OPG's DEEP GEOLOGIC

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FOR LOW & INTERMEDIATE LEVEL WASTE

Glacial Erosion Assessment

March 2011

Prepared by: B. Hallet

NWMO DGR-TR-2011-18

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AECOM Canada Ltd.			
Prepared by:	B. Hallet (University of Washington)		
Reviewed by:	W.R. Peltier (University of Toronto), R. Frizzell		
Approved by:	R.E.J. Leech		
Nuclear Waste Management Organization			
Reviewed by:	M. Jensen		
Accepted by:	M. Jensen		

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EXECUTIVE SUMMARY

This report examines the phenomenon of glacial erosion as it is a consideration in the safety and performance of Ontario Power Generation's proposed Deep Geological Repository for low and intermediate level radioactive waste beneath the Bruce nuclear site located in the Municipality of Kincardine, Ontario. The report reviews currently available knowledge, insights and data that provide a basis for assessing the maximum amount of glacial erosion likely to occur during the next glacial cycle at the Bruce nuclear site. As in the Long-Term Climate Change Study of Peltier (2011), the prudent assumption will be made that the region will be subjected to major glacial cycles, recurring every ~100,000 years, much as they have over the last ~900,000 years. Accordingly, the geologic data and reconstructions of the Laurentide ice-sheet (LIS) during the last major glaciation are used to assess the amount of erosion expected over the next glacial cycle.

The report looks at several independent types of geological evidence in order to assess the magnitude of total erosion, which would likely occur over one glacial cycle. These include, historical and recent regional estimates of Quaternary erosion associated with the LIS, empirical studies of glacial erosion on bedrock and sediment substrates in diverse settings and at different scales, extreme cases of erosion by ice and by catastrophic glacial outburst floods, theoretical considerations of glacial erosion and their application in a model of erosion by the LIS, evidence for erosion by subglacial meltwater and for the occurrence of a sediment cover over the bedrock, and relevant results from the University of Toronto Glacial Systems Model (UofT GSM) of Peltier (2011).

Deep excavation similar to that experienced during formation of the Great Lakes is unlikely to occur at the Bruce nuclear site. Future glaciations would tend to repeat the actions of their predecessors, by funneling into the lake basins and deepening them until the glacio-hydraulic conditions start to trap the sediments that shield the bedrock from further erosion. On the other hand, topographic highs tend to be maintained as they deflect ice, making them sites of relatively slow ice flow, low erosion rates, and sediment accumulation. The thick mantle of glacial outwash and till that was exposed by the retreating ice at the Bruce nuclear site show this to be a site of net deposition, like much of the terrain except for the Great Lakes themselves; it may have been eroded only slightly or not at all by the LIS.

Rather than relying heavily on mechanistic models of glacial erosion, the estimates of erosion for the Bruce nuclear site rely primarily on the UofT GSM (Peltier 2011) in combination with empirical results from studies of glacial erosion rates on a basin scale, maximum known amounts of erosion in various areas, and observations of glacial processes in the region of the Bruce nuclear site. The estimates are also guided by insights into glacial erosion processes and basal processes. These lines of evidence suggest that bedrock erosion on the time scale of one glacial cycle is likely to range between a few metres and a few tens of metres.

The state of understanding of erosion and other processes occurring at the base of ice-sheets is far from complete, and the conditions that control erosion are likely to vary with time and space in complex ways. The data and model results summarized in this report collectively point to a broad range of values for erosion at the Bruce nuclear site on a one million year (1 Myr) time scale. They range from ~300 m, the largest, most conservative amount to a few metres, and perhaps no erosion and net deposition. In view of the absence of topographic features or other known factors that would tend to localize erosion by ice or water over the Bruce nuclear site, and the absence of evidence of preferential past erosion over the site, a more realistic but still quite conservative site-specific estimate is 100 m for 1 Myr.

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1. GEOLOGICAL STUDIES IN THE LAURENTIDE ICE-SHEET AREA

This report examines the phenomena of glacial erosion as it may influence the safety and performance of Ontario Power Generation's proposed Deep Geological Repository for low and intermediate level radioactive waste beneath the Bruce nuclear site located in the Municipality of Kincardine. The report reviews currently available knowledge, insights and data that provide a basis for assessing the maximum amount of glacial erosion likely to occur during the next glacial cycle at the Bruce nuclear site. As in the Long-Term Climate Change Study of Peltier (2011) the prudent assumption will be made that the region will be subjected to major glacial cycles, recurring every ~100,000 years, much as they have over the last ~900,000 years. Accordingly, the geologic data and reconstructions of the Laurentide ice-sheet (LIS) during the last major glaciation are used to assess the amount of erosion expected over the next glacial cycle.

The amount of erosion by the former LIS has received considerable attention from the geological and glaciological community, as will be reviewed below, but the subject has proven challenging because subglacial erosion is inherently difficult to study for a number of reasons: 1) where the beds of former ice-sheets are exposed, past events are effaced with time as the surface erodes: a problem common to all eroding landscapes; 2) where these processes are active at present, that is, at the bottom of ice-sheets, they are difficult to observe; and 3) glacial erosion varies in time and space and depends on conditions at the bed, but bed conditions of former ice-sheets can only be inferred from sparse, residual geologic information.

Investigating this subject requires the use of all available information about the LIS that pertains to erosion, as well as studies of glacial erosion in other regions combined with theoretical considerations. Chapter 1 summarizes geological studies that use both traditional and new techniques in the area formerly occupied by the LIS. Chapter 2 discusses results of empirical studies of glacial erosion on bedrock and sediment substrates in diverse settings; it includes extreme cases of erosion by ice and by catastrophic glacial outburst floods. Theoretical considerations of glacial erosion and their application in a model of erosion by the LIS are reported in Chapter 3. Turning to the Bruce nuclear site region, Chapter 4 reviews considerable evidence for erosion by subglacial meltwater and for the occurrence of a sediment cover over the bedrock, which impacts bedrock erosion. Chapter 5 presents relevant results from the University of Toronto Glacial Systems Model (UofT GSM) of Peltier (2011) and uses them to compute likely ranges of total erosion for the next glacial cycle at the Bruce nuclear site. This chapter ends with a short discussion of the limitations of numerical models of erosion. Chapter 6 addresses other site specific transient phenomena associated with glacial erosion, and Chapter 7 summarizes the report findings. Figure 7.1 in Chapter 7 presents a compilation of estimated values of erosion or erosion rates as cited throughout the report.

1.1 Regional Estimates of the Amount of Erosion During the Quaternary

Geologists of the 19th century recognized that massive ice sheets had covered the northern continents in the relatively recent geologic past. They and their successors examined the streamlined and polished landscapes of these glaciated regions, and came to the intuitively reasonable conclusion that extensive glacial scouring had formed these landscapes. More recently, a number of studies have produced estimates of the amount of erosion by the LIS and the Fennoscandian ice sheet, with widely ranging results (e.g., White 1972, Gravenor 1975, Sugden 1976, Sugden 1978, Lidmar-Bergstrom 1997). They are briefly summarized in the next section. These studies provide estimates of total erosion generally averaged over the entire region formerly covered by the LIS for the duration of the Quaternary, roughly the last 2 million years (Myr) of earth history; the precise date marking the onset of Quaternary has been under discussion and was revised in June 2009 officially to 2.588 Myr. These studies do not, however, illuminate the spatial variability of the erosion and the amount of erosion that could have occurred locally. Moreover, they usually do not resolve the amount of erosion that occurred during the last 120,000 years of the last glaciation, often referred to as the Last Glacial Maximum (LGM), as opposed to the whole Quaternary period.

1.1.1 Geomorphic Studies of Relict Surfaces in Formerly Glaciated Area

Numerous examples of surviving pre-glacial regolith (weathered bedrock) in glaciated regions suggest that Plio-Pleistocene ice sheets may have done no more than remove a pre-existing blanket of deeply weathered bedrock, which formed in a temperate Tertiary climate acting on stable, low-relief cratonic surfaces over many millions of years, and lightly buff the surface of the underlying unweathered bedrock (Feininger 1971, Lidmar-Bergstrom 1997, Patterson and Boerboom 1999). This suggestion is also consistent with the results of studies of sub-aerial stream patterns, and other vestiges of the preglacial landscape (e.g., Sugden 1976). It is inherently difficult to quantify the amount of erosion using this approach, but most workers suggest that erosion through the Quaternary did not exceed 10 to 40 m (Figure 7.1) for both the LIS and the Fennoscandian ice sheets (Sugden 1976, Lidmar-Bergstrom 1997). An example of a local estimate of total erosion restricted to the LGM, of 6 to 20 m, was derived from a study of the Dubawnt dispersal train, glacial erratics of a distinctive rock type that are down glacier of their bedrock source, in western Nunavut (Donaldson 1965, Kaszycki and Shilts 1980).

1.1.2 Estimates Based on Studies of Sediment Volumes

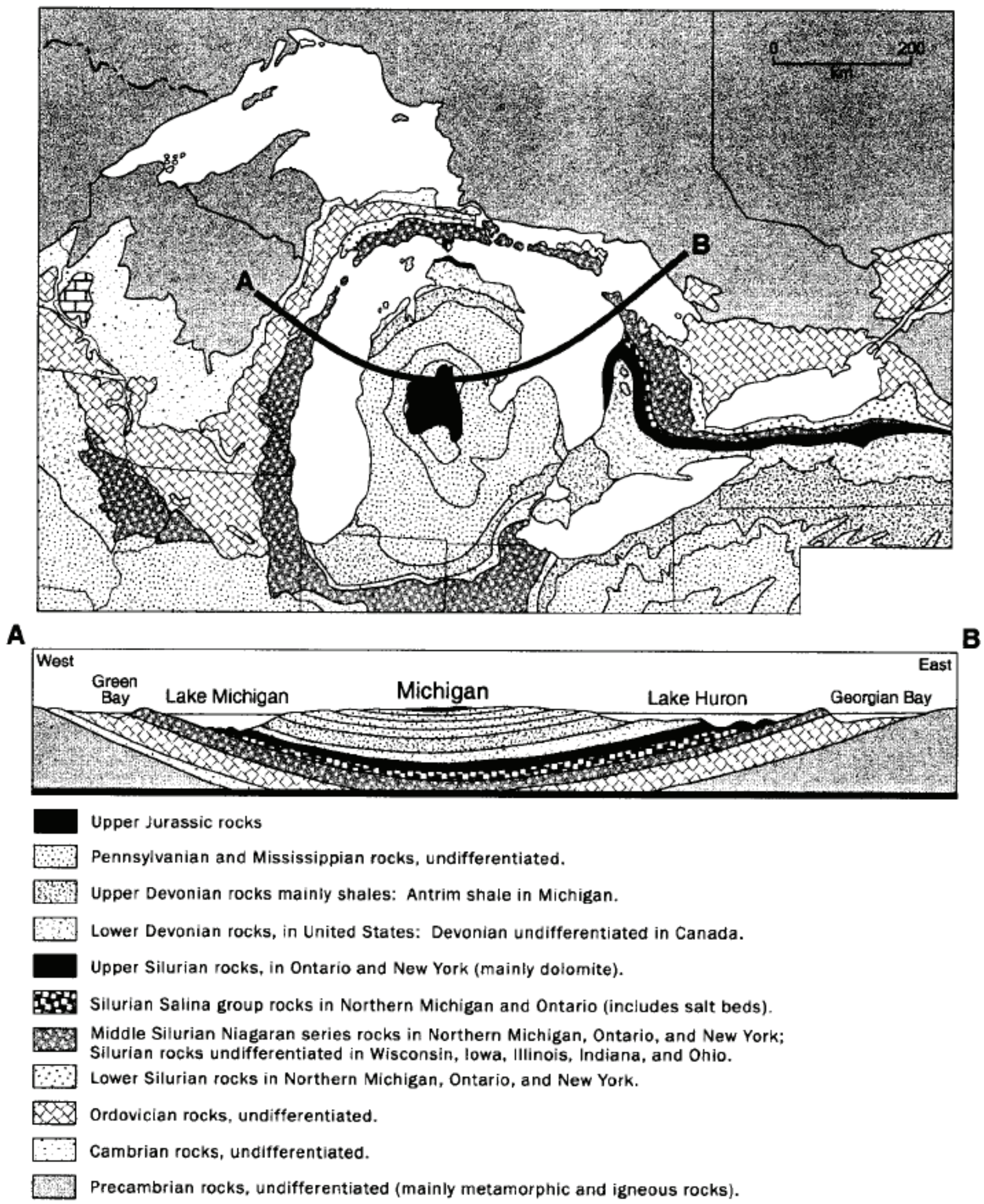
An alternate method for determining the cumulative erosion is to quantify the volume of sediment produced by an ice-sheet. Flint (1947) made one of the first efforts to map and determine the volume of all the terrestrial glacial sediment in North America, and concluded that all of the Plio-Pleistocene advances of the LIS had accomplished only a few tens of feet of erosion of the Canadian Shield. White (1972) pointed out that this ignored the much larger quantity of sediment deposited in the oceans, and revised the estimate upward by about an order of magnitude. Subsequently, Laine (1980, 1982) used North Atlantic deposits and Bell and Laine (1985) used all the marine sediment repositories of the LIS (excluding the Cordilleran Ice Sheet) to arrive at an average erosion of 120 m over 3 Myr (Figure 7.1). Bell and Laine (1985) consider this to be a minimum value although they make no allowance for non-glacial erosion. Hay et al. (1989), contending that in the Gulf of Mexico the depth of sediment of Laurentide provenance is greatly overestimated by Bell and Laine (1985), reduced this estimate of regional erosion to 80 m over the same time period (Figure 7.1).

A more recent estimate of long-term erosion rates under ice-sheets is available from Svalbard. Based on seismically mapped glacial sediments of the Isfjorden Fan, Elverhøi et al. (1995) estimate the regional erosion rate to average 0.2 millimetres per year (mm/yr) over the last glacial cycle, consistent with estimates from other Barents Sea basin sediments (Elverhøi et al. 1998). Over the LGM this amounts to a total erosion of 18 to 30 m (Figure 7.1), if the uncertainty in the erosion rate is estimated to be 25%.

1.2 Great Lakes: Spatial Patterns of Erosion and Deposition

According to Larson and Schaetzl (2001), the basins of all the Great Lakes “owe their origin mainly to channeling of ice flow along major bedrock valley systems that existed prior to glaciation, and to increased glacial scouring and erosion in areas of relatively weak bedrock.” This is particularly evident from (Figure 1.1), which shows parts of the Huron, Erie, and

Michigan basins conforming to the outcrop pattern of Devonian and Upper Silurian rocks that are, in large part, erodible shales and carbonates.



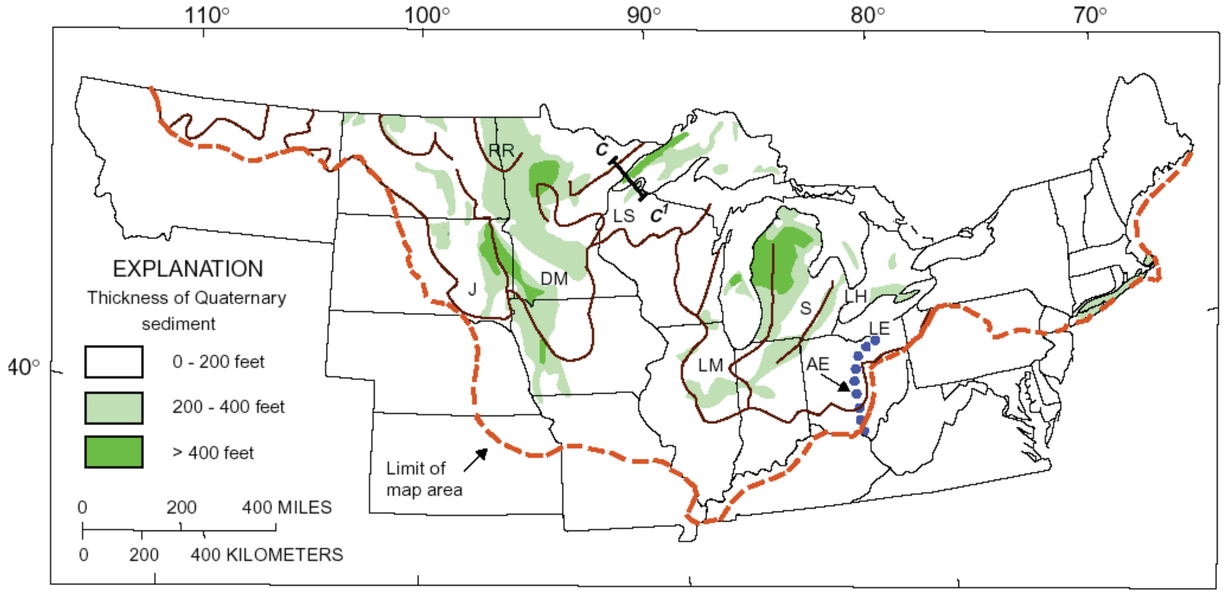
Notes: Different stippling and shading patterns are shown to differentiate one rock unit from another, and do not refer back to the key shown above (from Larson and Schaetzl 2001).

Figure 1.1: Bedrock Geology of the Great Lakes Watershed

They explain that during the Quaternary, preferential erosion excavated the deep basins of the Great Lakes. An upper limit of total amount of glacial erosion can be estimated if, as suggested by Larson and Schaetzl (2001), prior to the Quaternary water drained from the region west of the Great Lakes to the Atlantic through the Laurentian or similar channel. Assuming that was the case, elevations along the channel must have decreased monotonically eastward, and hence in this region where differential subsidence due to tectonics or other causes within the Quaternary period is highly unlikely, any bedrock elevations along this path that are distinctly lower than this former drainage path reflect subsequent erosion.

Figure 1.2 shows a simplified Quaternary sediment thickness map and the maximum extent of the late Wisconsinan glacial lobe. The deepest basin along this former fluvial drainage way is now occupied by Lake Superior; it has a floor about 400 m below its rocky outlet. Taking into account the sediments in the lake, the bedrock surface in the lake is at least 270 m deeper (Figure 1.3). The sum of these two values, 670 m, provides a measure of the most erosion that has occurred differentially in the Great Lakes region through the Quaternary. The absolute value of the maximum actual erosion could, however, differ from this value. Notably, it could be somewhat greater because: 1) the 270 m of sediment shown in Figure 1.3 taken from Soller (1992) may be exceeded elsewhere in Lake Superior; and 2) the bedrock outlet could also have been lowered by erosion. On the other hand, less erosion would be required to account for the 670 m depth, if a lake already existed at the site of Lake Superior prior to the glaciations. The ancient existence of such a lake and its mode of formation are poorly known, but there was ample time between the formation of the rocks underlying the Bruce nuclear site and the beginning of the Quaternary for significant vertical crustal motion to occur in this region. Returning to the 670 m, for which there is solid evidence and acknowledging the confounding effects just mentioned, it does provide a measure of the maximum erosion that is possible over a period of ~2 Myr. It should be recognized as a maximum, rather than a representative figure for the region, because it is derived from the deepest basin of the whole region occupied by the LIS, and hence most likely represents the most erosion in the region due, presumably, to a singular combination of particularly dynamic ice sliding on particularly erodible bedrock. Assuming that the excavation of Lake Superior took place over the last 2 Myr, the pace of differential glacial erosion has averaged 0.33 mm/yr (Figure 7.1).

