ERCB/AGS Open File Report 2012-16

Surficial Geology and Quaternary History of the Southern Lesser Slave Lake Area, Alberta (NTS 83O South)

Energy Resources Conservation Board

Alberta Geological Survey

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Contents

Figures

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Abstract

This report describes the distribution, character, and origin of surficial sediments south of Lesser Slave Lake in the Lesser Slave River (NTS 83O/SE) and Faust (NTS 83O/SW) map areas and outlines the Quaternary history of the region. This region consists of a narrow lowland that extends parallel to the lake and separates uplands rising to the north and south.

Two gravel sheets are identified in the southern Lesser Slave Lake area. Each sheet grades to a different base level and relates to two major cycles of fluvial aggradation and incision. These cycles are attributed to the evolution of regional, braided river systems draining eastwards from the Rocky Mountains since the early Quaternary.

Extensive till deposits document the advance of the Laurentide Ice Sheet across the southern Lesser Slave Lake area during late Wisconsinan glaciation. This advance covered the Lesser Slave basin and the surrounding uplands of the Swan Hills and Pelican Mountains, and deposited three major till facies across much of the southern Lesser Slave Lake area.

As the Laurentide Ice Sheet retreated northeastwards and down drainage from the southern Lesser Slave Lake area at ca. 12 ka BP, subaerial and glacial meltwater drainage systems converged towards this ice-blocked basin and formed a proglacial lake. Associated glaciolacustrine sediments form part of a larger expanse of proglacial lake sediments, which were deposited within Glacial Lake Peace at the end of the last glaciation. The extent of Glacial Lake Peace in the southern Lesser Slave Lake map area is documented by the distribution of glaciolacustrine sediments up to ~610 m asl. Glaciolacustrine sediments also occur locally at higher elevations, where lobes of the Laurentide Ice Sheet impounded lakes within larger river valleys during the early stages of deglaciation.

Following the regional retreat of the Laurentide Ice Sheet, the postglacial evolution of the Lesser Slave Lake area has been influenced by the accumulation of fine-grained alluvium, eolian sediment, and extensive peatlands, which mantle the glaciolacustrine deposits, particularly across the lowlands of the Lesser Slave Plain. Upland areas have been further influenced by slope failures and mass movements, with extensive colluvial deposits resulting from deep-seated landslides that fringe the steep slopes of the Pelican, Flat Top, and House mountains, and Grizzly Ridge.

1 Introduction

As part of an ongoing program, the Alberta Geological Survey (AGS) mapped the surficial geology of the Lesser Slave River (NTS 83O/SE) and Faust (NTS 83O/SW) map areas, in the southern half of the Lesser Slave Lake map area (NTS 83O; Figure 1). This work was published as AGS Surficial Geology Map 553 (Pawley, 2010) and Map 554 (Atkinson, 2010), respectively.

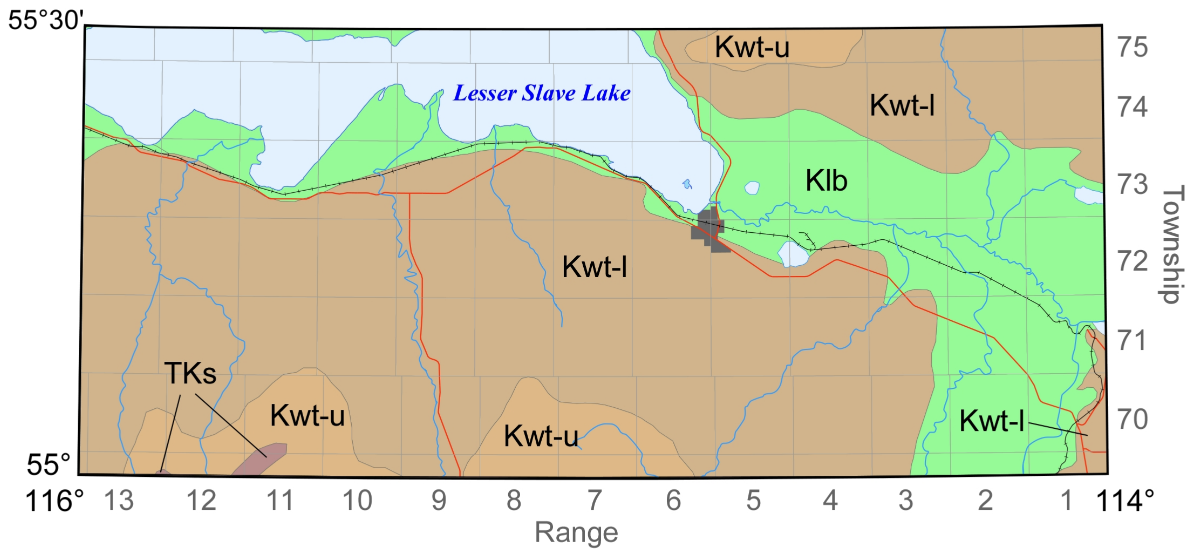
The mapping objectives were to describe the distribution, character, and origin of surficial sediments in the southern Lesser Slave Lake area. This report accompanies Maps 553 and 554 and presents a more detailed description of the characteristics, distribution, and genesis of these sediments and their associated landforms. The Quaternary history of the Lesser Slave Lake area is outlined, focusing on (1) the history of the preglacial river systems that occupied the Lesser Slave Lake basin through the early and middle Quaternary, and (2) the role of the Laurentide Ice Sheet in late Quaternary landscape evolution.

The surficial geology maps of the southern Lesser Slave Lake area were produced using mapping techniques that incorporate remote-sensing data, with an emphasis on the interpretation of airborne light detection and ranging (LiDAR) imagery. The LiDAR interpretations were augmented by observations from digital stereographic airphotos, orthographically corrected airphotos (1:60 000 scale), and satellite imagery (SPOT-5 and Landsat 7). These imagery-based mapping interpretations were performed directly in a geographic information system (ArcGIS), which improves the planimetric accuracy of the delineated landforms and enables the simultaneous mapping of multiple layers of remote-sensing data to highlight different characteristics of the terrain. Ground control was provided by 300 site descriptions, collected during field mapping in the summers of 2009 and 2010. Four-wheel-drive vehicles were used along roads and trails; more remote areas were accessed using a helicopter or an Argo 8x8 amphibious vehicle.

2 Bedrock Geology

The southern Lesser Slave Lake area is located in the north-central part of the Western Canada Sedimentary Basin (Mossop and Shetsen, 1994). It is underlain by Late Cretaceous sedimentary rocks that were deposited after 83 Ma, in a transition between marine and nonmarine environments during the regression of an interior seaway, termed the “Pakowki Sea” (Dawson et al., 1994; Hamilton et al., 1999; Figure 2). The nonmarine rocks of the lower Wapiti Formation form the uplands of the Swan Hills and Pelican Mountains, with lithologies dominated by grey feldspathic clayey sandstones and grey bentonitic mudstones with scattered coalbeds (Hamilton et al., 1999). In these uplands, the lower Wapiti Formation is unconformably overlain by up to 20 m of unconsolidated fluvial gravels, which cap the isolated plateaus of Marten, Flat Top, and House mountains and Grizzly Ridge. These deposits consist of clast-supported cobble gravel containing well-rounded quartzite clasts, traces of chert, and a coarse sandy oxidized matrix (Figure 3), and likely form a northern extension of the Oligocene Swan Hills Gravels (cf. Hamilton et al., 1999). The gravels represent the remnants of a large alluvial system that flowed across a gently inclined east to northeast dipping regional peneplain that formed a basinwide unconformity during the Oligocene to Miocene, following the Laramide Orogeny (Leckie, 1989, 2006; Leckie and Cheel, 1989; Leckie et al., 2004). The Lesser Slave Lake basin represents a trough that was eroded through the lower Wapiti Formation and into the underlying marine rocks of the Smoky Group. These rocks consist of dark grey shale and silty shale with ironstone partings/concretions and subcrop in a band striking west-northwest to east-southeast across the map area, where they underlie the lowlands surrounding Lesser Slave Lake.

Figure 1. Location of the southern Lesser Slave Lake map area, north-central Alberta.



Formation	Description
TKs Scollard Formation	**Paleocene and Late Cretaceous** grey feldspathic sandstone; dark grey bentonitic mudstone; thick coalbeds; nonmarine
Kwt-u Kwt-l Wapiti Group, upper/lower	**Late Cretaceous** grey, feldspathic, clayey sandstone; grey bentonitic mudstone and bentonite; scattered coalbeds; nonmarine
Klb Labiche Formation	**Early to Late Cretaceous** dark grey shale and silty shale; ironstone partings and concretions; silty fish-scale bearing beds in lower part

Figure 2. Bedrock geology of the southern Lesser Slave Lake area, north-central Alberta (Hamilton et al., 1999).

3 Physiography

The southern Lesser Slave Lake area occupies 7000 km^2 and is bounded by latitudes 55°00'N and 55°30'N, and longitudes 114°W and 116°W. The area straddles the Southern Alberta Uplands and Northern Alberta Uplands physiographic regions. These uplands are separated by the lowland of the Sturgeon and High Prairie plains, forming part of the Northern Alberta Lowlands, and the Lesser Slave Lowland, which is part of the Eastern Alberta Plains (Figure 4a; Pettapiece, 1986). The physiography of the region is dominated by Lesser Slave Lake, an elongate lake trending west-northwest to east-northeast that extends for 110 km across NTS 83O/SW and the western part of NTS 83O/SE (Figure 1). Lesser Slave Lake is the second largest lake in Alberta, covering 1170 km^2, and comprises two elliptical basins, each ~55 km long and up to 20 m deep. These basins are separated by The Narrows, a 3.8 km wide shoal where water depths decrease to typically less than 10 m. The lake drains eastwards along the Lesser Slave River into the Athabasca River. To the north of Lesser Slave Lake, the Utikuma Uplands include part of the Pelican Mountains, which rise to 940 m asl at Marten Mountain. The Swan Hills Upland extends across the southern part of the map area and includes the high plateaus of Flat Top Mountain (1040 m asl), Grizzly Ridge (1050 m asl), and House Mountain (1150 m asl; Figure 4b). A number of dendritic river systems drain these uplands, the most prominent being the Swan and Driftpile rivers, which drain northwards from the Swan Hills into Lesser Slave Lake, and the Driftwood and Fawcett rivers, which drain southwards from the Pelican Mountains into the Lesser Slave River. In the east of the southern Lesser Slave Lake area, the Athabasca River flows northwards across a wide lowland that is also occupied by the smaller Saulteaux River. The Athabasca River takes an acute eastward bend at the town of Smith and converges with the Lesser Slave River.

