Natural Late Holocene lake level fluctuations recorded in the Ipperwash strandplain, southern Lake Huron

by

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Author’s Declaration

I hereby declare that I am the sole author of this thesis. This is a true copy of the thesis, including any required final revisions, as accepted by my examiners.

I understand that my thesis may be made electronically available to the public.
Abstract

The Laurentian Great Lakes (LGL) are the largest system of surface freshwater on Earth. Three factors, glacial isostatic adjustment (GIA), outlet conveyance, and climate processes contribute to natural rises and falls in LGL lake level over geologic time. Studying the natural history of prehistoric lake levels preserved in coastal landforms helps determine the context of current lake levels and predict potential future lake level changes.

Detailed records of lake level change during the late Holocene are preserved in strandplains of beach ridges. Each beach ridge forms as a result of a lake level rise and fall over many decades and preserves a record of relative lake level elevation at the time of deposition. Multiple beach ridges within a single strandplain contain an account of relative lake level changes over the past 4,500 years.

This study examined beach ridges in the Ipperwash strandplain, southern Lake Huron, that uniquely preserves natural lake level fluctuations at the only unregulated outlet in the LGL, the Port Huron/Sarnia outlet of Lake Michigan-Huron, which is particularly susceptible to natural lake level fluctuations. The Ipperwash strandplain is the closest strandplain with the most number of beach ridges to the Port Huron/Sarnia outlet and therefore best records natural lake level fluctuations experienced at the Port Huron/Sarnia outlet of Lake Michigan-Huron.

The study of the Ipperwash strandplain beach ridges used many methods to derive measured elevations and modelled ages of ancient lake levels. Elevation data is combined with age data to create the Ipperwash paleohydrograph. Thirty-six basal foreshore elevations were used to reconstruct the elevation of ancient lake levels. Elevation data shows an oscillatory lake level fall from a maximum elevation of 181.0 m to a minimum elevation of 177.8 m. Ten optically stimulated luminescence ages were used to create a linear age model of the Ipperwash strandplain. The resultant age model shows a maximum age of 3520 years ago and a minimum age of 710 years ago.

The multi-millennium trend shows a net linear fall at an average rate of 7 cm/century for the entire Ipperwash paleohydrograph. This trend is interpreted as a record of the rate of GIA at Ipperwash relative to Lake Michigan-Huron's outlet. The multi-millennium trend suggests the rate of GIA at Ipperwash is 7 cm/century; however, estimates of GIA based on water gauge data suggest the rate
of GIA at Ipperwash is 0 cm/century. This discrepancy could result from an underestimation estimated from contoured water level gauge data for the rate of GIA at Ipperwash, erosion at the Port Huron/Sarnia outlet during the deposition of the Ipperwash strandplain and/or the Chicago outlet being dominant during the deposition of the Ipperwash strandplain.

The multi-millennium trend may also be expressed as two millennium trends shown as two vertically offset phases of lake-level lowering from 3520 to 2180 years ago and 2020 to 710 years ago. These age ranges correspond with the Algoma and 1700-high lake level phases in Lake Michigan. Millennium patterns at Ipperwash corresponds to regional climate records and may represent a climate signal. However, the rate of linear lake level lowering for the older lake level phase at Ipperwash corresponds with the difference in rates of GIA, based on water gauge data, between the Chicago outlet and the Ipperwash strandplain. Therefore, the millennium trends may represent either natural climate change or the abandonment of the Chicago outlet of Lake Michigan-Huron. Detailed sedimentologic and lake level records at the Port Huron/Sarnia and Chicago outlets are needed to resolve this controversy.

Centennial lake level fluctuations represent rises and falls in lake levels lasting an average of 208 years ± 114 years with an average amplitude of 0.8 ± 0.4 m about the linear millennium trends. The average timing of the centennial lake level fluctuation at Ipperwash are similar to centennial lake level fluctuations found in Lakes Superior and Michigan-Huron that are interpreted to represent climate driven lake level fluctuations.

Multi-decadal lake level fluctuations cause a single Ipperwash strandplain beach ridge to form average every 73 ± 35 years. The subsurface stratigraphy of Ipperwash beach ridges shows a similarity of other LGL beach ridges which are interpreted to form as a result of a climate driven lake level fluctuations over many decades.

The Ipperwash paleohydrograph provides the context needed to adjust all strandplain data in Lake Michigan-Huron to resolve basin-wide relative lake level changes related to GIA, outlet conveyance, and climate. In addition, the Ipperwash paleohydrograph suggest lake-level may rise and fall on a multi-decadal time scale contributing to erosion and setting the stage to create a new beach ridge, assuming the rate of sediment supply is maintained.
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Chapter 1

Introduction

The Laurentian Great Lakes (LGL) are the largest freshwater system in the world by surface area, and second largest by volume (Gronewold et al., 2013a). The LGL are a system of interconnected large lakes, as well as connecting and inflowing rivers, surrounding wetlands, and smaller water bodies, which extend along the border of the United States of America and Canada (Figure 1). At the head of the LGL, Lake Superior is the largest lake by volume, and outflows through the Sault Ste Marie rapids and locks (the Sault outlet) affecting lake levels in the lower basins (International Upper Great Lakes Study Board, 2009; 2012). Lakes Michigan and Huron are hydrologically connected and outflow through the only unregulated outlet (lacking any locks) in the entire LGL at Port Huron, Michigan and Sarnia, Ontario (Port Huron/Sarnia outlet) before draining into Lake Erie and then Lake Ontario then flowing down the 1,200 km-long St Lawrence River into the Atlantic Ocean (International Upper Great Lakes Study Board, 2009; 2012; Figure 1).

![Figure 1: The five Laurentian Great Lakes (LGL) are shared water bodies between the USA and Canada. Important former and active outlets are labelled. Today, the entire LGL drains into the Atlantic Ocean via the St Lawrence River. The study area for this thesis is located in southern Lake Huron, near the Port Huron/Sarnia outlet. Satellite imagery from Google Earth.](image)
The LGL contains nearly 20% of the world’s surface freshwater supply, and provides drinking water for forty million people in the USA and Canada (Gronewold et al., 2015). Many commercial industries rely on water from the LGL including fishing, recreation, cargo shipping, hydropower, manufacturing and agriculture. In 2010, a quarter of million jobs supported the shipping of 322.1 million metric tons of cargo through a 3,700 km long deep draft navigation system, the longest in the world, which links the LGL to the global sea lanes (Martin Associates, 2011). If politically combined, the eight US states and two Canadian provinces which at least partially lie within the LGL basin would constitute the third largest economy in the world (Porter, 2015).

Commercial industry in the LGL rely on consistent, or at least predictable, lake levels; however, lake levels in the LGL naturally fluctuate on many scales ranging from seconds to millennia. Lake level changes are influenced by various processes including weather systems, regional climate patterns, vertical ground movement, and human activities (Gronewold and Stow, 2014; Johnston et al., 2014). In modern times, lake level changes damage near shore structures (Meadows et al., 1997), impact wetland habitats (Wilcox et al., 2007), and affect commercial shipping (International Upper Great Lakes Study Board, 2009; 2012). Lake levels in Lake Michigan-Huron over the historic record (since 1860 CE) are recorded by water gauges which show lake levels have fluctuated up to 2 m on a decadal time scale primarily influenced by changes in regional precipitation and evaporation (Gronewold and Stow, 2014). Human modification and the relatively short historic lake level record means the historic record may not fully represent natural lake level patterns which are preserved in depositional coastal landforms. By studying preserved clues within preserved landforms geoscientists can reconstruct lake level fluctuations going back several millennia. These data provides a natural lake level record that can be used to calibrate models which predict future lake level changes, and can be used by shoreline communities and managers to understand the natural rates of erosion and deposition along a specific shoreline.

Of paramount interest is the outlet for Lake Huron, adjacent to Port Huron, Michigan and Sarnia, Ontario, which became the dominant outlet of Lake Huron during the Late Holocene (Thompson et al., 2014). No long-term lake level studies exist near this pivotal outlet (Johnston et al., 2014), therefore my thesis conducted a study of the Ipperwash strandplain, the nearest preserved lake level
record to the Port Huron/Sarnia outlet, to reconstruct natural lake level fluctuations during the Late Holocene.

### 1.1 Significance and objectives

An understanding natural coastal processes is important when making management decisions. Wave action, storms, tides, ice and geology contribute to dynamic coastal systems, and human populations living along a coastline are particularly vulnerable as shoreline position changes with lake level fluctuations.

The LGL contains over 4,500 miles of coastline (Gronewold et al., 2013a), yet when compared to ocean coastlines, LGL coastlines offer a unique setting due to minimal tidal influence, frequent lake level fluctuations over the past 10,000 years and vertical ground movement causing lake levels to relatively rise and fall within the same basin (Rawling and Hansen, 2014). To quantify risk along the LGL coastlines coastal vulnerability indexes, developed for ocean coastlines, have been applied to the Apostle Islands National Lakeshore on Lake Superior, Sleeping Bear Dunes National Lakeshore on Lake Michigan and Indiana Dunes National Lakeshore on Lake Michigan (Pendleton et al., 2010). More commonly, engineering and policy based shoreline management plans have been created for individual LGL shoreline segments (i.e. Reinders, F.J. and Associates, 1989; Au-Sable-Bayfield Conservation Authority, 2000). Both coastal vulnerability indexes and shoreline management plans use the historic lake level record to base decisions on. By studying the natural history of shorelines in the LGL Earth scientistes are able to reconstruct the natural lake level record as well as examine past shoreline behaviors preserved in coastal landforms. Natural records can then be used to improve coastal vulnerability indexes and shoreline management plans by incorporating natural lake level changes and shoreline behaviors.

Studying a specific LGL shoreline can provide insights into the natural history studied shoreline and the LGL as a whole. One example of a LGL shoreline is the Ipperwash strandplain, southern Lake Huron, which is the study site for this thesis (Figure 1). Indigenous peoples have occupied the region since time immemorial, and the Chippewa’s of Kettle and Stoney Point still occupy the area. Ipperwash Beach has been a favored site of cottage goers for well over a hundred years. In addition, the strandplain also hosts many important ecosystems such as some of Ontario’s last remaining
Carolinian Forests (Javala et al., 2015). Pinery Provincial Park in the central portion of the strandplain preserves some of the largest dunes in Ontario as well as diverse ecosystems, while allowing the public to camp, hike, and paddle in the park.

This thesis furthers previous investigations of the Ipperwash strandplain (Johnston, 1999) by examining subsurface stratigraphy of 40 beach ridges and dating 10 individual beach ridges. Subsurface stratigraphy and age data is used to understand the long-term, natural history of the Ipperwash strandplain by 1) defining the natural limits of lake level for the Ipperwash strandplain and 2) deducing the natural patterns of deposition/erosion along Ipperwash beach in relation to natural lake level change.
Chapter 2

Background

2.1 Glacial History and Glacial Isostatic Adjustment

The contemporary extent of the LGL were shaped during the net southward advances and net northward retreats of the Laurentide Ice Sheet. The Wisconsin glaciation was the last glacial advance to erode and deposit material in the LGL basins and reached its maximum extent sometime between 25,270 and 21,290 cal BP before a net northward retreat (Lewis, et al., 2008b). Glacial sediment was sourced from metamorphic and igneous rock of the Canadian Shield in the north and sedimentary rock of the Michigan structural basin in the south. Shoreline outcrops provide an easily available sediment source to a variety of depositional coastal features (Larson and Schaeztl, 2001).

As the ice thinned and retreated the depressed land began to slowly adjust. This adjustment termed glacial isostatic adjustment (GIA), was recognized by early researchers (Gilbert, 1898). Areas of thicker and longer lasting ice were depressed more, and are currently rebounding faster, than areas of thinner shorter-lasting ice. Ice was thicker and longer lasting in the northern portions of the LGL, so today, shore features created by former lake levels in the LGL are generally at higher elevation in the northern portions of the basin (Lewis et al., 2005; Drzyzga et al., 2012). In addition, rising northern outlets were eventually abandoned as outflow transferred to the modern lower southern outlets (Leverett and Taylor, 1915; Lewis, et al., 2008b; Johnston et al., 2014).

Researchers have used GPS and geologic data (Peltier et al., 2015) to estimate rates of GIA on a continental scale. These estimates have been further refined in the LGL using water gauge data (Mainville and Craymer, 2005; Figure 2). Lake level reconstructions from ancient shorelines indicate a similar general pattern of GIA, but suggest some values estimated from water level gauge stations underestimate the long-term rate of GIA shorelines (Baedke and Thompson, 2000; Johnston, et al. 2012). GIA estimations from ancient shorelines and water level gauge data show the ongoing adjustment of the ground surface for the LGL has led to a general southward tilt of the basins, meaning that southern shores are undergoing a long term relative lake-level rise while northern shores experience a relative lake-level fall (Figure 2). The zero isobase or line of no relative uplift or
subsidence has been interpreted to pass through southwestern Lake Superior, northern Lake Michigan and southern Lake Huron (Figure 2).

