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Geomorphology of the Tawatinaw Region

by

Elizabeth Claire Sjogren ©

A thesis submitted to the Faculty of Graduate Studies and Research in partial fulfilment of the requirements for the degree of Master of Science

Department of Earth and Atmospheric Sciences

Edmonton, Alberta

Fall, 1999

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled Geomorphology of the Tawatinaw Region submitted by Elizabeth Claire Sjogren in partial fulfilment of the requirements for the degree of Master of Science.

John Shaw

John Shaw (Supervisor)

R B Rains

R. Bruce Rains

Faye E. Hicks

Faye E. Hicks

20 August, 1999

Date

Abstract

The Tawatinaw valley, located approximately 100 kilometres north of Edmonton, operated as a tunnel channel/valley during the last glaciation. Its steep sides, rising long profile, sinuosity, hanging tributary valley, occupation by an underfit stream, and termination in a fan-like complex are all characteristics common to tunnel valleys.

The valley is dominated by sand; rhythmically bedded clay is present at the northern end. The valley opens up at its southern end into a fan-shaped deposit of glaciofluvial sand and gravel. Palaeocurrent data show water flow southward through the valley; profiles show that it rises to the south by approximately 100 metres. Flow in a southerly, uphill, direction provides indisputable evidence that subglacial meltwater flowed through the valley.

The following sequence is proposed: 1) the valley operated as a preglacial river, 2) water flowing in the Eastern tract of the Livingstone Lake megaflood event created the Athabasca fluting field and deposited massive sands high up the valley sides, 3) as the megaflood waned, meltwater occupied only the valley; the glaciofluvial complex formed, and 4) during ice sheet retreat, proglacial meltwater ponded in the deeper, northern end of the valley.

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I would first like to thank John Shaw, my supervisor, for accepting me as a graduate student and providing me the opportunity to pursue this dream. Thank you, John, for giving me the freedom to work on this project at my own pace and for allowing me to make it my own. The last few years have not been pleasant politics-wise, especially for you, and it would have been so easy simply to walk away - your unwavering belief in yourself and your work is amazing.

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CHAPTER 1: INTRODUCTION

1.0 Introduction

This chapter serves as an introduction to the project. It identifies the problem to be researched, and defines the objectives. This chapter also provides background information about the location of the study, regional setting, surficial deposits and features, and the bedrock geology.

1.1 Problem Statement

It has been proposed that north-central Alberta was overridden at least twice by continental ice sheets (e.g., St-Onge 1972; Roed 1975). The stratigraphy of till overlying "preglacial" gravels of the Empress Formation, and late Wisconsinan dates from the gravels, support Young et al.'s (1994) conclusion that pre-late Wisconsin glaciations did not reach the Edmonton area. Liverman et al. (1989) similarly concluded that only one glaciation, the late Wisconsinan, could be identified in the Peace River area. These findings are important to studies of glacial landforms in this area: if the hypotheses of Young et al. (1994) and Liverman et al. (1989) are correct, landforms in central Alberta reflect glacial dynamics and hydrology of the late Wisconsinan glaciation only.

Shaw et al. (1989), Rains et al. (1993), Sjogren and Rains (1995), Shaw et al. (1996) and Sjogren (1999) implicated catastrophic subglacial flooding events to explain the origins of landforms in parts of Alberta. This is in contrast to more conventional ideas which state that most glacial landforms were created by the direct action of ice (e.g., Gravenor and Meneley 1958, Gravenor and Kupsch 1959, Stalker 1960, Parizek 1969). The postulated main flood is also thought to have been responsible for large tracts of scoured bedrock. The flood is referred to as the "Livingstone Lake Event", and the study area lies within a main pathway of the proposed flood.

Tunnel valleys have been interpreted as products of ice-marginal fluvial erosion (e.g., Boulton and Hindmarsh 1987; Mooers 1989; Wingfield 1990). Alternatively, tunnel valleys are considered to be channels cut by catastrophic subglacial meltwater floods (e.g., Wright 1973; Shaw 1983; Shaw et al. 1989; Ehlers and Linke 1989; Brennand and Sharpe 1993; Piotrowski 1994; Brennand and Shaw 1994). Most theories agree that tunnel valleys are subglacial landforms, carved as meltwater flows beneath the glacier and carries sediment away. The challenge to the field scientist is to determine, on the evidence of geomorphology and sedimentology, the probable location of fluvial erosion relative to the ice front. It is obviously desirable to describe the method of formation of a landform, and to associate this process with the dynamics of the Laurentide Ice Sheet.

The primary problem to be addressed by this thesis is whether the Tawatinaw Valley formed as a tunnel valley. Also, the possible methods and timing of formation will be addressed. In so doing, the possible relationship between the valley and the "Livingstone Lake Event" will be investigated.

1.2 Description of Study Area

The primary feature of the study area is the Tawatinaw Valley, containing a small, meandering, north-northeasterly flowing stream. The valley was chosen for study because of its size, steep sides and the gravel/sand fan complex to its south - all of which imply vastly different conditions between the present day and the time of valley cutting. I. A. Campbell (pers. comm. 1992) described the valley as a potential tunnel valley, an intriguing hypothesis deemed worthy of investigation.

The study area is bounded by 112°45'W to 113°45'W Longitude and 54°N to 55°N Latitude (Figure 1.1). The southern limit of the study area is approximately 30 kilometres north of Edmonton, and its northern limit is approximately 30 kilometres north of Athabasca. These boundaries were chosen because they not only include the entire Tawatinaw Valley, but they also include much of the surrounding region which is of vital importance when considering the factors that were instrumental in the formation of the valley.

Universal Transverse Mercator coordinates (NAD27) will be used throughout the text when describing the locations of features. Also, in many instances, the locations of the features being discussed will also be described in terms of their location relative to known features such as the towns. Wherever possible, detailed maps will also be provided to aid in locating the features being discussed.

1.3 Regional Setting

Shetsen (1990) mapped the Quaternary geology of central Alberta. The

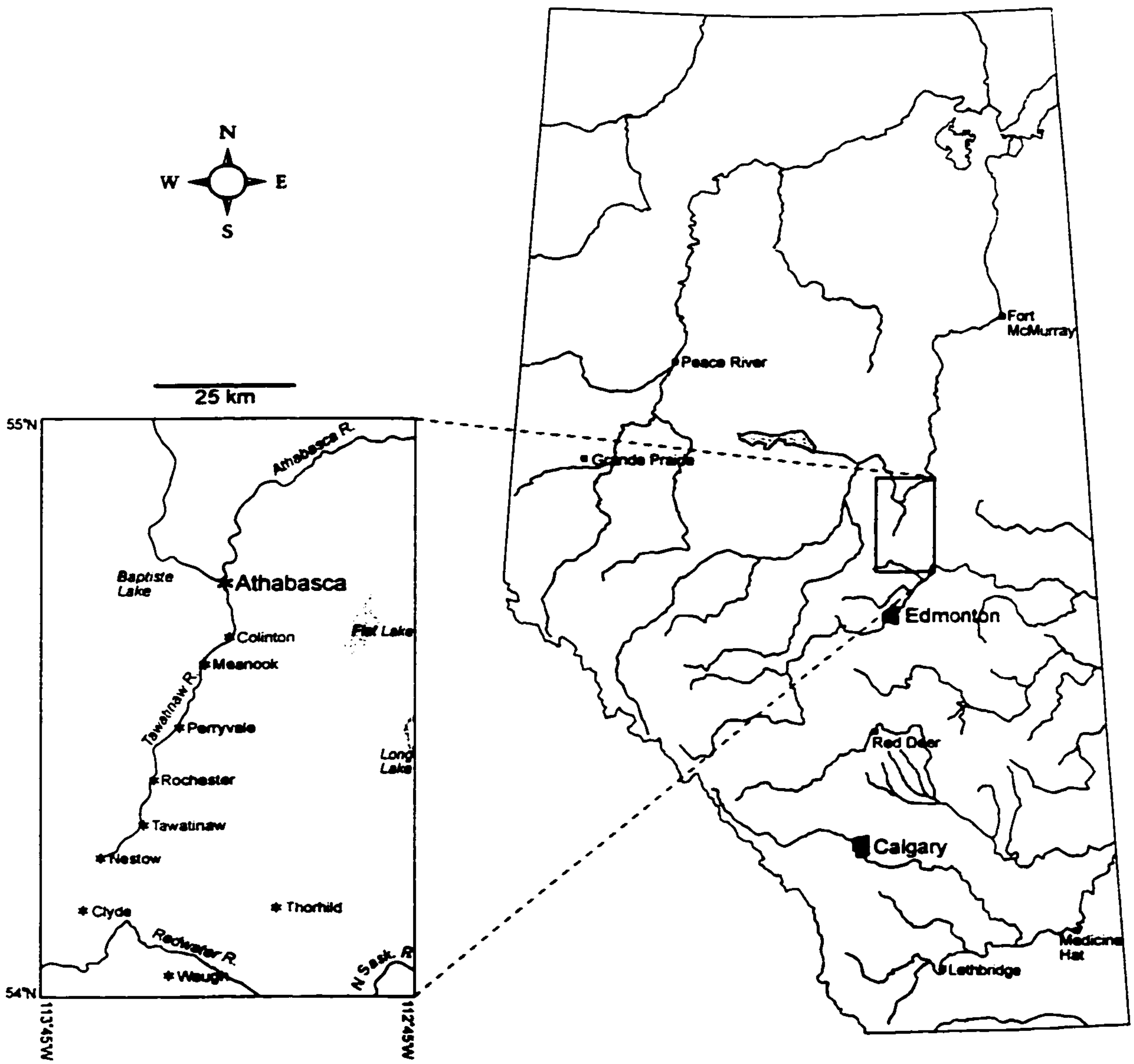


Figure 1.1 Map showing the location of the study area. The Tawatinaw valley is located approximately 100 kilometres north of Edmonton. Towns referred to in the text are labelled in the inset.

Edmonton area is mapped as being covered with lacustrine sediment, associated with Glacial Lake Edmonton (St-Onge 1972). To the west and east of the study area, morainal deposits are mapped. Shetsen (1990) breaks these down into draped and stagnation moraines, both of which are said to contain till and water-sorted material with thicknesses up to 20 metres. These morainal deposits range from flat to undulating to hummocky topography, sometimes conforming to the surface below and sometimes independent of the preglacial surface.

West of the study area, in the Westlock region, numerous NNE-SSW oriented flutes and NW-SE oriented spillways are found (Fraser 1991). Fraser (1991) concluded that the spillways formed parallel to the ice margin, and were thought to be the result of a sudden outburst (or outbursts) of huge volumes of water from a glacial lake. The spillways are steep sided, 200 metres to 2 kilometres across and are generally interconnected; where spillways meet, the channels widen (Fraser 1991). The flutes likely formed prior to the retreat of the ice front and the formation of the spillways.

Also in this region are several examples of ice-thrust moraine that consist of contorted bedrock, till and water-sorted material; this material was translocated by the Laurentide ice in a relatively intact state as thrust blocks or was deformed into folds (Shetsen 1990).

The Athabasca River flows from the west and curves around the northern part of the study area. The northward flowing Pembina River, located immediately west of the study area, is a tributary to the Athabasca River. The North

Saskatchewan River and the Sturgeon River drain eastward and bound the southern extent of the study area. The present-day Tawatinaw River drains northward into the Athabasca River. The Tawatinaw's source area is Helliwell Lake, located at UTM 295135, with additional input from Lebeaus Lake which occasionally drains into Helliwell Lake. The discharge of the Tawatinaw River is low, and the stream is clearly underfit.

Approximately 40 kilometres east of the Tawatinaw system is the White Earth Creek system. Similar to the Tawatinaw system, it consists of an underfit stream within a large meltwater channel which opens up into a broad, fan-like glaciofluvial complex at its southern extent (Figure 1.2). Both the Tawatinaw and White Earth glaciofluvial complexes are oriented generally NW-SE. Based upon their similar appearance, it is speculated that the Tawatinaw and White Earth valleys were formed in a similar manner.

Further to the east, the Lac La Biche region is dominated by fluted terrain. The region has been interpreted as having been overridden by an ice stream (Mougeot 1991, Wilson 1998, Patterson 1994) or marking the flow of a branch of the Livingstone Lake subglacial meltwater flood (Rains et al. 1993, Grant 1997). The obvious cross-cutting nature of two zones of flutes, a relatively long, thin zone (approximately 30 kilometres wide x 200 kilometres long) oriented NW-SE and a wider zone (approximately 100 kilometres by 100 kilometres) oriented NNE-SSW (Shetsen 1990), imply either a two-stage formation of the Lac La Biche flutes or the coalescing of two streams or flow paths.

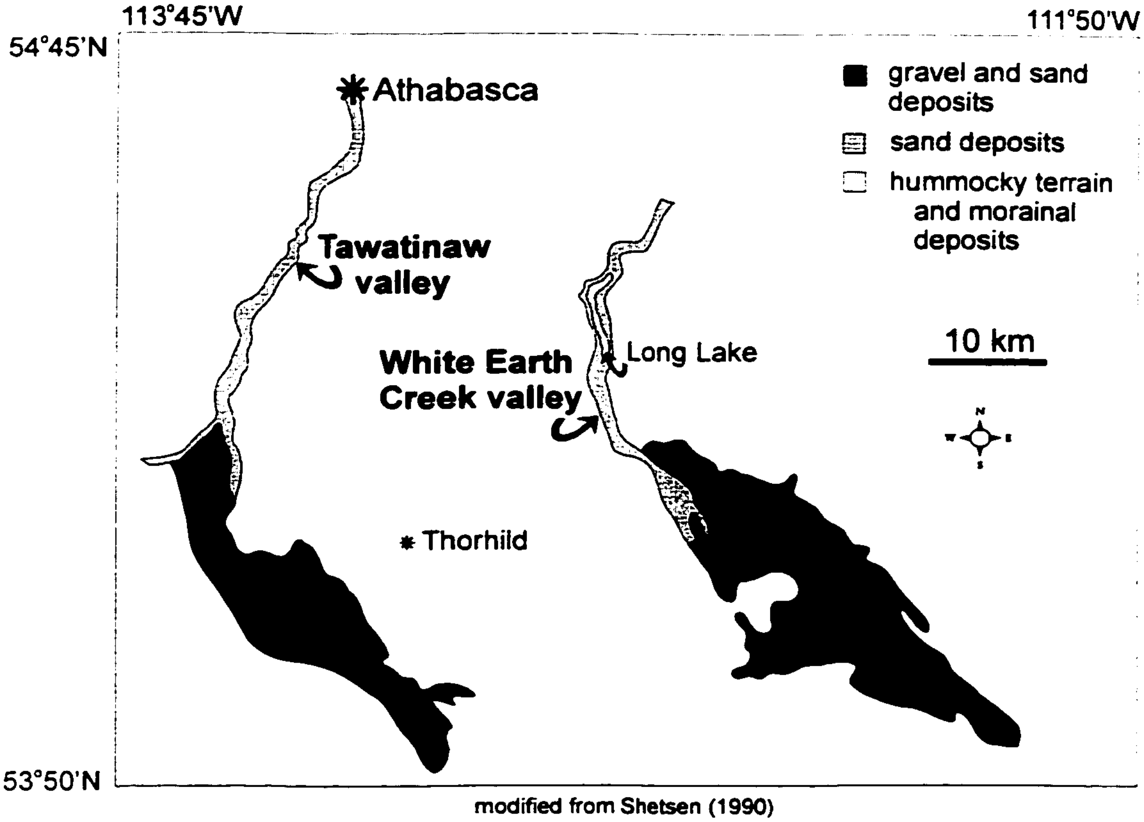


Figure 1.2 Generalized surficial geology map illustrating the similarity of form between the Tawatinaw valley and the White Earth Creek valley to the east. Both valleys are sinuous, steep sided and contain underfit streams. The White Earth Creek valley contains Long Lake at its base. Both valleys are dominated by glaciofluvial sands, and open up at their southern ends into fan-shaped deposits of gravel and sand. Both fan-shaped features deflect towards the southeast. It is speculated that this similarity indicates that the valleys and associated deposits were created by the same mechanism.

Rains et al. (1993) indicated that the region to the east of the study area (between the Tawatinaw and White Earth systems) had been scoured by a southward flowing branch of the Livingstone Lake flood. (Figure 1.3). They describe that the hummocky terrain belts may have been moulded significantly by subglacial meltwater sheets.

1.4 Surficial Deposits

Figure 1.4 illustrates the surficial geology of the study area. Within the Tawatinaw valley, sand is the predominant sediment. At the north end of the valley, glaciolacustrine sand, silt and clay are present, while further south (within the valley) glaciofluvial sand and gravel dominate. At the southern end of the valley, there was a small branch deflecting westward and a large, fan-shaped deposit of sand and gravel spreading southward from the mouth of the valley. Both the smaller westward and larger southward features consist of glaciofluvial sands and gravels. Kjearsgaard (1972) described part of this large glaciofluvial deposit as being spillway deposits. South of Nestow, to the immediate west of the large glaciofluvial complex, is a zone of sand, silt and clay that have possibly undergone aeolian transport. Detailed discussion of the sedimentology of the Tawatinaw valley and the associated glaciofluvial complex will be presented in Chapter 4.

As shown in Figure 1.4, the region surrounding the valley consists of morainal deposits. Shetsen (1990) used the term "stagnation moraine" for these deposits, which implies the process of sedimentation by melt-out. Shetsen (1990)

Figure 1.3 Proposed locations of megaflood paths in Alberta, as described by Rains et al. (1993) and modified from Figure 2, Shaw et al. (1996). Labelled are the western flow path (W), eastern flow path (E), and Lac La Biche flow path (L). The study area is outlined by a rectangle. It is located towards the outer western edge of the main eastern flow path. The hillshaded map of Alberta was created with 30 arc second data provided free of charge by the United States Geological Survey.

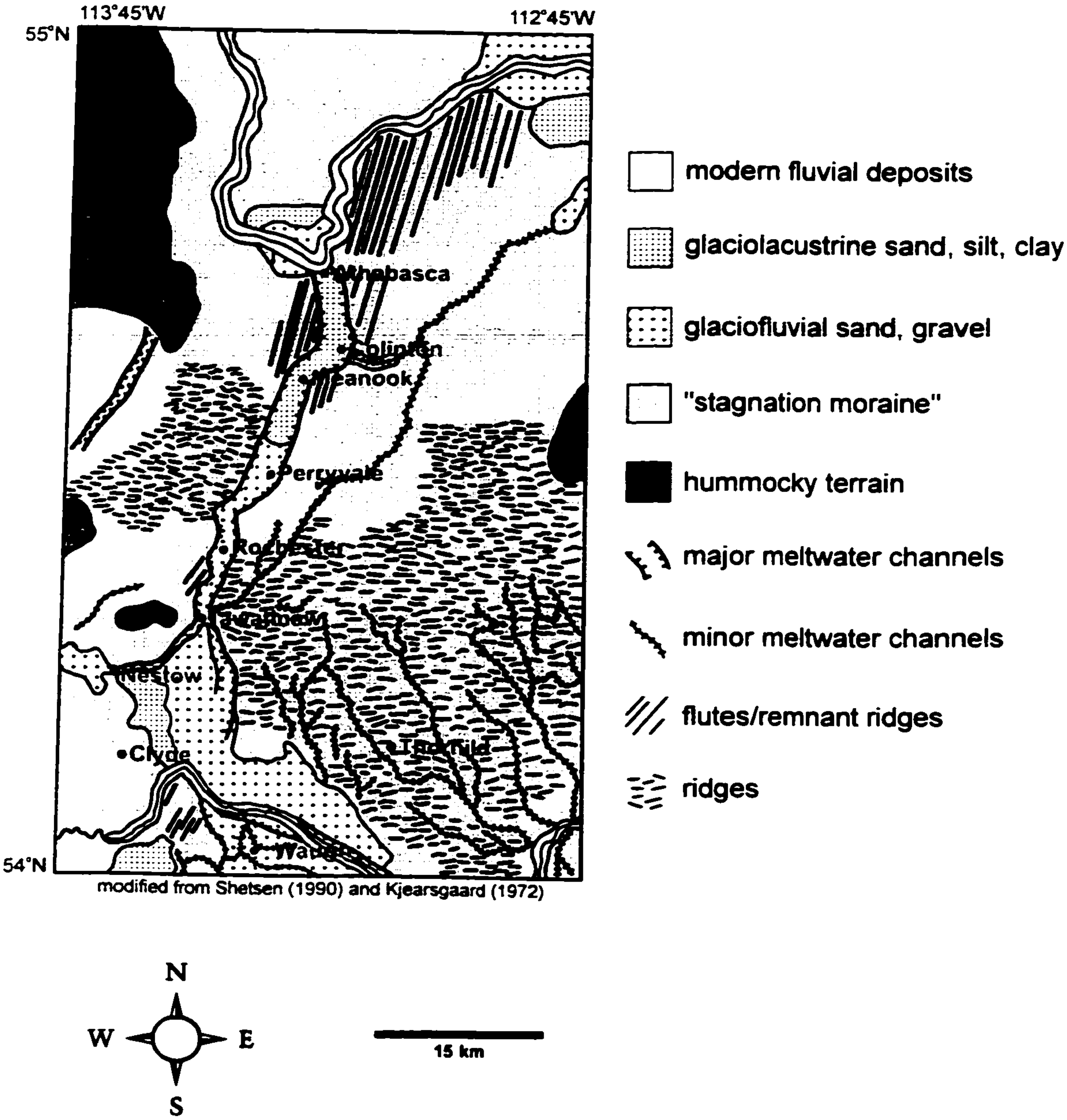


Figure 1.4 Map displaying the surficial geology of the study area. The valley is dominated by glaciofluvial sand and gravel deposits; glaciolacustrine deposits are found at the northern end of the valley. The valley opens up into a fan-like glaciofluvial complex to the south. The higher ground on either side of the valley consists of morainal deposits: sand, gravel, till, with some boulders and exposed bedrock. The Athabasca Flutes are located in the north of the study area. Transverse ridges, minor meltwater channels and zones of hummocky terrain are observed in the area.

defined stagnation moraine as being undulating topography consisting of till, sand and gravel, with some boulders and exposed bedrock. While this definition is consistent with observations to be described in Chapter 4, the term stagnation moraine is being avoided here because of the implication of process; the term ground moraine, which is the descriptor Kjearsgaard (1972) used for these deposits, is preferred.

Zones of hummocky terrain are identified in the region. These hummocks are said to consist of till, gravel and sand, and some bedrock; the local relief can be up to 10 metres (Shetsen 1990). Note again that her term "hummocky moraine" has been avoided and the less genetic term, terrain, has been substituted instead.

1.5 Surficial Features

The Athabasca fluting field is located to the north of the study area. These giant flutes are up to 20 kilometres long, 5 kilometres wide, and 90 metres high (Shaw et al. *in press*), and the field is 15 kilometres wide (Shaw et al. 1996). These ridges and troughs are erosional features, carved into both local sediment and bedrock (Shaw 1994, Shaw et al. 1996). The fluting troughs are parabolic in cross-section and contain some erosion marks; the intervening remnant ridges taper downflow (Shaw et al. *in press*).

1.6 Bedrock Geology

Figure 1.5 illustrates the bedrock geology of the study area. The Tawatinaw

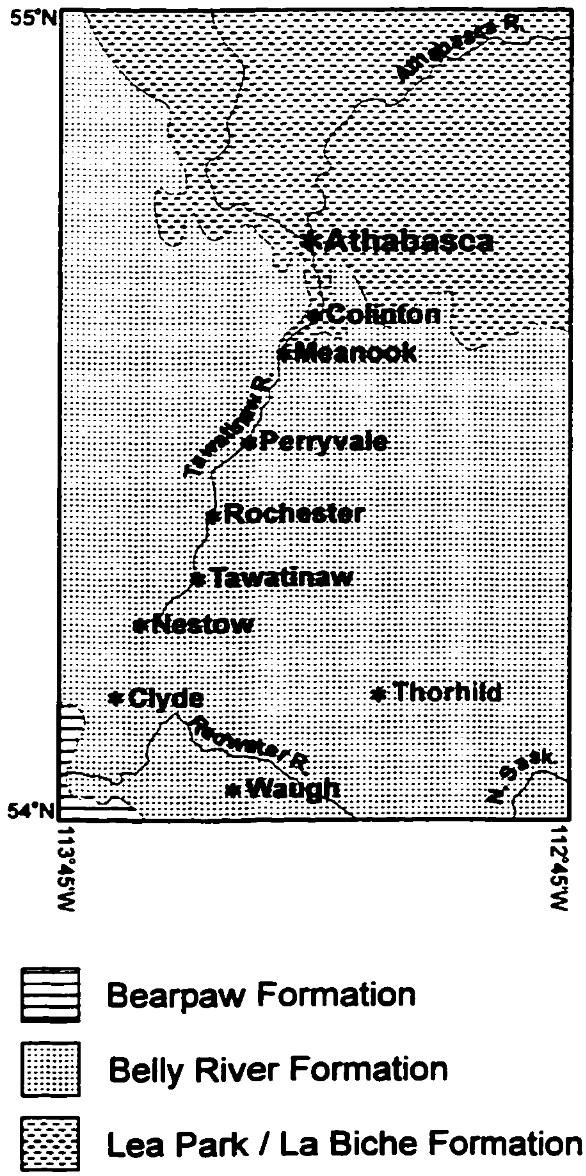


Figure 1.5 Bedrock geology of the study area. The bedrock dips to the southwest, exposing successively younger formations in that direction. The Lea Park Formation (also known as the La Biche Formation) is a dark grey marine shale of Cretaceous age. The Belly River Formation is a green-grey continental bedrock of Cretaceous age that consists of sandstone, shale, scattered coal beds and bentonite. The Bearpaw Formation is a dark grey marine shale that represents a later inundation of the same sea that deposited the Lea Park Formation.

map area is on the eastern limb of the Alberta syncline, otherwise known as the Central Alberta Basin (St-Onge 1972). The contact between the Lea Park and Belly River formations (St-Onge 1972; Jackson et al. 1981) sweeps diagonally across the study area. The bedrock dips gently to the southwest, so that successively younger formations appear in that direction (St-Onge 1972). The Belly River Formation, dominating the central part of the region, is a green-grey continental bedrock of Cretaceous age (Jackson et al. 1981) that consists of sandstone, shale, scattered coal beds and bentonite (St-Onge 1972). The underlying Lea Park Formation, located to the northeast of the study area, is a dark-grey marine shale also of Cretaceous age (Jackson et al. 1981). St-Onge (1972) calls this formation the La Biche Formation, and describes it as consisting of grey and dark-grey marine shales with lighter sandy streaks and a few concretions. In the southwestern corner of the study area, the edge of the Bearpaw Formation is identified. The Bearpaw Formation is a grey and dark grey marine shale which may include some brackish water equivalents (Kjearsgaard 1972). The Bearpaw Formation represents a later inundation of the same sea which deposited the Lea Park Formation.

