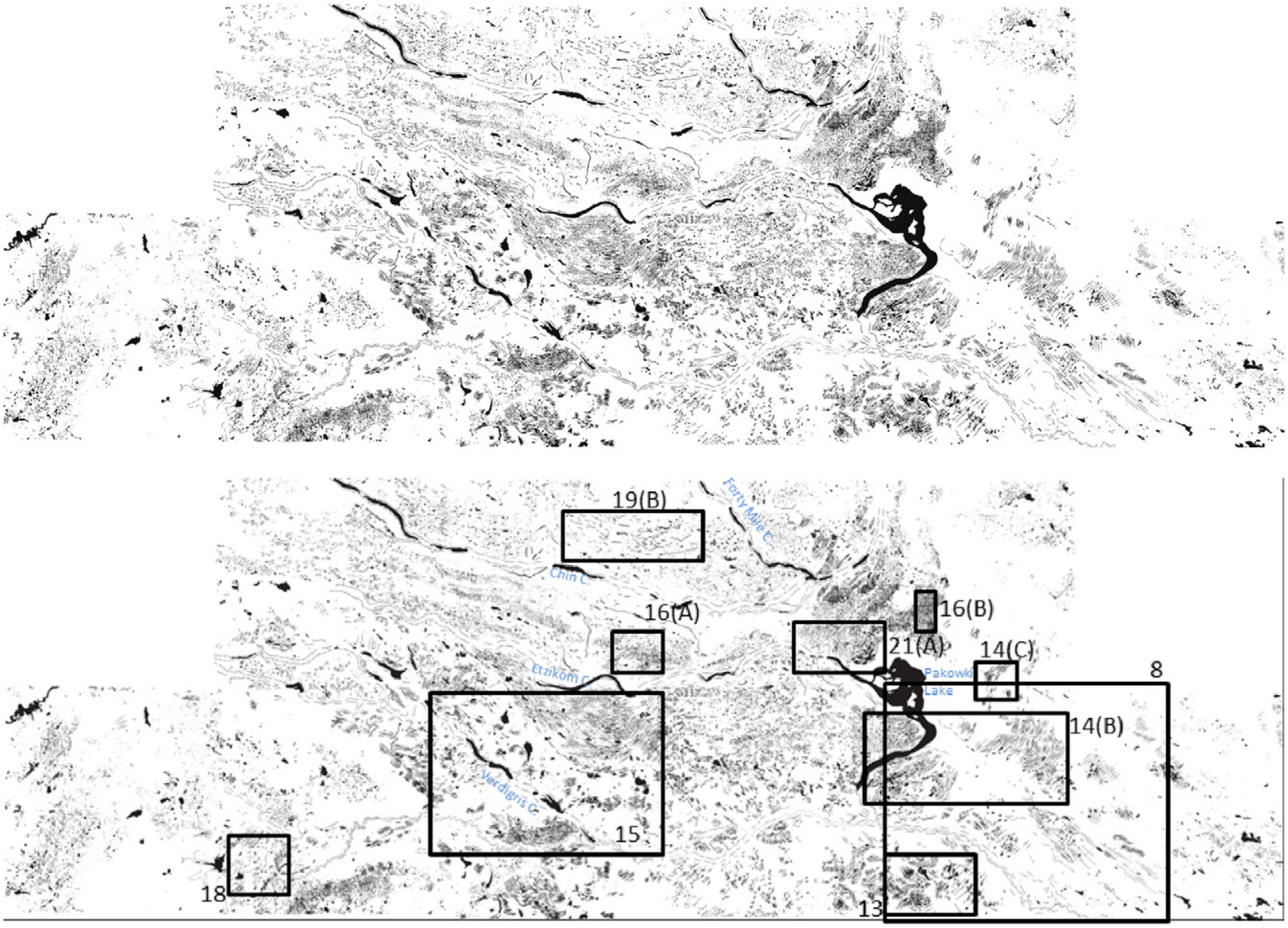


Fig. 11. Glacial geomorphology map of the landforms produced at the margin of the CAIS (lower map is low resolution version with locational information and upper map is high resolution version for viewing landform details). Black-shaded areas represent lakes and ponds and therefore demarcate the extent of meltwater channels/spillways and smaller scale depressions between hummocks and ridges. MTR type 2 crests are depicted as black arcuate lines, and the MTR type 1 crests of TR_10 are depicted as barbed lines. Flutings are represented by straight lines orientated oblique to transverse ridges. Large blank areas lying within bands of hummocks are predominantly occupied by flat-topped mounds, which are interpreted as ice-walled lake plains. Hatched broken lines depict the margins of major channels. The typical morphological details of the hummocky terrain (represented here by densely spaced small scale depressions) are illustrated and classified in Figs. 16 and 17, respectively.

have been compressive (Fig. 7D), and porewater pressures in the weak Cretaceous bedrock would have been elevated, a situation highly conducive to glacitectonism (Bluemle and Clayton, 1984; Aber et al., 1989; Tsui et al., 1989).

The MTR type 2 sequences (Fig. 12), primarily located in the Milk River drainage basin (Figs. 7D, E and 11), display an inset (en echelon) pattern that closely resembles that of push moraines presently developing at active temperate glaciers, for example at Breiðamerkurjökull and Fjallsjökull in Iceland (Price, 1970; Sharp, 1984; Boulton, 1986; Matthews et al., 1995; Krüger, 1996; Evans and Twigg, 2002; Evans, 2003; Evans and Hiemstra, 2005). These modern analogues have been used by Evans et al. (1999, 2008) and Evans (2003) to support the interpretation of the whole sequence of transverse ridges within the CAIS marginal area as recessional

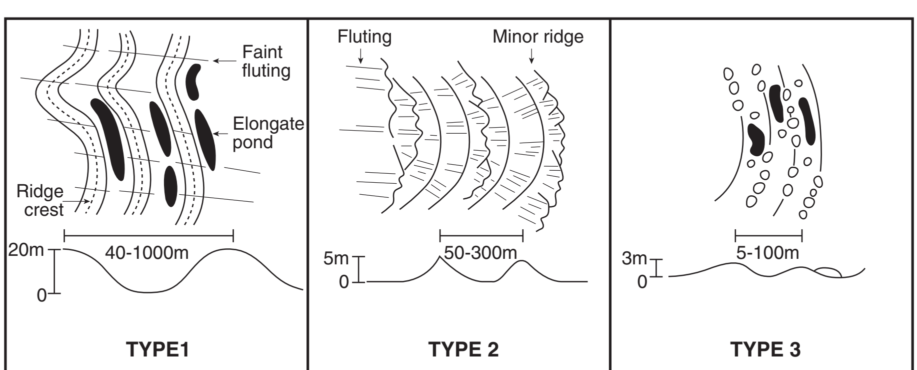


Fig. 12. Morphological characteristics of transverse ridge sets within the CAIS marginal zone. Type 1 ridges are symmetrical in form and have smoothed summits separated by partially water-filled depressions. Type 2 ridges have sharper crests and vary in wavelength. They also may display crenulate and partially overprinted planforms and are associated with short flutings developed on their ice-proximal slopes. Type 3 ridges are composed of numerous, strongly orientated hummocks and ridges separated by partially water-filled depressions with occasional hummocks. Ice flow in all cases is from the left.