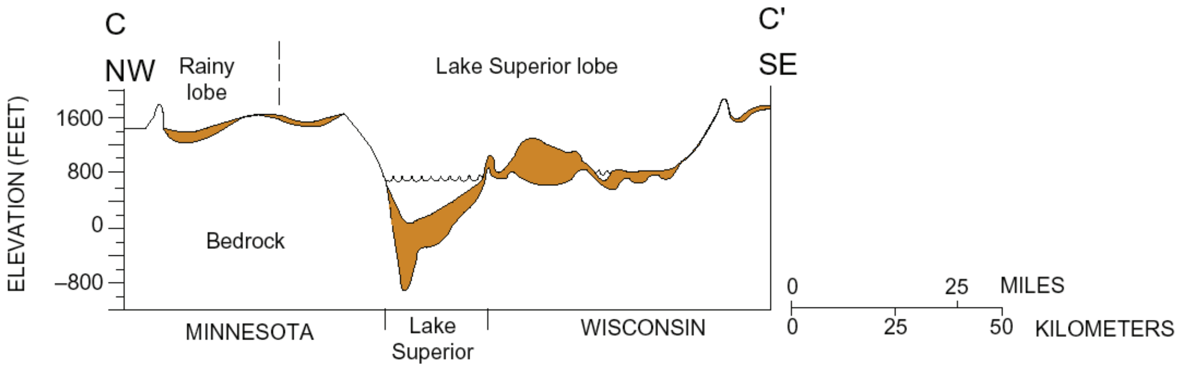
The combined effect of ice dynamics and bedrock properties on glacial deposition, as well as erosion, has long been recognized in the glacial geology of the Great Lakes basin region. Based on map patterns in the region, in concert with information on till texture, ice lobe pattern, and bedrock lithology, Soller and Packard (1998) suggest the following: “The differential resistance of bedrock to erosion by water and ice has partly controlled till textural distribution, especially in the Great Lakes basin. Mode of deposition (for example, interlobate area, basal ice, stagnating ice) also greatly affects the texture of the deposited till. Most of the largest areas of thick glacial sediment were late Wisconsinan interlobate areas on topographically high areas of the bedrock surface, whereas relatively thin deposits generally accumulated in the adjacent bedrock lowlands occupied by drainage and by ice lobes. The lithology of the bedrock and its resistance to erosion in part controlled the pattern of ice lobation and the distribution of thick sediment. The thickness of late Wisconsinan sediment accounts for only a minor part of the sedimentary sequence in the thick drift of these interlobate areas. It is likely, therefore, that once ice lobation had become established in an area, there was a tendency for ice lobes and interlobate areas to recur at roughly the same locations in successive glaciations. Thus, the general configuration of the bedrock surface may have been established in pre-Pleistocene time or after the earliest glaciations. Through successive glaciations, the bedrock topographic highs separating adjacent ice lobes continued to receive additional sediment from the lateral margins of each lobe, adding to the overall thickness of sediment in the interlobate area and further emphasizing topographic control on ice movement.”



Notes: Only the 200- and 400-ft thickness contours are shown (brown solid lines). Dashed "limit of map area," shows the maximum extent of glacial ice. LE, Lake Erie lobe; LH, Lake Huron lobe; S, Saginaw lobe; LM, Lake Michigan lobe; LS, Lake Superior lobe; RR, Red River lobe; DM, Des Moines lobe; J, James lobe. Cross-section C-C' is shown in Figure 1.3 (from Soller 1992).

Figure 1.2: Simplified Thickness Map of Quaternary Sediments Showing Late Wisconsinan Glacial Lobe Position

These conclusions have clear implications for the Bruce nuclear site, which is situated on a high ridge that owes its existence, at least in part, to modest erosion compared with the significant excavation that occurred immediately upglacier in Georgian Bay and downglacier in Lake Huron. Making the reasonable assumption that the region had very little relief in a northeast-to-southwest direction prior to the glaciations, which is consistent with much classic work on ancient low-relief surfaces in the region (Baker 1916, Collins 1925), the present topography and bathymetry demonstrate clearly that, whereas Georgian Bay and Lake Huron have been deeply excavated by ice, much less erosion must have occurred in the French River area to the northeast of Georgian Bay and the Bruce Peninsula. The contrast in amounts of erosion recorded in the topography more likely reflects differences in bedrock than in ice dynamics because ice to the southwest flowed right across this Georgian Bay and Lake Huron region. The erosional contrast between the Peninsula and the lakes does not shed much light on how much erosion occurred on the Peninsula, but because of its extremely low relief, the terrain on all sides of these lakes suggests that erosion of the Bruce Peninsula was small compared to the depth of the adjacent lakes. This would be consistent with Collins' (1925) suggestion of long ago, that the current landscape on the north side of Lake Huron closely resembles the low-relief Precambrian surface.



Notes: This shows relations among bedrock topography, Quaternary sediment thickness, and ice lobation. Colored areas represent Quaternary sediment. Location of section is shown in Figure 1.2. Brown color represents ~ 270 m of sediment filling in deepest portion of the southwest portion of Lake Superior and the thick sediment over the bedrock high, which was deposited between two sublobes of the Lake Superior lobe (from Soller 1992).

Figure 1.3: Cross-section of the Southwest Lake Superior Area

Figure 1.2 from Soller and Packard (1998) helps place the Bruce nuclear site in a broader regional context from the glacial geology point of view. It is located relatively close to the southern margins of the LIS. The vast expanses of bedrock to the north contrast with the region near the LIS margins where little or no bedrock erosion has occurred and where considerable glacial drift has accumulated. Hence, in general, net erosion at near-margin sites like the Bruce Peninsula tends to be low, considerably lower than the average for the LIS. First, the site was covered by temperate ice for only a small fraction of the duration of the glaciations (see Section 5.1). Second, when it was ice covered, bedrock erosion only occurred when: 1) the bedrock was not thickly mantled by drift; and 2) the ice sheet was not actively depositing till and other sediments. Many distal sites are thickly mantled with debris, including the region around the Bruce nuclear site (Figure 1.2). Moreover, where the drift thickness is much greater than the LGM drift, it is evident that no bedrock erosion occurred and little or no erosion of sediment occurred during LGM; the old sediment sequence was not entrained by the LIS but, instead, was covered by additional younger sediment.

Summing up, the present topography and bathymetry, together with the spatial distribution of glacial sediments, suggest that with future glaciations erosion of the Bruce Peninsula would continue to be limited, relative to the erosion that has occurred in the adjacent lakes, because of both the resistant bedrock, and the glaciological circumstances that make this region a site of intermittent or no erosion, and net deposition. Thus, the deep erosion evident in the basins of the Great Lakes themselves most likely results from continuous steering of each new advance of the LIS into these ever deepening depressions. The highlands, such as the Bruce Peninsula, are therefore regions protected from significant erosion by both this steering effect and the resistance of the local bedrock to erosion.

1.3 Cosmogenic Nuclide Studies

Cosmic rays hitting the earth surface produce distinct nuclides or isotopes (such as ^{10}Be, ^{26}Al, and ^{36}Cl) in rocks and other materials. Studies of these isotopes on both glacially eroded bedrock and glacial sediments provide a wealth of new measures and constraints on the amount of glacial erosion at specific sites, and over various periods of time. A number of investigators have studied the concentration of cosmogenic nuclides on surfaces eroded by the Laurentide and Fennoscandian ice sheets. Many of these studies are aimed at defining the chronology of glacial retreat by determining the exposure age. This is a simple exercise, in principle, because as soon as ice retreat exposes the bedrock to cosmic ray bombardment, cosmogenic nuclides progressively build up with time in the bedrock at a rate that is known. Hence, the concentration of cosmogenic nuclides becomes a measure of the duration of exposure while not shielded from cosmic rays by ice, and hence, of the duration of prior ice retreat and exposure of the surface to open skies and cosmic rays. The nuclide concentrations can be affected by: 1) shielding by water or sediments; and 2) the spontaneous decay of unstable nuclides. However, this latter effect generally has little significance on the time scale of interest here, 100,000 years, which is much shorter than the half-lives of the commonly used nuclides. For example, the half-lives of ^{10}Be and ^{26}Al are, respectively, 1.36 and 0.7 Myr.

Several studies, however, have shown that the concentration of cosmogenic nuclides is much higher than what could have built up since deglaciation, typically on the order of 10,000 years, indicating exposure of bedrock prior to the last period of local ice cover. Importantly for this report, cosmogenic nuclide production decreases exponentially with depth, generally reaching insignificant values 1 to 2 m below the rock surface, hence samples with significant prior exposure from any study site indicate that local bedrock erosion was very limited during the last period that the site was covered by ice. However, no cosmogenic nuclide data are available for the Bruce peninsula but considerable insight about the magnitude of glacial erosion can be gained from other sites in the region that were covered by the LIS.

1.3.1 Bedrock

Cosmogenic nuclide build-up does not provide a direct way to measure subglacial erosion rates, but any persistence of ^{10}Be and ^{26}Al from previous exposure episodes, during retreat phases of the Wisconsinan or prior periods of deglaciation, shows that subglacial erosion was less than a few metres during the last period of ice cover (Briner and Swanson 1998, Bierman et al. 1999, Davis et al. 1999, Fabel et al. 2002, Davis et al. 2006, Harbor et al. 2006, Li et al. 2005, Marquette et al. 2004, Phillips et al. 2006, Sugden et al. 2005).

Colgan et al. (2002) show that, at least at a few sites in south-central Wisconsin, USA, where bedrock was clearly abraded by the LIS near its southern margin during the LGM, the cumulative erosion was less than 1 m. These were on resistant bedrock outcrops that retained the glacial striae. On nearby sites, considerably more erosion occurred to remove at least the upper 2 or 3 m of bedrock that contained appreciable concentration of cosmogenic nuclides due to exposure to cosmic rays prior to the Wisconsinan advance. In view of the continuity of the exposed bedrock surfaces and close mutual proximity of many of the samples (at times within 100 m or much less), it is highly unlikely that any of the sites studied experienced considerably more erosion than 10 to 30 m. This bedrock information, taken collectively, leads to a range of 0.5 to 30 m (Figure 7.1).

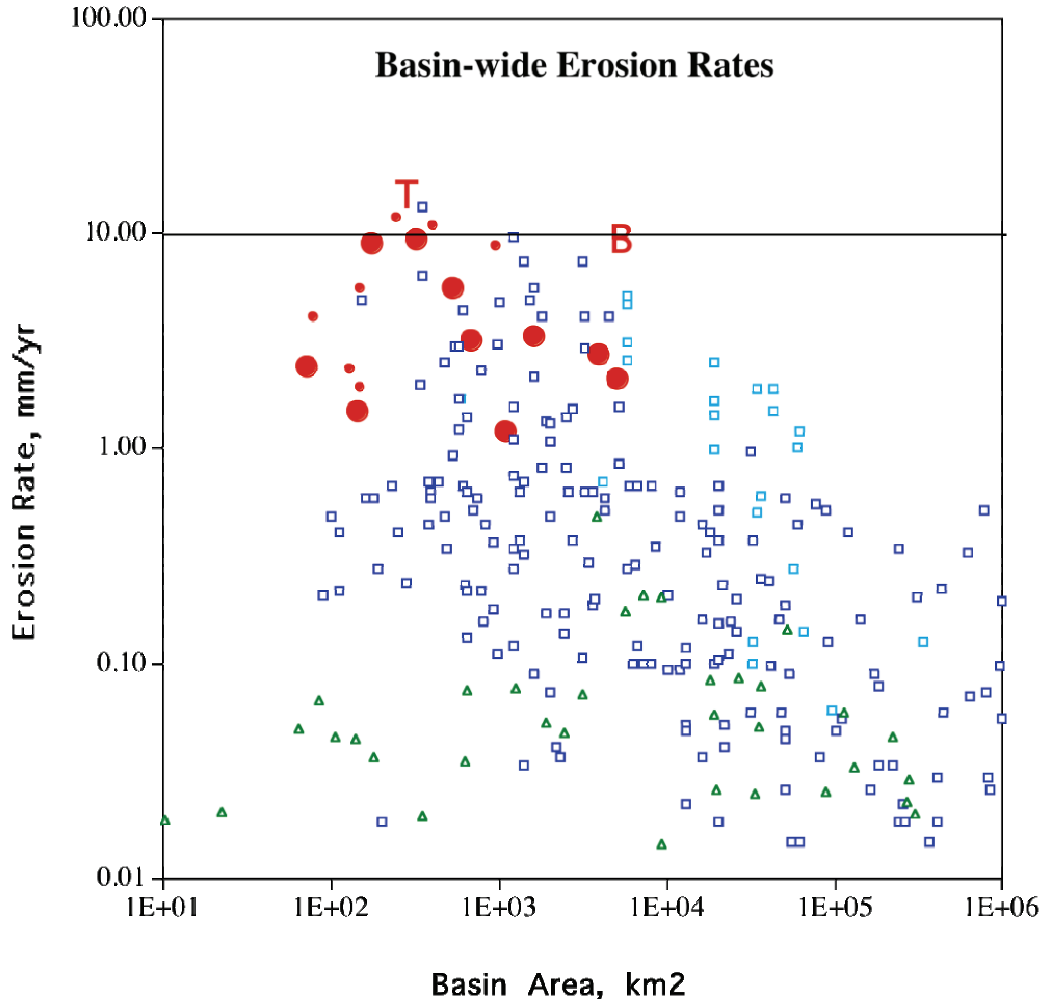
2. GLACIAL EROSION IN DIVERSE SETTINGS

2.1 Contemporary Erosion Rates for Alpine Glaciers

Glacial erosion is a principal issue in contemporary research on high-latitude climate change and landscape evolution in tectonically active mountain ranges (e.g., Champagnac et al. 2009, Egholm et al. 2009, Koppes and Montgomery 2009); hence it has received, and is continuing to receive considerable attention. For the purposes of this report the highest known rates of glacial erosion worldwide are of particular interest. These are found in the highest coastal mountain range in the world, the Chugach and Wrangell-St. Elias Mountains with peak elevations exceeding 5000 m. The high elevations of these mountains effectively intercept Pacific storms and induce heavy precipitation (2 to 3 m/yr, mostly as snow; Wilson and Overland 1987) that fuels some of the largest and most erosive valley glaciers on earth (Hallet et al. 1996). In many drainage basins, glacial coverage is practically complete. The rapidly uplifting range is being cut down by ice as fast as it grows (e.g., Bird 1996).

Glacial erosion rates for the last few centuries in coastal Alaska are estimated to range from 1 to 15 mm/yr (Figure 2.1). They overlap the high end of a recent, global compilation of ~400 glacial erosion rates (Figure 2.1) estimated using five different approaches, which range from 10^{-4} mm/yr to 10 mm/yr (Delmas et al. 2009). The higher rates from coastal Alaska generally exceed other regional rates worldwide, and as discussed below, they greatly exceed rates on geological time scales (100,000-1,000,000 years). The high glacial erosion rates from coastal Alaska appear sound and robust, however, for a number of reasons: 1) they represent a large data set gathered with well established techniques by a number of researchers; 2) they represent sediment yields over a vast region, rendering them reliable indicators of regional erosion; 3) in many cases the sediment volume estimates span many decades and, hence, are much less likely to reflect inherent seasonal or other high frequency variations than those from suspended sediment flux studies and other types of studies that typically last only a few years or decades; 4) the fjord sediments include bedload material that is rarely measured in studies of fluvial sediment fluxes; 5) many of the study areas are so heavily covered with glaciers that subaerial storage and input of sediment do not importantly influence estimates of sediment accumulation rate, and hence of glacial erosion rate; and 6) essentially all fjords studied are sites where the glaciers have undergone tens of kilometres of retreat with few or no minor interruptions, making it unlikely that sediment yields reflect the glacier recycling sediments as they over-ride terminal moraines or glacial drift.

It is noteworthy that sediment yields for the last few decades are up to 5 times greater than those corresponding to the 1-15 mm/yr range for the last centuries, presumably because they reflect a transient state of unusually rapid basal ice motion for the glaciers in the region (Koppes and Hallet 2002, Koppes et al. 2009). They have generally undergone spectacular retreats since the height of the Little Ice Age (ca. 1750-1850) due to accelerated ice motion, as well as calving. For example, glacial erosion rate for the basin of Tyndall Glacier averaged ~38 mm/yr for five decades. Accounting for the exceptionally dynamic and erosive state of Tyndall Glacier through the last century reduces the estimated erosion rate for time scales of centuries to ~15 mm/yr (Koppes and Hallet 2006).



Notes: The erosion rates averaged over Alaskan glacial basins are represented by the red circles; they tend to be above the rates from a world-wide set of basins that are largely free of glaciers (dark blue squares from Milliman and Syvitski 1992), a set of basins in the Himalaya (light blue squares) and a set of basins in British Columbia (green triangles from Slaymaker 1987). The T and B represent rates for the Tyndall and Bering glaciers, respectively. Figure modified from Hallet et al. (1996).

Figure 2.1: Comparison of Contemporary Sediment Yields from Alaskan and Other Worldwide Basins

2.2 Erosion Rates Decrease with Increasing Time Scale

The discussion above refers to glacial erosion rates averaged over decades to centuries, but these are considerably higher than known long-term rates, which are relevant for assessing the erosion potential at the Bruce nuclear site. Our own recent studies exemplify the contrast between current erosion rates and those sustained for time scales approaching or exceeding 10^5 years. Based on a sediment yield study, Koppes et al. (2009) reported exceptionally fast erosion averaging 39 ± 16 mm/yr over the past 50 years for Marinelli Glacier in Tierra del Fuego. Maximum Quaternary tectonic uplift rates in the region have been estimated at 1 mm/yr (e.g., Diriason et al. 1997). Such rapid erosion therefore cannot be sustained, or the Cordillera Darwin massif, which supports the glacier, would have been flattened within one glacial cycle. These rates must reflect a short-lived period of unusually rapid sediment excavation and rapid erosion. Similarly, the concurrent rapid thinning and recession of Marinelli Glacier must also be highly unusual, for at recent rates of retreat (over 13 km in 50 years) and thinning (over 200 m in 50 years) there would be no ice remaining in the basin within a century. It is noteworthy that this result, ~40 mm/yr over the past 50 years, is robust and based simply on the well defined sediment volumes in Marinelli Fjord determined from seismic profiles and the known retreat history.

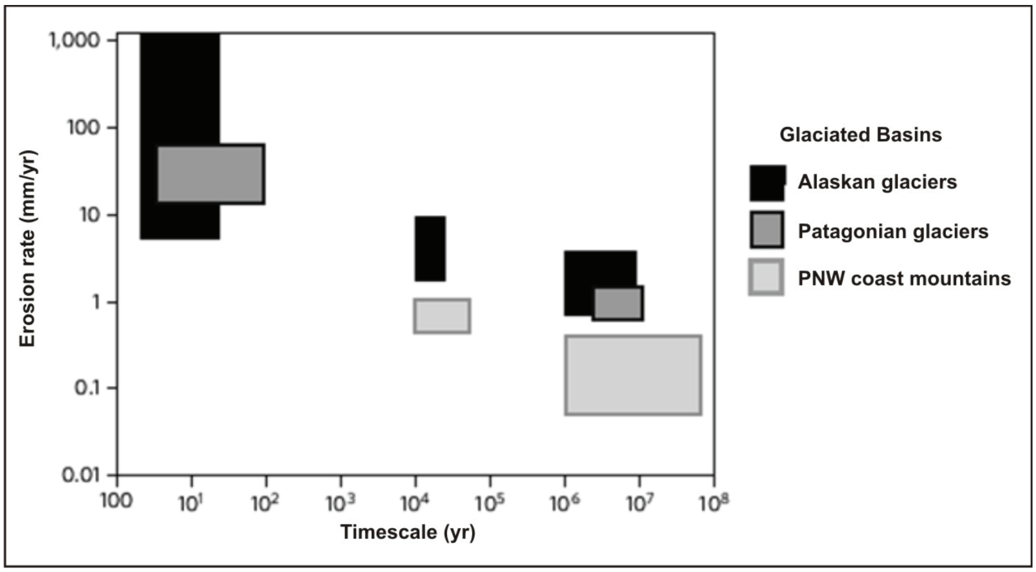
For Marinelli Glacier considerable recent data permits erosion rates to be defined unusually well for a range of timescales (Fernandez-Vasquez et al. 2009). They determined sediment volumes in the fjord in front of Marinelli Glacier using a dense grid of high- and low-frequency single channel seismic data and swath bathymetry along with piston and Kasten cores. Their results show dramatic differences in sediment delivery to the fjord and, by inference, in erosion rates averaged over the glaciated basins, for different timescales. Erosion rates at Marinelli Glacier decrease systematically as the time span over which erosion rates are averaged increases: ~40 mm/yr for the last 50 years, ~5.3 mm/yr for the last 350 years and ~0.5 mm/yr for the last 12,500 years. This trend of decreasing erosion rates appears to continue to time scales of millions of years, based on extensive new thermochronologic data from the Cordillera Darwin massif (Gombosi et al. 2009). The data show that only 600 to 1500 m of bedrock were exhumed in the last 30 Myr in this region, and suggest that about half of the exhumation occurred in the last 10 Myr. (“Exhumation” is generally used in thermochronologic studies to refer to the removal of rock from a region by any means. Erosion and tectonic removal of shallow crustal material or lateral extension of the crust are the primary modes of exhumation. In mountainous terrain subject to a wet climate and rapid incision by rivers and glaciers, tectonic exhumation is likely to be relatively insignificant, making it reasonable to use exhumation and erosion interchangeably). Due to the relative efficiency of glacial erosion, much of this exhumation erosion probably occurred during the Quaternary. An upper limit of the erosion rates averaged through the Quaternary, of 0.15 to 0.4 mm/yr, can be obtained by making the extreme assumption that the erosion in the last 10 Myr occurred entirely during the Quaternary. Collectively, the data from the Marinelli region illustrate how current erosion rates determined from sediment yields from tide-water glaciers exceed average long-term rates by orders of magnitude: the erosion rate averaged for the last 50 years is at least 100 to 270 times larger than the average Quaternary rate.