1.7 Objectives

As stated, the primary objective of this thesis is to determine whether or not the Tawatinaw valley developed as a tunnel valley. By examining the morphology and sedimentology of the valley, the intention is to address all possible mechanisms for formation. Based on preliminary study, it is hypothesised that the Tawatinaw valley is a tunnel valley and that it may be associated with some stage of the Livingstone Lake flood event. The primary objective of this thesis is to test this hypothesis by examining the morphological and sedimentological characteristics of the valley and surrounding region to see of they can be best explained by the flood hypothesis. At the same time, all traditional explanations for tunnel valley formation will be considered.

The nature of the glaciofluvial complex to the south of the valley will also be considered. It is thought that the complex formed as meltwater flowed toward the southeast as it left the constricted tunnel channel (Campbell et al. 1993) or, alternatively, by a stream flowing from the ice mass into a glacial lake (St-Onge 1972). A focus of this thesis will be to investigate the glaciofluvial deposits associated with the valley to determine whether they likely formed subglacially, proglacially, or ice marginally.

CHAPTER 2: LITERATURE REVIEW

2.0 Introduction

This chapter briefly summarizes the literature related to this thesis. It is subdivided into the following subject areas: (i) general glacial hydrology, (ii) glaciofluvial deposits, (iii) tunnel valleys, and (iv) glacial landforms. It is not possible to cite all works pertaining to these subjects, so emphasis is placed on landmark papers and recent publications. The intent of this chapter is to provide a solid background discussion of the current understanding of the nature of tunnel valleys and related glacial processes. The discussion and conclusions later in the thesis are, in part, based on these earlier works.

2.1 General Glacial Hydrology

2.1.1 Sheetflow

Fountain and Walder (1998) suggested that water at the base of a glacier will flow in a nonarborescent network of cavities and channels, and within a layer of permeable sediment. Weertman (1972) stated that water will flow in a thin sheet or film, sometimes less than millimetres thick, if the bed of the glacier is impermeable. Shaw (1996) inferred broad sheetfloods of turbulent subglacial meltwater. These sheetfloods are estimated to be on the order of 10 to 80 metres deep, 70 to 150 kilometres wide, and flowing at a rate of 5 to 10 m/s. Resulting estimates of discharge range between 0.6 and 6 X 10^7 m^3/s (Shaw et al. 1989). Shoemaker

(1995) estimated water sheetflood volumes to be on the order of 8,000 km^3, which is considerably less than the volume of 84,000 km^3 envisaged by Shaw et al. (1989) to form the Livingstone Lake drumlin field. Shoemaker (1991, 1992) discussed water sources for Laurentide sheetfloods, and considered a megasubglacial lake in the Hudson Bay basin, additional supraglacial inputs, and reverse floods from proglacial lakes.

The flow of water can be controlled by the ice surface slope, but if the slope of the bed is steep in another direction then the meltwater will likely not flow parallel to the flow direction of the ice (Shreve 1972, Weertman 1972). Shoemaker (1995) described that the pressure gradient due to the variable thickness of the ice sheet above the water sheet would drive the subglacial water toward the ice margin, even across an upward sloping bed. Sheetfloods would be unstable, and would become channelized over time (e.g., Walder 1982, Brennand and Shaw 1994, Sjogren and Rains 1995, Sjogren 1999).

2.1.2 Conduits

Conduits carry meltwater within and beneath glaciers and ice sheets. Englacial conduits can be found at any depth within the ice sheet or glacier, and may or may not be exposed at the surface. Subglacial conduits may be individuals or part of a larger, interconnected system. Hooke (1989) described an arborescent network of conduits ranging from millimetre-sized to larger passages. Röthlisberger and Lang (1987) stated that larger passages should be referred to as conduits, tunnels or channels, while smaller passages should be called tubes or veins. They refer to the work of Nye and Frank (1973), who described drainage through a system of veins that form at triple contacts of sutures between grains of ice.

Röthlisberger (1972) determined that, under steady-state conditions, there is an inverse relationship between pressure and discharge in conduits. If discharge were to decrease, this would cause a decrease in the energy of flow which would reduce the pressure on the conduit walls; the result would be that the conduit would decrease in size.

Shreve (1972) notes that conduit flow responds to the ice surface slope and the slope of the bed - the effect of the bed is 1/11 that of the ice surface slope. Thus, conduit orientation is influenced by hydrostatic forces, and may therefore flow uphill or diagonally across subglacial slopes.

Conduits at the glacier bed can take more than one form. Discrete subglacial drainage occurs when water is confined to only a few channels or conduits; distributed subglacial drainage refers to the situation when drainage occurs over much of the bed (Benn and Evans 1998). Discrete channel systems come in two types: Röthlisberger channels (R-channels) and Nye channels (N-channels), which are discussed below. Distributed systems can be a thin water film (e.g., Weertman 1972), a linked cavity network (e.g., Lliboutry 1976), a braided canal system (Walder and Fowler 1994), or can be mainly porewater (Darcian) flow within subglacial sediment or bedrock (Boulton and Jones 1979, cited in Benn and Evans 1998).

2.1.2.1 Röthlisberger Channels

R-channels, originally described by Röthlisberger (1972), are incised upwards into the base of the ice. They are similar to englacial conduits, except that they are floored by sediment or rock. Also similar to englacial conduits, they are kept open by the constant melting of the tunnel walls by frictional heat; the walls contract by ice creep resulting from the pressure differences between the tunnel and the walls (e.g., Röthlisberger 1972, Shreve 1985). R-channels can develop a branching network by capturing water from smaller channels, and they can be at (or near) atmospheric pressure if they are close to the ice margin or are fed from water from the surface. Because they originate subglacially under pressure, R-channels can flow uphill or across slopes; for ice sheets, water flow is generally approximately parallel to the direction of ice flow with only minor deviations due to topography (Shreve 1985). The flow path of the channels is controlled by the hydraulic gradient at the bed, which itself is more dependent on the surface slope of the ice sheet than the gradient of the bed.

The cross-sectional shape of the R-channels depends on the bed conditions (Röthlisberger 1972; Shreve 1972, 1985). The channels would be arched when flowing downhill or under melting conditions because frictional heating (and therefore melting rate) would be greatest towards the centre of the channel where water flow would be at maximum. The channels would be relatively wide and low under freezing conditions or when flowing uphill.

2.1.2.2 Nye Channels

N-channels are incised downwards into either bedrock or sediment. They can be single channels or form a braided network of channels covering a large area of the bed (Nye 1976, Walder and Hallet 1979). It is possible that the networks were once distributed systems, carrying water away from a subglacial source and gradually cutting into the bed (Walder and Hallet 1979). It has been observed that N-channels occur where topography has a strong influence on flow, such as within steep-sided valleys or rough glacier beds. The shape of the channel floor is controlled by the rate of erosion and removal of sediment; the ice-walled channel ceiling enlarges or contracts depending on melting conditions in much the same way as R-channels do (Alley 1989). The spacing of the channels depends on the amount of water to be removed and the efficiency of the drainage.

2.1.3 Subglacial Lakes

Wright (1973), Shoemaker (1991), and Munro-Stasiuk (*in press*), among others, believe that large subglacial lakes may have existed beneath the Laurentide ice sheet. Dowdeswell and Siegert (1999) document numerous subglacial lakes beneath the Antarctic ice sheet, and conclude that their findings may provide an analogue for conditions beneath Quaternary ice sheets. Subglacial lakes may have a role in glacial surging, flooding, and other ice-marginal (or non-ice-marginal) events. Subglacial cavities accumulate water when regions with a relatively high hydraulic potential surround regions with lower hydraulic potential (e.g., Nye 1976). Water moves from an area of higher hydraulic potential to an area of lower hydraulic potential until it reaches an equilibrium state, which in the case of subglacial cavities results in the formation of subglacial lakes. Nye (1976) stated that catastrophic drainage can occur as water drains from a subglacial lake by enlarging its drainage pathways to carry a vast amount of water.

2.1.4 Jökulhlaup

Jökulhlaup drainage is the periodic or occasional release of large amounts of stored water in catastrophic outburst floods (e.g., Nye 1973, Björnsson 1998, and Shaw et al. 1999). There is more than one situation that could result in triggering jökulhlaup floods, including: i) the growth and collapse of subglacial reservoirs, ii) the sudden drainage of an ice-dammed lake below or through the ice dam. Jökulhlaups usually occur during the melt season, when large amounts of meltwater have accumulated, but are not necessarily cyclic or annual events. Generally, large amounts of debris are carried by the floodwaters. Jökulhlaup discharges can be several orders of magnitude larger than regular meltwater drainage events. For example, a jökulhlaup flood event that is believed to have drained Glacial Lake Missoula and helped to form the Channeled Scabland in Washington is said to have peaked at 2.1×10^7 m^3/s, and certainly exceeded 3.0×10^6 m^3/s (Baker 1973, Clarke et al. 1984, Clarke 1986, Baker 1987).

2.1.5 Supraglacial Inputs

Supraglacial, or surface, meltwater can be considered to be an important factor in glacial hydrology. Paterson (1994) stated that supraglacial meltwater often represents the primary source of water reaching the bed of a temperate glacier. The water may percolate down to the bed via veins and englacial channels (e.g., Shreve 1985, Fountain and Walder 1998). An arborescent network of conduits, from millimetre-sized joining to form larger ones, could transport surficial meltwater to the bed (Hooke 1989, Fountain and Walder 1998). Hooke (1989) further described crevasses in the ablation area often cut across surface streams; often water drains from the crevasse in passages that lead farther down into the glacier. Hooke (1989) cites Weertman (1974), who stated that because water is denser than ice it is probable that water-filled crevasses may propagate downward until they reach the bed themselves. Moulins may also transport supraglacial meltwater to the bed. Some studies (e.g., Iken and Bindschadler 1986) have used dye tracing experiments to show a link between moulins and meltwater released at a glacier's terminus. Röthlisberger and Lang (1987) state that seepage should also be considered as a mechanism for transporting surface water to the bed.

2.1.6 Characteristics of fluids

A Reynolds number is a dimensionless ratio that expresses the ratio of inertial forces and viscous forces of a fluid. It determines whether the flow is laminar or turbulent. Laminar flow is when the fluid moves over a smooth surface in a series of thin layers sliding over one another and flowlines do not cross, and is more common for highly viscous fluids (Summerfield 1991). Turbulent flow involves flow velocity fluctuating in all directions. The flow is characterized by eddies and vortices, and there is a constant interchanging of water between zones of flow (Summerfield 1991). The Reynolds number is calculated as follows:

$$\mathrm{Re} = \frac{\rho v y}{\mu}$$

where ρ = density, v = velocity, y = a characteristic length and μ =viscosity, or (for a channel):

$$Re = \frac{vR}{\nu}$$

where v = velocity, R = hydraulic radius (area / wetted perimeter) and ν = kinematic viscosity (ratio between viscosity and density). The first formula shows that as velocity and depth increase, Re increases. A Reynolds value less than 500 reflects laminar flow, transitional values between 500 and 2500 represent flow with both laminar and turbulent elements and anything above 2500 indicates turbulent flow (Knighton 1984b).

A Froude number is another dimensionless ratio. It expresses the ratio of inertial forces and gravitational forces, and is calculated as follows:

$$Fr = \frac{v}{\sqrt{gy}}$$

where v = velocity, g = acceleration due to gravity and y = depth. The gravity component is created when water flowing over bed irregularities creates waves (ripples) which exert a weight or gravity force (Summerfield 1991). A Froude number < 1 represents subcritical (slow, gentle, tranquil) flow. In this case the wave velocity is greater than the flow velocity. A Froude number = 1 is critical flow, and a Froude number > 1 represents supercritical (fast, shooting) flow (E. Shaw 1983). In this case the velocity of flow is greater than the velocity of the waves. Supercritical flows are usually only temporary because there are large energy losses associated with this kind of flow.

Subcritical, critical and supercritical flows exist because changes in discharge can be handled by both changes in the depth and velocity of flow. The same discharge can be carried in a subcritical flow (deep, slow moving) as a supercritical flow (shallow, fast moving). Transitions between subcritical and supercritical flow are determined by velocity, and such transitions in flow regime can occur as a result of a change in the bed characteristics (Summerfield 1991). A sudden transition from supercritical to subcritical flow is called a hydraulic jump. Such a transition creates a standing wave and an increase in water depth. A transition from subcritical to supercritical flow results in a decrease in water depth land is called a hydraulic drop.

2.2 Glaciofluvial Deposits

Bedforms are created when sediment is organized into orderly patterns due to scour and deposition (Knighton 1984a). The bedforms are a response of the bed to varying flow conditions by adjusting in the vertical dimension (Knighton 1984a). Figure 2.1 shows the types of bedforms and their relevant flow conditions discussed below. Generally, as flow power/velocity increase there is a transition from plane beds to current ripples to dunes to plane beds and then antidunes.

Plane beds are subhorizontal surfaces that develop under two flow conditions: in slowly moving shallow water (lower phase) and more rapidly moving deeper water (upper phase) (Allen 1982a, Knighton 1984a). For the lower flow regime, plane beds develop from coarse to very coarse sand (0.8 - 2.0 millimetres),

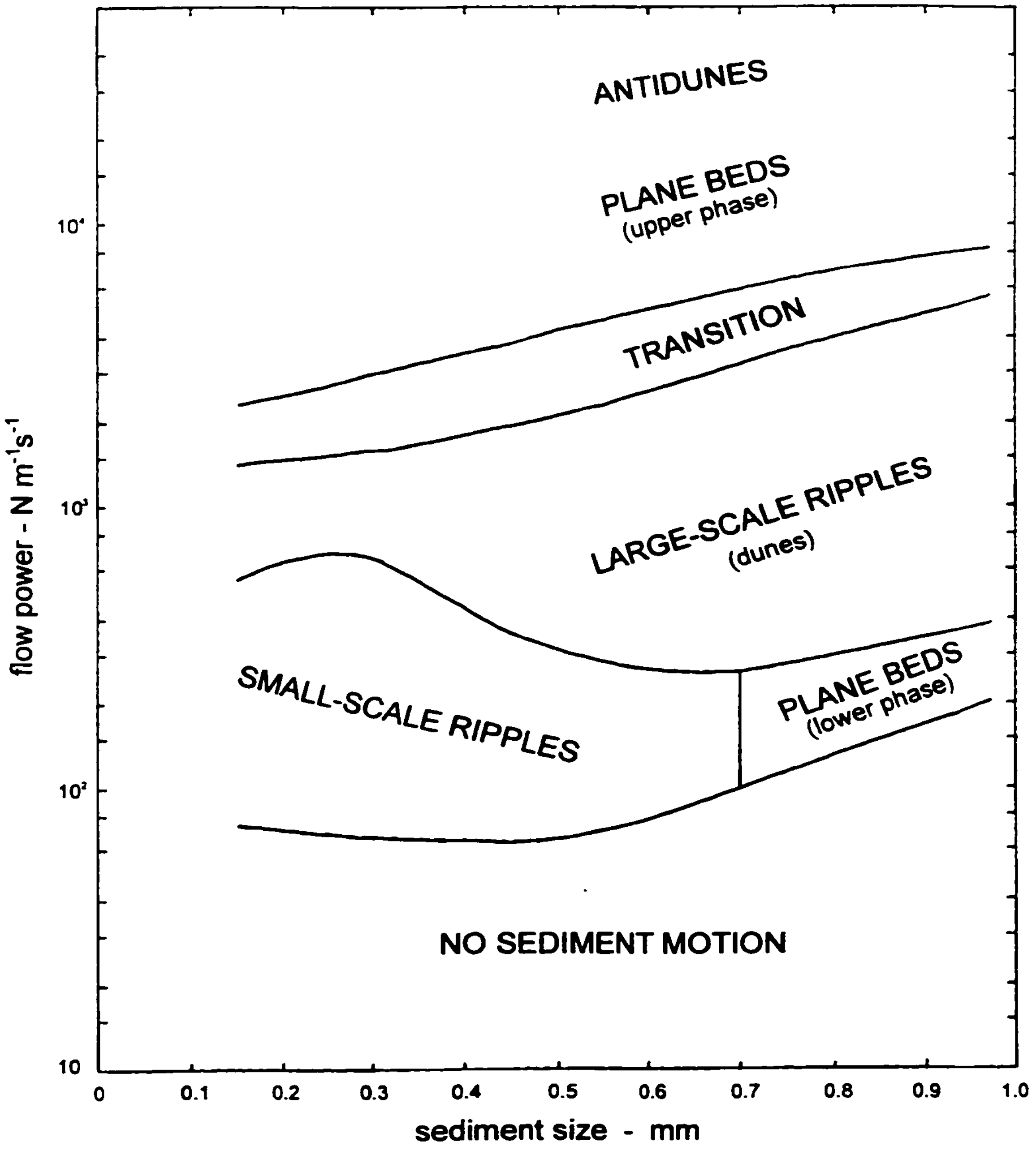


Figure 2.1 Types of bedforms, as a function of flow power and grain size. There is a sequence of forms with increasing intensity of flow: 1) plane bed without movement, 2) small-scale ripples, 3) large-scale ripples (dunes), 4) transitional forms from dunes to supercritical flow forms, 5) plane bed with movement, 6) standing sand waves, and 7) antidunes. Movement of sediment will not take place until a threshold stream power is exceeded, which varies with grain size. Modified from Figure 6.9, Allen (1968).

and are associated with relatively low rates of sedimentation and little downstream movement; under upper flow regime conditions, plane beds form in fine to medium grained sands (0.0625 - 1.0 millimetres) and are associated with greater downstream movement (Allen 1968).

Ripples are small-scale, often three-dimensional, transverse bedforms. The ripple crests are oriented at right angles to the flow, and may be long and relatively straight or have short, curving crests, depending on flow conditions (Allen 1982a). Ripple wavelength commonly varies between 10 and 60 centimetres and amplitude up to about 4 centimetres (Allen 1982a, 1985), and ripples are commonly restricted to sediment finer than 0.6 millimetres (Knighton 1984a). Ripple wavelength and height are controlled by sediment size and bed shear stress/roughness (Allen 1982a). Ripples are generally asymmetrical, with elongate, convex-up upstream faces and short, steeper downstream sides (Allen 1982a). Ripples migrate downstream as material is eroded from the upstream side and redeposited on the downstream side, and undergo vertical accretion as material is added from suspension (Knighton 1984a). Figure 2.2 illustrates the three types of climbing ripples: Type A, Type B and draped lamination (also known as Type S). Type A ripples are eroded on the stoss (upstream) side while Type B ripples have the stoss side preserved. Both Type A and Type B ripples form under migrating conditions, but for Type B ripples aggradation is more dominant; draped laminae form when aggradation is the only direction of growth (Ashley et al. 1985).

Dunes are transversely oriented bedforms, which are also referred to as

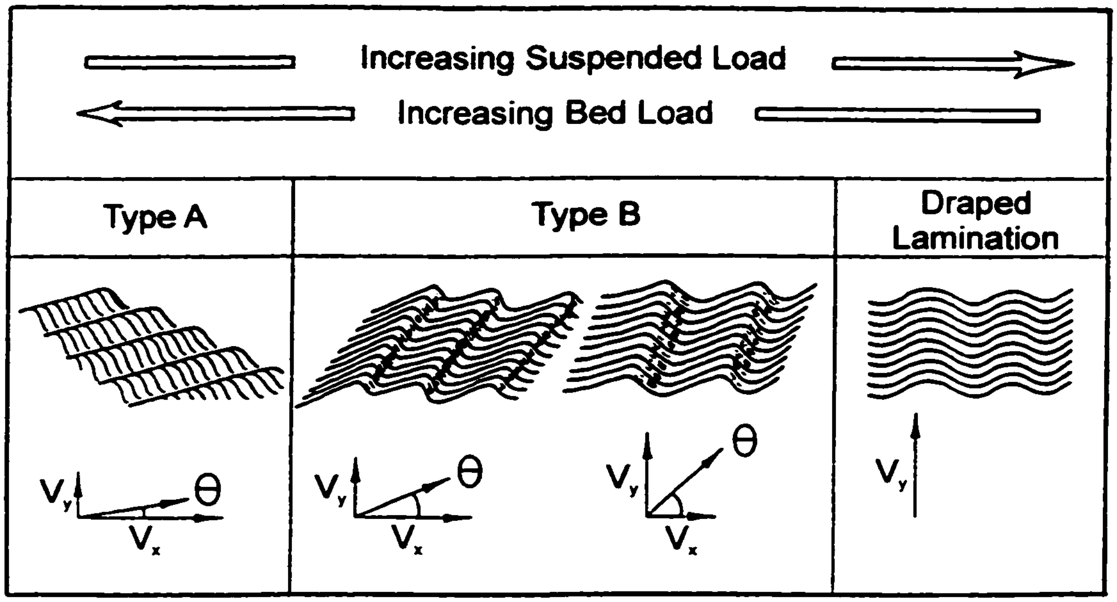


Figure 2.2 Types of climbing ripples. Type A ripples are eroded on the stoss side; in Type B ripples the stoss side is preserved. Type A ripples form under rapid migrating and low vertical aggradation conditions, while Type B ripples form under higher aggradation and slower migrating conditions. Draped laminae (also known as Type S ripples) form when aggradation is the only direction of growth. Figure modified from Figure 4-32, Ashley et al. (1985). V_y represents the mean aggradation rate, V_x represents the downstream migration rate, and Θ is the angle of climb.

bars, megaripples, large-scale ripples and sand waves (e.g., Allen 1982, Allen 1985, Knighton 1984a). They are typically asymmetrical in long profile; the lee side dips at around 30-35° while the stoss side is rarely inclined more than 5° (Allen 1982a). Dune wavelength can range from 1 metre up to 1 kilometre, depending on flow depth, grain size, and flow velocity; the amplitude can range from a few centimetres to over several metres, also depending on depth (Allen 1982a, Knighton 1984a). The wavelength/depth ratio is approximately 5 (Allen 1982a). Depending on flow conditions and sediment size, dunes can be 2-dimensional, with long, relatively straight crests, or 3-dimensional, with sinuous crests (Allen 1982a). Like ripples, dunes also migrate downcurrent (Knighton 1984a). Many dunes have smaller-scale ripples or plane beds superimposed on their upstream faces (Allen 1982b).

Cross-stratification patterns are generated by ripples and dunes (Allen 1982a). Cross-stratification refers to layers (texturally and/or compositionally distinct) inclined at an angle to the principal surfaces of the main stratification in a sediment (Allen 1982a); it has been referred to in the past as "diagonal bedding" (Allen 1982a citing Hall 1843). Cross-beds refer to strata greater than one centimetre and cross-laminae refer to strata thinner than one centimetre. Cross-strata are arranged in sets which are in turn grouped in cosets (Allen 1982a).

Antidunes are also oriented transverse to flow, but in these bedforms the steepest faces point in either the upstream or downstream direction (Allen 1982a). Antidunes are low amplitude bed waves in phase with the water-surface waves; sediment transport rate and flow velocity are high (Knighton 1984a). Antidunes may be long crested 2-dimensional features or short crested 3-dimensional features; 3-dimensional antidunes form under higher supercritical flow conditions than 2-dimensional antidunes, which form under critical and somewhat supercritical flow conditions (Allen 1982a). Antidunes migrate slowly either in the upstream or downstream direction, or remain stationary (Allen 1982a, Knighton 1984a).