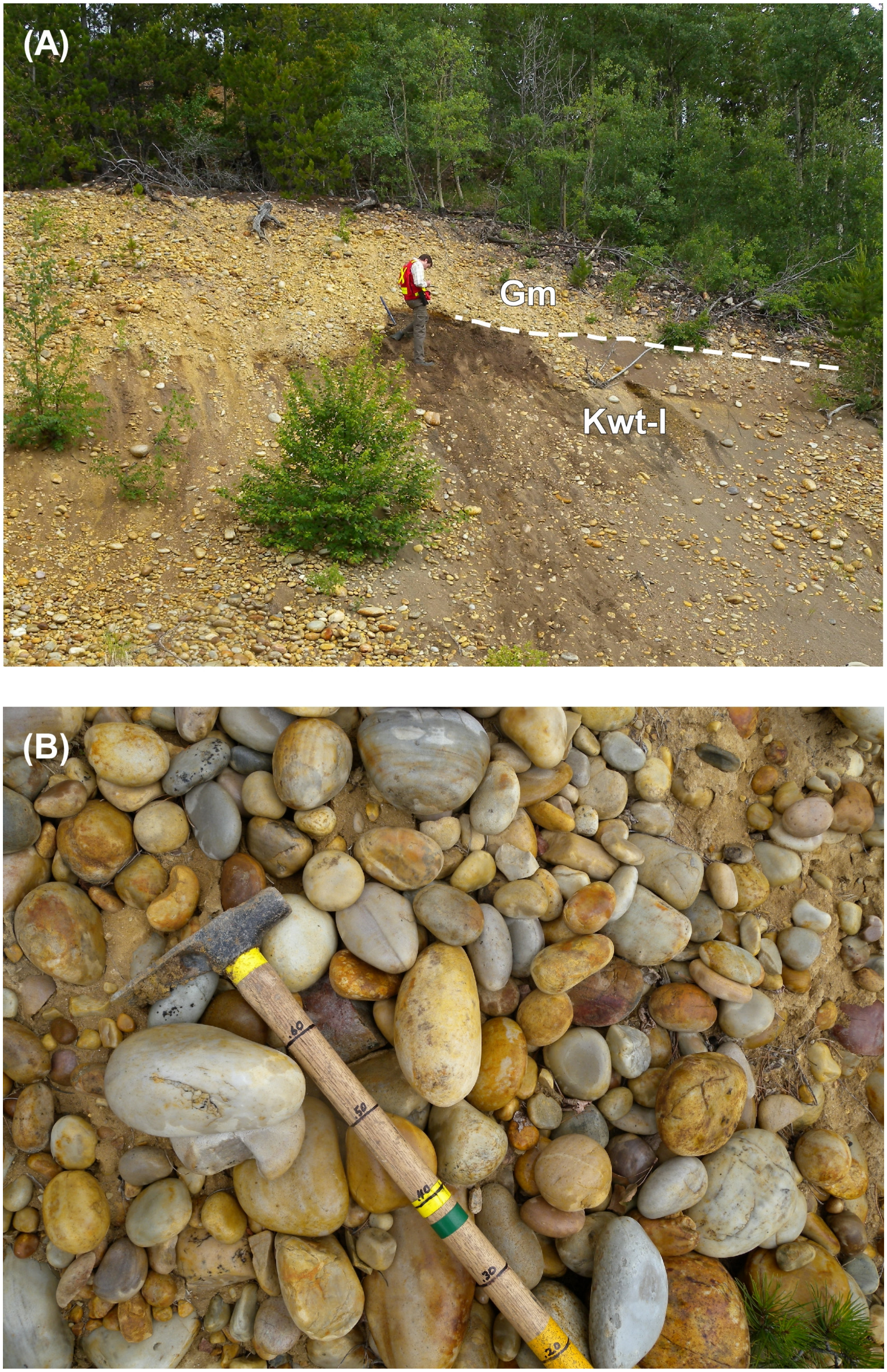
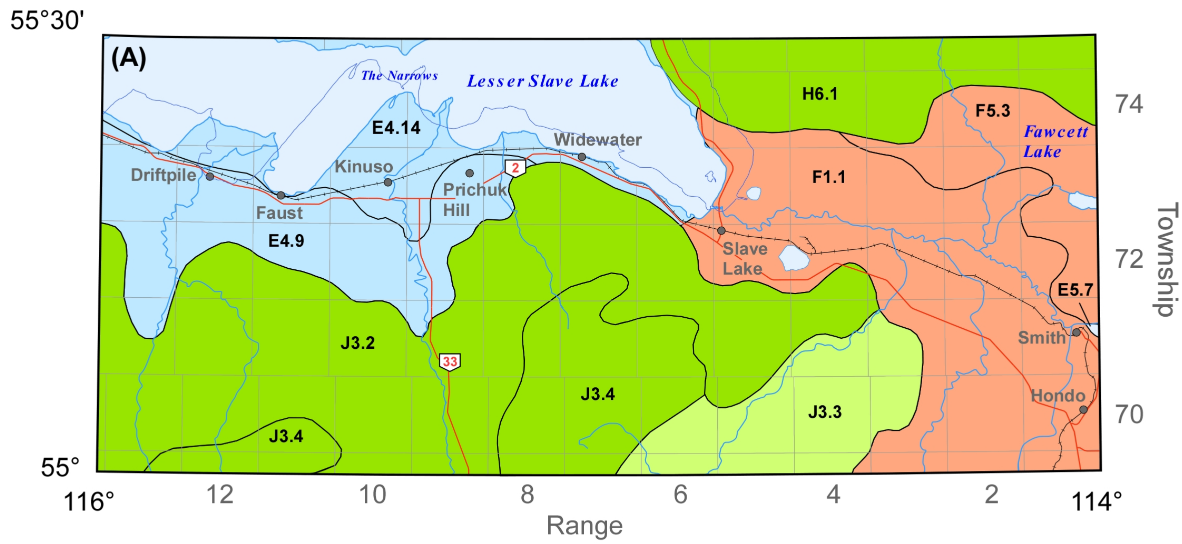


Figure 3. a) Unconsolidated fluvial gravel (Gm: gravels, massive) near Flat Top Mountain, directly overlying bedrock (Kwt-l: lower Wapiti Formation); b) consisting of well-rounded quartzite pebbles.



Region		Section		District	
E	Northern Alberta Lowlands (275–750 m asl)	E4	Peace River Lowland	E4.9	Sturgeon Plain
				E4.14	High Prairie Plain
		E5	Wabasca Lowland	E5.7	Athabasca Valley
F	Eastern Alberta Plains (450–950 m asl)	F1	Lesser Slave Lowland	F1.1	Lesser Slave Plain
		F5	Tawatinaw Plain	F5.3	Fawcett Plain
H	Northern Alberta Uplands (350–1050 m asl)	H6	Utikuma Uplands	H6.3	Pelican Mountains
J	Southern Alberta Uplands (600–1650 m asl)	J3	Swan Hills Upland	J3.2	Driftpile Benchland
				J3.3	Freeman Benchland
				J3.4	Swan Hills

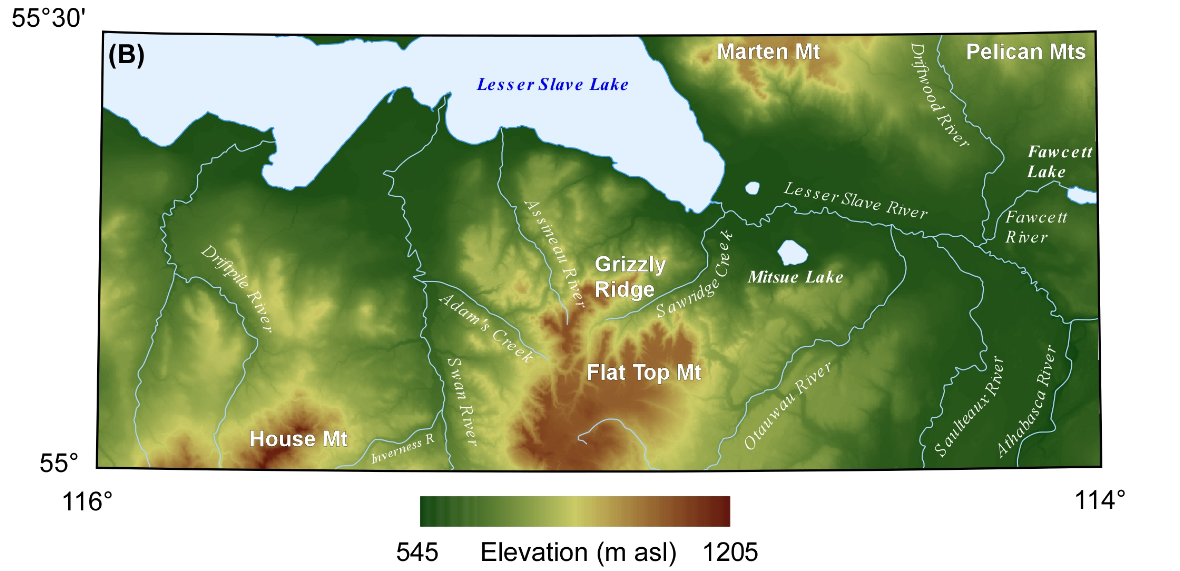


Figure 4. a) Physiographic regions; b) topography of the southern Lesser Slave Lake area, north-central Alberta (Pettapiece, 1986).

4 Vegetation

Lesser Slave Lake is in the mixed-wood section of the Northern Boreal Forest (Rowe, 1972). Near-continuous mixed conifer and deciduous forest covers the uplands with white spruce (*Picea glauca*) and quaking aspen (*Populus tremuloides*), together with smaller amounts of western balsam poplar (*Populus trichocarpa*), white birch (*Betula papyrifera*), and balsam fir (*Abies balsemea*). Lodgepole pine (*Pinus contorta*) and jack pine (*Pinus banksiana*) grow on Marten Mountain, Flat Top Mountain, and Grizzly Ridge, where the unconsolidated fluvial gravels provide a well-drained substrate. Lodgepole pine also occurs widely across areas of deglacial and Holocene sand dunes on the lowlands flanking the Lesser Slave and Athabasca rivers. The lowland surrounding Lesser Slave Lake is generally poorly drained and dominated by black spruce (*Picea mariana*), which is commonly associated with bogs of sphagnum moss and Labrador tea (*Ledum groenlandicum*). Areas covered by fen occur locally and are surrounded by marsh grasses, larch (*Larix spp.*), and sedges (*Cyperaceae spp.*).

5 Surficial Geology

The surficial geology maps of the region show the distribution of 10 genetic types of surficial material (Atkinson, 2010; Pawley, 2010; Figure 5). These materials and their associated landforms are described below in order of their relative ages (from oldest to youngest). A number of stratigraphic section logs in these materials establish the sedimentary succession in the region (Figure 6). The facies codes used in this report are from Benn and Evans (1998), modified from Eyles et al. (1983). Section 7 outlines the significance of these stratigraphic units to the Quaternary history of the southern Lesser Slave Lake area.

5.1 Unconsolidated Fluvial Gravels (Unit RT)

Unconsolidated fluvial gravels occupy the highest topographic position of surficial deposits in the southern Lesser Slave Lake area, occurring between 1000 and 1150 m asl. These deposits form sheets that cap the isolated plateaus of Marten, Flat Top, and House mountains and Grizzly Ridge (Figure 5). Because these plateaus are fringed by recessive slopes, these sheets are commonly mantled by colluvium along these slopes, therefore their architecture was not observed. This unit contains the same lithologies and occupies a similar topographic position as the Oligocene (late Paleogene) Swan Hills Gravels that cap the Swan Hills ~100 km to the southwest (Hamilton et al., 1999).

5.2 Preglacial Fluvial Deposits (Unit FP)

Preglacial fluvial deposits occur in two topographic settings in the southern Lesser Slave Lake area, subcropping at mid-level elevations (~650 to 720 m asl) on the moderately sloping northern flank of Grizzly Ridge and at lower elevations (~610 to 620 m asl) on the Sturgeon and High Prairie plains, adjacent to the Swan, Driftpile, and Driftwood rivers (Figure 4b). These deposits were observed in the field in sections beneath overlying stratigraphic units, and their distribution is indicated on the surficial geology maps (Atkinson, 2010; Pawley, 2010).

Lithologically, the deposits are similar to the pre-Quaternary fluvial gravels that cap the higher plateaus (unit RT), in that their clast content consists exclusively of well-rounded quartzite with traces of black chert. However, the preglacial fluvial deposits are usually finer grained and dominated by medium to large pebble-sized clasts, rather than the cobble-sized material prevalent within the pre-Quaternary gravels. Although pre-Quaternary fluvial deposits may be assigned to either of these surficial units (RT or FP), in the absence of dating and/or stratigraphic controls, they may actually relate to a range of multi-age, multi-event depositional units, rather than the topographically based single units presented in this report. The absence of clast lithologies derived from the Canadian Shield also distinguishes the preglacial fluvial deposits from younger glacial and postglacial gravel deposits in the report area.

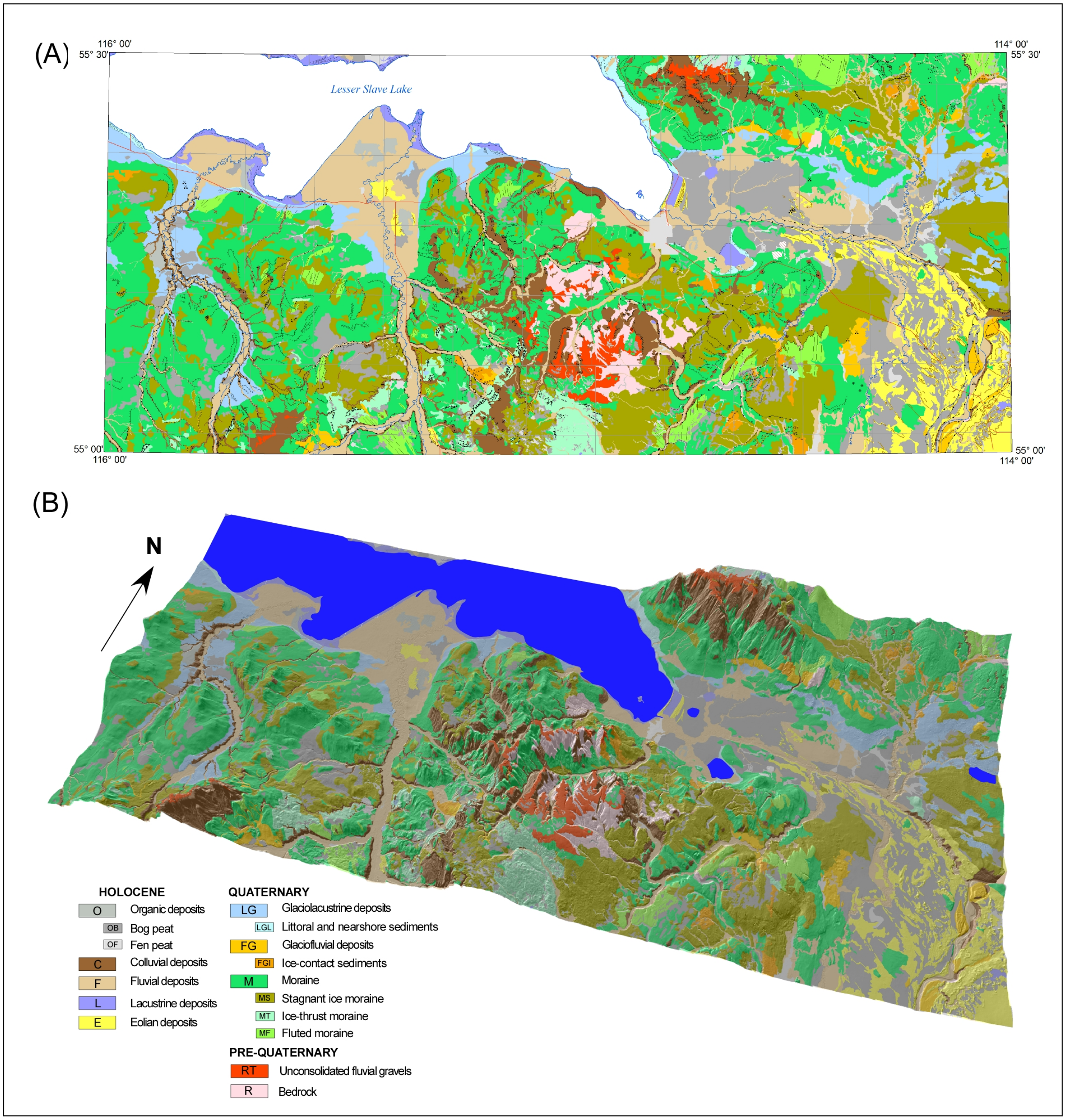


Figure 5. a) Surficial geology of the southern Lesser Slave Lake area (Atkinson, 2010; Pawley 2010); b) surficial geology draped on an oblique, shaded-relief light detection and ranging (LiDAR) digital elevation model with a northwestern light source and 7x vertical exaggeration.

