A COMPARATIVE ANALYSIS OF GLACIAL LANDFORMS:
SKEIÐARÁRSANDUR ICELAND AND NORTHWESTERN PENNSYLVANIA

A dissertation submitted
to Kent State University in partial
fulfillment of the requirements for the
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by

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This journey began as a way to get out of working in restaurants with a Bachelor's Degree in Education. Teaching at the college level soon became my goal and I enrolled in the Geology Master's program at Kent State University. In January of 2003, however, this journey nearly ended when my 12-year-old daughter, Jennifer, died unexpectedly. Therefore, I'd be remiss if I did not first thank my Master's advisor, Dr. Daniel Holm, and then geology department Chair, Dr. Donald Palmer, for encouraging me to continue with my education when quitting was all I wanted to do. The incredible support of my family, friends and professors during that terrible time was especially invaluable and I cannot thank all of them enough. The continued love and constant support of my family - especially that of my son, Stephen Arnold, and his dad, Matt Arnold, can never be measured or repaid. I am truly indebted to them. And although my daughter is gone, she inspired me in ways no one can ever imagine.

My friends - those that I've known for years and those that I've met recently - will never know how much their endless words of encouragement and countless hugs have meant to me. Thank you for picking me up and pushing me through. And last, but definitely not least, I must offer the biggest thanks to my friend and advisor, Dr. Mandy Munro-Stasiuk, who never gave up on me and who worked tirelessly down the stretch to help me to finish this degree. This journey took a lot longer than I anticipated, and she was there every step of the way. For that I am eternally grateful.

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I have reached my destination, but I am excited to embark on another journey as a professor, teaching geology and geography. I hope that this achievement inspires my son to never back down from a challenge and to always finish what he starts. I also hope that my journey from Bachelor's to PhD while raising a family, losing a child and working can be inspiring to other women who may be hesitant to start their education later in life.
CHAPTER 1: INTRODUCTION

1.1 RESEARCH STATEMENT

This study is a comparative analysis between glacial landforms known to have formed under glacial ice in south-central Iceland and the same landforms present in northwestern Pennsylvania. In the case of Iceland, all landforms formed in association with the Skeiðarárjökull glacier (an outlet glacier of the Vatnajökull Ice Cap) within the last 80 years and they are all on the outwash plain (Skeiðarársandur) of the glacier (Fig. 1.1). In northwestern Pennsylvania, the landforms were created near the last glacial maximum approximately 18,000 years ago in association with the Erie Lobe of the Laurentide Ice Sheet (LIS), and occur in what is now referred to as the Chautauqua Drumlin Field (Fig. 1.2). The suite of landforms at both locations is nearly identical and consists of drumlins, end moraines, fans and deltas, eskers, and tunnel valleys/channels. The scale of landforms between the two sites is, however, very different with those in Pennsylvania being more than an order of magnitude larger than those in Iceland.

At Skeiðarársandur, the creation of many of the landforms has been well documented as the result of glacial outburst floods, some catastrophic. In comparison it also has been hypothesized that massive floods generated from subglacial and ice-dammed lakes proximal to the LIS, during the last Ice Age (Fig. 1.3) were responsible for the creation of many of the same landforms across much of North America, including those in northwestern Pennsylvania. While modern flood analogs are small scale examples compared to what likely happened at the last great Ice
Figure 1.1. Satellite image showing the study site on Skeiðararsandur on the south side of the Vatnajökull Ice Cap.
Figure 1.2. Digital Elevation Model showing the terrain of the Chautauqua Drumlin Field (Modified from Saha, 2010)
Figure 1.3. North American ice sheets at Last Glacial Maximum, circa 18,000 years ago (Benn and Evans, 1998).
Sheets, they still offer the best illustration of subglacial meltwater behavior at the end of the last glacial maximum (LGM). This research therefore examines landforms and sediments in northwestern Pennsylvania and Iceland with the goal of building a better understanding of the origin of the ancient landforms.

For this research, ground-penetrating radar (GPR) was primarily used to investigate the landforms and their sediments in both locations. In Pennsylvania the structure of large valleys that dissect the Chautauqua Drumlin Field along with their associated landforms have traditionally been interpreted as proglacial meltwater features associated with the decay of the Erie Lobe of the LIS (Shepps et al., 1959). More recently they have been re-interpreted as being subglacial in origin (Munro-Stasiuk and Bradac, 2001; 2002) because the valley bottom profiles climb up to 50 m in the downflow direction and thus water had to be under pressure, under glacial ice, to force uphill flow. This re-interpretation rests not only on the uphill flow regime, but also on similarities to relict [non-actualistic or palimpsest] valleys interpreted as tunnel valleys in Alberta (Sjogren et al., 2002), Minnesota (Mooers, 1989) and Ontario (Barnett, 1998), rather than to modern or actualistic examples. This study therefore follows the logic of James Hutton’s concept of uniformitarianism (understanding ancient glacial landforms by understanding how modern glacial landforms relate to ongoing modern glacial processes) and it also follows process-form analysis, where details of landforms permit deductions about the processes that formed them (Clayton and Moore, 1974; Shaw, 2002). This study is therefore, to the best of my knowledge, the first that directly compares possible glacial flood sediments and landforms in ancient environments with those observed to have formed in modern environments. This study can therefore provide a greater insight into the origin of the LIS landforms.
1.2 SKEIÐARÀRSANDUR: CHARACTERISTICS AND PREVIOUS WORK

At present, 10% (11,200 km²) of Iceland is ice-covered and 60% of the ice overlies the active volcanic zone (Fig. 1.4; Björnsson, 2002). The combination of high rates of precipitation, frequent subglacial eruptions and enhanced geothermal heat flux, generates frequent outburst floods (jökulhlaups) (Björnsson, 1992; 2002; Einarsson et al., 1997). Approximately one third of all eruptions in Iceland during the 20th century were subglacial (10 major and five minor) (Björnsson, 2002). More than 80 volcanic eruptions occurred during the last 800 years at the Vatnajökull Ice Cap (Larsen et al., 1998). This region is especially complex because of the combination of ice along with the considerable volcanic activity associated with a segment of the Mid-Atlantic Ridge which crosses Iceland under the ice cap (Fig. 1.5). This paradoxical interaction between fire and ice has resulted in a number of large outburst floods and glacial surges, which in turn has created a multifaceted and dynamic landscape of superimposed erosional canyons and landforms resulting from the transportation and deposition of enormous quantities of sediment and ice blocks over vast outwash plains and sandur deltas. The power of these floods has threatened human populations, farms and hydroelectric power plants on glacier-fed rivers. They have damaged cultivated and vegetated areas, disrupted roads on the outwash plains, destroyed bridges (Fig. 1.6) and have even generated flood waves in coastal waters (Björnsson, 1996).
Figure 1.4. Location map of Iceland showing ice caps, the volcanic zone (The Palagonite Formation) and the central volcanoes (Björnsson, 2002).
Figure 1.5. Map showing location of Iceland in relation to Mid-Atlantic Ridge (www.marinebio.net).
Figure 1.6. Remnant of bridge that crossed Skeiðará River after the 1996 jökulhlaup.
There are six active ice caps in Iceland (Fig. 1.4): Mýrdalsjökull, Eyjafjallajökull, Langjökull, Drangajökull, Hofsjökull, and Vatnajökull, the largest of all of them. Vatnajökull has 30 outlet glaciers, one of which is Skeiðaràrjökull, the subject of this research. Each glacier has an extensive outwash plain, or sandur. At approximately 1,350 km², Skeiðaràrsandur is the largest active glacial outwash plain in the world, and extends from the terminus of Skeiðaràrjökull to the Icelandic coast (Fig. 1.3). The retreat of Skeiðaràrjökull has exposed, and subsequently modified, a considerable expanse of sedimentological and geomorphological landscapes relating to subglacial processes and land systems.

It is estimated that Skeiðaràrjökull reached its greatest extent in the 18th century and fluctuated around its maximum position until around 1890 when it started to retreat (Sigurðsson, 2005). A surge in 1929 overrode this older margin in several places (Sigurðsson, 2005) and the glacier has since retreated to its present position (3 km behind the 1929 moraine) with minor fluctuations.

The flood history at Vatnajökull is extensive with floods generated from ice-marginal lakes such as Graenalon (Fig. 1.1) and escape of volcanically-generated melt. The best documented flood was the November 1996 jökulhlaup that burst from the Vatnajökull Ice Cap onto Skeiðaràrsandur providing researchers with a rare opportunity to witness firsthand the geomorphic impact of a low-frequency, high-magnitude jökulhlaups. It was the highest-magnitude glacial flood ever measured (Smith et al., 2004) triggered by a volcanic eruption beneath the ice cap that began on September 30, 1996. Over the next month, 2.8 km³ of meltwater travelled subglacially into the Grímsvötn subglacial lake (Fig. 1.1) until it reached a critical level for drainage (Björnsson, 1997; 2002). The resulting jökulhlaup began in the Skeiðará River, the most
It drained Skeiðarârjökull on the morning of November 5, and reached a peak discharge of $45-53 \times 10^3$ m$^3$/s within 14 hours (Björnsson, 2002; Snorrason et al., 2002). After the release of water from Grimsvötn, the jökulhlaup spread as a high-pressure subglacial flood wave taking just 10.5 hours to reach the glacier snout (Jóhannesson, 2002; Flowers et al., 2004; Roberts, 2005).

Within ~48 hours of the initiation of the flood, the waning jökulhlaup left a suite of proglacial features, including transported giant clasts and ice blocks, massive aggradation deposits, eroded scarps, turbidite sequences, englacial sediment deposition and a massive supraglacial collapse embayment (Russell and Knudsen, 1999A, B; Russell et al., 1999, 2001; Smith et al., 2000; Gomez et al., 2000; Roberts et al., 2000; Roberts et al., 2001; Waller et al., 2001; Smith, 2002; Magilligan et al., 2002). This flood added an immense volume of sediment to the sandur which is characterized by massive, coarsening-upward sequences of clast-supported pebbles and cobbles, all interpreted as being deposited by a series of jökulhlaups (Maizels, 1991; 1993). Subglacial eruptions and/or geothermal heat creates meltwater which subsequently accumulates in the subglacial lake, Grimsvötn, where increasing hydrostatic pressure eventually causes a breach in an ice dam at the glacier bed and triggers a jökulhlaup (Fig. 1.7; Björnsson, 1992). Grimsvötn, located in the western part of Vatnajökull (Fig. 1.1), is the largest subglacial lake in Iceland and is considered one of the most active of the five volcanic systems that have been identified under Vatnajökull (Björnsson, 2002).
Figure 1.7. Schematic drawing of two main types of subglacial lakes; (A) a stable lake, (B) an unstable lake that drains in jökulhlaups (Björnsson, 2002).
The gap that has existed between the literature and tunnel valley formation within contemporary glacial systems was narrowed slightly when retreat of the southeast margin of Skeiðaràrjökull after the 1996 flood revealed a tunnel valley that had been excavated into moraine sediment on Skeiðaràrsandur. The valley ascends 11.5 m over a distance of 160 m (Fig. 1.8; Russell et al., 2007). This proved to be an important discovery in a contemporary environment as it provides a valuable analog to help decipher the role that meltwater plays in tunnel valley formation during outburst floods. Until now, this information had only been inferred from the Quaternary landscape. Although the Skeiðaràrjökull depositional system is small compared to Pleistocene glacial systems, frequent and well time-constrained outburst floods allows for one of the best modern analogs for high magnitude glaciofluvial erosional and depositional processes. The outbursts from Grímsvötn, and other Icelandic floods, have been well documented, therefore allowing for the association of particular glacial landsystems with specific glacier characteristics and dynamics.

This research was undertaken on the western side of Skeiðaràrsandur (Figs. 1.1 and 1.9), where the sandur is the widest and the terminal moraine complex sits within the proglacial zone (Figs. 1.9 and 1.10). The site was selected because of the number of features that exist within a rather small area. Initial reconnaissance revealed a feature that based on its characteristic radial shape and spatial relationship to a channel was proposed to be a delta. Other landforms at the site were identified as an esker and a drumlinized hill based on their distinctive shapes. Boulder lags are scattered across the sandur surface, which was comprised of poorly sorted sands and gravel.
Figure 1.8. (A). View from glacier looking south of tunnel channel revealed after 1996 jökulhlaup on Skeiðarársandur, Iceland. (B). View from south of tunnel valley revealed after 1996 jökulhlaup on Skeiðarársandur, Iceland (Russell et al., 2007).
Figure 1.9. Western edge of Skeiðaràrsandur, Iceland, showing the 1929 surge moraine and inset (white box) study site (Location of Figure 1.10)

Figure 1.10. Aerial photo of Skeiðaràrsandur, Iceland, highlighting drumlinized terrain (Iceland Anzel Survey, 1994).
1.3. NORTHWESTERN PENNSYLVANIA: CHARACTERISTICS AND PREVIOUS WORK

North America was glaciated by three ice sheets: the Cordilleran, Laurentide and Innuitian (Fig. 1.3). Around the Great Lakes, a number of stream ice lobes formed associated with each of the lakes. In northwestern Pennsylvania, the Erie Lobe spilled on land as a series of smaller sublobes (Fig. 1.11).

The landscape of northwestern Pennsylvania has been studied in detail since the mid-1800s. Lesley (1876) first recognized part of the landscape as having been glaciated with Chamberlin (1883) and Lewis (1884) doing the bulk of the groundwork in initial mapping. These landscapes were not examined in detail until the Second Pennsylvania Geological Survey was established in the late 1800's (Fleeger, 2005). Initial investigations mainly involved the identification and mapping of terminal moraines to determine the general extent of glaciations (Fig. 1.11). A host of other researchers have mapped glacial deposits and landforms and have presented an array of terminology and age determinations (Leverett, 1902, 1934; Shepps et al., 1959; White et al., 1969; Szabo and Totten, 1995; D'Urso, 2000), much of which has been summarized by Braun (2004) and Fleeger (2005).

Pennsylvania is now understood to have been glaciated in two areas: northwestern Pennsylvania (the subject of this research) which was glaciated by the Erie Lobe of the LIS; and northeastern Pennsylvania, which was glaciated by the Champlain Lobe (Fig. 1.12).

The majority of the research related to understanding the Quaternary history of northwestern Pennsylvania originates from the 1950’s and 1960’s work of George W. White, his students, and his colleagues. His pioneering efforts, which included maps of
the regional distribution of glacial deposits across northwestern Pennsylvania has helped to establish the formal lithostratigraphic positions for till units. This work was summarized and presented in a pair of publications: *Glacial Geology of Northwestern Pennsylvania* by Shepps et al. (1959) and then in *Pleistocene Stratigraphy of Northwestern Pennsylvania* by White et al. (1969).

Figure 1.11. Map showing location and extent of Erie and Grand River Lobes of the LIS (Shepps et al., 1959). Approximate location of study area shown in red square.
Figure 1.12. Ice lobes and directions of ice flow into Pennsylvania during the LIS (from Sevon et al., 1999). Approximate location of study area shown in red square.
At the margin of the LIS in northwestern Pennsylvania, ice moved down the Erie Basin as a major lobe known as the Erie Lobe and spread out into northwestern Pennsylvania twice during the Illinoian age and five times during the Wisconsin age of the Pleistocene Epoch (Fig. 1.11; Shepps et al., 1959). A minor ice advance by the Erie Lobe may have occurred as early as about 27.5 ka, but it probably did not extend much beyond the southern shore of Lake Erie (Fullerton, 1986). This was followed by another advance of ice into the Erie and Huron basins at about 22 ka (Karrow, 1984), and by 21 ka ice extended deep into southwestern Ohio and south central Indiana where it joined ice flowing south out of the Michigan basin (Fullerton, 1986). There, the ice margin wavered for approximately 4,000 years (Lowell et al., 1990). In northern Pennsylvania, northeastern Ohio, and western New York, the advancement of ice was dictated by the topography of the Appalachian Plateau which prevented the ice from penetrating too far inland (Muller and Calkin, 1993). Abundant striations indicate that the ice flowed generally southeastward (Leverett, 1902). The Grand River Lobe (Fig. 1.11), a sublobe of the Erie Lobe, covered northwestern Pennsylvania during the late Pleistocene (White, 1951). This lobe followed the Grand River lowland in northeastern Ohio and flowed outwards, into northwestern Pennsylvania (Shepps, 1955). Ice from the Grand River Lobe continued to advance through Pennsylvania from the northwest to the southeast, sculpting the landscape into the valleys that generally parallel the direction of the ice flow (Straffin and Grote, 2010) and are the focus of this study.