2.3 Tunnel Valleys

Tunnel valleys are subglacial landforms, carved by meltwater at the glacier bed. They are typically asymmetrical, undulating, elongate depressions, some greater than 100 kilometres long and 4 kilometres wide (Ó Cofaigh 1996). Many authors note that tunnel channels commonly terminate in outwash fans or similar aggradational features, often thought to indicate ice marginal positions (e.g., Mooers 1989, Wingfield 1990, Patterson 1994, Trisko 1996). Tunnel valleys often contain subglacial landforms (e.g., eskers) on their floors, and frequently form anastomosing networks (e.g., Brennand 1994, Shaw 1994). Boyd et al. (1988) noted the undulatory nature of the valley floors, the narrow, steep walls, and the presence of smaller hanging tributary valleys in many cases. They also describe the valleys they observed as comprising an integrated network. Embleton and King (1968) stated that an up-and-down longitudinal profile is often diagnostic of tunnel valleys. Booth and Hallet (1993) commented that tunnel valleys can be relatively straight, single features, but more commonly form dendritic or anastomosing networks. They also note that the valleys are often occupied by lakes, bogs, or underfit streams.

Many authors (e.g., Wright 1973, Mooers, 1989, Shaw and Gilbert 1990, Wingfield 1990, Gilbert and Shaw 1992, Brennand and Shaw 1994) describe similar characteristics for tunnel valleys. All these authors concur that the drainage occurred subglacially, under hydrostatic pressure. They differ, however, in their thoughts as to the location of the feature in relation to the ice margin. It is more widely accepted that tunnel valleys are ice-marginal landforms (e.g., Boulton and Hindmarsh 1987, Mooers 1989, Wingfield 1990). Another hypothesis is that tunnel valleys form as part of a catastrophic, channelized meltwater flood, possibly originating well back from the ice margin (e.g., Wright 1973, Shaw 1983, Shaw et al. 1989, Ehlers and Linke 1989, Brennand and Sharpe 1993, Piotrowski 1994, Brennand and Shaw 1994). The details of these hypotheses are discussed in sections 2.3.2, 2.3.3, and 2.3.4 of this chapter.

2.3.1 Tunnel Valleys vs. Tunnel Channels

Many authors use the terms "tunnel valley" and "tunnel channel" interchangeably. Mooers (1989) differentiated between the two. He applied the term "tunnel channel" to channels carrying bankfull flow, referring to the hypothesized catastrophic outburst of subglacial meltwater. The term "tunnel valley" refers to drainage by small subglacial streams not under bankfull conditions, probably at or near the ice margin.

Some authors invoke an entirely different term to refer to the features. Booth and Hallet (1993) referred to the valleys they studied as "channelways". Other terms used include: rinnen, rinnentaler, talrinnen, tunneltaler, and tunneldalen (e.g., Grube 1979, Ehlers and Linke 1989).

2.3.2 Ice Marginal, Time-transgressive, Steady-state Formation

One theory for tunnel valley formation has the landform developing progressively over time. Christiansen (1987) reported that the Verendrye valley in southwestern Saskatchewan formed time transgressively by headward erosion while the glacier margin remained stationary at the Glidden esker. He stated that the fining-upward sequence of sediments in the valley supports the hypothesis that it formed over time. He described the valley as an N-channel, a valley formed subglacially and eroded wholly into the glacier bed. He stated that the valley developed by headward erosion, and notes that there is no evidence that the ice margin retreated at the same time as only one esker is observed. He compared the Glidden esker to the "delta-moraine" of Embleton and King (1968), both of which are similar in appearance to outwash fans described by other authors.

The creep of deformable subglacial sediment from the sides and base into R-channels, under steady-state conditions, is another postulated method of formation of tunnel valleys (Ó Cofaigh 1996, citing examples from Boulton and Hindmarsh 1987 and Mullins and Hinchey 1989). Boulton and Hindmarsh (1997) envisaged the formation of channels at the ice margin where hydraulic gradient and discharge are high and effective pressure relatively low. High water pressure within an R-channel (initially a small subglacial conduit) results in the removal of sediment in contact with the channel, and eventually enough is removed to create a tunnel valley. Ó Cofaigh (1996) pointed out that if the channel were flowing at atmospheric pressure, the effective pressure may not necessarily be low, and the hydraulic gradient may not be linked to the surface slope of the ice. Figure 2.3 illustrates Boulton and Hindmarsh's (1987) hypothesis that produces tunnel valleys from the gradual expansion of subglacial conduits.

Mullins and Hinchey (1989) studied the Finger Lakes in New York, a radial system of tunnel valleys. They hypothesized that steady-state subglacial drainage removed large volumes of pressurized subglacial meltwater and sediment, eventually eroding valleys. Ó Cofaigh (1996) pointed out that the Finger Lakes are incised into bedrock, a fact not fully addressed by Mullins and Hinchey (1989).

Cox (1985) described several phases of channelling within the tunnel valleys. He reported that a majority of the channelling and deposition is associated with non-glacigenic fluvial activity, and that glacial and meltwater processes served only a modifying purpose. This is in contrast to previous work in East Anglia, which had concluded that the valleys had formed due to subglacial meltwater erosion (Woodland 1970). Woodland (1970) described tunnel valleys in Denmark, Germany and England. He stated that the outflow ends of the tunnel valleys nearly always coincide with a recognizable line of ice stand. He also described the presence of a large gravel fan spreading out from the abrupt end of a system of tunnel valleys observed in Denmark. These valleys are shallow in comparison to their width, and he envisaged a lateral migration of a longer term, smaller tunnel (R-channel) rather than a single, wide, stream-filled tunnel (Woodland 1970). The valleys observed in

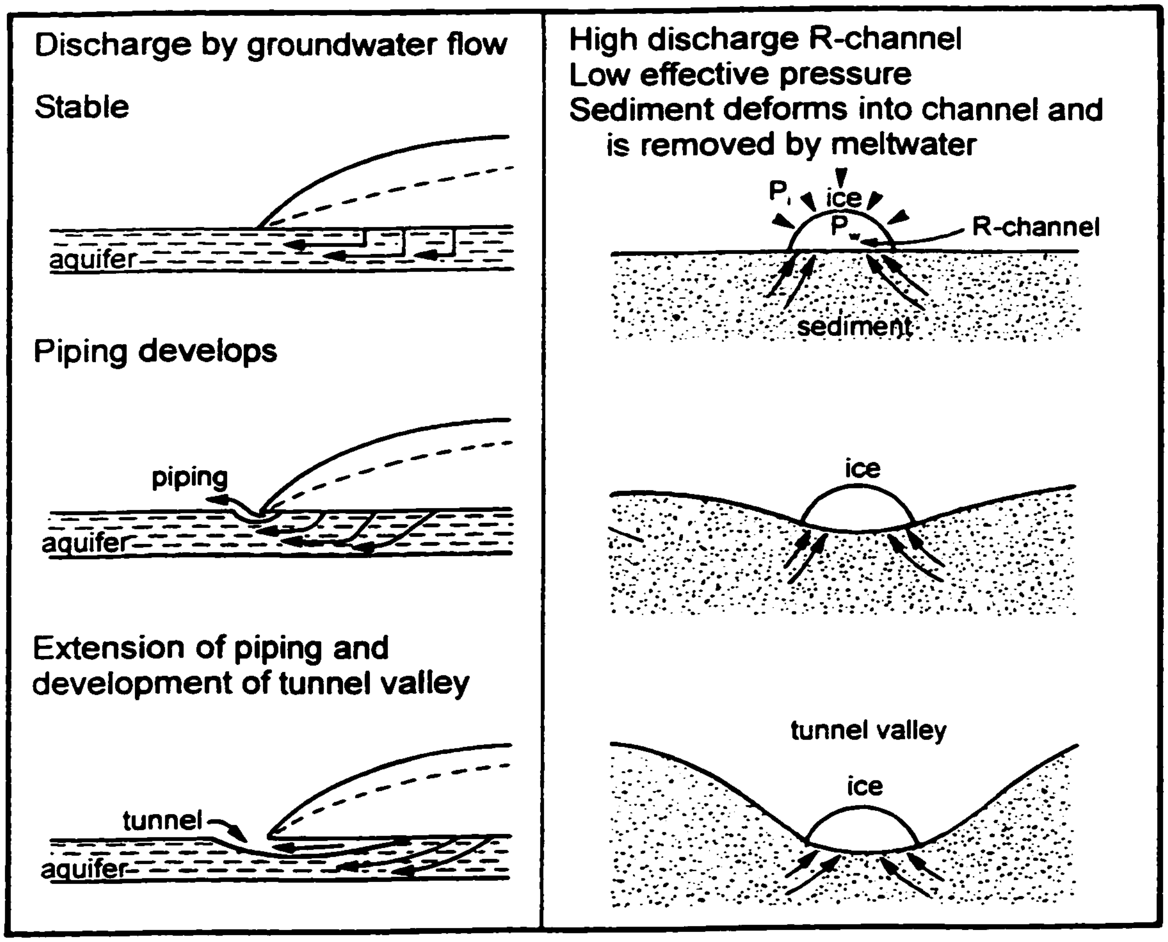


Figure 2.3 Boulton and Hindmarsh's (1987) hypothesis that tunnel valleys form as a result of the creep of sediment into R-channels. Under steady state conditions, sediment will be continually carried away by meltwater until the valley forms. Figure modified from Figure 3, O Cofaigh (1996) (in turn modified from Boulton and Hindmarsh 1987).

England are much narrower and deeper than their counterparts in Europe. He described steady-state conditions, whereby subglacial streams (under pressure) drain from beneath the ice, eroding an uneven base. As the ice retreated, the exposed portion of the valley became infilled with sediment. He stated that the fact that the valleys are deeper in their upper reaches support this hypothesis, as erosion continued longer there. The valleys filled with silts and clays after the meltwater lost its hydrostatic head and relatively stagnant water sat in the valleys.

2.3.3 Ice Marginal, Time-transgressive, Non-steady-state Formation

Other authors also hypothesize that tunnel valleys form near the ice margin over time, but they do not envisage steady-state conditions. Mooers (1989) invoked a recessional process for the valleys he studied. He described how, as the Superior Lobe retreated, new moulins and crevasses brought supraglacial meltwater to the base, thus extending pre-existing passages headwards and eroding tunnel valleys. He eliminated catastrophic outburst as a mechanism for formation of the channels because he observed that the outermost channels are deeper and wider, and there are more of them. Had flow decreased downflow, as would be expected under catastrophic conditions, the larger channels should have been located at the head of the system (Mooers 1989). Also, he observed that many of the channels are completely independent of one another, and reports no evidence to support branching or braiding within the system. He considered steady-state drainage, but concluded that the Superior lobe did not remain in place long enough for the channels to form. He concluded that tunnel valleys are composed of segments joined end to end. He stated that the younger eskers within the channels are indicators of the recessional nature of the system, as are the outwash fans at the channels' termini.

Wingfield (1990) described the jökulhlaup release of meltwater at the ice margin as being the formative process for the tunnel valleys he studied. He described the progressive headward collapse of an ice tunnel, eventually reaching the meltwater source (subglacial lake) which immediately drains. The result is an anastomosing, subaerial network of channels, fan-like in shape (Figure 2.4). Ó Cofaigh (1996) noted that there is no distal fining of sediments beyond the tunnel mouth, which would be expected, and that Wingfield's (1990) assumption of a subaerial proglacial environment may be flawed as in this case isostatic depression may have caused a relative high sea level along the ice margin.

Van Dijke and Veldkamp (1996) stressed the importance of supraglacial meltwater as a source for the volumes needed to form tunnel valleys. This is in contrast to Piotrowski (1997), who cited Reynaud (1987) and Echelmayer and Harrison (1990), who concluded that moulins would not penetrate deep enough to bring water to the bed. Van Dijke and Veldkamp (1996) proposed that tunnel valleys can be regarded as a series of elongated depressions aligned end-to-end. They concluded that the irregular beds of tunnel valleys, described as segments separated by thresholds, are an indication that the landform was created progressively over time. Periods of rapid retreat would cause the formation of tunnel valley-fed proglacial lakes (Figure 2.5). These periods of rapid retreat were

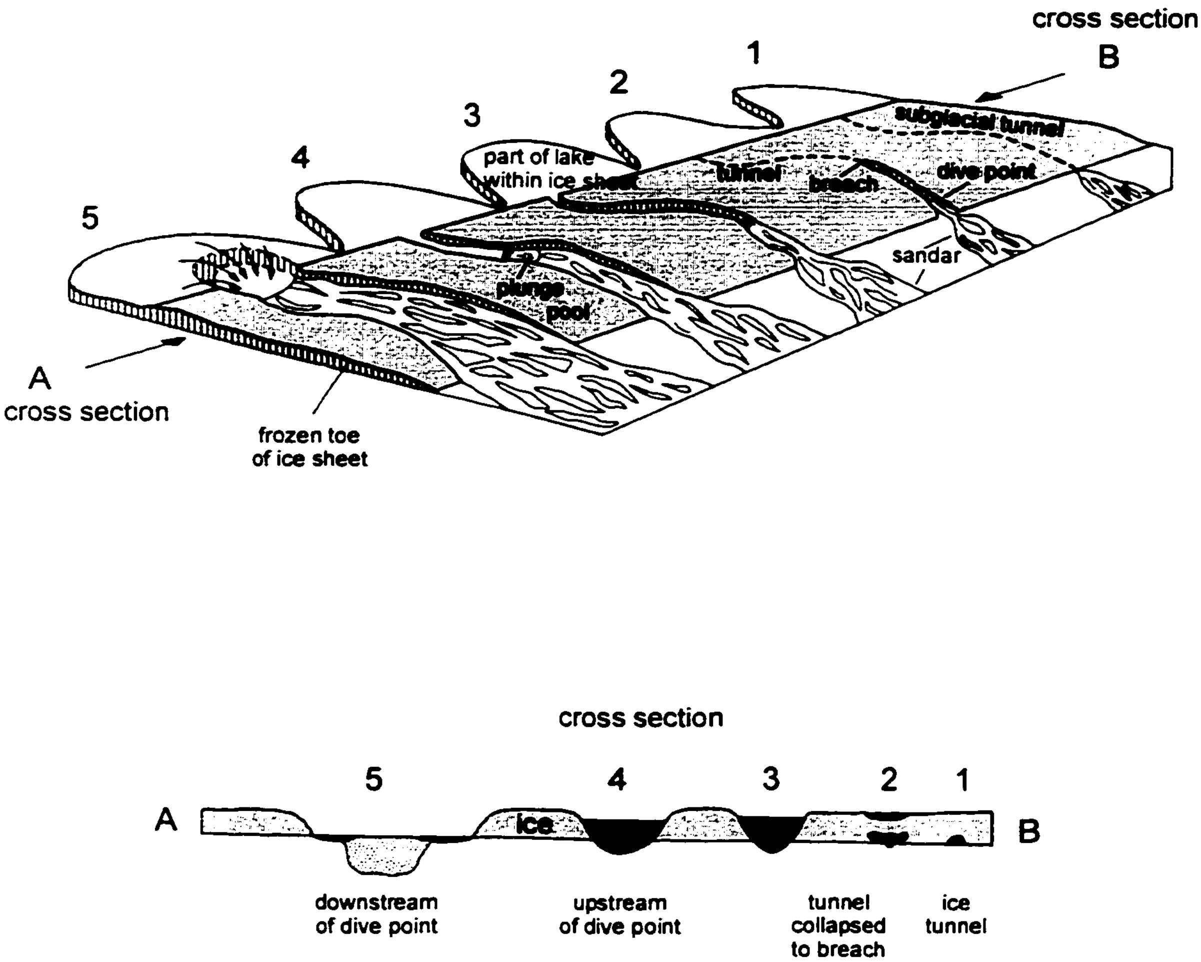


Figure 2.4 Wingfield's (1990) hypothesis on the formation of tunnel valleys. It is proposed that the gradual headward erosion of an ice tunnel will eventually trigger a jökulhlaup release of meltwater, resulting in an anastomosing network of channels. Figure modified from Figure 4, O Cofaigh (1996) (in turn modified from Wingfield 1990).

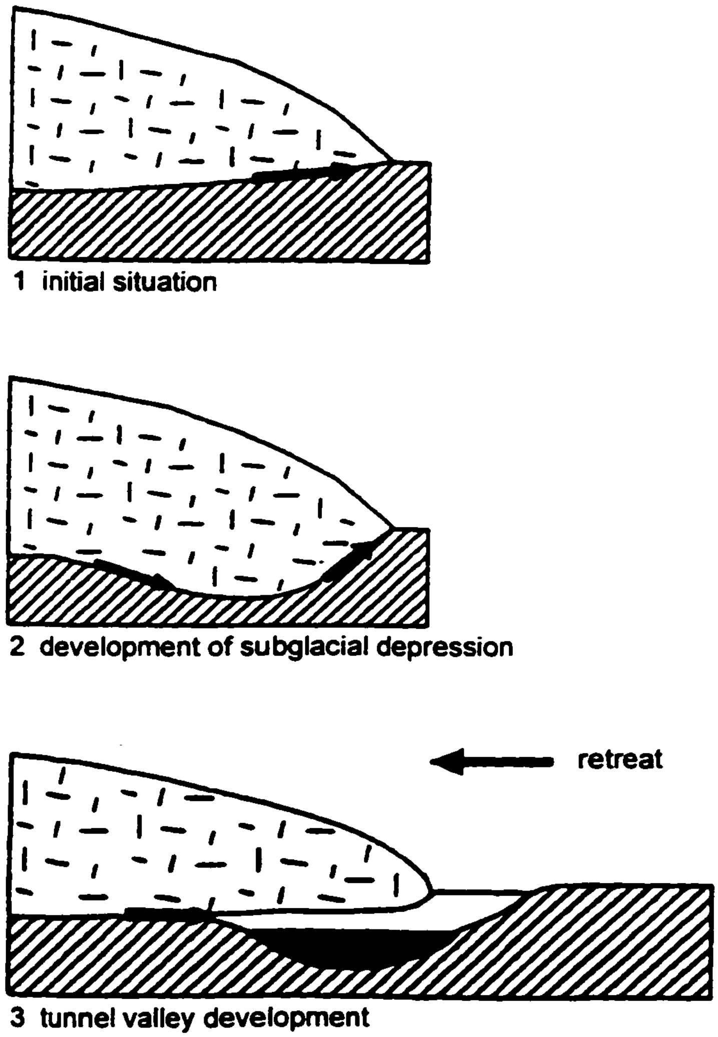


Figure 2.5 Van Dijke and Veldkamp's (1996) hypothesis on tunnel valley formation. They propose that large inputs of supraglacial meltwater caused by higher air temperatures result in the formation of elongate depressions by subglacial channels. Episodic periods of retreat are said to result in tunnel valleys with irregular beds. Modified from Figure 7, Van Dijke and Veldkamp (1996).

thought to be associated with periods of higher air temperature, which would cause increased meltwater production. The tunnel valleys form as a temporary response to conditions when supraglacial meltwater supply increases and is compensated for by subglacial channels.

2.3.4 Catastrophic Formation

Another, perhaps less widely accepted, hypothesis to explain the formation of tunnel valleys involves the catastrophic drainage of subglacial meltwater. The meltwater is described by some authors to be of flood scale (e.g., Brennand and Shaw 1994). Several authors propose that anastomosing tunnel valleys form simultaneously as the water becomes channelized (e.g., Boyd et al. 1988). This catastrophic drainage is said to be supported by the presence of drumlins, flutes, and smaller erosion features (e.g., p-forms and s-forms), also believed to have been carved by glacial meltwater (e.g., Shaw 1996).

Wright (1973) suggested that basal meltwater (formed from geothermal and/or frictional heat), possibly ponding in the Hudson Bay area for thousands of years, could have broken through the frozen toe of the Superior Lobe, and carved the tunnel valleys during its catastrophic release. He noted that as the flow waned, small eskers were formed within the valleys. His tunnel valleys trend sub-parallel to the contours of the surface, lending strong support to the conclusion that the valleys formed subglacially, their courses determined by the slope of the ice surface.

Brennand and Shaw (1994) suggested the progressive channelization of a meltwater sheet as the formative mechanism for tunnel channels. They hypothesized that gradual channelization and flow diversion processes resulted in these landforms. For bedrock terrain, Brennand and Shaw (1994) envisage a transition from a sheet of subglacial meltwater to a megachannel to tunnel channels (some interconnected and some unconnected). In drumlinized terrain, they see a starting point of either linked channels and broad cavities or a meltwater sheet which channelizes over basal highs to form tunnel channels. They concluded that overbank flow may result in the formation of smaller, secondary channels. They also interpreted transverse ridges within tunnel channels as bedforms generated by meltwater erosion and deposition, and eskers within tunnel channels may represent seasonally driven meltwater drainage.

Piotrowski (1997) stated that fine-grained, subglacial sediments with relatively low hydraulic conductivities may only drain about 25% of basal meltwater. He suggested that the remaining meltwater would be evacuated through tunnel valleys in catastrophic outburst events. Piotrowski (1997) proposed a cycle of ponding and catastrophic release through tunnel valleys. Piotrowski (1994), in describing the formation of the Bornhöved tunnel valley, proposed the following: the formation of a large subglacial meltwater reservoir may have caused an episode of surge; drainage was prevented because the margin of the ice was frozen to permafrost; when the glacier retreated, the water was catastrophically released at a rate estimated at about 3.75×10^3 m^3/s (Figure 2.6).

Ehlers and Linke (1989) stated that the infill they observed reflects episodic

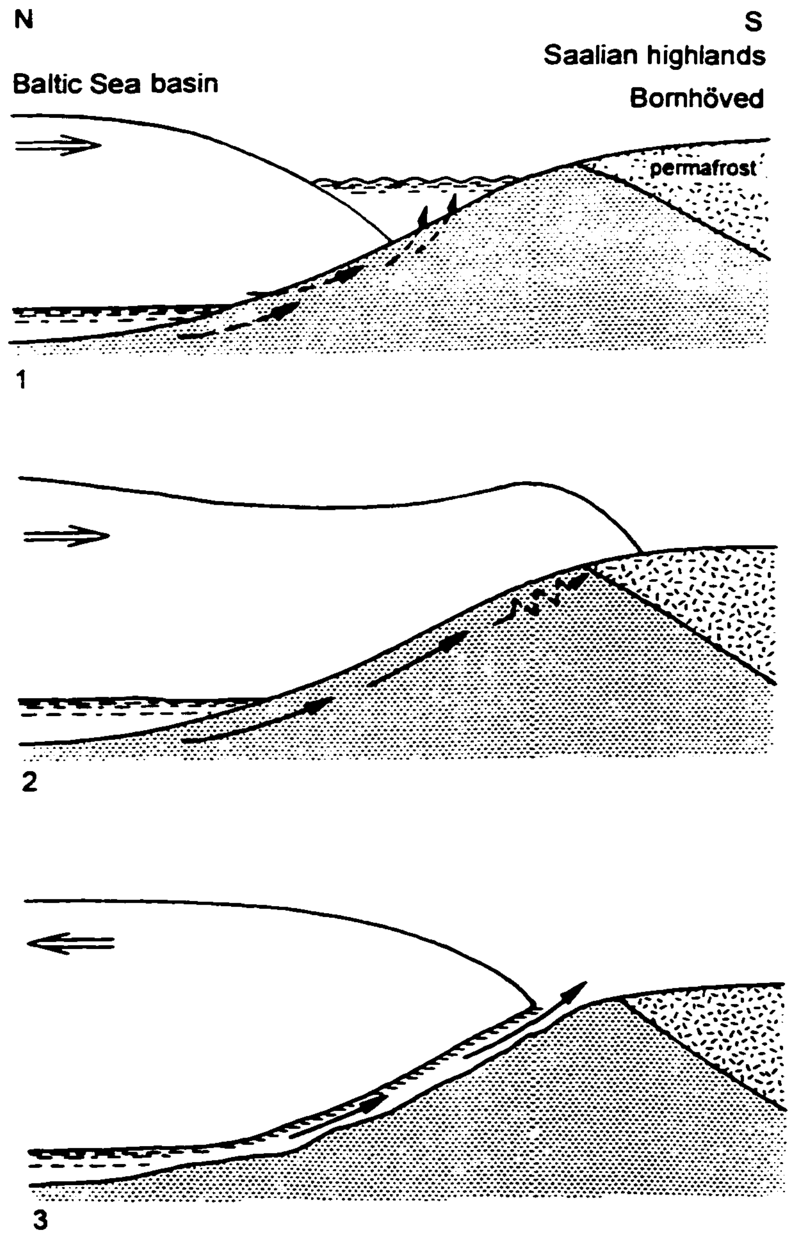


Figure 2.6 Piotrowski's (1994) tunnel valley formation hypothesis. He describes the advance of the ice sheet, with drainage of subglacial meltwater hampered by a proglacial lake. The sheet overrides a permafrost zone, which further prevents drainage. When the margin retreats, there is a rapid outburst of the subglacial meltwater which creates, or modifies, a tunnel valley. Modified from Figure 16, Piotrowski (1994).

outbursts of meltwater. They concluded that tunnel valley formation is the result of catastrophic meltwater release near the ice margin, and that later meltwater outbursts followed along the same paths as before. They stated that drainage follows depth contours as well as being controlled by hydraulic head. Ehlers and Linke (1989) also speculated that the channels may be resculptured and widened by ice.