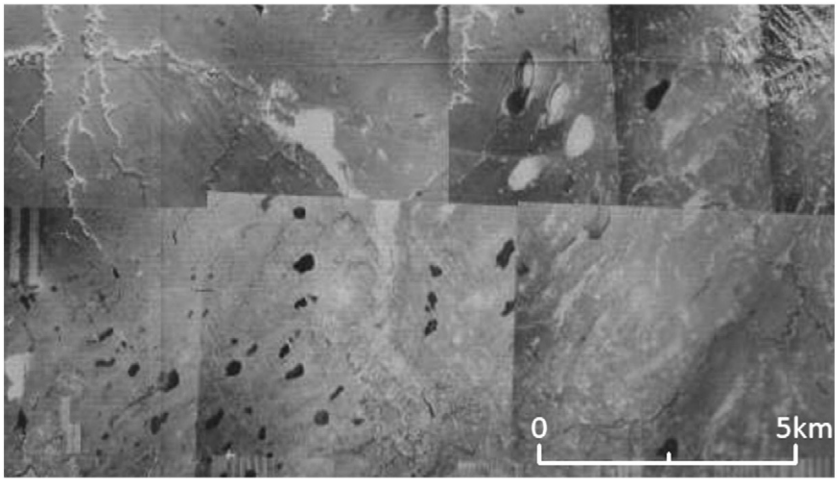


Fig. 13. Aerial photograph mosaic of transverse ridge sets MTR type 1 located in the SE corner of the CAIS marginal zone and overprinted by faint lineations not visible on the SRTM imagery. These ridges form the corrugated appearance of TR_10 (see Fig. 8 for location).

push moraines, a more specific genetic assessment than the previous conclusions of Westgate (1968) that the landforms represented 'washboard moraine', 'linear disintegration ridges' and 'ridged end moraine'. A recessional push moraine origin implies that the CAIS margin must have been warm based during landform construction, reflecting seasonal climate variability (Boulton, 1986; Evans and Twigg, 2002; Evans, 2003). We recognize that the concept of a seasonal climatic control on ice stream activity is novel and not uncontroversial, but the association of palaeo-ice stream landsystem signatures with those of active temperate glacier margins needs to be reconciled. Moreover, fast glacier flow and active temperate snout activity are not entirely incompatible, and hence we should not be surprised to find evidence of their juxtaposition in the landform record. Nonetheless the regional distribution of landform signatures, particularly the zonation of minor transverse ridges as depicted by the patterns in Fig. 2, potentially reveals varying spatial and temporal operational activity of ice streaming (see Section 6).

The origins of MTR type 3 (Fig. 12) are indicated by the style of hummock (see Section 5.3 below) visible within the linear assemblages that make up the component ridges. The individual hummocks that predominate within MTR type 3 vary between HT type 1 and HT type 2 hummocks, which are interpreted below as having formed supraglacially. This implies that significant englacial debris concentrations characterized the margin of the CAIS at the time of MTR type 3 formation. Debris provision could have been related to either englacial thrusting and stacking of debris-rich ice created by compressive flow against the reverse regional slope (Fig. 7A; Boulton, 1967, 1970; Ham and Attig, 1996; Hambrey et al., 1997, 1999; Glasser and Hambrey, 2003) or incremental stagnation (Eyles, 1979, 1983; Ham and Attig, 1996; Patterson, 1997; Jennings, 2006; Clayton et al., 2008; Bennett and Evans, 2012). In the case of incremental stagnation, the moraine linearity would be related to either the high preservation potential of controlled moraine (Gravenor and Kupsch, 1959; Johnson and

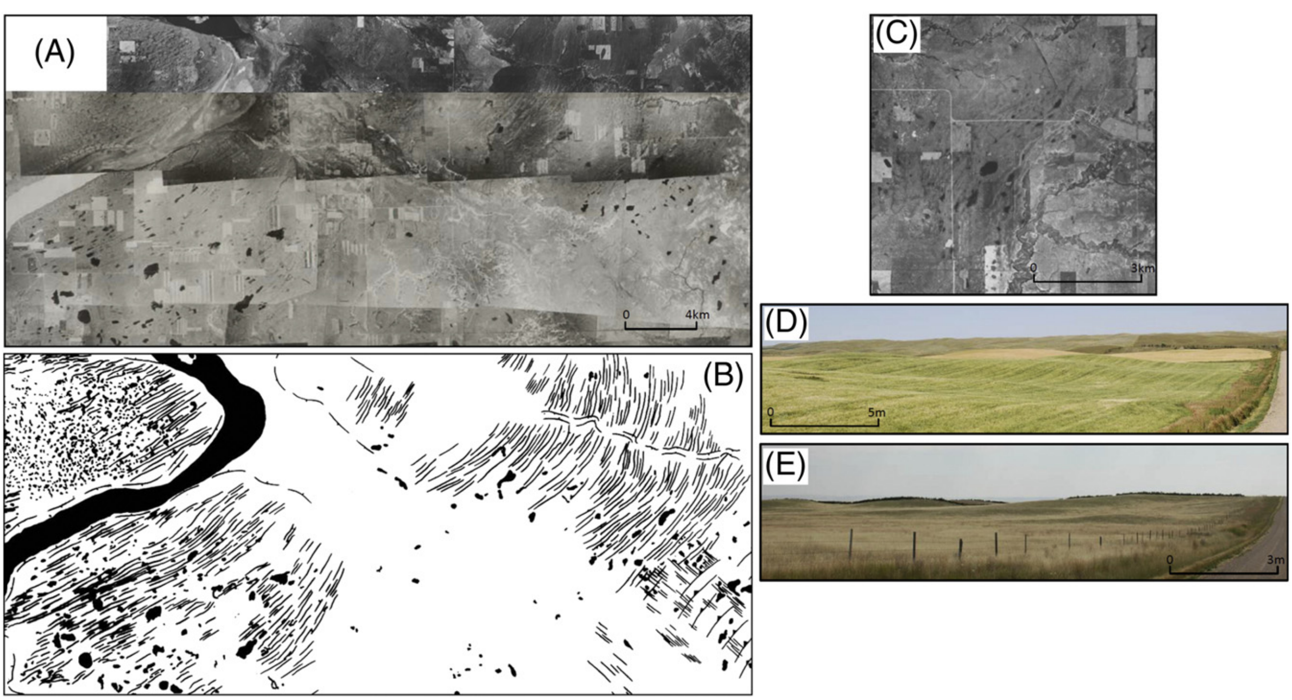


Fig. 14. MTR types 2 and 3: (A) aerial photograph mosaic and (B) geomorphology map extract of MTR type 2, located to the south of Pakowki Lake (see Fig. 11). The northwest corner of the image and map shows type 3 ridges blending into type 1 hummocky terrain; (C) type 2 ridges located 5 km to the north of the images in (A) and (B) (centre of image is 49°23.5′ N. and 110°44′ W.); (D) and (E) ground views showing the parallel, smooth-crested, and discontinuous nature of type 3 transverse ridges.