The erosion data in the preceding sections were obtained in fjords in front of tidewater glaciers. It is widely recognized that these types of glaciers undergo cycle of advance and retreat that reflect conditions at the ice front that dictate the rate of ice loss there by calving and submarine melting. Since much of a normal tidewater glacier cycle is spent in a quasi-stable advance mode (Meier and Post 1987), the recent phase of rapid retreat, rapid ice motion and associated erosion at Marinelli Glacier, is likely to be relatively short. During the much longer advance phase, the glacier first must evacuate the proglacial and subglacial sediment collected in the basin before it can erode its basin anew. Although total sediment yields would be high during such an advance due to excavation of subglacial sediments, the bedrock would be shielded from erosion as long as sediments overlie it.

Both the recent and long-term sediment yields and erosion rates at Marinelli Glacier are amongst the highest reported rates worldwide, similar to the largest Alaskan tidewater glaciers, many of which have also experienced drastic retreat, but which are located in a considerably more active tectonic setting (e.g., Powell 1991, Hallet et al. 1996, Koppes and Hallet 2002).

The similarity in the correlation between rapid erosion and glacier retreat in both hemispheres suggests that this correlation is universal for retreating tidewater glaciers. As these glaciers are all responding to a century of exceptionally rapid warming following the end of the Little Ice Age, the unusually rapid ice motion typical of the retreat phase most probably biases all recent rates of erosion by tidewater glaciers. The discrepancy between erosion rates and uplift rates in both of these regions underscores the transient state of such glaciated landscapes.

A wealth of exhumation data recently obtained from the Chugach-St. Elias Mountains reveal exhumation rates up to 5 mm/yr (Berger et al. 2008a and 2008b, Enkelmann 2008, 2009), precisely where glacial erosion rates reach 15 mm/yr over the last few centuries. For the purposes of this report, the 5 mm/yr can be regarded as the upper limit of known erosion rates on time scales 10^5 to 10^6 years for regions continuously occupied by fast moving temperate glaciers (see Figure 2.2). The Bruce nuclear site is not such a region, largely because of its geologic, tectonic and climatic setting, as well as its infrequent ice cover due to its location close to the margin of the LIS (as discussed subsequently).



Notes: Boxes represent ranges of erosion rates, including errors in estimating erosion rates (height) and timescale of measurement (width). Erosion rates measured from the same or nearby glaciated basins in Alaska, Patagonia, and the coast mountains of Washington State in the Pacific Northwest (PNW). See text for additional and more recent data from Patagonia (Tierra del Fuego) where erosion rates have averaged 0.15 to 0.4 mm/yr through the Quaternary (~2 Myr), hence the rates shown above for the 10^6 to 10^7 year interval are too high by nearly an order of magnitude (from Koppes and Montgomery 2009).

Figure 2.2: Comparison of Short-term and Long-term Erosion Rates from Glaciated and Fluvial Basins

2.3 Distinction between Bedrock and Sediment Erosion

In considerations of erosion rates under ice sheets it is important to distinguish between erosion of bedrock, and erosion of sediments. Because sediment is much more easily entrained, it would be inappropriate and misleading to utilize rates of sediment erosion as a measure of bedrock erosion. Herein, examples of erosion rates determined for unlithified sediments are provided, and the tendency for sediments to shield the bedrock from erosion is discussed.

A rare opportunity to examine the consequences of a glacier advancing over recently deposited sediments arose from the recent advance of Taku Glacier in southeast Alaska (Motyka and Post 1993). Early bathymetric measurements in 1890 showed a shallow bay in front of Taku Glacier. Since then, the glacier has advanced 8 km. Radar soundings show that within a few kilometres of the terminus, the glacier bed is ~100 m below the level of the 1890 sediment surface; hence, a substantial volume of sediment was removed as the ice over-rode the area. Contrasting the early bathymetry with 1989 soundings indicates that 0.5 to 0.6 km^3 of sediment were remobilized in about 99 years and rates at which the surface of the unlithified sediment was lowered by erosion, probably mostly due to entrainment by subglacial waters, reached up to 3 m/yr (Motyka et al. 2006).

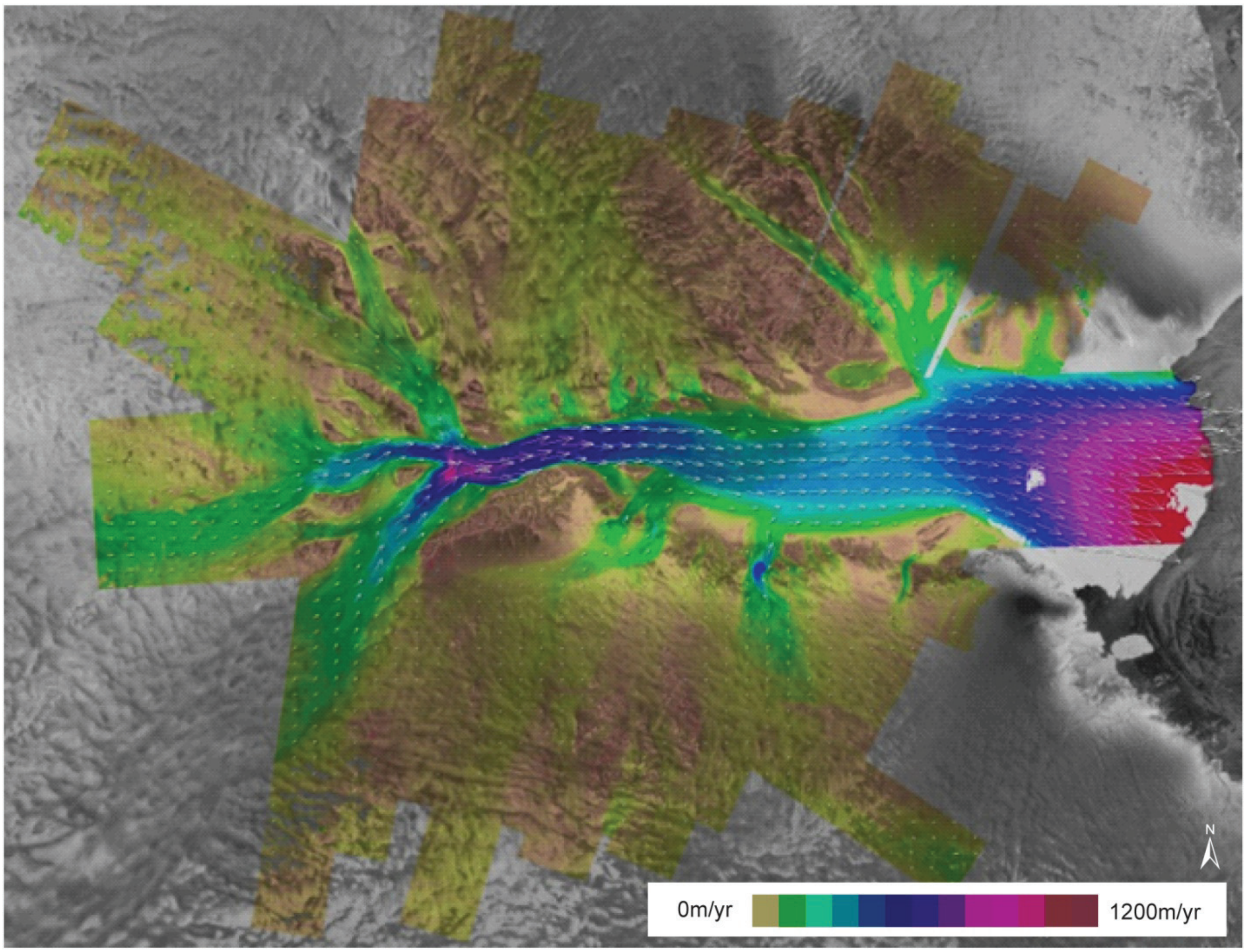
Booth (1994) reported another estimate of rates of sediment mobilization by ice or subglacial water, this one on a millennia time scale in a study of the evolution of topography in the Puget Sound area, western Washington, USA under the Cordilleran Ice Sheet during the LGM. Local stratigraphy suggests that, as ice advanced over the Puget Sound, it flowed across a plain of proglacial outwash, and subsequently eroded the valleys that dissect the Puget Sound area today. Based on the current bathymetry and topography, and on the distribution of remnants of the outwash surface, Booth (1994) estimates that broad tunnel valleys up to 300 to 400 m in depth were evacuated during the glacial advance. The glacial geologic record, and in particular numerous ^{14}C dates defining the brief advance and retreat history, indicate that only about 2500 years were available for this evacuation. Long-term average rates of sediment evacuation must have been ~0.4 km^3/yr, and the evidence suggests that subglacial water was the principal geomorphic agent evacuating the sediment (Booth 1994).

The rapid erosion described in this section has little direct connection with bedrock erosion. However, it highlights the potential of subglacial water to move sediment and by inference to erode bedrock, and the importance of distinguishing between incision of bedrock and incision in sediments. For example, the excavation of 400 m deep mega-channels in 2500 years in the Puget Sound imply an average erosion rate of 0.16 m/yr in glacial outwash and till, but says nothing about erosion of bedrock. Notably, upglacier of these mega-channels, glacial erosion of resistant bedrock has been shown to be as low as a fraction of a metre in the same time interval (Briner and Swanson 1998).

2.4 Extreme Amounts of Erosion: Empirical Upper Limit on Total Erosion

The deepest known excavations by glaciers and by catastrophic glacial floods on our planet provide, in a most general sense, measures of the most (largest total magnitude) glacial excavation that could be reasonably expected anywhere, including the Bruce Peninsula. Although the value of such measures is arguable, they benefit from being simple and empirical. These are reviewed before delving into processes in Chapter 3.

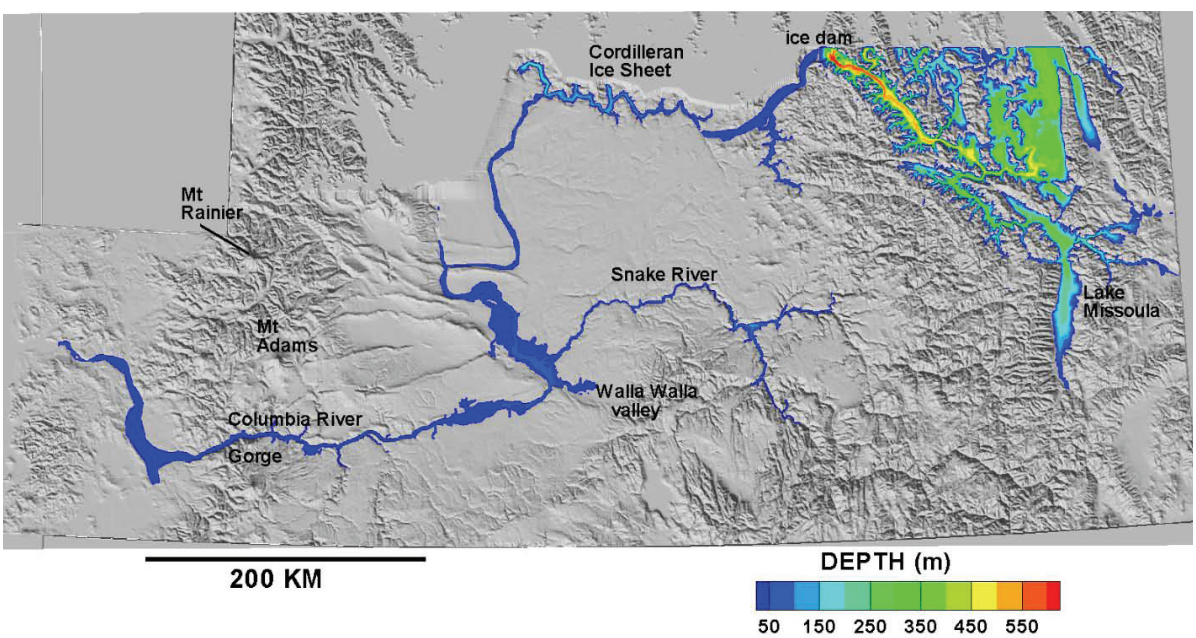
The greatest glacial over-deepening is probably under the largest glacier complex on earth, the Lambert glacier system, which drains a major portion of the East Antarctic ice sheet. Other distinguishing characteristics of this glacier system are its maximum ice thickness, which is likely a world record of 3,500 m, surface velocities peaking around 1 km/yr (Figure 2.3), and its remarkable longevity. According to Jamieson et al. (2005), the Lambert Graben has acted as the principal forcing factor upon erosional pathways in the region for over 118 Myr, initially under fluvial conditions and subsequently under ice sheet conditions over the last 34 Myr. Inferred minimum glacial and preglacial erosion rates are remarkably similar, ranging between ca. 1 and 2 m/Myr. These very low rates are surprising given the rapid motion of the ice, but they are consistent with the exceptional duration of focused erosion in this region, and they show that slowly eroding glaciers can excavate bedrock deeply provided they have sufficient time to do so, in this case tens of millions of years.



Notes: Obtained by using RADARSAT SAR imagery (Canadian Space Agency/NASA/Ohio State University, Jet Propulsion Laboratory, Alaska SAR facility).

Figure 2.3: Ice Velocity Vectors for the Lambert Glacier System

Widespread erosion by catastrophic floods emanating largely from former Lake Missoula is particularly well documented (e.g., Baker 1973, Atwater 1986, Denlinger and O'Connell 2010), and provides a relevant example of extreme localized erosion. Lake Missoula formed and drained repeatedly as the regional drainage was blocked by a massive ice lobe extending south from the Cordilleran ice sheet until the level of the lake rose sufficiently to destabilize the dam and produce the largest well-studied floods that have occurred worldwide. Of particular interest to this report is a much smaller lake, Pend Oreille, which is more than 300 m deep over a distance exceeding 20 km and width of nearly 3 km; it is located at the former outlet of Lake Missoula in Northern Idaho. This unusually deep lake basin was excavated by the floods precisely where ice dams up to 500 m high, formed (Figure 2.4) and failed repeatedly, attesting to the exceptional erosive power of the numerous catastrophic floods where they were funneled through a narrow spillway (Denlinger and O'Connor 2010). The depth to bedrock considerably exceeds the 300 m water depth because thick sediments cover the floor of the lake. Using 500 Hz seismic reflection profiles, Breckenridge and Sprenke (1997) estimated that "the net thickness of unconsolidated sediment at the deepest part of Lake Pend Oreille is about 490 m and that the bedrock basin beneath the lake extends to 214 m below sea level." Recent re-analysis of the original profiles suggests that the bedrock is even lower, extending down to 377 m below sea level. Making the reasonable assumption that the ancestral river draining this region drained westward, as it does now down through the Spokane River at an elevation of 468 m, the maximum depth of glacial outburst flood erosion through the entire Quaternary is about 785 m. This estimate is conservative because it ignores the incision required to erode the ancestral river at the site of Lake Pend Oreille to the level of the outflow from the basin along the Spokane River.



Notes: The modeled glacial Lake Missoula was formed by damming the Clark Fork of the Columbia River at the location labeled 'ice dam' (from Denlinger and O'Connell 2010).

Figure 2.4: Initial Conditions for Lake Missoula Flood Simulations

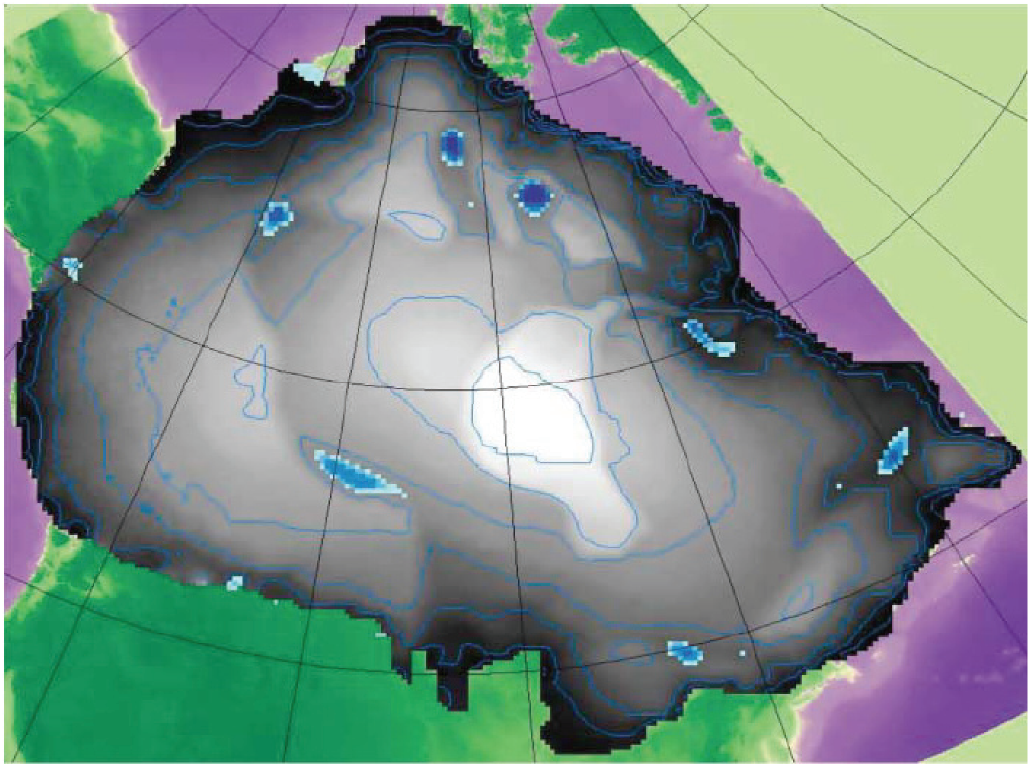
Considerable additional evidence, independent of the seismic data at Lake Pend Oreille, further demonstrates that outburst floods from Lake Missoula can incise bedrock deeply. For example, a recently documented bedrock depression, based on gravity data, that is about 500 m deep and several km wide 30-50 km southwest of Lake Pend Oreille. These examples of bedrock erosion by floodwaters are, as far as the author knows, unsurpassed worldwide. They are distinct from cases of similar, or even greater bedrock erosion from glaciated areas, where the bedrock is eroded by massive, persistent, and fast moving streams of ice that are steered down major valleys. A notable example, which is also in the region formerly covered by the Cordilleran ice sheet, is Okanagan Lake, in south-central British Columbia, where the bedrock floor is nearly 650 m below sea level and more than 2000 m below the rim of the surrounding plateau (Eyles et al. 1991)

Excellent exposures of massive flood deposits, separated by varves, reveal close to 100 catastrophic flood events during the LGM (Atwater 1986), indicating that the massive glacier dam of Lake Missoula failed and re-established repeatedly, every ~20 to 60 years, during the LGM, and in view of the number (~7) of LGM-caliber glaciations during the Quaternary, most probably occurred several hundred times through the Quaternary. Thus, deep bedrock erosion during each LGM-caliber glaciation, lasting a nominal 10^5 years, is estimated conservatively to range from 100 m to 133 m (800 m divided by 6-8). Such pronounced bedrock erosion results from a multitude of exceptionally large outburst floods, but only provided the water flow is highly localized and the floods are funneled persistently over a particular area throughout the period of interest by a deep valley a few kilometres wide. This condition for large outburst floods to be extremely erosive is not met, however, for the Bruce Peninsula or other low relief regions; evidence that the floods were not localized in such terrain comes from Ontario, Quebec, Alberta and the Northwest Territories, where Shaw (2002) has inferred “regional-scale flood tracts extending …several hundred kilometres in width” (see Section 4.1). For outburst floods that cover swaths orders of magnitude wider than the outlet of former Lake Missoula, erosion is expected to be at least an order of magnitude smaller than computed above for the Lake Missoula outlet (100 m to 133 m), especially if they were less frequent. Thus, given the low relief of the region around the Bruce nuclear site and the orientation of Bruce Peninsula generally transverse to the directions of both the ice flow and the inferred subglacial floods, which are not conducive to flow channelization, a reasonable maximum estimate of the bedrock erosion by flood water likely at the Bruce nuclear site in the next glacial cycle is about one tenth of the value for the Lake Missoula outlet; allowing for the additional uncertainty in relating erosion rates in different regions, this upper limit of the total erosion is estimated at 5 m to 25 m (Figure 7.1).