Booth and Hallet (1993) related their "channelways" to the episodic catastrophic release of water from ice marginal lakes. They noted that meltwater could pond behind topographic highs, and be catastrophically released when the water level rises to a sufficient height. They observed the presence of lacustrine sediments in the alpine valleys immediately upglacier from these high points, which supports their hypothesis. They noted that the drainage of one lake would likely trigger the drainage of all others downglacier. They speculated that the refilling time for the lakes could take only a few decades.

Boyd et al. (1988) showed how sudden discharge events could cause rapid increases in subglacial water pressure, which would result in the development of periods of unstable sheet flow followed by the formation of channels. They hypothesized that larger channels could grow at the expense of smaller ones. They calculated that the channels observed could support bankful discharge of 6.4×10^5 m^3/s. This volume of meltwater would require a sufficiently large subglacial depression to act as a reservoir; they proposed that the Scotian mid-shelf basins could be the water source. They speculated that a rise in sea level could be the trigger necessary to start the catastrophic release of the meltwater.

2.3.5 Systems vs. Single Channels

Tunnel valleys can exist as part of a networked system of several valleys (e.g., Boyd et al. 1988, Patterson 1994), as an anastomosed system (e.g., Grube 1983, Shaw and Gilbert 1990, Booth and Hallet 1993, Brennand and Shaw 1994) or as single entities (e.g., Christiansen 1987, Mooers 1989, Sjogren and Shaw 1996). Ó Cofaigh (1996) proposed that tunnel valleys may be polygenetic in origin.

2.4 Additional Glacial Landforms

2.4.1 Flutes

Flutes are troughs, found between regularly spaced elongate ridges. The ridges are typically low (<3 metres) and narrow (<3 metres), and are usually less than 100 metres long (Bennett and Glasser 1996). The fluted surface is oriented parallel to the direction of ice flow. Megaflutes bound ridges that are taller, broader, and longer than those between flutes. The length to width ratio is usually greater than 50:1.

The ridges may be composed of lodgement till or fluvial sands and gravels. They are often said to start from either a boulder, collection of boulders, or a bedrock obstacle at the up-ice end (e.g., Boulton 1976, Rose 1989). As the ice flows around the obstacle, a cavity or area of low pressure will form in its lee. Sediment may move into the low pressure area, forming a linear ridge behind the obstacle. The ridge grows in length, down-ice, as the low pressure area extends in front of the sediment ridge (Bennett and Glasser 1996). Subglacial meltwater flow within the cavity may accentuate the morphology of the ridge by eroding sediment along its flanks. Because not all the ridges have obstacles, and for additional reasons, other authors (e.g., Shaw 1996) proposed that fluted surfaces are formed by the erosive action of meltwater. Vortices are thought to bifurcate, forming troughs and leaving elongate ridges behind. Pollard et al. (1996) proposed that flutes and other erosional grooves are created when flow patterns are disrupted by forward-facing escarpments.

2.4.2 Ridges

Transverse ridges are common in glaciated areas. The ridges can range from 1 metre to well over 100 metres high. They normally consist of basal till and/or contorted bedrock. They form a distinct geomorphic pattern of subparallel ridges and depressions. Such ridges can form by a variety of mechanisms.

Small ridges are said to form by the squeezing of basal till into subglacial crevasses (e.g., Hoppe 1952). The ridges are preserved if the ice becomes cold-based immediately after ridge formation, if the ice stagnates, or if the ice melts out immediately after the formation (Bennet and Glasser 1996). The ridges may also be formed by glaciotectonic thrusting of sediment or bedrock blocks (e.g., Kupsch 1962, Westgate 1968, Tsui et al. 1989). Beaney (1998) proposes that transverse ridges may be primary bedforms, resulting from erosion beneath stationary waves in subglacial meltwater sheet flows.

2.4.3 Hummocky Terrain

Hummocky terrain is characterized by a series of rounded hills and depressions. The hillocks can vary in size and shape; most are 1 to 50 metres high and 25 to 300 metres wide, with slopes ranging from 1° to 25° (Munro and Shaw 1997). Such terrain is conventionally thought to have been created during the stagnation phase, by deposition of material during melt-out (e.g., Gravenor 1955, Gravenor and Kupsch 1959), or by ice-pressing causing basal sediment to squeeze up into mounds (e.g., Hoppe 1952, Stalker 1960). Recent work has begun to examine the possibility that the features are largely the result of meltwater erosion (e.g., Rains et al. 1993, Munro and Shaw 1997). Several researchers have noted that not all hummocks consist of sediment that could have originated by let-down (e.g., Hoppe 1952) and others have observed hummock exposures that also discount squeezing (e.g., Munro and Shaw 1997). The presence of truncated glaciolacustrine deposits (Munro and Shaw 1997) and undisturbed bedrock (Munro-Stasiuk and Sjogren 1998) can only be satisfactorily explained by erosional processes. While Munro and Shaw (1997) state that some hummocks are indisputably erosional forms, they cite examples of others that form by stagnation. It is therefore concluded that hummocky terrain is polygenetic.

2.4.4 Eskers

Eskers are glaciofluvial landforms formed by meltwater. They are said to form in a variety of settings: 1) by deposition in subglacial tunnels, 2) by deposition in englacial tunnels and then subsequent lowering, 3) by deposition in supraglacial channels and then subsequent lowering, and 4) by deposition in ice-walled re-entrants at the ice margin (Bennett and Glasser 1996). They are usually slightly sinuous ridges of glaciofluvial sediment, and tend to undulate in height along their length. Sometimes they have a beaded appearance, with wider sections at regular intervals along the length, and sometimes form a chain of short lengths of esker ridge between which the ridge is barely visible. The orientation is controlled by the ice surface slope and the water pressure within the glacier. Topography is not a major factor, and eskers therefore do not necessarily form in the downslope direction.

There are two types of eskers: single ridge eskers and braided/anastomosing eskers. Braided eskers consist of a network of ridges which bifurcate and merge. These are typically shorter in length, usually less than 1 kilometre, and are often associated with kame and kettle topography. Single ridge eskers vary from less than 1 kilometre to hundreds of kilometres in length. Long eskers are usually 400-700 metres wide and 40-50 metres high, while small eskers (<300 metres long) are usually 40-50 metres wide and only 10-20 metres high (Bennett and Glasser 1996).

Eskers are said to form when a tunnel becomes obstructed and deposition occurs (Warren and Ashley 1994). Brennand (1994) stated that eskers formed in anastomosing tunnels may form synchronously, with erosion, transportation and deposition happening at once, though not at the same place in the tunnel.

Eskers are typically composed of a core or base of poorly sorted sands and gravels (e.g., Banerjee and McDonald 1975). Above this, sorted sands and gravels are common. These are usually well rounded and have palaeocurrent orientations that parallel the length of the esker. Sometimes a thin veneer of till may cap the esker. Brennand (1994) proposed that alternating sand and gravel beds observed in some eskers may represent a response to seasonally driven episodic flood events.

Eskers are repeatedly reported to be observed within tunnel valleys (e.g., Brennand and Shaw 1994). They are subsequent landforms, formed during waning flow (Brennand and Sharpe 1993).

CHAPTER 3: MORPHOLOGY OF THE TAWATINAW VALLEY

3.0 Introduction

The Tawatinaw valley is a long, wide, shallow valley currently containing an underfit stream (the Tawatinaw River). Current drainage is northward into the Athabasca River, with an approximate discharge of 32-45 m^3/day (Borneuf 1973). This chapter presents a detailed description of the morphology of the valley, and describes the appearance of the landforms observed.

3.1 General Characteristics/Landform Associations

Figure 3.1 is a terrain model of the Tawatinaw Valley. It was created from twenty 1:20 000 DEMs (Digital Elevation Models) covering the region. The data were obtained from the Alberta Government, and were produced using measurements from aerial photographs sampled at 25 metres; the data are accurate such that 90% of the points are +/- 5 metres (Land Information Services Division 1988). A sun azimuth of 315° with an elevation of 25° from horizontal were used to simulate light and shadow in the image.

The NNE to SSW trending valley is clearly the dominant feature of the image. The sinuous nature of the valley is shown, and the Tawatinaw River is visible in places. The present-day river is very much an underfit stream, and given the resolution of the data (25 metres) it is not discernable to the south where the channel is only a few metres wide. The valley sides are relatively steep, with slopes around 5-6°. The valley is approximately 60 kilometres long and is, on average, 75

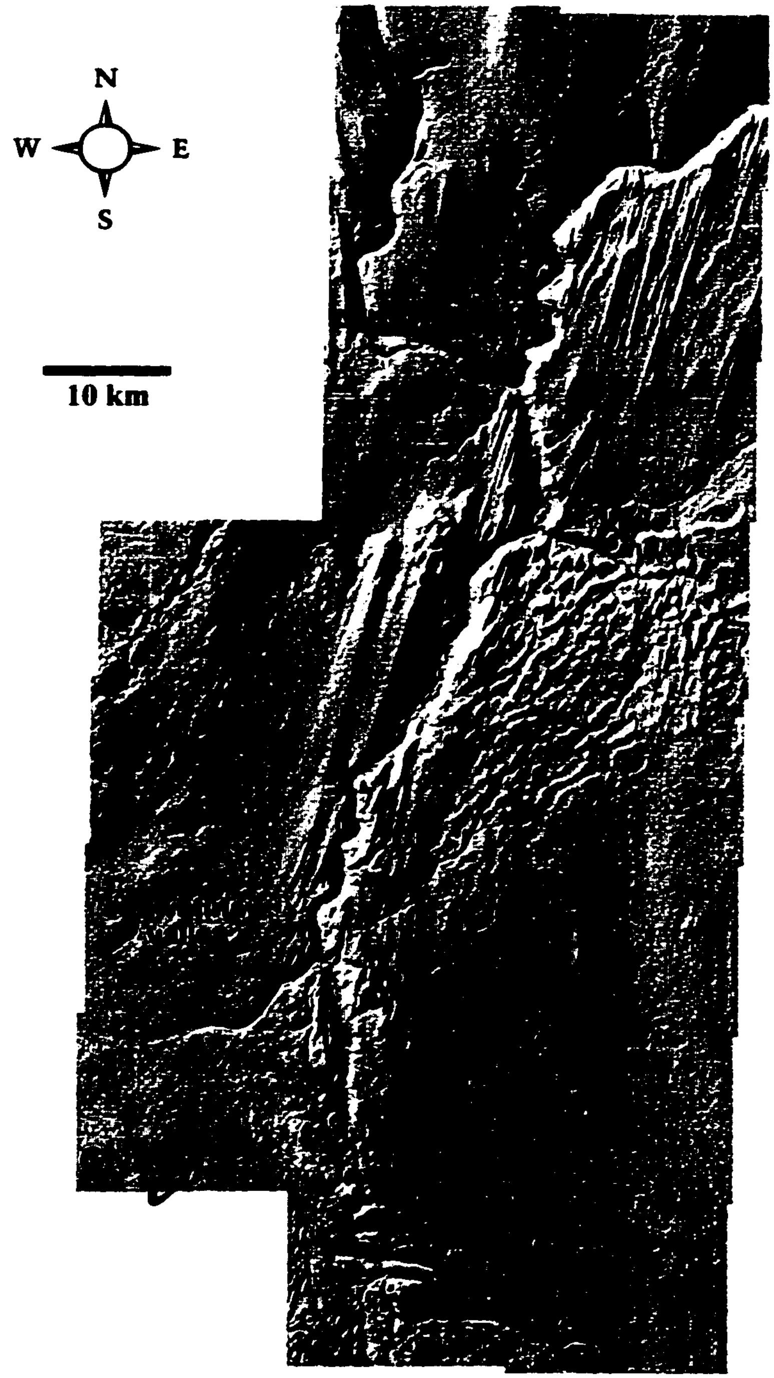


Figure 3.1 Hillshaded terrain model of the Tawatinaw valley. Created from twenty 1:20,000 Digital Elevation Models; a sun azimuth of 315° with an elevation of 25° from horizontal were used to simulate light and shadow. Note the sinuous nature of the valley. Also visible are the Athabasca flutes, ridged terrain, large subtle NW-SE elongate ridges to the southeast of the image, and erosional residuals within the glaciofluvial complex.

metres deep. It ranges from 1500 metres wide at the north end to 750 metres wide before it branches out into the glaciofluvial complex at the southern end. There are variations in width throughout the length of the valley. Little Pine Creek, in a smaller valley located to the east, is a tributary to the Tawatinaw valley.

Giant flutes are striking landscape elements in the north and northeast of the image. Many large, elongate ridges and a three-dimensional array of ridges and depressions characterize the landscape surrounding the Tawatinaw valley southwest from its junction with Pine Creek. These features are described in Chapter 1. In the southeast of the image some very large, extremely subtle, NW - SE trending, elongate ridges are present.

The sinuosity of the Tawatinaw valley, and the presence of bar-like residuals along the west-facing valley wall, are important aspects of the valley morphology. Figure 3.2 is a cropped hillshade of the valley, and arrows indicate two of these bars. The bars are usually teardrop shaped, and occur in a train along the length of the valley. The sediment of the bar indicated by the lower arrow is glaciofluvial granular sand; the palaeocurrent measurements taken from a trough cross-bedded exposure of this sediment are presented later. The east-facing valley wall is distinctly scalloped, and arrows point to two of these sculpted margins. The scallops coincide with the bars - they are almost directly opposite each other. To the east of the valley, numerous sinuous ridges are observed.

Figure 3.3 is a classed elevation representation of the same data used to create the terrain model in Figure 3.1; the hillshade is incorporated into the figure

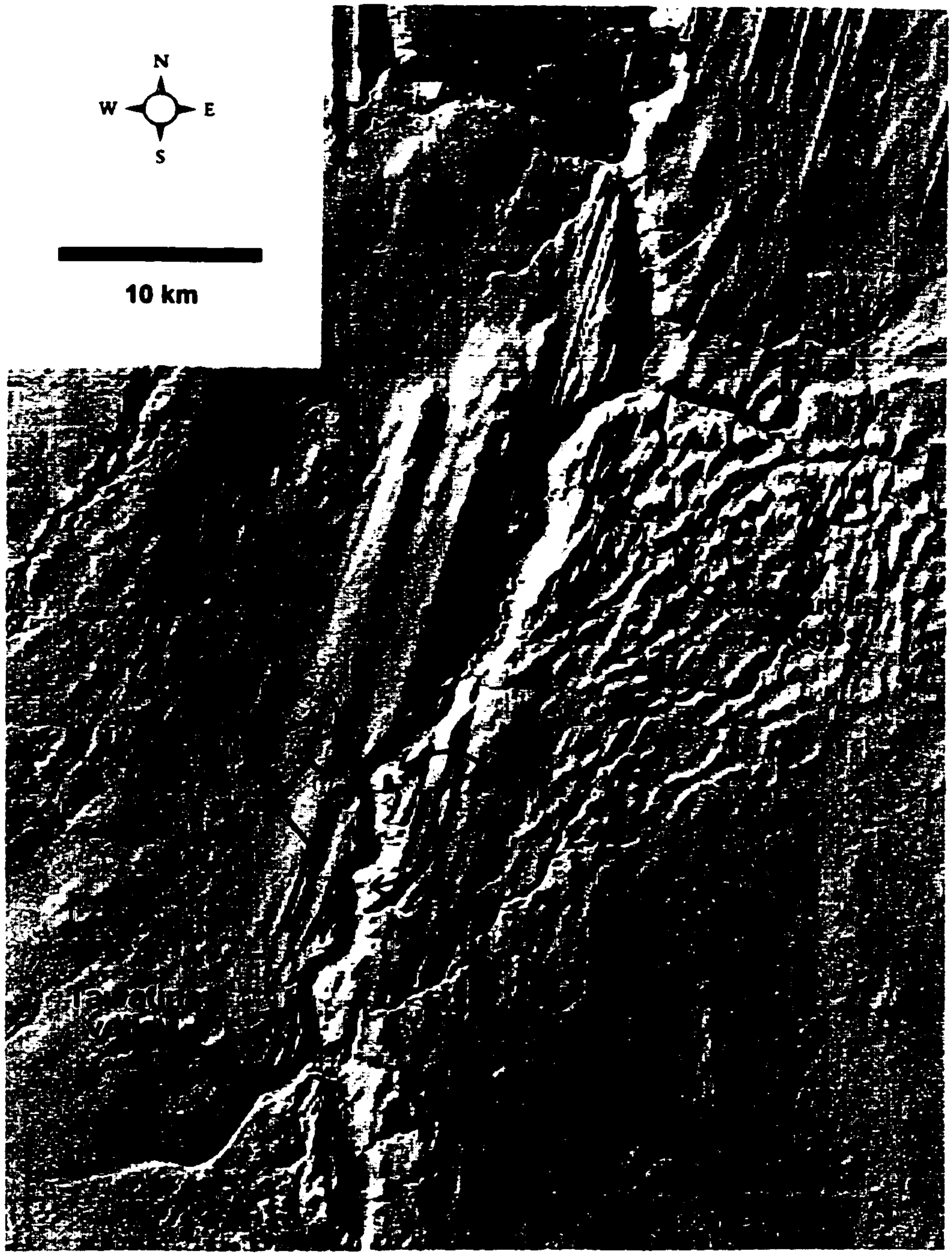


Figure 3.2 Hillshaded terrain model of the Tawatinaw valley, cropped from Figure 3.1. Note the scalloped appearance of the east-facing bank of the valley and the bar-like features against the west-facing valley wall, some of which are indicated with arrows. Sinuous ridges to the east of the valley are also indicated.

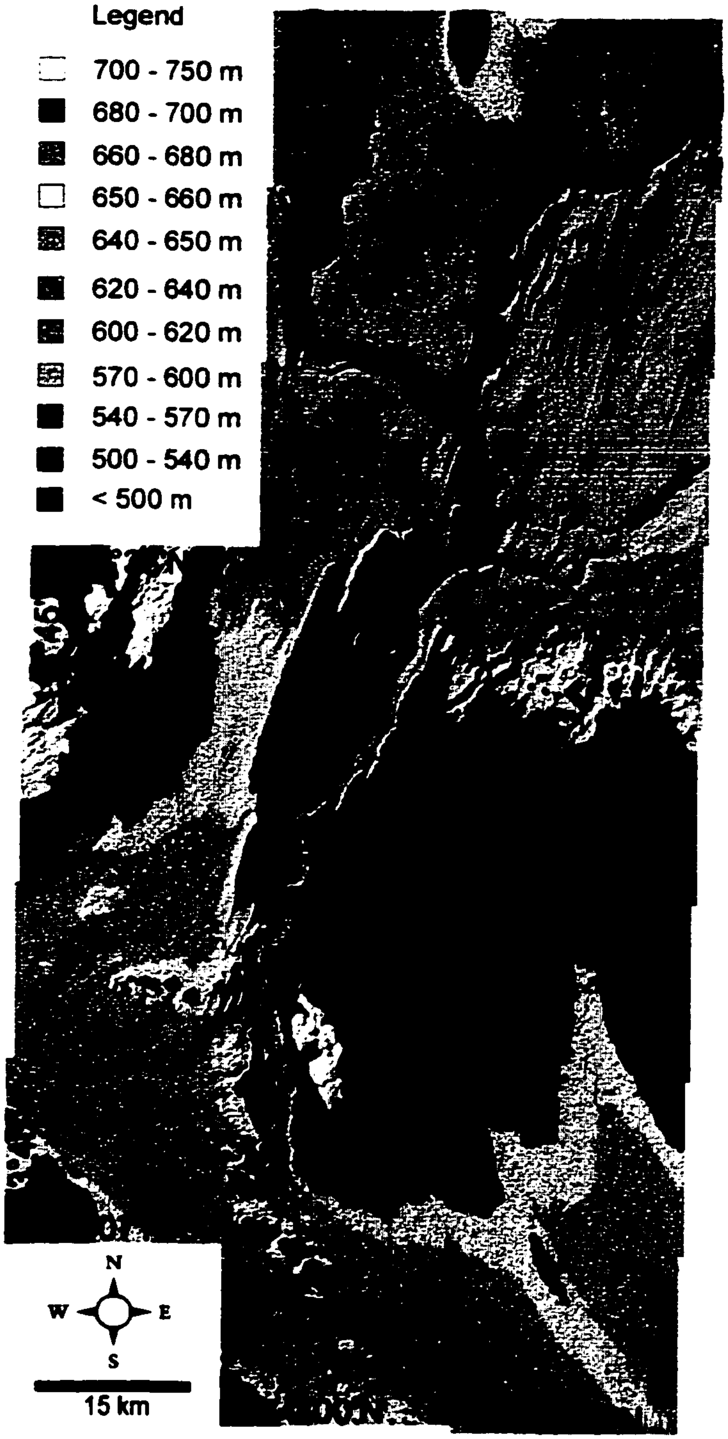


Figure 3.3 Classed elevation map with hillshade overlay. Elevations are expressed as metres above sea level. Note that the lowest part of the valley is at the northern end, with the thalweg of the valley rising southward. There is an increase in elevation of approximately 100 metres from north to south. Also note that the glaciofluvial complex does not occupy a depression - the larger branch straddles a high point before extending off to the southeast.

in order to modulate the colours so that absolute elevation, slope and aspect are all conveyed. This image clearly shows that the lowest part of the valley is at the northern end, with the thalweg rising southward. There is a steady rise of approximately 100 metres from the north to south. Where the valley reaches the glaciofluvial complex it bifurcates into a southwest branch and a larger north-south branch that extends up and over a high point before turning towards the southeast. This bifurcation is very important to reconstructing the evolution of the drainage system because palaeocurrent data taken from sediment within the valley (to be discussed in Chapter 4) indicate that water within the valley was flowing from north to south. This image indicates that any water flowing from north to south in this valley must have been flowing uphill under pressure - lending strong support to the idea that flow was subglacial.

The eastern channel carrying flow southward out of the valley into the glaciofluvial complex has a convex-up long profile. It is higher than the walls of the Tawatinaw valley at the northern end.

3.2 Valley Profiles

Figures 3.4 and 3.6 show a series of profiles (both long and transverse) of the Tawatinaw valley. These profiles were created using the 1:20 000 DEMs and choosing the endpoints of the profiles after importing the data into IDRISI.

Figure 3.4 is a profile down the axis of the valley. Clearly there is a rise from north to south. It is important to note that the long profile is down the axis of the

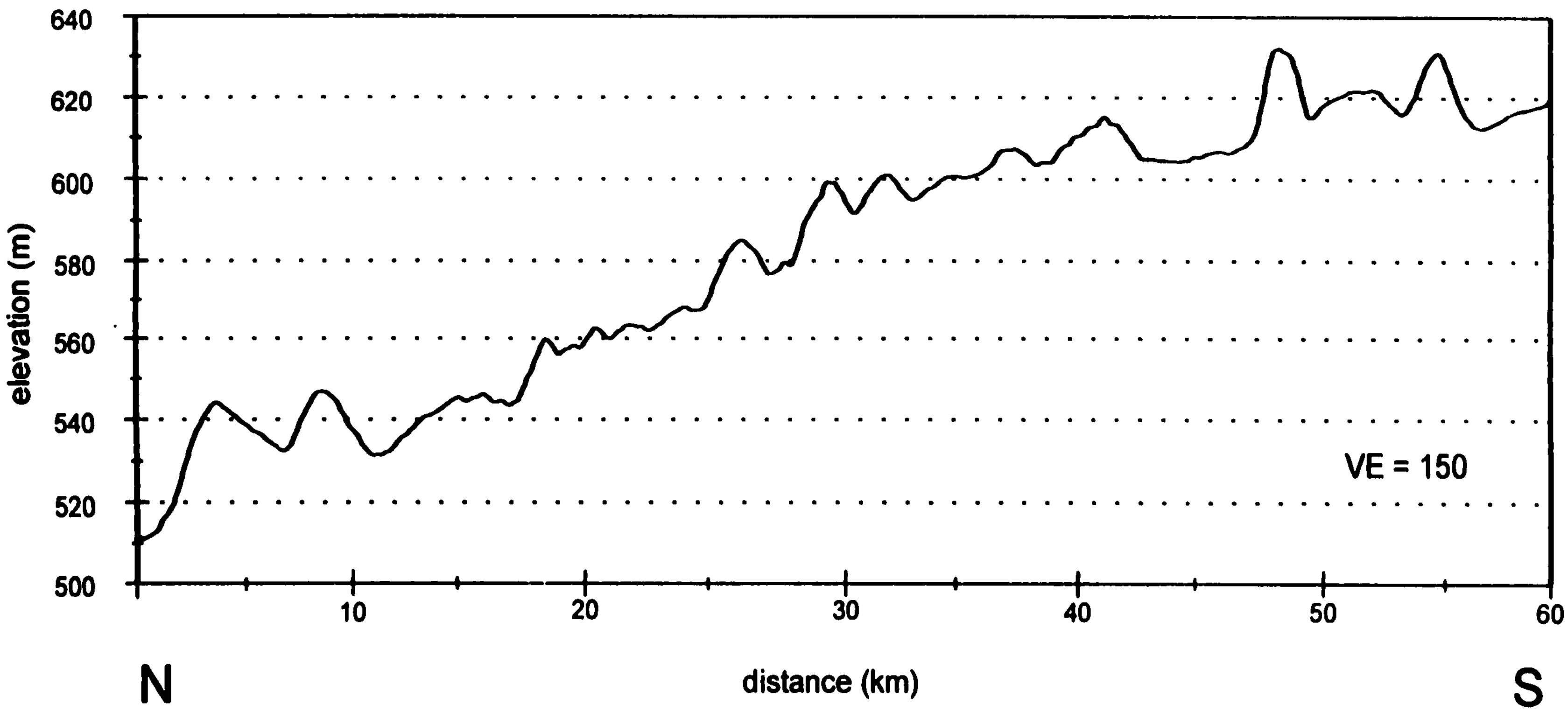


Figure 3.4 Longitudinal profile along the axis of the valley. The profile clearly indicates the rise, of over 100 metres, from north to south. The valley floor undulates with an amplitude of between approximately 10 and 20 metres. The undulations are superimposed on the rising profile.

valley, not the thalweg. The local variations in topography, superimposed on the long profile, represent the crossing points of bars and slip-off slopes.