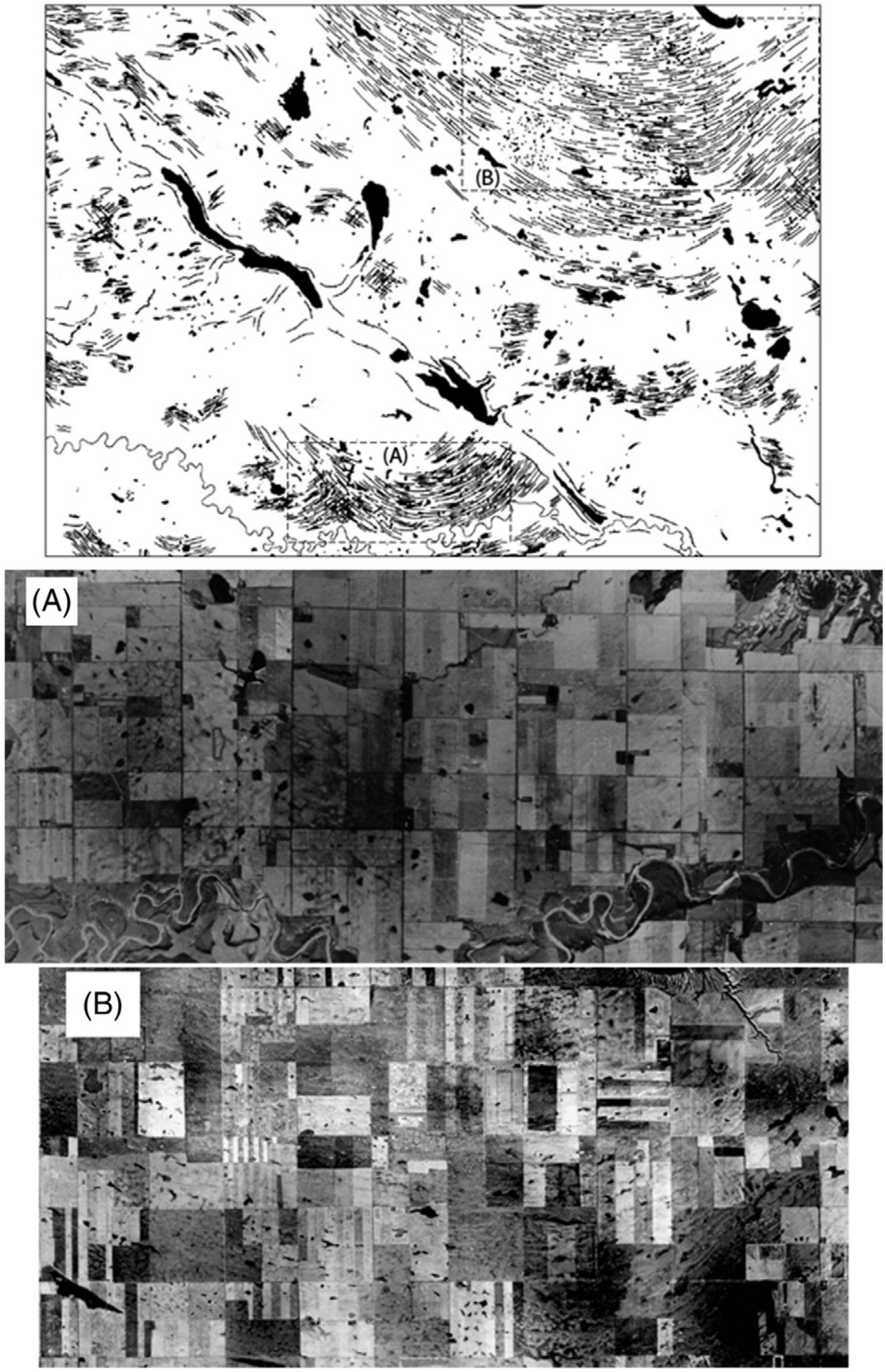


Fig. 15. Geomorphology map extract of MTR type 3 located in the central portion of the CAIS marginal zone (see Fig. 11). Individual hummocks and ridge segments are arranged contiguous with each other, giving rise to linearity in the landform record: (A) area located between Verdigris Coulee and the Milk River; (B) area located south of Crow Indian Lake in Etzikom Coulee.

Clayton, 2003), an unlikely scenario based upon modern analogues of controlled moraine development (Evans, 2009; Roberts et al., 2009), or active recession of a debris-charged ice margin brought about by warm polythermal conditions and accentuated by upslope advances (Evans, 2009). This is supported by the fact that, although MTR type 3 sequences are composed of contiguous linear hummock tracks and discontinuous ridges (Figs. 11, 12, 14, and 15), small-scale mapping (Figs. 2 and 11) shows clear inset sequences of MTR types 2 and 3, typical of active recession of both the CAIS and HPIS margins in southern Alberta (Fig. 4 illustrates that the minor ridges TR_1a are MTR types 2 and 3; cf. Evans, 2003) based upon modern analogues of active temperate and warm polythermal glaciers (Boulton, 1986; Evans and Twigg, 2002; Colgan et al., 2003; Evans, 2003; Evans and Hiemstra, 2005; Evans, 2009).

5.3. Hummocky terrain

The HT type 1 hummocks represent the largest proportion of hummocky terrain within the CAIS marginal area. Concentrations of HT type 1 hummocks occur around the Del Bonita highlands and in the lobate bands of hummocks north of Etzikom Coulée (Fig. 11), 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 HT type 1 hummocks are composed of till. A supraglacial origin for HT type 1 hummocks can be supported by simple form analogy (cf. Boulton, 1967; Clayton, 1967; Parizek, 1969; Boulton, 1972; Clayton and Moran, 1974; Eyles, 1979, 1983; Paul, 1983; Clayton et al., 1985; Johnson et al., 1995; Ham and Attig, 1996; Patterson, 1997, 1998; Mollard, 2000; Johnson and Clayton, 2003; Jennings, 2006), but their juxtaposition with active recessional moraines in lobate arcs of landform assemblages (Figs. 11 and 16) suggests that they were not associated with widespread ice stagnation. Differential melting and supraglacial debris reworking by continuous topographic reversal can be invoked to explain the irregular shapes and sizes of the hummocks when viewed at larger scales, although subglacial pressing of the soft substrate at the margin of the CAIS as proposed by Stalker (1960), Eyles et al. (1999), and Boone and Eyles (2001) could have been operating in the poorly drained conditions of the reversed proglacial slopes of the region (Klassen, 1989; Mollard, 2000). Nevertheless, the lobate arcuate appearance of HT type 1 hummocks when viewed at smaller scales has a strong resemblance to the controlled moraine reported by Evans (2009) and the hummock assemblages along the southern Laurentide Ice Sheet margins described by Colgan et al. (2003) and Johnson and Clayton (2003) as their 'Landsystem B'. The corollary is that, during early deglaciation, the edge of the CAIS was at times cold based and part of a polythermal ice sheet margin, beyond which there was a permafrost environment (Clayton et al., 2001; Bauder et al., 2005); several generations of ice wedge casts and associated ground ice forms around the Del Bonita highlands (Karlstrom, 1990) and the Cypress Hills (Westgate, 1968) verify ground ice development in the region, although the features are undated.

North of the CAIS marginal zone, HT type 1 hummocks are extensive and well developed, and therefore have been the subject of numerous investigations (e.g., Stalker, 1960; Munro-Stasiuk and Shaw, 1997; Eyles et al., 1999; Boone and Eyles, 2001; Evans et al., 2006). Comparison of Fig. 2 and existing maps (cf. Shetsen, 1984, 1987; Clark et al., 1996; Evans et al., 1999) shows that hummocky terrain mapping using SRTM data is capable of a high degree of precision. Because of their position between corridors of fast flowing ice lobes, the hummocks have been used to demarcate an *interlobate* terrain by Evans et al. (2008), but the more generic term *hummocky terrain* is preferred here. Nonetheless, the abrupt transition from smoothed topography (corridor) to hummocky terrain along the CAIS margin is interpreted as a change in subglacial regime and hence demarcates the flow path of the ice stream (cf. Dyke and Morris, 1988; Patterson, 1998; Evans et al., 2008; Ó Cofaigh et al., 2010). Glacitectonic evidence identified along the north shore of Travers Reservoir demonstrates that some linear hummocks and low amplitude ridges in hummocky terrain are in fact thrust block moraines (Evans et al., 2006) formed by ice flow from the north east, indicative of CAIS advance into the area after the HPIS had receded. The input from the HPIS is demarcated by flowsets Hfs_4 and 5 (Fig. 3) that flow into the McGregor moraine. Detailed investigation of this area by Evans et al. (2006) reveals that the hummocky terrain, when viewed at large scale, comprises inset recessional push ridges and associated arcuate zones of flutings similar to modern active temperate glacial landsystems (Evans et al., 1999; Evans and Twigg, 2002; Evans, 2003; Evans et al., 2006, 2008). The hummocky terrain therefore represents a less linear set of ice-marginal landforms to those with which it is laterally continuous in the HPIS trunk immediately to the west (Fig. 2). The reconstructed ice margins show that ice was flowing into the area from the northwest (Evans et al., 2006) and so most likely represent the termination of flowset Hfs_5.