The glaciated portion of northwestern Pennsylvania contains deposits that are Illinoian and Wisconsinan in age (Fig. 1.13; White, 1951). Till and outwash deposited during the late Pleistocene epoch (Leverett, 1902) overlie Devonian and Mississippian bedrock that gets progressively younger as you move southward (Shepps et al., 1959).
The till sheets that were deposited in the region characterize a sequence of alternating advances and retreats during this time period (Shepps et al., 1959). The Devonian bedrock includes mostly light gray shales, whereas the Mississippian units are composed of interbedded sandstones, siltstones, and shales (Richards et al., 1987). During each advance of the ice, materials were picked up locally within northwestern Pennsylvania and from places to the northeast. These materials were carried along by the ice and meltwater for varying distances and were finally deposited in one form or another. The greatest percentage of the material was laid down directly from the ice indiscriminately over the landscape in the form of ground or end moraines composed of till (Shepps et al., 1959). The remainder of the deposition was in the form of outwash from meltwater flowing under the ice, through fissures in the ice or issuing from the ice margin. These outwash deposits are variously known as kames, kame terraces and kame moraines and consist of well- to poorly-bedded sands, silts, clays, and gravels (Shepps et al., 1959). The Kent till, the oldest of the Wisconsinan tills (White, 1982), includes kames and eskers, as well as end, recessional, and ground moraines (Shepps et al., 1959). The Kent end moraine occurs just outside my study area but approximately half of the glaciated topography of northwestern Pennsylvania is composed of Kent ground moraine (Shepps et al., 1959).
Figure 1.13. Map showing glacial deposits of Pennsylvania. Commonwealth of Pennsylvania Department of Conservation and Natural Resources Bureau of Topographic and Geological Survey (Sevon and Braun, 2000).
Erie County is bordered to the north by Lake Erie and to the west by the Ohio state line and lies within the Appalachian Plateau physiographic province. Three physiographic divisions are recognized in Erie County (Fig. 1.14): Lake Plain, Escarpment Slope, and Upland Plateau (Tomikel and Shepps, 1967). The northern border of the Lake Plain is at the elevation of Lake Erie (mean lake elevation is 174 m) and extends inland to an elevation of about 244 m. The Escarpment Slope separates the nearly flat lake plain from the Upland Plateau, which is characterized by broad valleys with flat bottoms and relatively steep walls, which have since been identified as tunnel valleys. The transition from Lake Plain to Escarpment Slope in western Erie County is gradual with gentle changes in elevation and mixed surface features in both divisions. The maximum elevation of the escarpment slope is about 305 m, with the highest elevation of the Upland Plateau at slightly above 579 m in southeastern Erie County.
Figure 1.14. Physiographic provinces of Erie County, Pennsylvania. Borough of Edinboro and area southwest of Waterford shown in squares (Buckwalter et al., 1996).
The plateau is dominated by drumlins throughout this region, which range from 328 m to 2,831 m in length and from 140 m to 1,000 m in width, and most are oriented along a the mean direction of 150° (Saha et al., 2011). While not identified as drumlins in the early research of northwestern Pennsylvania, Shepps (1955) mapped what he referred to as a “special type of glacially-induced topography” in the area northeast of Edinboro in Crawford County. He noted that the surface consisted of a series of long, smooth, parallel ridges with intervening valleys, with the ridges consistently oriented in a N35°W direction and frequently unbroken for a kilometer or more (Fig. 1.15). He coined the term “corrugation topography” to describe these landforms because of the resemblance to corrugated cardboard. These features since have been identified as drumlins.
Figure 1.15. Orientation of surface features, now known as drumlins, in the study area referred to as 'corrugation topography' by Shepps (1955).
1.4 STRUCTURE OF THE REST OF THIS DISSERTATION

Chapter 2 - Literature Review on the Origin of a Subglacial Landform Suites

This chapter will examine the difference between erosional and depositional landforms, as well as the scale that pertains to landforms found in the glacial continuum of Munro-Stasiuk et al. (2009). It will focus on the role that meltwater plays in the formation of tunnel valleys, drumlins, kames and eskers and introduce the meltwater hypothesis of Shaw (1983). Previous research on tunnel valleys in other parts of North America will be presented, as well as research concerning landforms associated with outburst flooding in Iceland.

Chapter 3 - Methodology

This chapter will present the principles of ground penetrating radar and the methods of data collection and processing. It also will detail the use of the GPR software, EkkoView Deluxe and EkkoView2 in data processing, and the use of digital elevation models (DEMs), topographic maps and Google Earth in preparation for field work.

Chapter 4 - Iceland Data Analysis

This chapter includes photographs of Skeiðarársandur that note the location of individual transects selected for GPR profiling. Here, data that were deemed relevant for analysis are interpreted and discussed. Each GPR profile is annotated to highlight significant features which led to each interpretation.
Chapter 5 - Pennsylvania Data Analysis

Maps are presented to show the location of areas selected for GPR surveys. Lengthy profiles were chopped into smaller segments so that features could be more easily identifiable. These features were then noted on each profile and interpreted. Many data within this study were degraded as a result of human interference along roadways, i.e. bridges, culverts and buried cables. Some of the most interesting data were found within surveys at a privately owned and active gravel pit, southwest of Waterford and these are presented in more detail.

Chapter 6 - Discussion

This chapter revisits the rationale for this study and summarizes the information collected in Iceland and Pennsylvania. It addresses the similarities and differences between the two study sites and uses these as a basis of discussion of the formation of glacial features and form analogy.

Chapter 7 - Conclusion

The findings are summarized, and limitations of this research and suggestions for future work are presented.
CHAPTER 2: REVIEW OF THE ORIGINS OF SUBGLACIAL LANDFORMS

2.1 INTRODUCTION

The landforms studied in this dissertation are part of a subglacial-ice-marginal continuum that contains tunnel valleys/channels, drumlins and eskers behind the margin, and fans, deltas and moraines in front of the margin. The origin of this landform suite, especially the subglacial one is contentious, with the major controversies revolving around formation by direct erosion or deformation by ice, erosion or deposition by meltwater, erosion or deposition by “regular” seasonally fluctuating meltwater under the glacier (jökulhlaups), or various combinations of these processes. Formation of drumlins and tunnel valleys in particular still remains unclear. Understanding their origin is crucial for this dissertation, and both landforms are central to the highly controversial meltwater hypothesis introduced by Shaw (1983).

As a foundation to the rest of this dissertation, therefore, this chapter reviews the literature on the nature of subglacial environments, the meltwater hypothesis and the controversies surrounding it, and the origins of the landforms discussed in this dissertation.

2.2 SUBGLACIAL ENVIRONMENTS: EROSIONAL VERSUS DEPOSITIONAL LANDFORMS

Current reconstructions of paleoenvironments that address the dimensions, geometry and dynamics of former ice sheets are based mainly on glacial depositional landforms rather than glacial erosional landforms (Glasser and Bennett, 2004). This is due in part to a lack of a detailed understanding of the origin and significance of glacial erosional landforms since the conditions under which they form remain enigmatic. Glacial erosional processes involve the removal and transport of bedrock and/or
sediment which results in landform assemblages that are frequently observed in areas that were known to have been occupied by ice sheets and glaciers. Glasser and Bennett (2004) summarized erosional glacial landforms into three categories based on size: micro, meso, and macro, and provided summaries of the significance of each landform (Tables 2.1, 2.2, and 2.3). They deliberately omitted landforms such as drumlins, which have been interpreted to be both erosional and depositional.

### Table 2.1. Micro-scale landforms of glacial erosion and their significance for the reconstruction of former ice masses (Glasser and Bennett, 2004).

<table>
<thead>
<tr>
<th>Landform</th>
<th>Morphology</th>
<th>Glaciological significance</th>
</tr>
</thead>
</table>
| Striae            | Small grooves or scratches on bedrock surfaces, often continuous over several metres. Commonly associated with polished bedrock surfaces. Striae can be assigned to one of three morphological classes:  
Type 1 striae become progressively wider and deeper down-glacier until they end abruptly, often as deep steep-walled grooves.  
Type 2 striae start and terminate as faint, thin traces. They steadily broaden and deepen until they reach a maximum width and depth near their centre point.  
Type 3 striae begin abruptly as deep grooves, then become progressively narrower and shallower down-glacier. | Warm-based ice carrying a basal debris load  
Striae oriented in the direction of local ice flow  
Individual striae created by large (>1 mm) clasts  
Polishing caused by fine (sub-millimeter) fracture  
High clast-bed contact pressures (>1 MPa) inferred from clast penetration into bedrock and from intimate ice-bedrock contact.  
Type 1 striae indicate that the striating clast ploughed forward and downward, before either the striator point broke off the clast or the torque on the clast was sufficiently large to rotate it out of the groove.  
Type 2 striae indicate that a clast slided down until the maximum striation depth was reached and then steadily accelerated until the striator terminated at the same velocity as the ice.  
Type 3 striae indicate that a striator point contacted and indented the bed. The clast rotated with little displacement along the bed producing a low ploughing angle, so that a gradual reduction in indentation depth occurred as sliding proceeded. |
| Micro coag and tails | Small tails of rock formed in the lee of resistant crystals, grains or nodules. | Warm-based ice carrying a basal debris load  
Crag and tall indicates orientation and direction of ice flow  
High clast-bed contact pressures (>1 MPa) inferred from intimate ice-bedrock contact. |
| Bedrock gouges and cracks | Grooves and cracks, often crescentic in outline, cut into bedrock surfaces. Often occur as a series of gouges or cracks. | Warm-based ice carrying a basal debris load  
Direction of forward dip of bedrock fracture indicates the direction of local ice flow  
High clast-bed contact pressures (>1 MPa) inferred from evidence of bedrock fracturing.  
Temporal fluctuations in clast-bed contact pressures are implied because gouges and cracks often occur in series.  
Some evidence that the length of individual crescentic fractures and gouges increases forward with ice thickness, allowing estimates of former ice thickness. |
| p-forms and re-entrants | Smooth-walled sculpted depressions and channels cut in bedrock. Encourages a variety of morphological expressions including arched basins, basins, erosion marks, potholes, bowls, channels and grooves. | Landforms carrying striae indicate warm-based ice  
Carrying a basal debris load and high clast-bed contact pressures (>1 MPa) inferred from intimate ice-bedrock contact.  
Landforms where striae are absent indicate presence of abundant basal meltwater, possibly concentrated by catastrophic discharge. Low effective normal pressures inferred for these landforms. |
<table>
<thead>
<tr>
<th>Landform</th>
<th>Morphology</th>
<th>Glaciological significance</th>
</tr>
</thead>
<tbody>
<tr>
<td>Streamlined bedrock features</td>
<td>Streamlined bedrock eminences with abraded surfaces on all sides</td>
<td>Warm-based ice carrying a basal debris load</td>
</tr>
<tr>
<td>(‘whalebacks’)</td>
<td></td>
<td>High effective normal pressures (0.1-1 MPa) inferred from intimate icebedrock contact and cavity suppression</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Thick ice</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Low sliding velocity with little available basal meltwater</td>
</tr>
<tr>
<td>Stoss and lee forms (‘roches</td>
<td>Upstanding bedrock eminence with both abraded and quarried faces</td>
<td>Warm-based ice carrying a basal debris load</td>
</tr>
<tr>
<td>mouminies’)</td>
<td>Detailed morphology controlled by preglacial weathering characteristics and patterns of bedrock jointing</td>
<td>Low effective normal pressures (0.1-1 MPa) inferred from the presence of basal cavities</td>
</tr>
<tr>
<td>Rock grooves and rock basins</td>
<td>Smooth-walled sculpted depressions and channels cut in bedrock</td>
<td>Quarried faces indicate abundant basal meltwater with regular fluctuations in basal water pressure</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Rapid sliding velocity</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Some evidence that roches mouminies form under thin ice, e.g., possibly during ice sheet build-up and decay</td>
</tr>
<tr>
<td></td>
<td></td>
<td>May indicate direction and orientation of ice flow</td>
</tr>
<tr>
<td>Subglacial meltwater</td>
<td>Steep sided channels cut into bedrock or till</td>
<td>Warm-based ice carrying a basal debris load</td>
</tr>
<tr>
<td>channels</td>
<td>Channel orientation may be discordant with the local topography</td>
<td>Channel systems can be used to calculate former hydraulic potential gradient and therefore infer the regional pattern of subglacial drainage, and to estimate ice surface slope and ice thickness</td>
</tr>
<tr>
<td></td>
<td>Channels may have an irregular convex-up long profile</td>
<td>Calculations of palaeovelocities and palaeodischarges possible from measurements of channel shape, channel width and size of material transported by meltwater flow</td>
</tr>
<tr>
<td>Ice-marginal meltwater</td>
<td>Either a complete channel cross-section or a channel floor and one wall wherever the other wall was formed by ice (half channel)</td>
<td>Release of large quantities of supra- and/or subglacial meltwater</td>
</tr>
<tr>
<td>channels</td>
<td>Channels start and end abruptly</td>
<td>Channels indicate the location of the former ice margin</td>
</tr>
<tr>
<td></td>
<td>Often associated with other ice-marginal depositional landforms</td>
<td>Gradient of channel long profile may indicate that of the ice margin</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Calculations of palaeovelocities and palaeodischarges possible from measurements of channel shape, channel width and size of material transported</td>
</tr>
</tbody>
</table>

Table 2.2. Meso-scale landforms of glacial erosion (Glasser and Bennett, 2004).
<table>
<thead>
<tr>
<th>Landform</th>
<th>Morphology</th>
<th>Glaciological significance</th>
</tr>
</thead>
<tbody>
<tr>
<td>Regions of areal scour</td>
<td>Areas of low relief smoothed into streamlined</td>
<td>Warm-based ice carrying a basal debris load when the margin of the glacier recedes. Basal</td>
</tr>
<tr>
<td></td>
<td>erosions and basins, taking the form of roches</td>
<td>effective normal pressures (0.1–1 MPa) inferred from the presence of basal cavities caused by</td>
</tr>
<tr>
<td></td>
<td>moutonnées and windrows</td>
<td>large-scale changes in the glacial tributary system. Whirls and lee features may indicate</td>
</tr>
<tr>
<td></td>
<td>Rock surfaces may contain striae and bedrock</td>
<td>former basal conditions. Queried landforms indicate low effective normal pressures (0.1–1 MPa)</td>
</tr>
<tr>
<td></td>
<td>gouges and cracks</td>
<td>inferred from the presence of basal cavities with large-scale changes in the glacial tributary</td>
</tr>
<tr>
<td></td>
<td>Detailed morphology controlled by preglacial</td>
<td>system. Whirls and lee features may indicate former basal conditions. Queried landforms</td>
</tr>
<tr>
<td></td>
<td>weathering characteristics and patterns of bedrock</td>
<td>indicate low effective normal pressures (0.1–1 MPa) inferred from the presence of basal</td>
</tr>
<tr>
<td></td>
<td>jointing</td>
<td>cavities with large-scale changes in the glacial tributary system. Whirls and lee features</td>
</tr>
<tr>
<td></td>
<td></td>
<td>may indicate former basal conditions. Queried landforms indicate low effective normal</td>
</tr>
<tr>
<td>Glacial troughs</td>
<td>Deep valleys with smooth, polished, and</td>
<td>Warm-based ice carrying a basal debris load when the margin of the glacier recedes. Basal</td>
</tr>
<tr>
<td></td>
<td>steep-walled and flat-bottomed</td>
<td>effective normal pressures (0.1–1 MPa) inferred from the presence of basal cavities caused</td>
</tr>
<tr>
<td></td>
<td>Valley cross-sectional morphology can be</td>
<td>by large-scale changes in the glacial tributary system. Whirls and lee features may indicate</td>
</tr>
<tr>
<td></td>
<td>described by empirical power-law functions and</td>
<td>former basal conditions. Queried landforms indicate low effective normal pressures (0.1–1 MPa)</td>
</tr>
<tr>
<td></td>
<td>by second-order polynomials</td>
<td>inferred from the presence of basal cavities with large-scale changes in the glacial tributary</td>
</tr>
<tr>
<td>Cinques</td>
<td>Large bedrock hollows that open downslope</td>
<td>Warm-based ice and abundant meltwater. The cross-sectional area of a trough and its</td>
</tr>
<tr>
<td></td>
<td>and are bounded upslope by a cliff, steep</td>
<td>longitudinal profile may become calibrated to discharge over time, enabling estimates of the</td>
</tr>
<tr>
<td></td>
<td>slope or arcuate basin</td>
<td>mass balance condition of the glacier. Queried landforms indicate low effective normal</td>
</tr>
<tr>
<td></td>
<td></td>
<td>pressures (0.1–1 MPa) inferred from the presence of basal cavities with large-scale changes</td>
</tr>
<tr>
<td>Giant gullies and lee forms</td>
<td>Large upstanding bedrock hills or spurs with</td>
<td>Warm-based ice carrying a basal debris load when the margin of the glacier recedes. Basal</td>
</tr>
<tr>
<td></td>
<td>both-abraded and quarried faces</td>
<td>effective normal pressures (0.1–1 MPa) inferred from the presence of basal cavities caused</td>
</tr>
<tr>
<td></td>
<td>Detailed morphology controlled by preglacial</td>
<td>by large-scale changes in the glacial tributary system. Whirls and lee features may indicate</td>
</tr>
<tr>
<td></td>
<td>weathering characteristics and patterns of bedrock</td>
<td>former basal conditions. Queried landforms indicate low effective normal pressures (0.1–1 MPa)</td>
</tr>
<tr>
<td></td>
<td>jointing</td>
<td>inferred from the presence of basal cavities with large-scale changes in the glacial tributary</td>
</tr>
<tr>
<td>Tunnel valleys and tunnel</td>
<td>Large, sinuous, steep-sided valleys or</td>
<td>Warm-based ice carrying a basal debris load when the margin of the glacier recedes. Basal</td>
</tr>
<tr>
<td>channels</td>
<td>depressions that may contain enclosed basins in</td>
<td>effective normal pressures (0.1–1 MPa) inferred from the presence of basal cavities caused</td>
</tr>
<tr>
<td></td>
<td>their floor</td>
<td>by large-scale changes in the glacial tributary system. Whirls and lee features may indicate</td>
</tr>
<tr>
<td></td>
<td>Tunnel channels are incised into bedrock,</td>
<td>former basal conditions. Queried landforms indicate low effective normal pressures (0.1–1 MPa)</td>
</tr>
<tr>
<td></td>
<td>glaciogenic sediments on either pre-existing</td>
<td>inferred from the presence of basal cavities with large-scale changes in the glacial tributary</td>
</tr>
<tr>
<td></td>
<td>materials</td>
<td>system. Whirls and lee features may indicate former basal conditions. Queried landforms</td>
</tr>
<tr>
<td></td>
<td>Tunnel valleys are usually infilled with</td>
<td>indicate low effective normal pressures (0.1–1 MPa) inferred from the presence of basal</td>
</tr>
<tr>
<td></td>
<td>sediment and occur both on the continental shelf</td>
<td>cavities with large-scale changes in the glacial tributary system. Whirls and lee features</td>
</tr>
<tr>
<td></td>
<td>and in lowland areas</td>
<td>may indicate former basal conditions. Queried landforms indicate low effective normal</td>
</tr>
</tbody>
</table>

Table 2.3. Macro-scale landforms of glacial erosion (Glasser and Bennett, 2004).
Uncontested erosional glacial landforms include glacial valleys and roches moutonnées which form due to quarrying and abrasion mainly. Quarrying, also known as plucking, is the removal of large blocks from the bed. Abrasion is the wear of the surface that is created by the movement of isolated clasts in ice or the sliding of sediment over the bed. Roches moutonnées and whalebacks, for example, are products of the combination of quarrying and abrasion, and are created when cavities form between an ice sheet and its bed and are indicators of low effective basal pressures and high sliding velocities that are necessary for the separation of the ice–bed interface (Glasser and Bennett, 2004).