Figure 3.6 is a series of cross profiles taken down the length of the valley; Figure 3.5 indicates the locations of these profiles. The cross-profiles show the asymmetrical nature of the valley and the steep-sides of the valley, typical of tunnel valleys. In many profiles the underfit modern river can be identified, and the series again reflects the valley floor rising southward. Profiles 1 through 4 show the flutes on the outer valley surfaces. Slumping is present on the west-facing slope in Profile 5, explaining the more stepped nature of the valley sides. Profiles 5 through 7 show the bars located on the eastern side of the valley. Profile 9 is taken after the valley bifurcates at the southern end. The valley maintains a fairly uniform width throughout, though tending to narrow towards the south; variations in width are on the order of only approximately 100 metres.

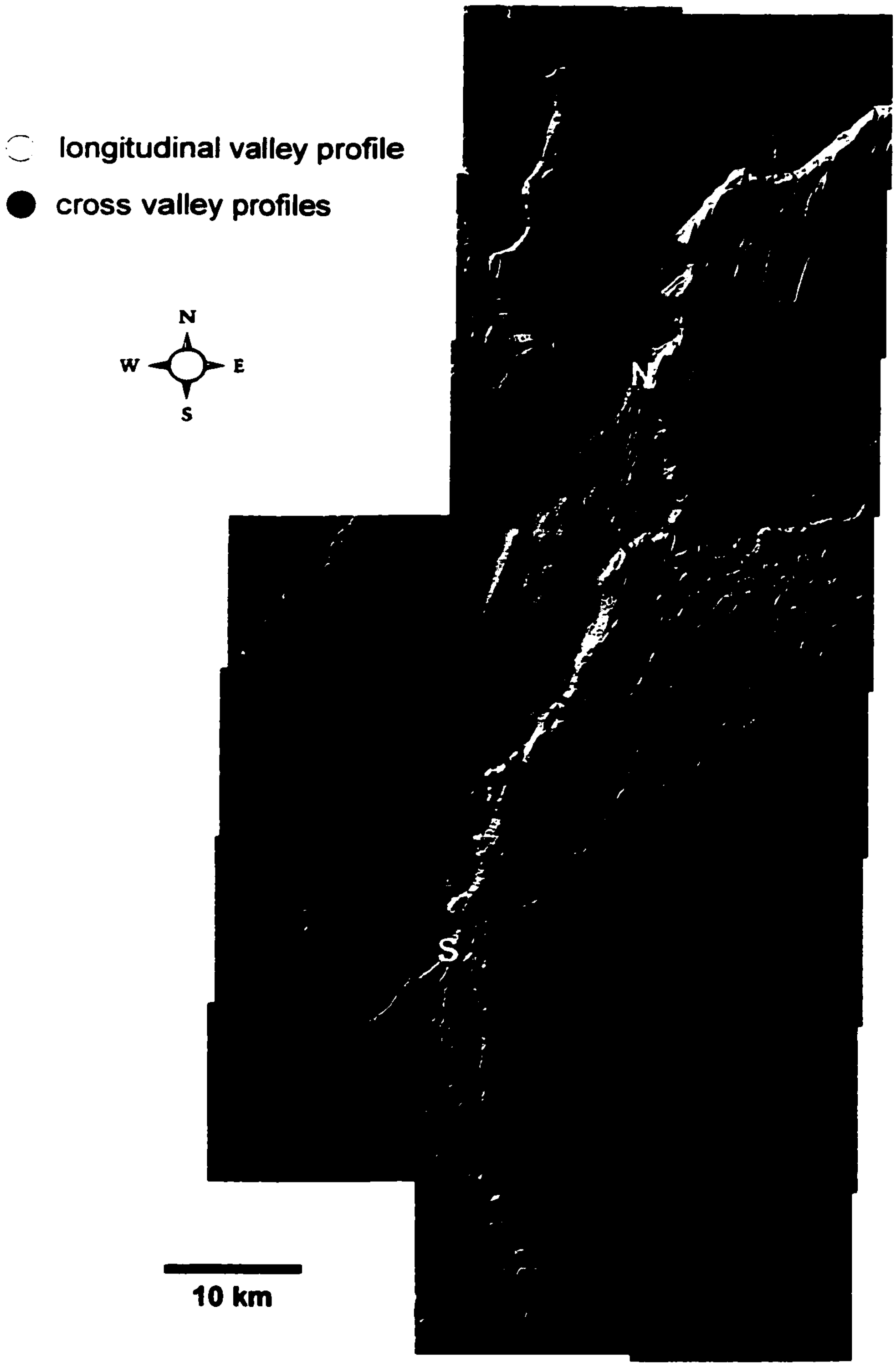
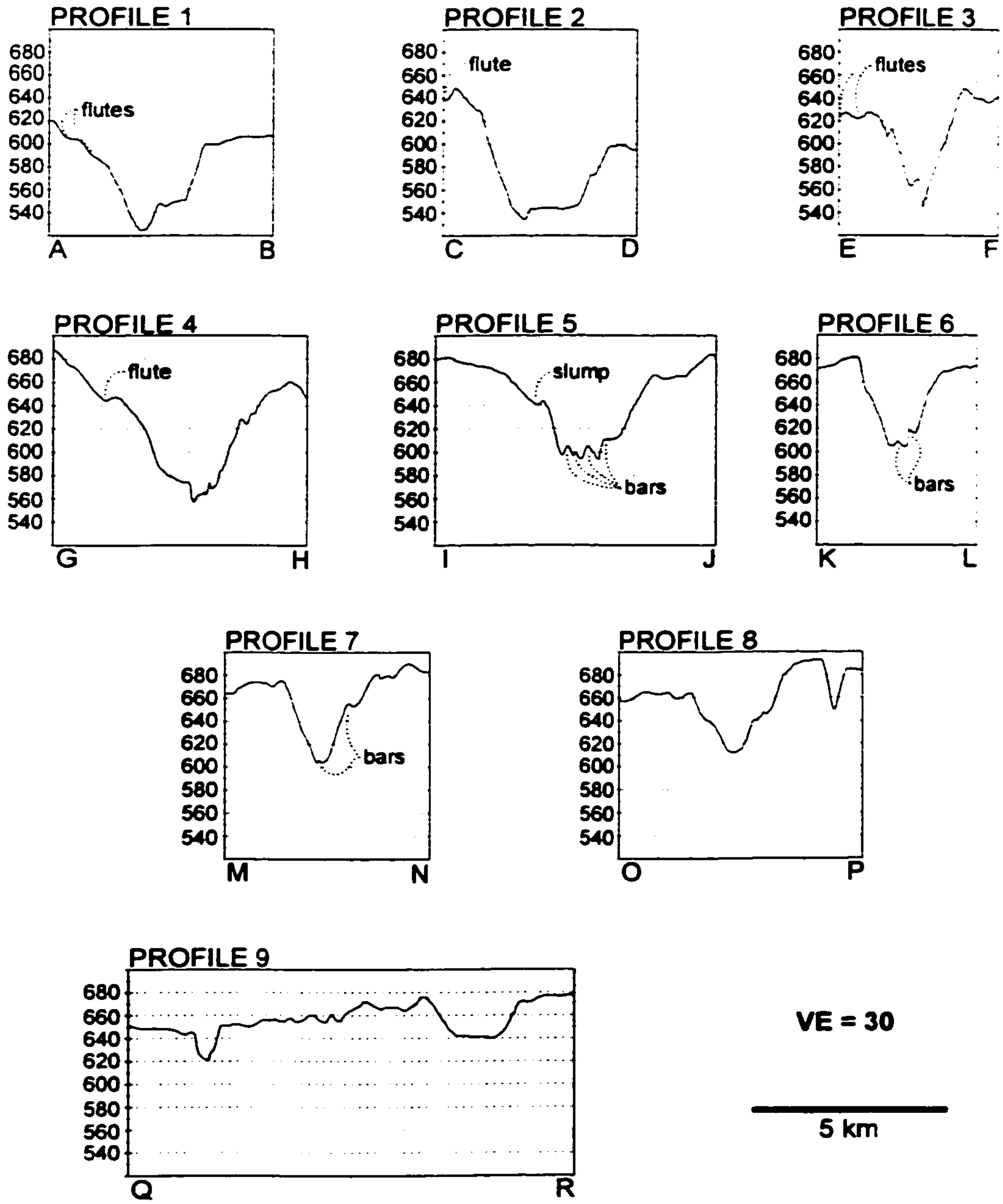


Figure 3.5 Locations of the profiles created. The long profile was taken from north to south, following the thalweg as closely as possible. Nine cross profiles were plotted, taken perpendicular to the valley. These profiles are presented in Figure 3.6.



Figue 3.6 Series of cross profiles taken along the length of the valley. Note the asymmetry and steep sides of the valley. The profiles again reflect the steady southward rise of the valley floor. The valley maintains fairly uniform width throughout. Profile 9 is taken after the valley bifurcates at the southern end. Profile 5 is taken in a region of suspected slumping. Figure 3.5 shows the locations of the profiles. Elevation is in metres above sea level.

CHAPTER 4: SEDIMENTOLOGY

4.0 Introduction

Sediments associated with the Tawatinaw valley are critical to understanding the glacial history. By examining the sediments within the valley, near its margins, and within the glaciofluvial complex to the south it is possible to pose a reconstruction of the probable events that occurred during the last glaciation. Some sediments will reflect localized deposition, while other sediments will provide indicators of more regional scale events.

4.1 Tawatinaw Valley

4.1.1 Valley Sediment Size Analysis

Twenty-two sediment samples were taken down the length of the valley. Figure 4.1 shows the locations. The samples were collected, as near the valley bottom as possible, at sites where sediment was exposed in undisturbed sections. While this number of samples is not statistically significant, it is possible to make some general interpretations. The samples were not evenly distributed along the length of the valley because there was limited availability of acceptable exposures. Most of the samples were taken in the central part of the valley. These data are presented in several tables and graphs, as part of an overall discussion of the valley sedimentology. Detailed site descriptions from within and around the valley will be presented in the rest of section 4.1, followed by descriptions of the sites examined

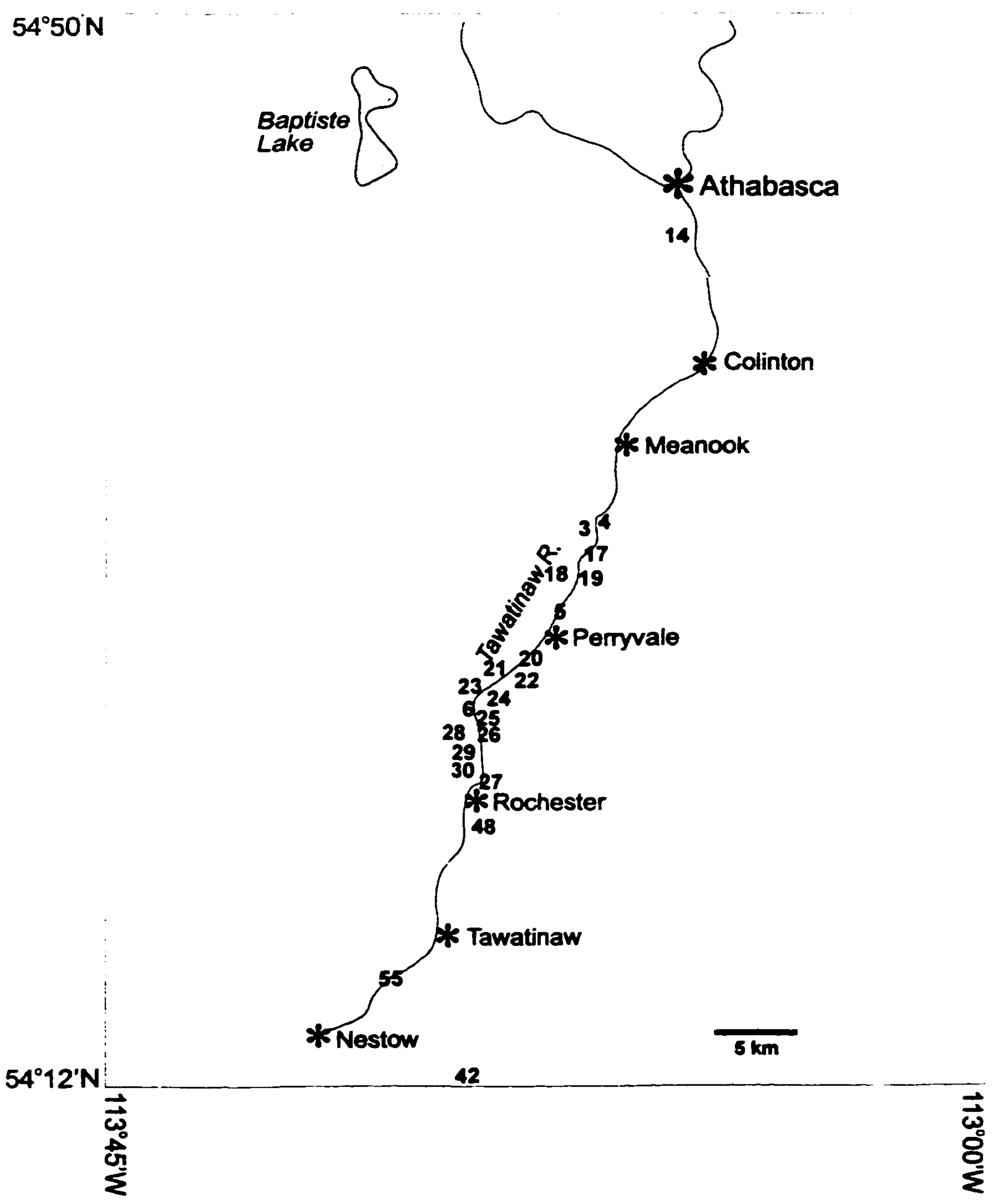


Figure 4.1 Locations of the twenty-two samples taken down the length of the valley. The samples were collected at sites where sediment was exposed in undisturbed sections, as close to the valley bottom as possible.

within the glaciofluvial complex in section 4.2.

Table 4.1 displays the percentages of pebbles, granules, very coarse sand, coarse sand, medium sand, fine sand, very fine sand, and silt and clay for each sample taken down the valley. Raw data containing sample weights and weights of each grain size per sample are presented in Appendix A, for both the valley samples and other important samples taken over the course of the project.

Figure 4.2 is a set of graphs plotting the percentage of the small fraction (silt, clay, and very fine sand) against distance down-valley. The first graph (A) shows the data for silt and clay content and the second graph (B) shows the data for very fine sand. Both clearly show that the sample taken farthest north up the valley is by far the one with the highest small particle content (18.8% silt and clay and 22.5% very fine sand). They also show a general decrease in small particle content progressively southward. The r^2 values of 0.4 and 0.3 are quite low, meaning that there is a high degree of variance between the samples. However, with the sample size of 22, there is still significance at the 0.01 level for both; 't' tests also conclude that both graphs show significant relationships, even down to the 0.001 level (Hicks, pers. comm. 1999).

Figure 4.3 is a set of graphs plotting the percentage of larger particles (granules and pebbles) against distance down valley. The first graph (A) shows the data for pebbles and the second graph (B) shows the data for granules. Both graphs indicate that the samples reflect a coarsening trend with distance down-valley. The regression line for the pebbles represents an r^2 value of 0.3; the

Sample #		distance down-valley	pebble %	granule %	v. c. sand %	coarse sand %	med. sand %	fine sand %	v. fine sand %	silt & clay %
14	V1	2.64	0.0	0.3	2.3	13.8	16.8	24.6	22.5	18.8
4	V2	23.69	0.0	1.3	2.1	11.6	46.7	21.3	5.8	11.0
3	V3	23.70	0.0	0.1	0.1	0.5	64.8	33.0	0.6	0.6
17	V4	28.00	7.0	10.4	14.3	39.3	23.2	3.8	0.8	0.7
18	V5	29.64	2.5	3.2	7.8	28.3	27.0	16.9	8.6	5.1
19	V6	29.65	3.7	6.1	6.4	34.3	45.9	2.5	0.5	0.4
5	V7	31.14	0.0	20.4	24.1	17.2	12.2	8.8	8.8	8.2
20	V8	32.19	0.0	0.6	1.9	43.9	53.0	0.7	0.1	0.1
21	V9	32.84	0.0	1.1	4.8	12.7	48.9	18.1	4.6	9.3
22	V10	32.85	0.0	0.1	0.3	1.9	19.7	75.0	1.9	0.9
23	V11	33.29	0.0	0.9	1.4	8.8	54.4	32.6	1.1	0.5
24	V12	33.30	0.0	0.1	0.1	2.6	71.2	24.1	0.9	0.7
6	V13	37.64	0.0	0.8	1.1	4.9	68.0	21.6	1.5	2.0
25	V14	38.44	0.0	0.2	1.2	31.4	62.8	3.9	0.2	0.1
26	V15	38.64	0.0	0.1	0.2	4.3	72.0	22.3	0.7	0.4
28	V16	39.64	5.3	2.6	4.9	11.6	29.2	35.4	7.4	3.2
29	V17	41.09	0.0	1.7	5.9	5.8	13.8	69.1	2.7	0.2
30	V18	41.44	44.2	13.3	12.8	18.9	5.9	2.7	1.1	0.9
27	V19	41.94	0.0	0.3	1.6	35.0	54.5	7.6	0.4	0.2
48	V20	43.79	10.2	5.7	15.5	34.2	23.2	6.7	2.1	2.2
55	V21	56.69	0.0	0.1	0.3	2.5	40.6	53.1	2.9	0.4
42	V22	66.95	53.3	16.5	9.7	8.7	8.1	2.7	0.7	0.3

Table 4.1 Table displaying the percentages of each grain size for the samples collected down the valley (distances are expressed in kilometres). As is conventional, the pebbles are <64mm to 4mm; the granules are <4mm to 2mm; the very coarse sand is <2mm to 1mm; the coarse sand is <1mm to 0.5mm; the medium sand is <0.5mm to 0.25mm; the fine sand is <0.25mm to 0.125mm; the very fine sand is <0.125mm to 0.0625mm; the silt and clay is <0.0625mm. The data are graphed in subsequent figures and are discussed in the text.

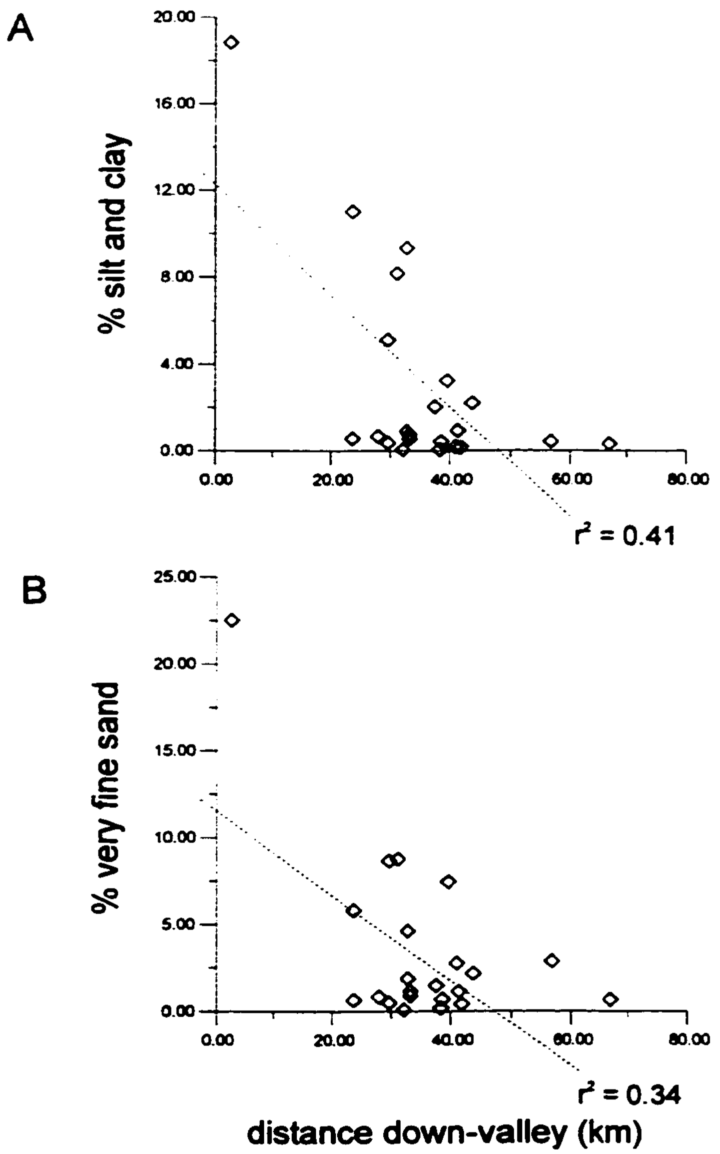


Figure 4.2 Set of graphs plotting the finest grain sizes measured in each of twenty-two samples against distance down-valley. Graph A plots the amount of silt and clay. The graph shows a general decrease in silt and clay content progressively southward; the sample taken farthest north up the valley is the one with by far the highest silt and clay content (18.8%). The regression line represents an r^2 value of 0.41. Graph B plots the amount of very fine sand. The regression line represents an r^2 value of 0.34, and also shows an overall general decrease in very fine sand content progressively southward. The sample farthest north again stands out, having by far the highest very fine sand content (22.5%).

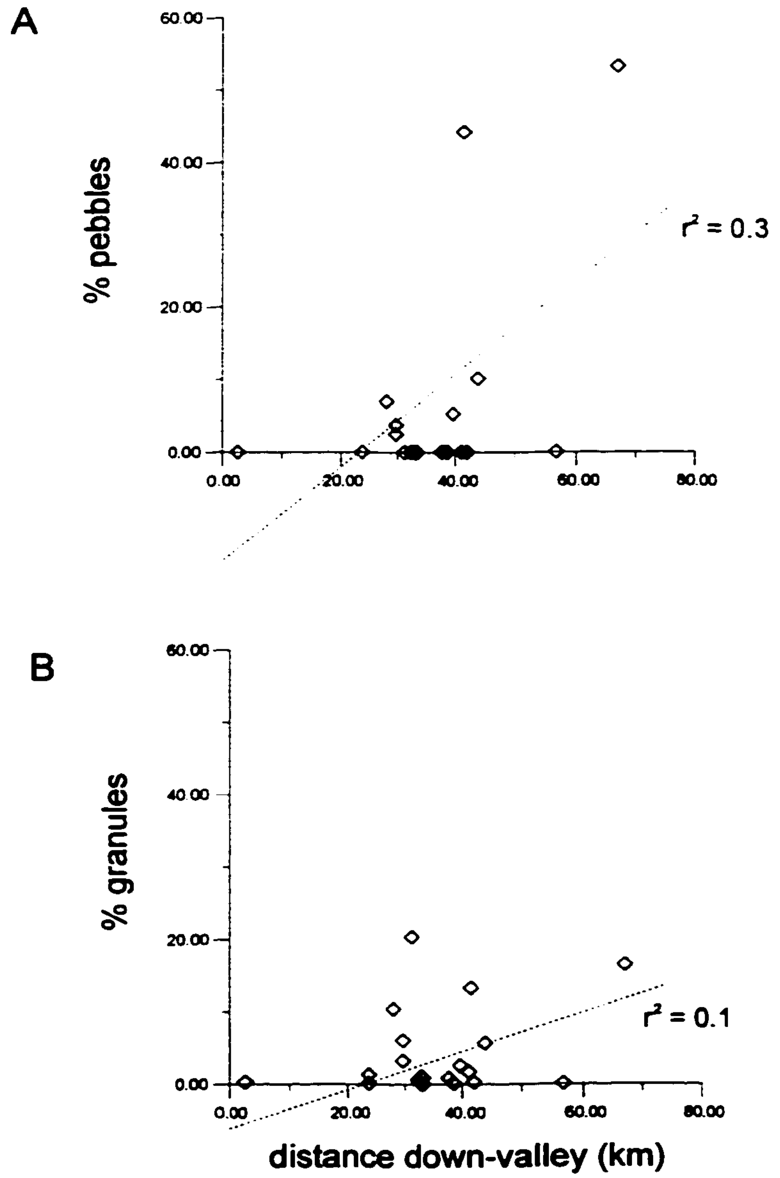


Figure 4.3 Set of graphs plotting the coarsest grains observed in the twenty-two valley samples against distance down-valley. Graph A plots the amount of pebbles against distance and graph B plots the amount of granules against distance. The regression lines for both plots show a general trend of increasing grain size progressively southward; the r^2 values are 0.3 and 0.1 respectively.

regression line for the granules represents a r^2 value of 0.1, which is extremely low. This means that while the graphs appear to show a coarsening trend, there is too much variance between the samples such that a definitive interpretation cannot be made. The correlation analysis for the pebbles is significant only at the 0.05 level, and a 't' test concludes that there is a significant relationship down to the 0.01 level only; all tests show that the trend shown in the granule graph is not statistically significant (Hicks, pers. comm. 1999).