![Figure 2: Rate of GIA in cm/century based on water level gauge data nested within ICE-3G rates of GIA. Note the zero relative isobase passes through southern Lake Huron near the Port Huron/Sarnia outlet. (Modified from Mainville and Craymer, 2005)](image)

2.2 Prehistoric Lakes

Synthesis of pro-glacial and post-glacial lakes in the LGL basin were first compiled by Leverett and Taylor (1915). Later synthesis includes Hough (1958), Karrow and Calkin (1985), Teller (1987), Larson and Schactzl (2001), Karrow and Lewis (2007), Kincare and Larson (2009), Clark et al. (2012), Lewis and King (2012), Johnston et al. (2014). A review of the pro-glacial and post-glacial lakes in the Huron basin by Lewis et al. (2008b) and refined by Lewis and Anderson (2012) compiles the current status of research exploring how changes in GIA, outlet conveyance, and climate affected basin wide lake level in ancestral Lake Huron. To elucidate variations in lake level, four well studied post-glacial lake phases in Lake Huron are briefly described: Algonquin highstand, Stanley lowstands and Mattawa highstands, Nipissing highstand, and the modern LGL (Figure 3).
2.2.1 Algonquin Highstand

The Algonquin highstand was the first hydrologically joined proglacial lake to occupy the Michigan-Huron basin and its shoreline can be traced around much of the basin (Lewis, et al., 2008b). The Algonquin highstand formed as the isostatically rising Kirkfield outlet (Figure 1) caused lake levels to slowly rise eventually reaching a highstand approximately 13,000 cal BP (Figure 3). At this time Lake Algonquin extended beyond the limits of lakes Superior, Huron and Michigan and was
bordered by the Laurentide Ice Sheet to the north (Figure 3). Following the Algonquin highstand, the retreating Laurentide Ice Sheet uncovered the isostatically depressed North Bay Outlet (Figure 1) and lake levels began to fell as Lake Algonquin began to drain through the North Bay Outlet.

2.2.2 Stanley Lowstands and Mattawa Highstands

As water began to flow through the isostatically depressed North Bay outlet and the Laurentide Ice Sheet retreated from the LGL basin, water levels fell to the Stanley lowstands (Lewis, et al., 2008a; Lewis et al., 2008b; Figure 3). A lowstand was first identified when a deep water unconformity was interpreted as being caused by erosion during a relatively low lake level (Hough 1962). During this time (~10,000 to 8,000 cal BP) lake levels were primarily controlled by the slow ascent of the North Bay outlet (Figure 1; 3). However, sedimentological evidence suggests rapid rises in lake levels collectively known as the Mattawa highstands (Lewis et al., 2005; 2008b). These short highstands are associated with either the glacial outburst floods from the melting ice sheet or overflows from Glacial Lake Agassiz causing lake levels to rapidly rise up to 60 m above the mean elevation of the Stanley lowstands (Lewis and Anderson, 1989; Breckenridge and Johnson, 2009).

However, climatic factors also played a pivotal role, lake level decreased to a point when the Superior, Michigan, Huron and Georgia Bay basins became disconnected between ~8,900 to 8,200 cal BP (Lewis, et al., 2008b, McCarthy and McAndrews, 2012). The disconnected lakes are commonly linked with an increasingly warm and dry climate and demonstrates the LGL's sensitivity to climate changes (Lewis, et al., 2008a).

2.2.3 Nipissing Highstand

The rise to the Nipissing highstand began with a transition to a wetter climate causing water to rise from the Stanley lowstand until drainage again flowed over the still rising North Bay outlet (Figure 1; Johnston et al., 2014). GIA caused the North Bay outlet to rise forming Lake Nipissing when the Huron, Michigan and Superior basins became confluent (Figure 3). However, the lake level rise to the Nipissing highstand, caused by GIA, was supplemented by persistent wet conditions over the LGL (Booth, et al., 2002). A rapid rise of 7.2 cm/year continued until 6,000 cal BP at which point the rate in lake level rise slowed to about 2.8 cm/year until 4,500 cal BP (Thompson et al., 2011) when the Nipissing reached a maximum of 183.3 m at the Port Huron/Sarnia outlet (Thompson et
al., 2014). The slowing of the lake level rise at 6,000 cal BP is attributed to the capture of the outlet by the Chicago and/or Port Huron outlet (Baedke and Thompson, 2000; Johnston et al., 2014).

2.2.4 Modern Lakes

Presently, Lake Superior drains through the Sault outlet into Lake Michigan-Huron which drains through the Port Huron/Sarnia outlet into Lake Erie (Figure 1). The modern configuration formed soon after the Nipissing highstand when lake levels decreased and began to fluctuate within the historic lake level range. Baedke and Thompson (2000) studied five strand plains of beach ridges around Lake Michigan using techniques developed by Thompson (1992) to reconstruct an outlet referenced record of lake level (paleohydrograph) since 4,500 cal BP (Figure 4). Following the Nipissing highstand, 4,500 cal BP, lake level underwent a rapid fall of 4.1 m until 3,400 cal BP (Baedke and Thompson, 2000). The end of the rapid fall is attributed to the abandonment of the Chicago outlet (Baedke and Thompson, 2000). Baedke and Thompson (2000) also propose lake levels rose and fell on a millennial rhythm over the next several thousand years. Lake levels rose from 3,300 to 3,000 cal BP, associated with the Algoma highstand, fell from 2,400 to 2,250 cal BP, and rose from 2,100 to 1,700 cal BP and fell from 1,700 to 1,000 cal BP (Figure 4). This millennial rise and fall is associated with changes in climate though the precise mechanism has not been identified (Baedke and Thompson, 2000). In Lake Huron, over the past millennium, lake level has fluctuated within historical measurements of 2.1 m (Lewis, et al., 2008a). Millennial oscillations were also found in Lake Superior and were a factor in the final separation of Lake Superior from Lake Huron-Michigan at approximately 1060 cal BP (Johnston et al., 2012).
2.2.5 Drivers of Post-Nipissing Lake Level Change

Thompson and Baedke (1995) identified three lake level patterns by examining the geomorphology and sedimentology in five Lake Michigan strandplains of beach ridges (Figure 4). Millennium, centennial and multi-decadal lake level patterns are quasi-periodic and superimposed on one another. A millennium oscillation is observed as sets of beach ridges and correspond to the Algoma in the upper LGL and the Sault and sub-Sault lake phases in the Superior basin (Johnston et al., 2012). Lake levels fell from the Nipissing to Algoma to modern lake levels has been associated with outlet incision at Port Huron/Sarnia outlet (Hough, 1962) and/or climate (Booth et al., 2002). A shorter-term centennial pattern with a quasi-periodicity of approximately 160 years are composed of groups of 4-6 ridges. A similar pattern, lasting 100-150 years, is attributed to climate (Fraser et al., 1990).
The shortest quasi-periodic pattern repeats about every 30 years and is the average amount of time it takes for an individual ridge to be deposited.

Climate has been suggested as the cause of the three lake level patterns identified by Baedke and Thompson (2000). A correlation between atmospheric circulation patterns and quasi-periodic lake level fluctuations has been investigated but a consensus has not yet been reached. Millennial climate oscillations over the North Atlantic (Bond et al., 1997) drive atmospheric circulation patterns over North America effecting climate (Viau et al., 2002). The transition to the sub-Sault phase in Lake Superior is attributed to drought over North America during the transition from the Medieval Climate Anomaly to the Little Ice Age (Johnston et al., 2012).

Quasi-periodic decadal and multi-decadal lake level oscillations are attributed to changes in atmospheric circulation patterns over the LGL (Cohn and Robinson, 1976; Polderman and Pryor, 2004; Hanrahan et al., 2009; Watras et al., 2014). Quasi-periodic multi-decadal lake level fluctuations have been attributed to either the Pacific Decadal Oscillation (Watras et al., 2014), the North Atlantic Oscillation (Hanrahan et al., 2009), changes in large scale atmospheric circulation over the arctic (Polderman and Pryor, 2004) or the intermodulation of two near decadal atmospheric oscillations over the North Atlantic (Hanrahan et al., 2010). The linkage between climate and lake level emphasize how susceptible the LGL is to climate change.

2.2.6 Historic lake level drivers

Historic drivers of lake level change are GIA, outlet conveyance, and climate with climate being the largest contributor (International Upper Great Lakes Study Board, 2009; 2012). The causes of climate driven historic lake level changes are examined and used to better understand possible climate influence on prehistoric lake levels.

The present area of the LGL basin contains roughly 33% surface water and 67% land (compared to other large lake basins which often contain 1-5% surface water), therefore, overlake evaporation, overlake precipitation as well as basin runoff are chief contributors to net basin supply and consequent lake level fluxes (Gronewold, et al., 2013b). Of particular consequence is the effect of seasonal changes in the frequency and intensity of weather patterns over the LGL (Argyilan and Forman, 2003; Polderman and Pryor, 2004). For instance, wide spread drought in North America
during the 1930s CE correlates with an extreme lake level low in Lake Huron caused by a reduction in spring and summer overlake precipitation and basin runoff (Argyilan and Forman, 2003). While the lake level low between the late 1990s CE and early 2010s CE has been related to increasing temperature reducing winter ice cover and subsequently increasing overlake evaporation (Gronewold and Stow, 2014). On the other hand, lake level rise in the 1980s CE is attributed to increased overlake precipitation and basin runoff during the autumn (Argyilan and Forman, 2003). While recent (2013-2014 CE) lake level rise is attributed to above average spring and fall overlake precipitation and basin runoff coupled with reduced summer and winter overlake evaporation caused by below average temperatures during those months (Gronewold et al., 2016). Over historic times, periods of persistent dry and warm climate correspond with lake level falls and periods of persistent wet and cool climate correspond with lake level rises (Argyilan and Forman, 2003; Gronewold et al., 2016). The relationship between high lake level with cool and wet climate and low lake levels with warm and dry climate is also postulated in the ancient lake level record (Fraser et al., 1975)

Numerous researchers have run computer simulation in an attempt to predict future LGL water levels under differing climate change scenarios and have calculated that lake levels are likely to fall as the climate warms (Lofgren et al., 2002; Angel and Kunkel, 2010; Hayhoe et al., 2010; Lofgren et al., 2001; MacKay and Seglenieks, 2013). However, since lake levels natural fluctuate it is important to have precise lake level records over prehistoric times to calculate natural lake level fluctuations and calibrate models predicting future lake level changes.

### 2.3 Beach Ridges

Comprehensive reviews of marine and lacustrine beach ridges are found in Taylor and Stone (1996), Otvos (2000), Hesp et al. (2005) and Tamura (2012). Beach ridges are common features on many depositional coastlines that are topographically expressed as elongated sand and/or gravel ridges running parallel or subparallel to the shoreline and are separated by intervening low areas called swales. Beach ridges are composed of sand, gravel, and/or shingles (flat cobbles common in some coastal areas) and contain a core of waterlain sediments capped by windblown sediments. A series of beach ridges attached to the mainland is termed a strandplain (McCubbin, 1981).
2.3.1 Beach Ridges and LGL Lake Levels
In the LGL, strandplains were noted and described by early researchers Goldthwait (1908) and Leverett and Taylor (1915). Decades later researchers began to utilize beach ridges to identify ancient lake levels (Larsen, 1985; Fraser et al., 1991). Thompson (1992) proposed the shoreface-foreshore contact (also called the basal foreshore) within an individual beach ridge correlates with lake level at the time of deposition. By coring and analyzing multiple beach ridges many ancient lake level elevations at one site are derived from one strandplain. Time constraints are attained through the dating of organic samples in swales or mineralogic sample in ridges. An age model is then created for each strandplain. Cross-strandplain elevations and ages are then used to produce a paleohydrograph extending back several millennia with a multi-decadal resolution.

2.3.2 Sedimentary Facies
In cores collected from the individual beach ridges, normally less than five meters in length and four inches in width, genetic sediment facies in sandy LGL beach ridges are dune, foreshore and upper shoreface sediments (Figure 5; Thompson, 1992; Baedke et al., 2004). Upper shoreface deposits typically consist of moderately sorted upper very fine to lower fine sand with some beds of coarse sand and gravel. Shoreface deposits may also contain silt and clay laminae. Sedimentary structures typically include horizontal to high angled parallel laminae, ripple bedding and cross stratification with the entire sequence coarsening upward. Foreshore deposits are typically 1.0 to 1.8 m thick and consist of moderately sorted, upper fine to cobble size particles (sediment size dependent on sediment supply). Sedimentary structures include horizontal and lakeward-dipping subhorizontal laminae defined by alternations in grain size, with coarser grains concentrated at the base. The lower coarse-grain portion is called the basal foreshore and correlates with lake level at the time of deposition. The contact between the foreshore and upper shoreface is typically sharp and easily identified (Thompson, 1992) but may locally appear homogeneous (Johnston et al., 2007). Dune facies are 0.5 to 4 m thick and consist of moderately to well sorted, lower medium to lower fine grain quartz sands. Dune deposits are usually unstratified, but may locally contain heavy mineral laminae and/or horizontal to steeply dipping laminae. The upper portion typically contain rootlets. The contact between dune and foreshore deposits is sharp to gradational (Thompson, 1992).
2.3.3 Depositional Model

Thompson and Baedke (1995) modified Curray’s (1964) model of shoreline behavior to provide a conceptual framework for the deposition of a beach ridge in response to lake level oscillations. The depositional or erosional behavior of a given coast is guided by the rate in water level change and the rate of sediment supply. Therefore, changes in the rate of lake-level rise or fall and changes in sediment supply will affect shoreline behavior. In an area where sediment supply exceeds the rate of lake level change the shoreline will prograde, build lakeward. Progradation can occur during rising lake levels if the rate of sediment supply exceeds the rate of lake level rise, but most readily happens as lake level falls.