The HT type 2 hummocks resemble the 'doughnut hummocks' or 'ring forms' that are common to many deglaciated ice sheet forelands

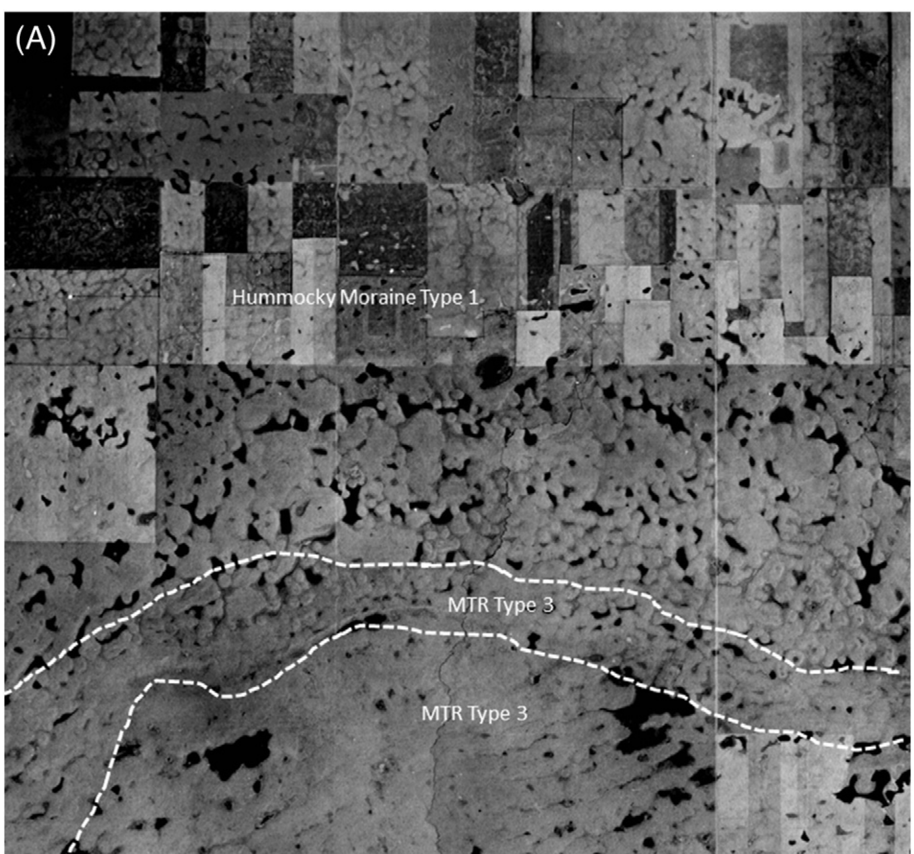


Fig. 16. Examples of HT type 1 and 2 hummocks (see Fig. 11 for locations): (A) predominantly HT type 1 hummocks north of Crow Indian Lake in Etzicom Coulee (centre of image is 49°26′ N and 111°39′ W). Note also the abrupt boundaries at the base of the image where the hummocky moraine changes to a narrow band of MTR type 3 which is in turn overprinted on another area of MTR type 3 with a more linear pattern; (B) predominantly HT type 2 hummocks north of Pakowki Lake (centre of image is 49°28′ N. and 110°54.5′ W). (see also Fig. 21B).

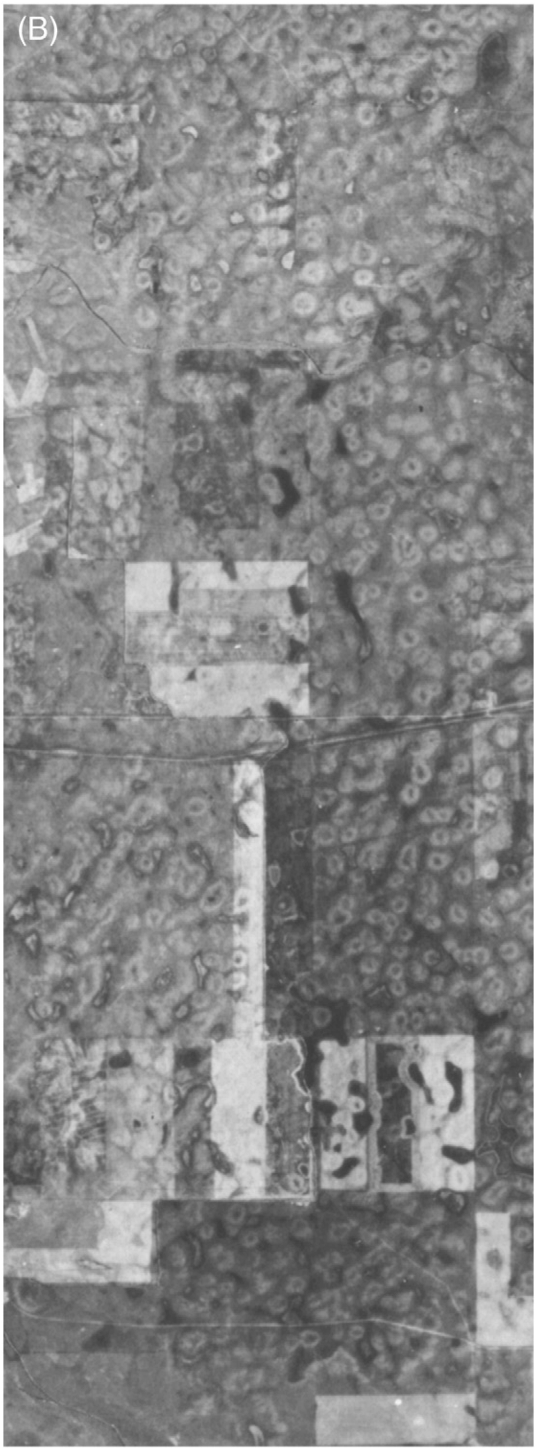


Fig. 16 (*continued*).

in mid-latitude North America and Europe (e.g., Gravenor and Kupsch, 1959; Bik, 1969; Parizek, 1969; Aartolahti, 1974; Lagerbäck, 1988; Boulton and Caban, 1995; Mollard, 2000; Colgan et al., 2003; Knudsen et al., 2006). Johnson and Clayton (2003) demonstrated that doughnut hummocks across North America are predominantly composed of clayey till, which they suggest is important to hummock formation. Several genetic models have been proposed, all of which regard the landforms as indicative of a stagnant glacial regime (Knudsen et al., 2006), but they remain poorly understood. Importantly, like HT type 1 hummocks, the fact that HT type 2 hummocks are often contiguous with push ridges appears to contradict the stagnation model. Because HT type 2 hummocks are contiguous with not only recessional push moraines but also HT type 1 and HT type 3 hummocks (see below), which are argued to be supraglacial in origin; therefore doughnut hummocks most likely also originated as supraglacial debris concentrations (controlled moraine) in a polythermal ice sheet margin. Alternative origins for HT type 2 hummocks include proglacial blowout features created by over-pressurized groundwater (Bluemle, 1993; Boulton and Caban, 1995; Evans et al., 1999; Evans, 2003, 2009) and subglacial pressing of saturated sediments (Gravenor and Kupsch, 1959; Stalker, 1960; Aartolahti, 1974; Eyles et al., 1999; Mollard, 2000; Boone and Eyles, 2001), although the latter would not produce linear chains of hummocks lying between arcuate push moraine ridges.

The conspicuous ridges that occur in association with HT type 2 hummocks (Fig. 18), and which are often continuous with hummock rims, indicate that rims and ridges are a product of the same formational processes. This could involve either (i) the elongation of hollows between controlled moraines during melt-out, giving rise to preferential deposition in linear chains of ice-walled channels or supraglacial trough fills (e.g., Thomas et al., 1985); and/or (ii) occasional ice-marginal pushing during the overall downwasting of a debris-charged snout upon which controlled moraine was developing (cf. Evans, 2009; Bennett et al., 2010; Bennett and Evans, 2012).

The HT type 3 hummocks closely resemble the ice-walled lake plains of the southern Laurentide Ice Sheet lobes in Minnesota, North Dakota, Wisconsin, Michigan, and 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 sedimentology and molluscs present within the enclosed deposits. Their presence therefore is unequivocally associated with supraglacial origins, the corollary of which is that any adjacent hummocky terrain is also of supraglacial origin (Johnson and Clayton, 2003; Clayton et al., 2008). The large sizes of the HT type 3 hummocks can be explained by their prolonged development after ice recession as a result of thick insulating debris cover (Attig, 1993; Clayton et al., 2001; Attig et al., 2003; Clayton et al., 2008), hence also their absence from 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, whereby the largest ice-walled lake plains are located around the Del Bonita highlands where permafrost features have also been recorded (Karlstrom, 1990); hence the production of debris-rich glacier ice by basal freeze-on and its development into a widespread supraglacial debris cover was possible. Verification of this proposed influence of permafrost conditions on ice sheet marginal debris entrainment processes requires the establishment of dating controls on the permafrost features of the region.