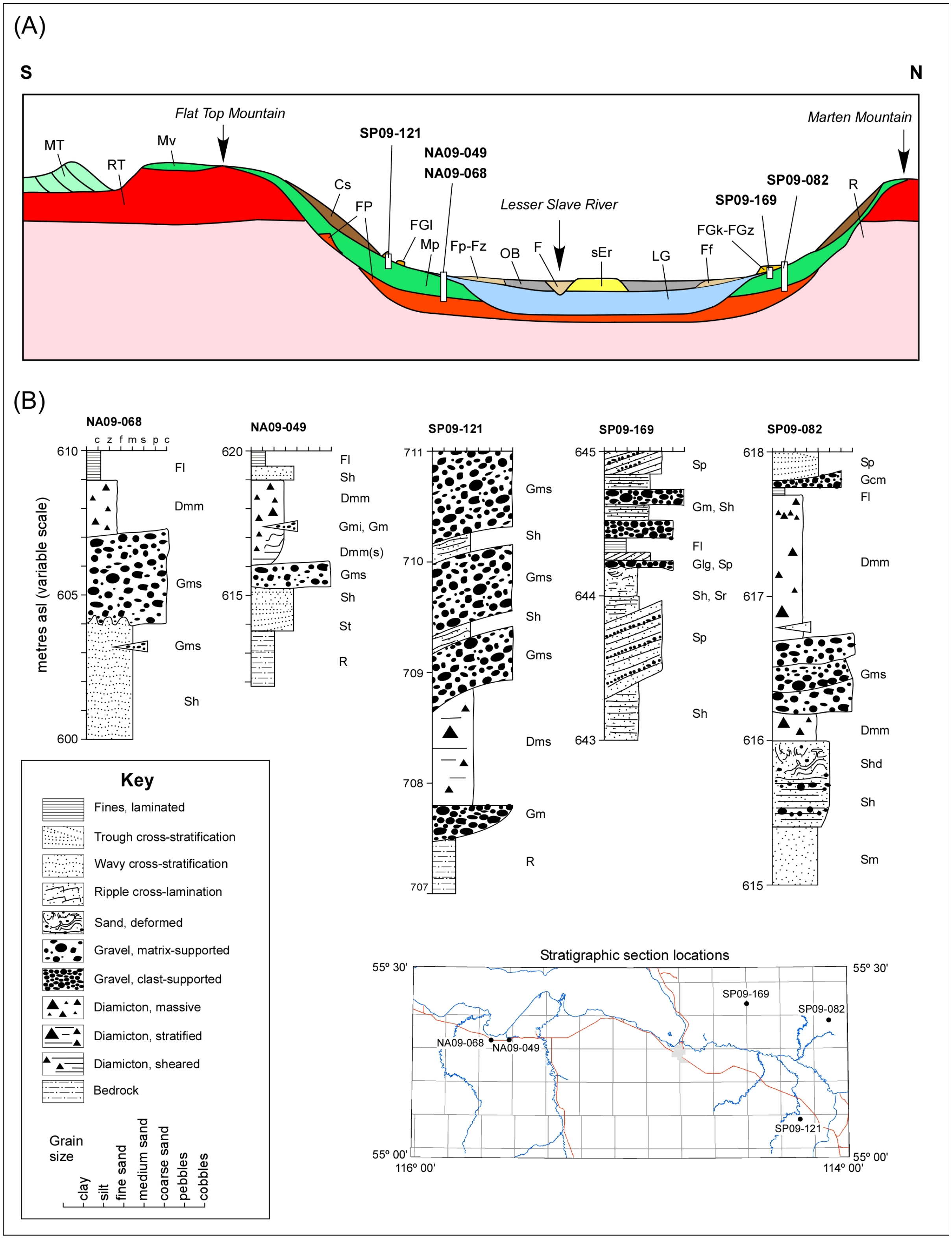


Figure 6. a) Schematic cross-section of the sedimentary succession in the southern Lesser Slave Lake area; b) vertical profiles showing the logged stratigraphic sections referred to in the text.

5.2.1 Mid-Level Fluvial Deposits

The mid-level fluvial deposits consist of an assemblage of sand and gravel that was logged between ~650 and 720 m asl in two gravel pits adjacent to Highway 2 between Widewater and Prichuk Hill (NA09-81 and NA09-185, respectively; Figure 4a). This lithofacies assemblage is up to 10 m thick and is composed of three sedimentary facies: (1) structureless to weakly imbricated cobble gravel (Gm, Gmi); (2) ~20 cm thick lenses of fine- to medium-grained, horizontally bedded sands (Sh); and (3) weakly imbricated, matrix-supported cobble gravel (Gsi, Gms; Figures 6 and 7).

Framework clasts within Gsi and Gms typically compose 60% of the lithofacies, with the matrix composed of coarse-grained sand. Basal contacts that separate these facies are defined by both erosional- and gradational-bounding surfaces. Clasts within the framework are oxidized and entirely composed of Cordilleran lithologies and are well rounded but display low sphericity. This lithofacies assemblage overlies lower Wapiti sandstone along an undulating contact, which is inclined gently to the east.

Figure 7. Matrix-supported cobble gravel (Gms) overlain by massive, matrix-supported diamict (Dmm) along an erosive contact (dashed line).

5.2.2 Lower-Level Fluvial Deposits

Lower-level fluvial deposits were logged at ~610 and 620 m asl in two gravel pits excavated on the lowlands of the High Prairie Plain (NA09-68 and NA09-49; Figures 6 and 8). This assemblage overlies lower Wapiti sandstone along a sharp, horizontal contact and comprises a 2 to >5 m thick sequence of interbedded pebble to cobble gravel and fine- to medium-grained sand. The lowest exposed unit of this assemblage is characterized by an ~2.5 m thick sand unit comprising trough cross-bedded and rhythmically bedded sand and silt with medium gravel interbeds, often only several clasts thick (St/Sh/Gms; Figure 9). This rhythmically bedded sand unit is overlain along a sharp contact that exhibits flame structures (Type B pillars: Lowe, 1975; Figure 10) by a laterally continuous, long-wavelength, concave-up gravel lens (Figure 11). Internally, this lens is composed of up to 3 m of massive, matrix-supported cobble to boulder gravel (Gms). Framework clasts within Gms typically compose 70% of the lithofacies, with the matrix composed of fine-grained sand. Clasts within the framework consist exclusively of Cordilleran lithologies and are well rounded but display moderate to low sphericity. Elongate clasts (low b/a axis ratios) are commonly disaggregated along high-angle fractures (Figure 10). This lithofacies assemblage, as well as the associated sedimentary succession, appears to occur widely across the Sturgeon Plain and the Driftpile Benchland and was observed along the flanks of a number of tributaries of the Swan River, including the Inverness River (Figures 4 and 12).

In the east of the southern Lesser Slave Lake area, fluvial deposits occurring at low elevations (615 to 618 m asl) were logged in a gravel pit excavated 5 km to the north of Fawcett Lake (SP09-082; Figures 4 and 6). The fluvial deposits exposed at the site are at least 3 m thick and consist of massive gravel with minor units of interbedded diamicton and horizontally laminated sand. The gravel units are matrix supported (Gms), with a matrix of poorly sorted silty sand, and with clasts representing at least 60% of the facies. These show large variations in modal grain size between beds, varying from medium pebbles to small cobbles. Some beds show crude vertical grading. Individual beds are separated by erosional contacts, with gently inclined cross-cutting surfaces causing some beds to cut out laterally across the section. Finer grained facies form only a minor component of the assemblage and consist of a horizontally bedded sand unit (Sh), which is deformed into density-driven loading and flame structures along the contact with an overlying 20 cm thick bed of silty diamicton (Dmm; Figure 6) in the lower part of the exposed sequence. The diamicton matrix has a silty texture and a light brownish-grey colour, similar to that of the overlying Gms facies, and contains a laterally variable concentration of clasts, exclusively consisting of quartzite and chert rock types with no Canadian Shield varieties present. The unit also has a disorganized structure and unordered clast fabric. The lateral extent of this sedimentary succession across the Fawcett Plain remains unknown because the units are buried by overlying till and glaciofluvial deposits (Figure 6).

Figure 8. Oblique, east-facing view of two gravel pits excavated into the High Prairie Plain. Inset shows the location of the logged section exposed at NA09-68.

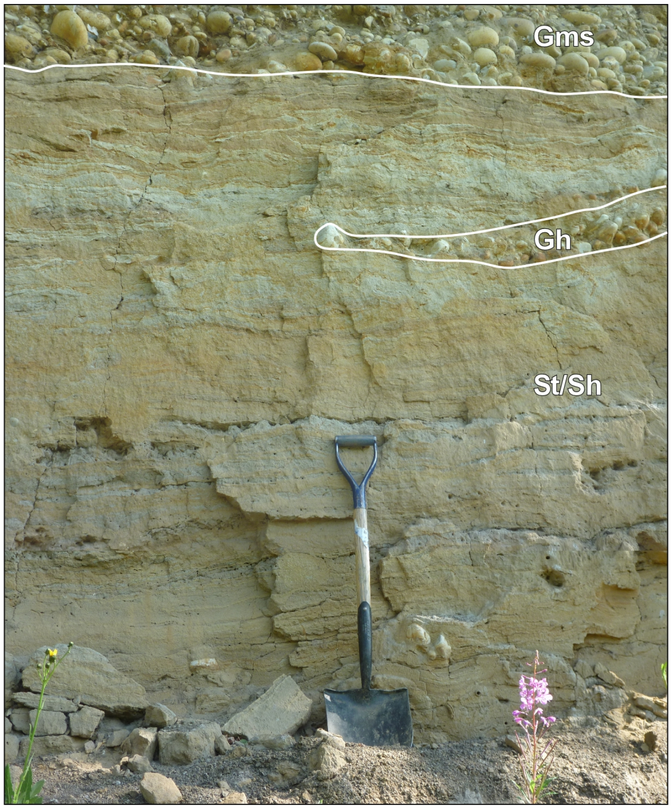


Figure 9. Lowest lithofacies exposed at site NA09-68, comprising an ~2.5 m thick sand unit exhibiting trough cross-bedded and rhythmically bedded sand and silt (St/Sh) with medium gravel interbeds (Gh/Gms).

Figure 10. Evidence of soft-sediment deformation, demonstrated by flame structure, along the contact between rhythmically bedded sand (Sr/Sh) and overlying matrix-supported gravel (Gms). Arrows denote fractured quartzite cobbles along high-angle faults in this preglacial fluvial deposit.

Figure 11. Laterally continuous, concave-up lens of matrix-supported cobble to boulder gravel (Gms) overlying rhythmically bedded sand (Sr/Sh) and capped by a strongly columnar-jointed, matrix-supported Laurentide till (Dmm).