Multi-decadal rises and falls in lake levels create beach ridges (Thompson and Baedke, 1995; Figure 6). When lake level change transitions from falling to rising lake levels the shoreline may eventually begin to erode. Then as the lake level rise reaches a peak elevation the rate of lake level change decreases and the shoreline will begin to aggrade, build upward. As lake levels fall, the shoreline will prograde forming the lakeward adjacent swale. Repetition of multi-decadal lake-level rises and falls have been shown to create a strandplain of beach ridges if sediment supply and accommodation space is sufficient for deposition (Figure 6; Thompson and Baedke, 1995; Johnston et al 2007).
Figure 6: Conceptual model of LGL beach ridge development shows how beach ridges develop in response to lake level fluctuations and sediment supply. Colors show correlations between graphs and diagrams. A simple rise or fall in lake level will affect the rate of lake level change (A) and a change in the rate of sediment supply will affect shoreline behavior (B) together the rate of lake level change and sediment supply dictates how the coastline responds to lake level fluctuations (modified from Johnston et al., 2007). Diagrams of strandplain cross sections show how the beach responds to changing rates of lake level change (C).
Ground penetrating radar (GPR) has been used to understand stratigraphic patterns across several beach ridges in the LGL. Johnston et al. (2007) used GPR on a Lake Superior strandplain to image cross strandplain stratigraphy and identify a systemic pattern in LGL beach ridges to link a conceptual model of beach ridge formation with the preserved record. The water table is often imaged in GPR profiles as a continuous horizontal reflection. Beach ridge foreshore deposits are imaged as sigmoidal reflections that build upward and lakeward from a concave up reflection extending from below a beach ridge crest to the ground surface in the next landward adjacent swale. The reflections imaged across multiple beach ridges show the preserved stratigraphic patterns resulting from the deposition of multiple beach ridges.

Changes in the rate of lake level fluctuation and sediment supply cause individual LGL shorelines to erode and deposit over geologic time. This causes beach ridges to form in embayments along the LGL over a multi-decadal time scale, eventually forming a strandplain of many beach ridges. Because individual beach ridges form as a result of lake level fluctuations (and sediment supply) they preserve a lake level elevation at the time of deposition. Strandplains of many beach ridges can therefore be used to reconstruct lake level fluctuations over geologic time.
Chapter 3
Study Area

3.1 Lake Huron

Lake Huron is located at the downstream terminus of the upper LGL (considered lakes Superior, Michigan, and Huron). The major source of inflow to Lake Huron is from Lake Superior through the St Mary’s River and the Sault outlet. Outflow from the Lake Huron and the entire upper LGL discharges through the Port Huron/Sarnia outlet in southern Lake Huron, and into the St Clair River, Lake St Clair and the Detroit River before draining into Lake Erie (Figure 1). Lakes Michigan and Huron are hydrologically joined through the Straits of Mackinaw. Flow through the Straits of Mackinaw is typically from west to east, though periodic changes in weather conditions over each lake disrupts net flow through the Straits (Saylor and Sloss, 1976; Saylor et al., 1991). Lake Michigan and Lake Huron are often considered separate basin because of their narrow connection (Lewis et al., 2008b). Lake Huron's surface area is 49,600 km², while Lake Huron's total drainage area is 194,000 km². The Lake's maximum depth is 229 m and its total water volume is 3,540 km³ (Lewis, et al., 2008b). Lake Huron's mean yearly elevation based on water gauges (1918-2015 CE) is 176.42 m. Lake Huron has undergone noticeable lake level fluctuations, for instance a maximum lake level yearly average high was reached in 1986 at 177.29 m while a lake level minimum was reached in 1964 at 175.68 m (Gronwold, et al., 2013b).

The potential for outlet conveyance, a change in the water carrying capacity of a lake's outflow channel(s), could cause changes in lake level and prevent deep draft vessels passage through the outlet. From the Port Huron/Sarnia outlet to Lake Erie the average elevation drop is 3 m over 130 km. The relatively minor drop in elevation allows ships to pass from Lake Huron to Lake Erie without the necessity of locks. This is partly why the Port Huron/Sarnia outlet is the only unregulated outlet (lacking any locks or dams) in the entire LGL. The lack of any regulatory structures means cargo ships must vary their weight depending on lake level, lightening their load during low lake levels. The St Clair River channel has been dredged in the past to allow for passage of deep draft vessels during low lake levels. The last major dredging of the St Clair River occurred in 1963 and contributed to a permanent reduction in Lake Huron lake levels (International Upper Great Lakes Study Board, 2009). Since then an average decline of 23 cm has occurred in the difference of lake levels between Lake Huron and Lake Erie. Baird & Associates (2005) attributed
this drop to downcutting of the St Clair River, though further investigations identified active bedforms on the river bottom suggesting the river has not downcut in the recent past (Czuba et al., 2011). With the completion of the International Upper Great Lakes Study (International Upper Great Lakes Study Board, 2009; 2012) the decline in lake level between 1963 and 2006 was attributed in part to a change in outlet conveyance (7-14 cm), GIA (4-5 cm); however, the majority of recent lake level change was attributed to climate change (9-17 cm).

However, there is evidence of deposition in the Port Huron/Sarnia outlet in the form of spits (Campbell, 2016). The presence of the Port Huron/Sarnia spits means that sediment has accumulated in the Port Huron/Sarnia outlet in the past and could potentially accumulate in the outlet in the future. If sediment accumulates in the Port Huron/Sarnia outlet in the future it could prevent the passage of deep draft vessels. Quantifying past lake level fluctuations in southern Lake Huron will provide details needed to determine the timing and context of deposition of the Port Huron/Sarnia spits.

The study site for this thesis is strategically located in southern Lake Huron near the Port Huron/Sarnia outlet (Figure 1). Ipperwash is the closest strandplain to the Port Huron/Sarnia outlet with the most number of preserved beach ridges that document many past lake level fluctuations. Since the Ipperwash strandplain is located near the outlet (~40 km), the Ipperwash beach ridges preserve conditions experienced at the Port Huron/Sarnia outlet in geologic past.

### 3.2 The Thedford Embayment

The Thedford embayment is located approximately 42 km northeast of the Port Huron/Sarnia outlet, 175 km west of Hamilton, Ontario and 130 km northeast of Detroit, Michigan. Towns in the area include the coastal towns of Grand Bend, Port Franks, Ipperwash Beach, and Kettle and Stoney Point First Nation communities, and the town of Thedford is located at the inland margin of the embayment (Figure 7).

Bedrock consists of the Middle Devonian Dundee Formation, Hamilton Group and Kettle Point Formation (Cooper, 1974). The only bedrock exposures are on Kettle Point and Stoney Points where the Kettle Point Formation outcrops before extending as offshore ridges. Between Port
Franks and Grand Bend a buried bedrock valley, called the Ipperwash trough, exists and is interpreted to have been eroded by subglacial/glacial meltwater (Gao, 2011).

Till exposures in the area primarily consist of St Joseph till (Cooper, 1974). The St Joseph till is a grey to yellowish-brown silt to clayey silt till with few pebbles (Cooper, 1974) and when exposed on the shoreline forms easily erodible bluffs (Au-Sable-Bayfield Conservation Authority, 2000). Erosion of bluffs of St Joseph till northeast of the Thedford embayment has historically supplied sediment to the littoral system at a rate of 68,000 m$^3$/year (Reinders and Associates, 1989).

The Thedford Embayment comprises the area between an elevated wave cut bluff and the modern shoreline. Between Kettle Point and Stoney Point, the embayment extends 2 km landward from the modern shoreline and then increases to a maximum of 10 km landward from the modern shoreline in a broad arc between Port Franks and Grand Bend (Figure 7). Within the landward portion of the embayment at least two large gravel bars have been related to either the Nipissing or Algonquin lake phases (Cooper, 1979). Further studies postulated the gravel bars to contain an Algonquin age core capped by Nipissing deposits based on the identification of a paleosol in several gravel pits (Karrow et al., 1980).

### 3.3 The Ipperwash strandplain

The Ipperwash strandplain consist of a 2 km wide strip of land between a prominent wave cut bluff/dune ridge landward from the modern Lake Huron shoreline and extends 25 km from Kettle Point northeast to Grand Bend (Figure 7). The northeastern and southwestern portions of the Ipperwash strandplain contain beach ridges while the central portion is masked by dunes in Pinery Provincial Park (Eyles and Meulendyk, 2012). The landward limit of the strandplain is a 10 m high bluff/dune ridge formed during the Nipissing (Cooper, 1979).

The southern Ipperwash strandplain, near Ipperwash Beach, contains a well-defined strandplain (Figure 8) and up to 40 individual beach ridges (Johnston, 1999). Drainage in the Ipperwash Beach area flows to the north-east along the natural Duffus Drain and manmade Ipperwash Drain (Figure 8). Road access is along West Ipperwash, and Ipperwash Road with numerous, hiking, ATV, and two-track trail allowing for relatively easy access to parts of the Ipperwash strandplain (Figure 8).
This thesis focuses on a surveyed shore perpendicular transect across the southern (Ipperwash Beach area) Ipperwash strandplain starting at the modern beach and ending at the Nipissing bluff approximately 2.2 km landward from the modern shoreline to reconstruct past relative lake levels preserved in beach ridges. The southern Ipperwash strandplain is an ideal location to study past lake level fluctuations for several reasons. 1) The Ipperwash strandplain is predicted to have a rate of GIA similar to the rate of GIA experienced at the Port Huron/Sarnia outlet (Figure 2). Therefore, the Ipperwash strandplain beach ridges will provide detailed information about lake level fluctuations at the outlet. 2) the 40 Ipperwash strandplain beach ridges are, at present, well preserved and have little dune covering allowing for easy access to waterlain sediments (Johnston, 1999). This thesis extends well beyond the Johnston (1999) topographic survey by coring 40 Ipperwash beach ridges to uniquely derive accurate elevations of past lake level stages and deriving the first age model for the Ipperwash strandplain to produce the most detailed record of post-Nipissing lake level fluctuations at Ipperwash created to date. The Ipperwash paleohydrograph provides a lake level record of past lake level fluctuations at Lake Huron’s outlet which is required to understand past conditions (GIA, outlet conveyance, climate) in the Lake Huron basin and at specific sites around Lake Huron.
Figure 7: The Thedford Embayment, southern Lake Huron, contains several ancient shorelines. The Ipperwash strandplain extends from Kettle Point to Grand Bend and extends from the Nipissing shoreline to the modern beach. Shorelines in Pinery Provincial Park are covered by large dunes, while beach ridges are exposed on the northern and southern ends of the Ipperwash strandplain. The study area is located on the southern end of the Ipperwash strandplain, near Ipperwash Beach. Elevation data from Southwestern Ontario Orthimagery Project 2015.

Figure 8: Aerial image of the southern Ipperwash strandplain with outline of surveyed area, major drainages and the Nipissing bluff. Road access is along W Ipperwash Road, Ipperwash Road, and Army Camp Road. Satellite imagery from Google Earth.
Chapter 4

Methods

To create an Ipperwash paleohydrograph, diverse methods were used to glean relevant data from the Ipperwash strandplain. These methods include field observations, satellite image interpretation, ground penetrating radar (GPR), topographic surveying, vibracoring, sediment analysis, and optical dating. These data were synthesized and systematically interpreted to create an Ipperwash relative paleohydrograph.

4.1 Field Observations and Satellite Image Interpretation

Field notes were collected along the surveyed transect from the modern beach landward 2.2 km to the Nipissing bluff (Figure 8). Field observations include estimations of swale width, water depth in swales, ridge width, ridge height, and ridge crest topography as well as notes describing the type and distribution of vegetation and human modifications. Width estimates from field observations were compared to satellite imagery available through GoogleEarth. Comments on GPR profile locations, core locations and the nature of vibracoring (i.e. how well the vibracore penetrated the subsurface) were also recorded.

4.2 Ground Penetrating Radar (GPR)

GPR is a noninvasive method to image the shallow subsurface. GPR utilizes electromagnetic (EM) waves which reflect from changes in dielectric properties of sediments (Jol and Bristow, 2003). The GPR set-up used in this study consists of a backpack mounted computer/console connected to a transmitter and receiver antennae with fiber optic cables. The GPR system was a pulseEKKO 100 with 100 and 200 MHz antennae and a 1000 V transmitter. For all transects, data was collected in step mode with a step size of 0.5 m or 0.1 m and antennae separation of 1.0 m or 0.5 m for 100 and 200 MHz antennae respectively. Relief measurements were collected using an optical surveying level to geometrically adjust the radar profile to reflect changes in topography. The data was processed using EKKO_Project using dewow filtering, vertical and horizontal averaging and automatic gain control.
Estimates of propagation velocities of the EM wave through the sediment are needed to calculate depth. Common midpoint surveys (CMPs) were collected parallel to several transects to estimate propagation velocity through the sediment. However, CMP velocities were adjusted based on lithology and depth to water table.