6. Discussion

6.1. Overview and chronology

The regional glacial geomorphology of southern Alberta primarily records the deglacial dynamics of the southwest margin of the Laurentide Ice Sheet, within which three major ice streams (HPIS, CAIS of Evans et al. (2008), and Ice Stream 1 or east lobe of Ó Cofaigh et al. (2010) and Shetsen (1984), respectively) coalesced and flowed against the north-easterly dipping topography, thereby damming proglacial lakes and diverting regional drainage during advance and retreat (Shetsen, 1984; Evans, 2000; Evans et al., 2008). In combination with the available deglacial chronology for the region (cf. Westgate, 1968; Clayton and Moran, 1982; Dyke and Prest, 1987; Kulig, 1996), the ice-marginal landforms are now used to chart the broad pattern of ice sheet retreat (Fig. 20).

Although the existing chronology is not well constrained by absolute dates, we acknowledge Westgate's (1968) five distinct

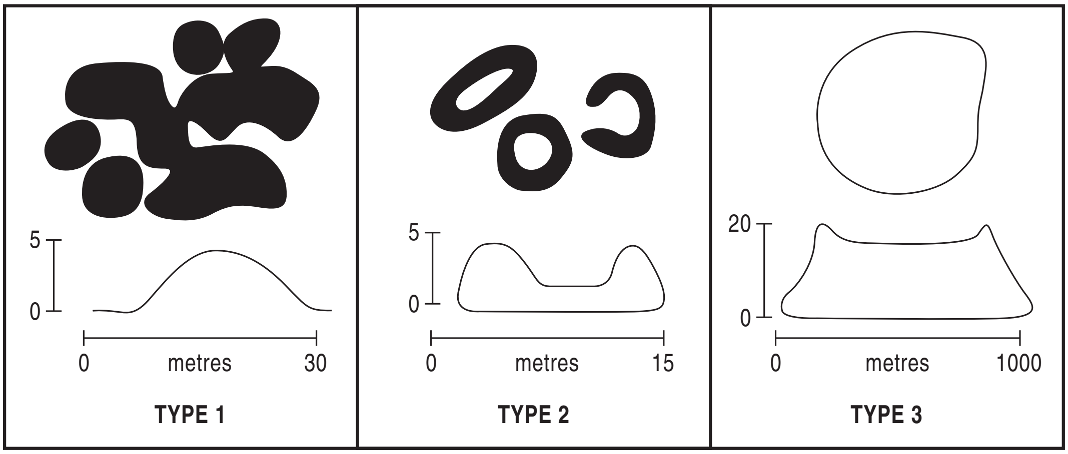


Fig. 17. Morphological characteristics of hummocks within the Lethbridge moraine sequence. The dimensions reflect the largest features in each class.

morphostratigraphic units (Elkwater drift; Wild Horse drift; Pakowki drift; Etzikom drift; Oldman drift), each of which has been taken to represent a readvance limit in southeast Alberta based on petrography and morphology. 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 15–20 m 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. 20) is marked by the outer extent of the push moraines to the southeast of Lake Pakowki and runs along the northern tip of the Milk River and north around the Cypress Hills (Westgate, 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 (15 ka BP; Kulig, 1996). The Pakowki advance (Fig. 20), not recognized in Christiansen's (1979) or Dyke and Prest's (1987) deglacial sequences, most likely occurred between 14 and 13.5 ka BP (Kulig, 1996) and relates to Clayton and Moran's (1982) stages F–H. The Etzikom drift limit is interpreted as the Lethbridge moraine limit of Stalker (1977) and is marked in Figs. 11 and 20 by the broad band of hummocky terrain just north of Etzikom Coulee. This ice margin maintained its position along the Lethbridge moraine until around 12.3 ka BP (Stage I, Clayton and Moran, 1982; Dyke and Prest, 1987; Kulig, 1996). The Oldman drift limit (Fig. 20) is located just south of the Oldman River. Importantly, the correlation between the thrust ridges at Travers Reservoir (Evans et al., 2006) and the Oldman limit suggests that they were formed during this readvance episode. The corollary is that the HPIS had already receded further to the north. This readvance (stages J–L; Clayton and Moran, 1982) most likely occurred just after 12 ka BP. Based on the regional geomorphology map (Fig. 2), we suggest that a further readvance occurred (Vauxhall advance), the limit of which is marked by the Vauxhall Ridge and must have occurred sometime after 12 ka BP. Evans (2000) suggested that the CAIS margin had receded to the north of the study area by 12 ka BP. Based on the Vauxhall advance evidence, the CAIS must have receded later than that proposed by Evans (2000). Importantly, Dyke and Prest (1987) placed the ice sheet margin to the 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 readvance of the CAIS after which time it receded rapidly (Evans, 2000). The exact timing of the HPIS and east lobe retreat are unclear, but the HPIS had likely receded somewhere north of Bow River by 12 ka BP. The various stages depicted in Fig. 20 are now related to the pattern of landform assemblages identified by this research and then used to infer phases of ice streaming (switch on) versus the steady state or normal-fast flow that produces active temperate glacial landsystems or the polythermal to cold based snout conditions that produce controlled moraine.

6.2. Landsystem model for the CAIS margin in the Milk River drainage basin

Various parts of the ice stream beds of western Canada have been interpreted previously as manifestations of specific landsystems based upon similarities with modern analogues; for example, Evans et al. (1999, 2008) identified an active temperate landform signature in the HPIS imprint and a surging signal in the Lac La Biche ice stream. Additionally, switches in basal thermal regime have been invoked by Evans (2009) to explain inset suites of different moraine types associated with the recession of the HPIS margin in the McGregor moraine belt. Thermal regime switches and intermittent surges during recession have been proposed elsewhere in reconstructions of southern Laurentide Ice Sheet palaeoglaciology. For example, Colgan et al. (2003) identified three characteristic landsystems that they interpret as the imprint of an ice lobe with changing recessional dynamics. The outermost landsystem of a drumlinized zone grading into moderate- to high-relief moraines and ice-walled lake plains represents a polythermal ice sheet margin with sliding and deforming bed processes giving way to a marginal frozen toe zone. Inside this landsystem lie fluted till plains and low relief push moraines, a landsystem indicative of active temperate ice recession. This in turn gives way to a landsystem indicative of surging activity. At a regional scale, Evans et al. (1999, 2008) and Evans (2009) have promoted similar temporal and spatial variability in ice stream landform imprints in Alberta, but the large-scale mapping reported here allows a finer resolution record of such changes to be elucidated for ice sheet margins during the early stages of deglaciation.

The juxtaposition of the transverse ridge moraine types of southern Alberta as presented in Figs. 8, 11, and 13 is illustrated specifically for MTR types 2 and 3 in Fig. 21A and then used in Fig. 21B to construct a conceptual landsystem model for terrestrial terminating ice stream margins. This model provides a diagrammatic representation of the continuum of landforms visible at the former margins of the CAIS between the Milk River drainage basin and the Oldman/Bow River drainage basins. The MTR type 2 are interpreted as active recessional push moraines that document temperate snout conditions and hence indicate that the lobate ice sheet margin was responding to seasonal climate drivers. Subglacially fluted tills and esker networks were active at these times. Hummocky moraine arcs containing ice-walled lake plains, kame mounds, and short esker segments represent cold-based lobe margins when controlled moraine was constructed by widespread freeze-on and stacking of basal debris-rich ice sequences. Between these two ends of the landform continuum lie moraine arcs composed of aligned hummocks and ponds (MTR type 3), indicative of polythermal margins that probably responded to intermediate timescale (decadal) climate drivers. The geomorphology of the CAIS in the areas located south of the Milk River and north of the Bow River is significantly different to that portrayed in Fig. 21, being dominated by major transverse