Figure 4.4 is a series of three graphs plotting coarse sand, medium sand and fine sand against distance down valley. They show no obvious coarsening or fining of sand-sized particles down the length of the valley. Individual samples range from very little to relatively high proportions of sand within a few kilometres of each other. The r^2 values of zero or near zero reflect this variability.

4.1.2 Valley Sands

The sediment type observed most frequently within the Tawatinaw valley was sand. Bedding was not observed in most of the exposures, though ripples and cross-beds were present at some locations. Many of the sand exposures were orange/brown in colour, though others are more brownish; the orange/brown sands are believed to have been stained by iron weathering. Commonly, these sands contained granules, and in some instances were associated with a surface deposit/lag of larger clasts. The major sand deposits observed within the valley are discussed below.

At the edge of a steep slope overlooking the valley from the east (UTM

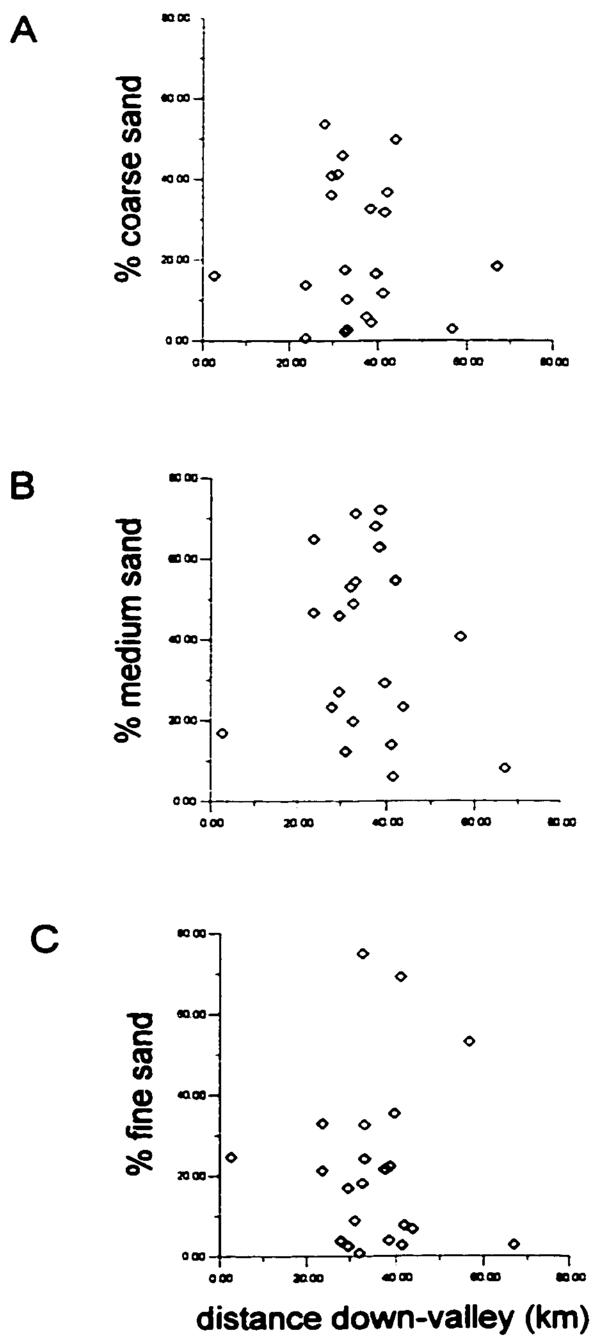


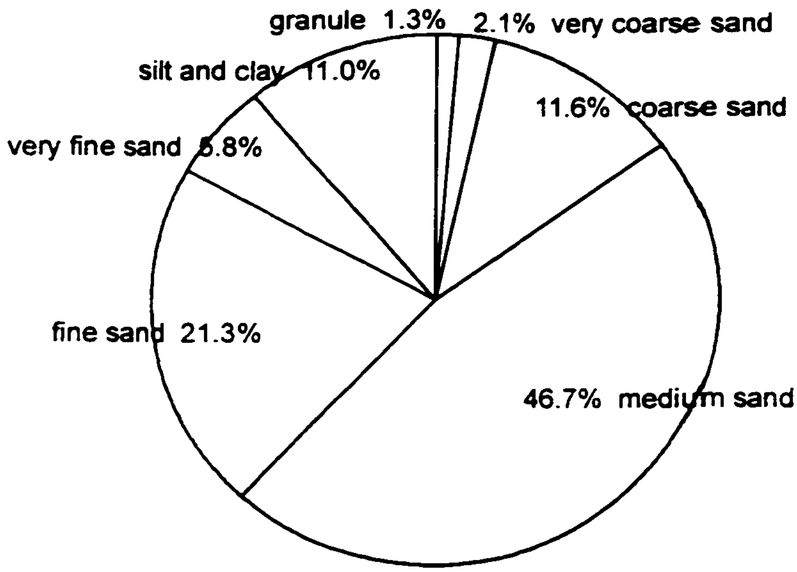
Figure 4.4 Series of graphs plotting intermediate grain sizes against distance down-valley. Graph A shows coarse sand, graph B medium sand, and graph C fine sand. All three plots show that there is no distinct fining or coarsening trend of these particle sizes within the valley. The r^2 values of zero or near zero for all three grain sizes reflect the variance between the samples.

485445), the entire surface consisted of orange/brown sand. The site was located approximately 6 kilometres south of Meanook, and was an abandoned pit. The sand deposit was extensive. It was exposed from the crest of the valley side, just west of the road, to approximately halfway down the valley wall. Its north-south extent was several hundred metres. Throughout the region for many kilometres, however, similar orange/brown sand was exposed at the side of the road, so the deposit was considered to be on the order of at least 5 kilometres in extent. Figure 4.5 is a photo showing the sand, which contains some granules. No bedding was observed. Samples of the sediment show it to be dominated by medium-grained sand. Figure 4.6 shows the results of the sieve analysis.

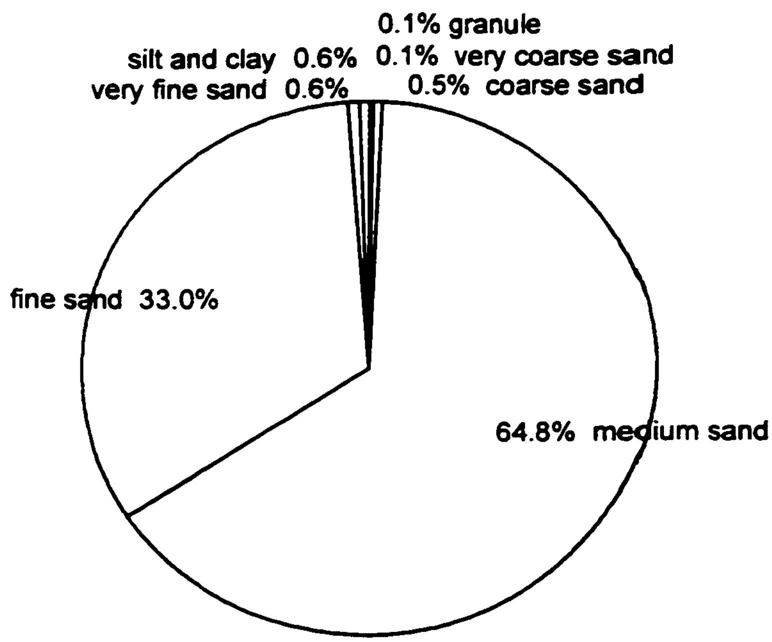
Throughout the Perryvale area, red/brown sand containing small granules was observed in numerous roadcuts along the valley floor. This was a continuation of the sand deposit described from the Meanook area, likely with more iron content. Samples taken (presented in Figure 4.7) show them to be dominated by medium-grained sand. Gravel, cobbles and small boulders were also frequently observed at the surface, above the sand deposit.

One exposure north of Rochester (UTM 419339) along the east-facing wall contained trough cross-bedded granular-sand. This exposure was in one of the bars described in section 3.1. The exposure was approximately 3 metres high, and extended for approximately 15 metres (Figure 4.8) Palaeocurrent measurements were taken from the crossbeds and are plotted in Figure 4.9. The palaeocurrent information indicates that the flow that deposited the sediments was towards the

Figure 4.5 Photo of portion of extensive sand deposit located near Meanook. The photo is taken at a site approximately 25 metres down the valley side. The deposit is believed to extend at least 5 kilometres down the valley, and the exposure shows that it has a depth of many metres. The deposit has no bedding structures. The medium-sized sand contains small granules. The results of the sieve analysis of this sediment are presented in Figure 4.6.



Sample 4 (V2)



Sample 3 (V3)

Figure 4.6 Set of pie charts graphing sieve analysis of two samples taken from extensive sand deposit located approximately 6 kilometres south of Meanook. The deposit is visible from just below the crest of the slope of the eastern edge of the valley, and exends down the slope to at least halfway down the valley wall. The deposit is considered to be on the order of at least 5 kilometres in north-south extent. No bedding is evident. The samples show that the deposit is dominated by medium-grained sand.

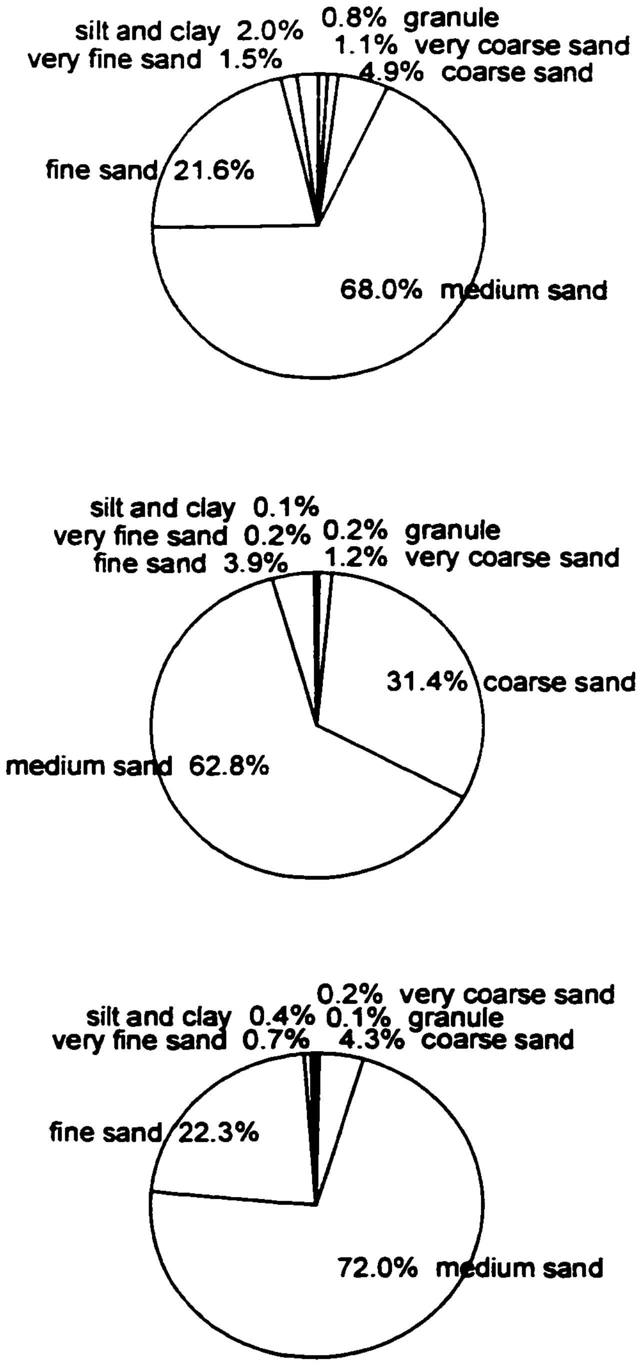


Figure 4.7 Three samples (originally numbered 6, 25, and 26) collected in the Perryvale area. The samples are dominated by medium-grained sand.

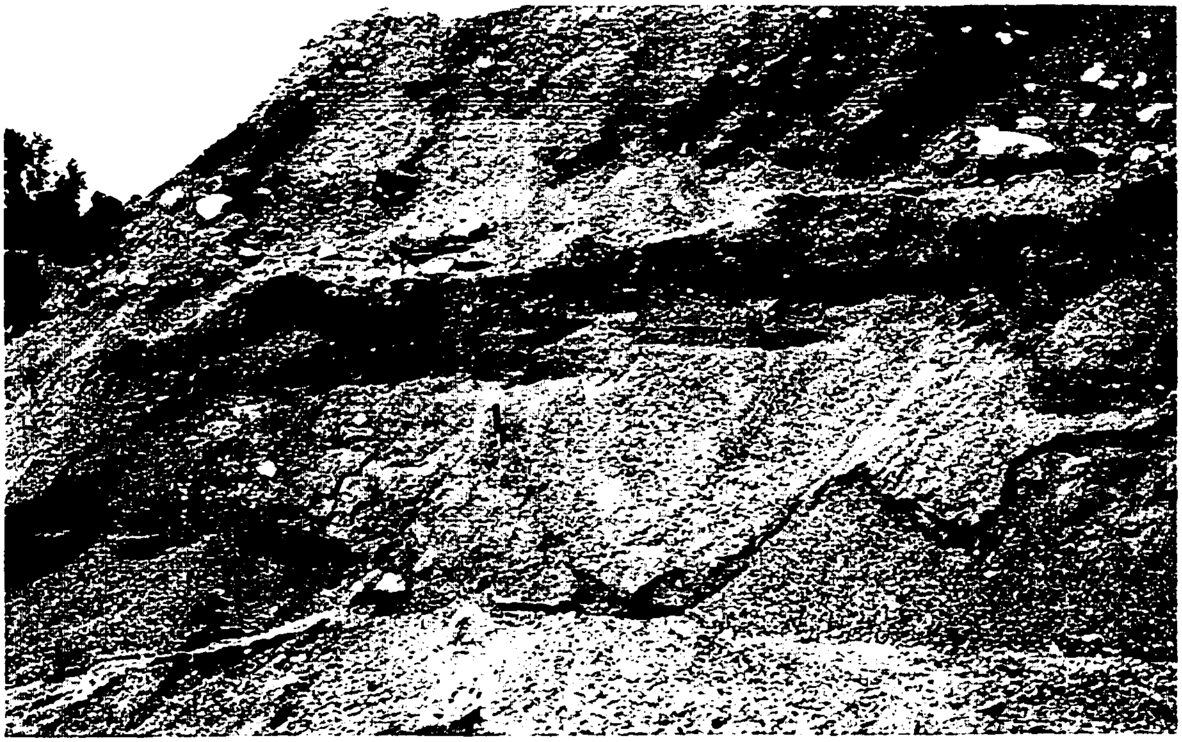


Figure 4.8 Photo of trough cross-bedded sands near Rochester. Note that the exposure had undergone much slumping in time following that in which the exposure was studied. The sediment is a granular sand. The exposure is approximately 3 metres high and extends for approximately 15 metres. Paleocurrent measurements were taken from the crossbeds and are illustrated in Figure 4.9. Note that the material above and behind the exposure is not *in situ*.

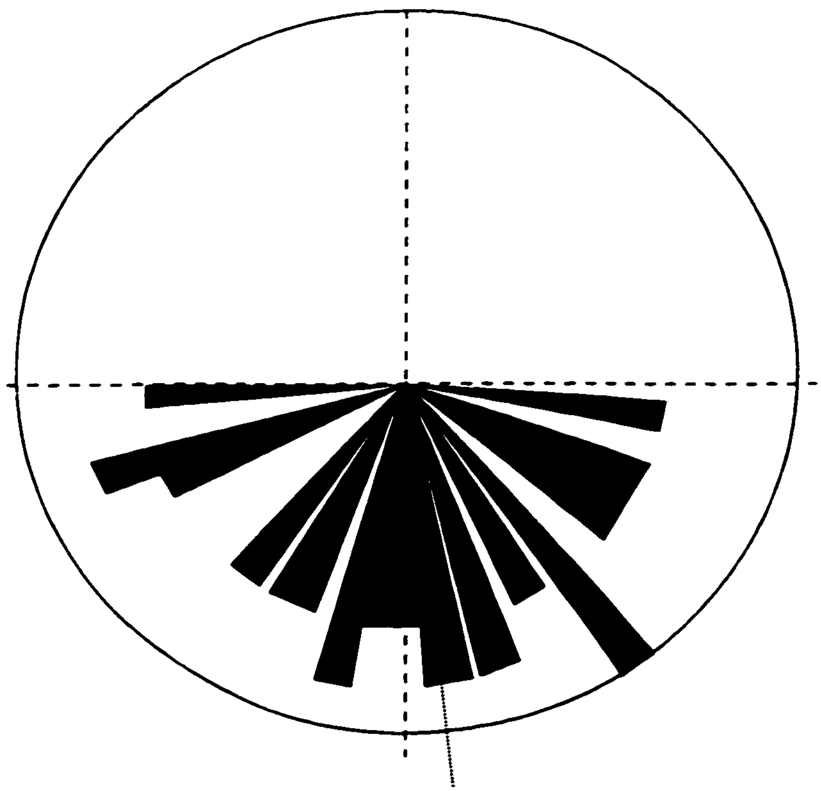


Figure 4.9 Palaeocurrent meaurements taken from a trough cross-bedded granular-sand exposure located north of Rochester. The pit is located high up the valley side. The cross-bedded sands are stratigraphically above plane-bedded/massive sands and below imbricated/unstructured cobbles and surface boulders. The palaeocurrents of the cross-beds indicate that the direction of flow was generally toward the south (number of observations = 20).

south. Elsewhere in the pit, plane bedded sands were located beneath a chaotic sand/pebble/cobble unit. These sands were found well below a surface unit of imbricated cobbles and surface boulders. Figure 4.10 is a photo of these plane-bedded sands.

Another exposure containing bedded sand was observed near Rochester (UTM 412289). This exposure was nearer the valley floor, to the east. The exposure was approximately 7 metres high, and showed trough cross-bedded sands with superimposed ripples in the mid- and upper section; discontinuous plane beds were observed below. The exposure was fractured or faulted in many places. Figure 4.11a plots the palaeocurrents measured in the ripples. The palaeocurrents indicate that they were deposited in a multitude of directions. They were interpreted to have formed in localized eddies.

Immediately to the south (UTM 408288), another exposure of fractured cross-beds and ripples was present. This exposure was approximately 10 metres high and was also situated near the valley floor to the east of the present stream. The palaeocurrents of the ripples are shown in Figure 4.11b, and again the multidirectional form was interpreted to represent local changes in current direction.

It was noted that progressively southward through the valley, the quantity of pebbles, cobbles and gravels observed at the surface visibly increased. These larger clasts were typically subrounded to subangular, and were believed to be associated with the glaciofluvial complex at the southern end of the valley.

The presence of sand throughout the valley, including high up the valley

Figure 4.10 Photo of plane-bedded sands, near Rochester. These sands are found in the same pit as the cross-bedded granular sands shown in Figure 4.8. They are observed below a chaotic sand/pebble/cobble unit which is elevationally below the level at which the cross-bedded unit was found.

a

b

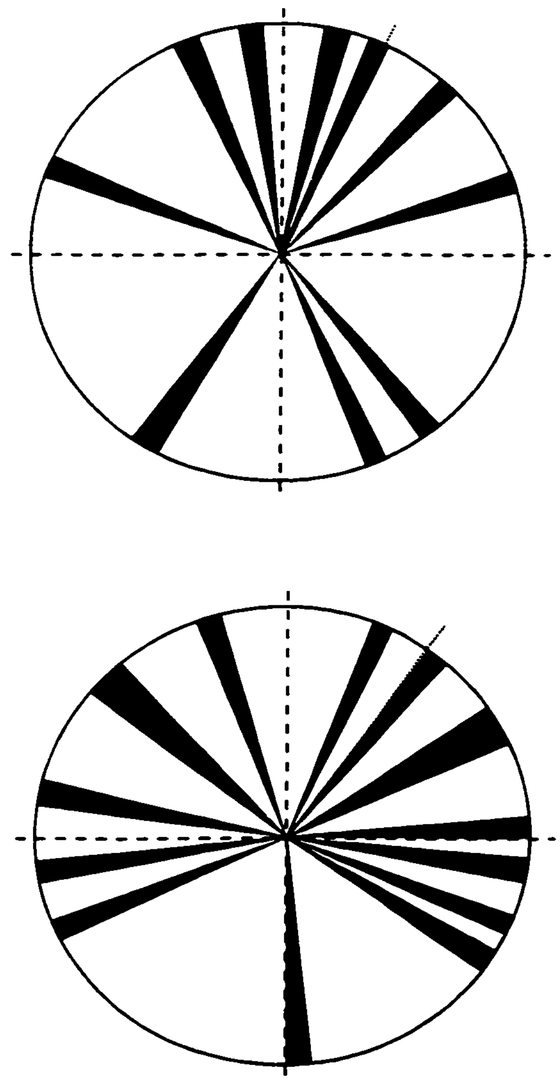


Figure 4.11 Set of rose diagrams showing the palaeocurrent measurements taken from two ripple cross-laminated sand exposures. The multidirectional nature of the ripples is interpreted as showing that they were deposited in localized eddies, and do not reflect a large-scale flow pattern (number of observations (a) = 10, number of observations (b) = 13).

sides as observed in many exposures, is important. The deposition of sand at the floor of the valley and along the walls could be interpreted as an indication that the valley was, at least for a while, at near bankfull conditions. As expressed in section 2.3.1, a subglacial tunnel at bankfull conditions is customarily referred to as a tunnel channel rather than a tunnel valley (though many researchers still use the terms interchangeably).

4.1.3 Other Sediments Observed in the Valley

As stated, the Tawatinaw valley is dominated by sand deposits, commonly containing granules and in some cases overlain by larger clasts. Other sediments observed in the valley include: clays and finer sands, bedrock, and diamict/till. These are discussed in turn below. Sediments observed in the remnant ridges and on the surfaces to either side of the valley are discussed in section 4.3.3.

4.1.3.1 Clays in the North

Immediately south-southeast of the town of Athabasca was a large, abandoned pit (UTM 544633). The pit contained a large deposit of rhythmically bedded clays, most of which had been removed, though some clay was still present on the floor of the pit. The walls of the pit were up to two metres high, and it covered an area of several hundred square metres. It was interpreted that these clays were probably deposited in a glacial lake, formed when water ponded in front of the glacier as it retreated. The water would have collected in this northern portion of the valley as it was the lowest point (refer to Figure 3.2). If this interpretation is correct, this is evidence that the valley formed prior to the last stages of deglaciation as the valley had to pre-exist in order to fill with ponded water.

Sample 14/V1, taken from near the side of the road approximately 3 kilometres south of Athabasca (UTM 536633), was found to have a very high clay and silt content in comparison to the other valley samples; the sieve analysis of this sample is presented in Figure 4.12. This sample is the anomalous point in Figures 4.2 and 4.3. If these sediments were deposited in a proglacial lake, then they are unrelated to the rest of the valley sands.

4.1.3.2 Bedrock

To the east of Rochester, approximately one quarter up the valley side (UTM 399273), bedrock was observed by the side of the road. The bedrock was stratigraphically below the granular sand deposit that dominated the slope; it was concluded that the bedrock was part of the original valley wall, prior to sand deposition in the tunnel valley/channel. No other bedrock was exposed within the valley.

4.1.3.3 Till / Diamict

At several locations in the Meanook area, diamict was observed. The sediment was clay-like, and yellowish brown in colour. It was crumbly, and contained subangular to subrounded clasts (of predominantly quartzites and granites) ranging in size from less than a centimetre to up to 40 centimetres. Sand stringers were observed in some exposures, and the sediment was deformed in some places.