Roches moutonnées (Fig. 2.1) are sculpted landforms with a convex shape. The lee side of the feature is composed of unconsolidated materials as a result of water that seeps into fractures and subsequently experiences alternating freeze-thaw cycles (Sugden, 1976). Whalebacks (Fig. 2.2) also have a convex shape and are formed in a similar manner to roches moutonnées, with the key physical difference being the lack of unconsolidated materials on the lee side of whalebacks.

Figure 2.1. Schematic showing formation of roches moutonnées (https://macwiki.mcmaster.ca/clip/index.php/Main/rochessmoutonnées).
Three main hypotheses have been used to explain the origin of the features commonly referred to as p-forms (plastic forms) and s-forms (sculpted forms). These are (1) formation by quarrying (Boulton, 1974; Goldthwait, 1979), (2) formation by abrasion (Gjessing, 1965), and (3) formation by melt water (Shaw, 1983; Sharpe and Shaw, 1989; Kor et al., 1991; Pair, 1997; Sawagaki and Hirakawa, 1997; Glasser and Hambrey, 1998; Glasser and Nicholson, 1998). The challenge to determine which of the different mechanisms is responsible for the formation of the various glacial features relies on how well each approach explains the individual landform characteristics, while still keeping the basic physical principles that must be involved in their creation (Shaw, 2002).

Other landforms that result from glacial abrasion include striae, grooves, microcrag and tails, bedrock gouges and cracks (Table 2.1). Striae are small grooves or scratches associated with polished bedrock surfaces, while grooves are smooth-walled
channels sculpted into bedrock. Micro-crag and tails are small tails of rock that are formed on the lee side of resistant materials. Gouges and cracks often have a crescentic outline that often occur in a series due to fluctuations in the ice contact with the bed. Each of these forms are indicators of warm-based ice that is carrying a basal debris load.

Glacial meltwater is produced by the melting of surface ice due to heat transfer from contact with air and rain at temperatures above freezing point, and from the energy of incoming solar radiation, or by the melting of basal ice due to pressure, friction, and geothermal heat. Meltwater is important in the evacuation of sediment and rock fragments produced by processes such as abrasion and fracturing. In addition, meltwater flowing in subglacial channels or as films at the ice-rock interface and the sediment carried by meltwater cause erosion of bedrock or sediments by mechanical and chemical processes. Landforms that originate from meltwater erosion include subglacial and ice-marginal meltwater channels and drumlins. The effectiveness of meltwater as an agent of erosion depends on: (1) the susceptibility of the bedrock involved; (2) the water velocity and level of turbulent flow; and (3) the quantity of sediment in transport (Glasser and Bennett, 2004). It is possible to calculate paleovelocity and paleodischage from measurements related to channel shape, width and the size of the materials that are transported by the meltwater. Bedrock erosional forms were pivotal in establishing Shaw’s meltwater hypothesis (1983) since their form is widely interpreted as having a fluvial origin.
2.3 THE MELTWATER HYPOTHESIS

The meltwater hypothesis of Shaw (1983) attributes some subglacial landforms to the erosive effects of catastrophic floods that are postulated to have been considerably larger than any floods in history. It is an alternative argument to current hypotheses on bedforms that invoke direct glacial action or the action of subglacial deforming beds. Shaw initiated the subglacial megaflooding theory after viewing aerial photographs of landforms in northern Saskatchewan that were previously interpreted as eolian sand dunes. He realized that these hills were in fact a type of drumlin. Shaw noted that drumlins were similar in form to inverted erosional marks produced by turbulently flowing water. Fieldwork eventually revealed that these drumlins were subglacial cavity fills of water-deposited gravel and sand that were filled as the meltwater flows slowed (Fig. 1.5; Shaw and Kvill, 1984; Shaw et al., 1989; Shaw, 1994, 1996). Shaw and Gilbert (1990) then went on to map drumlin fields in Ontario and New York to reconstruct patterns of regional subglacial meltwater flow (Fig. 2.3). From there, the megaflood theory has evolved into a controversial model hypothesizing that swarms of drumlins are the result of large amounts of meltwater released during catastrophic flooding events that scoured the bedrock as the water flowed under the last great ice sheets (Shaw, 1983). However, Benn and Evans (2005) go as far as to assert that “most Quaternary scientists give little or no credence to the megaflood interpretation, and it conflicts with an overwhelming body of modern research on past and present ice sheet beds.” They do not, however, provide evidence as to why the interpretation conflicts with an overwhelming body of modern research. This also has led to revisiting interpretations of other associated landforms such as tunnel valleys.
Figure 2.3. Regional flow patterns in southern Ontario and northern New York State of Algonquin and Ontarian subglacial flood events interpreted from drumlin long axis orientations (Shaw and Gilbert, 1990).
2.4 THE ROLE OF MELTWATER

While features resulting from quarrying and abrasion have been widely recognized and accepted as being the direct result of processes involving glacial ice moving over its bed, the role that water plays in these enigmatic glacial processes continues to draw a considerable amount of attention and has created a great deal of controversy which will be discussed later. Since Shaw’s landmark paper in 1983, attention has migrated away from the effects of direct glacial abrasion towards water as a major land-forming agent. The sheer volume of water released at the end of the LGM makes the significance of meltwater and other glaciofluvial processes hard to ignore. The volume of the LIS was $33 \times 10^6 \text{ km}^3$, the bulk of which was released as water in just over 10,000 years (Eyles, 2006). Glacial meltwater is the key to glaciers being able to move and to erode the landscape and is responsible for transporting significant volumes of sediment to the proglacial environment (Fig. 2.4). These sediments and landforms dominate large tracts of the glacial landscape in North America. Subglacial meltwater systems were instrumental in reworking large volumes of glacioclastic sediment into marine basins and it has been estimated that less than 6% of the total volume of these sediments produced during the Pleistocene remains on land (Eyles, 2006).
Figure 2.4. Schematic drawing showing relationship between tunnel valley to ice margin and proglacial fan (Hooke and Jennings, 2006).
Sources of meltwater and the drainage routes on top of (supraglacial), through (englacial) and under (subglacial) temperate and sub-polar glaciers are well documented. The largest volume of water that forms within the glacial system is the result of thawing ice during the warmer spring and summer seasons. This melt can result in large supraglacial rivers. Water that is not carried away on top of the ice will eventually infiltrate into the ice through complex plumbing systems within conduits called moulins (Fig. 2.5). This process is analogous to karst systems that are found within limestone formations. Meltwater also can be produced by the heat that is generated by friction as the water moves through the moulins and other conduits. In the early 1970’s, Röethlisberger and Nye defined two basic types of subglacial meltwater channel conduits (Fig. 2.5). The first, R-channels (Röthlisberger, 1972), are passages eroded upwards into the basal ice of a glacier, with little or no erosion of the substrate. Passages incised downward into the substrate, which may be sediment and/or bedrock, are called N-channels (Nye, 1973).

It also has been suggested that during periods of reduced melting of ice in the fall and winter seasons, that meltwater conduits within the glacier close off causing water to be stored (Hodgkins, 2001). Renewed input of water during the following spring and summer seasons will then reopen the conduits by enhancing the rate of deformation of basal ice. In the event that meltwater does not infiltrate the substrate nor does it evacuate at the ice margin, meltwater may be stored subglacially in enclosed basins such is the case in Iceland’s subglacial caldera lake, Grímsvötn and in Antarctica’s Lake Vostok. When ice-dammed lakes drain, the water may enter into the subglacial and englacial drainage system and carry the water away.
Figure 2.5. Meltwater drainage in temperate ice (modified from Brennand, 2000).
An enormous volume of research has addressed a meltwater genesis for a large set of subglacial landforms including s-forms, drumlins, fluting, hummocky terrain, Rogen moraine, megaripples, tunnel valleys, eskers and streamlined hills (Munro-Stasiuk et al., 2009). These landforms provide clues as to the extent, dynamics and history of former ice sheets (Kleman et al., 1997; Clark, 1999; Clark and Meehan, 2001; Boulton et al., 2001). It has been proposed that many streamlined landforms, drumlins for example, are actually part of a meltwater landform continuum that contains forms of many shapes and sizes (Munro-Stasiuk et al., 2009). Many s-forms have been reproduced in flumes (e.g., Allen, 1971; Shaw and Sharpe, 1987) and many similar forms also have been found in strictly fluvial environments (e.g. Karcz, 1968; Tinkler, 1993; Whipple et al., 2000) lending credence to the meltwater hypothesis for landform genesis.

The variety of features recognized within the meltwater landform continuum of Munro-Stasiuk et al. (2009) has been divided into four categories based on size: micro forms, meso forms, macro forms and mega forms. Micro forms (Table 2.1), such as muschelbrüche and sichelwannen, spindle forms, pot holes and rat tails, typically range in size from a few millimeters to a few tens of meters (e.g. Kor et al., 1991; Munro-Stasiuk et al., 2005). Drumlins, fluting, Rogen moraines, hummocky terrain, eskers and tunnel valleys are considered to be meso forms (Table 2.2). These occur on the scale of tens of meters to tens of kilometers in length and tens to hundreds of meters in width. Streamlined hills, reentrant valleys and bedrock rises are classified as macro forms (Table 2.3).
2.5 THE CONTROVERSIES

The controversy and debate surrounding the formation of tunnel valleys and other subglacial landforms is fueled by, but not limited to, the following questions:

- Was their formation the result of catastrophic releases of meltwater or steady-state flows?
- Were features formed synchronously or time-transgressively?
- Was it a sheet flow or channelized flow? Or both?
- What volume of water was necessary to form these landforms?
- What was the velocity and discharge of the water?
- Where could such large volumes of water have been stored?
- What would have been the climatic consequences of such flooding events?

It has been suggested that tunnel valleys and other landforms may have formed under a variety of conditions. The subglacial meltwater erosion hypotheses proposed for tunnel valley formation can largely be divided into two variants: (1) the steady state, progressive formation (e.g. Boulton and Hindmarsh, 1987; Mooers, 1989; Smed, 1998; Huuse and Lykke-Andersen, 2000; Praeg, 2003); and (2) the sudden formation by catastrophic outbursts (jökulhlaups) of subglacially accumulated meltwater (e.g. Wright, 1973; Ehlers and Linke, 1989; Brennand and Shaw, 1994; Patterson, 1994; Piotrowski, 1994; Björnsson, 1996; Clayton et al., 1999; Beaney, 2002). The former includes theories of slow removal of sediment, sediment deformation and small migrating erosive channels that requires several hundred to several thousand years (Mooers, 1989). The sediment deformation theory as inferred by Boulton and Hindmarsh (1987), predicts sediment creep towards small R-channels (Röthlisberger, 1972) accommodating steady-
state meltwater flow. The creeping sediments are gradually washed out by the flowing meltwater and, if this process continues for a long period, the deformable bed will be lowered and a large valley occupied by ice will form. Based on field relationships in a large system of tunnel valleys formed by the Superior Lobe in Central Minnesota, Mooers (1989) suggests that the dominant source of the water responsible for formation was seasonal meltwater from the glacier surface that reached the bed through moulins and crevasses (Fig. 2.5).

Mooers (1989) has suggested that tunnel valley networks are a result of relatively modest, steady state meltwater discharges that occurred over long periods of time, but agrees that discharges would have to be large enough to transport the large clasts found in some proglacial fans. Mooers (1989) further suggests that the influence of large, but not catastrophic, inputs from surface meltwater could be responsible for the sedimentary deposits found to be associated with tunnel valleys. However, others argue that because of the large amounts of sediment and clasts that are larger than 2 m in diameter found in some outwash fans, steady-state meltwater releases may not be sufficient to explain tunnel valley deposits (Fig. 2.6; Piotrowski 1994; Cutler et al., 2002).
Figure 2.6. Large boulders found in the Bryant Pit, northeastern Wisconsin (Cutler et al., 2002). Note person (circled in red) for scale.
Contrary to the steady-state theory, other tunnel valleys have been interpreted to have been formed by sudden releases of subglacially stored meltwater that simultaneously occupied entire channel systems with bank-full discharges (e.g. Brennand and Shaw, 1994; Beaney and Shaw, 2000; Beaney, 2002). Tweed and Russell (1999) acknowledge a host of mechanisms that regularly result in glacial outburst flooding. Perhaps the simplest cause of these outbursts is the overspill of ice-dammed lake water, initiated when the water depth in an ice-dammed lake exceeds the thickness of the damming ice and overtops the dam. Ice-dam flotation is another mechanism that can set off the drainage of ice-dammed lakes. Flotation potential is determined by the ratio of lake water depth to ice-dam height and density characteristics. Marcus (1960) and Higgins (1970) declared that during the drainage of ice-dammed lakes, water may enter the subglacial and englacial drainage system of the glacier forming the retaining ice dam, and carry the discharge away from the lake, a process they referred to as siphoning.

The catastrophic release hypothesis is favored by numerous scientists, who suggest an outburst triggered by the drainage of a surficial or subglacial reservoir, or even an extreme rainfall or melt event (Wright, 1973; Attig et al., 1989; Clayton et al., 1999). The strongest indications for catastrophic outbursts to be responsible for tunnel valley incision are: (1) the occurrence of boulder accumulations in outwash fans; (2) the anastomosing valley patterns and (3) bedforms indicative of subglacial meltwater floods (Wright, 1973; Ehlers and Linke, 1989; Brennand and Shaw, 1994; Piotrowski, 1994; Björnsson, 1996; Clayton et al., 1999; Beaney and Shaw, 2000; Beaney, 2002; Cutler et al., 2002). The formation of tunnel valleys as a result of large releases of subglacial meltwater, even during catastrophic flooding events, has been proposed for the Superior
Lobe in Minnesota (Wright, 1973); for lobes in Wisconsin (Attig et al., 1989; Clayton et al., 1999; Johnson, 1999); and Michigan (Fisher et al., 2005) and for other parts of the southern margin of the LIS (Brennand and Shaw, 1994). Down flow from the ice margin, these valleys are found to evolve into fluvially eroded landscapes or open onto proglacial outwash fans (Fig. 2.4; Hooke and Jennings, 2006), which are comprised of material that range up to boulder size (Patterson, 1994; Piotrowski, 1994). Cutler et al. (2002) documented clasts with intermediate axes of up to 2 m in gravel pits found in Wisconsin (Fig. 2.6). The sediments found in the proglacial outwash fans contain valuable information as they provide insight into the characteristics of the water that passed through the valleys as well as to their origin and the overall hydrology of the southern margin of the LIS (Cutler et al., 2002).

2.6 TUNNEL VALLEY GENESIS

Most studies regarding tunnel valley genesis have focused largely on the morphology of the valleys and the impact of subglacial drainage on the flow of the ice (Domack et al., 2006; Lewis et al., 2006). The relationship between meltwater, sediments and landforms is poorly defined and it is important to understand that sediments recovered from and observed within the valleys have the potential to reveal important information on valley formation and drainage processes. It has been suggested that hyperconcentrated flows associated with catastrophic discharge of water result in the deposition of massive and graded sands and gravels (Ó Cofaigh, 1996; Cutler et al., 2002; Fisher and Taylor, 2002). In general, however, the sediment facies associated with subglacial drainage networks has been comparatively under-studied (Benn and Evans, 1998).
Tunnel valleys are known to truncate drumlins and eskers, and if drumlins relate to erosional marks extending over the ice bed, this sequence represents a transition from a subglacial sheet flood to channelized flow in broad N-channels to flow in narrow R-channels (Shaw, 2002). This is considered to be the expected sequence for meltwater flow starting as a sheet flood in a subglacial environment and would indicate that a considerable amount of meltwater was involved. The existence and stability of sheet floods in combination with the requirement that such large quantities of water could be stored either subglacially or supraglacially leads to two of the controversies involved in the megaflood theory.