Several regional lines of evidence further suggest that localization of water flow and deep excavation of bedrock is highly unlikely in the Bruce Peninsula region: 1) the dynamic nature and gentle surface of the ice lobes in the southern portion of the LIS (Evatt et al. 2006, Peltier 2011) make it unlikely that the surface topography of the ice localized water flow narrowly over any particular region for an extended period (Figure 2.5); 2) the bedrock erosional features and other bedforms in the Bruce Peninsula region suggest that the catastrophic outburst floods were widespread (e.g., Kor et al. 1991, Kor and Cowell 1998, Shaw 2002); they covered vast low-relief areas rather than narrow channels where erosion could be considerable; and 3) the general lack of large-scale channels paralleling ice flow direction to the southwest suggest that the total erosion did not approach hundreds of metres through the Quaternary (Figure 2.5).

Nevertheless a network of shallow, irregular channels is clearly evident in the topography of the French River area and the Bruce Peninsula, and the bathymetry of Georgian Bay, reflecting modest localized erosion on a smaller scale. Together with the pervasive glacial scouring of the Bruce Peninsula (Kor and Cowell 1998), there are clear signs of widespread glacial erosion. The total depth of erosion, however, is likely small, on the order of the 1 to 30 m scale of the micro-relief characteristic of rock erosional features in the French River area and the Bruce

Peninsula, and of irregular channels in Georgian Bay (see Chapter 4 for further discussion and illustrations).



Notes: Ice sheet thickness shown in grays, highest potential being white. Model shows no tendency to channel subglacial waters (from Evatt et al. 2006).

Figure 2.5: Modeled LGM Hydraulic Potential Surface

3. THEORETICAL CONSIDERATIONS OF GLACIAL EROSION PROCESSES

3.1 Glacial Erosion: Mechanisms

The mechanisms of glacial erosion are widely recognized; they consist of abrasion, quarrying (plucking), and mechanical erosion by meltwater. Subglacial dissolution has also been recognized (Hallet 1976), but the efficiency of this process, as an erosional mechanism is generally not significant in comparison to the mechanical processes.

Abrasion has received the most attention from a theoretical standpoint (Boulton 1974; Hallet 1979, 1981) because the process of abrasion by rocks, which are forced against the bedrock by the ice and entrained by the sliding motion, is relatively straightforward. Hallet (1979) developed a simple equation describing abrasion rate: $Å = \alpha F_c v_r C$, where α is a constant, and C and v_r are the rock concentration (number of rocks in contact with the bed per unit area) and rock speed, respectively. The constant α is known as the attritivity of the rock; it depends on the relative hardness of the abrader and bedrock, shape of the point contacting the bed, and grain size distribution of the abraded debris. Due to the importance of viscous forces in defining the contact force, v_r and F_c (contact force) both scale with sliding velocity (Hallet 1979).

Denoting the coefficient of friction as μ, the work done by one rock in frictional motion over the bed per unit time is $\mu F_c v_p$. Hence, the total work done (energy dissipated) per unit time per unit area on friction/abrasion is $\mu F_c v_p C$. Thus, the rate of glacial abrasion ($Å = \alpha F_c v_p C$) is proportional to the rate at which work is being done, that is the power dissipated, in rock/rock friction at the glacier bed. This simple result will be used in the calculation of abrasion based on the energy dissipation at the base of the LIS.

Quarrying has also been addressed theoretically in several studies, with a focus on how sufficient differential stresses can occur at the glacier bed to fracture bedrock (Iverson 1991, Hallet 1996, Hildes 2001, and Hildes et al. 2004). These studies are useful in illustrating the mechanism and its controls; they show that: 1) the sliding velocity and basal effective pressure (both the average value and the temporal variations in the difference between the ice pressure and the water pressure at the glacier bed) are the key glaciological variables; and 2) the resistance to fracture and the existence of pre-existing cracks are the key lithologic controls.

Meltwater can cause significant erosion, with conspicuous indications of widespread erosion of bedrock and sediments arising during massive subglacial outburst floods, as discussed earlier (e.g., Shaw 2002, Booth 1994). This subglacial fluvial erosion has not been the subject of theoretical studies. Nevertheless it is well worth careful consideration in the context of this report not only for completeness but, especially, because some of the clearest and most spectacular evidence for subglacial floods that has been recognized worldwide is in the southern Ontario region. This evidence is summarized and discussed in Chapter 4.

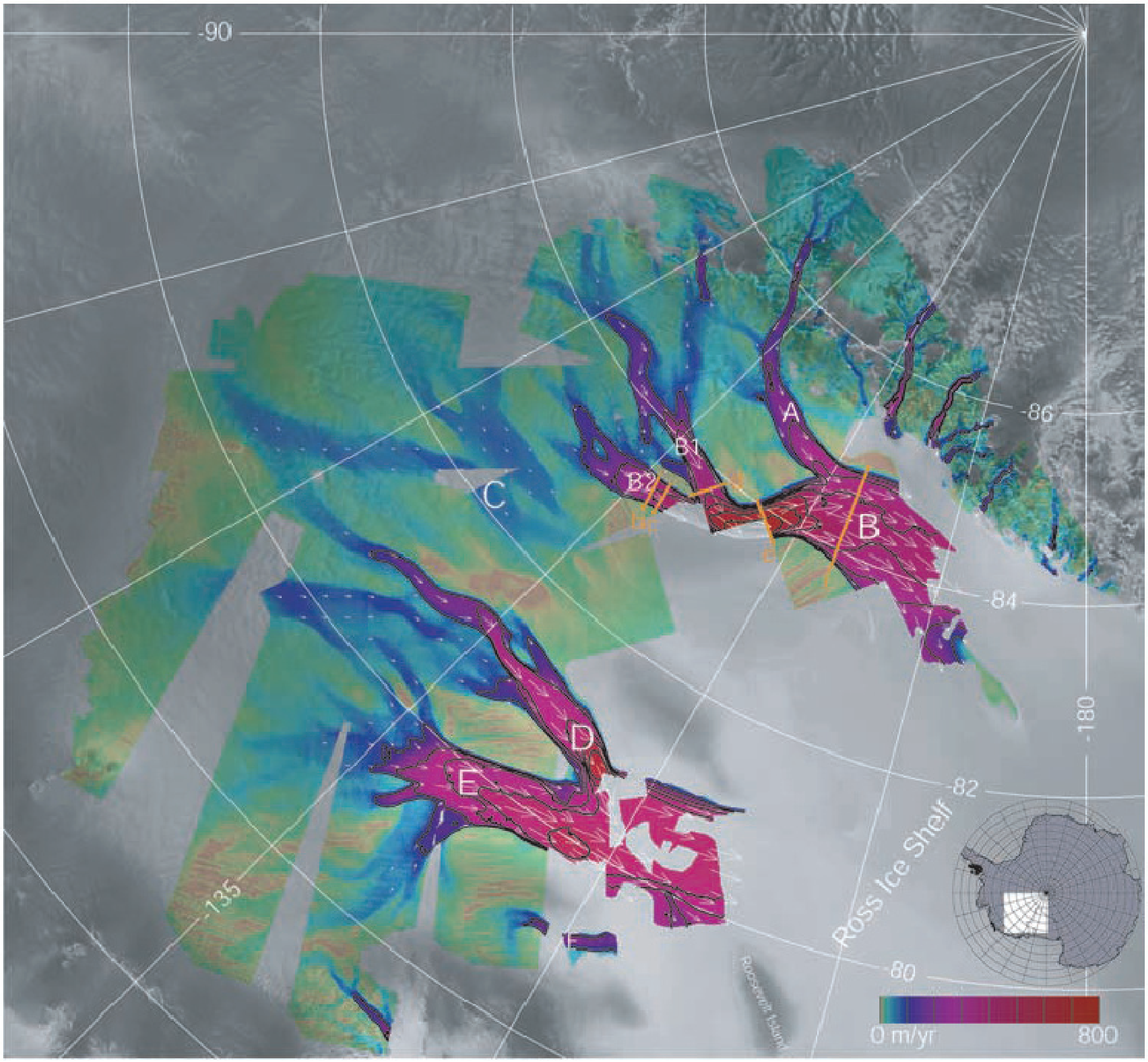
Meltwaters also affect the bedrock erosion by all glacial processes because they evacuate the eroded debris, keeping the bedrock exposed to erosive ice and water. Evacuation of debris can be rate limiting in glacial erosion, especially for the excavation of basins. Such excavation is possible until water and sediment can no longer be evacuated downglacier due either to the hydraulic head gradient being insufficient to drive subglacial water out of the basin, or to the lack of adequate subglacial pathways for water flow. Under temperate glaciers, water that is driven uphill fast decompresses rapidly, and tends to freeze because it is super cooled. The freezing tends to block subglacial pathways for both water and the entrained sediment; the resulting sediment accumulation shields the bed from erosion. This stabilizing feedback in glacier-bed erosion has been examined by Alley et al. (2003) who "find that the long profiles of beds of highly erosive glaciers tend towards steady-state angles opposed to and slightly more than 50 per cent steeper than the overlying ice-air surface slopes, and that additional subglacial deepening must be enabled by non-glacial processes." The implications for the study area are that excavation of the basins of the Great Lakes has been, or will eventually be limited by the ability to drive water and water-entrained sediments out of the basins.

3.2 Basal Energy Dissipation and Glacial Erosion

The basal sliding velocity is widely recognized on theoretical grounds as a primary factor controlling the rate of erosion (e.g., Boulton 1974, Hallet 1979, Egholm et al. 2009). Rapid basal motion, however, does not necessarily imply rapid erosion because the fast motion may reflect decoupling between the glacier and the bedrock. For example, ice streams occur in areas where the basal temperature reaches the pressure melting point and the water pressure approaches the ice pressure; hence the ice is near flotation. The resulting low effective pressure permits rapid basal motion due to the deformation of soft subglacial sediments, and/or the formation of extensive water-filled cavities that separate the ice from bedrock. Rapid basal ice motion occurs when the glacier bed offers little resistance to ice motion; this is widely recognized in nature as illustrated, for example, in Figure 3.1 (Joughin et al. 2004), and is well simulated in ice sheet models (e.g., Peltier 2011). Under the ice streams of West Antarctica, for example, basal shear stresses are typically on the order of 10^3-10^4 Pa (Figure 3.2, Joughin et al. 2004), which are very small compared to 10^5 Pa, the shear stress typically found at the base of alpine glaciers (Paterson 1981). The following discussion outlines a simple and effective means of representing the dependence of the glacial erosion rate on both the speed of basal motion and the strength of the glacial coupling.

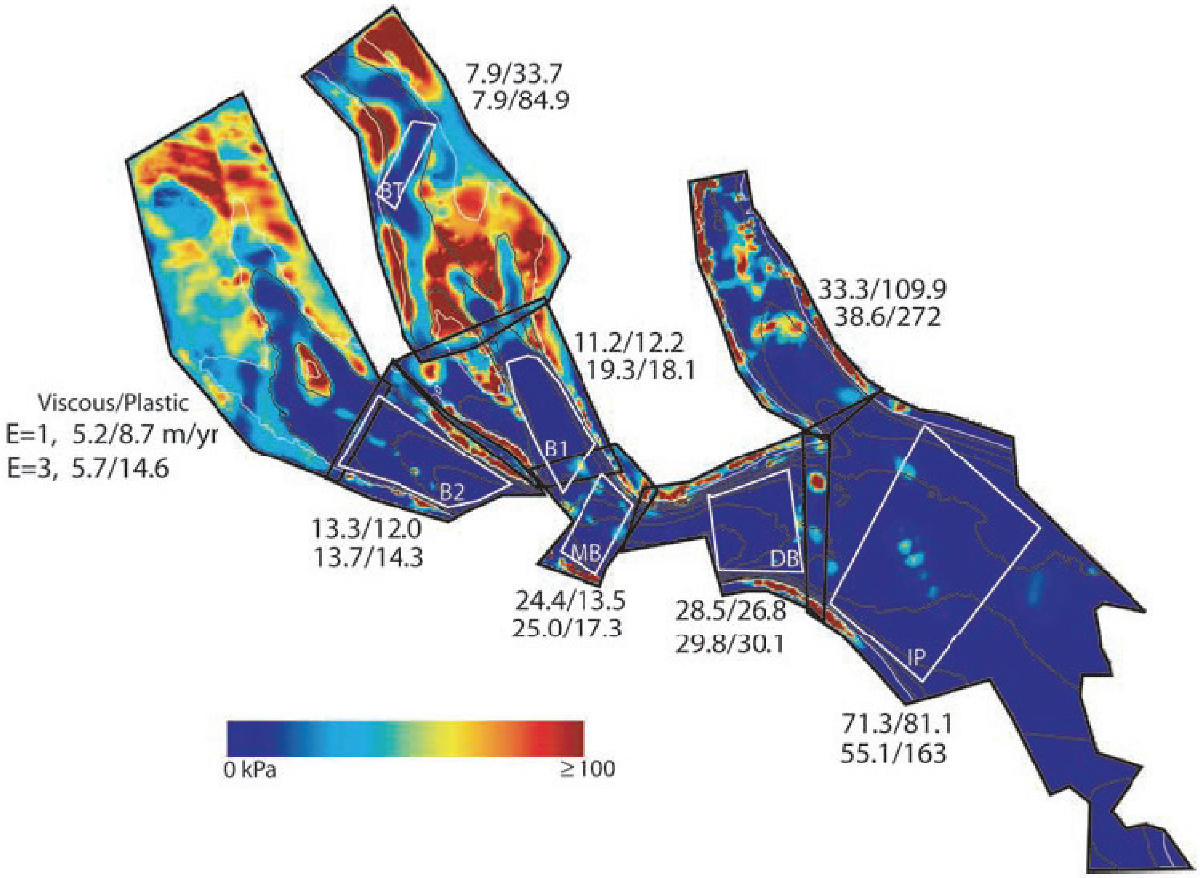
The product, $U\tau$, of the sliding velocity and the basal shear stress is a natural combination of these key parameters that bear on erosion rate. For example, as mentioned above, it is evident from explicit models of glacial abrasion (Hallet 1979) that the abrasion rate scales with the amount of energy that is spent by the sliding ice, per unit time and per unit area, on rock-to-rock friction at the glacier bed. This rock frictional energy is expected to increase with the total energy expenditure due to motion over the bed, which can be shown to scale with $U\tau$, as well as, the number of rock fragments in the ice contacting the bed per unit area, and factors representing the angularity and relative hardness of the rock fragments. Glacial plucking and subglacial water erosion are more complex processes, but they too will tend to increase with the amount of energy dissipated per unit time and per unit area at the bed, which we term basal power, P: glacial erosion due to all mechanisms combined is expected to stop as basal power vanishes, and to increase with $U\tau$ in a way that can be determined empirically. Importantly for this report, results compiled in Peltier (2011) make estimates of $U\tau$ for the Bruce nuclear site particularly convenient because they include time series of basal melting rate, M, which also scales linearly with basal power. These provide a basis for quantitatively assessing erosion rates and total erosion over the study site.

This approach benefits from actual data on glacier dynamics and erosion, and decreases the reliance on theoretical approaches which are not as adequate. This is because absolute rates of erosion cannot be calculated reliably from first principles due to: 1) the complex set of processes involved in subglacial erosion; and 2) poorly known basal conditions and bed properties (roughness, bedrock jointing and hardness, till cover, subglacial hydrology) that generally vary both spatially and temporally. The use of the empirical relationship between long-term erosion rates and basal power together with modeled basal energy dissipation for the LIS provides a sound basis for estimating erosion rates in the study region.



Notes: Results are determined from satellite measurements (from Joughin et al. 2004).

Figure 3.1: Surface Speeds of Ice Streams Draining the West Antarctic Ice Sheet



Notes: Ice stream B as labelled in Figure 3.1, draining the west Antarctic ice sheet (from Joughin et al. 2004).

Figure 3.2: Calculated Basal Shear Stress Distribution for Ice Stream B

According to theoretical studies of glacial abrasion, the erosion rate scales with basal power, increasing from zero when there is no motion on the bed or the glacier is decoupled from the bed by a lubricating water layer, to a finite rate that is best determined empirically from field data. To define the relationship between erosion rate and basal power most clearly, it is instructive to focus on areas where ice motion and erosion rates are high and readily measured. Two such areas exist where considerable recent research permits definition of the range of representative basal power and maximum long-term erosion rates. Two major glaciers in these areas, the Malaspina in coastal Alaska and the San Rafael in Patagonia, have been characterized sufficiently to define the basal power where exhumation rates are known to be high. The basal shear stress of temperate alpine glaciers, including San Rafael, is generally close to 10^5 Pa. For the Malaspina section where exhumation rates are high, the Seward Throat, the shear stress ranges from 0.9 to 2.4 x 10^5 Pa based on finite-element calculations and satellite-based measurements of surface velocity. Representative basal velocities range from 400 to 1400 m/yr for the Malaspina, based on the calculations just mentioned, and 600 to 1200 m/yr for the San Rafael, based on surface velocities and a simpler calculation. The long-term erosion rates, determined from exhumation studies using low-temperature thermochronology, reach up to 5 and 2 mm/yr, respectively, for the Malaspina

(Berger et al. 2008a, 2008b, Enkelmann 2008, 2009) and San Rafael regions (Koppes et al. 2006). Taken together, these data suggest that the ratio of erosion rate (m/yr) to basal power ($J/m^2/yr$) ranges from 1.5 x 10^{-11} to 1.4 x 10^{-10} (1/Pa); these results will be used to calculate erosion at the Bruce nuclear site in Section 5.1.

Caution is in order in using these calculated values of erosion because of a substantial inherent difficulty in calibrating the use of glacier basal power based on current glacial characteristics to calculate rates of long-term erosion. Whereas, basal power can be estimated well for present day glaciers, long-term erosion rates reflect exhumation by any process over a period of time that is so long (much longer that the 10^5 years typical of individual Quaternary glaciations) that glaciers vary tremendously in size and shape. The data from the Malaspina region, and more broadly the St. Elias-Chugach range, in coastal Alaska highlight this difficulty because high exhumation rates are not confined to the vicinity of fast moving glaciers. Rather, a range-parallel band of rapid exhumation crosses a series of basins some of which currently contain no glaciers (Berger et al. 2008a, 2008b); during the LGM, however the whole St. Elias-Chugach range was thickly mantled with ice.

The approach taken here, which assumes that erosion rates scale with power, is very similar to the one taken in contemporary basin-scale models of erosion by alpine glaciers, which generally assume simply that erosion rates scale with basal ice speed (e.g., Harbor et al. 1988, Oerlemans 1984, Tomkin and Roe 2007, Herman and Braun 2008, Egholm et al. 2009). The approaches are similar because the basal shear stress varies over a narrow range for contemporary temperate valley glaciers (Paterson 1981). With regard to erosion by the LIS, however, basal power is preferable to basal speed because shear stress at the base of ice sheets may vary from values around 10^5 Pa, comparable to those typical of alpine valley glaciers, to values up to two orders of magnitude lower, where the lubricating effect of extensive bodies of water or soft-sediments reduce erosion rates, as well as basal shear stresses. In such regions, erosion rates would likely show little or no correlation with sliding speed, but would still be expected to scale with the basal power.