Fabric data collected from an exposure approximately 1 kilometre east of

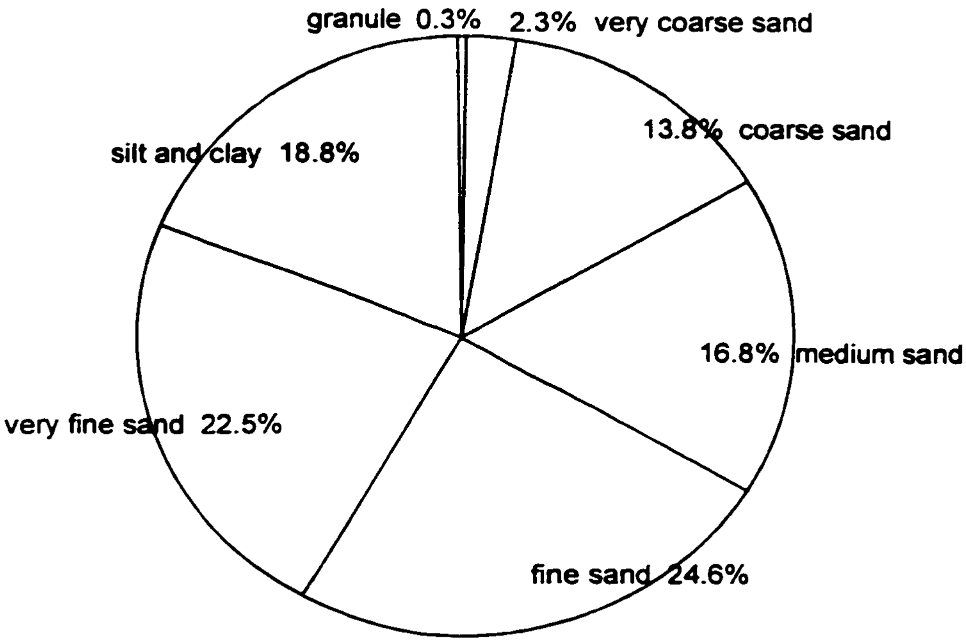


Figure 4.12 Pie chart graphing sieve analysis of a sample taken from a site in close proximity to a clay pit at the northern end of the valley. The sample is the only one in the valley that contains a high proportion of silt and clay particles (18.8%).

Meanook (UTM 507502) are presented in the stereoplot shown in Figure 4.13. The exposure was located near the crest of the valley wall, and appeared to be deformed. The contoured diagram has a bimodal distribution (Hicock et al. 1996). The principle eigenvalue of 0.496 is fairly low (due to this bimodality), and leads to the conclusion that the deposit is either not a till or is a till altered by deformation. A lodgement till normally has a 1st eigenvalue of at least 0.6, while a meltout till usually has a value of approximately 0.8 or greater; a fabric with a very low value is more likely a flow deposit (Dowdeswell and Sharp 1986, Hicock et al. 1996). Because the exposure was observed to appear deformed and because it had the visual characteristics associated with till, it is concluded that the original deposit was likely a till.

Immediately west of Perryvale (UTM 452383), close to the valley bottom, till was observed in an exposure by the side of the road. Figure 4.14 is a photo of the exposure, showing a pod of till within a sand and cobble matrix. In the same exposure, some bedded sand was observed, as well as massive sand containing pebbles and cobbles, and a deposit of boulders at the top. Because the till was observed in a pod, it was believed that the material was transported and deposited within the valley.

4.2 Glaciofluvial Complex

As described in Chapter 1, the glaciofluvial complex is a large, fan-shaped feature that extends from the southern end of the valley, deflecting in a

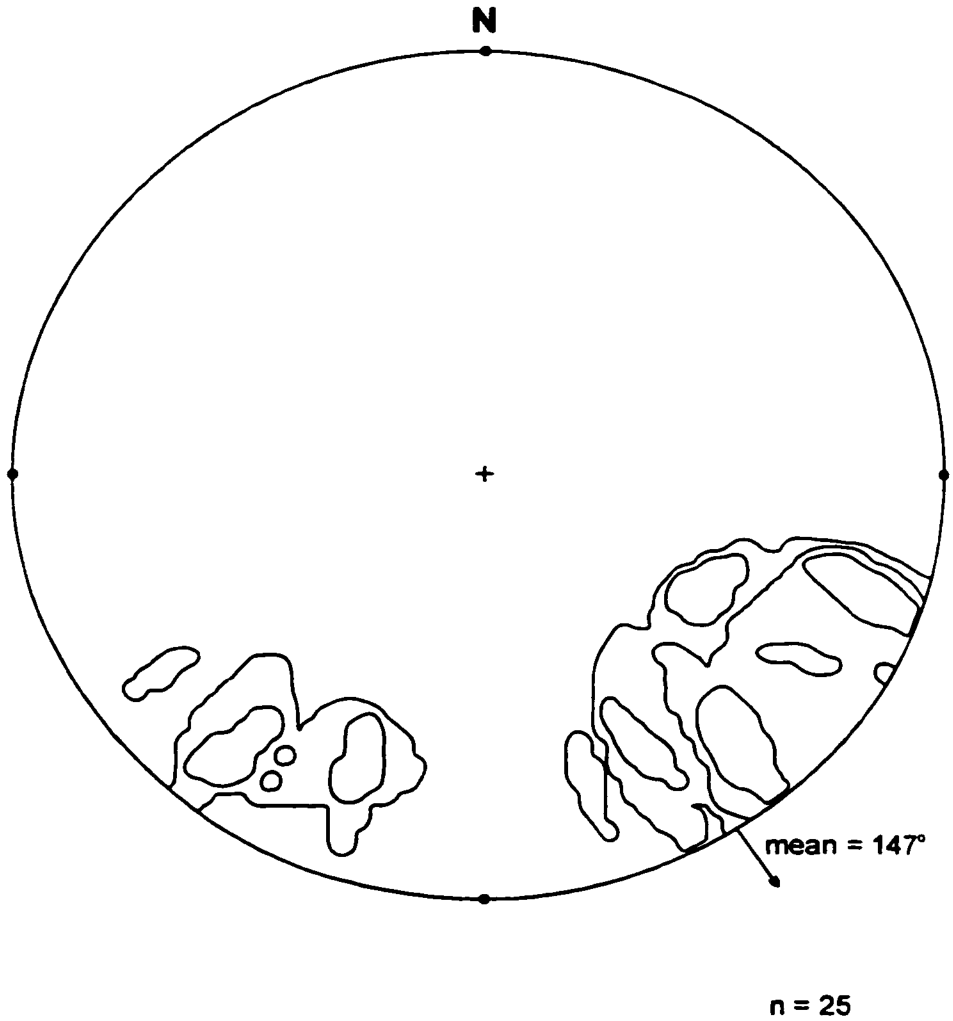


Figure 4.13 Stereonet plotting the fabric measured from a diamict section near Meanook. The exposure is located close to the crest of the eastern valley wall, and appears to be somewhat deformed. The principle eigenvalue of 0.5 is very low - tills normally have a 1st eigenvalue of at least 0.6. It is therefore concluded that this deposit is either not a till, or a till that has been deformed.

PHOTO A

PHOTO B

Figure 4.14 Photos of exposure immediately west of Perryvale. A pod of till is observed near the top of the exposure (Photo A). The exposure is dominated by a diamict of sediment ranging from fine sand to cobbles. A layer of boulders is observed near the top of the exposure (Photo B.). A small pod of bedded sand is also observed, and appears as though it has undergone some deformation.

southeasterly direction. The complex was dominated by gravel deposits (some showing bedding structures and others massive). Sands (structured and unstructured), boulders, and aeolian deposits were also observed. The remainder of this section presents descriptions of some of the sediments observed in the complex.

4.2.1 Cross-bedded Gravels

Numerous exposures throughout the region revealed trough cross-bedded gravel-dominated deposits. Figure 4.15 shows palaeocurrent data collected from three such exposures; the Lafarge pit (a) was located at UTM 403978, the Westlock Sand and Gravel pit was located at UTM 424921, and the Zilinski pit was located at UTM 429933 (all were approximately 10 kilometres southeast of Clyde, which is indicated on several figures). The measurements of the foreset beds clearly indicated that the water that deposited the sediment was flowing in a southeasterly direction. The three exposures described above were the largest and best-exposed examples of cross-bedded gravels in the glaciofluvial complex. Other, smaller and/or more poorly exposed, exposures also revealed cross-bedded gravels indicating flow in generally south, east or southeasterly directions.

4.2.2 Non-structured Gravels

Massive gravels were also observed throughout the glaciofluvial complex. The gravels were commonly contained within a coarse, sandy matrix, and exposures were on the order of at least several metres high. No bedding structures were observed on the exposed surfaces, and the clasts did not appear to have a

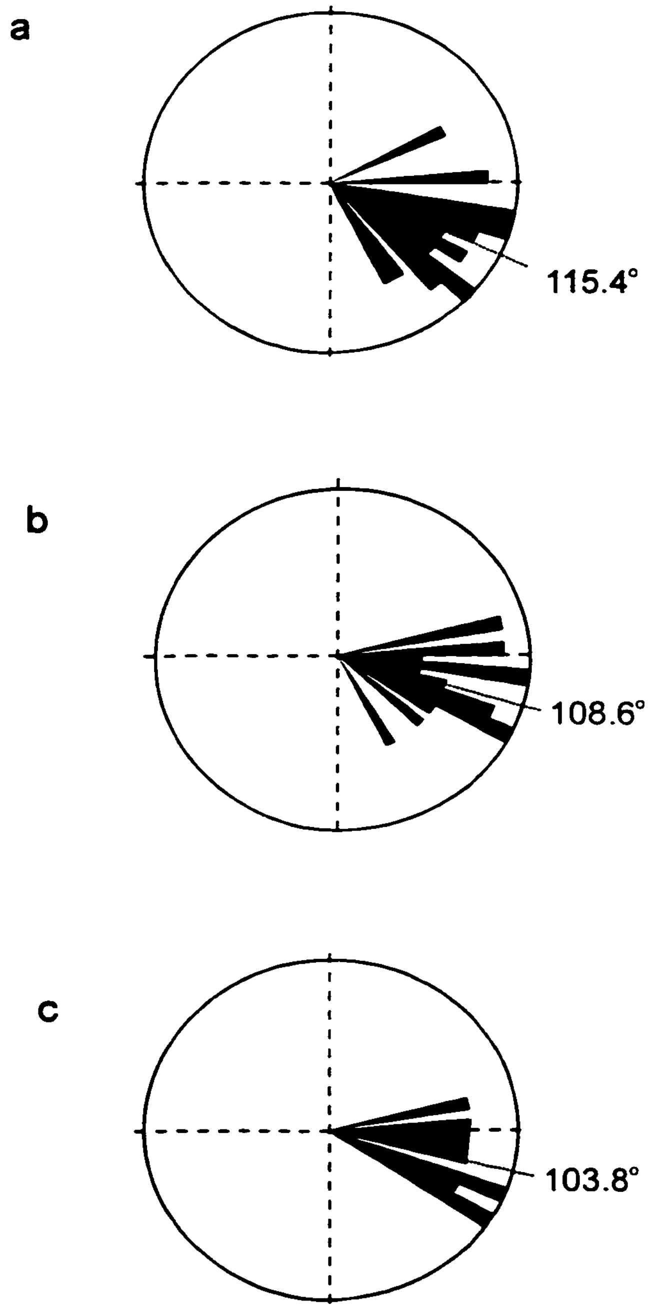


Figure 4.15 Set of rose diagrams showing the palaeocurrents measured from three pits in the glaciofluvial complex to the south of the valley (the Lafarge pit, a, the Westlock Sand & Gravel pit, b, and the Zilinski pit, c. All three exposures show that the sediment was deposited by flow in a southeasterly direction.

preferred orientation.

4.2.3 Other Sediments Observed in the Glaciofluvial Complex

4.2.3.1 Sands

While gravel contained within coarse sand matrices was by far the dominant sediment within the glaciofluvial complex, sand deposits were also observed. In some cases, the sand was stratigraphically below the gravels. Some of the sand had bedding structures (plane beds and/or ripples), while some was massive. At some locations the unstructured sand features were suspected to have been translocated by aeolian processes.

4.2.3.1.1 Bedded Sands

Several exposures in the glaciofluvial complex contained complex bedded sand deposits. One of the larger exposures, located at UTM 360079, was an abandoned pit site. It was on the order of 10 metres high and approximately 30 metres long, and a portion of it is shown in the photo to the right of Figure 4.16. The sketch to the left of the photo is a generalized composite of the sediments observed throughout the section. The lowest unit observed revealed large foresets of sand; the same unit also contained Type A and Type B ripples. It is likely that the ripples were superimposed on the foresets, as part of a large subaqueous dune. Above this, a discontinuous laminated clay deposit was present. The next unit was trough cross-bedded silty sand, which visibly coarsened and graded into a plane bedded sand unit above. Type A and Type B ripples were observed above the plane beds. Palaeocurrent measurements in this ripple cross-laminated unit reflected

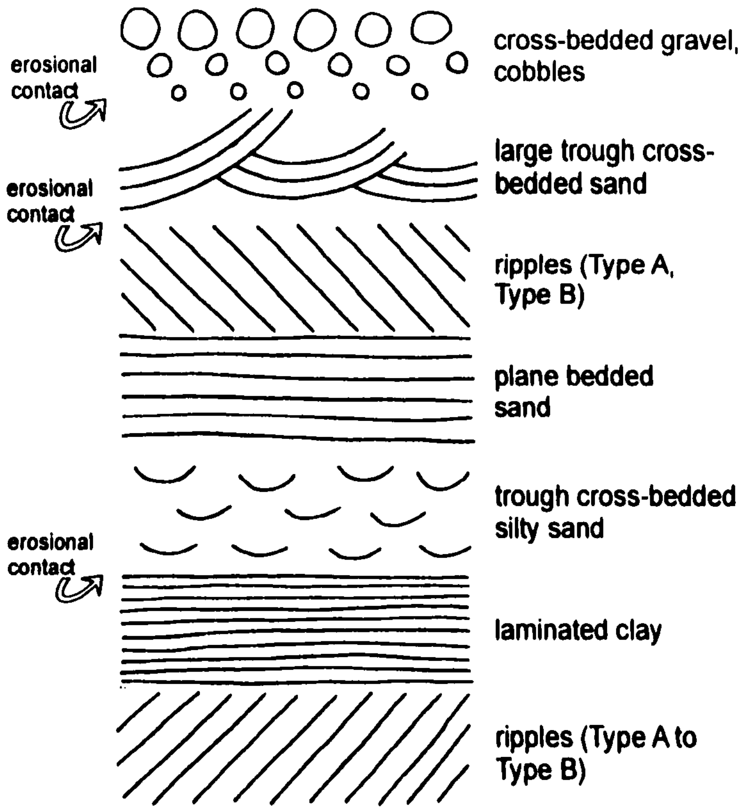


Figure 4.16 Photo of portion of large sand deposit located at the northern end of the glaciofluvial complex. It is approximately 10 metres high, and extends for approximately 30 metres. The exposure contains cross-bedded gravels above trough cross-bedded coarse sand, in turn above above zones of plane beds and ripples in finer sands. Observations of the palaeocurrents within the various cross-laminated ripple sets reveal multidirectional flow. The composite sketch to the left summarizes the stratigraphy of the sediments observed in the exposure. The units grade from one to the next, except for the large trough cross-bedded sands and the cross-bedded gravels/cobbles which have erosional contacts.

multidirectional flow, and it was believed that these structures were the products of localized flow. Large trough cross-beds of coarse sand were present above the rippled unit, with an erosional contact between the two. Cross bedded gravels and cobbles were above the large troughs, again with an erosional contact.

Exposures on a smaller scale, some as complex as the one described above and some simpler, were observed throughout the glaciofluvial complex. Many show gradations between Type A and Type B ripples and plane beds, while others reveal only a single bedding structure.

These complex sand exposures show that flow in the area was variable. Flow of subcritical stream power would have been necessary to form the dune structure and current ripples observed at the base of the large exposure pictured in Figure 4.16 and described above (Allen 1982). Flow power then dropped to form the laminated clay unit. Flow power gradually increased to form the trough cross-bedded fine sands, the plane bedded sands, and the Type A and Type B ripples; then there was a marked increase in power to near supercritical flow to form the large trough cross-bedded sands and the cross-bedded coarse clasts above.

4.2.3.1.2 Aeolian Sands

Throughout the area toward the western edge of the glaciofluvial complex, sand deposits have a dune-like shape in airphotos; the hills of sand are almost crescent shaped. It is believed that many of the sand deposits in this region represent aeolian transportation of glaciofluvial sands related to the tunnel valley (Shetsen 1990).

4.2.3.2 Boulders

Boulders were occasionally observed at the surface in the glaciofluvial complex. Most were observed in positions that were almost definitely not *in situ*. They were most commonly subangular and some were striated. The boulders were possibly unrelated to the flow that deposited the gravels and sands in the complex; they could have been left behind during ice melt.

4.3 Sediments Observed in Other Locations in the Study Area

4.3.1 Bedrock to the South

The entire surface of the region to the south of the glaciofluvial complex (from the Waugh area south to near the Bon Accord/Gibbons area) was dominated by bedrock exposures. This bedrock shows that the region is virtually devoid of glacial or glaciofluvial deposits, and is believed to have been scoured clean - likely by meltwater.

4.3.2 Boulders to the East

Throughout the region to the east of the Tawatinaw valley, boulders of varying sizes were seen scattered over the surface. The boulders ranged in size from 50 centimetres up to a metre or more. They were observed in nearly every untreed field/pasture in differing densities. Those boulders closely examined tended to be striated, and were believed to have been left behind at the surface after the retreat of the ice.

4.3.3 Either Side of the Valley Walls and Remnant Ridges to West

A sediment sample taken from the top of a hill at the western edge of the Tawatinaw valley (UTM 536627) was found to be predominantly sand with many granules and pebbles, most quite angular. This high area appeared to be either a terrace or slump block above the main part of the valley. Several small, semi-angular boulders with a-axes on the order of 15 - 20 centimetres were present.

Immediately north of Colinton (UTM 528552), above the valley to the west where the remnant ridges are, bedrock was observed to the side of the road. In an exposure cutting through one of the remnant ridges just to the south, diamict pods, massive sands, upturned crossbedded sands, cobbles, and fracturing were observed.

The crest of another remnant ridge was cross-cut further south (UTM 509545). The exposure was dominated by sandy diamict, with subangular boulders up to 80 centimetres in diameter sitting at the surface as a lag.

On either side of the valley walls, sediments were consistently sandy/silty matrices with fairly angular granules, pebbles and cobbles. Boulders were frequently observed, ranging in size up to 40 centimetres. These sediments were chaotic, with no sorting or bedding observed.

Two kilometres west of Colinton (UTM 525545), a huge, striated boulder was observed. This 2 metre high boulder was sitting on a hillside composed of a sandy matrix containing granules, pebbles and cobbles; a roadcut has exposed a cross-section through the side of one of the remnant ridges above the valley. There was a core of silty-clayey sediment with increasing grain size upwards; boulders 60 - 100 centimetres in diameter capped the top. The clasts were subrounded to subangular. Figure 4.17 is a photo of the largest boulder, perched on the side of the hill. Bedrock blocks were present near the base of the hill.

Table 4.2 displays the dimensions of several large boulders observed in the Rochester pit, described in section 4.1.2; the pit is located in one of the bars described in section 3.1. Striations were present on many of the boulders. The larger boulders were more angular than the smaller ones sighted in the same area. Stratigraphically, throughout this area finer sediments were observed below coarser deposits. Some parts of the pit were dominated by sand (with pebbles, cobbles) while there were pods of large cobbles in other places. In many instances, finer sands/pebbles/gravels were capped by coarser clasts.

A bedded gravel/cobble layer was observed in several exposures. The cobbles were on the order of 20 - 50 centimetres, and their imbrication (seen in multiple exposures) reflected a flow direction from the north. The clasts were rounded, and percussion marks were frequently observed. Since the layer was only one metre thick, it was determined that the flow which carried the large clasts was of short duration or primarily erosional except for the last stage.

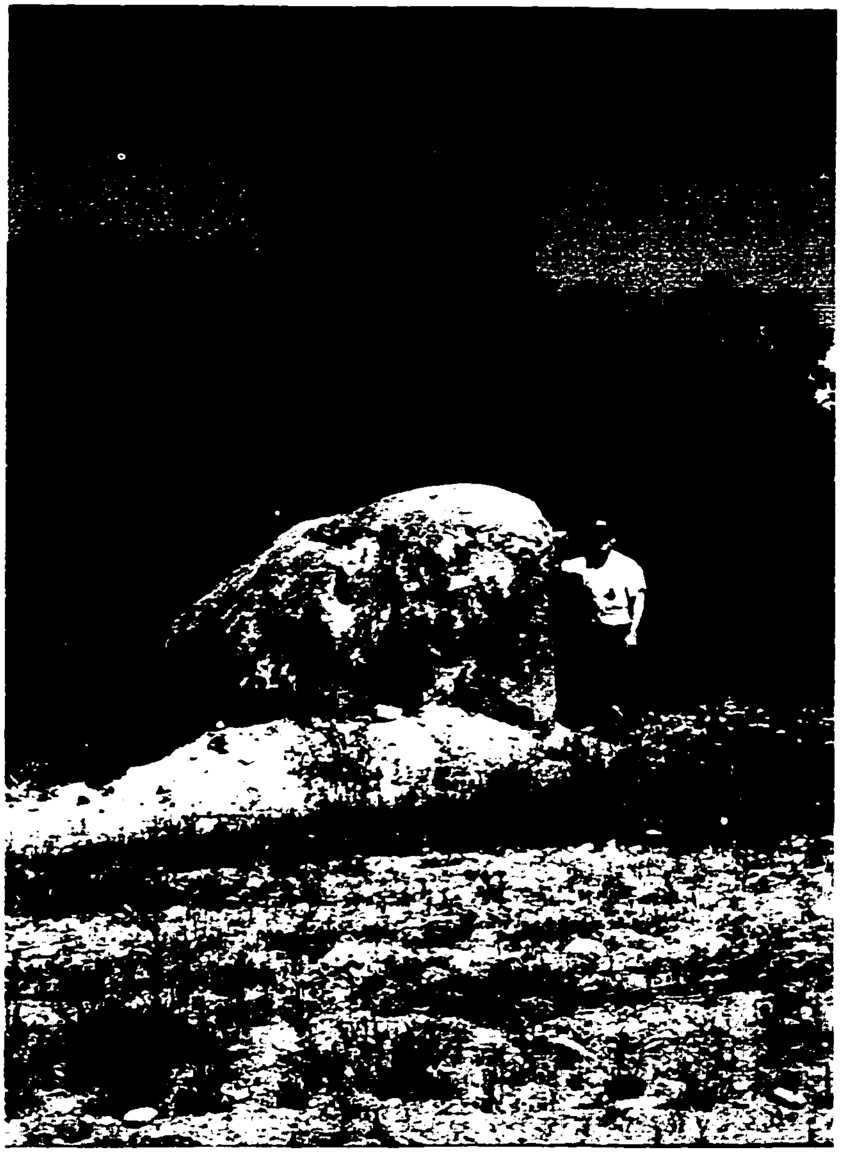


Figure 4.17 Photo of large, striated boulder observed at the surface of remnant ridge above the west side of the valley. The ridge consists of a sandy matrix containing granules, pebbles and cobbles. It appears to have a core of finer material, with the grain sizes coarsening to the outer edges of the ridge. Boulders up to one metre in size cap the top, with the boulder in the photo being the largest, at approximately two metres wide. The clasts are subrounded to subangular in shape.

lengths of axes (cm)				
a	b	c	rock type	shape
115	99	46	gabbro	subangular
133	110	69	homeblende granite	angular
129	87	51	homeblende granite	subangular
104	64	58	granite pegmatite	subrounded
106	76	53	granite pegmatite	subrounded
138	124	85	granite pegmatite	angular
143	143	69	sandstone (local bedrock)	subangular
129	101	62	granite biotite	subrounded
253	138	115	granite pegmatite	angular
232	159	150	homeblende granite	angular

Table 4.2 The dimensions of several large boulders observed in a pit near Rochester. Most of these boulders are angular to subangular, many are striated, and all are composed of Shield material except for one (which is composed of local terrestrial bedrock). None of these boulders was *in situ*. It is proposed that they were likely left at the surface during melt-out, as it is suspected that they are too big to have been deposited by the meltwater flowing through the Tawatinaw valley.

CHAPTER 5: DISCUSSION AND CONCLUSIONS

5.1 Discussion

Data presented on the morphology of the Tawatinaw valley have produced strong evidence in support of the hypothesis that the valley operated as a tunnel valley during the last glaciation. The valley has many of the characteristics common to tunnel valleys: long and relatively straight, steep sides, rising-downflow long profile, smaller hanging tributary valleys, occupied by an underfit stream, and termination in a fan. The sedimentological evidence indicates that water flowed in a southerly (uphill) direction, providing indisputable evidence that subglacial meltwater flowed through the valley.

Isostatic rebound in the region has not been estimated to date. The main body of the Laurentide ice sheet was located to the northeast of the study area (Dyke and Prest 1987) with the region around Hudson Bay undergoing the most depression. Isostatic depression would actually have marginally increased the northwards slope of the valley. Thus, flow was driven up a steeper gradient than that illustrated in Figure 3.4. Therefore the conclusion that water flowing southward in the Tawatinaw Valley must have been under hydrostatic pressure is sound, and cannot be explained by isostatic rebound to account for present day gradients.