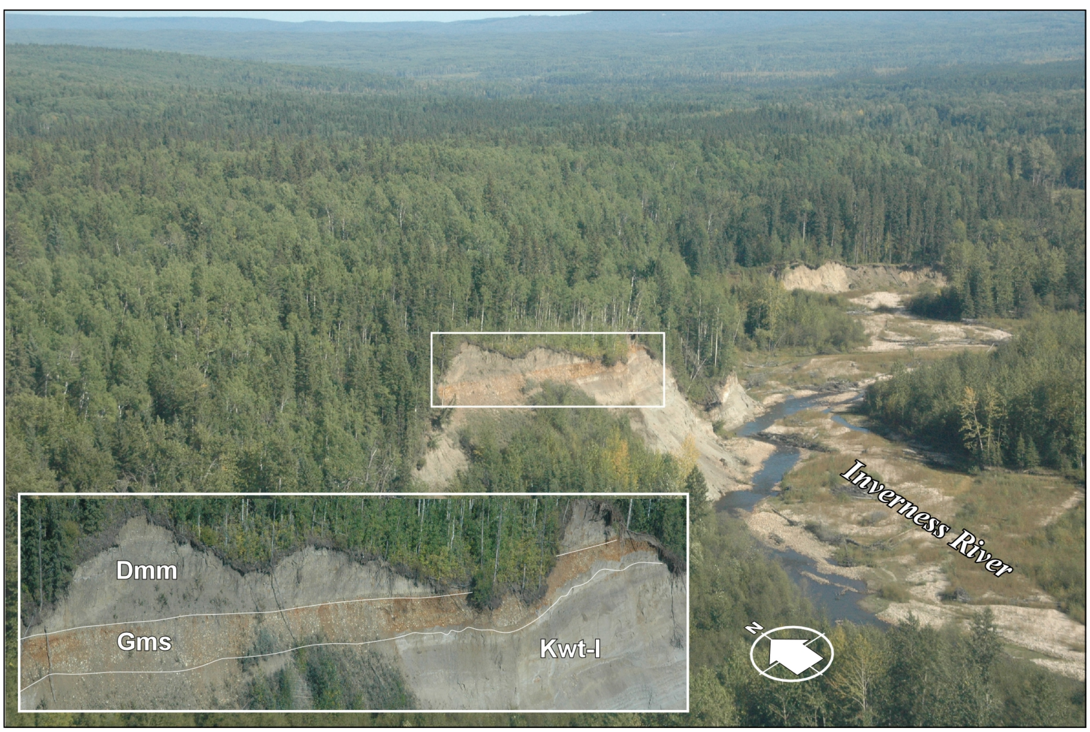


Figure 12. Sedimentary succession exposed along a flank of the Inverness River. Inset shows contacts and thickness of the matrix-supported gravel (Gms) and massive, matrix-supported diamict (Dmm) overlying lower Wapiti sandstone (Kwt-l).

5.3 Moraine (Units M, MS, MT, MF)

Moraine forms the most widespread surficial material across the southern Lesser Slave Lake area (Figure 5). Moraine veneer (Mv) is mapped wherever till is too thin to obscure the relief of the underlying unit (<2 m thick). Typically, moraine veneer occurs across the gravel-capped uplands of Flat Top, Marten, and House mountains and Grizzly Ridge (Figure 6a). Moraine plain (Mp) is mapped wherever till obscures the topography and structure of the underlying unit (≥2 m thick) and occurs in areas of lower relief and on the mid- to low-level slopes flanking Pelican, Flat Top, and House mountains and Grizzly Ridge (Figures 6a and 13).

Figure 13. Moraine plain (unit Mp) composed of dark grey, clast-poor (3%-5%), sandy-silt diamict; site SP09-081.

In areas with moderate to steep slopes, moraine veneers and plains are commonly dissected by gullies that are inferred to have formed by Holocene fluvial dissection and mass wasting processes. Moraine veneers and plains on low to moderate slopes commonly exhibit a suite of irregular undulations that document the retreat of Laurentide ice lobes during regional deglaciation. Areas of hummocky moraine are commonly associated with dissected and gullied moraine, likely resulting from erosion and reworking by glacial meltwater and periglacial rivers, as well as the modern drainage of ephemeral streams during spring snowmelt.

Stagnant-ice moraine (MS) is common in moderately high relief terrain and comprises a variety of landforms that in places increase the local relief by 2 to 5 m (Figure 14). These landforms include hummocky topography consisting of chaotic to crudely organized ridges and depressions and 'doughnut moraine' consisting of sharply defined, circular to ellipsoid rings of till with central depressions that are commonly filled with postglacial organic deposits and/or glaciolacustrine sediments. Stagnant-ice moraine extensively fringes glaciolacustrine sediments to the east of Lesser Slave Lake where doughnut-type forms are present, as well as areas covered by reticular and oval to crudely diamond-shaped patterns of ridge networks (up to 3 m high), which surround central depressions filled with glaciolacustrine sediments. These ridge network patterns diminish downslope into the flat glaciolacustrine plain and upslope into till-dominated, stagnant-ice moraine.

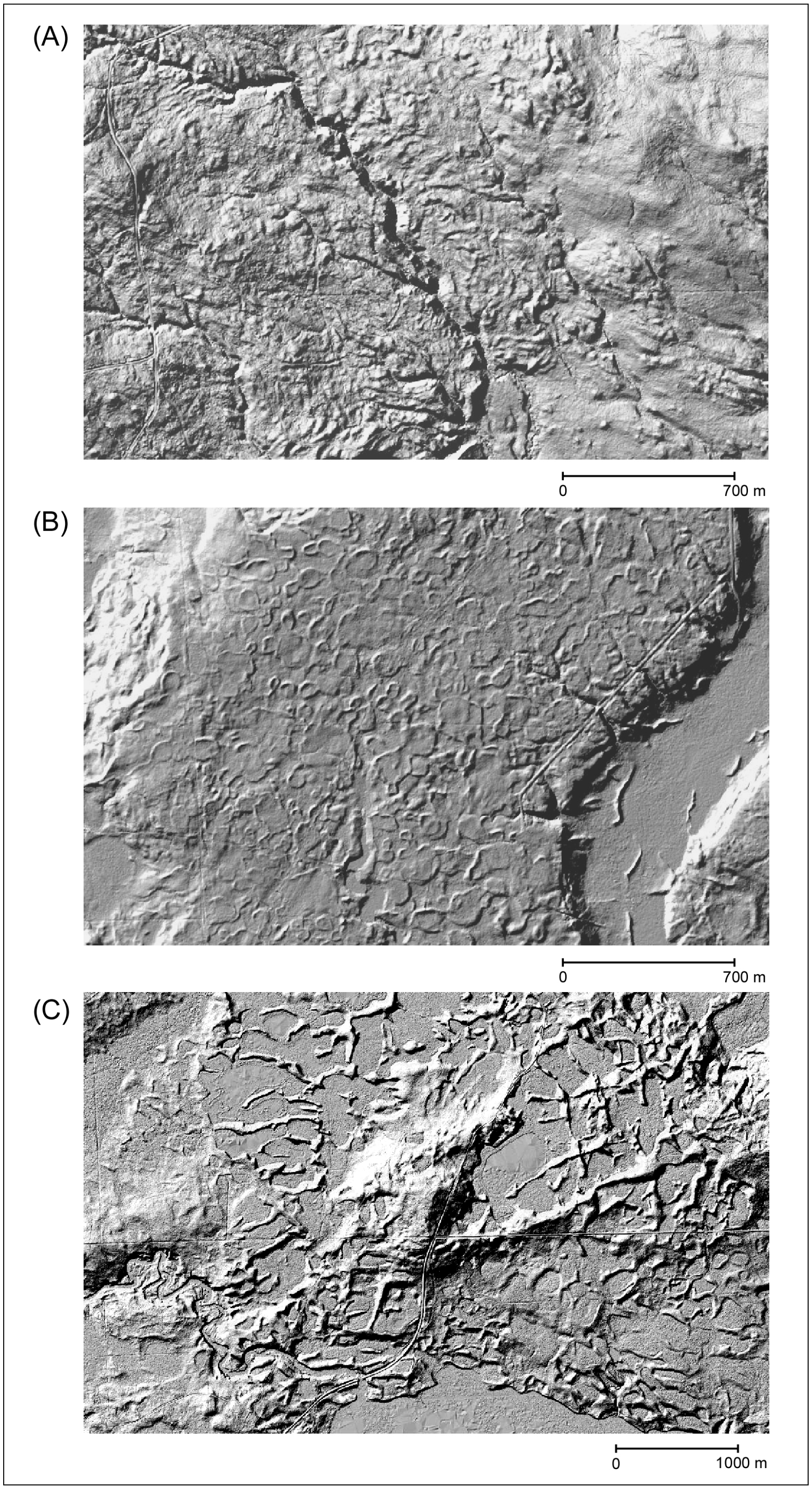


Figure 14. Examples of ice-stagnation landforms as shown on light detection and ranging (LiDAR) relief-shaded imagery: a) hummocky topography with chaotic ridges and depressions; b) typical doughnut moraine with circular to ellipsoid rim ridges and central depressions; c) reticulated patterns of ridge networks.

Ice-thrust moraine (MT) is a locally significant landform, imparting moderate to high relief (2 to 5 m) on the slopes flanking the Swan River and on the upland plateau to the south of Flat Top Mountain (Figure 4). On the eastern slope of House Mountain, ice-thrust moraine is characterized by high-relief, undulating terrain containing numerous bedrock outcrops interspersed with till veneer. Cross-sections exposed by road cuts through this terrain reveal inclined and folded bedrock strata, interpreted to be the result of glaciotectonic deformation (Figure 15). Similarly, glaciotectonic compression is considered to be responsible for suites of sharp-crested, concentric Cretaceous bedrock ridges fringing the northwest slopes of House Mountain and Grizzly Ridge.

Figure 15. Glaciotectonically deformed beds of grey bentonitic mudstone (shown with dashed line) within exposed lower Wapiti Formation sandstone on the eastern flank of House Mountain.

Locally, areas of moraine contain tracts of fluted and streamlined bedforms. Fluted moraine (MF) extends across parts of the Pelican Mountains and is characterized by parallel flutes up to 3 km long, 50 m wide, and 2 m high, and oriented between 195° and 205°, with some local variation of up to 20°. Other evidence for glacial moulding of the bed includes subtle streamlined parallel ridges that are present in the south of the study area and strike with a similar northwest-southeast orientation. Moraine proximal to the south shore of Lesser Slave Lake exhibits a number of widely dispersed, low-gradient streamlined tracts (1 to 5 km wide) containing suites of parallel, low-amplitude flutings. Individual flutings are up to 1 km long and 35 m wide, with orientations ranging from 200° to 230°. Farther south, streamlined tracts exhibit more variable orientations (130° to 200°), evidently due to the southeastward deflection of ice around the flanks of House Mountain and Grizzly Ridge and the resulting flow convergence along the Swan River valley. In a number of areas, fluting tracts either emanate from undulating ice-thrust moraine or terminate at glaciotectonized bedrock ridges.

5.3.1 Till Sedimentology, Composition, and Distribution

A field-based assessment of till characteristics demonstrates that three till lithofacies are found in the southern Lesser Slave Lake area (Figure 6). The first lithofacies comprises a 1 to 2.5 m thick sequence of very dark greyish-brown (2.5Y 3/2), massive, matrix-supported, clayey-silt diamict (Dmm[s]; Figure 16). The clast content, which includes Cretaceous, Paleozoic, and Canadian Shield rock types, is typically 2% to 5% and comprises mainly subangular granules, with occasional subrounded and glacially faceted pebbles. This till typically occurs proximal to the south shore of Lesser Slave Lake and adjacent to lowland basins, including the upper reaches of river valleys and local topographic depressions.

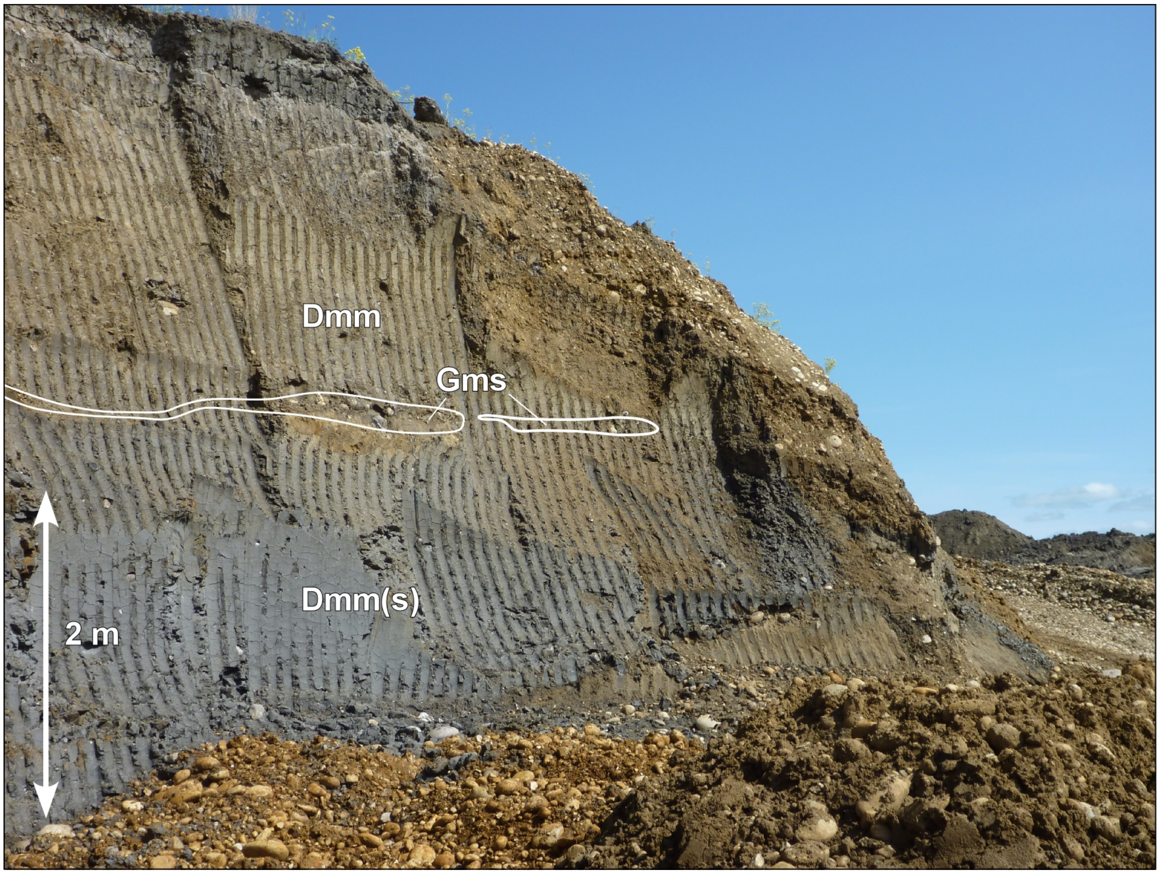


Figure 16. Two till lithofacies exposed in a gravel pit adjacent to the Driftpile River comprising clayey-silt till (Dmm[s]) overlain by sandy-silt till (Dmm).