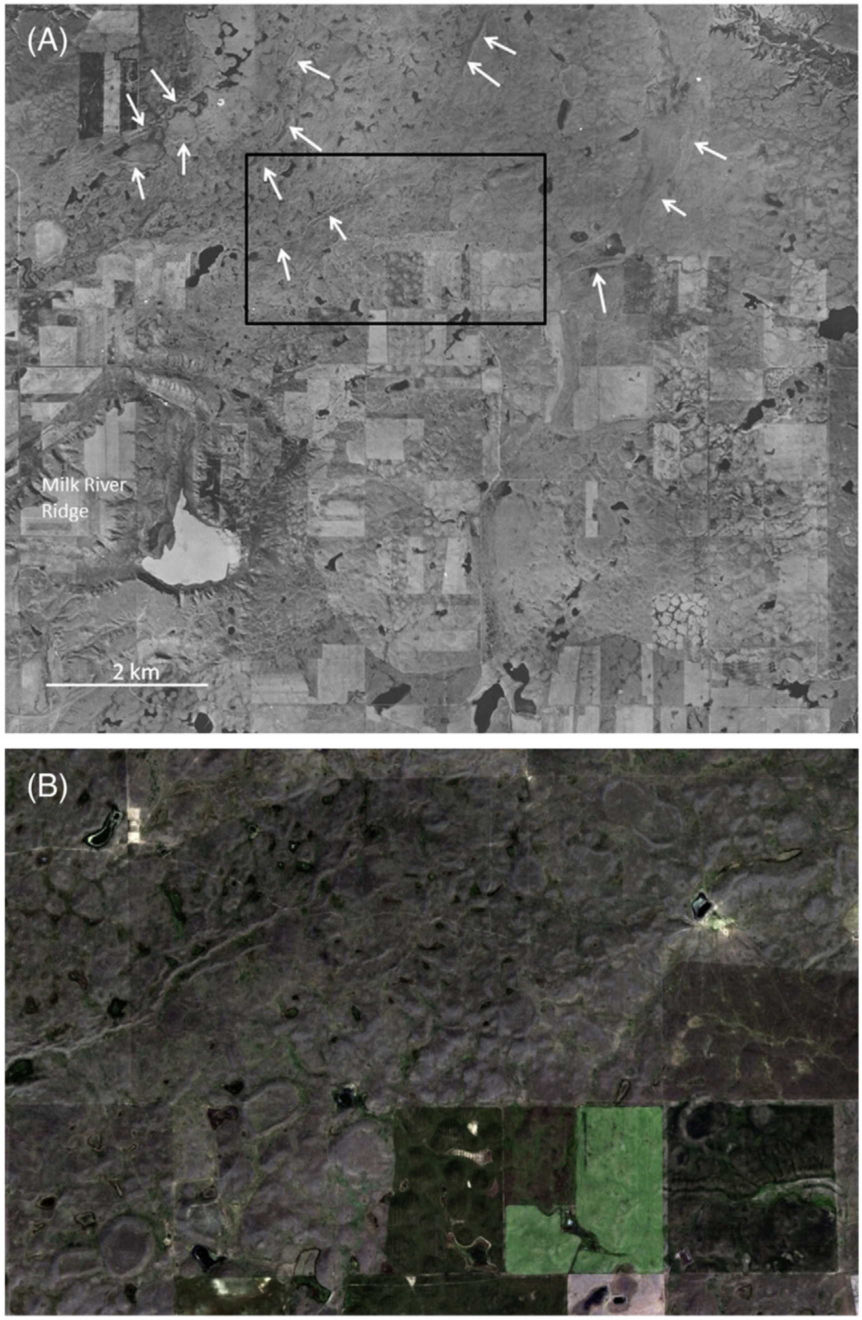


Fig. 18. (A). Examples of hummocky terrain on an aerial photograph mosaic of an area of the Milk River Ridge located near Del Bonita, showing the juxtaposition of all three hummock types. Also within the image are the ridges (highlighted by the white arrows) that run through some hummocky terrain bands. Note that here they run between HT type 2 hummocks and, in places, constitute parts of the hummock rims (centre of image is 49°04.5′ N. and 112°37′ W.). Area of landform details in (B) are identified by black box. (B) Google Earth image of area in box in (A) showing details of ice-walled lake plains and HT type 3 hummocky moraine in addition to prominent subsidiary ridges.

ridges (thrust moraines) and long flutings arranged in flowsets (Fig. 2). Similarly, the geomorphology of the HPIS contains landform assemblages like those portrayed in Fig. 21 only in the areas covered by TR_1a and TR_2 and within the central McGregor Moraine.

6.3. Dynamics of the Alberta terrestrial-terminating ice stream lobes

The Alberta ice streams flowed over a substrate composed of Cretaceous and Tertiary sediments, consisting of poorly consolidated clay, sand, and silt. The Cretaceous beds in particular are prone to glacitectonic folding and thrusting conditioned by a high bentonite content, which is reflected by the quantity and size of thrust features within southern Alberta. Additionally, the drainage conditions caused by swelling clays will have almost certainly created elevated porewater pressures and localized impermeable substrates, giving rise in turn to fast glacier flow (Clayton et al., 1985; Fisher et al., 1985; Klassen, 1989; Clark, 1994; Evans et al., 2008). Bedrock highs, many of which are residual Tertiary gravel-capped monadnocks, will likely have created resistance to ice flow (e.g., Alley, 1993; Joughin et al., 2001; Price et al., 2002; Stokes et al., 2007) and caused localized compression, highlighted by the presence of thrust ridges at such locations. Additionally, the reverse gradient of the easterly dipping bedrock surface will have initiated significant marginal compressive flow that also would have resulted in glacitectonic disturbance and well-developed controlled moraine on debris-charged snouts. The region is thereby an ancient exemplar of geologic setting exerting strong controls on the

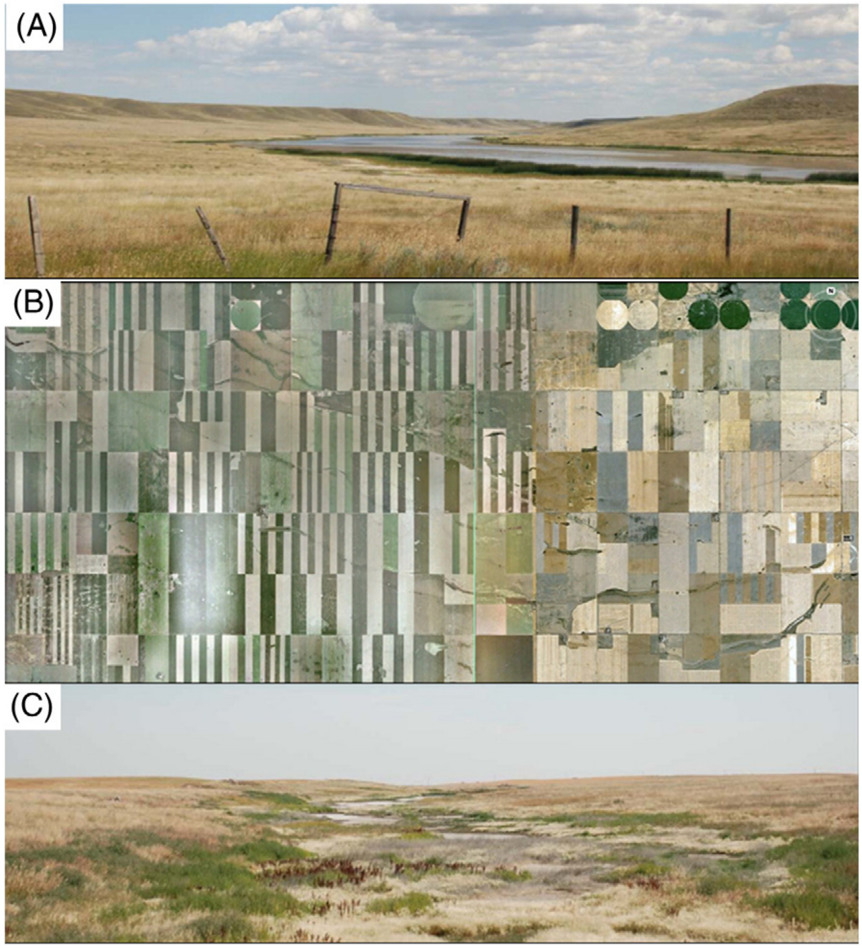


Fig. 19. Details of meltwater channels and spillways: (A) view eastward along Etzikom Coulée; (B) Google Earth image of the network of channels to the north of Chin Coulée (centre of image is 49°37.5′ N. and 111°38′ W); (C) ground view of shallow channels in the Google Earth image.

location and flow dynamics of ice streams (Anandakrishnan et al., 1998; Bell et al., 1998; Bamber et al., 2006), although we cannot ascertain whether fast ice motion occurred through deformation or sliding or a combination of the two. Numerous till units and upice thickening till wedges within southern Alberta (Westgate, 1968; Evans and Campbell, 1992; Evans et al., 2008, 2012) are consistent with the theory of subglacial deformation (Alley, 1991; Boulton, 1996a,b), although Evans et al. (2008) argued that the presence of large subglacial channels and thin tills overlying thin stratified sediments and shale bedrock along the CAIS trunk indicates that at some stage in this area deformation was subordinate to sliding.