The pioneering work of Shoemaker (1992) on sheet floods drew criticism from Walder (1994) who argued that sheet floods were unstable and quickly broke down into channelized flow. Shoemaker (1994) responded that Walder’s calculations were based on laminar flow and that turbulent flows, such as outburst floods, have the ability to maintain sheet flows longer than laminar flows. Munro-Stasiuk et al. (2009) compare the formation of subglacial landforms with other notable forms that have recognized to have been formed by turbulent flows, such as yardangs which are widely accepted as the work of wind (Fig. 2.7A; Greeley and Iversen, 1985), sastrugi in snow and ice that result from wind sculpting (Fig. 2.7B; e.g. Herzfeld et al., 2003), erosional marks produced by tsunami waves (Fig. 2.7C; Bryant and Young, 1996), and the loess hills of the Channeled Scablands that were formed by enormous outpourings of water from Glacial Lake Missoula and the Cordilleran Ice Sheet system (Fig. 2.7D; Bretz, 1959; Baker, 1978; Shaw et al., 2000). The common factor for the forms is that they are considered to be the products of turbulent flow which contained vortices (Munro-Stasiuk et al., 2009).
Figure 2.7. Examples of landforms produced by turbulent flows. (A) Yardangs (image from Google Earth); (B) Sastrugi in snow (courtesy, the Royal Geographical Society); (C) Erosional marks produced by tsunami waves (courtesy of E. Bryant); (D) Streamlined loess hill in the Channeled Scablands. (From Munro-Stasiuk et al., 2009)
In general the origin of the valleys that exist at the southern edge of the LIS is not well understood and remains highly controversial in spite of a considerable amount of research. The controversy surrounding the debate will be discussed in more detail later, but includes such things as the amount of water necessary to form the valleys, where the water could have been stored and was their formation the result of catastrophic outbursts or steady-state conditions. Such debate is important as interpretations have a direct bearing on reconstructions of Late Wisconsinan ice-sheet dynamics and hydrology. Difficulties in accessing the subglacial environment of modern glacier systems, as well as the lack of a detailed sedimentary record as a consequence of the burial of sediments, have impeded direct research of this highly complex sedimentary system and created the controversy. These valleys have been called tunnel valleys (Ussing, 1903), tunnel channels (Woldstedt, 1923), valley channels (Cepek, 1968), and linear incisions (Ehlers and Wingfield, 1991). Not only is there controversy surrounding their formation, but also regarding the nomenclature used to describe them. Usage of the term ‘tunnel valley’ is not entirely straightforward and there are several who object to its use. For example, a ‘valley’ is supposed to have a falling thalweg, but tunnel valleys defy this feature (Ehlers and Wingfield, 1991). A ‘valley’ also is expected to be occupied by a small stream on the valley floor only and not by a bank-full flow implied in the term ‘channel’. The terms ‘tunnel valley’ and ‘tunnel channel’ are often used synonymously, but Clayton et al. (1999) proposed that ‘channel’ should be reserved for elongate depressions subglacially carved by bank-full discharges, whereas the term ‘valley’ should be used for elongate depressions carved by small subglacial conduits that are considerably narrower than the depression.
Tunnel valleys have been widely regarded as being carved by meltwater (Brennand and Shaw, 1994; Beaney and Shaw, 2000; Beaney, 2002; Munro-Stasiuk et al., 2009) and it also has been argued that drumlins have a meltwater origin (Munro-Stasiuk et al., 2009). The close association of drumlins, tunnel valleys and eskers points to a sequence: (A) formation of drumlins by broad meltwater flows; (B) flow concentration in channels and the formation of tunnel valleys as the sheet flow breaks down; (C) concentration of flow into conduits eroded upwards into the ice bed and the formation of eskers (Munro-Stasiuk et al., 2009).

Tunnel valleys often are found to truncate subglacial bedforms, contain subglacial landforms (e.g. eskers), follow uphill slopes and are the result of large subglacial rivers that carried meltwater and sediment from under past ice sheets (e.g., Wright 1973; Rampton 2000). Valleys usually are 10’s of km long, km’s wide, and 10’s-100 m deep. They often have flat bottoms and radial or anabranching valley systems that have been cut into bedrock or glaciogenic sediment, terminate in proglacial fans, and may be empty, partially filled or buried by sediments (Munro-Stasiuk et al., 2009). Many tunnel valleys ascend adverse bed topography suggesting that subglacial flow within the channels was pressurized and flowing down a hydraulic gradient controlled by ice-surface elevation (Clayton et al., 1999; Johnson, 1999; Cutler et al., 2002; Hooke and Jennings, 2006; Jørgensen and Sandersen, 2006).
2.7 DRUMLINS, ESKERS AND KAMES

Drumlins are the epitome of subglacial landforms and are considered to be one of the classic symbols of continental glaciations. They are easily recognizable in the glacial landscape and are often found in swarms, or drumlin fields, which may contain many hundreds, sometimes thousands of these landforms, as is the case in the Chautauqua Drumlin Field (Fig. 1.2). Drumlins are asymmetrical in plan view, highest at their proximal blunt ends, and taper in the downflow direction on their lee sides (Fig. 2.8). As with other glacial bedforms, the conditions under which they form are most often obstructed and inaccessible. This has led to a considerable amount of controversy regarding their formation. Several researchers, including Shaw (1983, 1996), Shaw and Kvill (1984), Shaw et al. (1989) and Rains et al. (1993) have concluded that subglacial megafloods were responsible for carving drumlins, flutings, meltwater channels and a variety of other associated landforms beneath the southwest margin of the LIS. The distribution of drumlins on interfluves between tunnel valleys, both of which are part of an unconformity (Sharpe et al., 2004), suggests that both drumlins and tunnel valleys are products of erosion by the same agent (Munro-Stasiuk et al., 2009). The meltwater hypothesis for drumlin formation proposes that drumlins are created either by direct fluvial erosion of the bed, or by fluvial infilling of cavities formed as erosional marks in glacier beds (Fig. 2.9; Shaw, 1996).
Figure 2.8. The classic streamlined drumlin with blunt stoss side and tapered lee side (Munro-Stasiuk et al., 2009).
Figure 2.9. A model illustrating the components of the meltwater hypothesis. The bedforms result from interactions between the ice bed, the ground surface and the broad meltwater flow (Shaw, 2002).
In addition to tunnel valleys and drumlins, this study also investigates the relationship of eskers to these landforms. Eskers are glaciofluvial ridges of stratified sand and gravel that are inferred to be the casts of subglacial channels (R-channels; Röthlisberger, 1972) or reentrants into the ice front (Banerjee and McDonald, 1975). The distribution, morphology, and sedimentology of eskers have been instrumental in determining the dynamics and hydrology of former ice sheets that covered North America and Europe (e.g., Shreve, 1985; Clark and Walder, 1994; Brennand, 2000; and Boulton et al., 2009). Hypotheses regarding their origin have been based primarily on morphological and sedimentary analyses of ancient eskers (e.g. Brennand and Shaw, 1996) or on glaciological theory (Clark and Walder, 1994; and Boulton et al., 2009). Most studies conclude that eskers form during multiple cycles of deposition over long periods of time, with the role of outburst floods being overlooked (Burke et al., 2010).

The presence and distribution of eskers is determined by a combination of factors: (1) meltwater supply (location of crevasses, moulins and reservoirs); (2) sediment supply; (3) the nature of the basal substrate (as it controls the style of the subglacial plumbing system; Clark and Walder, 1994); and (4) the presence and/or drainage of proglacial water bodies (Brennand, 2000). They are commonly found at the margins of former ice sheets or glaciers in the Northern Hemisphere and can be up to several hundred kilometers long. The formation of these long, narrow, often sinuous ridges requires a considerable amount of sediment and large volumes of meltwater (Fig. 2.10). Eskers, which are usually much smaller than the tunnel valleys in which they occur, link valley formation to the subglacial environment and implies a change from erosional to depositional conditions (Kehew et al., 2013) during ice retreat or stagnation (Fig. 2.11).
Fig. 2.10. Sketch by Mannerfelt (1945) showing the form, internal structure and sediments in a model esker (www.landforms.eu).

Figure 2.11. Schematic showing processes related to formation of eskers (www.landforms.eu).
As with drumlins and eskers, kames also are formed by the deposition of sediment via glacial meltwater as the ice becomes stagnant or recedes (Figs. 2.12). Kames often occur in association with kettle holes in kame and kettle topography. Kettle holes may be filled with water, and represent areas of subsidence caused by the melting of buried ice. Kame and kettle topography may exhibit some degree of lineation (Benn and Evans, 1998). The association with tunnel valleys and eskers indicates their formation proximal to the ice margin in situations where there are large volumes of meltwater and sediment. Kames are predominantly composed of bedded and sorted sand and gravel, but often have sharp lateral variations that indicate rapid changes in flow velocity. Kame deposits often contain foreset bedding as the result of deposition of sediment in proglacial lakes (Fig. 2.13), and may have internal folding and faulting due to the removal of the supporting ice.
Figure 2.12. Illustration of the development of kames, which begins with the melting of stagnant ice in a valley that has a marginal stream along one side and a marginal lake along the other side. After ice melts, a kame terrace with kettles remains along the valley wall, an esker where a tunnel formerly existed, two kames where the stream flowed on the ice, and a kame delta where sediment built into a marginal lake (after A. N. Strahler, 1973).
Figure 2.13. Schematic showing processes related to formation of kame and kettle topography. Note development of foreset beds as water flows under the ice into a proglacial lake (www.landforms.eu).
The interpretation of the sequence of events related to the formation of this suite of landforms has evolved from sedimentological analysis and comparisons of subglacial landforms with erosional forms produced both by experiments in wind tunnels and flumes and in the boundary layers of contemporary fluvial and aeolian environments (Evans et al., 1999). Due to the enigmatic nature of the formation of these features, it is possible that there may never be enough evidence to convincingly explain their origin, making it necessary to base interpretations of these landforms and the related glacier dynamics on answers provided by comparing and analyzing modern analogs.

Form analogy has proved to be instrumental in understanding the processes involved with past events. James Hutton’s concept of uniformitarianism - The present is the key to the past - dates back to the late 18th century. With advances in technology and in the field of glacial sedimentology has come a variety of interpretations pertaining to landform genesis (Mickelson and Colgan, 2003). A paper by Clayton and Moran (1974) established the idea of the process-form model in glacial geology. They emphasized that the modern glacially-derived landscape form is directly related to the glacial processes active in the past. Therefore if we study glacial processes occurring today and compare these with past ice behavior, we should see that the processes are analogous. However, the challenge to determine which of the different approaches is responsible for the formation of the variety of glacial features lies in how well each approach explains the individual landform characteristics, while still keeping sight of the basic physical principles that must be involved in their creation (Shaw, 2002).
2.8 MODERN ANALOGS

In spite of the wealth of literature concerning Quaternary tunnel valley development, until recently there were no examples of their formation within contemporary glacial systems. This gap has commonly been attributed to a lack of appropriate modern conditions for tunnel valley genesis or due to the inaccessibility of the subglacial environment (Jørgensen and Sandersen, 2006). However, subsequent retreat of the southeast margin of Skeiðarárjökull following the 1996 jökulhlaup, has revealed the first modern example of a tunnel valley (Fig. 1.8A and B) excavated into the surrounding moraine sediment, which demonstrates that glacier outburst floods are capable of generating tunnel valleys (Russell et al., 2007). This jökulhlaup had a sufficiently rapid onset to induce water pressures high enough to force large volumes of water to ascend from a heavily over-deepened glacier basin (Russell et al., 2007). Russell et al. (2007) suggest that the presence of proglacial lakes will increase the variability of geomorphologic response and sedimentary signature of glacier outburst floods over distances as little as a few hundred meters. These floods also have the potential to produce other large-scale landforms, such as drumlins and eskers, which is supported by estimates of the dimensions of paleofloods associated with the LIS and mapped by erosional and depositional evidence (Shaw and Kvill, 1984; Rains et al., 1993; Shaw, 1994, 1996).

The 1996 Skeiðarárjökull jökulhlaup left an extensive suite of features that can be studied and related to Quaternary landform assemblages. In the case of tunnel valleys/channels, Russell et al. (2007) note that similarities between the Sæluhúsakvísl, Iceland, tunnel valley and Quaternary tunnel valley systems include: up-glacier slopes; associated outwash fans displaying rapid downstream fining and apex incision;
dissection of end moraines and a characteristic ‘box’ shape (Clayton et al., 1999; Cutler et al., 2002; Kozlowski et al., 2005).

During the jökulhlaup, meltwater flows within the Sæluhúskvísl channel were noted to have left a considerable number of ice blocks when compared to adjacent outlets on the sandur (Fig. 2.14; Russell et al., 2007). The widespread break up of glacier margins during jökulhlaups makes ice-block release a characteristic jökulhlaup impact (Roberts, 2005), with floodwaters often depositing large (~30 m diameter) blocks of ice across the proglacial zone (Burke et al., 2010). Evidence for large-scale erosion of the glacier substrate during the flood takes the form of numerous intraclasts or rip-ups found in en- and proglacial jökulhlaup deposits (Fig. 2.15) as well as a heavily-kettled proglacial outwash fan (Fig. 2.16; Russell et al., 2007). The tunnel valley that formed as a result of this outburst flood has confirmed that a close relationship exists between tunnel valley erosion and proglacial outwash deposition.

The onset of this outburst flood was significant enough to induce water pressures that were high enough to force large volumes of water up from a heavily over-deepened glacier basin (Russell et al., 2007). Many Quaternary tunnel valleys ascend to ice margins from over-deepened basins (Clayton and others, 1999; Cutler and others, 2002; Jørgensen and Sandersen, 2006) and it is posited that over-deepened basins may actually enhance the formation of tunnel valleys since the pressurized, ascending subglacial meltwater will seek the most efficient route through highly erodible and complex glacier substrate (Russell et al., 2007). Several researchers also propose that the presence of an over-deepened basin may help to retain meltwater within subglacial lakes (e.g. Shoemaker, 1992; Alley et al., 2006; Domack et al., 2006), providing a source for release during tunnel valley-forming outburst events. This study by Russell et al.
(2007) reports on the significance of meltwater outbursts as agents of tunnel valley formation and provides valuable modern analog for a process which has until now only been inferred from the Quaternary record.

Figure 2.14. Photograph showing Skeiðaràrsandur in 2007 looking east. Dr. Andy Russell is holding a picture of the sandur, which is east of the site used in this study, after the 1996 jökulhlaup showing large ice blocks that have since melted, creating depressions seen in the distance.
Figure 2.15. Photograph of rip-up clast of till on the outwash fan at the site of the 1996 jökulhlaup, Skeiðarârsandur, Iceland. Photograph taken in 2007.

Figure 2.16. Photograph taken in 2007 of numerous kettle holes that resulted from the 1996 jökulhlaup at Skeiðarârsandur, Iceland.
Eskers, as is the case with other subglacial landforms, have been used to gather information regarding the dynamics and paleohydrology of large ice sheets. Interpretations of esker genesis have been attributed to both steady-state and transient deposition within persistent streams (e.g. Banerjee and McDonald, 1975; Gorrell and Shaw, 1991; Brennand, 1994; Warren and Ashley, 1994; and Boulton et al., 2007A, B). Skeiðararjökull and Sölheimajökull, Iceland, provide the only modern examples of eskers known to be deposited during jökulhlaups (Russell et al., 2006). Burke et al. (2008) used GPR to identify the controls on the large-scale sedimentary architecture of a proto-esker deposited during the 1996 jökulhlaup at Skeiðararjökull (Fig. 2.17). They use the term proto-esker since buried ice remained beneath the landform at the time of their study and the landform had yet to attain its post-depositional geometry. Based on the GPR surveys completed on the proto-esker, they concluded that a single high-magnitude jökulhlaup is capable of forming an esker on a large scale (Burke et al., 2008).
Figure 2.17. Photograph taken in 2007 of esker formed during the 1996 jökulhlaup at Skeiðarársandur, Iceland.
Antarctica also exhibits characteristic meltwater features. Valley systems in southern Victoria Land have been determined to be the result of massive meltwater outbursts comparable in scale to the largest lake drainage events in the Northern Hemisphere (Fig. 2.18; Denton and Sugden, 2005). This valley system is 5 km long and 5 km wide with valleys that are as much as 100 m deep and have anastomosing patterns with up-and-down long profiles, characteristic of pressurized subglacial meltwater as has been interpreted for features found in contact with the LIS. The valleys also are associated with large potholes and plunge pools, some littered with large blocks of dolerite. Denton and Sugden (2005) attribute the outbursts to the accumulation of water in a topographic depression beneath areas of the ice sheet with basal ice at the pressure melting point. If the volume of water and the ice surface gradient crosses a critical threshold, the water will propagate down-ice at the ice–rock interface under the influence of ambient pressures within the ice (Shreve, 1972).
Figure 2.18. Photograph of channel system that forms the Labyrinth in western Wright Valley, Victoria Land, Antarctica. Channels are cut in dolerite. Individual channels are up to 100 m deep. Some potholes on the interfluves are littered with dolerite blocks, some of which are imbricated. An area described as a channeled scabland is to the left (Denton and Sugden, 2005).
2.9 WATER STORAGE

The best known modern examples of subglacial lakes are in Iceland and Antarctica, but areas where substantial volumes of meltwater may have been stored within the Quaternary environment remains a source of debate.

In Iceland, as discussed previously, water related to the 1996 outburst flood was stored in the subglacial lake Grímsvötn. In Antarctica, the discovery of approximately 145 subglacial lakes under the ice sheet (Fig. 2.19) has increased awareness surrounding the possibility that large volumes of subglacial meltwater also may have been stored under past ice sheets. Antarctic subglacial lakes were first identified by Robin et al. (1970) after airborne radio echo sounding (RES) investigations of the ice-sheet interior. Measurements derived from these studies have identified the total volume of water stored as lakes beneath the ice sheet and provides an analogue for conditions at the base of the large ice sheets which developed over mid- and high northern latitudes during Quaternary glacial conditions. It is possible that comparable water systems may have existed beneath the LIS, which was similar in size to the modern Antarctic Ice Sheet and overrode numerous deep bedrock basins suitable for development of subglacial lakes (Evatt et al., 2006).
Figure 2.19. Map showing location of lakes beneath the Antarctic Ice sheet (marked by triangles, identified from 60 MHz airborne radio-echo sounding records (Siegert et al., 1996).
The largest Antarctic subglacial lake, Lake Vostok, exists beneath 4 km of ice (Eyles, 2006) under Vostok Station. It is approximately 230 km long, up to 50 km wide and 500 m deep with an estimated surface area of 14,000 km$^2$, making it one of the world’s largest lakes (Siegert, 1999) comparable to Lake Ontario (Kapitsa et al., 1996). Recent analysis of satellite data has demonstrated the movement of considerable volumes of water beneath today’s Antarctic ice sheets (Gray et al., 2005; Wingham et al., 2006; Fricker et al., 2007) potentially in well-organized channelized systems (Wingham et al., 2006; and Siergert et al., 2007). Wingham et al. (2006) discovered that lakes drained subglacially very rapidly from one to another, which infers that lakes beneath the Pleistocene ice sheets may also have drained intermittently (Munro-Stasiuk, 1999). These observations indicate that large meltwater reservoirs may have filled along flood paths and then released as outburst floods.