The second lithofacies comprises a 1 to 5 m thick sequence of very dark grey (2.5Y 3/1), massive, matrix-supported, sandy-silt diamict (Dmm; Figures 16 and 17), which is usually altered to greyish brown (10YR 4/2) to a depth of 3 m by oxic groundwater conditions. This till has a high bulk density and commonly exhibits a fissile structure, characterized by well-developed vertical dilation joints (Figure 17). The till is generally clast poor (2% to 5%) and contains subangular granules to subrounded pebbles of quartzite, ironstone, partly disaggregated mudstones, and green sandstone streaks, together with Canadian Shield rock types. This till typically occurs in areas of lower relief, primarily on mid- to low-level slopes flanking Pelican, Flat Top and House mountains and Grizzly Ridge. Coal clasts commonly occur in this till, particularly across the Pelican Mountains and northern Swan Hills, where coalbeds in the local bedrock have been incorporated into the till through subglacial shearing (Figure 18). The till does not usually have any significant calcium carbonate content, although reprecipitated calcium carbonate nodules occur within the illuviated B soil horizon. In rare cases, the till has a weak hydrocarbon odour due to the incorporation of bitumen-bearing rocks located up-glacier.

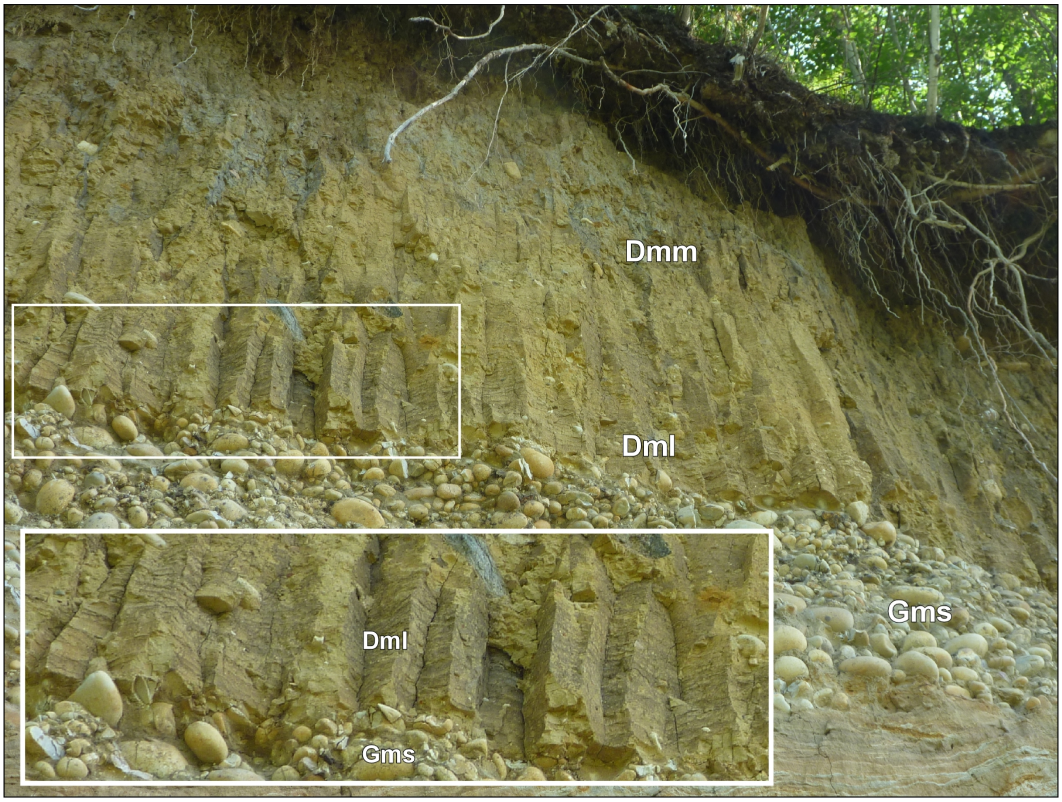


Figure 17. Well-developed vertical dilation joints within sandy-silt till (Dml, Dmm) exposed in a gravel pit west of the town of Faust at site NA09-68.

Figure 18. Glaciotectonically deformed coal incorporated into the base of a clast-rich till; site SP09-121.

In sections exposed within two gravel pits adjacent to the Driftpile River, the clayey-silt till is overlain by the sandy-silt till. The base of the clayey-silt till is marked by a shear amalgamation zone (Dmm[s]; Figure 19) within which attenuated clayey diamict pods separate deformed sand (Sm) and gravel beds (Gmi/Gm) that were incorporated from the underlying lower-level preglacial fluvial deposit (Unit FP; Figure 6). The overlying sandy-silt till (Dmm) is separated from the clayey-silt till (Dmm[s]) by a diffuse contact, which in places is marked by a discontinuous horizon of well-rounded quartzite cobbles (Gms) one to five clasts thick (Figure 16). Evidence of shear amalgamation is also present along this contact, with the basal 40 cm of the till containing horizontal bands of clay that are inferred to have been sheared from the underlying clayey till (Dml; Figure 17).

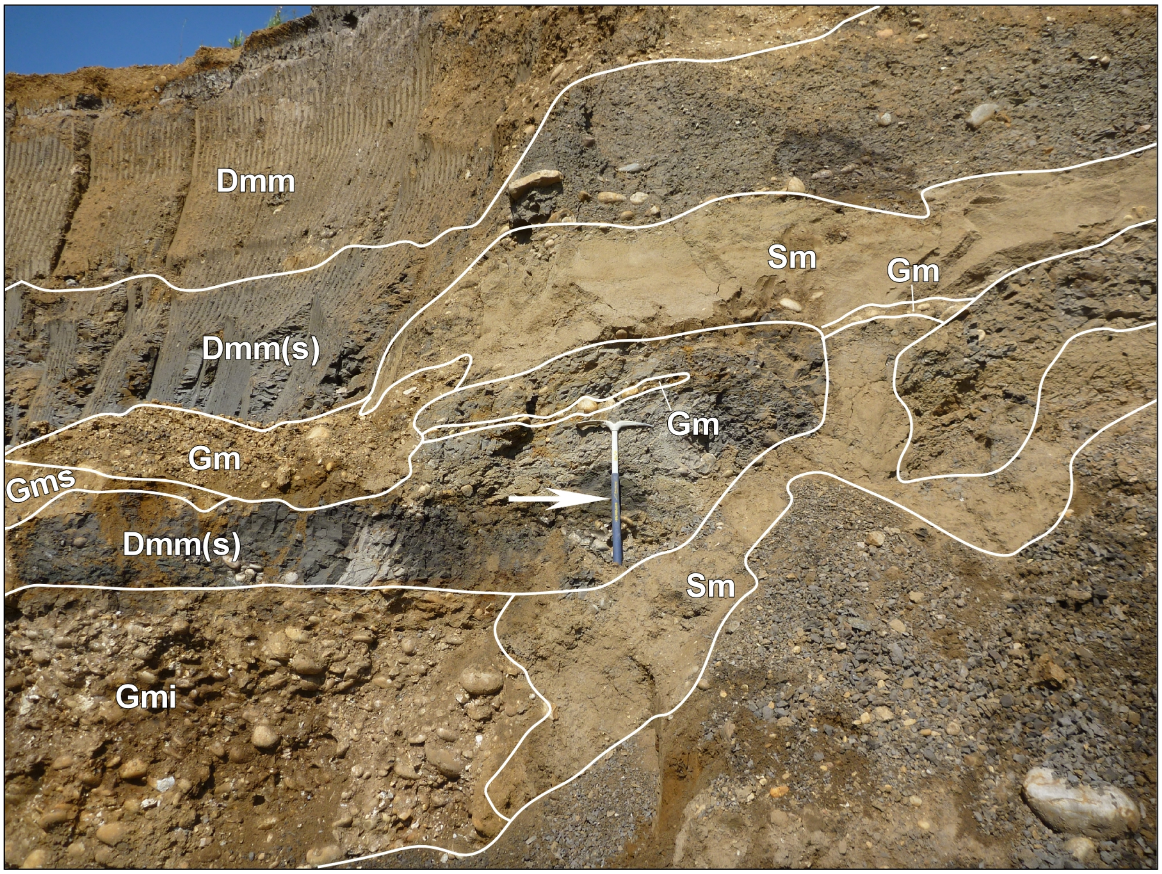


Figure 19. Attenuated gravel and diamict lithofacies within a subglacial shear amalgamation zone exposed in a gravel pit adjacent to the mouth of the Driftpile River.

The third lithofacies consists of a brown (10YR 4/3), clast-rich (30% to 50%), matrix-supported, sandy-silt diamict. Clast content is dominated by well-rounded quartzite cobbles eroded from underlying gravel (units RT and FP), although subangular to subrounded granules and cobbles of Cretaceous, Paleozoic and Canadian Shield rock types are also present. This till facies typically occurs as moraine veneer across Flat Top, Marten, and House mountains and Grizzly Ridge.

5.4 Glaciofluvial Deposits (Units FG, FGI) and Associated Landforms

5.4.1 Glaciofluvial Terraces, Outwash Fans, and Ice-Contact Deposits

Glaciofluvial deposits occur within three types of sediment-landform associations across the study area. Major deposits of moderate- to well-sorted sand and gravel occur along the Athabasca River in the area surrounding the towns of Hondo and Smith (Figure 4a). These deposits form a glaciofluvial terrace between 580 and 590 m asl that originated as bank-attached bars within the inner part of meander bends or as mid-channel bars that were bounded by paleochannels. These paleochannels scoured the lowland in several places between the Athabasca, Saulteaux, and Lesser Slave rivers, although infilling by organic material has subdued their morphology.

Sandar and glaciofluvial deltas occur at the termini of a number of meltwater channels emanating from Pelican and House mountains and Grizzly Ridge. Deposits within these landforms consist of horizontally and cross-bedded, fine- to coarse-grained pebbly outwash sand with minor well-rounded boulder to cobble lags (site SP09-169; Figure 6). Far-travelled clasts from the Canadian Shield are abundant within these deposits, in addition to locally derived quartzite clasts. Several areas of pitted outwash (FGk) occur within an east-west trending deposit fringing the southern flank of the Pelican Mountains and include extensive deposits of moderately sorted, coarse-grained pebbly sand. Numerous kettle holes occur in the surface of the outwash plain, where they were formed by the melting of ice blocks, buried within the outwash plain.

Ice-contact deposits (FGI), including eskers, kames, sandar, and pitted outwash, occur locally on the plateaus and flanks of Pelican, Flat Top, and House mountains and Grizzly Ridge (Figure 5). The eskers range from single, isolated ridges to anastomosing ridge complexes composed predominantly of sand and gravel and extend for up to 2 km across glaciofluvial outwash plains and areas of stagnant-ice moraine. Kames also occur in association with eskers and stagnant-ice moraine and consist of assemblages of irregular hills and hummocks with 5 to 10 m relief. A section through one of these kames on the upland surrounding the Otauwau River (Figure 4b) reveals several sheets of poorly sorted, pebble to cobble, matrix-supported, sandy gravel (Gms; Figure 20a) overlying a clast-rich diamicton (Dms) along a sharp, undulating contact. Clasts within the gravel beds represent ~60% of the facies and are supported within a matrix of poorly sorted, silty coarse-grained sand (Figure 20b). Lithologies include a mixture of quartzite and Canadian Shield lithologies, along with a small component of coal and mudrock. Thin, silty sand beds separate the thicker gravel sheets and contain silt laminations and clasts of sandstone and coal (Sh). Structurally, the sand and gravel assemblage shows evidence of gravity-induced deformation, with the bedding situated close to the margin of the landform being downwarped towards the north (Figure 20c). Examination of sections in a number of other ice-contact deposits, forming crude ridges, hills and hummocks across mid-level slopes surrounding the upland of Flat Top Mountain, reveal that these landforms include a similar range of sedimentary facies.