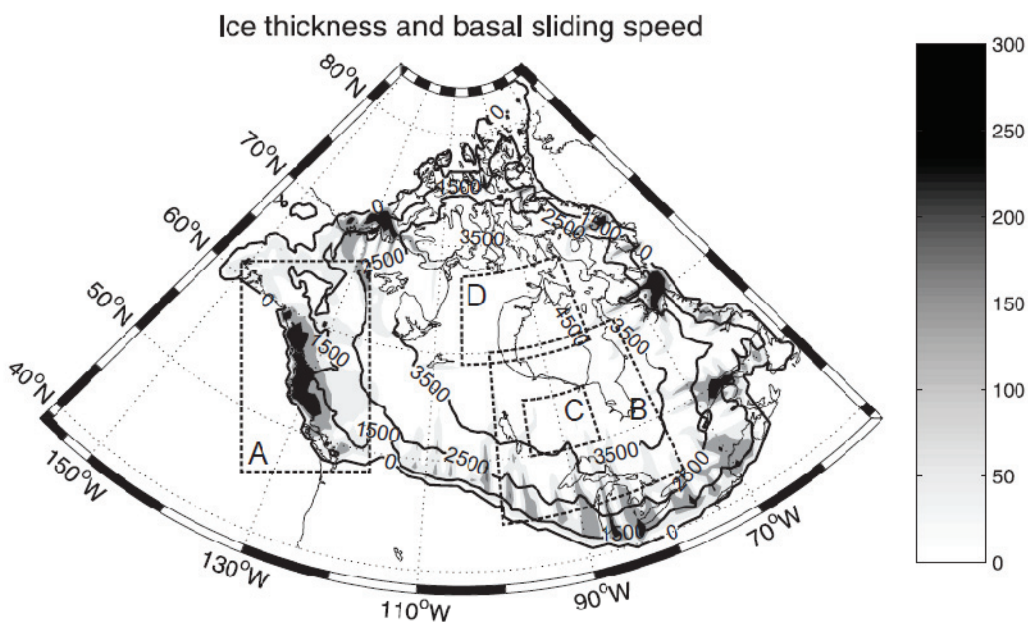
3.3 Model of Basal Processes for the North American Ice Sheet

In a seminal study, Hildes et al. (2004) present "results from the first large-scale physically based model of subglacial processes driven by ice dynamical, thermal and hydrological models…A detailed lithological description of the bed permits geologically based parameter assignments and allows distinctive lithologies to be used as natural tracers. The ice sheet model…, driven by estimates of paleoclimate, is used to predict ice dynamics and thermodynamics and includes a thermally regulated sliding scheme with an enhancement over soft-bedded regions. Subglacial water pressure, a critical input to the subglacial process model, is calculated using a hydrological model for coupled basal and groundwater drainage (Flowers 2000, Flowers and Clarke 2002). For both abrasion and quarrying, the dynamical and hydrological input fields prove to be crucial, with abrasion strongly dependent on sliding speed and quarrying highly sensitive to the water pressure relative to ice pressure." Figures 3.3 and 3.4 are from the Hildes et al. (2004) study.

Hildes et al. (2004) calculate the cumulative erosion over a full glacial cycle (~120,000 years) for the LIS (Figure 3.5). Averaged over the model domain, it ranged from 0.41 m to 0.58 m for two representative models. A maximum model estimate, obtained by prescribing no initial sediment cover (fully exposed bedrock) was 1.63 m of total erosion. These results suggest rather modest erosion during the growth and decay of the LIS.

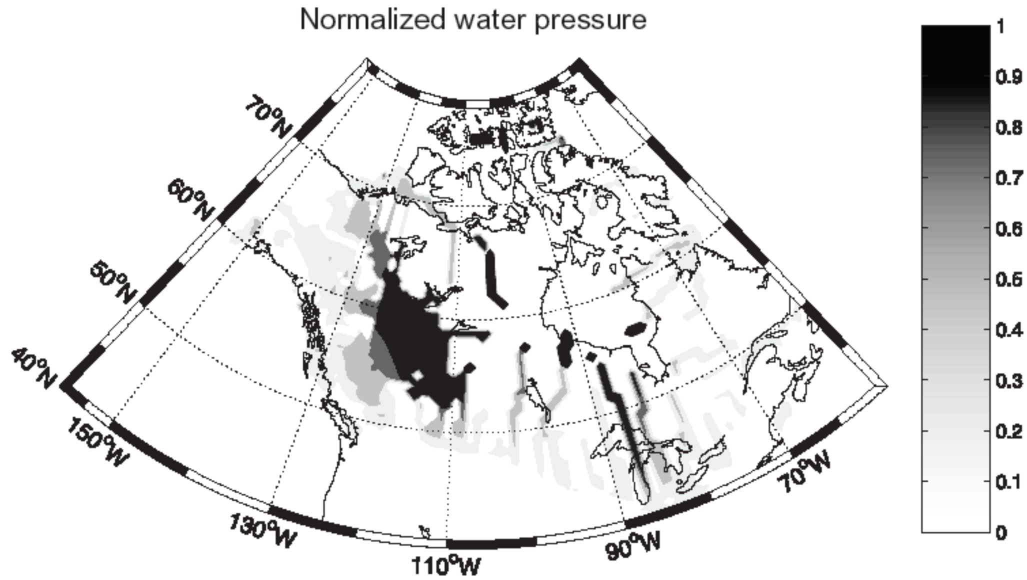
The model presented by Hildes et al. (2004) is most instructive. It is also ambitious and it constitutes a substantial advance in the field. In addition to calculating bedrock erosion, they compare model results and geologic data and examine the spatial distribution of debris in glacial drift within a sound glaciological model. Hildes et al. (2004) point out, however, a number of reasons why these model results underestimate the actual erosion by the LIS. They write, "comparison of mapped Hudson Bay Paleozoic carbonate dispersal with model results shows that the predicted debris distribution is too extensive and the volume of deposited sediment too high. This suggests that either sliding speeds in this region are too high or that the central ice dome structure is incorrect. In contrast, the simulated transport of Dubawnt Group sediment is underestimated relative to the observed dispersal, echoing the suggestion that the Laurentide dome structure is inaccurately modeled. A better match between predictions and observations would result if Keewatin and Labrador Domes were appropriately developed." The latter is well known to be a significant problem as originally conjectured by Dyke and Prest (1987) and demonstrated by Peltier (2002) on the basis of the analysis of modern space geodetic constraints, derived from measurements by both Global Positioning System (GPS) and Very Long Baseline Interferometry (VLBI), which enable one to accurately locate the positions of LGM local extrema in ice thickness.

Caution is in order, however, when considering the magnitude of the calculated erosion over the LGM because the treatment of the hydrology and diverse basal processes – subglacial erosion, sediment entrainment, englacial mixing, advective transport and deposition – and the choice of model parameters are generally poorly constrained, lack solid validation, and involve considerable idealization.



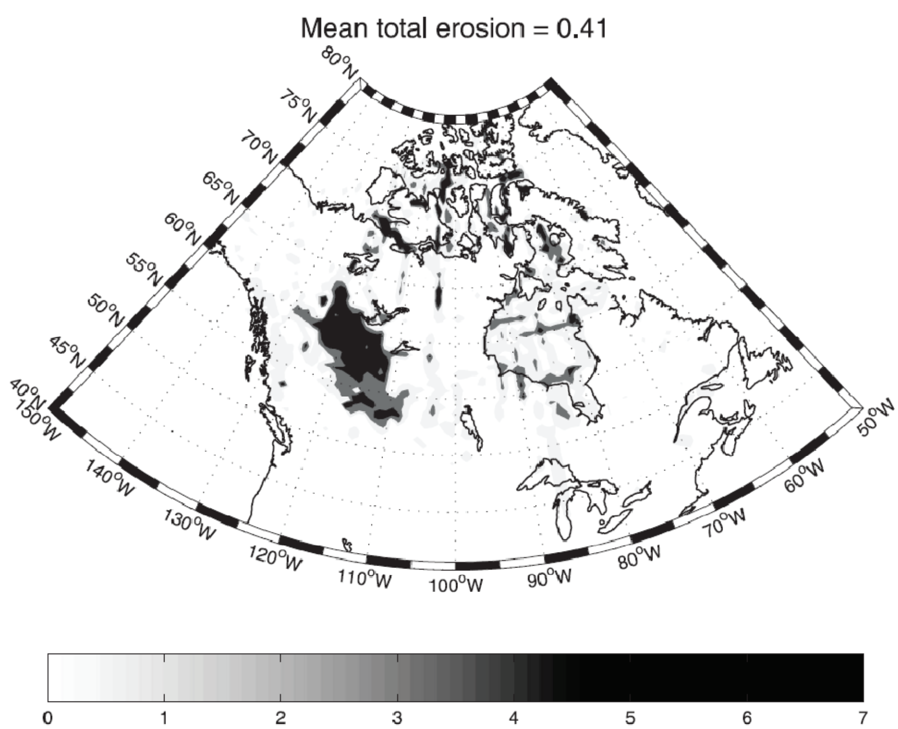
Notes: The grey scale shows sliding speed in metres/year and the contours show ice thickness in metres. Figure is from Hildes et al. (2004).

Figure 3.3: Estimated Ice Thickness and Sliding Speed Computed for the North American Ice Sheet at the LGM



Notes: The grey scale shows the ratio of water sheet to ice overburden pressure at the base of the ice sheet (from Hildes et al. 2004).

Figure 3.4: Estimates of Basal Water Pressure at the LGM



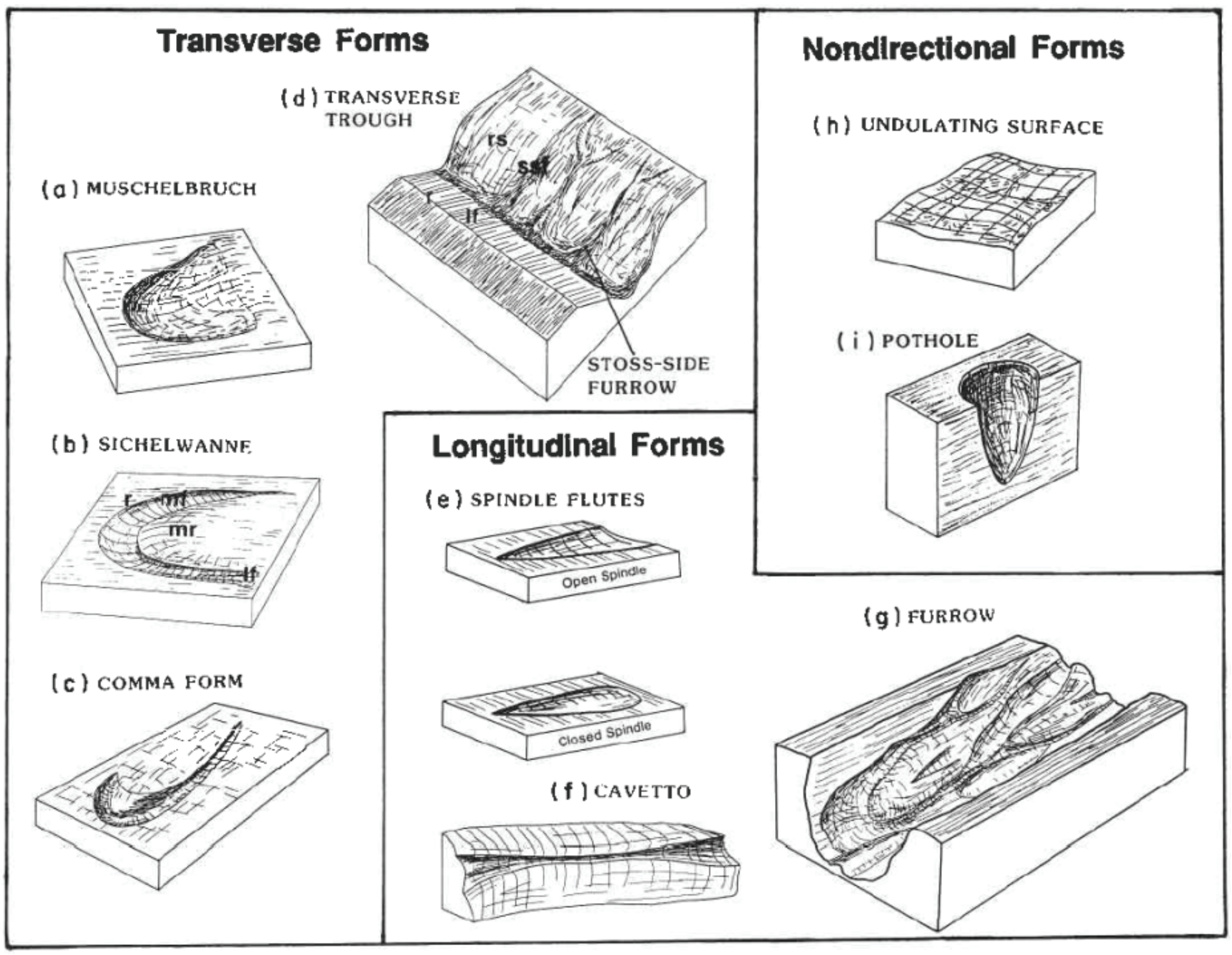
Notes: Grey scale is in metres of total erosion. Figure is from Hildes et al. (2004).

Figure 3.5: Total Erosion Through the LGM

4. GEOLOGICAL OBSERVATIONS: BRUCE PENINSULA AND SOUTHERN ONTARIO

4.1 Erosion by Subglacial Water and Ice in Southern Ontario

John Shaw, in collaboration with a small group of glacial geologists over the last 25 years, has developed a convincing case for the "meltwater hypothesis", which evokes "enormous outburst floods for the formation of subglacial bedforms". Shaw (2002) describes "regional-scale flood tracts in Ontario, Quebec, Alberta and the Northwest Territories extending over 1000 km in length and several hundred kilometres in width" marked by a rich suite of subglacial bedforms, which he unifies under the "meltwater hypothesis". He argues that these bedforms are very broadly defined and include drumlins, bedrock erosional marks, fluting, Rogen moraines, hummocky terrain, and transverse ridges.



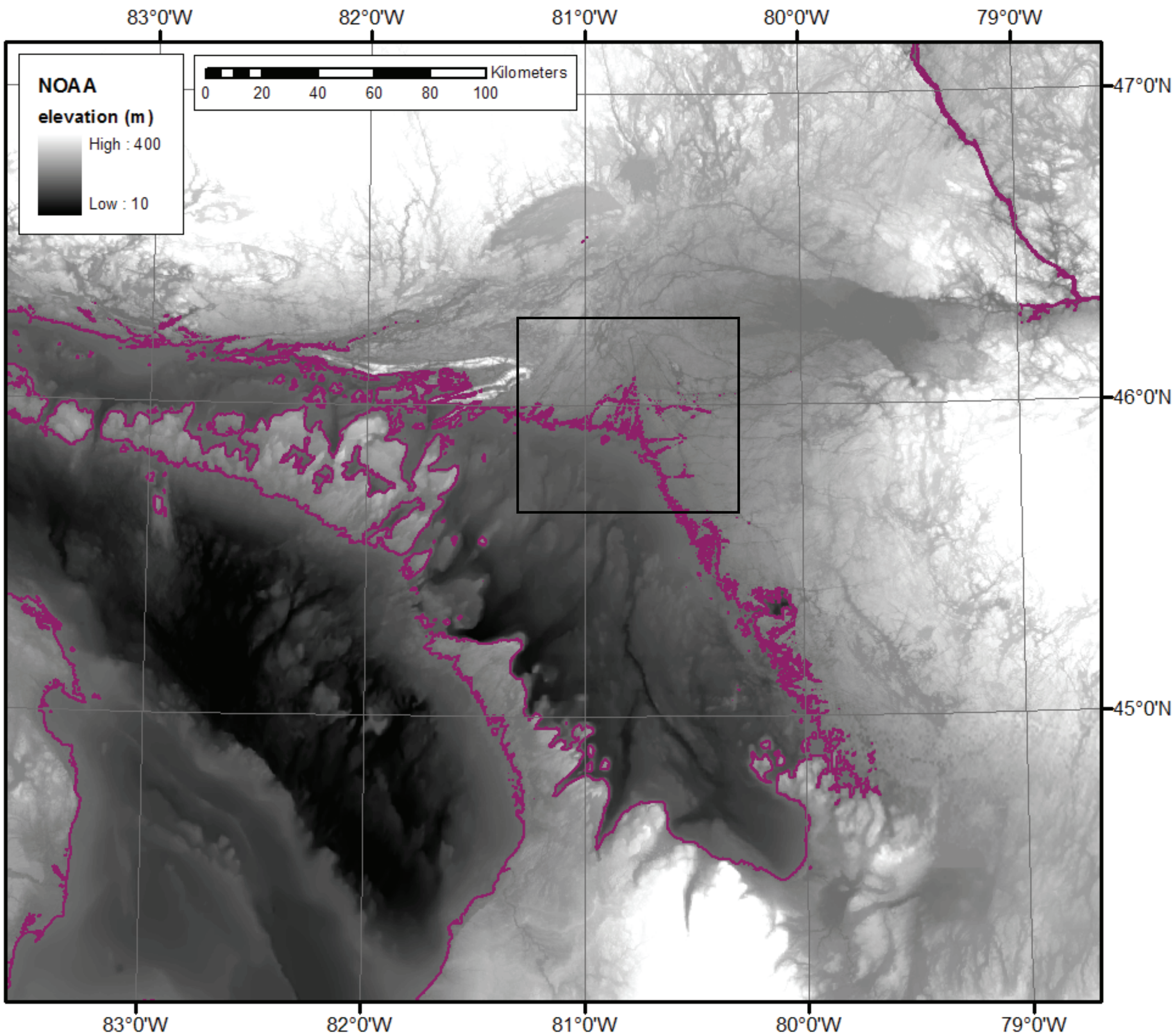
Notes: Flow direction from left to right in all but non-directional forms (from Kor et al. 1991).

Figure 4.1: Atlas of S-forms Identified and Sketched from the French River Complex

Importantly for this report, “the most compelling of all papers on the meltwater hypothesis”, Kor et al. (1991), arose from research around the French River area, on the northern shore of Georgian Bay and directly upglacier of the Bruce Peninsula. According to Shaw (2002), it “shows that the classification [of erosional features in bedrock] is workable; gives a comprehensive explanation of the erosional features as meltwater forms; and clearly demonstrates formation under a single, unidirectional sheet flow”. Kor and Cowell (1998) followed the direction of these flows southwestward across Georgian Bay and over Bruce Peninsula, where they found “spectacular suites of erosional marks indicating a continuous flow from the French River” over a vast region. Shaw (2002) writes “the meltwater event thought to have eroded the bedrock forms at French River and the Bruce Peninsula is on a spatial scale of drumlin fields”, that is on a scale of 1000 km^2.

These ideas are not accepted universally, however, and they have been challenged by a number of scientists, including prominent glacial geologists. Notably, Benn and Evans (2005) comments sharply that “a large number of papers have been published interpreting a wide range of subglacial landforms in North America as products of subglacial megafloods…The sheer volume of peer-reviewed publications promoting the ‘megaflood interpretation’…may lend it an aura of respectability in the eyes of those unfamiliar with the evidence. However, 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”. In this context, the superb field evidence for intense glacial erosion of bedrock by ice or subglacial waters in the region of the Bruce nuclear site merits special attention. Kor and Cowell (1998) reported convincing evidence for erosion by water, including potholes and sculpted forms (s-forms) in bedrock (Figure 4.1) over a broad range of spatial scales and well sorted deposits of cobbles and small boulders. Among the bedrock erosional features that are naturally exposed, they highlight the occurrence of highly elongated s-forms and flutes ranging in size from centimetres to tens or a hundred metres, and interpret these as the obvious products of erosion by highly turbulent subglacial flows sweeping to the southwest across the Bruce Peninsula. Although, at present, there is no consensus on the formative process producing highly elongated s-forms and flutes, there is no doubt that they reflect pervasive bedrock scouring under the LIS across the Bruce Peninsula.