Noting that jökulhlaups often are the result of water released from subglacial lakes, Alley et al. (2004) have promoted a hypothesis pertaining to the processes for the accumulation and evacuation of water from subglacial reservoirs. Acknowledging that glaciers frequently erode and occupy over-deepened basins, Alley et al. (2004) note that climatically or dynamically driven glacier fluctuations will allow for the retreat from and subsequent readvancement of the ice into these basins, which often are occupied by water. They infer that if the cooling of the atmosphere is sufficient, an ice shelf may form during readvancement, followed by ice-shelf thickening and grounding of the ice on the sill. Grounding of the ice would then result in water trapped in the basin and restrain ice flow so that the ice thickens over the water and raises its pressure. The authors go on to say that the base of the frozen-on ice then is likely to warm, which would lead to thawing and perhaps loss of the reversed ice-air surface slope. This would initiate the release of
at least some of the trapped water and speed up the ice flow creating an outburst flood with an ice surge (Fig. 2.20).

Figure 2.20. Sketch of the main steps in Alley et al. (2004) hypothesis. (1) Situation before ice-shelf grounding; (2) The situation after ice-shelf grounding on the sill, with reversal of the ice-air interface and increasing ice thickness over the subglacial lake; and (3) Release of the outburst flood.
In Quaternary landscapes, Beaney and Hicks (2000) attribute the southeast Alberta, Canada, landscape to the drainage of Livingstone Lake and Shoemaker (1999) hypothesized that many of the Great Lake basins could have stored enough water to form tunnel valleys and associated features in that region. On several occasions, lake levels were affected by catastrophic influx of meltwater from glacial Lake Agassiz which developed outside of the Great Lakes basin (Teller and Thorleifson 1983; Teller 1985; Leverington et al., 2000). Evatt et al. (2006) note that the subglacial terrain of the former LIS is relatively flat, and the presently existing large lakes (the Great Lakes, Great Bear Lake, and Great Slave Lake) are the best candidates for large former subglacial lakes (Fig. 2.21). Other researchers have stated that catastrophic drainage of Lake Agassiz and Lake Algonquin was responsible for the formation of drumlins, tunnel valleys and other features found at the southern limit of the LIS.
Figure 2.21. A ‘likelihood map’ of subglacial lakes of the LIS. Darkness of blue approximates increased likelihood of a subglacial lake in that position. Existing large lakes are also shown (stippled blue) as these likely housed subglacial lakes during the LGM (Evatt et al., 2006).
The Channeled Scablands of eastern Washington are thought to have been caused by a succession of catastrophic floods from the ice-dammed proglacial Lake Missoula (Bretz 1923; 1969), from an overspill of Lake Bonnevile (Malde, 1968), and from jökulhlaups of other ice-dammed lakes in the Columbia system (Baker and Bunker, 1985). Outbursts from the Cordilleran Ice Sheet itself also could have been involved (Shaw et al., 2000). When water in glacial Lake Missoula reached a critical depth, the ice dam became buoyant, tunnels opened beneath the dam, and the lake drained catastrophically (Waitt, 1985). These floods are thought to have been some of the largest on Earth, causing the gouged terrain that remains today.
2.10 CLIMATIC CONSEQUENCES

Questions arise relating to the climatic conditions that would have been associated with the release of meltwater from floods of such magnitude. Runoff from the melting of the LIS to the Gulf of Mexico represents the largest and best-documented meltwater episode in the marine record (Brown and Kennett, 1998). Approximately $18 \times 10^6 \text{ km}^3$ of meltwater flowed through the Mississippi River into the Gulf of Mexico between 14.0 and 11.0 ka, therefore contributing to a rapid rise in sea level (Teller, 1990; Fairbanks, 1989). Evatt et al. (2006) propose that a sequence of rapid climatic change events over central Greenland during the last ice age, termed Dansgaard-Oeschger events, are linked to rapid changes in the North Atlantic Ocean circulation driven by changes to the freshwater input to the North Atlantic Ocean. They suggest that the most likely source of freshwater is from subglacial or proglacial lakes in the form of jökulhlaups. High volume Quaternary jökulhlaups are thought to have forced abrupt climatic changes by disrupting ocean circulation patterns with the addition of the cold, fresh water (Clark et al., 2001; Fisher et al., 2002).

2.11 SUMMARY

The studies mentioned above have been concerned mainly with high energy, short-duration meltwater outbursts, which do not generally occur across the entire ice front, but which may occur on a regional to continental scale (i.e. Björnsson, 1974; Shaw, 1989; Shaw and Gilbert, 1990; Rains et al., 1993; Russell and Knudsen, 1999). Knight (2002) noted that these high-magnitude outbursts are only one end-member of a range of subglacial meltwater events that can occur on different spatial and temporal scales. Lower-magnitude meltwater events, which may be subglacially confined and occur over longer time periods (months to decades rather than hours to weeks), have
been less well studied. However, Knight (2002) notes that on the basis of their longer existence, these events may be more characteristic of the mean subglacial hydrological conditions of temperate ice sheets.

Whatever the mechanism, whether there were bank-full or less than bank-full conditions, catastrophic or time-transgressive release, it is generally believed that tunnel valleys served as major subglacial drainage pathways for large volumes of meltwater, and are thought to play a considerable role for the entire hydraulic system beneath glaciers. Because glacier behavior largely reflects the subglacial hydraulic regime, the understanding of how tunnel valleys and associated landforms are created, is crucial for the reconstruction and understanding of former ice sheets. However, the inaccessibility of the subglacial environment and current climatic conditions makes it necessary to rely on form analogy and inferences in this reconstructive attempt. The relationship between large channels cut by the flow of subglacial meltwater and other aspects of the subglacial drainage system, such as areas of ice-bed contact, areas of ice-bed separation and precipitate-filled depressions provides vital information relevant to the reconstruction of former subglacial drainage conditions (Walder and Hallet, 1979; Sharp et al., 1989).
CHAPTER 3: METHODOLOGY

3.1 INTRODUCTION

In this study, Google Earth, topographic maps, Digital Elevation Models (DEM)s, aerial photography and the GIS application package *RiverTools*, were initially used to locate and measure a variety of parameters relating to the tunnel valleys and topography in northwestern Pennsylvania. DEMs are digital representations of cartographic information in a raster form. They consist of a sampled array of elevations for a number of ground positions at regularly spaced intervals. This information was obtained from the National Elevation Dataset (NED) of the United States Geological Survey, a seamless DEM for the United States based on 7.5' topographic mapping. *RiverTools* is a GIS application for analysis and visualization of digital terrain, watersheds and river networks. The software has the ability to extract drainage network patterns and analyze hydrologic data from very large DEMs. It provides accurate measurements of river and basin characteristics such as upstream area, channel lengths, elevation drops and slope.

3.2 PRE-FIELDWORK PREPARATION

Prior to entering the field, as much spatial and topographic information as possible was gathered about each site. The methods were slightly different for the two sites since the data available are different for each location. For Pennsylvania field investigations, aerial photographs (Fig. 3.1), topographic maps (Figs. 3.2 and 3.3) and Google Earth (Figs. 3.4 and 3.5) were utilized extensively. Before traveling to Iceland, data mostly were obtained via moderately high resolution aerial photos taken in 1997 (Figs. 3.6 and 3.7). No DEM data or recent high-resolution imagery exists for the site.
Since USGS topographic maps use contour lines to show elevation, these maps provided useful information about slope steepness, location of towns, and gravel pits, and details of access roads. Since many topographic maps are quite old, it was necessary to verify gravel pits with Google Earth. It was not uncommon to locate a gravel pit on the topographic map only to find that it no longer existed or to find a relatively new pit that was not depicted on topographic maps from several decades ago, as was the case with the quarry surveyed southwest of Waterford.
Figure 3.1. Aerial photograph of Erie and Crawford counties, PA, showing quadrangles for possible field investigations (PASDA).
Figure 3.2. USGS topographic map of study area. Select drumlins east of study area highlighted with red lines to note parallel alignment of features. Valley in which Edinboro sits highlighted with red lines. Gravel pits circled in red.

Figure 3.3. USGS topographic map of site southwest of Waterford. Large red circle indicates private quarry used in study. Topographic contours show that quarry sits within a valley and a drumlin exists to the northwest.
Figure 3.4. Google Earth image of site southwest of Waterford that shows location of gravel pits in the area and their relationship to the quarry used in this study.

Figure 3.5. Google Earth image zoomed into location of Glover quarry.
Figure 3.6. 1997 aerial photograph of the western edge of Skeiðarásandur with inset of study area.
Figure 3.7. Close-up of study area showing major drumlinized surface. To the south is the 1929 surge moraine and to the north in the 1996 flood channel.
Initially, 20 USGS 7.5-minute quadrangles located in Erie and Crawford counties were selected as possible research sites (Fig. 3.1). After field reconnaissance, areas near Edinboro and Waterford were selected for the use of ground-penetrating radar surveys due to accessibility, their relationship to the valleys and the permission from a quarry owner to survey his property. In contrast to the use of topographic maps and other media to determine the Pennsylvania study area, the Icelandic study area was chosen based on the recommendation of Dr. Andrew J. Russell, School of Geography, Politics and Sociology, University of Newcastle upon Tyne, United Kingdom. Dr. Russell has done considerable research in Iceland and is extremely familiar with features on Skeiðaràrsandur. He recommended this site because of the existence of a channel in the area.

3.3 DEM ANALYSIS

The DEM of the Chautauqua Drumlin Field (Fig. 3.8) shows an anabranching network of valleys that dissect the drumlinized ridges throughout northwestern Pennsylvania and western New York. This anabranching pattern meets one of the criteria used to identify tunnel valleys. To determine other morphological parameters that would identify these features as tunnel valleys, such as uphill longitudinal profiles, data from the DEM were input into the RiverTools software. A total of six valleys were analyzed (Fig. 3.8) to determine cross-sectional (Fig. 3.9) and longitudinal profiles (Fig. 3.10).
Figure 3.8. DEM identifying six valleys analyzed in RiverTools to determine shape and longitudinal profiles. Black squares denote location of Edinboro and Waterford sites.
Figure 3.9. West to east cross section of channel near Edinboro calculated using DEM and RiverTools application.

Figure 3.10. North to south longitudinal profile of valley A created using the DEM and RiverTools application. Note uphill slope.
3.4 PRINCIPLES OF GPR

GPR is a non-invasive geophysical tool used for imaging and accessing information regarding subsurface sediments and features. Such information is very important for different types of studies, varying from those related to archeological research to those studying geological elements of bedrock (Aranha et al., 2002) and sorted sediments such as fluvial, glaciofluvial, aeolian, or coastal bar deposits (e.g. Froese et al., 2005; Moore et al., 2004). In archeological studies, GPR has been successful at locating buried structures (Sternberg and McGill, 1995; Hruska and Fuchs, 1999; Leckebusch, 2003), graves (Schultz, 2007), and urns (Cezar et al., 2001). GPR can image stratigraphy in shallow subsurface sand, gravel, and peat with the best results obtained in clean quartz-rich, thick, clastic sediments that are free of silt and clay (Jol and Smith, 1991). In sedimentary geology, GPR is used primarily for stratigraphic studies where near-continuous, high-resolution profiles aid in the ability to determine (1) stratigraphic architecture, (2) sand-body geometry, and (3) correlation and quantification of sedimentary structures (Bristow and Jol, 2003). GPR also has been utilized to determine the thickness of glacier ice, structure and bed configuration (e.g. Moorman and Michel, 2000), the structure of karst terrain (e.g. Chamberlain et al., 2000), water table configuration (e.g. Doolittle et al., 2006; Turesson, 2006), soil water content (Huisman et al., 2001), environmental contamination (Davis and Annan, 1989; Lawton and Jol, 1994), and the determination of permafrost thickness (Arcone and Delange, 1987; Fisher et al., 1989).

GPR obtains subsurface images based upon the principles of electromagnetic theory, in which pulses of electromagnetic radiation, mostly in the UHF (Ultra High Frequency) and the VHF (Very High Frequency) microwave bands, are transmitted into
the ground and are absorbed, reflected or attenuated by the subsurface materials. The receiver will then record the amount and strength of the reflected waves and measure the time a signal needs to travel to an interface and back to the surface. However, since the waves have the ability to be absorbed, attenuated, and reflected by subsurface structures, the time and strength of the signal at the time it is recorded can vary considerably. Interfaces become detectable with GPR when there are changes in soil composition and compaction, as well as rock type.

The depth to which radar energy can penetrate and the amount of resolution that can be expected is dependent upon the electrical conductivity of the ground combined with the frequency, and therefore the wavelength, of the radar energy transmitted (Conyers and Goodman, 1997). Standard GPR antennae propagate radar energy that varies in frequency from 10 to 1000 MHz. Higher frequencies give higher resolution results, but do not penetrate far into the ground, whereas lower frequencies penetrate further into ground but have lower resolution. Low frequency antennae (10-120 MHz) are capable of generating long wavelength radar energy that has the ability to penetrate up to 50 m in certain conditions, but can only resolve very large buried features (Conyers and Cameron, 1998). In contrast, the maximum depth of penetration of a 900 MHz antenna is about one meter or less in typical materials, but its generated reflections can resolve features with a maximum dimension of a few centimeters (Conyers and Cameron, 1998). Therefore, a tradeoff exists between the depth of penetration and the subsurface resolution. Optimal depth penetration is achieved in dry sandy soils or massive dry materials such as granite, limestone, and concrete (Leckebusch, 2003) and can sometimes reach almost 70 m in depth (Smith and Jol, 1995). In contrast, in moist and/or clay-laden soils and those with high electrical conductivity, penetration may be
only a few centimeters. However, fortunate combinations of antenna frequencies, and the nature of the material being studied, GPR has been shown to characterize sediments and bedrock up to almost 70 m deep (Smith and Jol, 1995).

In this study, GPR data were collected using a pulseEKKO Pro system with interchangeable antennae. Icelandic surveys were collected using 50 and/or 100 MHz antennae, which in this environment, allowed depth penetration to about 18 m with the 50 MHz antennae and 9 m with 100 MHz. In Pennsylvania, where the majority of transects were run on roadways over asphalt, 100 MHz antennae were mounted on a pushable cart, and achieved a depth penetration to about 9 m.

The relative dielectric constant, an expression of the extent to which a material concentrates electric flux, is the most important physical parameter of the subsurface material being imaged (Leckebusch, 2003). Different dielectric constants will result in the emitting waves traveling through materials at different speeds and reflections appear at boundaries with different dielectric constants. In this study all data were calibrated to an average wave velocity of 0.100 m/ns which is the average for sand, the main sediment type across the sandur. Data were captured in reflection mode, with the transmitter and receiver separated by 1 m (the standard separation for 100 MHz), and a step size of 25 cm, while a step size of 50 cm was used for the 50 MHz antennae. The step size is the distance that the antennae are moved between traces. All data have a permanent DE-WOW filter applied, which removes unwanted low frequency interference. Additionally, an Automatic Gain Control (AGC) of 500 was applied to the dataset. With AGC gain, each data trace is processed such that the average signal is computed over a time window and then the data point at the center of the window is amplified (or attenuated) by the ratio of the desired output value to the average signal amplitude.
(Sensors and Software, 2003). All data were topographically corrected and analyzed in EkkoView2 and EkkoView Deluxe. Elevations necessary for topographic corrections for Pennsylvania data were determined through the use of Google Earth, while Iceland elevations were calculated by simple field surveys.

3.5 GPR DATA COLLECTION

GPR data were collected throughout July 2008 in Iceland on Skeiðarårsandur with the help of volunteers working with EarthWatch Institute. A total of 107 transects were completed over the entire site and were run perpendicular and parallel to predicted meltwater flow paths. Only transects associated with a few specific areas within the study area were actually analyzed for this study (Fig. 3.11). All data were calibrated to a wave velocity of 0.100 m/ns which provided an overall depth range of approximately 3.5-18 m. All data were collected over the rugged terrain using three volunteers to operate the system in step mode (Fig. 3.12), with 1 m spacing between each trace (each pulse of radar sent into the ground) and a step size of 25 cm. In step mode, the GPR unit is positioned, data are acquired, and the equipment is then moved to the next position, and then data are acquired again. This is repeated until data are collected from all desired locations.
Figure 3.11. Close up of study site on Skeiðaràrsandur with location of major landforms chosen for analysis.
Figure 3.12. Photograph of EarthWatch Institute volunteers collecting GPR data in step mode on Skeiðarársandur, Iceland.
In Pennsylvania, where data were collected on open roads over considerable distances, the pulseEKKO Pro and Digital Video Logger (DVL) were mounted on the accompanying SmartCart, a self-contained mobile platform that allows for fast data acquisition (Fig. 3.13) and then pushed along the roadside, using a car in front of the cart and a car behind for safety.

Figure 3.13. The pulseEKKO Pro and Digital Video Logger (DVL) mounted on the SmartCart.
3.6 GPR DATA ANALYSIS

In Pennsylvania, GPR transects were several thousand meters in length. In order to be able to determine the subsurface structure of each line, data were topographically corrected and “chopped” into smaller segments using EKKOView Deluxe. An example of data processing is given in figure 3.14 which shows the 3200 m long, west-east transect, followed by a series of chopped images. The small, numbered triangles on image are fiduciary marks that denote specific features at the surface, i.e. bridges, culverts, intersecting roads, etc.

![Figure 3.14. Above, entire Line00-Crane Road transect near Edinboro, PA. The following images are smaller segments of Line00-Crane Road, to illustrate the process of ‘chopping’ in EKKOView Deluxe.](image)
Data were then viewed in EkkoView2 which allows visual manipulation by changing the color pallet so that features may be made more readily identifiable (Fig. 3.15). In this study, most images were viewed using either the grey or bone setting on the color palette.