5.4.2 Meltwater Channels

The Swan and Driftpile rivers, together with their larger tributaries, currently occupy relatively small channels that meander through wide (up to ~2.5 km), flat-floored, misfit valleys. Consequently, these rivers, as well as the abandoned channels flanking the Athabasca River, are interpreted as proglacial meltwater channels that evolved during regional deglaciation. Elsewhere, the uplands of the Swan Hills and the Pelican Mountains are dissected by numerous valleys, which form broad, steep-sided troughs that extend orthogonally across local slopes (Figure 21). These valleys are aligned perpendicular to the fluting tracts and have high-sinuosity meanders, probably reflecting ice-marginal control on the channel position. However, the outer meander bends are deeply scoured and are filled by elongate lakes, including Florida Lake and Parker Lake. These channels also crosscut other abandoned channels or are in places dissected by modern rivers, which flow normal to the local slope. Valley-long profiles are undulating with uphill sections. The valley flanks are extensively draped by stagnant-ice moraine, including deformed doughnut moraine in which the circular rims have been attenuated downslope, reflecting the downwasting of supraglacial debris into the channel. These valleys are similarly interpreted as glacial meltwater channels; however, their supraglacial sediment infill demonstrates that they occupied subglacial positions prior to abandonment. These channels, therefore, are unlikely to have originated solely from incision by proglacial meltwater systems following deglaciation. Although a polyphase origin cannot be discounted, these channels’ characteristics suggest that this type of channel was eroded by subglacial meltwater driven by the hydrostatic gradient of the overlying ice sheet (cf., Ó Cofaigh, 1996).

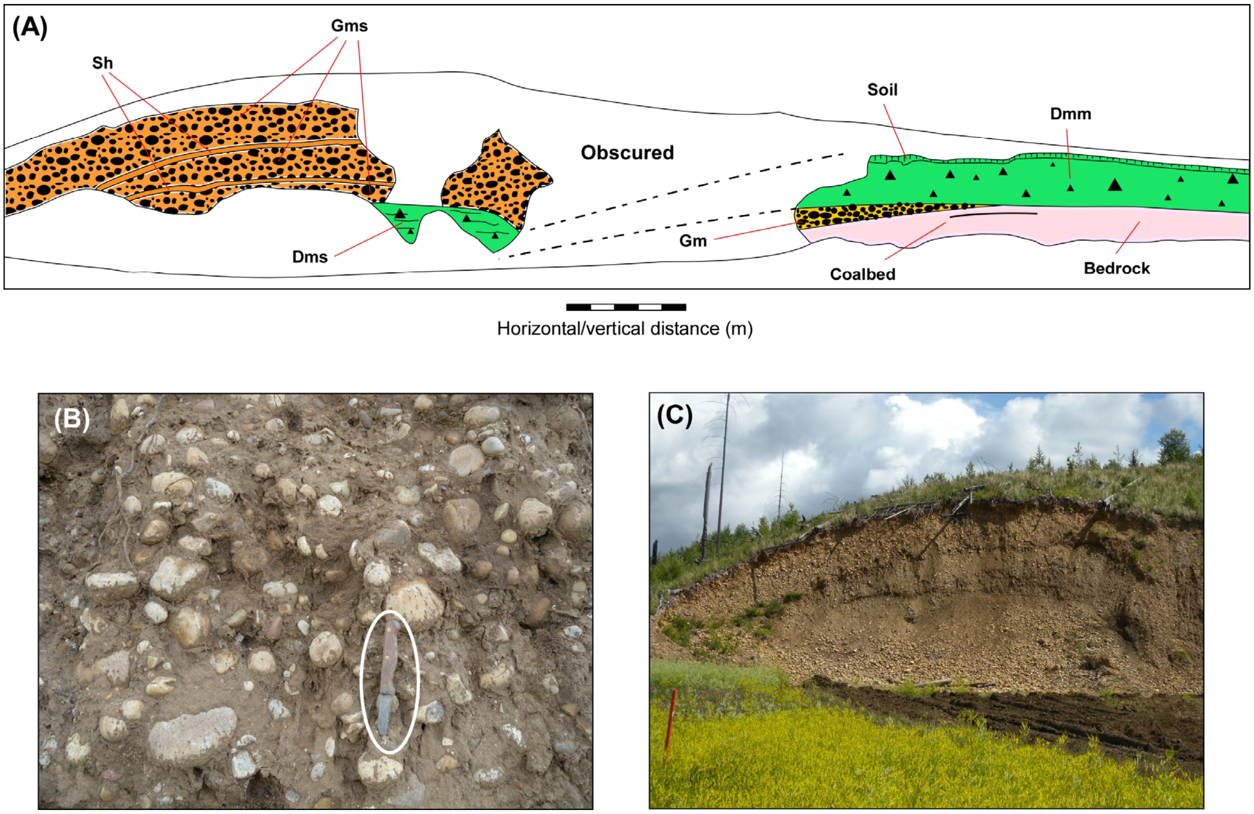


Figure 20. Ice-contact deposits at site SP09-121: a) section diagram through a kame showing bedrock overlain by till and ice-contact fluvial deposits; b) internal composition of the kame, dominated by matrix-supported gravels (knife highlighted for scale); c) downwarping of beds along the northern flank of the kame.

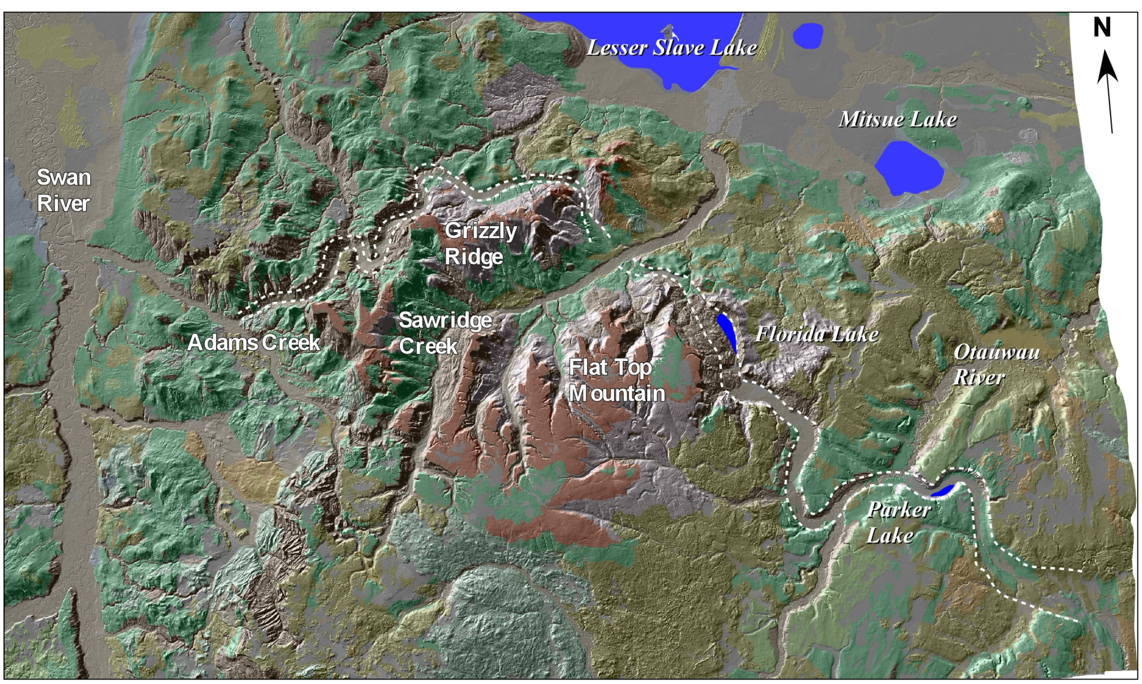


Figure 21. Light detection and ranging (LiDAR) image of a major meltwater channel dissecting the uplands of the Swan Hills and Flat Top Mountain.

5.5 Glaciolacustrine Deposits (Units LG, LGL)

Glaciolacustrine deposits (unit LG) typically blanket the underlying units, forming narrow plains fringing the shore of Lesser Slave Lake. Low to moderately sloping moraine on the west side of the Swan River valley is onlapped by glaciolacustrine sediments to 615 m asl. These sediments overlie till along a sharp, horizontal boundary and consist of massive to thickly laminated, rhythmically bedded couplets of silty clay and fine sandy silt with occasional dropstones (Flv; Figure 22). Individual sandy-silt–clay couplets vary in thickness from 0.5 to 1 cm. Clay laminae range from 0.1 to 0.5 cm thick, whereas the thickness of the sandy-silt component of the couplet is more variable, ranging from 0.5 to 1 cm. The clay laminae overlie the silt laminae with either a gradational or sharp contact. The clay laminations also contain bundles of very thin silt laminae, demonstrating variable energy conditions and/or sediment supply during deposition.

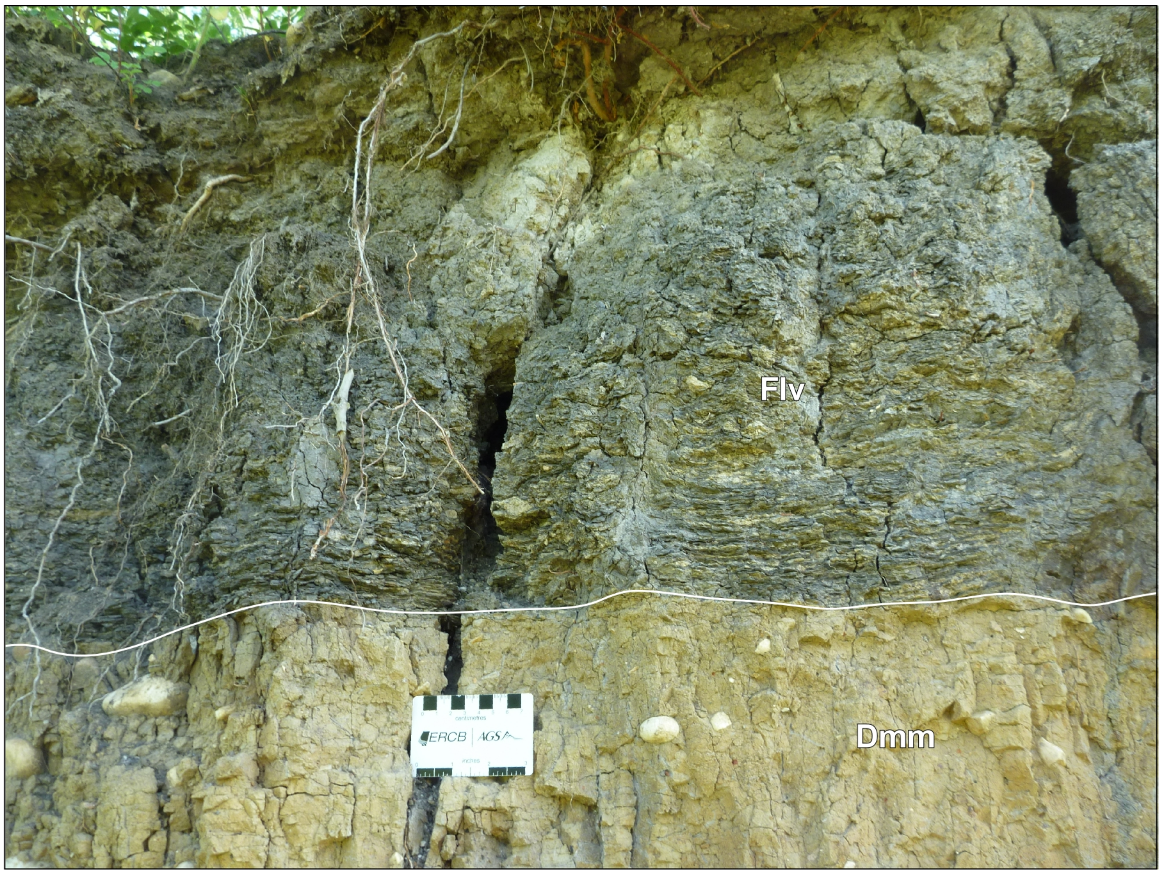


Figure 22. Rhythmically bedded glaciolacustrine sediment (Flv) overlying till (Dmm) along a sharp, horizontal contact.

Glaciolacustrine deposits mapped between Faust and Kinuso exhibit parallel tracts of subtle flutings. These low-amplitude flutes are oriented ~210°, parallel with the fluted tracts that occur on moraine plain immediately to the south.

More localized glaciolacustrine deposits occur across mid-level slopes of upland areas. Towards the headwaters of the Driftpile River on the northern flank of House Mountain, glaciolacustrine sediment mantles till up to 855 m asl (Figure 5). In the northeast, glaciolacustrine sediment onlaps till on the southern flank of the Pelican Mountains up to 630 m asl where it is associated with a glaciofluvial fan delta. Much of this unit has a hummocky morphology, likely due to melting of stranded ice blocks and differential melting and collapse of massive ground-ice inclusions (cf., Dyke et al., 1992), and frequently adjoins areas of doughnut or stagnation moraine (cf., Evans, 2003).