A clear change in landform assemblages from south to north along the axis of the CAIS documents a temporal change in ice stream/lobe dynamics. Initial advance of the CAIS was responsible for the glacitectonic construction and overriding of large transverse ridges in bedrock (cupola hills). Although the dynamics of the CAIS during Laurentide Ice Sheet advance are difficult to reconstruct, the construction of large thrust moraines is most commonly associated with surging glacier snouts and therefore this mode of flow during advance cannot be ruled out.

During deglaciation, the dynamics of the CAIS switched from fast flow/streaming to steady state normal flow toward a lobate margin with a changing submarginal thermal regime. This is recorded by the arcuate bands of MTR types 2–3 and hummocky terrain located between the preglacial divide (Milk River Ridge) and the Bow River catchment. Specifically, the sequential south to north change from hummocky terrain to MTR type 2 to MTR type 3 in this area records a temporal switch in ice marginal characteristics, from cold polythermal to temperate and then to warm polythermal (cf. Colgan et al., 2003; Evans, 2009). A similar switch in submarginal thermal characteristics has been proposed for the HPIS by Benn and Evans (2006) and Evans (2009) to explain a south to north change in moraine characteristics. Based upon the chronology of ice sheet recession presented in Fig. 20, the switch to temperate conditions apparently occurred at the margins of both the CAIS and HPIS at similar times during deglaciation, indicating a potential regional climatic control.

The summary of landform spatial changes presented in Fig. 21B and the regional pattern of landforms depicted in Fig. 2 potentially record the switching on and off of palaeo-ice streaming as well as the changing thermal conditions and dynamics at the margins of more steady state

Table 2
Palaeo-ice stream criteria of the CAIS and HPIS compared to the schema proposed by Stokes and Clark (1999, 2001).

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. Glacitectonic and geotechnical evidence of pervasively deformed till	Yes	Yes
8. Submarine till delta or sediment fan (trough-mouth fan)	NA[a]	NA[a]

[a] Large arcuate assemblages of moraines and thick, complex sequences of tills and associated glacigenic sediments reported at the former HPIS and CAIS margins by Evans et al. (2008, 2012) are likely to be the terrestrial equivalents of trough-mouth fans.

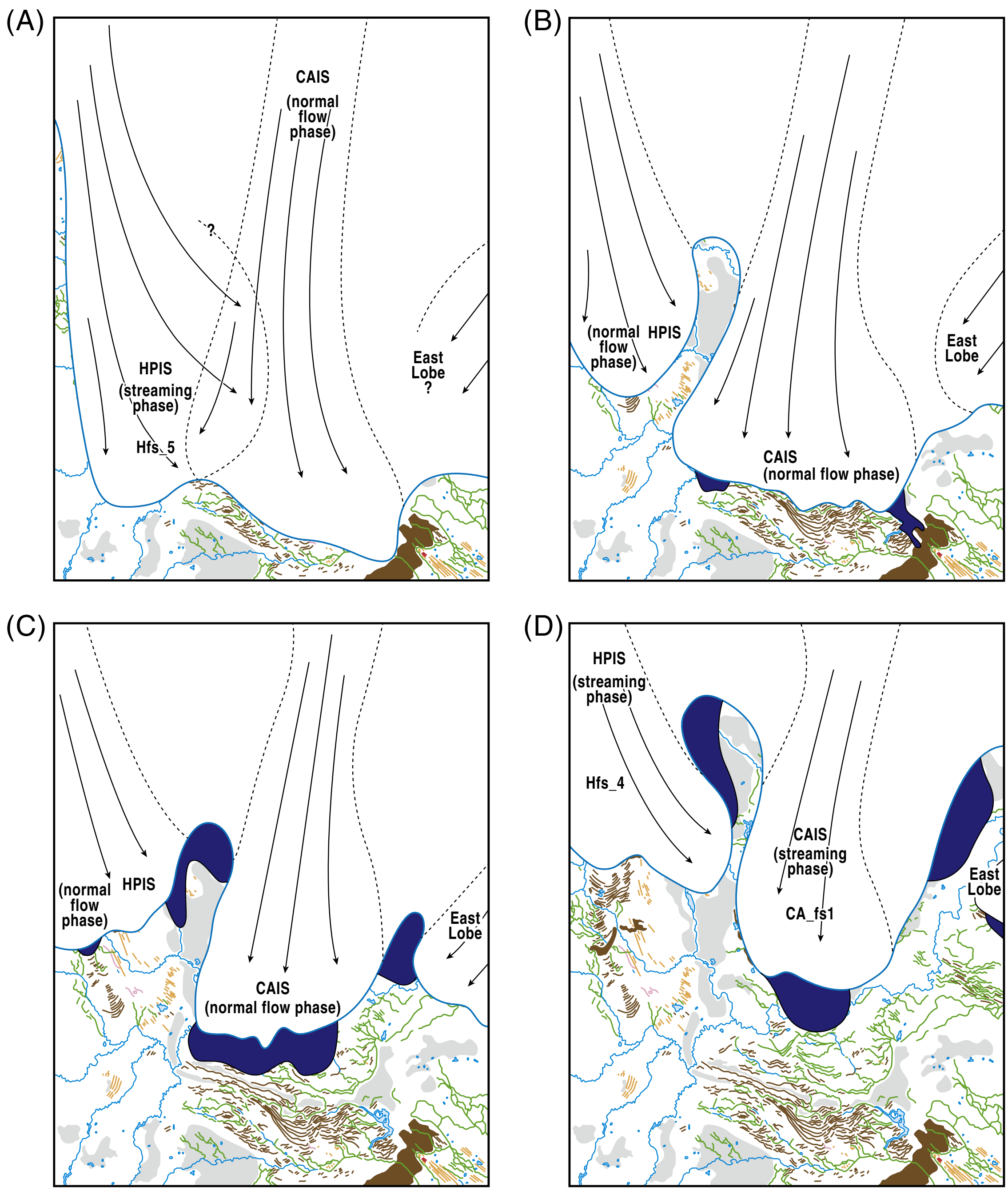


Fig. 20. Reconstructed generalized palaeoglaciology of the southern Alberta ice streams/lobes during deglaciation based on published chronologies (Westgate, 1968; Clayton and Moran, 1982; Dyke and Prest, 1987; Kulig, 1996) and constrained by geomorphology presented in this paper: (A) Pakowki advance limit around 14–13.5 ka BP. Note that this involves two substages, with a later substage involving the advance of the CAIS over an earlier HPIS imprint; (B) Etzikom limit located along the Lethbridge moraine at around 12.3 ka BP; (C) Oldman limit at approximately 12 ka BP; (D) Vauxhall limit tentatively dated at around 11.7 ka BP. The reconstructed position of the HPIS is based solely on geomorphology and so the chronology of the marginal positions is speculative. The proglacial lakes are minimal reconstructions based upon previous work by Westgate (1968), Shetsen (1987), and Evans (2000).

normal-fast flow lobes within the southwest Laurentide Ice Sheet. Ice streaming in the CAIS was most vigorous during the development of CAfs_1 and CAfs_2, but switched off during recession of the ice margin from the Milk River to the Bow River drainage basins when the snout underwent changes between active temperate, polythermal, and cold-based conditions in response to climate drivers. Ice streaming in the HPIS was switched on to produce each of the flowsets Hfs_5 to Hfs_1, but was undergoing active to polythermal marginal recession when receding northward through the area covered by the extensive minor moraine sequence TR_2 and TR_1a.

A contrasting landform assemblage north of the Bow River basin documents a further change in CAIS dynamics, wherein overridden thrust moraines, megaflutings (CAfs_1), and a fluted esker network lie inside the Vauxhall Ridge. This assemblage is interpreted as the imprint of a fast flow/streaming event, a precursor to the surges that constructed thrust moraines and crevasse-squeeze ridges to the north of the study area (Evans et al., 1999, 2008).