Figure 3.15. Examples of how line data can be displayed in EKKOView2 using two different color schemes. A - bone; B - seismic.
Raw data were imported into EkkoView Deluxe for processing. During processing a DE-WOW filter was applied, a gray or bone color scheme was again used to make viewing the data easier, and an automatic gain control (AGC) of 500 was used to enhance reflectors. Some gained GPR data were then topographically corrected using information from simple field surveys or elevations determined through Google Earth. Other transects were not topographically corrected due to slight differences in elevation and the inability to adequately survey these areas.

Figures 3.16A, B and C represent three stages of processing data for analysis: (A) shows Line 12 as unprocessed data; (B) after AGC of 500 has been applied; and (C) after topographic correction.
Figure 3.16A. Unprocessed GPR data.

Figure 3.16B. GPR data with AGC of 500 applied.

Figure 3.16C. GPR data after AGC and topographic correction were applied.
4.1 INTRODUCTION

Figure 4.1 shows the entire Skeiðaràrsandur site as viewed from the moraine complex which borders the study area to the south. Skeiðaràrjökull can be seen in the background. Identification of the features labeled in this figure will be justified below.

Figure 4.1. View of study area looking north from moraine complex at Skeiðaràrsandur, Iceland.
4.2 INITIAL TRENCHING

Prior to running GPR, a trench, approximately 1.5 m deep and 15 m long, was excavated to get a sense of the subsurface structure (Fig. 4.2). Figure 4.3 shows a lower unit of stratified sand that is sharply overlain by poorly sorted sand and imbricated, sub-rounded to -sub-angular gravel. The trench reveals a boulder that was deposited within the lower unit of stratified sand (Fig. 4.4).

Figure 4.2. Trench used to examine sedimentary architecture prior to running GPR on Skeiðaràrsandur.
Figure 4.3. Sedimentary structure of the subsurface at Skeiðarârsandur.

Figure 4.4. Boulder deposited within stratified sand at bottom of section.
4.3 THE DELTA

Figure 4.5A. View of delta with moraine complex in the background, looking southeast, with paleoflow approximately to the south along Line 12.

Figure 4.5B. View of delta looking west. Dashed line highlights delta front.
The feature labeled “delta” in Fig. 4.1 was identified as a delta on the basis of its triangular fan shape, its flat top, its steepened front, and its position at the downstream end of the channel. Line 12 is a 92 m transect from north to south that parallels water flow (Fig. 4.5A, B), and was surveyed using 50 and 100 MHz antennae. Depth of wave penetration was approximately 10 m with a wave velocity of 0.100 m/ns with the 100 MHz antennae. The GPR cross section produced in EkkoView Deluxe (Fig. 4.6) has been topographically corrected and shows high-angle, down-flow dipping reflections, or foreset beds. Discontinuous undular reflections are also visible at depth. Since this transect continued up to the moraine complex, moraine sediments that dip in the opposite direction from the foreset beds also are visible as well as an older moraine complex that has been overlapped by foreset beds. Toward the southern end of the transect, an approximately 20 m concave-up feature possibly represents a channel that has since been filled with sediment.
Figure 4.6. GPR survey of delta Line 12 parallel to flow, from north to south.
A 25 m east-west transect of the delta (Fig. 4.7) also was surveyed using 50 and 100 MHz antennae. The 100 MHz image shows a depth of wave penetration to approximately 9 m and wave velocity equaled 0.100 m/ns. In this image, which has been topographically corrected, moderate and high reflectors indicate nearly continuous subhorizontal undulating beds of sand and gravel that follow topography.
Figure 4.7. GPR survey of delta perpendicular to flow, from east to west.
4.4 THE CHANNEL

Figure 4.8A. View of channel looking north toward Skeiðarárjökull.

Figure 4.8B. View of channel looking south toward the moraine complex.
A pair of surveys running parallel and perpendicular to water flow also were conducted through the channel that sits north of the delta (Fig. 4.8A, B). Data from the 100 MHz antennae were utilized for analysis with depth of wave penetration reaching to approximately 9 m with an average wave velocity of 0.100 m/ns. Line 27 was surveyed over 88 m from north to south and the cross section (Fig. 4.9) has been corrected for topography. The uphill longitudinal profile indicates that pressurized water was flowing under the ice. A buried set of foresets, with steep dips indicates that another delta (delta X) may have existed behind the current delta prior to the channel formation. There also is evidence of ice-block slumps and an old moraine complex at the northern part of the delta image where the channel and delta intersect, indicated by the high angle beds which dip in the opposite direction. These features indicate that the most recent proglacial environment is multi-generational and has succeeded an earlier version of the same thing except for reforming slightly farther down-current.
Figure 4.9. GPR survey of channel parallel to flow which ran north to south. Notice the uphill longitudinal profile despite the southward dipping foresets.
Line 29 (Fig. 4.10) was surveyed over 54 m perpendicular to flow from west to east. After topographic correction, reflections from the 100 MHz antennae change from dipping eastward on the west side to dipping westward on the east side, principally paralleling the modern channel outline. However, there are also one or two possible buried channels (the eastern possibly is poorly defined and may just show sediment lapping onto the main channel margin).
Figure 4.10. GPR survey of channel perpendicular to flow which ran north to south.
4.5 EASTERN AND WESTERN RIDGES

The ridges that parallel the channel (Fig. 4.11) were surveyed parallel and perpendicular to paleoflow and data from the 50 MHz antennae were utilized for analysis. Wave penetration reached to approximately 8 m with an average wave velocity of 0.100 m/ns. Topographically the ridges varied in shape and size, with the eastern ridge being considerably larger.

Figure 4.11. View of eastern and western ridges looking northwest from the moraine complex.
Line 10 on the eastern ridge was surveyed over 180 m from south to north and the processed image (Fig. 4.12) has been corrected for topography. The cross section shows reflections that are interpreted to be large ice-block slumps that range in size from approximately 10-20 m in diameter. There also is evidence of a buried channel with poor reflections at its base, which is likely due to the presence of till or clay that can inhibit reflections. This cross section also shows high-angle, downflow-dipping beds, which are likely foreset beds from the older delta X at the southern end of the transect. The northern end of the transect contains discontinuous low-angle beds that slope gently northward (upflow) into the basin that will be discussed later.
Figure 4.12. GPR survey of eastern ridge parallel to flow.
Line 11 on the eastern ridge was surveyed over 150 m from west to east, but after data collection, traces were reversed using EkkoView Deluxe so that the final image could be viewed as if the transect was surveyed from east to west (Fig. 4.13). This image has been corrected for topography. Radar waves penetrated to a depth of approximately 8 m with moderate reflections. The cross section shows predominantly subhorizontal beds that follow topography with a small buried channel. Foreset beds are indicated on the western edge of the image, likely related to the older delta.
Figure 4.13. GPR survey of eastern ridge perpendicular to flow which ran north to south.
Line 13 on the western ridge was surveyed using 50 MHz antennae over 270 m from north to south, and continued midway to the existing moraine complex. The cross section (Fig. 4.14) indicates several large ice-block slumps that are 20 to 40 m in diameter and what appears to be a buried channel. The feature labeled and interpreted as an ice-block slump at the southern end of the section, may actually be sediments that are part of the existing moraine complex.
Line 14 on the western ridge was surveyed using 100 MHz antennae over approximately 52 m from west to east through the channel. The cross section (Fig. 4.15) has been corrected for topography. Again, the cross section is highlighted by several large ice-block slumps that are up to 10 m in diameter. These slumps have created an undulating surface near the bottom of the section. The topography clearly shows the location of the channel and the western edge of the ridge that borders the channel to the east.
Figure 4.15. GPR survey of western ridge perpendicular to flow, which ran north to south.
Several transects were surveyed throughout the basin and most exhibited very similar characteristics, therefore one transect that ran from north to south and one that was surveyed from west to east were chosen to represent this area (Fig. 4.16). The 50 and 100 MHz antennae both were utilized but the 100 MHz series were selected for analysis of the north-south transect and the 50 MHz used for the west to east transect. Radar waves penetrated to approximately 5 m with the 100 MHz antennae and to 15 m with the 50 MHz with strong reflections.
Line 53 (Fig. 4.17) was surveyed directly down the center of the basin from north to south over 115 m. The transect was not corrected for topography but there was an obvious uphill slope from the northern to southern end of the transect of approximately 1 m. This transect proves to be very interesting as it appears to contain a large hydrofracture or fault at approximately 45 m from the beginning of the transect. Hydrofracture, or water escape, features have been found and studied in the area to the west of this study site. These features probably were formed during the rising stage of jökulhlaups near the ice front when rapidly increasing water pressure drove water from over pressured subglacial channels and/or aquifers up through the overlying ice and possibly through till that already had been deposited (Munro-Stasiuk et al., 2009). However, since there is evidence of large ice-block slumps throughout the area, the near vertical feature may be the result of another large ice-block slump. The arching feature at the bottom of the image between 45 and 90 m actually appears to be part of the fault/hydrofracture feature and therefore could be a combination of processes resulting from the fault/hydrofracture on its north side and slumping sediments to the south.
Figure 4.17: GPR transect of basin from north to south. Notice large feature that may be a hydrofracture or fault near center of image.
Line 02 (Fig. 4.18) was surveyed over 100m from west to east across the large basin and was topographically corrected. There appears to be an old buried channel on the eastern side of the basin that has been infilled, but there are very few features of interest in this cross section other than several subhorizontal beds of sand and gravel that onlap onto the sides of the depression and, higher up, simply drape the sides of the basin, therefore showing simple vertical aggradation and basin fill, as befitting a lake.
Figure 4.18. GPR transect of basin from west to east.
4.7 THE ESKER

The esker was located on the western side of the site, approximately midway between Skeiðarárjökull and the moraine complex (Fig. 4.19). Simple site surveys were conducted to determine elevations for topographic profiles which have been added to the data. North-south and east-west transects were surveyed. The 100 MHz antennae were utilized for both transects and cross sections exhibit strong reflectors to more than 8 m in depth.
In Line 22, which was surveyed from east to west, perpendicular to paleoflow, slumping sediments once again dominate this cross section (Fig. 4.20). However, these slumps are likely related to the lateral collapse of the feature rather than from melting ice blocks as have been common throughout the rest of the site. In spite of the considerable amount of collapse that has occurred along the edges of the esker, the core sediments are intact and are what would be expected to be seen in cross section.
Line 22: E-W Esker (100 MHz)

Figure 4.20. GPR survey of esker perpendicular to flow which ran north to south.
Interesting features found in the 145 m long north-south trending Line 20 (Fig. 4.21) may be identified as pseudoanticlinal macroforms as described by Brennand (2000). Features like this can occur along geometrically uniform, narrow reaches of esker ridges and can be explained by the operation of secondary currents or vortices of similar power within narrow reaches of a synchronous R-channel (Shreve, 1972). Alternatively, they may also be merely the result of ice-block slumps as has been seen throughout the entire site and in this image.
Figure 4.21. GPR survey of esker, parallel to flow which ran north to south.
Bordering the western side of the site was a large ridge rippled with drumlins (Fig. 4.22). A topographic survey was not completed due to the size of the ridge, the number of drumlins, and because there were only minor changes in slope as you moved over each drumlin. Two transects were completed using the 100 MHz antennae; the north-south transect covered 180 m and the west-east line covered approximately 100 m. Radar waves penetrated to depths of over 8 m with moderate to strong reflectors.

Survey Line 06 (Fig. 4.23) was oriented from north to south, basically traversing one long drumlin. Again, large-ice block slumps, the largest of which is approximately 40 m in diameter, are prevalent within the transect. The northernmost slumping may be responsible for the fault that borders it. Near the southern end of the transect, a deeply buried feature possibly could be the remnant of an old surface or perhaps a channel that has since been filled with slump sediments.
Figure 4.23. GPR survey of drumlinized ridge parallel to flow which ran north to south.
Line 05 which trends west to east over the drumlinized ridge (Fig. 4.24) once again contains an abundance of chaotic reflections indicative of large ice-block slumps that range in size from approximately 15 to 20 m in diameter. A fault at the midpoint of the transect is again most likely related to the slumping sediments on either side. At the surface this area exhibits characteristic drumlin shapes, but subsurface horizons do not conform to surface topography because of extensive disruption by slumping.
Figure 4.24. GPR survey of drumlinized ridge from west to east perpendicular to flow which ran north to south.
4.9 SUMMARY

The site at Skeiðaràrsandur Iceland has classic features associated with subglacial and proglacial environments, such as a delta, tunnel channel, esker, drumlins and a large moraine complex. Except for the moraine complex, each feature was surveyed using GPR. Prior to surveying the area, a trench was excavated to reveal the stratigraphy of the subsurface. Not surprisingly, characteristic sand and gravels commonly found in glacial environments were found to a depth of 1.5 m within the trench.

GPR investigation of the delta revealed characteristic foreset beds prograding from the mouth of the channel. Surveys of the channel also contained foreset beds, suggesting the existence of an older delta buried behind the current feature. The channel exhibits an uphill longitudinal profile, indicative of pressurized subglacial meltwater that was forced upslope. Transects conducted over the ridges bordering the channel also have angled beds at their southern extent consistent with foreset beds of a buried delta. One very interesting feature within the basin was a possible hydrofracture or water-escape feature, although it might merely be a fault caused by slumping sediments that were prevalent in the area. Water-escape features have been identified in the Súla region of Skeiðaràrsandur by other researchers, whereas large slumps (formed as either the result of melting ice blocks or pressure release from removal of the overlying ice) have been found in every survey and have made the notably subjective interpretation process even more challenging.

Reflections within the north-south trending transect of the esker may be pseudoanticlinal macroforms as described by Brennand (2000) or again, merely the result of an abundance of slumps within the section. The drumlinized ridge surveys also
revealed mostly chaotic reflections associated with large slumps, which in turn resulted in a fault.
5.1 INTRODUCTION

In July of 2011, a series of GPR transects were collected from four sites near Edinboro, Pennsylvania. All transects, except those from the Glover quarry, were run across roadways using 100 MHz antennae with a signal depth to approximately 8 m. However, many transects contain interferences that compromise the integrity of the data due to buried cables, culverts and bridges. Therefore, only surveys that indicate obvious natural features were selected for analysis. Transects in Pennsylvania were up to 4,000 m long (in Iceland they were less than 300 m). Therefore it was necessary to chop the long transects from Pennsylvania into smaller segments to make subsurface features more readily visible. Figure 5.1 shows the entire study area around Edinboro where four transects were collected. Some of the most revealing data come from the area east of Edinboro, southwest of Waterford, in the Glover quarry area near French Creek Road.
Figure 5.1. USGS topographic image of Pennsylvania study site showing location of transects.
5.2 CRANE ROAD

Line 00-Crane Road (Figs. 5.2 and 5.3) was surveyed across the glacial spillway shown in Figure 1.2A, from just east of Interstate 79 to State Route 99 over a distance of 3,253 m. Radar signals penetrated to depths of approximately 8 m and were most likely inhibited due to the presence of till and/or bedrock. Fiduciary marks were inserted during data collection to note the existence of obvious features at the surface that may have skewed the GPR signals, such as bridges, buried cables, driveways and culverts. The Crane Road transect was chopped into five segments but only two will be discussed due to a lack of any significant subsurface features. Figure 5.4 is a photograph taken at the beginning of the survey, just east of Interstate 79. In the distance the eastern ridge of the valley can be seen.
Fig. 5.2. Cross section of entire transect surveyed across Crane Road from East to West.
Figure 5.3. USGS topographic map showing exact location of Crane Road transect in relation to Edinboro.
The first Crane Road segment that was chosen for analysis begins approximately 1,400 m from the beginning of the survey and was chopped after 400 m. Two images are shown below. The first (Fig. 5.5) was corrected for topography, and the second (Fig. 5.6) was not so that features are more readily visible. Note that the vertical scale is significantly exaggerated.