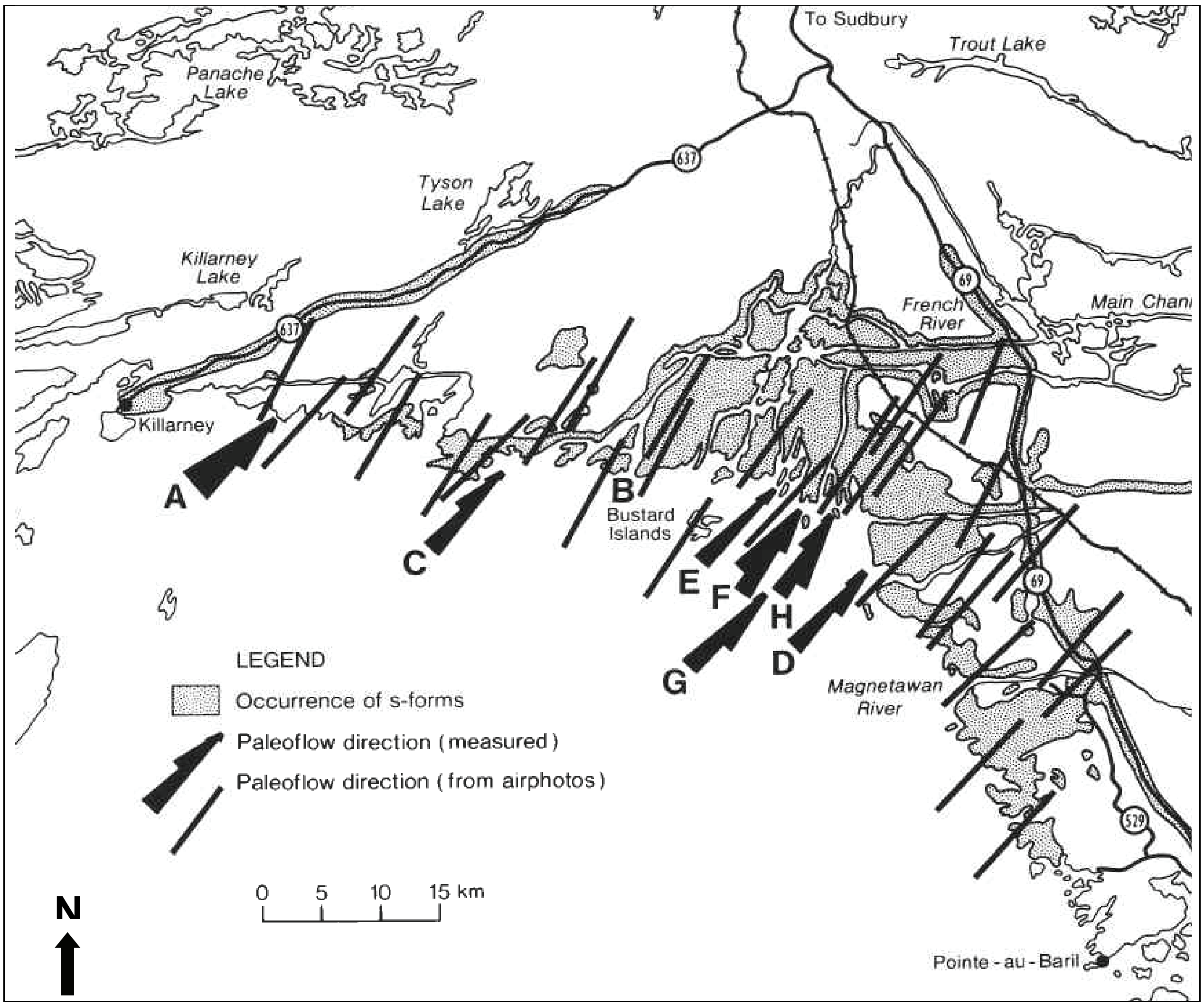
A critical issue for this report is the magnitude of the total erosion where the bedrock surface shows the spectacular elongated s-forms and flutes. Digital elevation models of the area are quite helpful in this regard. They show widespread stripping of the bedrock of the shield, and reveal a low-relief surface crisscrossed with shallow valleys and troughs (Figure 4.2). Surprisingly, these structurally or lithologically controlled depressions seem largely independent of the ice or water flow direction, which is exceptionally well defined. Kor et al. (1991) report a total of 178 flow indicators showing that the erosive flow across the whole study area, which exceeds 100 km in width, moved in a remarkably uniform direction to the southwest, with an average azimuth of 221° (Figure 4.3).



Notes: Gray tones show the elevation and depth of the Georgian Bay area. The purple line delineates the shorelines. The box shows the French River area, and approximates the location of the Kor et al. (1991) study area shown in Figure 4.3. This image reveals a maze of shallow channels incised into the Canadian Shield in the French River region and across the Bruce Peninsula with intervening slightly sinuous channels on the floor of Georgian Bay. The dearth of major channels or linear troughs parallel that parallel the ice flow direction to the SW suggests that the glacial removal of bedrock in the French River region was modest despite the bedrock surface displaying spectacular micro-morphology sculpted by ice and subglacial water.

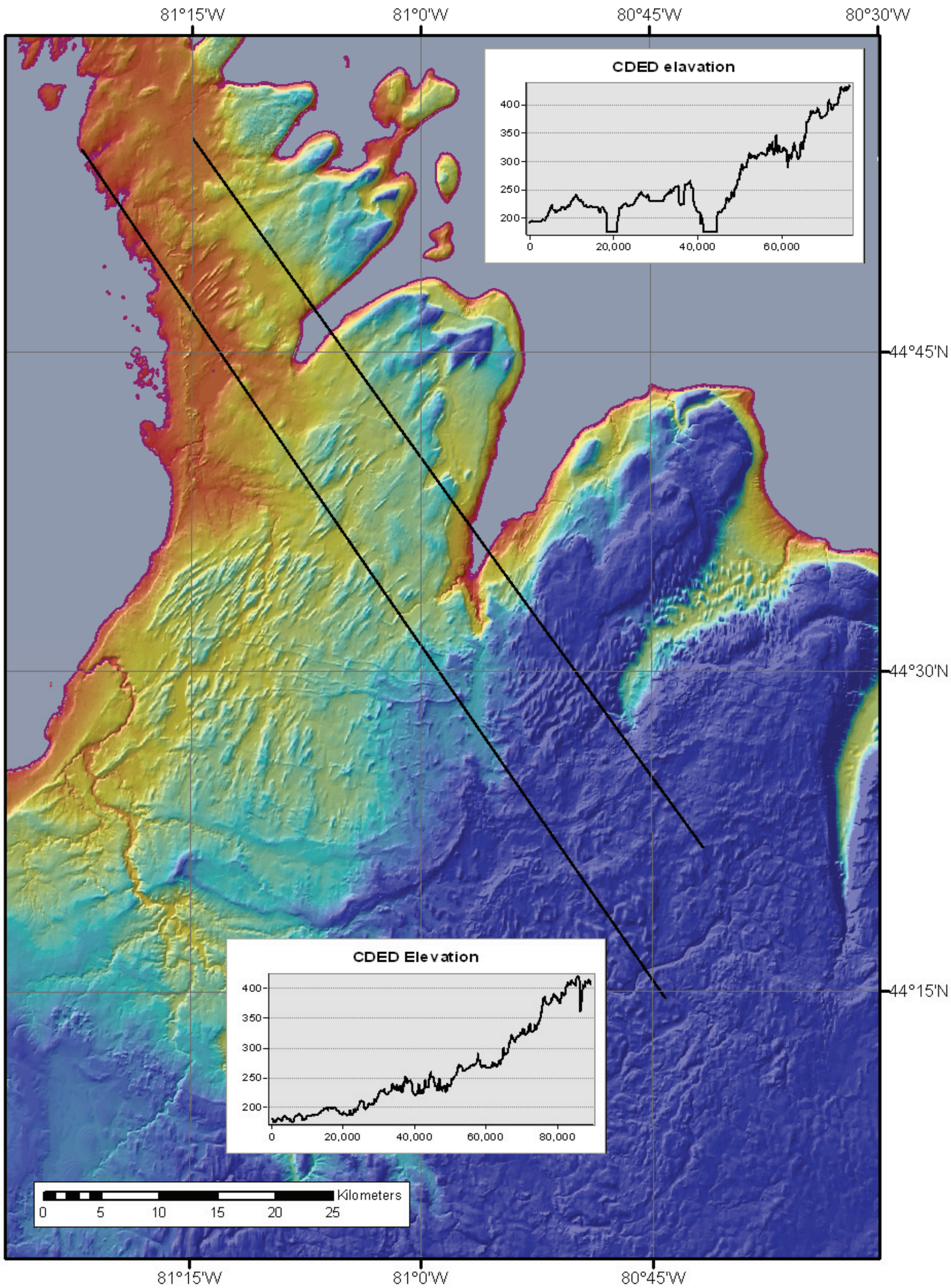
Figure 4.2: Topography and Bathymetry from the Georgian Bay Area

The lack of prominent topographic features (Figure 4.2) paralleling the southwest-trending flow features, Kor et al.'s (1991) s-forms and flutes, is noteworthy for this report. It suggests strongly that although large-scale subglacial flooding and/or sliding ice sculpted the bedrock surface distinctly, the total erosion of bedrock was limited otherwise pathways for the water would be incised more distinctly; any pathways exceeding perhaps ~10 m in depth would be clearly evident as channels oriented northeast-southwest on the Digital Elevation Model (DEM). The resistant shield rocks of the French River area undoubtedly contributed to this limited incision, but it appears that the total amount of erosion was also modest at sites to the south, where the sedimentary bedrock is likely much less resistant to glacial erosion. In particular, whereas erosional features on various scales on the Bruce Peninsula clearly reflect scouring by former ice and/or water flows to the southwest (Kor and Cowell 1998), DEM's show very little relief transverse to the former flow direction (Figure 4.4).



Notes: Figure is from Kor et al. (1991; their Figure 6) and shows observed distribution and orientation of s-forms, which all reflect former flow to the southwest. Mean directions by site location are as follows: site A: n = 20, mean = 223°; site C: n = 17, mean = 228°; site D: n = 22, mean = 223°; site E: n = 30, mean = 225°; site F: n = 54, mean = 221°; site G: n = 15, mean = 213°; site H: n = 20, mean = 223°; totals: n = 178, mean =221°. No measurements were taken at site B.

Figure 4.3: Observed Distribution and Orientation of S-forms, French River Area, Northeast Section of Georgian Bay



Notes: The plot shows flow-parallel rock-cored drumlins and large-scale lineations from northeast to southwest. Aside from the relatively high remnants of the Niagara Escarpment, shown in purple tones, the relief is very low, at most a few tens of metres over tens of kilometres roughly transverse to the flow direction, as seen in the two inset profiles (along trace of the northwest-northwest trending black lines) that show elevation (vertical axis) versus horizontal distance in metres.

Figure 4.4: Digital Elevation Model of the Northern Bruce Peninsula

4.2 Small-scale Features on Scoured Bedrock on Bruce Peninsula: Fresh Exposures

Fresh exposures of glacially abraded bedrock surfaces reveal considerable information about basal processes that is not available from natural exposures because most of the detailed surface features have been lost due to weathering in the thousands of years since they were exposed by ice retreat and emergence from postglacial lakes.

Extensive dolostone surfaces that were recently exposed from under a protective ~3 m-layer of lacustrine sediments in an active quarry, are very smooth and pervasively covered with striations (Figure 4.5). They show no hint of active erosional processes (quarrying and meltwater erosion) other than abrasion. Generally, all striations are parallel to one another and they were produced by the steady rectilinear motion to the southwest of ice-entrained clasts in sustained contact with the bed. Several observations suggest that relatively little bedrock was removed by the glacial scouring and that the contact forces between the bed and abrading clasts were relatively small.

1. Multiple basal ice flow directions are occasionally seen (Figure 4.5), indicating that the 1 to 10 mm of erosion required to erase older striations did not occur in the time span for a reorganization of the basal flow direction, which could be centuries or more according to the ice sheet model.
2. Many prominent striations are remarkably continuous and straight, showing that striating clasts were not rolling. This suggests, in turn, that contact forces between clasts and the bed were modest otherwise the corresponding frictional forces and resulting torques on the sliding clasts would cause them to roll. The lack of clast rotation also point to low shearing rates in the basal ice and, hence, relatively low basal shear stress and limited energy available to erode the bed.
3. Friction cracks and lunate fractures are exceptionally rare. As these require large local differential stresses to form, the observation that they are exceptionally rare also suggests that contact forces were unusually small and that large blocks were rare or rarely in contact with the bed.
4. Most striations are distinct but very shallow, indicating relatively light clast/bed contact.
5. Pervasively striated surfaces have a low relief (<1 m) similar to the underlying bedding planes, which would be surprising if much erosion had occurred.

Taken collectively this evidence demonstrates that, in the last phase of the glaciation at this site, debris firmly held in ice scoured the bedrock pervasively. The net erosion is most likely to have been slow and quite limited in magnitude. The quarry exposures suggest that erosion was rather uniform spatially for an area 100 m on a side or more, and this suggestion can be extended to much of the Bruce Peninsula because of its very low relief transverse to the flow direction (with the exception of the reentrants). The low transverse relief suggests that total erosion was modest; say tens of metres or less. This is because natural spatial variations in the conditions at the base of the LIS and in the bedrock characteristics that influence its resistance to erosion by ice or water would be expected lead to significant spatial differences in erosion rates and, hence, to substantial transverse relief if the differential erosion had been sustained long enough to erode a substantial thickness of bedrock, say more than a few, or a few tens of, metres.

The lack of obvious evidence of erosion by vigorous meltwater floods does not indicate whether or not they occurred. To be consistent with the quarry observations, however, their occurrence would have had to be limited to an early part of the LGM to allow for the surface to be pervasively striated. Additional evidence suggest that the ice was in close contact with the glacier bed late in the glaciation, as signs of dissolution of the dolostone at the lee of bed protuberances seem to reflect sustained, isolated water filled cavities of modest extent (~1 m^2).

Notes: Left: weathered flutes on naturally exposed surface near the northern tip of the peninsula. Center: freshly exposed limestone surface entirely covered with long, distinct striations, Adair quarry. Right: Freshly exposed striations to the southwest crossing an older deep striation. All photos are views down the former ice flow direction.

Figure 4.5: Photographs Showing Glacially Scoured Bedrock, Bruce Peninsula

4.3 Tunnel Valleys, Reentrants and Other Large-scale Topographic Features

Large-scale systems of tunnel valleys in southern Ontario have received considerable attention, and according to most of the recent work, they suggest widespread erosion by a “catastrophically released subglacial meltwater sheet”. For example, Shaw and Gilbert (1990) use the association of bedrock flutings, re-entrants in escarpments and tunnel valleys in southern Ontario and northern New York State to infer two major subglacial meltwater floods across the region. Along similar lines, Brennand and Shaw (1994) described three landforms in south-central Ontario – tunnel channels, mega-channels and late-stage sheet-flow scours – and interpreted them as evidence for the progressive channelization of meltwater during the collapse of a “catastrophically released subglacial meltwater sheet”. Although these water sheets likely scoured bedrock and sediment over vast areas, the total amount of erosion is not generally known. Tunnel valley depths reported by Russell et al. (2002), for example, can exceed 200 m in sediments; incision in bedrock is probably modest in comparison but does not seem to have been determined (see Section 2.3).

A large number of distinct reentrants in the Niagara Escarpment have attracted attention, and have been interpreted as demonstrating large-scale erosion by ice (Straw 1968) or by meltwater floods (Kor and Cowell 1998). The length of the reentrants, which reach over 16.5 km at Owen Sound, provides a measure of the differential erosion. This measure is not, however, a measure of the total erosion because slight differential erosion of the land surface in this region creates large reentrants due to the very shallow dip of the bedrock of the Escarpment. Using the 25 to 30 ft/mile (4.7 – 5.7 m/km) south-southwest dip reported by Straw (1968) and perfectly planar bedding, only 78 to 94 m of differential vertical erosion would produce Owen Sound, and an additional 128 m of erosion would be needed to account for the depth of the sound. Thus about 200 m of differential bedrock erosion through the Quaternary would suffice to create the largest reentrant, assuming no pre-glacial incision of the valley, which seems unlikely. If the whole Escarpment has retreated westward significantly, more erosion would be required; based on an estimate of 35 km between the tip of Owen Sound and the inferred pre-glacial position of the Escarpment (Straw 1968) and the depth of the sound, 300 to 330 m of erosion would be needed.

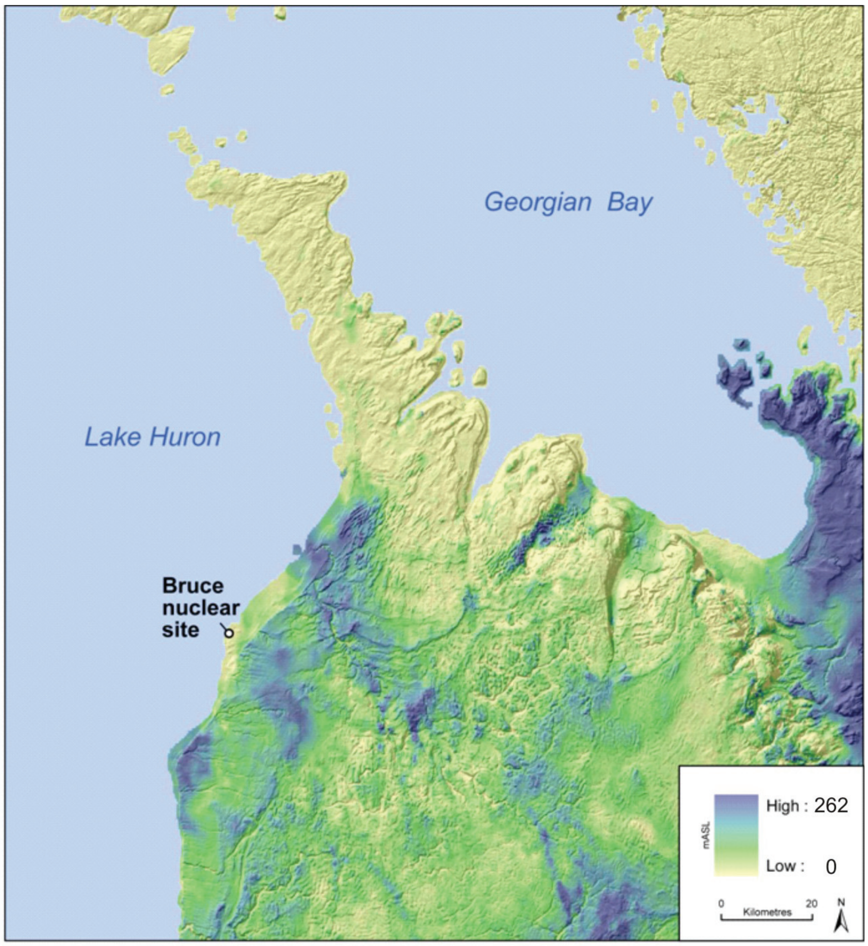
Several factors are likely to be responsible for the differential erosion that created the reentrant. Erosion was presumably localized because pre-existing topography steered the ice and subglacial water to these sites and/or the bedrock was more pervasively fractured in the reentrants. The latter possibilities could be addressed by examining the spatial variability of fracture density along the Bruce Peninsula. For example, Figure 4.6 shows a high fracture density at the head of one of the major reentrants relative to an interior site recently exposed by quarry activities. The lower fracture density could, of course, be in part due to the quarry exposure being more recent than the road cut.

Notes: Left: a few hundred metres North of Wiarton at the head of one of the major reentrants, Colpoy's Bay. Right, very few fractures are visible in the upper ~2 m of dolostone being sawed at the quarry a few kilometres southwest of the head of Hope Bay.

Figure 4.6: Photographs Showing Variability of Near Surface Bedrock Fracture Density

4.4 Distribution of Sediment Cover over Bedrock

Because of the potential for till and other sediments to effectively shield the bedrock from erosion, especially if this sediment cover is not shearing appreciably, it is important to consider the occurrence of the sediment cover and how it generally behaved under the LIS. Aside from the entrainment of single mineral or rock particles by sliding ice, sediments at the bed of a glacier can be mobilized as they are sheared by ice motion or entrained by subglacial waters. Although subglacial shearing of sediments has been discussed extensively and inferred to occur widely (e.g., Alley et al. 1986, Alley et al. 1989, Engelhardt et al. 1990, Humphrey et al. 1993), this process has been documented only in very few sites, primarily under thin ice along the margin of Icelandic glaciers (Boulton 1979, Boulton and Hindmarsh, 1987). The universality and importance of till shearing in contributing to glacial motion has been questioned more recently (e.g., Piotrowski et al. 2001, Hooyer et al. 2008) but, despite promising recent advances including those resulting from the experimental and field study of the magnetic fabric of sheared till (Hooyer et al. 2008), the questions remain largely unanswered as little information is available regarding the magnitude of the shear deformation in subglacial debris, and the thickness of the shearing debris. Nevertheless, soft-bed deformation is generally incorporated as a mechanism to accelerate basal motion in models of the LIS (e.g., Peltier 2011, Jenson et al. 1995, 1996) and sediment transport (Boulton 1996a).