7. Conclusions

Glacial geomorphological mapping from SRTM and Landsat ETM+ imagery and aerial photographs of southern Alberta has facilitated the identification of diagnostic landforms and landform assemblages (landsystem model) indicative of terrestrial-terminating ice stream

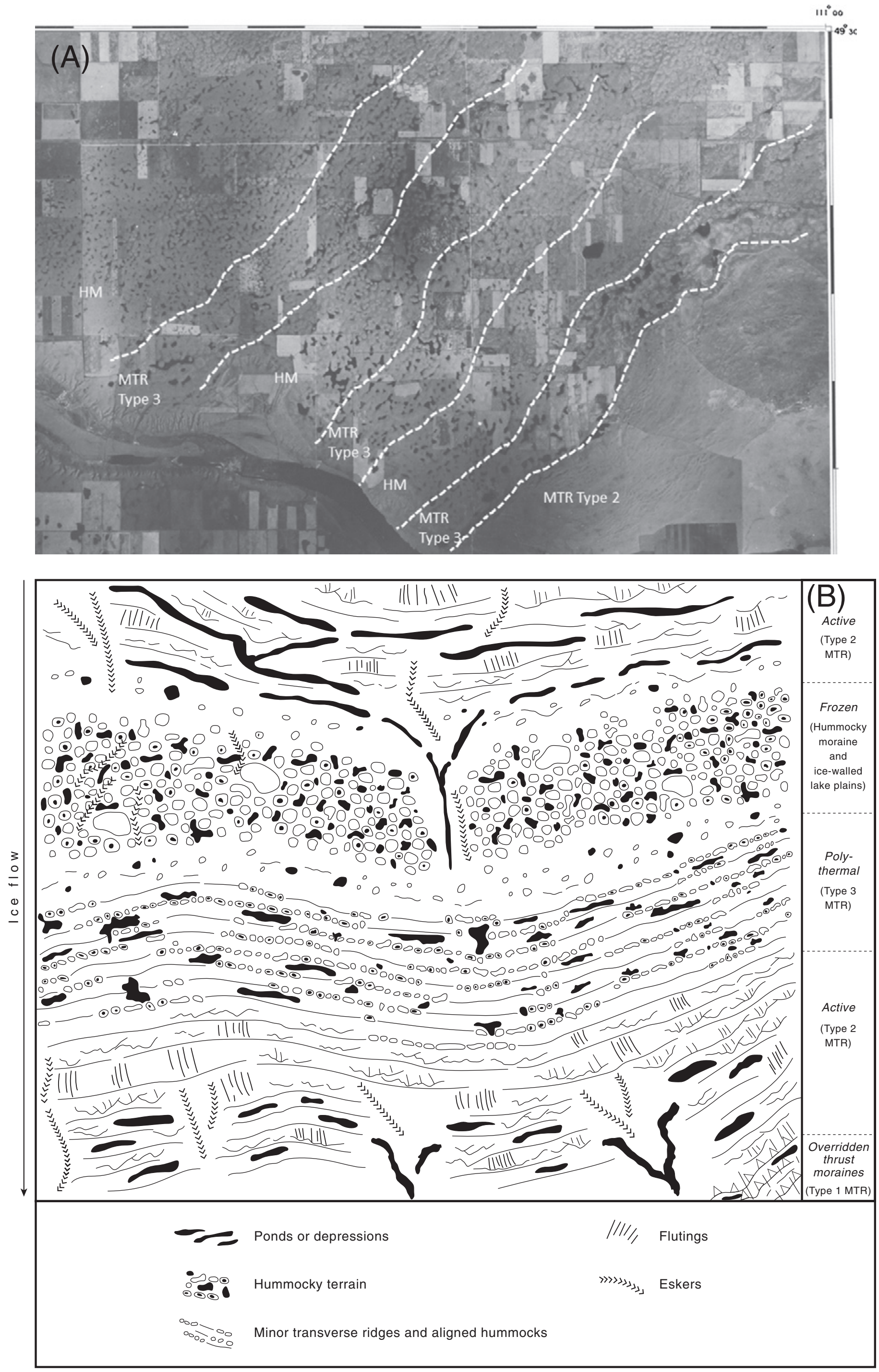


Fig. 21. CAIS marginal end moraine zonation/landsystem model: (A) aerial photograph mosaic of the area to the north of Pakowki Lake (see location on Fig. 11), showing the gradation from MTR type 2 in the southeast corner of the image to MTR type 3 and then to hummocky moraine with intermittent bands of MTR type 3 in a northwesterly direction; (B) conceptual model of the continuum of landforms created by the margin of the CAIS when it was located in the Milk River drainage basin (see text for details).

margins with lobate snouts. Spatial variability in landform type appears to reflect changes in palaeo-ice stream activity and snout basal thermal regimes, which are potentially linked to regional climate controls at the southwest margin of the Laurentide Ice Sheet but further research on this linkage is now warranted.

Small-scale mapping case studies of the High Plains (HPIS) and Central Alberta (CAIS) palaeo-ice stream tracks reveal distinct inset sequences of fan-shaped flowsets indicative of receding lobate ice stream margins. The lobate margins are recorded also by large, often glacially overridden transverse moraine ridges (commonly constructed through the glacitectonic thrusting of bedrock during ice streaming) and smaller, closely spaced inset sequences of recessional push moraines and hummocky moraine arcs (minor transverse ridges) produced during periods of alternating steady state normal-fast flow and cold-polythermal snout conditions. The locations of some MSGL on the downice sides of high points on ice stream beds are consistent with a groove-ploughing origin for lineations, especially in the case of the megafluting complex at the centre of CAfs_1 that appears as a flat-topped ridge with a grooved summit. During deglaciation, the dynamics of the CAIS in particular switched from fast flow/streaming (CAfs_2) to steady state normal-fast flow toward a lobate margin, which was subject to changing submarginal thermal regimes as recorded by the arcuate bands of MTR types 2–3 and hummocky terrain located between the preglacial divide (Milk River Ridge) and the Bow River catchment. Ice stream flow in the CAIS switched on again (CAfs_1) once the margin had retreated into the Red Deer/Bow River drainage basins.

Large-scale mapping of the southern limits of the CAIS reveals a complex glacial geomorphology relating to ice lobe marginal recession, comprising minor transverse ridges (MTR types 2–3), hummocky terrain (HT types 1–3), flutings, and meltwater channels/spillways. MTR type 1 likely originated through glacitectonic thrusting and have been glacially overrun and moderately streamlined. MTR type 2 sequences are recessional push moraines similar to those developing at modern active temperate glacier snouts. MTR type 3 document moraine construction by incremental stagnation because they occur in association with hummocky terrain. This localized close association of the various types of hummocky terrain with push moraine assemblages as well as proglacial permafrost features indicates that they are not ice stagnation landforms but rather the products of supraglacial controlled deposition on a polythermal ice sheet margin, where the HT type 3 hummocks represent former ice-walled lake plains.

The ice sheet marginal thermal regime switches indicated by the spatially variable landform assemblages in southern Alberta are consistent with palaeoglaciological reconstructions proposed for other ice stream lobate margins of the southern Laurentide Ice Sheet, where alternate cold, polythermal, and temperate marginal conditions sequentially gave way to more dynamic and surging activity. The sequential south to north change from hummocky terrain to MTR type 2 to MTR type 3 within the Lethbridge moraine and on the northern slopes of the Milk River Ridge records a temporal switch in CAIS marginal characteristics, from cold polythermal to temperate and then to warm polythermal. This is similar to patterns previously identified for the HPIS at approximately the same time based upon the available regional morphochronology and hence indicates a potential regional climatic control on ice sheet marginal activity. To the north of the Lethbridge Moraine, the landform assemblage of the Bow and Red Deer river basins, comprising overridden thrust moraines, megaflutings (CAfs_1), and a fluted esker complex lying inside the Vauxhall Ridge, records a later fast flow/streaming event. This was the precursor to the later ice stream surges that constructed the large thrust moraines and other surge-diagnostic landforms in central Alberta.

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