For the Ipperwash strandplain, GPR was used as an exploratory method to determine the sedimentary architecture across multiple beach ridges. Defining the sedimentary architecture distinguishes natural and anthropogenically modified sediment to determine the viability of collecting near-surface foreshore sediments from vibracoring. In addition, the internal stratigraphy of the Ipperwash strandplain beach ridges were compared to another GPR transect through LGL beach ridges (Johnston et al., 2007).

4.3 Topographic Surveying

A Sokkia total station survey was used to deduce the precise elevation of every beach ridge crest and swale, core location, and OSL pit. Elevations were corrected to a well-established datum in the LGL’s (International Great Lakes Datum 1985 or IGLD85) by using a geodetic survey benchmark in Grand Bend to measure the water level elevation of Lake Huron at Grand Bend, then using this water level elevation at Ipperwash as a known elevation to link the Ipperwash transect to IGLD85. Waves were dampened to better measure water level elevation by using a large plastic cylinder with a tube at its base to allow for the passage of water. Lake level elevation was also compared to the lake level recorded at the water level gauge station in Goderich, Ontario (62 km NNW of Ipperwash), to ensure elevations were consistent. A 0.05 m discrepancy is observed between Grand Bend and Goderich. Elevations, relative to IGLD85 were established for each core site, OSL pit and beach ridge crest and swale.
To develop a topographic profile across the Ipperwash strandplain a shore perpendicular transect was constructed. However, cores were not collected in a strict shore perpendicular orientation and the surveyed transect following core locations. A Garmin handheld GPS was used to determine core locations and these locations were then laterally adjusted to a shore perpendicular transect, roughly equidistant from all core locations, drawn in QGIS (Figure 9). Crest and swale midpoints were determined along the transect by examining aerial photographs. By adjusting GPS based core locations to a shore perpendicular transect and determining crest and swale locations along the transect, a topographic profile of the Ipperwash strandplain was created.

4.4 Vibracoring

Vibracoring is a proven method to extract subsurface sediment from beach ridges and identify relic beach facies (Thompson, 1991; 1992). A 3 m tall, land-based vibracore system was used to collect cores 1-4 m in length. Cores were collected from the lakeward margin of each, individual beach ridge to minimize the amount of dune sand recovered and penetrate deep enough to collect the contact between foreshore and upper shoreface facies. The lakeward side of each core was marked on each core tube so the orientation of sedimentary structures could be examined once the cores were returned to the laboratory and opened.

4.5 Lab Analysis

Cores were analyzed following the methods of Thompson (1991; 1992). Cores, once returned to the lab, were split, described (for visual grain size estimates, sedimentary structures, composition),
logged, photographed, preserved and sampled. Grain size was determined through visual estimation, and laser diffraction. Graphs of grain size distribution were produced to identify grain size changes within cores and across multiple cores. Latex peels were used to enhance sedimentary structures and produce a permanent record of the cores. Once sampled and described, facies were interpreted based on visual descriptions, grain size distribution and sedimentary structures.

Core samples, of approximately 2 cm$^3$ were collected every ~20 cm starting below the soil horizon. Additional samples were also collected above and below visibly identifiable contacts and in areas with unique sediments or stratigraphy. Over 200 samples from cores were analyzed to determine the grain size using the Malvern3000 laser diffractor system at the Indiana State Geological Survey. Samples collected on the modern beach were analyzed for grain size using a Fritsch laser diffractor system at the University of Waterloo. Laser diffraction is a proven method to analyze particle size of very fine grained to coarse sand sized, naturally occurring sediments (Sperazza, et al., 2004; Di Stefano et al., 2009). Analyzed samples were collected from cores chosen to bracket visually described facies and better define the contact between facies.

Once analyzed, samples were statistically described using GRADISTAT v. 08 (Blott and Pye, 2010). Statistical parameters were calculated using the Folk and Ward (1957) method. Final statistics were displayed in phi and described.

Sedimentary structures and statistical parameters are used to define facies interpreted as dune, foreshore and shoreface deposits. The study of other LGL strandplains suggests mean grain size, sorting, skewness and coarsest 1% (D(99)) are the most useful grain size statistics to differentiate facies (Thompson et al., 1997; Baedke and Thompson, 2000; Johnston et al., 2004, 2012, 2014). The contact between foreshore and shoreface sediments serves as a proxy for ancient lake level for each individual beach ridge (Thompson, 1992). Modern beach facies are used as an analogue and used to interpret ancient beach ridge facies.

4.6 Optical Stimulated Luminescence

Optical stimulated luminescence (OSL) is a technique to measure when certain minerals, typically quartz, feldspar or aluminum oxides, were last exposed to sunlight. OSL relies on mineral grains
being sufficiently exposed to sunlight to “bleach” the grains. Once bleached, the minerals absorb radiation from the decay of surrounding radioactive isotopes. This radiation is causes the trapping of electrons within imperfections of the crystal lattice. Stimulating samples with certain wavelengths of light and measuring released energy allows for the calculation of burial time (Aitken, 1998). OSL is a proven dating method in both marine (Murray-Wallace et al., 2002) and lacustrine (Argyilan et al., 2005) coastal landforms. In the LGL OSL has been used in recent studies as an alternative to $^{14}$C dating of organics found in the swales of beach ridges (Argyilan et al., 2005) to develop paleohydrographs for lake basins (Johnston et al., 2012). Samples were collected from approximately 1 m below the crest of beach ridges. Temporal and monetary constraints allow only every third to fifth beach ridge in the Ipperwash strandplain to be age-dated using OSL methods.

The OSL samples from the Ipperwash strandplain were sent to Dr Lepper at the Optical Dating and Dosimetry Lab at North Dakota State University. OSL ages were determined from quartz grains and “derived from data collected from between 93 and 144 individual aliquots per field sample and represent over 12,000 individual OSL measurements” (Lepper, 2017). OSL ages are in calendar years before 2017 CE, but the abbreviation BP is not used to avoid confusion with other published ages reported as calendar years before 1950 CE.
Chapter 5
Results and Discussion

Field data and samples were collected from the Ipperwash strandplain during the 2015 and 2016 field seasons. Data collected includes field observations, 14 GPR profiles totaling over 1 km in length, a 2 km survey line, 6 sediment samples from the modern Ipperwash beach, 40 vibracores with an average depth of penetration of 2.82 m ± 0.68 m, and 10 OSL samples collected in a roughly shore normal orientation across the Ipperwash strandplain (Figure 10).

These data are used to develop a relative paleohydrograph for the Ipperwash strandplain. Field observations were compared to satellite imagery and used to qualitatively divide the survey into segments based on ground surface observations. GPR is used as an exploratory tool used to describe subsurface stratigraphy. Topographic surveying allows the precise elevation of sample locations to be determined and to examine topographic patterns across the Ipperwash strandplain. Sediment samples from the modern beach are used to describe modern shoreline sediment facies. Vibracores from individual ancient shorelines are used to describe ancient shoreline sediment facies. Modern and ancient facies are interpreted either as dune, foreshore or shoreface deposits which are splayed laterally along the modern shoreline and vertically in cores from ancient shorelines. The contact between the foreshore and shoreface facies is of paramount interest since it approximates the elevation of the modern and ancient lake level when individual beach ridges
formed. OSL is used to date individual shorelines and develop an age model to reconstruct the timing of natural lake level oscillations over the past several millennia.

5.1 Field Observations

A field survey was collected in the summer of 2016 beginning at the modern Lake Huron shoreline and extending in a roughly shore perpendicular orientation 2200 m to the Nipissing bluff. The field survey can be qualitatively divided into 8 segments based on field notes and verified by satellite imagery. The 8 segments are described and separated based on swale width, swale water depth, ridge width, ridge height/crest topography, vegetation, and human modification (Table 1).

Segment A extends from the modern shoreline to 209 m landward and contains approximately 3 ridges and the modern foredune. Swales and ridges are variable in width. Swales are dry and ridges are up to 4 m high. Vegetation is dominated by beach grasses and has been modified by humans including privately owned homes and cottages, publicly maintained beach access paths and washrooms, as well as roads and underground utilities. Cores were not collected from this segment due to human modification.

Segment B extends from 209 m to 354 m landward from the modern shoreline and contains 4 ridges from which cores 2001-2004 were collected. Swale and ridge width and height are unknown due to human modification. The area crossed by the survey is a former parking lot, and is currently used for hiking and ATV/dirt biking. Elsewhere on the strandplain, segment B contain residential housing. With the exception of a central swale, swales have been infilled but ridge crests are still apparent in certain areas. The segment is open with patches of juniper bushes with cedar trees along a central swale.

Segment C extends from 354 m to 513 m landward from the modern shoreline and contains 3 ridges from which cores 2005-2007 were collected. Wide widths are observed for swales (10-30 m) and ridges (15-30 m). Swales contain deep water (~0.5 m) and ridges are low (~1 m above swales) with relatively flat crests. Swales are well vegetated with closely spaced, short, thin, woody plants with cedar and birch trees at the swale-ridge contact. Ridge crests are covered by grasses and juniper bushes. Human modification is limited to hiking trails.
Segment D extends from 513 m to 736 m landward from the modern shoreline and contains 6 ridges from which cores 2008-2013 were collected. Swale and ridge width is variable. Swales are dry and ridges are high (up to 2 m) and steadily rise landward often with bench topography (i.e. a low ridge followed by a high ridge with no discernable swale between ridges). The area is dominated by Carolinian Forest, a particular forest type dominated by deciduous trees. Human modification is limited to hiking trails in most areas, but the segment is split by the Ipperwash Drain which was cut parallel to ridge crests.

Segment E extends from 736 m to 951 m landward from the modern shoreline and contains 4 ridges from which cores 2014-2017 were collected. Wide widths are observed for swales (15-30 m) and ridges (20-30 m). Swales are dry to shallow (~0.2 m water depth), and ridges are variable height with relatively flat to hummocky crests. The segment is dominated by Carolinian Forest. Human modification is minimal except for a cleared area on the crest of ridge 2017 which contains a concrete foundation, presumable the base of an old cabin.

Segment F extends from 951 m to 1379 m landward from the modern shoreline and contains 12 ridges from which cores 2018-2029 were collected. Narrow widths are observed for swales (5-20 m), but ridge width is variable. Swales are dry to shallow (~0.5 m water depth), and ridges are variable height with relatively flat to hummocky crests. Segment F is dominated by Carolinian Forest. Human modification is minimal, limited to two-track trails and selective logging.

Segment G extends from 1379 m to 1785 m landward from the modern shoreline and contains 4 ridges from which cores 2030-2038 were collected. Wide widths are observed for swales (15-40 m) and ridges (20-40 m). Swales are shallow to deep (0.2-1.0 m water depth), and ridges are variable height often with bench topography. The segment is dominated by Carolinian Forest. Human modification is minimal, limited to a two-track trail and selective logging.

Segment H extends from 1785 m to 2200 m landward from the modern shoreline, ending at the Nipissing bluff identified by Cooper (1979) and contains ~8 ridges from which cores 2038-2040 were collected from the lakeward most two ridges. Ridges landward of the Ponderosa Pines Golf Course were not cored. Ridge and swale width and height are variable and rise steadily landward.
Swales are dry to shallow (0-0.2 m water depth), and ridges are variable height often with hummocky topography. The segment is dominated by Carolinian Forest. Human modification is minimal, limited two-track trails, except for the area occupied by the Ponderosa Pines Golf Course.