5.1.1 Tunnel Valley or Tunnel Channel?

The difference between the sediments within the valley and those in the areas stretching away from the upper valley sides shows that the water was confined within the valley during the depositional phase. There is no evidence for overbank flow, but because sands are observed high up the valley sides, it is proposed that the Tawatinaw valley was an active tunnel channel, operating at least temporarily at full capacity.

5.1.2 Age of the Tawatinaw Valley

The present-day "river" occupying the valley is an underfit stream. Pawlowicz and Fenton (1995) mapped the Tawatinaw as a preglacial valley. If it did exist preglacially, the valley was either much smaller than at present, or was occupied by a significantly larger river. As the neighbouring Athabasca river system has been shown to have existed preglacially (Pawlowicz and Fenton 1995), it is quite likely that a stream in the Tawatinaw region existed as a tributary to it. There was insufficient catchment area for the preglacial Tawatinaw River to have been large enough to have carved a valley as big as the one seen today. It is therefore proposed that the Tawatinaw valley was adopted and enlarged significantly by subglacial meltwaters flowing beneath the last ice sheet.

It is proposed that the Tawatinaw valley was active near the end of the last glaciation. This is supported by the fact that very little till was observed in the sediments, and that which was observed appeared to have been transported in chunks, likely by the same meltwater that carried the sands and gravels. If the valley was occupied only in a prior glaciation or during the early stages of the last, different sediments would be expected.

5.1.3 Consideration of the Various Hypotheses for Tunnel Valley/Channel Formation

The Tawatinaw valley demonstrates no morphological evidence for having been formed in stages over lengthy time. If the Tawatinaw tunnel valley/channel were indeed a preglacial valley, adopted and enlarged by glacial meltwater, this would eliminate time-transgressive mechanisms for formation. The presence of the glaciofluvial complex at the southern end of the valley is similar in appearance to the fans described by Wingfield (1990) and Paterson(1994). However, they both infer time-transgressive formation. There is no evidence in the Tawatinaw valley for progressive headwards collapse of an ice tunnel.

Piotrowski's (1997) model for tunnel valley formation - catastrophic evacuation of subglacial meltwater in catastrophic outburst events triggered by ice retreat - is the most fitting of the hypotheses described in section 2.3.4. Because the Tawatinaw valley stands alone (as opposed to being part of a network of channels), the models proposed by Boyd et al. (1988) and Brennand and Shaw (1994) - progressive channelization of an unstable meltwater sheet - do not appear to be entirely applicable.

5.1.4 Relationship to the Livingstone Lake Megaflood

With respect to the Tawatinaw valley's connection to the megaflood studied by Rains et al. (1993), Shaw et al. (1996), and others, it is useful to examine the province-wide DEM presented in Figure 1.3. The proposed megaflood flow paths are easily seen as interconnected, sinuous erosion zones. The Tawatinaw valley and the neighbouring White Earth valley parallel the outer margins of the Eastern flow path, immediately before the bifurcation into the lower Western and Eastern paths. The valleys' relationship to the flood event(s) can be envisaged. It is possible that the valleys represent waning stage flow from the megaflood. As the volume of water decreased, it would have channelised into lower elevations - and pre-existing valleys would have provided ideal paths for the water.

Pollard et al. (1996) observed that flutes often form downflow from topographic steps. This is seen around the Tawatinaw valley, where the Athabasca flutings are seen south of the Athabasca River and southwest of the Tawatinaw River valley. This indicates that the Tawatinaw valley must have existed, in a form similar to its present appearance, when the flutes formed in order to act as a step. The formation of these flutes has been linked to the megaflood (e.g., Rains et al. 1993, Shaw et al. 1996). It is therefore considered that the tunnel valley may have been active during an earlier stage of the flood event as well as during waning flow. Perhaps the pre-tunnel valley feature was large enough to create the steps necessary to trigger formation of the flutes, as described by Pollard et al. (1996).

Brennand and Sharpe (1993) proposed a similar sequence of events to explain the evolution of a tunnel channel and its association with adjacent flutes and drumlinoid ridges. Their hypothesis is illustrated in Figure 5.1. The deep scours, discontinuous thalweg, sculpted margins, and flutes on the downflow side of the valley wall described by Brennand and Sharpe (1993) are very similar in appearance to the features of the Tawatinaw valley, as seen in the cropped hillshade presented in Figure 3.2. Brennand and Sharpe (1993) interpreted that the

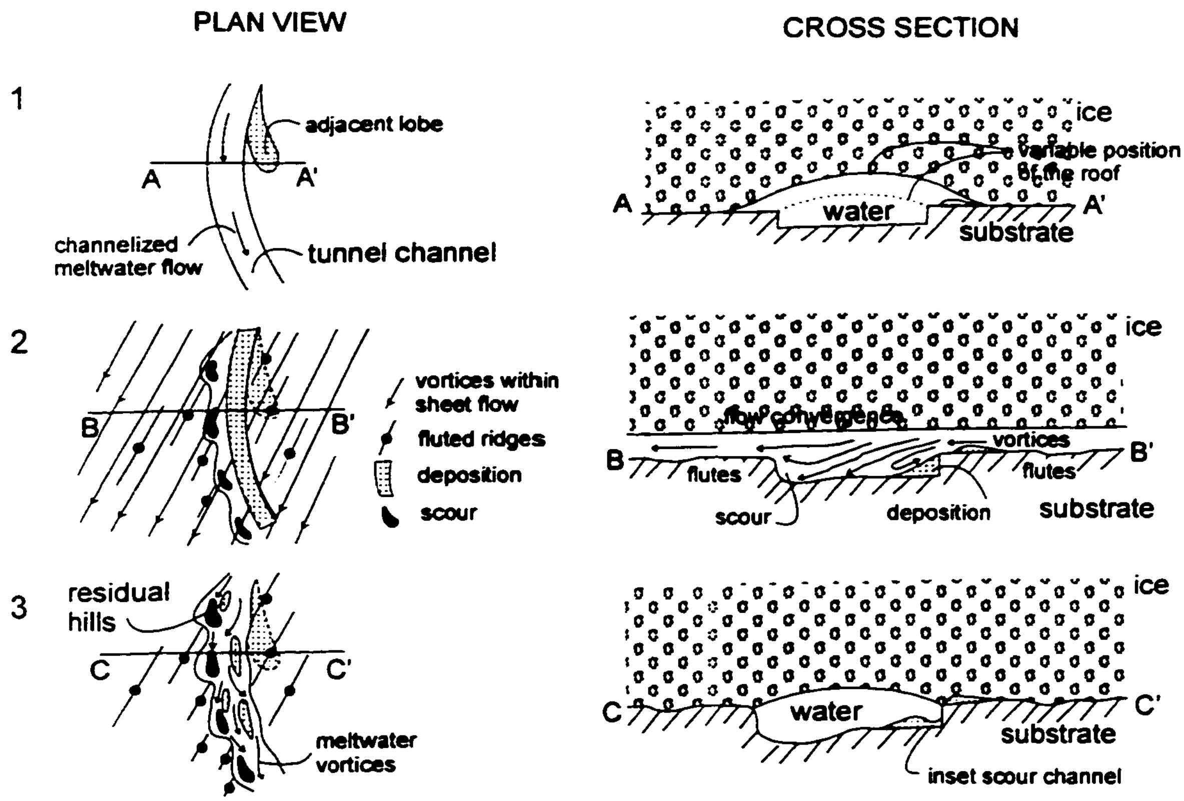


Figure 5.1 Series of diagrams illustrating the evolution of a tunnel channel: 1) tunnel channel (R/N channel) cuts into sediment; overbank deposition creates an adjacent lobe of glaciofluvial sediments. 2) subglacial meltwater sheet flow creates streamlined terrain, scalloped margins of tunnel channel, and crescentic scours and eddy deposits within the channel. 3) the tunnel channel continues to erode, forming shallower scour channels and residual hills. Modified from Figure 6, Brennand and Sharpe (1993).

tunnel channel was the product of complex flow of catastrophic subglacial meltwater. They envisaged the following sequence of events: 1) The channel must have been formed prior to the formation of the flutes, therefore they described the formation of a channel by an earlier channelized subglacial meltwater flow. In the case of the Tawatinaw valley, this stage may not have occurred, as it is possible that the valley existed in some form preglacially. Also during this early phase, as Brennand and Sharpe (1993) described, high meltwater discharges causing localized lifting of the ice from the bed to form a glaciofluvially deposited splay lobe. 2) During the second phase, Brennand and Sharpe (1993) described the formation of streamlined landforms above the channel by a subglacial meltwater sheet. They noted that flow separation and deposition would be expected in the obliquely oriented channel; impingement of vortices on the channel wall may have resulted in the scalloped margin, and flow may also have resulted in crescentic scours. 3) In the third phase, waning flow was confined to the tunnel channel, creating streamlined residual forms. Brennand and Sharpe (1993) went on to describe that as flow within the tunnel channel ceased, ice may have moved into the channel. They also noted that in all likelihood, the tunnel channel probably carried several separate discharge events.

Brennand and Sharpe's (1993) model is very relevant to the Tawatinaw valley. The tunnel channel described by Brennand and Sharpe (1993) is very similar in appearance to the Tawatinaw valley. Both are sinuous, with scallops along one wall and bars/residuals along the other. One significant difference is that the sheet flow described by Brennand and Sharpe (1993) is at a greater angle to the valley than the one was to the Tawatinaw. The Athabasca flutes run almost parallel to the Tawatinaw valley. The sequence of events described by Brennand and Sharpe (1993), however, explain the forms seen in the Tawatinaw area very well.

5.1.5 The Glaciofluvial Complex

The glaciofluvial complex to the south of the valley was either deposited into a subglacial cavity or lake, or subaerially into either a glacial lake or simply out onto the surface. In order for the glaciofluvial complex to have formed subaerially, the tunnel valley/channel must have been located in a position close to or at the ice margin. For the glaciofluvial complex to have formed into a subglacial cavity or lake, the tunnel valley/channel could have formed anywhere beneath the ice sheet. By associating the tunnel valley/channel with the proposed megaflood, the glaciofluvial complex is then, by default, implied to have been formed subglacially. The fact that the meltwater flowing out of the Tawatinaw valley in the eastern branch flowed uphill for the first few kilometres also indicates that the glaciofluvial complex was not formed subaerially.

The sequence of deposits illustrated in Figure 4.16 and described in section 4.2.3.1.1 (the complex bedded sand exposure located to the north of the glaciofluvial complex near the south end of the valley) can be interpreted in several ways. The upward gradation of units formed by increasingly powerful flow may simply represent increasing flow power out of the valley over time. They could also have resulted from lateral shifting of the eastern branch coming out of the valley - the lower flow power deposits forming when the main channel was further away, and the higher flow power deposits appearing as the channel came closer. Another possibility is that only the upper two units are associated with the tunnel channel, and the others were deposited at another time.

5.1.6 The Bars in the Valley

The bars in the valley, which consisted of cross-laminated and cross-bedded ripples, are interpreted as possibly being bedforms such as large subaqueous dunes. It is also possible that they are erosional residuals, as proposed by Brennand and Sharpe (1993).

5.2 Conclusions

Based upon the analysis of field data and image interpretation, it is concluded that the Tawatinaw valley operated as a tunnel valley/channel during the last glaciation. The source for the meltwater is presumed to be either waning-stage flow from the Livingstone Lake event or from a smaller source such as a subglacial lake. It is believed that water charged southward through the valley; then when it became less constricted the water spread out, deflecting southeastward. As water under normal flow conditions naturally follows the easiest (lowest) path, and since this southeastern deflection involves some uphill flow, it is believed that some sort of obstruction (such as grounded ice) existed to the southwest (Holden 1993). Because the flow was no longer constricted within the valley and was able to fan outwards, the large sediments being carried within were rapidly deposited. Many of the gravel deposits within the glaciofluvial complex are extensive chaotic deposits, reflecting rapid deposition. The crossbedded gravels observed in some exposures within the glaciofluvial complex indicate that some deposition was structured in nature.

5.2.1 Proposed Sequence of Events

The following sequence of events are proposed for the late glacial activity in the Tawatinaw area:

1. The Tawatinaw valley operated as a preglacial river, possibly utilized as a tunnel valley/channel early in the Late Wisconsinan.
2. The Eastern tract of the Livingstone Lake megaflood event swept through the region, creating the Athabasca fluting field; the preglacial Tawatinaw valley (or premegaflood Tawatinaw tunnel valley/channel) acted as a step to trigger the vortices needed to form the flutes. The massive sands located high up the valley sides were likely deposited at this time. The glaciofluvial complex began to be deposited as the meltwater exited the valley, spreading into a fan-like shape once it was no longer constricted.
3. As the megaflood flow waned, ice recoupled to the higher ground and the meltwater occupied only the Tawatinaw valley, incising deposits and meandering to follow the scallops on the west side. The glaciofluvial complex continued to be deposited as the meltwater

exited the valley, spreading into a fan-like shape once it was no longer constricted.

4. During retreat of the ice sheet, proglacial meltwater ponded in the deeper, northern end of the valley creating the rhythmically bedded clay deposits.

5.3 Future Research Objectives

With any study of the geomorphic history of a region, there are always more data to collect and new ideas to investigate. This project left the following avenues open for further study:

- GPR data collection

  Ground Penetrating Radar would be a useful tool for examining the glaciofluvial gravels to the south of the valley. Some preliminary GPR data were collected, and indicated that it would clearly pick up on bedding structures within the sediment. This would be a good way of confirming that the sediments were deposited by south-easterly flowing water, as indicated by the few good exposures that were observed in the area. GPR imagery was not included in this thesis, but will be an integral part of future research.

- Study of the neighbouring White Earth Creek valley

  As observed in the discussion of the regional setting of the study area, it is interesting to note that the White Earth Creek system is almost identical in appearance to the Tawatinaw system. Originally, it was proposed that this project would examine both valleys, since they likely formed in the same manner. It was decided to focus on the one feature, but in future it would be interesting to study the White Earth Creek area in order to see if the observations support the conclusions drawn regarding the genesis of the Tawatinaw system.

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Appendix A
Samples Processed Through Sieves

sample #		start weight		pebble -2ϕ 4mm		granule -1ϕ 2mm		v. c. sand 0ϕ 1mm		c. sand +1ϕ 0.5mm		m. sand +2ϕ 0.25mm		f. sand +3ϕ 0.125mm		v. f. sand +4ϕ 0.0625mm		silt & clay	
orig.	new	g	%	g	%	g	%	g	%	g	%	g	%	g	%	g	%	g	%
1		151.7	100	0.0	0.0	8.3	5.4	13.3	8.8	37.7	24.8	70.0	46.2	19.4	12.8	1.7	1.1	1.2	0.8
2		159.6	100	0.0	0.0	8.6	5.4	12.0	7.5	23.3	14.6	32.5	20.4	26.8	16.8	25.6	16.0	29.5	18.5
3	V3	153.6	100	0.0	0.0	0.1	0.1	0.2	0.1	0.8	0.5	99.6	64.8	50.7	33.0	1.0	0.6	0.9	0.6
4	V2	155.1	100	0.0	0.0	2.0	1.3	3.2	2.1	18.0	11.6	72.4	46.7	33.0	21.3	9.0	5.8	17.1	11.0
5	V7	153.9	100	0.0	0.0	31.3	20.4	37.1	24.1	26.4	17.2	18.9	12.2	13.6	8.8	13.5	8.8	12.6	8.2
6	V13	152.8	100	0.0	0.0	1.3	0.8	1.6	1.1	7.4	4.9	103.9	68.0	32.9	21.6	2.2	1.5	3.1	2.0
7		151.4	100	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	2.0	1.3	84.8	56.0	55.3	36.5	8.1	5.3
8		154.7	100	0.0	0.0	0.0	0.0	0.0	0.0	0.2	0.1	7.3	4.7	115.9	75.0	27.8	18.0	3.0	2.0
9		151.2	100	0.0	0.0	0.0	0.0	0.2	0.1	6.9	4.6	86.7	57.4	54.4	36.0	1.8	1.2	0.8	0.5
10		154.7	100	0.0	0.0	0.2	0.1	2.4	1.6	49.3	31.9	94.7	61.2	3.3	2.1	1.8	1.1	2.6	1.7
11		152.4	100	0.0	0.0	0.5	0.3	0.8	0.5	43.1	28.3	101.8	66.8	2.2	1.5	1.2	0.8	2.2	1.5
12		151.1	100	0.0	0.0	0.0	0.0	0.9	0.6	32.5	21.5	109.7	72.6	6.0	3.9	1.2	0.8	0.8	0.5
13		155.7	100	0.0	0.0	0.4	0.2	7.3	4.7	20.8	13.3	74.0	47.5	35.8	23.0	12.8	8.2	4.4	2.8
14	V1	156.6	100	0.0	0.0	0.5	0.3	3.6	2.3	21.6	13.8	26.3	16.8	38.5	24.6	35.2	22.5	29.5	18.8
15		155.7	100	32.2	20.7	32.6	20.9	24.4	15.6	29.3	18.8	27.2	17.5	7.6	4.9	1.5	1.0	0.8	0.5
17	V4	157.6	100	11.1	7.0	16.4	10.4	22.6	14.3	61.9	39.3	36.6	23.2	6.0	3.8	1.3	0.8	1.0	0.7
18	V5	151.5	100	3.7	2.5	4.9	3.2	11.8	7.8	42.9	28.3	41.0	27.0	25.6	16.9	13.1	8.6	7.8	5.1
19	V6	152.2	100	5.7	3.7	9.2	6.1	9.8	6.4	52.3	34.3	69.8	45.9	3.7	2.5	0.7	0.5	0.6	0.4
20	V8	155.7	100	0.0	0.0	0.9	0.6	3.0	1.9	68.3	43.9	82.6	53.0	1.2	0.7	0.1	0.1	0.1	0.1

sample #		start weight		pebble -2ϕ 4mm		granule -1ϕ 2mm		v. c. sand 0ϕ 1mm		c. sand +1ϕ 0.5mm		m. sand +2ϕ 0.25mm		f. sand +3ϕ 0.125mm		v. f. sand +4ϕ 0.0625mm		silt & clay	
orig.	new	g	%	g	%	g	%	g	%	g	%	g	%	g	%	g	%	g	%
21	V9	153.8	100	0.0	0.0	1.6	1.1	7.4	4.8	19.5	12.7	75.2	48.9	27.8	18.1	7.1	4.6	14.3	9.3
22	V10	156.8	100	0.0	0.0	0.1	0.1	0.5	0.3	2.9	1.9	30.7	19.7	116.9	75.0	2.9	1.9	1.4	0.9
23	V11	153.3	100	0.0	0.0	1.3	0.9	2.2	1.4	13.5	8.8	83.3	54.4	50.0	32.6	1.7	1.1	0.8	0.5
24	V12	159.8	100	0.0	0.0	0.1	0.1	0.1	0.1	4.1	2.6	113.7	71.2	38.6	24.1	1.4	0.9	1.2	0.7
25	V14	153.4	100	0.0	0.0	0.3	0.2	1.9	1.2	48.2	31.4	96.4	62.8	6.0	3.9	0.3	0.2	0.1	0.1
26	V15	152.5	100	0.0	0.0	0.1	0.1	0.3	0.2	6.5	4.3	109.8	72.0	34.0	22.3	1.0	0.7	0.6	0.4
27	V19	156.7	100	0.0	0.0	0.4	0.3	2.6	1.6	54.9	35.0	85.4	54.5	12.0	7.6	0.6	0.4	0.3	0.2
28	V16	150.3	100	7.9	5.3	3.9	2.6	7.3	4.9	17.5	11.6	43.9	29.2	53.1	35.4	11.2	7.4	4.9	3.2
29	V17	151.6	100	0.0	0.0	2.6	1.7	9.0	5.9	8.8	5.8	21.0	13.8	104.8	69.1	4.2	2.7	0.3	0.2
30	V18	386.5	100	170.8	44.2	51.6	13.3	49.5	12.8	73.0	18.9	23.0	5.9	10.3	2.7	4.4	1.1	3.6	0.9
32		152.0	100	0.0	0.0	0.1	0.0	0.2	0.1	9.1	6.0	93.8	61.8	43.2	28.5	4.6	3.0	0.6	0.4
33		155.3	100	0.0	0.0	0.8	0.5	0.7	0.5	6.0	3.9	36.4	23.4	95.9	61.7	10.6	6.8	3.9	2.5
34		158.4	100	14.0	8.9	1.0	0.6	5.0	3.2	61.6	38.9	71.3	45.0	3.8	2.4	0.5	0.3	0.6	0.4
35		287.5	100	9.7	3.4	17.9	6.2	81.4	28.3	124.2	43.2	47.1	16.4	4.3	1.5	1.1	0.4	0.9	0.3
36		400.0	100	143.3	35.8	77.2	19.3	25.0	6.3	52.8	13.2	77.0	19.2	19.6	4.9	2.9	0.7	1.8	0.5
37		241.3	100	46.5	19.3	29.9	12.4	12.2	5.1	26.0	10.8	108.1	44.8	14.7	6.1	1.8	0.7	1.9	0.8
38		150.2	100	0.0	0.0	0.0	0.0	0.0	0.0	0.4	0.3	4.4	2.9	102.6	68.3	35.6	23.7	5.8	3.8
39		152.2	100	0.0	0.0	0.0	0.0	0.0	0.0	1.3	0.9	84.3	55.4	64.1	42.1	1.7	1.1	0.3	0.2
40		152.8	100	0.0	0.0	0.0	0.0	0.1	0.0	0.3	0.2	1.1	0.7	6.3	4.1	70.3	46.0	74.5	48.7

sample #		start weight		pebble -2ϕ 4mm		granule -1ϕ 2mm		v. c. sand 0ϕ 1mm		c. sand +1ϕ 0.5mm		m. sand +2ϕ 0.25mm		f. sand +3ϕ 0.125mm		v. f. sand +4ϕ 0.0625mm		silt & clay	
orig.	new	g	%	g	%	g	%	g	%	g	%	g	%	g	%	g	%	g	%
41		156.3	100	0.0	0.0	0.0	0.0	0.2	0.1	3.2	2.1	55.3	35.4	83.0	53.1	12.0	7.7	1.7	1.1
42	V22	547.3	100	291.6	53.3	90.2	16.5	52.8	9.7	47.4	8.7	44.1	8.1	15.0	2.7	3.6	0.7	1.7	0.3
43		155.4	100	10.5	6.7	1.8	1.2	8.5	5.5	62.1	40.0	63.2	40.7	6.1	4.0	1.4	0.9	1.3	0.8
44		154.5	100	1.2	0.8	0.0	0.0	0.9	0.6	29.8	19.3	96.2	62.3	21.1	13.7	2.6	1.7	1.9	1.2
45		400.5	100	152.9	38.2	66.2	16.5	62.9	15.7	67.3	16.8	37.2	9.3	9.2	2.3	2.9	0.7	1.6	0.4
46		154.4	100	0.0	0.0	0.1	0.1	1.0	0.6	18.6	12.0	115.7	74.9	17.1	11.1	0.9	0.6	0.9	0.6
47		150.3	100	0.0	0.0	0.6	0.4	0.9	0.6	2.2	1.4	18.8	12.5	101.4	67.5	18.1	12.1	7.5	5.0
48	V20	161.1	100	16.4	10.2	9.2	5.7	25.0	15.5	55.0	34.2	37.4	23.2	10.9	6.7	3.5	2.1	3.5	2.2
49		150.0	100	1.9	1.3	2.2	1.4	6.4	4.2	17.2	11.4	29.0	19.3	42.8	28.6	27.7	18.4	21.9	14.6
50		112.6	100	2.9	2.5	1.5	1.3	10.3	9.2	24.9	22.2	25.6	22.8	20.3	18.1	14.7	13.1	11.2	10.0
53		158.9	100	0.0	0.0	0.0	0.0	0.0	0.0	0.5	0.3	36.4	22.9	93.7	59.0	25.8	16.2	1.9	1.2
54		151.6	100	0.0	0.0	0.1	0.1	0.3	0.2	3.1	2.1	55.7	36.7	74.5	49.2	16.6	10.9	0.8	0.6
55	V21	152.0	100	0.0	0.0	0.1	0.1	0.4	0.3	3.8	2.5	61.7	40.6	80.7	53.1	4.3	2.9	0.6	0.4
57		158.7	100	18.1	11.4	17.9	11.3	33.3	21.0	59.9	37.7	27.7	17.4	1.3	0.8	0.1	0.1	0.2	0.1
58		159.5	100	0.0	0.0	3.8	2.4	29.1	18.3	91.2	57.2	34.2	21.4	0.6	0.4	0.1	0.1	0.3	0.2
59		151.6	100	4.4	2.9	13.7	9.0	21.9	14.4	55.6	36.7	52.3	34.5	3.0	2.0	0.2	0.1	0.3	0.2
60		150.2	100	3.9	2.6	2.5	1.7	9.5	6.3	20.9	13.9	48.7	32.4	45.6	30.4	14.6	9.7	3.4	2.3