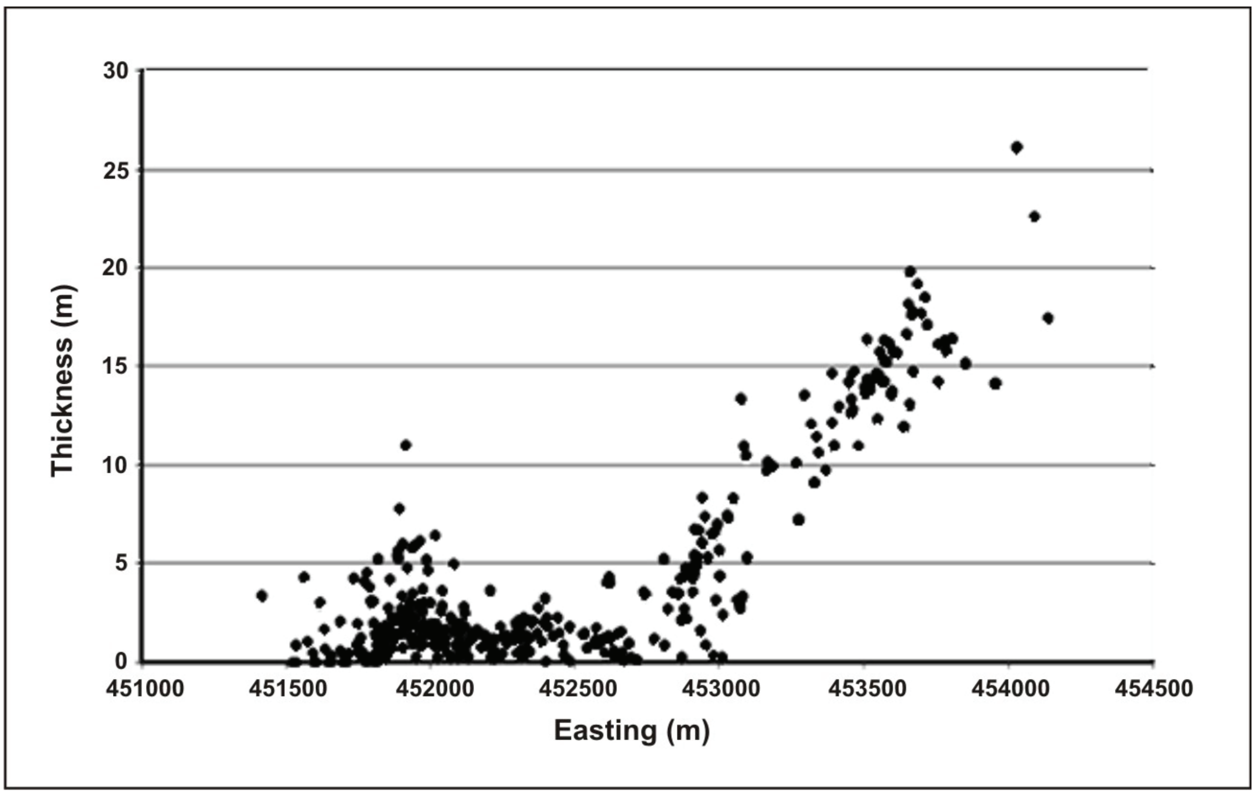


Note: Image taken from Google Earth image with Ontario Geological Survey Data.

Figure 4.7: Overburden Thickness (in metres) in the Region of the Bruce Nuclear Site

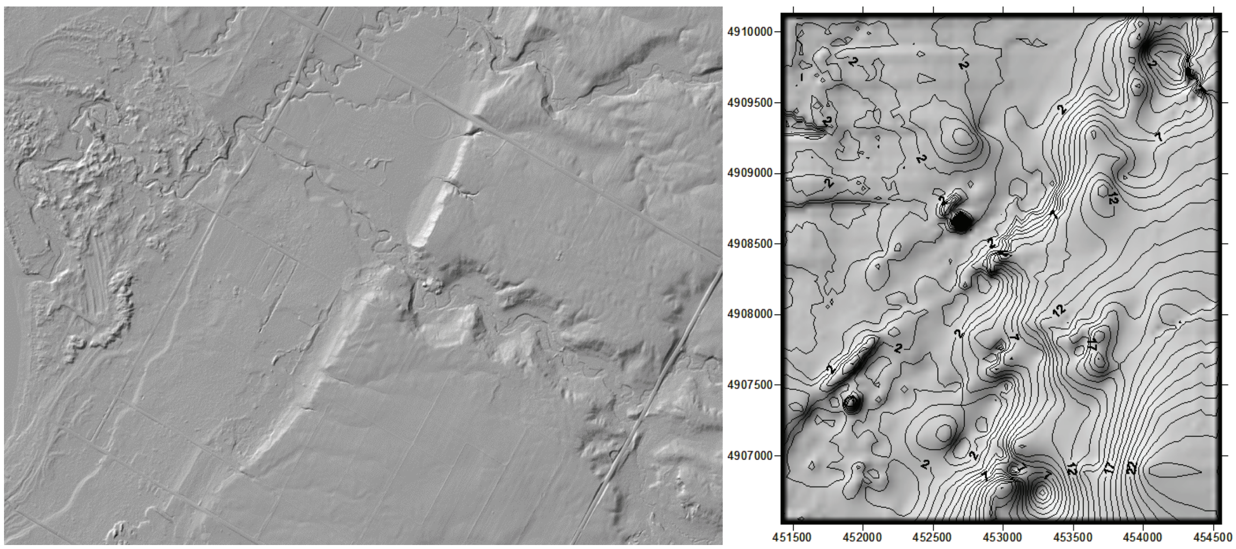
The shielding effects of a sediment cover on bedrock is widely recognized, and has direct relevance for glacial erosion at the Bruce nuclear site because of the thick overburden found in the immediate vicinity (Figures 4.7, 4.8 and 4.9). The effectiveness of this shielding is difficult to assess quantitatively. In the model described in Section 3.3 of erosion by the LIS, Hildes et al. (2004) represent the shielding with a factor multiplying the abrasion rate for bare bedrock that decreases exponentially with increasing sediment thickness. With the simple parameterization in the model and the choice of parameters representing this shielding (their Table 5), relative to erosion rates for bare bedrock, erosion slows significantly with a few metres of cover and vanishes with tens of metres of cover. The depth and amount of shearing in till are poorly known (e.g., Piotrowski et al. 2001, Hooyer et al. 2008), however, and likely depend sensitively on subglacial conditions and till rheology. If significant shearing occurs only very locally, say within a layer of sediment 0.1 to 1 m thick directly under the sole of the ice sheet, as has been reported for a contemporary ice stream (Piotrowski et al. 2001), the bedrock below that layer would be entirely shielded from erosion.

In the vicinity of the Bruce nuclear site, the overburden thins westward from 10 to 20 m to near zero at the shoreline (Figure 4.8), and most of the drop in thickness occurs along a northeast-southwest trend that corresponds closely to the Algonquian Bluff, which marks a former lake shoreline that is distinct on the LiDAR DEM (Figure 4.9). It is likely, therefore, that 10 to 20 m of drift separated the ice from the bedrock, at least near the end of the glaciation, and that well after the ice pulled back, energetic wave activity in Lake Huron washed away much of the drift that mantled the bedrock. Hence, the bedrock at the Bruce nuclear site was shielded from active glacial erosion during part of the last glaciation.



Note: Overburden thickness values are from the region represented by the DEM shown in Figure 4.9 (left). They are also shown in map view in Figure 4.9 (right).

Figure 4.8: West to East Profile of Overburden Thickness at the Bruce Nuclear Site



Notes: Left image is a LiDAR-generated shaded relief DEM of part of the Bruce nuclear site. Note the prominent scarp (illuminated from the west), the Algonquin Bluff, which marks a former post-glacial lake shoreline. Right image is a contour map of the overburden thickness (in metres).

Figure 4.9: Digital Elevation Model and Overburden Thickness at the Bruce Nuclear Site

The alignment of small scales erosional features on bedrock (striations, lineations, etc.) show that the broader study region experienced strong northeast to southwest flow of ice and water. Nevertheless, the slight channelized erosion evident in the bathymetry of the Georgian Bay area and the low relief of the surrounding terrain, including the Bruce Peninsula, suggest strongly that the amount of bedrock erosion was modest, on the order of tens of metres or less. Possible factors limiting the amount of erosion in study region include: 1) limited duration of ice occupation over this site due to its position close to the margin of LIS; 2) the significant sedimentary cover or glacial drift that served to protect the underlying bedrock from excavation by the ice flow; 3) the relative resistance of the local bedrock to glacial erosion; and 4) the lack of favorably-oriented major valleys that could localize ice flow and erosion.

5. NUMERICAL ESTIMATES OF GLACIAL EROSION AT THE BRUCE NUCLEAR SITE

5.1 The University of Toronto Glacial Systems Model

The University of Toronto Glacial Systems Model (UofT GSM) provides a rich quantitative framework for assessing glacial erosion rates for the site of interest, as it addresses a number of variables that bear directly on glacial erosion. These include basal temperature, basal ice speed due to both till deformation and sliding, basal shear stress, and rate of meltwater production due to viscous and frictional dissipation at the glacier bed. The latter is of particular interest because it permits calculation of erosion rates using empirical data relating the erosion rate to the rate of energy dissipation at the glacier bed.

Although a number of ice sheet models exist and are under development (e.g., Zweck and Huybrechts 2005, Bueler and Brown 2009), the UofT GSM is the natural one to use for assessing the maximum erosion over the Bruce Peninsula. According to Peltier (2011) "The model…has been under continuous development for the past two decades…[It] is unique in the research literature of this field for several reasons, the most important of which for present purposes is that it is equipped with the numerical apparatus required to calibrate it against a wide range of observational constraints using a statistically powerful Bayesian methodology. This model has its origins in the early work of Deblonde and Peltier (1991, 1993), Deblonde et al. (1993) and Tarasov and Peltier (1997, 1999). The most recent developments of this structure, as described in Tarasov and Peltier (2006, 2007), are highly relevant to the applications to be described in this Report, especially that by Tarasov and Peltier (2007) in which a detailed discussion of improvement to the methods employed to compute the evolution of permafrost extent and depth during a glacial cycle is provided."

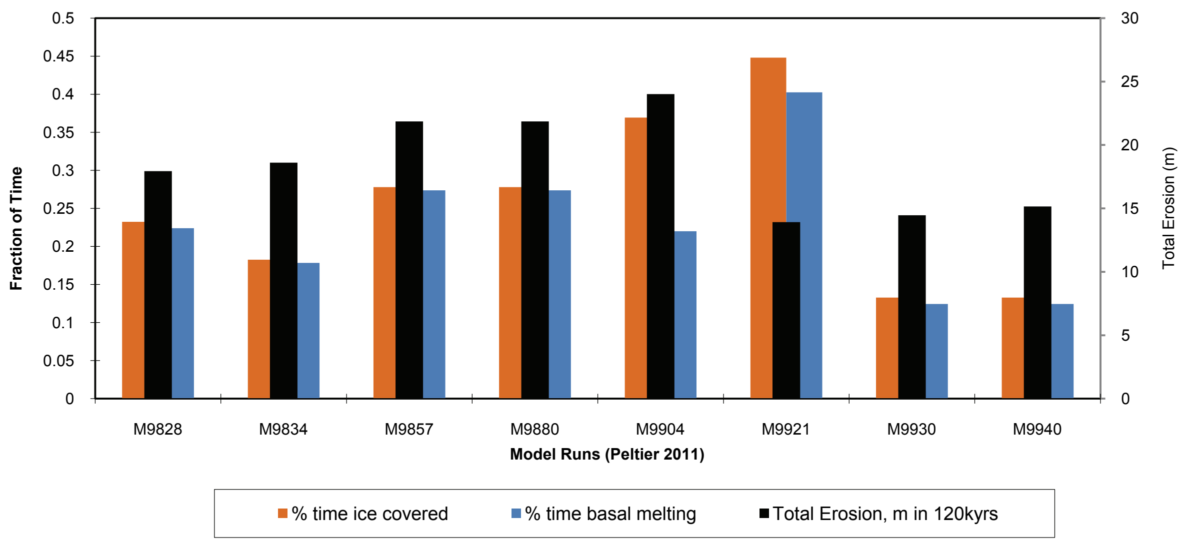
Before using this model to calculate erosion magnitudes and rates, it is important to note that this and other ice sheet models have received a lot of attention; a number of significant accomplishments and limitations have been recognized. For an example, the European Ice Sheet Modeling Initiative (EISMINT) reports the inter-comparison of ten operational ice sheet models, including the UofT GSM, and shows that the models are quite compatible in terms of producing similar ice sheet profiles and volumes (Payne et al. 2000). Considerable variation in the treatment of basal processes exists between models and, hence, in their ability to generate fast flow and to explain low-sloping ice sheet profiles inferred from isostatic and geological evidence. Enhanced basal sliding or deformation of soft sediments is often invoked as important sources of uncertainty in modeling the northern hemisphere ice-sheets and as potential mechanisms of instability (Clark 1994, Marshall et al. 2000). Related instability mechanisms, which draw the ice-sheet margins down towards the end of a glacial cycle, have been discussed in connection with isostatic adjustments to the time-varying ice loading (Zweck and Huybrechts 2005). Large uncertainties are also associated with the incorporation of previously unrecognized processes. The latter includes the penetration of surface meltwater to the base of an ice-sheet through ~1 km of cold ice, which creates a mechanism for the rapid basal motion and response of ice flow to climate change (Das et al. 2008).

Acknowledging the limitations of the ice-sheet modeling and the rapid ongoing advances in this field (e.g., Bueler and Brown 2009), the UofT GSM can be used with considerable confidence because 1) it has been particularly well documented and customized to be consistent with a diversity of regional data (Tarasov and Peltier 1997, 1999, 2006, 2007); and 2) because it gains much independent support for use with a focus on the Bruce Peninsula from substantial data sets in the Great Lakes region. For example, Braun et al. (2008) employ only subsets of the data from 55 tide-gauge and 70 GPS sites available, and compare the data "to predictions of

70 GIA models by combining three ice-load histories (ICE-3G, 4G and 5G [from 3 generations of the UofT GSM]) with a broad range of 1D and 3D viscosity models." The authors show that the data are well satisfied using the UofT GSM, especially the ICE-3G model. More generally, the UofT GSM has figured prominently over the last decade in "important advancements in the resolution and accuracy of forward modeling of GIA (glacial isostatic adjustment)…derived from the advanced observing systems onboard ongoing satellite missions (e.g., CHAMP, GRACE) and, more recently, from progress in realizing a geodetically stable reference frame (ITRF2005), from which GPS observations…, a variety of T/P-altimetry, tide-gauge, terrestrial-gravity and GPS data combinations take advantage" (Ivins and Wolf 2008).

Herein, erosion rates will be calculated using specific UofT GSM results that: 1) define the duration of glacial cover and temperate basal conditions; and 2) the rate of basal melting, hence basal energy dissipation, which is expected to control erosion rates. This and other models will also provide indications of the likelihood of channeling flow and erosion over the Bruce Peninsula. Erosion rates will be compared to those calculated by Hildes et al. (2004).

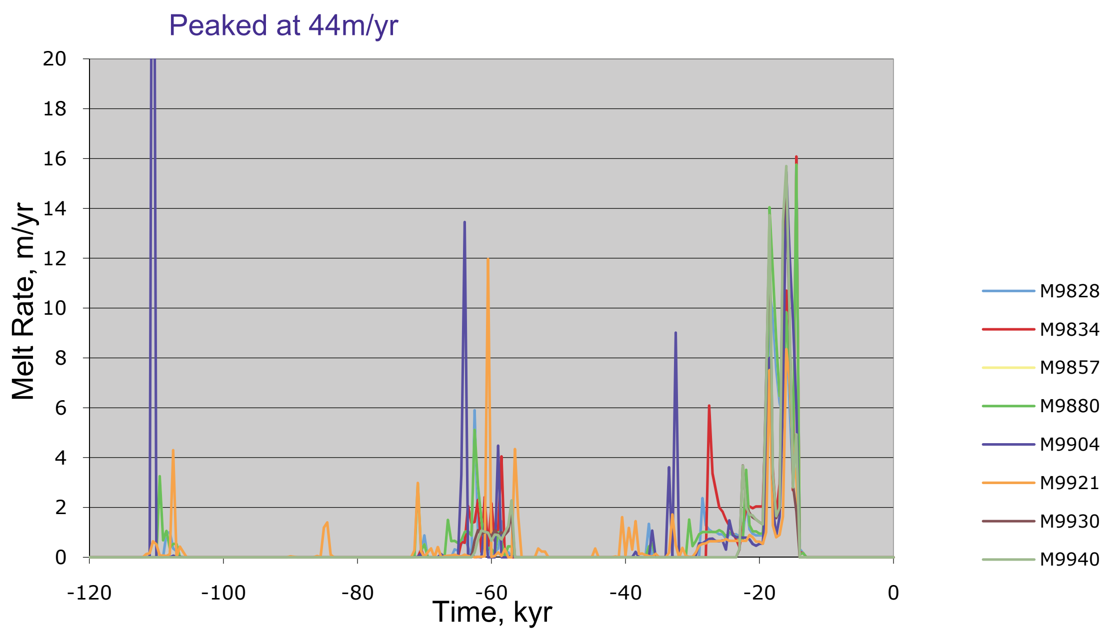
Figure 5.1 shows that for most of the LGM the LIS did not cover the Bruce nuclear site. On average the site was covered by ice only about a quarter of the time, which is consistent with independent results for another LIS model (Marshall et al. 2002, their Figure 7); for the whole series of runs, this varied from about 13% to 45%. For most runs, the base of the ice was at the pressure meting point when it covered the Bruce nuclear site, but for runs M9904 and M9921 (Peltier 2011) the base of the ice was frozen to the bed for an appreciable fraction of the time the LIS covered the site.



Notes: Left vertical axis shows the fraction of time that the Bruce nuclear site is covered with ice (orange bars), and covered with basal ice at the melting point (blue bars) for each of 8 UofT GSM model runs by Peltier (2011). The right axis is the total cumulative glacial erosion in 120,000 years calculated using the melting rates for the model runs (black bars); see text for discussion.

Figure 5.1: Fraction of Time the Bruce Peninsula Is Covered with Ice and Is at the Melting Point

Figure 5.2 illustrates the calculated basal melting rates as a function of time for the LGM, starting 120,000 years ago for the whole series of runs. It also shows when temperate ice covered the Bruce nuclear site as melting is only possible under these conditions. It is noteworthy that the model produces a complex history of LIS fluctuations including relatively short-lived events that invite comparison with the geological records for the study area. Taking a closer look at the period 10,000 to 30,000 years ago (Figure 5.3), most of the model runs, including run M9930 (shown as heavy brown line) that tends to outperform the others (Peltier 2011), suggest that two major advances of the LIS with rapid basal motion occurred, about 16,000 and 18,500 years ago, although the model chronology may differ considerably with data at particular sites, especially close to the margins of the LIS where distinct lobes had their own fluctuations. The most recent advance and the preceding retreat could well correspond to the local stratigraphy, which is well exposed a few kilometres east of the Bruce nuclear site (Figure 5.4), with a thick till unit, presumably the St. Joseph Till, overlying stratified sands and gravels. This outwash may well have formed between the two advances, during the Mackinan interstadial, either subaerially or subglacially. A lower till, the Catfish Creek Till or its equivalent, was not exposed at this site but it is widespread in the area.

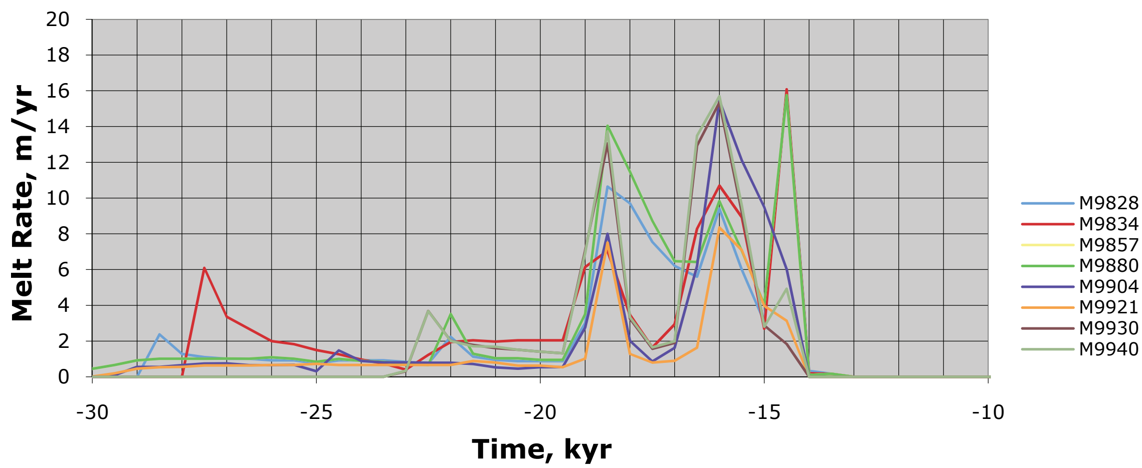


Notes: Basal melting rates calculated for each of 8 UofT GSM model runs by Peltier (2011). Rate is proportional to the product of the velocity and shear stress at the base of the LIS (Peltier 2011).