**Table 1: Characteristics used to qualitatively divide the Ipperwash strandplain into 8 segments**

<table>
<thead>
<tr>
<th>Transect Segment</th>
<th>A</th>
<th>B</th>
<th>C</th>
<th>D</th>
<th>E</th>
<th>F</th>
<th>G</th>
<th>H</th>
</tr>
</thead>
<tbody>
<tr>
<td>Number of Ridges</td>
<td>~3</td>
<td>4</td>
<td>3</td>
<td>6</td>
<td>4</td>
<td>12</td>
<td>10</td>
<td>~6</td>
</tr>
<tr>
<td>Distance Landward</td>
<td>0-209 m</td>
<td>209-354 m</td>
<td>354-513 m</td>
<td>351-736 m</td>
<td>736-951 m</td>
<td>951-1379 m</td>
<td>1379-1785 m</td>
<td>1785 m - 2200 m Nipissing Bluff</td>
</tr>
<tr>
<td>Swale Width</td>
<td>Variable</td>
<td>Unknown</td>
<td>Wide (10-30 m)</td>
<td>Variable</td>
<td>Wide (15-30 m)</td>
<td>Narrow (5-20 m)</td>
<td>Wide (15-40 m)</td>
<td>Variable</td>
</tr>
<tr>
<td>Swale Water Depth</td>
<td>Dry Swales</td>
<td>Dry Swales (infilled)</td>
<td>Deep wetlands (~0.5 m water depth)</td>
<td>Dry Swales</td>
<td>Shallow (0-0.2 m water depth)</td>
<td>Shallow (0-0.5 m water depth)</td>
<td>Shallow to deep (0.2-1 m water depth)</td>
<td>Dry to Shallow (0-0.2 cm)</td>
</tr>
<tr>
<td>Ridge Width</td>
<td>Variable</td>
<td>Unknown</td>
<td>Wide (15-30 m)</td>
<td>Variable</td>
<td>Wide (20-60 m)</td>
<td>Variable</td>
<td>Wide (20-40 m)</td>
<td>Variable</td>
</tr>
<tr>
<td>Ridge Crest Height/Topography</td>
<td>High (up to ~4 m) dunes</td>
<td>Unknown</td>
<td>Flat low crests (~1 m)</td>
<td>Steadily rising, high (up to ~2m) crests often with bench like topography</td>
<td>Hummocky to flat topography</td>
<td>Variable</td>
<td>Variable often with bench like topography</td>
<td>Hummocky</td>
</tr>
<tr>
<td>Vegetation</td>
<td>Beach grasses</td>
<td>Open with cedars in central swale</td>
<td>Vegetated wetlands, cedars at wetland edge, open crests</td>
<td>Carolinian Forest</td>
<td>Carolinian Forest</td>
<td>Carolinian Forest</td>
<td>Carolinian Forest</td>
<td></td>
</tr>
<tr>
<td>Human Modification</td>
<td>Privately owned houses and publicly maintained beach access points and washrooms</td>
<td>Former parking lot, ATV trails. Swales infilled but ridges still apparent</td>
<td>Minimal</td>
<td>Hiking trails. Cut through by Ipperwash Drain</td>
<td>Minimal expect for cleared area on ridge 2017</td>
<td>Minimal, selective logging and two-track trails</td>
<td>Minimal, selective logging and two-track trails</td>
<td>Minimal, two-track trails, with exception of Ponderosa Pines Golf Course</td>
</tr>
</tbody>
</table>
5.2 Radar Stratigraphy

Seven GPR profiles totaling 1069 m in length were collected in shore normal orientations across the modern beach (1 profile) and ancient shorelines (6 profiles) as well as 4 GPR profiles totaling 205 m in length were collected shore parallel on the modern beach (Figure 11). Relief measurements were collected with a Topcon laser level to geometrically adjust each GPR profile to changes in ground surface elevation.

![Figure 11: Location of GPR profiles collected on the Ipperwash strandplain.](image)

EM wave propagation velocity is needed to estimate depth and is determined with CMPs. However, velocity is adjusted based on lithology. For example, the BEAN-CMP was collected on the modern beach, above a near surface water table, and used to calculate a velocity of 0.05 m/ns, a velocity typical of saturated sand (Jol and Bristow, 2003). However, since the majority of the BEANR profile was above the water table a velocity typical of dry sands (0.15 m/ns) was used. All other CMPs were collected along transects well above the water table and show a velocity of 0.1 m/ns which is the velocity used to estimate depth in all other GPR profiles. Profiles collected with 100 MHz antennae show reflections up to 10 m depth while profiles collected with 200 MHz antennae show reflections 5-8 m depth but with a greater resolution.
Profile were collected in a shore perpendicular orientation across the modern beach and the most lakeward beach ridge. GPR profiles consist of multiple reflections which are grouped into reflection patterns based on configuration, continuity, amplitude and terminations. Reflection patterns are described using radar stratigraphic terminology (van Heteren et al., 1998; Jol and Bristow, 2003), which is a modification of seismic stratigraphic terminology developed by Mitchum (1977). Reflection patterns are then grouped into radar facies. Radar facies are a mappable sedimentary unity composed of reflection patterns which differ from adjacent facies (Mitchum, 1977).

The GPR profile shown in Figure 12 resolves all radar facies common in the Ipperwash strandplain. Modern and ancient shorelines in the Ipperwash strandplain show five reflections patterns grouped into three radar facies (Figure 13). Reflection pattern A extends from 5-10 m depth and reflections consist of horizontal to lakeward dipping (apparent angle less than 10°), continuous reflections. Reflection pattern B extends for 0-5 m depth and consists of sigmoidal, concave up reflections which truncate radar facies A and approach an asymptote at ~5 m depth. Reflection pattern C is at 0-2 m depth and consists of a single horizontal continuous reflection. Reflection pattern D occurs at 0-2 m depth, is discontinuous separated by the relief of beach ridges and consists of lakeward dipping to landward dipping to horizontal continuous reflections. Reflection pattern E occurs at 0-2 m depth, is discontinuous separated by the relief of beach ridges and consists of undulating to landward dipping semi-continuous reflections.

GPR profiles are interpreted based on comparison to other GPR studies in similar environments (van Heteren et al., 1998; Johnston et al., 2007). Ipperwash profiles are limited to 10 m depth, however other 100 MHz GPR surveys in similar environments have resolved structures up to 37 m depth (Smith and Jol, 1995). Therefore, the depth of penetration at Ipperwash is likely limited. GPR signals are limited by sediments of high conductivity, such as fine grain sediments (Jol and Bristow, 2003). There are three potential sources of fine grain sediments in the Ipperwash area: shale bedrock, St Joseph till, and offshore silts. Any of these could potentially limit the depth of penetration, however, local well log records describe underlying material as “blue clay” at 11 m depth which suggest either St Joseph till or offshore silts. The lower radar facies consists of reflection patterns A and B and is interpreted as shoreface and foreshore sediments. Reflection pattern A is interpreted as progradational and aggradational beach sand and reflection pattern B is
interpreted as a ravinement surface. Reflection pattern C is interpreted as the water table which is commonly exposed at surface in the swales of beach ridge. The first upper radar facies consists of reflection pattern D and is interpreted as disturbed areas because the facies is observed under paths allowing access to the beach. The second upper radar facies consists of reflection pattern E is interpreted as dune sediments and is found internally under the crest of beach ridges. Radar facies are similar to radar facies described in other LGL strandplains; therefore, the model for LGL strandplain formation (Thompson and Baedke, 1995; Johnston et al., 2007) can be applied to Ipperwash.

Figure 12: Processed (BEAN-2) and interpreted (BEAN-B) GPR profile collected in a shore perpendicular orientation across the modern beach. Color scheme of interpretations shown in Figure 13.
Figure 13: Generalized radar stratigraphic cross section across two, beach ridges based on analysis of modern and ancient Ipperwash beach ridges. Vertical extravagation 3x: Radar reflection patterns are described and interpreted to represent dune, disturbed areas, water table, erosional surface, foreshore and shoreface sediments.

5.3 Topographic Survey

The topographic survey for this thesis collected elevations of ridge crests, swales, core surface location and OSL pit locations during the 2016 field season. The 2016 survey began at the modern shoreline and ended at core 2040 near the Ponderosa Pines Golf Course. The 2016 survey yielded results that decrease lakeward with a maximum of 186.5 m at 1772 m landward and a minimum of 179.3 m at 355 m landward. The transect consistently intersects the known peak Nipissing water level elevation of 183.3 m at the Port Huron/Sarnia outlet (Thompson et al., 2014) starting 800 m landward from the modern shoreline; however, previous research (Cooper, 1974; Johnston, 1999) indicates the peak Nipissing water level extended to a bluff 2.2 km landward from the modern Ipperwash shoreline. The discrepancy in the location of the peak Nipissing water level elevation lead to the comparison of the 2016 survey to other elevations measurements collected along the same transect (Figure 14) because accurate core elevations are needed to determine basal foreshore elevations and generate a paleohydrograph.

Two data sources (Johnston, 1999; southwestern Ontario orthophotography project, SWOOP2015) provide independent surface elevation data in the study area. Johnston (1999) used an optical level to survey a transect across the Ipperwash strandplain and adjusted ground surface ridge crest and swale elevation measurements to a nearby Geodetic Survey of Canada benchmark to convert measurements to a well-established LGL datum, IGLD85. SWOOP 2015 is a 2 m horizontal
resolution DEM (CGVD28) based on orthophotographs. Since SWOOP 2015 is based on photographs it only preserves the surface elevation of water bodies and uses a “steam-rolling” algorithm which “allowed for some raised features to be reduced closer to ‘bare-earth’ elevations (e.g. small buildings, small blocks of forest cover)” (SWOOP 2015 User Guide). The height of trees and any standing water must be accounted for when interpreting the SWOOP 2015 DEM. At Ipperwash, the landward portion of the strandplain is more densely forested and many of the swales contain between 0.1-1.0 m of standing water (Table 1).

The Johnston (1999) survey began at the then modern (1999) shoreline and only roughly parallels the 2016 survey. The topographic profile based on SWOOP 2015 was generated in QGIS along the same shore perpendicular line used to generate the 2016 topographic profile. However, the 2 m raster cell resolution of the DEM and “steam-rolling” algorithms simplifies and averages over an area larger than the Johnston (1999) survey. Lateral and vertical elevation differences in all three data sets were first visually assessed to determine where the transect elevations diverged (Figure 14). For all transects, ridge crest heights were used for comparison to negate the effect of standing water in SWOOP 2015 data. Crests location laterally vary by up to ten meters throughout the strandplain (Figure 14). The lateral offset is attributed to differences in horizontal data resolution.

Johnston (1999), SWOOP 2015 and 2016 survey elevations show good agreement up to 630 m distance landward, roughly plotting within a meter above or below one another (Figure 14). Johnston (1999) diverges from the 2016 survey elevations at 630 m landward consistently plotting ~1 m below 2016 elevations from 630-690 m and consistently plotting ~2.5 m below the 2016 elevations from 690 m to the landward margin of the survey. SWOOP 2015 elevations consistently plot ~3.5 m below the 2016 survey elevations from 660 m to the landward margin of the strandplain. Johnston (1999) elevations consistently plot ~1 m above SWOOP 2015 elevations from 660 m to the landward margin of the strandplain. The visual analysis suggests the elevations are consistently offset.

Core elevations from the 2016 survey and GPS based core locations overlaid on the SWOOP 2015 DEM were also compared to verify if elevations are consistently offset (Figure 15). Core elevations between the 2016 survey and the SWOOP 2015 DEM differ across the entire strandplain with a range of 5.4 m, a mean of 2.4 m and a standard deviation of 1.5 m (Figure 15). However, a clear
offset is noted. From 209-666 m landward, elevation differences range by 2.1 m with a mean difference of 0.2 m and a standard deviation of 0.6 m. From 699-1785 m landward, elevation differences range by 2.4 m about a mean difference of 3.3 m and a standard deviation of 0.6 m. Values on both sides of the offset show similar ranges and identical standard deviations, but have means that differ by 3.1 m showing that the elevations are consistently vertically offset.

Figure 14: Comparison of cross strandplain elevation profiles developed from 2016 surveyed elevations, Johnston (1999), and drawn from the SWOOP 2015 DEM. Also plotted are core elevations and the Nipissing elevation from Thompson et al. (2014). A discrepancy between 2016 elevations, Johnston (1999) and SWOOP 2015 is minimal up to 630 m distance landward but then diverges.

Figure 15: Comparison of the difference between core elevations determined from 2016 surveyed elevations and from the SWOOP 2015 DEM. Two groups of elevations (209-666 m and 699-1785 m landward) are consistently offset. Groups have a similar standard deviation and range but have means that differ by 3.1 m.

In order to adjust the 2016 survey which is needed to determine core elevations, the most realistic survey was determined. Johnston’s (1999) survey is considered the most realistic elevation survey because Johnston’s (1999) directly surveyed the ground surface. The Johnston (1999) survey is used
to adjust the 2016 survey that is consistently offset, most likely due to human or instrument error. The SWOOP 2015 survey is also adjusted to match the Johnston (1999) survey. The “steam-rolling” algorithm used in the SWOOP 2015 is suspected to have overestimated the height of forest cover in the landward portion of the strandplain.

All elevations are vertically adjusted to better match the Johnston (1999) survey. Consistency in the vertical offset of elevations allow for the adjustment of each dataset to better match one another by subtraction of a set value for a portion of the dataset (Figure 16). SWOOP 2015 elevations are adjusted by adding 1 m in elevation to all points landward of 660 m from the modern shoreline. 2016 surveyed elevations were adjusted by subtracting 1 m between 630 and 690 m landward from the modern shoreline (affecting 1 core, 2011) and subtracting 2.5 m for all points beyond 690 m landward from the modern shoreline (affecting 19 cores, 2012-2040). Once adjusted, all profiles show good agreement across the entire length of the survey (Figure 16). Adjusted 2016 survey elevations were used to determine core elevations and in all other analysis for this thesis.

![Adjusted topographic profile shows good agreement across the entire length of the 2016 survey. Profiles were adjusted by subtracting 1 m from 2016 surveyed elevations between 630 and 690 m landward and 2.5 m from 690 m landward to the landward margin of the survey. SWOOP 2015 elevations were adjusted by adding 1 m to points landward of 660 m.](image)

Figure 16: Adjusted topographic profile shows good agreement across the entire length of the 2016 survey. Profiles were adjusted by subtracting 1 m from 2016 surveyed elevations between 630 and 690 m landward and 2.5 m from 690 m landward to the landward margin of the survey. SWOOP 2015 elevations were adjusted by adding 1 m to points landward of 660 m.
5.4 Modern Shoreline Facies
Surface elevations across the modern beach were surveyed along the same transect as the cores (Figure 10). Surface sediment samples were collected along a 6 m transect beginning offshore and continuing ending on the foredune crest. Sediment samples were analyzed for grain size, sorting and skewness (Figure 17) using the University of Waterloo’s Fritisch laser diffraction system. The survey is divided into three facies, dune, foreshore and shoreface, based on surface topography, sedimentary structures, mean grain size, sorting and skewness. The foreshore extends from the swash zone (the area of breaking waves) to the maximum wave runup on the modern beach. Dune facies extend landward from foreshore facies and shoreface facies extend lakeward from the foreshore facies. The plunge point exists at the foreshore-shoreface contact and can be used as a lake level proxy for ancient shorelines (Thompson, 1992).