Littoral and nearshore sediments (unit LGL) occur around Lesser Slave Lake. These consist of fine- to medium-grained, well-sorted sands and pebbly sands that were deposited along former shorelines fringing a progressively diminishing glacial lake. Along the northeast margin of the lake, these sediments consist only of a cobbly, wave-washed veneer that was winnowed from the underlying till. These wave-washed surfaces are associated with well-developed, wave-cut benches that were preserved following glacial lake recession.

5.6 Lacustrine Deposits (Unit L)

Lacustrine deposits are a minor unit in the southern Lesser Slave Lake area, comprising fine-grained, crudely laminated silty sands around the margins of Lesser Slave Lake (Figure 5). These sediments were deposited by postglacial littoral processes, producing a suite of landforms that include low-amplitude foreshore berms, nearshore bars, and spits.

5.7 Eolian Deposits (Unit E)

Eolian deposits are distributed extensively across the Lesser Slave Lowland between the Athabasca and Saulteaux river valleys and on the backshore zone abutting the eastern margin of Lesser Slave Lake (Figure 5). The dunes in the Lesser Slave Lowland consist of well-sorted, medium-grained sand and form a system of parabolic and compound dunes that have long trailing arms. The distinctive morphology of these dunes arises when vegetation stabilizes part of the dune during migration.

Linear dunes occur immediately to the east of Lesser Slave Lake near Devonshire Beach and in areas of Lesser Slave Lake Provincial Park. These dunes have several areas of blowouts and are associated with the backshore zone of the modern beach. Two further sets of linear dunes occur to the east of the modern beach and mark former beach positions during the recession of Lesser Slave Lake through the Holocene.

5.8 Fluvial Deposits (Unit F)

Alluvium is widespread across the map area and occurs in several depositional settings, including in a number of alluvial fans and plains that emerge from the Pelican Mountains and spread out across the low-relief glaciolacustrine plain to the east of Lesser Slave Lake. These sediments consist of laminated to crudely bedded, brownish silty sand with a moderate to high calcium carbonate content. The southern margin of Lesser Slave Lake is fringed by a number of alluvial plains deposited by rivers emanating from the Swan Hills, together with smaller creeks draining local uplands. The most prominent plains occur at the mouths of the Swan and Driftpile rivers and were deposited by deltaic progradation into the Lesser Slave Lake basin that began during regional deglaciation and continues to present. There are at least 3 m of sediments consisting of brown (10YR 4/3), poorly sorted silt and sandy silt with crude horizontal laminations and occasional organic-rich horizons that were deposited by suspension settling within overbank flows following periodic flooding.

Alluvium is also present in paired terraces along the Athabasca and Lesser Slave rivers. Along the Athabasca River, two paired terraces are present at 555–557 and 560–565 m asl. The lower terrace is separated from the modern flood zone by up to 5 m high bluffs. Significant gravel deposits are associated with these terraces, with the upper terrace being actively quarried along the eastern bank of the Athabasca River near the town of Hondo. These deposits contain quartzite clasts and Canadian Shield–derived rock types that were reworked from Laurentide glacial deposits.

Alluvium occurs along the rivers and creeks in the southern Lesser Slave Lake area, including the Driftpile, Swan, Otauwau, Driftwood, Lesser Slave, and Athabasca rivers. These deposits consist of sand and cobbles within the modern channel and floodplain, which is bisected by meandering channels 10 to 300 m wide. The remaining sand and gravel in these rivers occurs in paired and unpaired terraces along the margins of these channels.

5.9 Colluvial Deposits (Unit C)

Colluvium results from the downslope movements of materials, including till, glaciolacustrine sediment, preglacial gravel, and bedrock, and is a significant surficial unit in the map area, covering ~520 km^2. These movements result from the structural weakness or gravitational movement of the materials, fluvial erosion at the base of the slope leading to mass wasting, and/or are related to surface runoff and groundwater conditions. Colluvium mantles upland slopes and channel margins in a number of locations across the map area, and includes minor slumps and earth flows, and large, deep-seated landslides.

5.10 Organic Deposits (Units O, OB, OF)

Wetlands and associated peat deposits are mainly developed on the poorly drained substrates of fine-grained till and glaciolacustrine sediments. Bogs occur either as small, isolated depressions in upland areas or as more extensive features that overlie glaciolacustrine deposits in low-lying areas surrounding Lesser Slave Lake. Bogs consist of mixed sphagnum and wood peats in areas where slightly improved drainage conditions allow black spruce forest to develop. Fen peat forms only a minor component of this system, with small areas forming in low-lying regions covered with grasses and sedges between peat bogs.

6 Economic and Engineering Geology

6.1 Surficial Deposits

Till excavated from borrow pits is widely used as a construction material for roadbase and well pads, as well as for forestry access roads across uplands south of Slave Lake.

The pre-Quaternary gravels capping the uplands of Flat Top, Marten, and House mountains and Grizzly Ridge represent a potentially large granular aggregate resource. These gravels are dominated by quartzite cobbles and boulders, which may be suitable to be crushed for some industrial and construction applications. Road access to these deposits is limited and the gravels are not currently quarried. In contrast, a number of active quarries use the preglacial fluvial gravels that crop out at lower elevations along the southern shore of Lesser Slave Lake, adjacent to Highway 2. Fluvial terrace gravels are also being quarried along the east bank of the Athabasca River near Hondo.

Eolian sands are extracted in a number of small quarries in the southern Lesser Slave Lake area. These sands are used locally as construction material for well pads and associated access roads that cross bog and fen peats in the Lesser Slave Lowland.

6.2 Geological Hazards

Deep-seated landslides (Unit Cs in Figure 6a) occur on steep slopes fringing Pelican, Flat Top, and House mountains and Grizzly Ridge. They also commonly occur along the margins of major meltwater channels, where overdeepening steepens valley walls sufficiently to promote slope failure. For example, Florida Lake, which occupies a major, overdeepened meltwater channel, is the site of a large, deep-seated landslide that buried the western margin of the channel. These movements are frequently retrogressive and include rotational slides that comprise multiple back-tilted arcuate blocks, as well as local slumps and earth flows.

7 Quaternary History

7.1 Preglacial River Systems

Preglacial fluvial deposits described in this report document the evolution of two Quaternary drainage systems across the southern Lesser Slave Lake area. A third and considerably older drainage system is represented by unconsolidated fluvial gravels overlying the remnants of an east- to northeast-sloping peneplain that formed a basinwide unconformity during the Oligocene to Miocene following the Laramide Orogeny (Leckie, 1989, 2006; Leckie and Cheel, 1989; Leckie et al., 2004).

During uplift and associated changes in base level, eastward-flowing Quaternary river systems progressively incised this peneplain, leaving gravel-capped uplands within a lower-elevation regional planation surface. The mid-level preglacial fluvial deposits reflect drainage across this regional planation surface, with deposition occurring within braid bar and channel systems dominated by gravel sheets produced by the aggradation of extensive longitudinal bars (lithofacies Gm, Gmi) during high energy discharge conditions. Interbedded lenses of horizontally bedded sand (lithofacies Sh) represent only a minor component of this sedimentary succession and reflect the abandonment of secondary channels that modified the bar surfaces during falling discharge conditions. Further incision of the regional planation surface established the High Prairie paleovalley, which extends eastwards across the map area, along the axis of the Lesser Slave Lake basin (Atkinson and Lyster, 2010a). This paleovalley forms part of a larger, regionally integrated paleovalley system that spans north-central Alberta and is now infilled with up to 150 m of stratified and nonstratified sediment (Atkinson and Lyster, 2010b).

The internal architecture of the lower-level preglacial fluvial deposits indicates the superposition of morphologically and texturally distinct internal elements due to the downstream migration of accretionary bedforms across former channel positions. The trough cross-bedded and rhythmically bedded sand and silt with medium gravel interbeds (Sh/Gms; Figure 11) was deposited within distributary channels of an eastward-flowing braided river system during low-flow regimes. The truncation of this channel sandbody by massive, matrix-supported cobble to boulder gravel (Gms) documents the downstream accretion of a longitudinal bar during the upper-flow regime of a channel flood across the braidplain. This gravel facies represents progradation within diffuse sheets and highlights the lateral mobility of high-energy bedload channels within braided rivers systems (Hein and Walker, 1977; Miall, 1985; Stokes and Mather, 2000). Such mobility is emphasized by soft-sediment deformation that was probably caused by the rapid loading of saturated sand by the overlying gravel facies.

Along the southern flank of the Pelican Mountains, channel sands similarly exhibit evidence of soft sediment deformation, although in this setting, they are overlain by a silty, structurally disorganized diamicton that is interpreted as a mass flow facies, with unordered clast orientation resulting from a plug-like cohesive debris flow process (Nemec and Steel, 1984). Overlying matrix-supported gravel units exhibiting some vertical grading are also likely to reflect a mass flow origin, although they formed through a more turbulent and less fluid, noncohesive mass flow process (Nemec and Steel, 1984). This fluvial style may reflect the proximity of the site to the steep slopes of the Pelican Mountains immediately to the north. Although the age of the lower-level fluvial sediments is uncertain, they lie within the same stratigraphic and topographic position as a regionally extensive gravel sheet mapped along the northern flank of the Peace River, known in Alberta as the “Grimshaw gravels” (Chlachula and Leslie, 1998; Atkinson and Paulen, 2010) and in British Columbia as the “high planation surface” (Hartman and Clague, 2008). Consequently, these gravels are tentatively assigned a Neogene to early Quaternary age since they were deposited prior to the incision of the Peace River Lowlands and the establishment of the ancestral Peace River and High Prairie paleovalleys.

The distribution of fluvial deposits surrounding the lower reaches of the Swan and Driftwood river valleys relates to a younger aggradational phase, which is attributed to local fluvial systems that drained the Swan Hills and Pelican Mountains. Although the age of these sediments is uncertain, they are graded to a base level that is close to the modern drainage and are therefore assigned to the late Quaternary.

7.2 Glacial History, Ice Sheet Dynamics, and Till Genesis

The three till lithofacies described in this report are interpreted as the product of the advection and vertical accretion of sediment associated with a late Wisconsinan advance of the Laurentide Ice Sheet. Collectively, this interpretation accords with the glacial stratigraphy described elsewhere in the Peace River Lowland (Liverman et al., 1989; Catto et al., 1996; Atkinson, 2009; Atkinson and Paulen, 2010; Morgan et al., 2012). Till deposition commenced during the advance of the Laurentide Ice Sheet across the region at ca. 22 ka BP (Dyke, 2004). This till represents an amalgam of Cretaceous bedrock and pre-existing surficial materials, primarily glaciolacustrine sediments. These glaciolacustrine sediments subcrop throughout the Peace River Lowland and were deposited within a regionally extensive ice-marginal lake impounded by the Laurentide Ice Sheet as it advanced southwards (up drainage), blocking the Shaftesbury, Notekiwin and High Prairie paleovalleys (Hartman and Clague, 2008; Atkinson and Lyster, 2010a; Atkinson and Paulen, 2010). As the ice sheet advanced across the Lesser Slave Lake basin, overridden glaciolacustrine sediments would have been advected downflow by subglacial deformation, contributing to the clayey-silt till facies (Dmm[s]) described in this report. The diffuse base of this clayey-silt till documents the incorporation and amalgamation of preglacial sand and gravel within the shear zone of the overriding subglacial deforming layer.

Subglacial shear is typically associated with an increase in porewater pressure, resulting in sediment expansion and a lowering of the cohesive and frictional strength of the till (Alley, 1991). Such dilation inhibits the formation of a coherent till-matrix framework, which reduces ice overburden pressure and facilitates further till dilation, enhancing ice flow (Iverson et al., 2003). This relationship between till deformation and ice flow is reflected in the sediment-landform associations south of Lesser Slave Lake and around the eastern flank of House Mountain, where tracts of fluted moraine occur primarily in areas of clay-rich till.

The overlying sandy-silt till (Dmm) exhibits a lower density and higher porosity than the clayey-silt till (Dmm[s]). These characteristics suggest that these till lithofacies have a different provenance, likely due to the progressive downflow exhaustion of glaciolacustrine sediment from the Lesser Slave Lake basin. This exhaustion is recorded by clay bands at the base of sandy-silt till at NA09-68, which relate to the attenuation and amalgamation of glaciolacustrine sediment within the deforming layer of the overriding sandy-silt till (cf., Benn and Evans, 1996). Due to textural differences, the sandy-silt till, which is interpreted as part of a regionally dispersed till lithofacies (cf., Fenton, 2008; Atkinson, 2009; Atkinson and Paulen, 2010), likely dilated more rapidly and was less likely to experience collapse and solidification than the underlying clayey-silt till (Evans et al., 2006). Consequently, the intertill horizon of quartzite boulders exposed in the gravel pit adjacent to the mouth of the Driftpile River (Figure 16) may represent a boulder pavement along the transition between these two rheologically distinctive deformation tills. In this case, clasts within the sandy-silt till may have sunk through a weak deforming layer until they reached a level where the till matrix could support their weight, which was at the top of the clayey-silt till (cf., Clark, 1991).