Figure 5.2: Basal Melting Rate for the Bruce Nuclear Site

These model results can be used to estimate total erosion in a number of ways. The simplest and, perhaps, most conservative estimate is that, whenever the LIS covered the Bruce nuclear site and its base was at the pressure melting point, it was eroding bedrock at the highest long-term rate known worldwide, which is estimated to be 5 mm/yr (Figure 2.2). If we consider all the model runs shown in Figure 5.2, on average the LIS was eroding 23% of the time during the last 120,000 years (maximum, 40%, and minimum, 12%). Provided it was eroding at the maximum known long-term rate, 5 mm/yr, bedrock erosion over 100,000 years would total

114 m (maximum, 201 m, and minimum, 62 m). This estimate of erosion is inherently flawed for a number of reasons; including the unreasonable assumption that erosion at the study site occurs at a rate equal to the fastest long-term glacial erosion known from any part of the world, which is in the St. Elias range of Alaska, an active compressional orogen where the tectonic, climatic and glacial circumstances conspire to produce and sustain this rapid erosion.



Notes: This plot reveals the ability of the models to simulate two major advances between 14,000 and 30,000 years ago and other short-lived LIS fluctuations. Basal melting rate is from Peltier (2011).

Figure 5.3: Basal Melting Rate for the Bruce Nuclear Site from 10,000 to 30,000 years ago

Because all of these circumstances contrast with those in southern Ontario, it is critical to take into account the site-specific factors that most directly affect glacial erosion rates. Basal velocities and melting rates derived from the UofT GSM, are extremely helpful in this regard. The melting rate permits calculation of the average power dissipation at the bed for each model run. Combining these empirical results summarized in Section 4.2 with the estimates of basal power derived from the UofT GSM (Peltier 2011) produces a realistic estimate of the total potential erosion of bedrock over 100,000 years: the average for all runs is 14 m (maximum: 33 m, minimum: 2 m; Figure 7.1).

It is important to note, however, that these calculations of bedrock erosion greatly overestimate the likely erosion at the Bruce nuclear site, hence the reference to “potential” erosion above. First, both calculations assume that the sliding ice has direct access to the bed, whereas the bedrock was protected by a thick layer of sediment at least part of the time during which little or no erosion would have occurred. Second, the more realistic erosion calculation is based on a substantial overestimate of the rate of energy expenditure at the bedrock surface. This rate of energy expenditure is derived from the melting rates calculated by the UofT GSM as the product of the basal velocity and shear stress, but because basal motion in the model arises primarily from soft-bed deformation and secondarily from sliding over bedrock, the rate of energy expenditure at the bedrock surface is only a small fraction of the basal energy dissipation that is used to calculate melting rates. Both of these issues, the protective effect of the sediments and the energy dissipation at the bedrock surface, are addressed next.

Note: Exposure is a few kilometres east of the Bruce nuclear site (S. Davies, for scale, is 1.85 m tall).

Figure 5.4: Quarry Exposure of Massive Till above Coarse Glacial Outwash

The first effect, the duration of protection of the bedrock from erosion by a sediment layer, is important but difficult to quantify. The model results for the period 10,000 to 30,000 years ago, together with the stratigraphy, illustrate the issue well. Assuming the outwash shown in Figure 5.4 was deposited during the interstadial, around 17,500 or 18,000 years ago, and noting that the intact primary bedding in the outwash precludes appreciable shearing in the outwash, erosion of bedrock by the LIS at this site would have stopped 17,500 or 18,000 years ago or earlier if older sediments covered the bedrock. Moreover, shortly after that period over 5 m of till were deposited over this site. Assuming that the sediment cover prevents all bedrock erosion after 18,000 years ago, the previously calculated total expected erosion would decrease to an average of 39% (maximum, 46%, and minimum, 32%) of the total erosion for the period 10,000 to 30,000 years ago, which was determined without taking into account the shielding of the bed. For the entire 120,000 year period, the total erosion is still impacted significantly by the 4,000 years of shielding; it is reduced to 54% (maximum, 70%, and minimum, 46%) of the total erosion previously calculated. The revised estimate of the total erosion over 100,000 years is now 8 m (maximum: 22 m, and minimum: 1 m; Figure 7.1). It is important to note that this total erosion should be recognized as an upper limit because the bedrock could have been shielded from erosion for the entire period, resulting in zero erosion.

The second effect can also impact estimates of erosion substantially. For very soft beds, that is a sediment substrate that shears easily, most of the computed basal motion is due to bed deformation, and hence, most of the energy at the bed of the glacier is dissipated in the sediment, leaving a small fraction of that energy to fuel glacier sliding, to move rocks in frictional contact with the bedrock, and to erode the bedrock where ice has access to the bed on a subgrid scale. The fraction of the total basal energy that is dissipated at the bedrock surface per unit time can be assessed, knowing how the basal motion was computed in the UofT GSM (Tarasov and Peltier 2004, Peltier 2011) as the sum of the bed deformation and sliding. This fraction increases roughly linearly with the effective viscosity of the sediment layer, μ, from 0.5% to 35% of the total basal energy for the approximate range of viscosities, 1.1×10^9 to 45×10^9 Pa s, used by Peltier (2011). Based on these results, the erosion is reduced to 35% of the values presented in the previous paragraph, yielding a total erosion of 2.7 m over 100,000 years (maximum: 7.7 m, minimum: 0.4 m; Figure 7.1). A much further reduction could be well justified but is unwarranted, at this stage, in view of the aims of this report, which focus attention on the highest likely rates of erosion.

5.2 Limitations of Models of Erosion

Because of the considerable simplifications made in theoretical models of erosion mechanisms, and the extreme sensitivity of the fracture process to poorly known basal characteristics (the magnitude of the differential stresses on a bed of poorly known geometry, as well as its variation in time and space, and the characteristics of preexisting cracks in the bedrock), the models are not well suited for calculating absolute values of erosion rates. A further limitation of these studies is that they do not address all principal aspects of glacial erosion. Notably, erosion by subglacial meltwater under normal conditions of high discharge during the melt season, as well as during extreme outburst floods, has not been examined quantitatively and simulated numerically. Even the flow of subglacial water, without consideration of bedrock erosion and sediment transport, is seldom treated explicitly in ice-sheet models; the rare exceptions include the glacial hydrology models of Flowers (2000) and Flowers and Clarke (2002) and the recent models of meltwater recharge of subglacial aquifers and flow along the bed toward the glacial margin (Lemieux et al. 2008a, 2008b, 2008c). Another limitation of erosion models is that they have yet to incorporate the effects of important recent findings of high amplitude, transient stress singularities occurring at the glacier bed. Notably, Kavanaugh (2009) recorded >7000 subglacial water pressure pulses, with magnitudes reaching nearly 3 times the pressure required to float the glacier during a 231 day period. He suggests, “these hydraulic transients, when coupled to the high-magnitude mechanical stress transients associated with these events, could play an important role in the cracking and erosion of subglacial bedrock.” Indeed, subglacial hydraulic transients figure prominently in theoretical work on glacial quarrying (Iverson 1991, Hallet 1996).

These cautionary words pertain to all models of erosion, including the most current model of glacial erosion on the scale of the LIS by Hildes et al. (2004), which was introduced in Section 3.3. Hence, rather than relying heavily on mechanistic models of glacial erosion, the estimates of erosion for the Bruce nuclear site, presented in Section 5.1, rely primarily on the UofT GSM in combination with empirical results from studies of glacial erosion rates on a basin scale, maximum known amounts of erosion in various areas, and glacial geologic observations in the region of the Bruce nuclear site. These estimates are also guided by insights into glacial erosion processes and basal processes.

6. OTHER SITE-SPECIFIC TRANSIENT PHENOMENA

Ice streams are exceedingly important in discussions of ice-sheet volume and sea level change because they modulate, to a large extent, the delivery of ice from the major ice-sheets to the oceans. They remain poorly understood, however, and constitute major challenges for ice-sheet models, including the UofT GSM. The occurrence and location of ice streams can vary greatly with runs from the UofT GSM having slightly different choices of model parameters, and hence are inherently not predictable.

Although they are a major concern in terms of modeling the dynamics of ice discharge and sea level change ice streams have relatively little potential to cause rapid or deep erosion. As discussed at the end of Section 3.2, the basal conditions that decrease resistance to ice motion (Figure 3.2) and make the rapid flow of ice streams possible (Figure 3.1), may inhibit or preclude active erosion because, under these conditions, the bedrock is largely decoupled from the moving ice by an intervening soft sediment layer or discontinuous layer of pressurized water. In other words, the general inability of ice-sheet models to accurately predict the regions in which ice streams may form in regions of low relief does not undermine the general conclusion that only modest glacial erosion can be expected to occur under these circumstances at the Bruce nuclear site. The reason for this is that fast moving ice-streams are of necessity very weakly coupled to the bed and therefore are ineffective “eroders” of the bedrock substrate. For mountainous areas or other regions of considerable relief, topographic steering could be very effective in localizing ice streams generally over major valleys or depressions such as the Great Lakes; such steering would naturally lead to relatively fast ice motion over the depressions and relatively slow motion over topographic highs, such as the Bruce Peninsula.

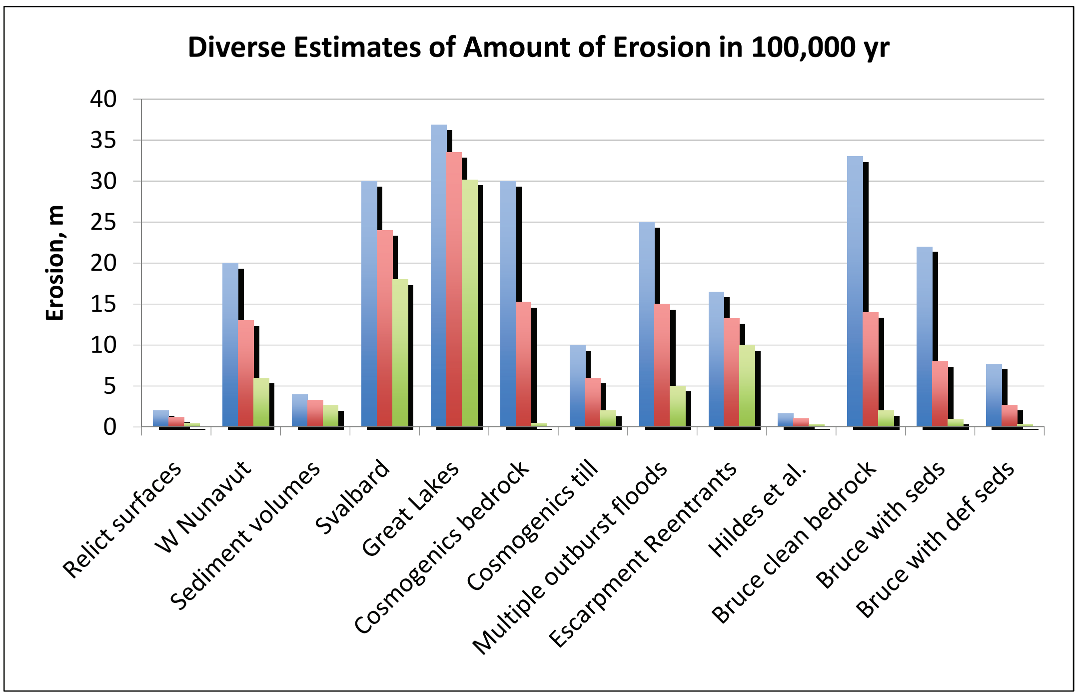
Water in moulins (narrow, tubular chutes through which water enters a glacier from the surface) and other glacial channels that reach the bed can cause rapid, highly localized erosion, in the form of potholes reaching depths on the order of ~10 to 20 m (Figure 6.1). They are not likely to cause large-scale excavations of the glacier bed, because they act so locally (usually <10 m), and many may not reach the bottom of the glacier as energetic flows. Moreover, they move with the ice, and hence cannot sustain erosion over any portion of the bed. In alpine settings or other areas of high relief, crevasses that are caused by ice flow over major irregularities of the bedrock topography can localize water flow in the same region of the bed year after year where crevasses continue to form at the same location, permitting water to cascades toward the bed. Such circumstances, however, are unlikely with thick ice moving over a substrate with little relief.

Note: Exceptionally deep and numerous potholes occur in Interstate State Park, Minnesota where the glacial St. Croix River rushed through the area, forming a series of potholes.

Figure 6.1: Photograph of the Bottomless Pit, a Glacial Pothole

7. SUMMARY

Thirteen estimates of total erosion (Figure 7.1) emanating from decades of research in various disciplines by diverse earth scientists support the conclusion that erosion would not exceed many tens of metres during the next glacial cycle to occur at the Bruce nuclear site. It is noteworthy that eleven of these estimates are independent of one another. The four values on the right in Figure 7.1 are model estimates: the first from the Hildes et al. (2004) model for the entire LIS, and the three others were calculated herein from results of eight representative model runs of the University of Toronto Glacial Systems Model (UofT GSM) of Peltier (2011), interpolated to the Bruce nuclear site.



Notes: Diverse estimates of bedrock erosion indicate that during one glacial cycle, on a 100,000-year time scale, the total erosion at the Bruce nuclear site is likely to be less than 35 m. The blue, red and green bars represent maximum, average, and minimum erosion estimates, respectively. From left to right, the first four results are from geological and geochemical studies in Chapter 1; the multiple outburst flood value is derived for the catastrophic Lake Missoula flood in Section 2.4; the Escarpment reentrant value is derived in Section 4.3. The remaining estimates, described in Chapter 5, are derived from ice-sheet models: a value averaged over the entire North American ice-sheet by Hildes et al. (2004) in Section 3.3 and three values reflecting different assumptions in using results of an ensemble of 8 representative runs of the UofT GSM (Peltier 2011).

Figure 7.1: Estimates of Bedrock Erosion Amounts over One Glacial Cycle

The estimates are quite distinct from one another and caution is in order when presenting them in one figure because they are not strictly comparable. For example, whereas the sediment volumes approach (third entry from left in Figure 7.1) provides a measure of erosion by all glacial erosional processes over continental scale regions at a pace characteristic of the last 2 Myr, a study of cosmogenic nuclides on striated bedrock (sixth entry from left in Figure 7.1) addresses only abrasion and does so on a bedrock surface a few square metres in extent for a period of time that is a hundred times shorter. Numerical results for the Bruce nuclear site estimate total erosion over this time span to range from 33 m to 1 m. It is very likely that erosion will be towards the lower end of this range because: 1) bedrock would be shielded from erosion by a sediment cover for much of the time; and 2) erosion rates were intentionally overestimated by assuming that rapid basal motion was due entirely to fast sliding rather than sediment deformation.

The deep excavation of the Great Lakes, especially Lake Superior, suggests that over the Quaternary erosion of bedrock by the Laurentide Ice-sheet (LIS) has exceeded 600 m, at least in one particular area. This obviously raises the question of whether similar glacial excavation could occur at the Bruce nuclear site. The answer to this critical question is no, or extremely unlikely. Confidence in this answer is considerable, because the erosion at Lake Superior, which is exceptionally deep for the Great Lakes region, must reflect singular conditions that conspired, and will in the future conspire, to focus erosion in that particular area. They include highly erodible bedrock, and continuous steering of each new advance of the LIS into this ever deepening depression due to preferential glacial erosion. The natural steering of actively flowing ice lobes into basins and valleys that are generally aligned with the ice flow direction has the opposite effect on highlands such as Bruce Peninsula; they erode relatively slowly and often become sites of deposition where the ice tends to slow down.

These various factors have long been recognized. For example, quoting Soller and Packard (1998): “For many areas, it is likely that once ice lobation had become established, the ice lobes and interlobate areas recurred at the same general positions in successive glaciations, causing a gradual buildup of sediment volume in the interlobate areas…Through successive glaciations, bedrock topographic highs separating adjacent ice lobes received additional sediment from the lateral margins of each lobe, adding to the overall thickness of sediment in the interlobate area and further establishing topographic control on ice movement.” Hence, future glaciations would tend to repeat the actions of their predecessors, by funneling into the lake basins and deepening them until the glacio-hydraulic conditions starts to trap the sediments, and maintaining the topographic highs where the ice flow velocities and erosion rates are relatively low, and where sediments tend to accumulate. The thick mantle of glacial outwash and till that was exposed by the retreating ice at the Bruce nuclear site show this to be a site of net deposition, like much of the terrain except for the Great Lakes themselves; it may have been eroded only slightly or not at all by the LIS.

The estimates summarized in Figure 7.1, are largely derived from state-of-the-art studies, high-resolution photographs and DEMs, and sophisticated thermo-mechanical ice-sheet models. They are not novel, however. Rather, they echo classic studies of the region including Baker (1916), who remarked, “Glaciation, of course, eroded both the Precambrian and the later rocks, and has served merely to freshen the existing topography of the country but has not produced it”. Regarding the Bruce nuclear site more specifically, Collins (1925) described the relief in the Manitoulin area and eastward on the north shore: “Its individual hill (of Lorraine quartzite) which have stood since before Paleozoic time, have gently sloping sides and rounded tops. Seen across the archipelago that fringes the north coast of Lake Huron this ancient range of snow white hills forms the dominating feature in what is perhaps the most picturesque part of Ontario”. Interestingly, the eastern edge of Collins’ (1925) study area bordered on French River, the center of the pathways of inferred catastrophic floods (Kor et al. 1991) that swept across that area and scoured the Bruce Peninsula. Evidently catastrophic floods may have raced across that landscape, but if they did, the traces they left were subtle (Figure 4.2), suggesting modest overall removal of bedrock in recent (Quaternary) time (Collins 1925).

Summing up, the state of understanding of ice-sheet constructions, erosion and other processes occurring at the base of ice-sheets is far from complete, and the subglacial conditions that control erosion are likely to vary with time and space in complex ways, hence the magnitude of erosion over the next glacial cycle cannot be assessed with precision. Many lines of evidence, however, point to a conclusion that bedrock erosion on this time scale is likely to range between a few metres and a few tens of metres (Figure 7.1). They include diverse geologic evidence, the present topography and bathymetry (e.g., Figure 4.2) that reflect only modest preferential erosion, the spatial distribution of glacial sediments, and results of two independent computer models. Collectively, these lines of evidence indicate that with future glaciations the total erosion over the Bruce Peninsula would continue to be modest (<35 m in 100,000 years) and much smaller than the erosion in the adjacent lakes, because of both the resistant bedrock, and the glaciological and topographic characteristics that make this region a site of relatively slow, intermittent erosion (if any), and net deposition.

Extending this line of reasoning to the one million year (1 Myr) time scale may be useful for some applications although it introduces additional uncertainty because of the inherent inability to predict conditions that are that much further removed from the present. Using the last 1 Myr of earth history as the most natural guide for anticipating the next 1 Myr, it is noteworthy that this period was punctuated by nine major Northern Hemisphere glaciations comparable, in terms of ice-sheet volume and extent, to the LGM based on the global marine isotopic record of ice volumes (e.g., see Figure 2.1A in Peltier 2011). The data and model results summarized in this report for total erosion during one 100,000 year glacial cycle collectively point to a broad range of values for total erosion at the Bruce nuclear site on a 1 Myr time scale. They range from ~300 m, the largest, most conservative amount to a few metres, and perhaps no erosion and net deposition. In view of the absence of topographic features or other known factors that would tend to localize erosion by ice or water over the Bruce nuclear site, and the absence of evidence of preferential past erosion over the site, a more realistic but still quite conservative site-specific estimate is 100 m for 1 Myr.

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9. ABBREVIATIONS, ACRONYMS AND UNITS

CHAMP	Challenging Mini-Satellite Payload
DEM	Digital elevation model
EISMINT	European Ice Sheet Modeling Initiative
GPS	Global Positioning System
GRACE	Gravity Recovery and Climate Experiment
ITRF	International Terrestrial Reference Frame
LGM	Last Glacial Maximum
LiDAR	Light detection and ranging
LIS	Laurentide Ice Sheet
m	metre
Myr	million years
P	basal power
Pa	Pascal
PNW	Pacific Northwest
SAR	Synthetic Aperture Radar
s-forms	sculpted forms
UofT GSM	University of Toronto Glacial Systems Model
VLBI	Very Long Baseline Interferometry
yr	year