In total, surface elevations decrease lakeward 2.9 m, with a total elevation decrease of 2.4 m occurring over the landward most 4 m, across the dune and foreshore. A 1.0 m tall erosional scarp separates the lakeward dipping foreshore surface from the undulating dune surface 1.0 m landward from the plunge point. From the plunge point lakeward elevations decrease a maximum of 0.5 m, however, a 0.4 m fall is observed over the first 0.1 m, near to the plunge point.

![Figure 17: Plot of topographic survey and grain size results (mean, sorting and skewness) from the modern Ipperwash Beach showing the foreshore-shoreface contact as an abrupt change in grain size parameters.](image-url)
All modern beach samples consisted of predominately sand sized particles. However, variations occur between dune, foreshore and shoreface facies (Table 2). Shoreface sediments consist of fine sand (2.425-2.509 Φ), are poorly sorted (1.072-1.190), are very fine skewed (0.454-0.557 Φ) and contain ripple marks. Foreshore sediments contain gravel size shell fragments; however these sediments were sieved out and only those particles finer than 2 mm were analyzed because they consisted of mollusk shells (presumably invasive zebra or quagga mussels) not found in cores of ancient shorelines. Foreshore sediments consist of medium sand (1.168 Φ), are moderately sorted (0.875) and are fine skewed (0.152 Φ). Dune sediments consist of fine sand (2.688-2.652 Φ), are poorly sorted (1.006-1.009) and are very fine skewed (0.551-0.560 Φ).

<table>
<thead>
<tr>
<th>Topography</th>
<th>Shoreface Lakeward dipping</th>
<th>Foreshore None</th>
<th>Dune Undulating</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sedimentary Structures</td>
<td>Ripple marks</td>
<td>None</td>
<td>Bioturbation</td>
</tr>
<tr>
<td>Mean Grain Size</td>
<td>Fine sand</td>
<td>Medium Sand</td>
<td>Fine Sand</td>
</tr>
<tr>
<td>Sorting</td>
<td>Poorly sorted</td>
<td>Moderately sorted</td>
<td>Poorly sorted</td>
</tr>
<tr>
<td>Skewness</td>
<td>Very fine skewed</td>
<td>Fine skewed</td>
<td>Very fine skewed</td>
</tr>
</tbody>
</table>

### 5.5 Ancient Shoreline Facies

Forty cores were collected on the lakeward margin of each individual beach ridge in the Ipperwash strandplain except for the ridges nearest the modern shoreline which are modified by human activity (developed for permanent residence, paved roads, and dirt trails). Core depth of penetration ranges from 1.37-3.36 m with an average of 2.82 m. Sedimentary structures and color (determined by visual core descriptions, photographs and latex peels), and mean grain size, coarsest 1%, sorting, and skewness determine with the Indiana Geological Survey Malvern3000 laser diffraction system are used to define sediment facies. Cores were systematically interpreted and divided into dune, foreshore and shoreface facies based on changes in grain size statistics, visual descriptions, and color observed in core, photographs and latex peels (Appendix G). Core 2025 (Figure 18) is considered a typical core and described in detail to explain how sedimentary facies are defined and differentiated.
Figure 18: Visual descriptions and grain size statistics from core 2025. Shoreface sediments (274-163 cm), foreshore sediments (163-119 cm) and dune sediments (119-0 cm) are characterized by differing sedimentary structures and grain size parameter. The abrupt contact between foreshore and shoreface sediments is used as a proxy for the ancient lake level elevation when that beach ridge formed.

In core, shoreface facies are the lowest facies and typically include horizontal laminae, ripple marks and thin (1-2 mm wide) organic stringers and are noted as having a grey color. Statistically, the shoreface is characterized by a fine sand (mean grain size 2.76-2.72 $\Phi$) with the coarsest 1% of medium sand ($D(99)$ of 1.75 to 1.89 $\Phi$), is well sorted (sorting 0.41-0.45) and symmetrically skewed (skewness 0.95-0.96 $\Phi$).

Foreshore facies contain horizontal to lakeward dipping laminae and a basal bed of coarse grain mineralogic and shell (gastropod) fragments with an abrupt lower boundary and are noted as having a brown color. Statistically, the foreshore is characterized by a fine sand (mean grain size 2.54-2.19 $\Phi$) with the coarsest 1% normally grading from fine to very coarse sand ($D(99)$ of 1.52 to 0.57 $\Phi$), from well sorted to moderately well sorted (sorting 0.46-0.98) and from symmetric to coarse skewed (skewness 0.95-0.96 $\Phi$).
Dune facies are the uppermost facies and typically contain massive sand and noted as having a light brown color. Statistically, the dune is characterized by fine sand (mean grain size 2.75-2.54 Φ) with the coarsest 1% of medium sand (D(99) of 1.82 to 1.52 Φ), is well sorted (sorting 0.43-0.46) and is symmetric to fine skewed (skewness 0.24-0.02 Φ). The foreshore dune contact is typically gradational, however, the appearance of the laminae is considered the top of the foreshore when calculating foreshore thickness because dune sediment is commonly structureless.

In summary, the shoreface, foreshore and dune facies are identified based on sedimentary structures, color (determined by visual core descriptions, photographs and latex peels), and grain size parameters: mean grain size, coarsest 1%, sorting, and skewness (Table 3). Shoreface sediments are typically fine, well sorted and contain poorly to well defined laminae and is typically grey in color. Foreshore sediments are normally graded, moderately to well sorted, lakeward dipping to horizontal laminated with brown color and a basal layer of coarse to medium grain shell and mineralogic fragments. Dune sediments are fine grained, well sorted, structureless and light brown in color. Of the 40 cores collected, 36 yielded shoreface-foreshore contacts (basal foreshore depths). The 4 cores which did not yield basal foreshore depth likely did not penetrate deep enough to reach the basal foreshore. The 36 basal foreshore depths were subtracted from core location elevations to yield basal foreshore elevations (the approximate elevation of past lake levels when each beach ridge formed).

Table 3: Common characteristics used to differentiate core sediment facies

<table>
<thead>
<tr>
<th>Sedimentary structure</th>
<th>Mean grain size</th>
<th>Coarsest 1%</th>
<th>Sorting</th>
<th>Skewness</th>
<th>Color</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dune</td>
<td>Structureless</td>
<td>Fine sand</td>
<td>Medium Sand</td>
<td>Well sorted</td>
<td>Symmetrical to fine skewed</td>
</tr>
<tr>
<td>Foreshore</td>
<td>Lakeward dipping to horizontal laminae with basal bed of shell and mineralogic fragments</td>
<td>Fine sand</td>
<td>Normally grading from medium to very coarse sand</td>
<td>Well to moderately sorted</td>
<td>Symmetrical to coarse skewed</td>
</tr>
<tr>
<td>Shoreface</td>
<td>Horizontal laminae, ripple marks, organic stringers</td>
<td>Fine sand</td>
<td>Medium Sand</td>
<td>Well sorted</td>
<td>symmetrical</td>
</tr>
</tbody>
</table>
5.6 Cross Strandplain Geomorphic and Sedimentological Characteristics

Across the Ipperwash strandplain, changes occur in basal foreshore elevations, foreshore thickness, foreshore average grain size, beach ridge width (core to core spacing), and ridge height (swale-crest difference). These changes are quantified and statistically described.

Ground surface elevations (Figure 19) show a net increase from the lakeward to landward margin of the strandplain from a minimum of 178.1 m at the bottom of the Ipperwash Drain 700 m landward from the modern shoreline to a maximum of 184.0 m on the crest of the landward most ridge. The most lakeward ~3 beach ridges are not included because of intense human modification and were not cored. The topographic survey and core facies (Figure 19) were used to quantify the geomorphic and stratigraphic characteristics of the strandplain respectively (Table 4).

Figure 19: Surface elevation, core sediment facies and OSL ages from the Ipperwash strandplain. All data is plotted with respect to distance from the modern shoreline with ridge numbers labelled.

Basal foreshore elevations (Figure 19; Table 4) have a mean of 179.2 m with a standard deviation of 0.7, but show a net increase landward (slope 0.0006, y-intercept 178.3 ± 0.4) with a maximum near the landward margin of the strandplain (180.99 m at 1746 m landward) and a minimum near the lakeward margin of the strandplain (177.8 m elevation at 312 m landward from modern shoreline). Foreshore thicknesses (Figure 19; Table 4) have a mean thickness of 0.9 m with a standard deviation of 0.4 with a relatively horizontal trendline (slope 0.00005, y-intercept 0.7 ± 0.2). The maximum foreshore thickness is 2.0 m thick at 1637 m landward from modern shoreline and the minimum is
0.2 m thick at 817 m landward from modern shoreline. Foreshore average grain sizes (Figure 19; Table 4) have a mean of 2.0 Φ phi with a standard deviation of 0.1, but show a net increase landward (slope 0.0002, y-intercept 2.4 ± 0.1) with a maximum of 2.7 Φ at 873 m landward and a minimum of 2.0 Φ at 312 m landward.

Beach ridge width (Figure 19; Table 4) have a mean of 40.4 m with a standard deviation of 0.7 and show a net decrease landward (slope -0.009, y-intercept 50.8 ± 8.1), but have a maximum (91.1 m at 1043 m landward) and minimum (7.5 m at 1239.7 m landward) near the center of the strandplain. Ridge heights (Figure 19; Table 4) have a mean of 1.4 m with a standard deviation of 19.7 and have a landward decreasing trendline (slope -0.00003, y-intercept 1.4 ± 0.3), but have a maximum (3.0 m height at 603 m landward) and minimum (0.02 m at 627 m landward) near the center of the strandplain.

<table>
<thead>
<tr>
<th>Basal Foreshore Elevation (m)</th>
<th>x-value of (maximum y-value)</th>
<th>x-value of (minimum y-value)</th>
<th>Mean</th>
<th>Standard Deviation</th>
<th>Trend line equation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Basal foreshore elevations and foreshore thickness have lakeward sloping trendlines through all data points. Foreshore average grain size, beach ridge width and ridge height have relatively horizontal trendlines which vary within 1 standard deviation of the mean. No visually apparent correlation is observed between quantified geomorphic and sedimentological characteristics.</td>
<td>Basal foreshore elevations and foreshore thickness have lakeward sloping trendlines through all data points. Foreshore average grain size, beach ridge width and ridge height have relatively horizontal trendlines which vary within 1 standard deviation of the mean. No visually apparent correlation is observed between quantified geomorphic and sedimentological characteristics.</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
5.6.1 Ridge Groups and Interpretations

A pattern is noted in the variations of parameters above and below one standard deviation (Figure 20). Groups of 3 to 7 ridges are defined by variations above and below 1 standard deviation. Patterns in beach ridge groupings are interpreted to represent relatively short-term fluctuations in lake level likely driven by climate (Baedke and Thompson, 2000). The five geomorphic and sedimentological parameters (Table 4) are described in regards to what past conditions the parameter likely record.

Basal foreshore elevation is a record of lake level at the time of deposition (Thompson, 1992). The lakeward sloping trendline (Figure 20) suggests that lake level has undergone a net regression during the deposition of the Ipperwash strandplain. However, the effects of GIA contributing to this relative, long-term lake level fall in elevation will be discussed in detail later.

Foreshore thickness is a factor of wave energy/height (Howard and Reineck, 1981) and wave energy/height is a factor of wind direction, speed, fetch (Komar, 1998) collectively these parameters are referred to as wave climate (Johnston et al. 2007). Wave climate describes the average condition of waves in a certain location at a certain time, and foreshore thickness is related to wave climate when a specific beach ridge was deposited (Johnston et al., 2004).

Foreshore average grain size shows similar patterns to foreshore thickness (Figure 20) indicating a change in wave climate (Fox et al., 1966; Komar, 1998) or a change in sediment supply (Johnston et al., 2007). The close agreement between foreshore thickness and foreshore average grain size at Ipperwash is interpreted as a record of a climate oscillation.