The cross section shows significant interference from a bridge, but reflections also show what appears to be an abundance of large stacked boulders east of the bridge. However reflections within 2 m of the surface may be anthropogenic (buried cables for example).
Figure 5.5. Segment 1 of Crane Road with topographic correction.
Figure 5.6. Segment 1 of Crane Road without topographic correction.
The second chopped segment of Crane Road begins at 1,740 m from the beginning of the survey and continues to 2,140 m. This segment was near the midpoint of the survey where the transect goes up and then back downslope rather sharply as can be seen in the photograph taken from the top of ridge looking east (Fig. 5.7). Both cross sections (Figs. 5.8 and 5.9) appear to show an abundance of large boulders, though it is possible that they may be culverts or other manmade structures. Strong reflections to the east represent the intersection of Lay Road with Crane Road.
Figure 5.8: Segment 2 of the Crane Road transect with topographic correction.
Figure 5.9. Segment 2 of the Crane Road transect without topographic correction.
5.3 BLYSTONE AND JERICHO ROAD

Line 02, which lies to the south of Edinboro just across the Erie County line in Crawford County, begins at the intersection of Blystone Road and Irish Road and was surveyed from south to north. (Fig. 5.10) At the intersection of Blystone Road and State Route 99, the survey turned east, following Jericho Road. This transect was then continued until Jericho Road ended at Mount Pleasant Road (Fig. 5.11). The survey covered 2,942 m and the cross section of the entire transect shows that Jericho Road gradually slopes into a valley. The only significant feature in this profile is near the Blystone Road - State Route 99 intersection where a quarry exists. The Jericho Road segment has no significant features other than those that are anthropogenic.
Fig. 5.10. USGS topographic map showing location of Blystone and Jericho Roads which lie to the south of Edinboro in Crawford Ct.
Figure 5.11. Cross section of entire length of the transect surveyed from south to north and then from west to east across Blystone and Jericho Roads, which are south of Edinboro.
The most obvious reflections from the Blystone Road segment of Line 02 that was chosen for analysis were found where the road paralleled a quarry (Figs. 5.12 and 5.13). Unfortunately, the quarry was fenced and a survey was not possible. The northern part of the area adjacent to the quarry shows strong reflections that indicates the existence of boulders and layers of sand and gravel. The southern end of the segment, which was farmland, exhibits continuous, nearly horizontal undulating reflections likely to indicate shallow layers of sand and gravel in parallel beds overlying till. At the very northernmost part of the transect, reflections show the existence of materials used in constructing State Route 99.
Figure 5.12. Blystone Road segment with topographic correction.
Line 02 - Blystone Road – 0m to 440m (100 MHz)  

Intersection  
Rt. 99  

Quarry area

Figure 5.13. Blystone Road segment without topographic correction.
5.4 OLD STATE ROUTE 99

Line 08, which is located to the north of Edinboro, was surveyed over a distance of 4,493 m and was selected since it trends north-south and runs through the valley. The cross section of the entire transect shows that Old State Route 99 has an uphill long profile (Fig. 5.14), which is what would be expected with pressurized water flowing under the ice sheet and has been confirmed in images produced in RiverTools (Fig. 3.8). Figure 5.15 shows the location of the transect. There were many places where the radar signals were skewed because of anthropogenic features, but at the end of the transect the area adjacent to a known active gravel pit exhibited interesting natural features.
Figure 5.14. Cross section of entire length of Old Rt. 99 transect surveyed from north to south.
Figure 5.15. USGS topographic map showing location of Old Rt. 99 transect.
The first segment of Old Route 99 that was chosen for interpretation was chopped from 500 to 1,100 m (Figs. 5.16 and 5.17). Fiduciary marks 4, 5 and 6 were added during the survey to indicate the location of three houses, so it is possible that there is some human interference near the surface, reflected as discontinuous undulating lines which include buried fiber optic cables, gas lines, etc. Arching features reflected at more than 5 m deep are likely large boulders.
Figure 5.16: GPR survey of segment 1 of Old Route 99 with topographic correction.
Figure 5.17: GPR survey of segment 1 of Old Route 99 without topographic correction.
The second segment (Figs. 5.18 and 5.19) was chosen for analysis in spite of two bridges and an intersection creating significant interference. Fiduciary mark 14 was noted because of an exposure of till at the surface (Fig. 5.20). However, the signal penetrated through the till and picked up the reflection of boulders at approximately 3 m. After the second bridge between the 2,360 m and 2,440 m section, there are undulating, discontinuous, dipping features that could be anthropogenic since the area is bordered on both ends by a bridge and intersection. However, there is the possibility that there are boulders at depth and possibly a fault at approximately the 2,390 m mark.
Figure 5.18. GPR survey of segment 2 of Old Route 99 with topographic correction.
The third segment of Line 08 that was chosen for analysis was chopped from 2,580-3,190 m (Figs. 5.21 and 5.22) and has a series of three fiduciary marks that again represent houses along the survey. It was noted that these homes all had large rounded boulders in their yards and lining the roadway. However, there also were signs denoting buried fiber optic cables and gas lines that may actually account for some of the reflections seen in cross section.

For nearly 400 m across the section, discontinuous undulating beds with sharp peaks are obvious within 3 m of the surface. These could represent shallow beds of sand and gravel or manmade features. From the northern end of the profile an undulating continuous reflection of approximately 200 m and from a depth of 3 m to 8 m then back to 2 m can be seen (Fig. 5.23). This reflection likely represents the bottom of
a kettle lake lined with clay that has since infilled with large amounts of silt and capped with sand, gravel and possibly till.

At the end of the transect, a continuous, nearly horizontal reflection overlies a host of chaotic reflections, possibly the existence of a sand and gravel pod. This segment is located just north of an area that has identified gravel pits. In the image that hasn’t been corrected for topography, there is a vertical reflection that may be a fault. In several areas throughout the section, large boulders are visible at depths of more than 6 m, indicating that till may not be present in this segment.
Figure 5.21. GPR survey of segment 3 of Old Route 99 with topographic correction.
Figure 5.22. GPR survey of segment 3 of Old Route 99 without topographic correction.
Figure 5.23. Northern end of segment 3 of Old Route 99 zoomed in to undulating feature.
The USGS topographic map below (Fig. 5.24) shows the location of a property that is for sale at the intersection of Zessinger Road and its proximity to the Tallman family gravel pit that will be discussed in the next section. The fourth segment of Line 08 (Figs. 5.25 and 5.26) intersects with Zessinger Road, which has a real estate sign advertising frontal property with commercial gravel pit development. The chaotic reflections at the southern end of this profile are likely the sand and gravel adjacent to this property.
Figure 5.24. USGS topographic map showing location of property along Old Route 99 designated for gravel pit development and its proximity to a known gravel pit.
Figure 5.25. GPR survey of segment 4 of Old State Route 99 with topographic correction.
Figure 5.26. GPR survey of segment 4 of Old State Route 99 without topographic correction.
The last segment of Line 08 (Figs. 5.27 and 5.28) has significant reflections that are related to an active gravel pit that sits west of Old State Route 99. The area surrounding the family home between the road and the gravel pit was highlighted by a vast array of large to small erratics. The owners noted that their property also contained a large peat bog (Fig. 5.24), which they were told had “no bottom.” Time conflicts with the owners did not allow for a GPR survey of the gravel pit itself, but a tour of the area revealed piles of gravel and cobbles, as well as outcrops of till (Fig. 5.29).
Figure 5.27: GPR survey of segment 5 of Old State Route 99 with topographic correction.
Figure 5.28: GPR survey of segment 5 of Old State Route 99 without topographic correction.
Figure 5.29. Photographs of erratics, gravel and till found on the Tallman family gravel pit property.
5.5 FRENCH CREEK ROAD

Lines 03 and 04 were surveyed east of Edinboro and southwest of Waterford. This area borders French Creek (Fig. 5.30). Line 03 was surveyed from south to north, and Line 04 began where French Creek Road curves to the east. Line 04 was surveyed across a flat-floored valley (Fig. 5.31). The total distance covered between the two surveys was 3,100 m with 100 MHz antennae. Radar signals penetrated to depths of approximately 5 m. Elevations were acquired via Google Earth and data in Figures 5.32 and 5.33 were corrected for topography. Annotated images were not corrected for topography since there was very little change in elevation across the survey. I was able to talk to the owner of the quarry that was just east of French Creek Road and was granted permission to survey his entire property. These surveys account for some of the most interesting data in the area.
Figure 5.30. USGS topographic map of the area east of Edinboro and southwest of Waterford showing exact location of French Creek Road transects, including the location of the quarry.
Figure 5.31. Photograph of Line 04 looking west across the flat-floored valley. Large drumlin seen in distance.
Figure 5.32. GPR survey of French Creek Road from south to north corrected for topography.
Figure 5.33. GPR survey of French Creek Road from west to east corrected for topography.
The Line 03 transect (Fig. 5.32) was surveyed from south to north over 1,931 m, parallel to French Creek to the west and the Glover quarry to the east. While reflections throughout the transect revealed many areas that contain boulders, the section between 1,000 and 1,700 m was chopped for analysis since it was the section that bordered the quarry (Fig. 5.34). Strong reflections indicate significant amounts of sand and gravel, as well as many large boulders visible to 9 m, with very little anthropogenic interference.
Figure 5.34. GPR survey of section of French Creek Road bordering quarry without topographic correction.
Line 04 was surveyed from west to east over a distance of 1,167 m (Fig. 5.33). The western side of this survey lies directly to the north of the quarry. Reflections in this profile are shallow, only reaching to approximately 4 m, which may indicate the presence of till or bedrock. The first section chosen to analyze was chopped from 210 m to 510 m and has distinct undulating reflections that are indicative of foreset facies that dip to the west and east (Fig. 5.35). The second section (Fig. 5.36) has a wide, chaotic assemblage to the east that spans more than 110 m and is believed to be sand and gravel. Unlike Line 03, there are no visible boulder reflections across this survey.
Figure 5.35. GPR survey of section 1 of French Creek Road north of the quarry without topographic correction.
Figure 5.36: GPR survey of section 2 of French Creek Road without topographic correction.
5.6 GLOVER TRUCKING QUARRY

The quarry is approximately 500 m from west to east and 300 m from north to south (Fig. 5.37). Twelve surveys were conducted with 100 MHz antennae, attaining a signal depth from 4 to 6 m. The area in the northwest corner, shown in the red circle below, has previously been excavated, leaving this section approximately 10 m below the original surface (Fig. 5.38). The area circled in the northeast corner is a ridge that sits approximately 50 m from the surveyed surface (Fig. 5.39) and contains a variety of facies containing sand and gravel (Figs. 5.40). These facies varied in location from massive, imbricated gravels interbedded with fine grained, stratified sands to a continuous, horizontal, poorly-sorted cobble/boulder layer, indicating a variety of flow regimes within the area. A series of foreset beds that have been truncated and overlain by till, also are visible on the ridge that parallels Line 11 (Fig. 5.41).
Figure 5.37. Google Earth image of Glover quarry.
Figure 5.38. View of area containing Lines 15, 16 and 17 looking east-northeast.

Figure 5.39. View of outcropping ridge that borders Lines 05 and 06 looking southwest.
Figure 5.40. Sand and gravel assemblages found on ridge that borders Lines 05 and 06.
Figure 5.41. View of foreset beds in outcropping ridge that parallels Line 11. Beds have been truncated and overlain by till.
The Google Earth image shows the location of Lines 16 and 17 in the northwest corner of the quarry (Fig. 5.42). Line 16 was surveyed from east to west, but traces were reversed in EkkoView Deluxe so that the profile reads from west to east and is easier to compare with the Line 17 profile. Line 16 has strong, nearly horizontal reflections that continue for approximately 30 m and then dip sharply to the east, indicating a series of foreset beds (Fig. 5.43). These beds can be traced to approximately 7 m deep, even though reflections are increasingly weaker with depth. On the eastern side of Line 17 (Fig. 5.44), another series of steeply dipping beds exist and can be traced to approximately 5 m. A series of weak reflections on the western side of the transect also may be foreset beds that dip to the east.

Figure 5.42. Google Earth image showing location of Lines 16 and 17 in Glover quarry.
Figure 5.43. GPR profile of Line 16 collected at Glover quarry. Traces were reversed in EkkoView Deluxe.
LINE 17 – Quarry (100 MHz) - Not corrected for topography

Figure 5.44. GPR profile of Line 17 collected at Glover quarry.
The Google Earth image below shows the location of Lines 15 and 18 in the northwest corner of the quarry (Fig. 5.45). Strong continuous, dipping reflections in both profiles once again represent the presence of foreset beds, with those in Line 15 having a second set overlying an older section (Fig. 5.46). The foresets in Line 15 dip sharply to the southwest, while those in the western part of the Line 18 transect dip to the west-southwest.

Figure 5.45. Google Earth image showing location of Lines 15 and 18 in Glover quarry.
Figure 5.46. GPR profile of Line 15 Collected at Glover quarry.
LINE 18 – Quarry (100 MHz) – Not corrected for topography

Figure 5.47. GPR profile of Line 18 collected at Glover quarry.
The Google Earth image shows the location of Lines 10 and 13 (Fig. 5.48) that were surveyed at the southern end of the quarry from east to west. However, traces were reversed in EkkoView Deluxe so that all profiles trend in the same direction. Line 10 exhibits strong reflections to 5 m and a weaker signal can be seen to approximately 7 m (Fig. 5.49). Foreset beds are again visible within this profile, as well as reflections that indicate a channel that dips deep within the section. On the western edge of the image, a unique feature that spans approximately 30 m and exhibits sharp reflections at its top, could be representative of an area where water was forced upward and till squeezed into the void. Reflections in Line 13 (Fig. 5.50) are significantly weaker than those found in Line 10, but foresets and a channel can be identified.

Figure 5.48. Google Earth image showing location of Lines 13 and 10 in Glover quarry.
Figure 5.49 GPR profile of Line 10 collected at Glover quarry.

LINE 10 – Quarry (100 MHz) - Not corrected for topography

Channel or slump in middle of foreset beds

Water escape feature?
Figure 5.50. GPR profile of Line 13 collected at Glover quarry.
Lines 11 and 12 are located at the southwest corner of the quarry (Fig. 5.51) and exhibit strong reflections to approximately 6 m. Line 11 was surveyed from west to east, parallel to an outcrop that has visible foreset beds that dip to the east. The subsurface image has reflections that indicate foreset beds that dip to the west (Fig. 5.52). The north-south trending Line 12 has strong, nearly horizontal reflections that also mirror the outcrop bordering the section (Fig. 5.53).

![Google Earth image showing location of Lines 11 and 12 in Glover quarry.](image-url)
LINE 11 – Quarry (100 MHz) - Not corrected for topography

Figure 5.52. GPR profile of Line 11 collected at Glover quarry. Photograph shows outcrop containing foreset beds.
LINE 12 – Quarry (100 MHZ) - Not corrected for topography

Figure 5.53: GPR profile of Line 12 collected at Glover quarry. Photograph of outcrop which runs parallel with Line 12 shows continuous horizontal bedding consistent with GPR survey.
Line 09 was surveyed from north to south on the eastern side of the quarry (Fig. 5.54). Strong reflections are seen at the northern and southern ends of the section and reveal undulating, continuous beds that dip to the south (Fig. 5.55). Reflections in the center of the section are very weak and could indicate a former channel that is underlain by till or bedrock.

Figure 5.54. Google Earth image showing location of Line 09 in Glover quarry.
Figure 5.55: GPR profile of Line 09 collected at Glover quarry.
Lines 05 and 06 were surveyed in the northeastern section of the quarry and parallel the large outcropping ridge (Fig. 5.56), which based on field observations, is likely to be the remnant of a kame. Line 05 exhibits shallow, undulating, discontinuous reflections that reach only to approximately 3 m (Fig. 5.57). Line 06 (Fig. 5.58) also contains shallow, undulating, discontinuous reflections but exhibits reflections that indicate the presence of foreset beds and both profiles exhibit features that are consistent with sand and gravel assemblages seen within the outcrop (Fig. 5.40). These assemblages can be compared to material exposed in a trench at Skeiðararsandur, Iceland (Fig.4.4).

Figure 5.56. Google Earth image showing location of Lines 05 and 06 in Glover quarry.
Figure 5.57. GPR profile of Line 05 collected at Glover quarry.
Figure 5.58. GPR profile of Line 06 collected at Glover quarry.
5.7 SUMMARY

The Edinboro area was chosen as the study site because of its location within one of the large valleys that dissect northwestern Pennsylvania and the area is known to have extensive deposits of sand and gravel. Since the majority of data were collected along roadways, there was significant interference in many of the sections due to buried cables, culverts, gas lines, bridges, etc. These interferences made data analysis difficult in many instances.

The complete profile of Line 00 - Crane Road shows the general shape of the valley, but other than large boulders, this section provided little information about the subsurface. Most reflections were deemed to be anthropogenic, which was to be expected. Till was observed to outcrop in several places along the transect and may have inhibited reflections in some areas.

Line 02 - Blystone Road was surveyed along an area that has an active quarry, and reflections indicated the existence of sand, gravel and boulders. Reflections south of the quarry showed nearly horizontal, continuous reflections that, based on the proximity to the quarry, are likely to be parallel layers of sand and gravel. GPR signals in the southern end of this transect were shallow possibly due to the existence of till.

Data from Line 08, the north-south trending Old State Route 99, illustrate an uphill longitudinal profile and show several areas that are underlain by pockets of sand and gravel. Many large sub-angular to sub-rounded erratics were observed at the surface along the entire length of the transect and also are reflected in cross section. Several areas of till also were observed along the roadway. An active gravel pit and property designated for commercial gravel pit development are present at the southern end of the transect. At the Tallman family home and gravel pit area, sand, gravel and till
were abundant as were boulders in a range of sizes. These results suggest that sediment in this area is outwash carried by meltwater flowing in subglacial channels or fissures within the ice and deposited at the front of the ice. It also is possible that sediment was deposited in kames that may have disrupted water flow. Personal communication from the Tallman family noted that a large peat bog also exists on their property, which also can be seen on Google Earth. If drainage was disrupted in this area, a lake may have formed and eventually have been filled with silt, clay and peat.

Data from Line 03, Line 04 and those collected from the Glover quarry southwest of Waterford, have reflections that suggest deposition of sediment in a kame that disrupted water flow and formed a lake that led to the creation of a delta. Surveys from French Creek Road that border the quarry have reflections that indicate pockets of sand and gravel, as well as a significant number of boulders. In addition, Line 04, which was surveyed across a flat-floored valley, also has dipping reflections that suggest the presence of foreset beds. This may represent the surface of lake deposits or glacial outwash deposits laid down by streams at the ice front. Within the quarry, several surveys also exhibit foreset beds, consistent with features that would be found as sediments were deposited in a lake. The image below (Fig. 5.59) shows the general direction of the dipping foreset beds within transects in the quarry.

The outcropping ridge at the northeast corner of the quarry has sand and gravel assemblages consistent with what would be found in kames Bedded and sorted sand and gravel (Fig. 40B, C) indicate rapid changes in flow velocity and sands clasts (Fig. 40D) are beds of saturated sands that were frozen before being transported. Figure 5.40E contains a fold and 5.40F illustrates a fault that formed as the kame likely slumped after deposition.
Figure 5.59. Composite image showing general direction of dipping foreset beds within Glover quarry.
CHAPTER 6: DISCUSSION

6.1 INTRODUCTION

The glacial landscape and sedimentological characteristics associated with northwestern Pennsylvania have been poorly studied in comparison to the glacial history of Iceland. There also is a considerable difference in scale as the Chautauqua Drumlín Field in Pennsylvania spans an area almost twice the size of Skeiðarársandur in Iceland and the landforms themselves are more than an order of magnitude larger in Pennsylvania. However, many of the same features have been identified in both regions: drumlins, channels/valleys, eskers, all of which have been considered part of a continuum involving subglacial meltwater (Munro-Stasiuk et al., 2009). The rationale for this study was to compare glacially-derived landforms in Iceland with those found in northwestern Pennsylvania based solely on form analogy and sedimentology, meaning that the glacial processes we can observe today are directly related to the glacial processes active in the past. Therefore if we compare the ancient processes with the modern, documented processes, we should see that they are analogous in spite of differences in scale.