Sediment-landform associations across upland areas surrounding Flat Top Mountain demonstrate that hummocky moraine occurs primarily in areas of sandy-silt till. The limited thickness of this till lithofacies across these uplands may be due to the shear strength of the underlying coarse-grained material, which is more resistant to deformation and could have impeded the efficient transportation and deposition of subglacial material (Boulton, 1996; Boulton and Dobbie, 1998). Alternatively, these till thickness variations may be because higher parts of the glacier bed extended into cleaner ice or till mantling these uplands experienced enhanced erosion (cf., Dyke et al., 1992). Nevertheless, the widespread distribution of bedrock outcrops, particularly around the fringes of Flat Top Mountain and Grizzly Ridge, indicate that subglacial deposition by the Laurentide Ice Sheet was limited across these uplands.

Ice-thrust moraine on the slopes flanking the Swan River evidently formed during initial ice advance and was subsequently overridden, as indicated by the moraine's undulating morphology and the occurrence of glaciotectonically deformed bedrock strata. Flutings emanating from this undulating ice-thrust moraine likely relate to the downflow grooving of the substrate by material scoured from the moraine (cf., Tulaczyk et al., 2001; Evans et al., 2008). This process would have been augmented by the convergence of southward-flowing ice around the flanks of House Mountain and Grizzly Ridge and its subsequent acceleration along the Swan River valley. The suite of sharp-crested, concentric ice-thrust moraines fringing the northwest slopes of House Mountain and Grizzly Ridge are inferred to have been deposited during ice recession and may be related to the surge of a local lobe of Laurentide ice southwards along the Swan River valley (cf., Evans et al., 1999).

7.3 Deglaciation and Glacial Lake Peace

Retreat of the Laurentide Ice Sheet from the Lesser Slave Lake area at ca. 12 ka BP (St-Onge, 1972; Dyke et al., 2003; Squires et al., 2006) is recorded by a widespread suite of ice-marginal and proglacial sediment-landform associations. In the southern Lesser Slave Lake area, this retreat is evident by the extent of ice-stagnation sediments and landforms, including large areas covered by well-developed doughnut moraine. The genesis and glaciological significance of doughnut moraine is still debated. However, most hypotheses invoke the melting of stagnant ice, with the landform forming either by topographic inversion due to the slumping of supraglacial debris into holes or ice-walled lakes within stagnant ice (Clayton, 1967; Parizek, 1969; Gravenor and Kupsch, 1959; Johnson et al., 1995; Johnson and Clayton, 2003; Clayton et al., 2008) or the pressing of dead-ice blocks into the plastic substrate (Hoppe, 1952; Stalker, 1960; 1973; Eyles et al., 1999; Boone and Eyles, 2001).

An alternative interpretation, which does not necessarily associate the landform with ice-marginal stagnation, proposes that some doughnut moraines originated from proglacial diapiric extrusion of sediment in response to the injection of overpressured subglacial groundwater beyond the ice margin (Mollard, 1983, 1984; Boulton and Caban, 1995).

The association of doughnut moraine with kames and eskers supports the interpretation that deglaciation was characterized by widespread ice stagnation. Ice-stagnation landforms deposited across upland areas likely reflect the melting of residual ice masses that would have persisted following regional deglaciation. The kames formed in locations where glaciofluvial sediments were reworked by supraglacial mass flows. Downwarped bedding observed in one of the landforms indicates that deposition occurred in an ice-contact position, since the sediments evidently collapsed after they lost buttressing ice support during melting. The association between eskers and areas of doughnut moraine suggests that the eskers formed when supraglacial meltwater breached sub/englacial conduits or formed supraglacial channels. Sediments within these drainage pathways were unlikely to be preserved in areas where the ice sheet actively retreated. Their association with doughnut moraine, therefore, supports an ice-stagnation origin for the landform assemblage. The absence of well-developed glaciofluvial fans at the termini of the eskers also indicates that they represented minor meltwater outlets, which were probably short lived.

At lower elevations, stagnant-ice moraine is widely associated with reticular to crudely diamond-shaped ridge networks, with the intervening basins filled with glaciolacustrine sediments that overlie till. The ridge networks likely reflect a structural control on deposition through crevasse and fracture patterns within the ice, with ridge material being derived from supraglacial sources, or through the subglacial squeezing of water-saturated till into basal ice fractures (Gravenor and Kupsch, 1959). The transition of ridge networks upslope into till-dominated, stagnant-ice moraine and downslope into more extensive glaciolacustrine deposits indicates that these deglacial landforms were associated with thin glacier ice in the shallow shore zone of a glacial lake, which periodically flooded the ice surface. During subsequent breakup, ice blocks would become stranded, sinking into the soft clay substrate (Mollard, 2000).

As the Laurentide Ice Sheet retreated northeastwards and down drainage from the southern Lesser Slave Lake area, it exposed a broad lake basin and adjoining lowlands. Subaerial and glacial meltwater drainage systems that converged towards this ice-dammed basin formed a proglacial lake, which is documented by widespread glaciolacustrine sediments across the study area. These glaciolacustrine sediments form part of a larger expanse of proglacial lake sediments deposited within Glacial Lake Peace at the end of the last glaciation (Mathews, 1980; Paulen, 2004a, b; Fenton, 2008; Atkinson, 2008; Atkinson and Paulen, 2010).

Glacial Lake Peace was a large, time-transgressive proglacial lake that submerged the Peace River Lowland, from British Columbia in the west (Mathews, 1980) to High Level in the north (Plouffe et al., 2007). This lake occupied successive proglacial positions as the retreating Laurentide Ice Sheet exposed progressively lower outlets to the north and east. Previous paleoshoreline studies have suggested that the Lesser Slave Lake area was inundated from the late Clayhurst to the Indian Creek stages of Glacial Lake Peace (Mathews, 1980). During these stages, Glacial Lake Peace extended up to 300 km north across the Peace River Lowland and 175 km west of the map area (NTS 83O/S).

The highest glaciolacustrine sediments described in the southern Lesser Slave Lake area (855 m asl) occur along the Driftpile River. These are interpreted to have been deposited when a lobe of Laurentide ice retreated northwards along the Driftpile River valley, impounding local drainage and forming a small, short-lived proglacial lake. As the margin of the Laurentide Ice Sheet continued to retreat northwards from the Swan Hills Upland, the larger expanse of the Lesser Slave Lake basin became inundated by Glacial Lake Peace. Deposits associated with Glacial Lake Peace in the southern Lesser Slave Lake area document the suspension settling of fine-grained sediment within a low-energy, glaciolacustrine environment. Rhythmically bedded clays showing sandy-silt–clay couplets are widespread and reflect sedimentation within deeper parts of the lake basin, where slight variations in energy conditions and/or sediment supply would have resulted in the deposition of the silt-clay couplets. A regionally well-developed washing limit on the slope separating the Driftpile Benchland from the Sturgeon Plain records a maximum Glacial Lake Peace shoreline of 610 m asl. This paleoshoreline correlates with the early Clayhurst stage of Mathews (1980).

Previous research has demonstrated that during the recession of Glacial Lake Peace, the regional retreat of the Laurentide Ice Sheet was punctuated by local readvances (Henderson, 1959; Paulen, 2004a, b; Fenton, 2008; Atkinson, 2008; Atkinson and Paulen, 2010). Evidence of a southward readvance of a Laurentide ice lobe across glaciolacustrine sediments has been reported immediately to the west in the High Prairie map area (Atkinson, 2008). Consequently, the fluted and glaciotectonized glaciolacustrine sediments described in this report are inferred to relate to a more widespread regional readvance that extended at least 40 km farther south, covering the High Prairie Plain and the northern Sturgeon Plain. This regional readvance also likely explains the occurrence of till plains at low elevations across these plains, indicating that they lay inside the ice margin during this lake stage. Likewise, the general absence of proglacial lake sediments and widespread occurrence of low-elevation surface till and stagnant-ice moraine east of the town of Slave Lake suggests that this area lay inside the ice margin during the final Glacial Lake Peace stage, when the eastern limit of a contiguous Glacial Lake Peace was situated close to Mitsue Lake. As Laurentide ice finally retreated off the Lesser Slave Plain, northwards onto the Pelican Mountains, the Lesser Slave River valley was exposed, opening an outlet along the Lesser Slave Plain to the Athabasca River.

7.4 Postglacial History

Following the local drainage of Glacial Lake Peace and final retreat of Laurentide ice, fine-grained alluvium aggraded across the Sturgeon, Fawcett, and Lesser Slave plains, mantling the glaciolacustrine sediment. The thickness and extent of the inactive alluvium across these plains suggests that much of it accumulated during the early Holocene, when rivers and creeks would have been characterized by significantly higher sedimentation rates because of snowmelt from adjacent uplands, melting permafrost and meltwater from residual ice masses occupying upland plateaus after regional deglaciation. Recent fluvial processes are responsible for the ongoing aggradation and incision of sand and gravel along rivers draining the Pelican Mountains and Swan Hills Upland. A small amount of incision along the Lesser Slave River has left a paired terrace perched above the modern channel. Eolian processes have left a significant imprint on the southern Lesser Slave Lake landscape, with extensive dunes and sand sheets occurring across the Lesser Slave Plain. Minor eolian deposits also occur on the Sturgeon Plain. The orientation of the dune crests across the Lesser Slave Plain indicates that paleowinds from the northwest deposited the dunes, derived presumably from sand from the exposed lake basin.

An optically stimulated luminescence date on a dune at Hondo produced an age of 14.2 ± 0.7 ka, placing their accumulation within analytical uncertainties of deglacial ages in the region (Wolfe et al., 2004). These dunes form part of a system of eolian deposits in Alberta that accumulated when strong northwesterly katabatic winds were funnelled southeastwards between the retreating margins of the Laurentide and Cordilleran ice sheets (Wolfe et al., 2004; Wolfe et al., 2007). The long trailing arms of the dunes also suggest that they developed when some surface vegetation was present. Minor eolian deposits on the Sturgeon Plain, close to the mouth of the Swan River valley, are also considered to have accumulated at this time, although their southwest alignment suggests that paleowinds would have been funnelled down the Swan River valley, potentially due to the katabatic effect of a residual ice mass on the Swan Hills Upland.

During the Holocene, extensive wetlands developed across the southern Lesser Slave Lake area, particularly across the lowlands of the Lesser Slave Plain. These wetlands mostly consist of bog peatlands with minor areas of fen that accumulated after the deposition of the dune sands and continue to accumulate through to the present day. These wetlands are ecologically important and play a role in recharging groundwater systems.

The postglacial evolution of the southern Lesser Slave Lake area has been further influenced by slope failures and mass movements. Many of these slope failures likely initiated in the early Holocene, representing the paraglacial adjustment of the landscape and degradation of oversteepened slopes following deglaciation.

8 Summary

Preglacial fluvial deposits identified in the southern Lesser Slave Lake area document the evolution of Quaternary braided river systems draining eastwards across the remnants of an east- to northeast-sloping peneplain that formed a basinwide unconformity following the Laramide Orogeny (Leckie, 1989, 2006; Leckie and Cheel, 1989; Leckie et al., 2004).

Widespread deposits of till and associated streamlined glacial landforms document the southward advance of Laurentide ice across the southern Lesser Slave Lake area. A succession of up to three till facies, in places overlying preglacial fluvial deposits, indicates that the Lesser Slave basin and the surrounding uplands of the Swan Hills and Pelican Mountains were inundated by the Laurentide Ice Sheet during late Wisconsinan glaciation (cf., Liverman et al., 1989; Catto et al., 1996; Atkinson, 2009; Atkinson and Paulen, 2010; Morgan et al., 2012).

The Laurentide Ice Sheet retreated from the southern Lesser Slave Lake area at ca. 12 ka BP (St-Onge, 1972; Dyke et al., 2003; Squires et al., 2006). This retreat is recorded by a widespread suite of ice-marginal and proglacial sediment-landform associations, including large areas of well-developed doughnut moraine. The association of doughnut moraine with glaciofluvial landforms supports the interpretation that deglaciation was characterized by widespread ice stagnation, partly related to the melting of residual ice masses that persisted after regional deglaciation.

As the Laurentide Ice Sheet retreated northeastwards and down drainage from the southern Lesser Slave Lake area, it exposed a broad lake basin and adjoining lowlands. Subaerial and glacial meltwater drainage systems converged towards this ice-blocked basin, forming a proglacial lake. Associated glaciolacustrine sediments form part of a larger expanse of proglacial lake sediments, which were deposited within Glacial Lake Peace at the end of the last glaciation.

Following the retreat of the Laurentide Ice Sheet from the region, the postglacial evolution of the Lesser Slave Lake area has been influenced by the accumulation of fine-grained alluvium, eolian sediment, and extensive peatlands, which mantle the glaciolacustrine deposits, particularly across the lowlands of the Lesser Slave Plain. Upland areas have been further influenced by slope failures and mass movements associated with deep-seated landslides that fringe the steep slopes of the Pelican, Flat Top, and House mountains and Grizzly Ridge.

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