Beach ridge width is related to shoreline behavior. Simplistically, beach ridge width is a factor of the rate of lake level change and sediment supply as well as predepositional slope (Thompson and Baedke, 1995; Figure 5). A wide beach ridge can therefore be thought to represent a period of relatively significant progradation, and a narrow ridge can be though to represent a period of relatively minor progradation. However, due to varying erosion rates along the coast even a narrow ridge may be the result of a period of significant progradation of the coast and subsequent erosion removed much of the progradational record. Beach ridge groupings based on beach ridge width are well expressed across the entire strandplain and suggest an oscillation in shoreline behavior across several beach ridges during the deposition of the Ipperwash strandplain (Figure 20).
Beach ridge height is also related to shoreline behavior and variable dune cap (Figure 5). Since the core of a beach ridge is built by aggradation the height of a beach ridge is a factor of how long the shoreline aggrades (Figure 5). However, the presence of the dune cap complicates matters. Wind-blown sands accumulate on top of the beach ridge and increase the height of the beach ridge. Beach ridge height shows well defined beach ridge groups lakeward of 1200 m (Figure 20), this suggests oscillations in shoreline behavior and/or climate during the deposition of the Ipperwash strandplain.
Figure 20: Geomorphic and sedimentologic characteristics across the Ipperwash strandplain showing aerial imagery with OSL sample location labeled. Eight segments (A-H) were qualitatively identified based on field observations. Groups of 3 to 7 ridges identified as rises and falls below 1 standard deviation are also identified and bracketed. Satellite imagery from Google Earth.
5.7 Age Modelling

Ten OSL ages (in calendar years before 2017 CE) were collected approximately 1 m below the ground surface of ridge crests on every third to fifth ridge. OSL samples were collected in an attempt to bracket ridge segments based on field observations (Figure 20). Error is reported as the standard error of equivalent dose (Dₑ) distributions divided by the environmental dose rate (Figure 21). This method of error reporting allows for a more direct comparison of OSL measurement to other dating methods (such as radiocarbon) by placing emphasis on the variability of OSL Dₑ measurements (Lepper et al., 2011). Propagated age uncertainty (Aitkens, 1985), which takes into account geologic uncertainties is also reported (Figure 21).

Ages sequentially decrease from a maximum landward most (sample 2038, collected 1746 m landward from the modern shoreline) age of 3490 ± 110 to a minimum landward most (sample 2001, collected 209 m landward from the modern shoreline) age of 650 ± 20 years (Figure 19; 21).

Beach ridge number and distance the from modern shoreline were recorded for each OSL sample location. When beach ridge number is plotted against age, the average rate of beach ridge development is calculated. When distance landward from the modern shoreline is plotted against age, the average rate of long-term progradation is calculated. Distance landward is used to create the Ipperwash paleohydrograph because 1) beach ridges are variably spaced across the strandplain so plotting the ages against beach ridge number misrepresents the data and 2) during surveying it is possible that beach ridges were missed, buried, misinterpreted (i.e. dunes ridges misinterpreted as beach ridges), or discontinuous meaning the sampled ridges may over or under represent the true amount of beach ridges.
Figure 21: OSL element data used for dose rate calculation and OSL age results (Modified from Lepper, 2017).

Age models were created to extrapolate the 10 OSL ages collected in the Ipperwash strandplain to assign ages to the 36 recovered basal foreshore elevations used to create the Ipperwash paleohydrograph. To identify outliers and produce the most representative and realistic age model for the Ipperwash strandplain, two age models were created.

Age model A was developed as an initial examination of a linear relationship between individual ages connected sequentially (Figure 22). Slopes are calculated from the mean sample age of individual ages. Age model A shows slopes (in years per meter) ranging from 0.7 to 4.5 with an average slope between all 10 ages of 1.9 years/m ± 1.1 years. The maximum and minimum slope between ages occurs between the landward most three ages. And the lakeward most four ages have a consistent slope between ages, within 0.4 year/m of each other. This shows that slopes (i.e. the
average progradation rate) are most variable in the landward most portion of the strandplain and least variable in the lakeward portion of the strandplain (Figure 22).

**Figure 22:** Age model A connects all individual ages sequentially and shows slopes between individual ages. The average slope is 1.9 years/m and a standard error of 1.1.

Age model B uses all ten ages in a single linear regression (Figure 23). For age model B $r^2$ is 0.98, with a slope of $1.8 \pm 0.09$ years/m + $98.5 \pm 107.0$ years. Significance F is $6.31 \times 10^{-8}$ and the average of all residuals is 98.9 years. All age error bars lie at least partially within the confidence intervals and wholly within the prediction intervals (Figure 23). The $r^2$ and confidence intervals suggest this age model is a good representation of the data. Ages within age model B are generally younger than reported ages in age model A (Table 5).

**Figure 23:** Age model B uses a linear regression through all ages. The resultant equation is $y = 1.8 \pm 0.09x + 98.5 \pm 107.0$ and has a $r^2$ is 0.98.
### Table 5: Reported and modelled ages

<table>
<thead>
<tr>
<th>Ridge Number</th>
<th>Distance from modern shoreline (m)</th>
<th>Reported Age/Age Model A</th>
<th>Age Model B</th>
<th>Age Model A – Age Model B</th>
</tr>
</thead>
<tbody>
<tr>
<td>01</td>
<td>209.2</td>
<td>650 ± 20</td>
<td>475</td>
<td>175</td>
</tr>
<tr>
<td>05</td>
<td>412.6</td>
<td>860 ± 40</td>
<td>841</td>
<td>19</td>
</tr>
<tr>
<td>09</td>
<td>602.6</td>
<td>1120 ± 40</td>
<td>1183</td>
<td>-63</td>
</tr>
<tr>
<td>14</td>
<td>816.9</td>
<td>1410 ± 60</td>
<td>1568</td>
<td>-158</td>
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<tr>
<td>18</td>
<td>1042.6</td>
<td>1860 ± 60</td>
<td>1975</td>
<td>-115</td>
</tr>
<tr>
<td>22</td>
<td>1151.1</td>
<td>2180 ± 80</td>
<td>2170</td>
<td>10</td>
</tr>
<tr>
<td>25</td>
<td>1232.3</td>
<td>2300 ± 80</td>
<td>2317</td>
<td>-17</td>
</tr>
<tr>
<td>30</td>
<td>1412.0</td>
<td>2650 ± 90</td>
<td>2640</td>
<td>10</td>
</tr>
<tr>
<td>34</td>
<td>1587.7</td>
<td>2780 ± 100</td>
<td>2956</td>
<td>-176</td>
</tr>
<tr>
<td>38</td>
<td>1746.4</td>
<td>3490 ± 110</td>
<td>3242</td>
<td>248</td>
</tr>
</tbody>
</table>

#### 5.8 Developing the Ipperwash Paleohydrograph

An Ipperwash relative paleohydrograph can be developed by using measured elevations and modelled ages. Measured basal foreshore elevations (the ancient lake level proxy) are determined by analysis of vibracores from individual beach ridges. OSL ages are used to create an age model to assign ages to individual beach ridges. Since both basal foreshore elevations and modelled ages are plotted against distance landward from the modern shoreline, distance landward is used to relate elevation and age data.

Two Ipperwash relative paleohydrographs, one based on age model A (paleohydrograph A) and the other based on age model B (paleohydrograph B), were developed. Paleohydrographs are visually similar (Figure 24). F-tests, which test the equality of variances (Davis, 2002), are used to statistically compare paleohydrographs A and B. F-tests show the F-statistic to be less than the F Critical one-tail value; therefore, the null hypothesis is accepted, and the variance of the two populations is statistically similar (Table 6).
Figure 24: Ipperwash paleohydrographs developed from age model A and B.

### Table 6: F-Test Two-Sample for Variances

<table>
<thead>
<tr>
<th></th>
<th>Paleohydrograph A</th>
<th>Paleohydrograph B</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>1973.35</td>
<td>2156.63</td>
</tr>
<tr>
<td>Variance</td>
<td>559372.90</td>
<td>558099.12</td>
</tr>
<tr>
<td>Observations</td>
<td>34</td>
<td>34</td>
</tr>
<tr>
<td>df</td>
<td>33</td>
<td>33</td>
</tr>
<tr>
<td>F-Statistic</td>
<td>1.002282357</td>
<td></td>
</tr>
<tr>
<td>P(F&lt;=f) one-tail</td>
<td>0.497407417</td>
<td></td>
</tr>
<tr>
<td>F Critical one-tail</td>
<td>1.787821747</td>
<td></td>
</tr>
<tr>
<td>F-Statistic &lt; F critical one-tail</td>
<td></td>
<td>Variance of two populations equal</td>
</tr>
</tbody>
</table>

Though the age models are statistically similar, geologically the age models have different implications. Age model A implies a variable long-term rate of progradation through time. Age model B implies a constant long-term progradation rate for the Ipperwash strandplain.

Age model B is used to reconstruct the most realistic Ipperwash paleohydrograph because 1) slopes between individual ages in age model A are similar, with most slopes falling within 2.2 year/m of one another suggesting a linear relationship across the entire strandplain, 2) age model A does not include the landward most two ridges since the landward most OSL sample was collected on ridge 38 (1746 m landward from the modern shoreline), 3) a linear model serves as a good, simple approximation of beach ridge age for the Ipperwash strandplain, and 4) age model B allows for the comparison to other LGL strandplain paleohydrographs since other paleohydrographs used linear age models (Baedke and Thompson, 2000; Johnston et al., 2012).
Chapter 6

The Ipperwash paleohydrograph

Thirty-six basal foreshore elevations and ten OSL ages were combined to create the Ipperwash paleohydrograph. Patterns within the Ipperwash paleohydrograph are described and interpreted in the context of known drivers of lake level change during historic times (International Upper Great Lakes Study Board, 2009; 2012) and geologic times (Baedke and Thompson, 2000; Johnston et al., 2012).

The Ipperwash paleohydrograph chronicles relative lake level fluctuations from 3520 to 710 years ago (Figure 25). The maximum relative lake level is 181.0 m which occurred 3450 years ago and the minimum relative lake level is 177.8 m which occurred 850 years ago. Prehistoric lake level elevations are higher than historic lake level elevations. Historic (since 99 years ago or 1918 CE) lake levels in Lake Huron have fluctuated between a lake-wide yearly average high of 177.3 m 31 years ago (or 1986 CE) and a lake-wide yearly average low of 175.7 m 53 years ago (or 1964 CE) (Gronewold et al., 2013a). Recent International Joint Commission Studies have concluded historic lake level changes has been driven by GIA, outlet conveyance, and climate with the largest contributing factor being climate (International Upper Great Lakes Study Board, 2009; 2012). GIA, outlet conveyance and climate are also important drivers of prehistoric lake level fluctuations (Johnston et al., 2014). Therefore, GIA, outlet conveyance, and climate are briefly reviewed to give context to interpretations of patterns within the Ipperwash paleohydrograph. It should be noted that though the terms millennium, centennial, and decadal are used these terms should not be taken as an exact length of time, but rather as a rough estimate of natural patterns observed through geologic time.

The Ipperwash paleohydrograph is compared to the Lake Michigan paleohydrograph which reflects lake levels at the Port Huron/Sarnia outlet (Figure 4; 26). The Lake Michigan paleohydrograph is an outlet-referenced paleohydrograph created by combing strandplain data from 5 sites around Lake Michigan and adjusting overlapping strandplain paleohydrographs to reconstruct lake level fluctuations in one hydrologically connected lake (Lake Michigan-Huron) at the Port Huron/Sarnia outlet. The Lake Michigan paleohydrograph shows that following the Nipissing highstand, 4,500 cal BP, lake level underwent a rapid fall of 4.1 m until 3,400 cal BP (Baedke and Thompson, 2000). The end of the rapid fall is attributed to the abandonment of the Chicago outlet (Baedke and
Thompson, 2000). Baedke and Thompson (2000) also propose lake levels rose and fell on a millennial rhythm over the next several thousand years. Lake levels rose from 3,300 to 3,000 cal BP, associated with the Algoma highstand, fell from 2,400 to 2,250 cal BP, and rose from 2,100 to 1,700 cal BP, associated with the 1,700 high, and fell from 1,700 to 1,000 cal BP (Figure 4; 26).

Ipperwash is the nearest preserved strandplain with the most beach ridges to the Port Huron/Sarnia outlet. The rate of GIA between the Ipperwash strandplain and the Port Huron/Sarnia outlet is expected to be similar (Figure 2; Mainville and Craymer, 2005). The Ipperwash paleohydrograph and the outlet referenced Lake Michigan paleohydrograph are expected to be similar as both reconstruct lake levels experienced at the Port Huron/Sarnia outlet. However, the Ipperwash paleohydrograph consistently plots above the Lake Michigan paleohydrograph (Figure 26). Johnston et al. (2012) and Thompson et al. (2014) suggested too much GIA was collectively removed when creating one outlet-referenced paleohydrograph for Lake Michigan. This thesis reevaluates the Lake Michigan paleohydrograph using the Ipperwash paleohydrograph.

The Ipperwash paleohydrograph is also compared to the historic lake level record for Lake Huron to determine the relationship between historic and prehistoric lake levels which can be used to more accurately predict potential future lake level changes at Ipperwash.

Figure 25: Ipperwash paleohydrograph compared to Lake Huron’s annual average historic lake level. The Ipperwash paleohydrograph records lake level fluctuations from 3520 to 710 years ago and shows a net lake level decrease from a high of 181.0 m 3450 years ago to a low of 177.8 m 850 years ago.
Figure 26: Historic hydrograph, Ipperwash paleohydrograph and the Lake Michigan paleohydrograph (Baedke and Thompson, 2000). If the Lake Michigan paleohydrograph is adjusted based on the rate of GIA at Ipperwash the two graphs plot more closely.