Several researchers, including Shaw (1983, 1996), Shaw and Kvill (1984), Shaw et al. (1989) and Rains et al. (1993) have concluded that subglacial megafloods were responsible for carving drumlins, flutings, meltwater channels and a variety of other associated landforms beneath the southwest margin of the Laurentide Ice Sheet. The distribution of drumlins on interfluves between tunnel valleys suggests that both drumlins and tunnel valleys are products of erosion by the same agent (Munro-Stasiuk et al., 2009). Tunnel valleys have been widely regarded as being carved by meltwater and it also has been argued that drumlins have a meltwater origin. This interpretation has
involved sedimentological analysis and comparisons of subglacial landforms with erosional forms produced both by experiments in wind tunnels and flumes and in the boundary layers of contemporary fluvial and aeolian environments (Evans et al., 1999). Controversy surrounds this topic and includes questions regarding the volume of water necessary to create such large suites of landforms as well as to the location of where that water could have been stored.

As mentioned, both Iceland and Pennsylvania exhibit landforms associated with formation by subglacial meltwater near the glacial margin (Fig. 6.1). However, Iceland has modern features, many of which were deposited or sculpted during outburst floods in the last two decades, while the features in Pennsylvania are related to the decay of the Laurentide Ice Sheet. Were processes during that decay time-transgressive or catastrophic? The subglacial meltwater hypotheses proposed for tunnel valley formation has two variants: (1) the steady state, progressive formation (e.g. Boulton and Hindmarsh, 1987; Mooers, 1989; Smed, 1998; Huuse and Lykke-Andersen, 2000; Praeg, 2003), and (2) the sudden formation by catastrophic outbursts (jökulhlaups) of subglacially accumulated meltwater (e.g. Wright, 1973; Ehlers and Linke, 1989; Brennand and Shaw, 1994; Patterson, 1994; Piotrowski, 1994; Björnsson, 1996; Clayton et al., 1999; Beaney, 2002). Steady state scenarios rely on slow processes (albeit large enough to move large boulders) operating over several hundred to several thousand years (Mooers, 1989), whereas jökulhlaup theories consider time frames on the order of days to weeks, perhaps with multiple well-spaced events, where sudden releases of subglacially-stored meltwater simultaneously occupy entire channel systems by bank-full discharges (Brennand and Shaw, 1994; Beaney and Shaw, 2000; Beaney, 2002),
thereby accounting for the large amounts of sediment and clasts larger than 2 m in diameter found in some outwash fans (Piotrowski, 1994; Cutler et al., 2002).

Figure 6.1. (A) Aerial photograph of Skeidararsandur Iceland showing relationship of study site (shown in square) to glacial margin; (B) DEM of northwestern Pennsylvania with dotted line indicating glacial margin and its proximity to the area used in this study.
In Pennsylvania, the ice advanced several times and resulted in materials being picked up both locally and from places to the northeast. The greatest percentage of the material was laid down directly from the ice in the form of ground or end moraines composed of till, with the remainder of the deposition in the form of outwash from meltwater flowing under the ice, through fissures in the ice or from the ice margin (Shepps et al., 1959). These outwash deposits are variously known as kames, kame terraces, kame moraines, eskers, or valley trains and consist of well- to poorly-bedded sands, silts, clays, and gravels (Tomikel and Shepps, 1967). In Iceland, composition of sandur sediments range from a variety of non-flood facies to massive, coarsening-upward sequences of clast-supported pebbles and cobbles that have been interpreted as being deposited by a series of jökulhlaups (Maizels, 1991, 1993).

6.2 SIMILARITIES

Both study sites have surficial features that are posited to have been formed under similar circumstances. In this study, GPR was utilized in an attempt to determine the subsurface architecture related to processes in glacial environments that would create these features and to determine if there was any commonality. The DEM of the Chautauqua Drumlin Field in Pennsylvania (Fig. 1.2) highlights an anabranching network of large tunnel valleys that dissect the Chautauqua Drumlin Field that spans into upstate New York. These valleys have uphill longitudinal profiles, indicative of pressurized subglacial meltwater. At the Iceland study site, a satellite image highlights a drumlinized surface (Fig. 1.10) and researchers have identified a modern tunnel channel that was formed after the 1996 outburst flood on eastern margin of Skeiðarársandur (Russell et al., 2007). Another tunnel channel with an uphill longitudinal profile also was identified in
this study on the western part Skeiðarðarásandur. The existence of moraines, deltas, eskers and similar sedimentological assemblages also contribute to the use of form analogy here.

Glover quarry, east of Edinboro and southwest of Waterford in Pennsylvania, was selected for study because of its location within a large valley (Fig. 1.2). It is obvious that the quarry, which at present is approximately 500 m by 300 m, is representative of only a small area within the cross section of this valley, as topographic maps (Fig. 6.2) confirm several gravel pits within the area neighboring the Glover quarry. These pits are situated in an obvious line that trends toward the northeast.

Figure 6.2. USGS topographic maps highlighting location of Glover quarry in relation to other gravel pits in the vicinity. These quarries may or may not be active based on the date of the topographic map.
Two outcrops at different elevations within the Glover quarry provided the opportunity to observe the sedimentological assemblages of sand and gravel (Fig. 5.40) and to compare those assemblages to sediments found in a trench dug at Skeiðararsandur (Figs. 4.3 and 4.4). Both areas exhibit coarsening upward sequences, bedded sands, and poorly-sorted, unconsolidated sands and gravels. Boulders also were present in both areas, as was till, although more till was visible throughout Pennsylvania than was found in Iceland.

GPR data in Pennsylvania was compromised by anthropogenic features related to roadways, bridges, culverts and buried cables. The Iceland study site was not affected by human interference, but analysis of many cross sections proved difficult due to an abundance of disrupted sediments related to melting ice blocks or slumping. The most obvious similarities between the two sites were features identified through GPR that are interpreted to be foreset beds. Upon arrival at the Iceland site, a surficial feature that sloped in a radial pattern away from the mouth of a channel on the southern margin was presumed to be an outwash fan or delta. Data analysis on this feature confirmed that it was a delta, as prograding foreset beds dipping southward were unmistakable in cross section (Fig. 4.6). Other transects within the Iceland survey also exhibited foreset bedding, including the channel that sits directly north of the delta (Fig. 4.9). Foreset beds also were evident in cross section at Glover quarry but were found to dip in different directions throughout the entire site. Foreset bedding also was visible at the surface on the eastern edge of the lower of the two outcrops at the quarry (Fig. 5.52), which confirmed the features identified in GPR imagery. Figure 5.59 provides an overview of the directionality of the foresets observed in the quarry, either via GPR cross sections or surficial exposures.
Based on data collected in Pennsylvania, it is surmised that water flowed from the north-northwest through the large valley and sediment was deposited as outwash in and around the study area. This information correlates with the DEM image of the area. In the quarry, foresets are present in sections that sit approximately 60 m lower than the top of the ridge and are representative of water in a subglacial channel carrying sediments into a proglacial lake prior to their deposition. These foresets are posited to be part of a large kame delta that may encompass several kilometers. Kames also have been identified on Skeiðarârsandur, although to the east of the site in this study (Fig. 6.3) and are similar in shape to the feature found at Glover quarry (Fig. 6.4).
Figure 6.3. Photograph of a kame on Skeiðarársandur taken in 2007. This feature was found east of this study site. Notice the proximity to the glacial margin.

Figure 6.4. Photograph of feature at Glover quarry identified as part of a kame based on data collected during this study.
The difference in scale between the two areas and the location of other gravel pits in the vicinity, as well as several kettle lakes that have been identified in the area (Fig. 6.5), leads to the conclusion that the features found in the Pennsylvania quarry are a subset of a much larger suite of kames and deltas, whereas the feature surveyed in Iceland was likely the entire delta.
While the difference in scale really did not inhibit the identification and comparison of the delta, the considerable difference in magnitude made comparing the channel found in Iceland with the valley surveyed in Pennsylvania difficult. The north-south transect, Line 08, that was surveyed in Pennsylvania was nearly 5,000 m long, whereas the channel in Iceland was only 90 m long. The only similarity between the two features was the general uphill longitudinal profile. However, the channel in Iceland opened into the delta and at the end of the Pennsylvania Line 08 survey, there was an area with considerable amounts of sand and gravel in the Tallman family gravel pit and the land adjacent that was advertised for sale. There also were many large boulders visible at the surface and in GPR images of the subsurface. These sediments may have been carried and deposited by meltwater within fissures or be related to a delta or fan. Surveys were not conducted in those gravel pits to confirm or deny either scenario. And while boulders were seen in subsurface images at Skeiðarásandur, few were seen at the surface and those that were visible were composed of till.

While an esker and a drumlinized ridge were surveyed in Iceland, neither was surveyed in Pennsylvania, although they are visible on DEMs and topographic maps. And although these features were easily identifiable at the surface in Iceland due to their characteristic shapes, the subsurface investigation failed to support the observations due to a large number of slumps, related to either melting ice blocks or the release of overburden pressure as the ice retreated.

6.3 DIFFERENCES

Except for the existence of ice block slumps, many of the surveys conducted in Iceland revealed nearly continuous, horizontal to subhorizontal reflections indicative of
bedding planes constructed of sand and gravel. Due to the interference of asphalt, and other anthropogenic features in Pennsylvania, it was difficult to determine if it was sand and/or gravel that were being reflected or if it was something anthropogenic. Foreset beds were discovered in the Iceland channel, but none along the transect of the valley in Pennsylvania. Along the roadway transects in Pennsylvania, some of the sand and gravel assemblages found in the area were dispersed in pockets that resulted in chaotic reflections as opposed to the reflections at Glover quarry and Skeiðarársandur where profiles exhibited many strong, horizontal to subhorizontal reflections. The existence of till and bedrock in Pennsylvania also was more prevalent than in Iceland, as were large boulders. Taking scale into consideration, the pockets of sand and gravel in Pennsylvania are possible of the same magnitude as the Iceland site.

6.4 SUMMARY

This study proposed the use of form analogy to determine the mechanisms responsible for creating the glacial landforms that exist in Iceland, a modern environment, and those found in northwestern Pennsylvania, an ancient environment related to the end of the LGM. Tunnel valleys and drumlins are the most readily identifiable features in northwestern Pennsylvania and their formation remains enigmatic. And although recent research (Russel et al., 2007) has revealed a modern tunnel valley in Iceland, the scale of tunnel valleys in Pennsylvania compared with the one analyzed in this study in Iceland, make the use of form analogy difficult in this scenario. Surveys through a valley in Pennsylvania measured nearly 5,000 m long and was only a small fraction of the entire valley length. In comparison, the channel in Iceland measured just 90 m. The difficulty related to data collection and interpretation
was enhanced by the abundance of human-induced interferences within the Pennsylvania surveys.

This study set out to determine the mechanisms responsible for the formation of the network of tunnel valleys in northwestern Pennsylvania and to determine if form analogy was a viable method for this determination. This study proved that the application of form analogy is useful in some instances. Scale and manmade interferences made it difficult to compare valleys in Pennsylvania to the one targeted by this study in Iceland, inhibiting the use form analogy in this case. However, as the study site at Skeiðarásandur and the area in and around Glover quarry were nearly the same size, form analogy in this instance proved to be successful and valuable in the interpretation of the features studied.

At the site in Iceland selected for this study, there was no evidence that would indicate that features in this area were created by a catastrophic release of meltwater, unlike features that were found after the 1996 outburst flood on the eastern side of the sandur, such as giant clasts, ice blocks, massive aggradation deposits, eroded scarps and turbidite sequences (Smith et al., 2006). While GPR profiles reflect boulders at depth in this study, the outwash fan is dominated by bedded sand and gravel and no boulders were found at the surface of the fan. Based on the existence of a delta, foreset beds within the channel, the dominance of nearly continuous, subhorizontal bedded sand and gravel, the esker and numerous ice block slumps, this area is likely kame and kettle topography (Fig. 6.6). The area directly in front of the moraine complex also offers characteristic kettle and kame features although it was not surveyed (Fig. 6.7).
Figure 6.6. Photograph of study site at Skeiðarârsandur highlighting kame and kettle topography.

Figure 6.7. Photograph showing kettle and kame topography on the northern side of the moraine complex at Skeiðarârsandur, Iceland. Delta and channel are visible in the foreground.
CHAPTER 7: CONCLUSIONS

7.1 CONCLUDING REMARKS

The purpose of this study was to apply form and sediment analogy to features on Skeiðarârsandur, Iceland, to those in northwestern Pennsylvania, in an effort to determine the processes involved in their formation. Landforms found on the modern sandur in Iceland are known to be the result of meltwater processes; these include a small uphill trending channel, a delta, drumlins, an esker and several kames. Features in Pennsylvania are similar, although on a much larger scale and include tunnel valleys, drumlins, outwash plains, eskers and kames. Sediment assemblages at both study sites also were very similar. Therefore, the question that this study addressed is related to the feasibility of using form analogy to determine if landforms in Pennsylvania were the result of steady-state meltwater conditions or from catastrophic outburst events.

Preliminary information regarding the physical characteristics of the tunnel valley network in Pennsylvania indicated that it was likely carved by a catastrophic release of meltwater that may have been stored in the Lake Erie basin. However, this study could not confirm nor deny either scenario for its formation. Scale was definitely problematic, limiting the usefulness of form analogy on the valleys.

While Skeiðarârsandur is known to have undergone significant modification as a result of many documented jökulhlaups, this study site was west of the area subjected to the 1996 flood. Data analysis from this section of Skeiðarârsandur suggests that significant portions of the landscape were created and/or modified by glacial meltwater, however, not necessarily from outburst flooding. Data show areas that exhibit foreset beds indicative of deposition in a deltaic environment at the ice margin, likely at the
bottom of kames. A trench excavated on the sandur exposed graded beds of sand and gravel as would be expected in this environment and GPR reflections confirm continuous and discontinuous, horizontal to subhorizontal beds throughout the area. It is widely accepted that eskers and drumlins are part of a suite of landforms created by subglacial meltwater. Their presence on the sandur confirms a meltwater genesis for the landforms found here.

Sedimentary evidence found at Glover quarry in Pennsylvania exhibits similar characteristics to that found on Skeiðararsandur. Sediment assemblages included graded beds of sand and gravel that range from very fine sands to boulders. Outcrops on the property also contained sand clasts which are remnants of transported frozen blocks that were thawed within other bedding. There also were small faults and folds, likely the result of the sediments within the kames slumping after deposition. Foreset bedding was prominent in several subsurface cross sections and also was visible at the surface in a quarry wall. In addition to foresets, GPR surveys also identified nearly continuous subhorizontal graded beds of sands and gravel that also suggests deposition in an outwash fan or delta near the ice margin, likely related the lower part of the kame. Topographic maps identify a series of gravel pits that exist on a southwest-northeast trending line that includes Glover quarry and most likely indicates a much-larger feature that aligns with the former ice margin.

7.2 LIMITATIONS OF THIS STUDY

While the initial investigation into the formation of the valleys indicated a catastrophic meltwater event, the expanse of the network made data collection in this study difficult. The inability to adequately survey these valleys with GPR leaves
considerable gaps in this investigation. In addition to the size of the study area, the influence of urbanization also affected data collection. There are several cities within the valleys and they also contain many of the major roadways, making GPR surveys not only difficult but also dangerous. Data that were collected on roadways were significantly influenced by features such as bridges, culverts, buried cables, gas lines, etc. Attempts to contact a few property owners for permission to survey their properties were unsuccessful.

7.3 SUGGESTIONS FOR FUTURE RESEARCH

The site on Skeiðarârsandur selected for this study was located west of Sæluhúsakvísl where the tunnel valley identified by Russell et al. (2007) was carved by the 1996 jökulhlaup. While data related to the Sæluhúsakvísl site are available, the inability to collect adequate data in Pennsylvania made comparison difficult except for the uphill trending longitudinal profiles. Since the size of the Chautauqua tunnel valley network is responsible for the holes in this study, future research should include other methods of collecting viable GPR data within the valleys. It would also prove beneficial to collect data in areas other than roadways to eliminate some, if not all, of the anthropogenic interferences.

While Edinboro was selected because of its location within a valley in Pennsylvania, it was located near the midpoint of the valley and information for this study could have been improved if other locations closer to the ice margin (Kent moraine) would have been surveyed. If the valleys were carved by a catastrophic release of meltwater, an outwash fan dominated by large, rounded boulders that have been found in other studies, would have aided in the validity of this hypothesis. Boulders
were identified by GPR reflections throughout the study area, and some were seen along roadways, however only a few were seen at the surface in the quarry. This does not mean that a significant amount of boulders were not present in the quarry, only that they may have been removed for sale. While the sedimentary structure of the quarry has led to the identification of kame topography, a GPR survey south of the site may provide information relating to the body of water that the outwash fan emptied into. It also would be beneficial to survey the line of quarries that exist to the east and west of Glover quarry to confirm that they also are kames and if they are part of a kame delta or delta terrace.

Expansion of GPR surveys on Skeiðaràrsandur to include the area directly north of the moraine complex, in front of the delta, could have provided more detailed information regarding what has been posited as kame and kettle topography here. The area was initially thought to be part of the moraine complex, but is now believed to be composed of kames and kettle holes. In addition to the GPR surveys within the study site, it would have proved useful to have examined the subsurface in more detail by excavating trenches on the esker and drumlin ridge to compare the results of the GPR findings. Two outcrops in the Glover quarry allowed for comparison of subsurface images, leading to the conclusion that the quarry was a kame. In Iceland, GPR profiles can be compared only to the surficial expression of features.
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