6.1 GIA, outlet conveyance and climate

GIA is the ongoing rate of land surface elevation change (expressed in cm/century) during and following glacial advances and retreats which influence relative lake level change in the LGL. Various methods have been used to estimate the rate of GIA in and around the LGL. Geologic data has been used in some areas of the LGL to estimate rates of GIA in a single basin relative to that basin's outlet (Lewis, 1970; Baedke and Thompson, 2000; Johnston et al., 2012). The analysis of geologic data suggests a linear rate of GIA over at least the past 3500 years and is interpreted as the longest pattern in LGL paleohydrographs (Johnston et al., 2014). Various geologic and GPS datasets have been used to model the absolute rate of GIA at a continental scale (Peltier et al., 2015), and these models have been further refined in the LGL through the inclusion of the lake level gauge record since 1918 CE (Mainville and Craymer, 2005; Figure 2) which forms the basis for rates of GIA in used in international management plans (International Upper Great Lakes Study Board, 2009; 2012). Though records vary on the exact rate of GIA at a precise location, all records show the same general pattern of GIA, the land surface rising (up to 54 cm/century) in the northern LGL and subsiding (down to -27 cm/century) in the southern LGL. Based upon the most recent GIA models at continental scales (Peltier, et al., 2015) and over the LGL (Mainville and Craymer, 2005) the rate of GIA is zero cm/century near the Port Huron/Sarnia outlet relative to a point representing the center of the Earth. Ipperwash is located 40 km from the Port Huron/Sarnia outlet, and is also estimated to have a rate of GIA of zero cm/century. In other words, the ground surface at
Ipperwash is not moving up or down relative to a fixed point in the center of the Earth and is identical to the ground surface at the Port Huron/Sarnia outlet through geologic time.

Outlet conveyance is a change in the water carrying capacity of a lake's outflow channel(s) due to channel activation or abandonment, or when the active channel experiences erosion or sedimentation. For example, Hough (1958) suggested the fall from the peak Nipissing, approximately 4500 years ago, was a result of natural erosion at the Port Huron/Sarnia outlet, and Johnston et al., (2007) suggested the outlet at Sault Ste Marie began regulating Lake Superior's lake level at 1100 calendar years ago. During historic times only a relatively small portion of the lake level change has been attributed to erosion, caused by dredging, in the Port Huron/Sarnia outlet (International Upper Great Lakes Study Board, 2009). These examples show the importance of a lake's outlet in regulating lake level in the LGL. A recent study by Campbell (2016) initiated investigation of natural changes in outlet conveyance preserved in three depositional coastal landforms (spits) located in the Port Huron/Sarnia outlet that were deposited sometime in the last 4500 years. Since the Port Huron/Sarnia spits lie at a similar elevation to the beach ridges in the Ipperwash strandplain (i.e. between the modern lake and the peak Nipissing), deposition of the spits likely occurred between 3520 and 710 years ago, the period of record in the Ipperwash paleohydrograph. The ages and subsurface stratigraphy of the Port Huron/Sarnia spits is therefore very important to determine past natural times of outlet conveyance caused by longshore drift and sedimentation in the Port Huron/Sarnia outlet that may have restricted outflow from Lake Michigan-Huron. No studies have investigated outlet conveyance caused by sedimentation in the Port Huron/Sarnia outlet. The Ipperwash paleohydrograph provides the context needed to interpret the Port Huron/Sarnia spits and how the spits may relate to natural outlet conveyance through geologic time.

Climate also influences LGL lake levels. Historic data and paleoclimate data suggests lake level fluctuate on the order of many decades (Argyilan and Forman, 2003; Gronwold and Stow, 2014) and many centuries (Fraser et al., 1990) and may be related to climate. Warm and dry climate have been related to low lake levels during historic (Argyilan and Forman, 2003; Gronewold and Stow, 2014) and prehistoric times (Fraser et al., 1990; Lewis et al., 2008a). Cool and wet climate have been related to high lake levels during historic (Gronwold et al., 2016) and prehistoric times (Fraser et al., 1990; Booth et al., 2002). Lows and highs in the Ipperwash paleohydrograph may therefore record at least
a partial climate record. Analysis of 40 paleoclimate records (Shuman and Marsieck 2016) show mid-latitude North America (i.e. the belt in which the LGL predominately lie) experienced relatively dry and warm conditions from 2900 to 2100 cal BP then a rapid transition to relatively wet and cool climate conditions which last until 1800 cal BP.

GIA, outlet conveyance and climate all affect lake levels, but with different temporal and lateral scales in a lake basin (Johnston, et al. 2014). GIA is an ongoing, linear, geologic process unique to different sites around a lake basin. Outlet conveyance also occurs over historic and prehistoric time scales but should be experienced and preserved at all strandplain sites around a lake basin. Similar to outlet conveyance, climate naturally changes on time scales ranging from decades to millennium and effects lake levels in an entire lake basin.

The Ipperwash paleohydrograph is analyzed to discern trends and patterns in lake level fluctuations. Multi-millennium, millennium, centennial, and multi-decadal patterns within the Ipperwash paleohydrograph are interpreted to represent GIA, outlet conveyance, and/or climate.

6.2 Multi-millennium pattern

A long term, multi-millennium pattern identified by using all 36 data points in the Ipperwash paleohydrograph is simplistically described by a single linear regression. The resultant equation is $y = 0.0007x + 177.8$ and $r^2$ is 0.56 (Figure 27). Standard error of the slope is 0.001 and standard error of the y-intercept is 0.2. The slope of the trend line suggests a net lake level fall at a rate of 7 cm/century with oscillations above and below the trend line. The multi-millennium pattern shows a net, linear, relative lake level fall of 1.8 m from 3520 to 710 years ago.
Figure 27: A linear regression through the entire paleohydrograph shows a relatively well confined pattern spanning the entire paleohydrograph.

GIA is regarded as the driver of the longest trend in LGL paleohydrographs (Johnston, et al., 2014) and is therefore interpreted as the dominant driver of the multi-millennium lake level lowering at Ipperwash. The multi-millennium linear pattern calculated form Ipperwash beach ridges suggests a rate of GIA near $7 \pm 1$ cm/century. However, the rate of GIA, based on the current understanding in published literature, at Ipperwash is expected to be 0 cm/century (Mainville and Craymer, 2005). This difference in rates of GIA at Ipperwash could stem from three reasons or a combination of the three reasons. 1) Interpolation between water gauge stations underestimates the rate of GIA at Ipperwash. This could stem from insufficient water gauges near Ipperwash (the closest water gauge station is located at Lakeport, Michigan, approximately 43 km from Ipperwash) to accurately resolve the rate of GIA at Ipperwash. 2) The rate of lake level lowering at Ipperwash could stem from erosion at the Port Huron/Sarnia outlet during the deposition of the Ipperwash strandplain. If this is the case, all strandplains around Lake Michigan-Huron should reflect this rate of lake level lowering. However, strandplains in Lake Michigan show a varying rate of lake level change through geologic time (Thompson and Baedke, 1997). In addition, recent studies have found no active erosion in the Port Huron/Sarnia outlet in historic times (International Upper Great Lakes Study Board, 2009). 3) An outlet other than Port Huron/Sarnia was active and dominant during the time period recorded in the Ipperwash paleohydrograph. The Chicago outlet is the most likely outlet to be active during the deposition of the Ipperwash strandplain and interpretation of the millennium patterns within the Ipperwash paleohydrograph (discussed below) may support the dominance of the Chicago outlet during the time period recorded in the Ipperwash paleohydrograph.
Interestingly, applying a consistent rate of 7 cm/century to the outlet-referenced Lake Michigan paleohydrograph (Baedke and Thompson, 2000) helps the Lake Michigan and Ipperwash paleohydrographs match more closely (Figure 26). This argues for a readjustment of the Lake Michigan paleohydrograph to better represent the conditions experienced at the Port Huron/Sarnia outlet and is the first verification of an idea presented in previous studies which suggested too much GIA was removed when creating the Lake Michigan paleohydrograph (Johnston et al., 2012; Thompson et al., 2014).

When the multi-millennium long, linear pattern is extended to the present, the confidence interval minimum (177.3 m) intersects the historic yearly average lake level high (177.3 m) 31 years ago or 1986 CE (Figure 27). The intersection of the lower confidence interval of the multi-millennium pattern in the Ipperwash paleohydrograph and the historic annual Lake Huron lake level high, may support that the multi-millennium pattern is represented in the historic record. The multi-millennium pattern extended above the historic record potentially because subsurface elevation measure in Ipperwash beach ridges record multi-decadal lake level highs (Thompson and Baedke, 1995; Johnston et al., 2007). To equate the historic and prehistoric lake level records the multi-decadal lake level high (occurring in 1986 CE or 31 years ago) is close to the statistical range calculated from Ipperwash beach ridges (Figure 27).

6.3 Millennium patterns

Field observations and geomorphic and sedimentologic data suggests a break between a landward and lakeward set of beach ridges. The landward set of Ipperwash beach ridges consists of cores 2040 - 2014 (1785 – 817 m landward from the modern shoreline) and is characterized by swales filled with shallow to deep water (0 – 1 m deep), well defined beach ridge groups based on foreshore thickness and average grain size, and beach ridge width that is typically narrower than the lakeward set of ridges (Table 1; Figure 20). The lakeward set of Ipperwash beach ridges consists of cores 2013 – 2001 (737 to 209 m landward from the modern shoreline) and is characterized by either dry swales or swales filled with deep water (~1 m deep), well defined beach ridge groups based on ridge
height, and beach ridge with that is typically wider than the landward set of ridge (Table 1; Figure 20).

When the Ipperwash paleohydrograph is divided based on field observations, geomorphic and sedimentological evidence, two lowering patterns are observed each lasting approximately 1300 years. Patterns are vertically offset by 0.5 m over 160 years. These patterns are termed millennium patterns for the sake of simplicity and to differentiate them from the multi-millennium patterns which spans several millennium. The oldest period of relative lake level record lasts from 3520 to 2180 years ago, has an equation of $y = 0.001x + 175.9$ with an $r^2$ of 0.76. Standard error of the slope is 0.0002 and standard error of the y intercept is 0.5. The youngest period of relative lake level record lasts from 2020 to 710 years ago, has an equation of $y = -0.0006x + 177.9$ with an $r^2$ of 0.19. Standard error of the slope is 0.0004 and standard error of the y intercept is 0.5. The two patterns are vertically offset by 0.5 m.

![Figure 28: Ipperwash paleohydrograph with GLA removed and divided into two linear regressions which suggests two phases of lake level lowering. The two periods of lake level lowering are vertically offset by 0.5 m and a period of 160 years.](image)

The millennial pattern in the Ipperwash paleohydrograph suggests two periods of oscillatory relative lake level fall separated by a 160 period (between 2180 and 2020 years ago) showing a vertical offset of 0.5 m between the millennium trends. The oldest period lasted from 3520 to 2180 years ago and resulted in a mean net lake level fall of 1.7 m. The youngest period lasted from 2020 to 710 years ago and resulted in a mean net lake level fall of 0.8 m.
The millennial patterns compare with lake phases in the Lake Michigan paleohydrograph. The oldest millennium trend in the Ipperwash paleohydrograph roughly relates to the Algoma phase in Lake Michigan strandplain data using radiocarbon dates from 3400 to 2300 cal BP (Baedke and Thompson, 2000; Figure 26) and Lake Superior strandplain data using OSL dates from 2800 to 2000 cal years BP (Johnston et al., 2012). A vertical offset between the two millennium patterns in the Ipperwash paleohydrograph is observed over a period lasting from 2180 to 2020 years ago. This vertical offset between millennium trends at Ipperwash may relate to rising water levels during the transition from the Algoma phase to the 1700-high phase recorded in strandplain data of Lake Michigan (Baedke and Thompson, 2000; Figure 4) and the Algoma and Sault phases recorded in strandplain data of Lake Superior (Johnston et al., 2012). Farrand (1962) suggested Lake Superior separated from Lake Michigan-Huron at this time of low water level and Lake Superior became its own lake, as it is today elevated by a bedrock sill above Lake Michigan-Huron. But Johnston et al. (2012) suggests this time period only represented a short time period of separation and Lake Superior's final separation occurred closer to 1100 calendar years ago.

Extrapolating from published reports of a relationship between climate and lake level in geologic (Fraser et al., 1990; Lewis et al., 2008a) and historic (Argyilan Forman, 2003; Gronewold et al., 2016) times in the LGL, falls in lake level may be related to a relatively dry and warm climate and lake level rises may be related to wet and cool climate. To investigate this relation further the paleoclimate record for the LGL is examined. Shuman and Marsicek (2016) show mid-latitude North America experienced relatively dry and warm conditions from 2900 to 2100 cal BP and relatively wet and cool conditions from 2100 to 1800 cal BP with rapid changes occurring at 2100 cal BP. The period of the Ipperwash paleohydrograph lasting from 3520 to 2180 years ago is a period of lake level fall and corresponds with dry and warm conditions from 2900 to 2100 cal BP. Additionally, from 2180 to 2020 years ago the Ipperwash paleohydrograph records a relative lake level rise between the two millennium trends which corresponds with wet and cool conditions lasting from 2100 to 1800 cal BP. Since time periods of lake level falls and rises correspond with periods of dry and warm, and wet and cool climates respectively, climate is a possible dominant driver of millennium patterns in relative lake levels.

Alternatively, part of the two millennium trends may be a result of a long-term change in outlet conveyance that can be resolved through evaluation of rates of GIA between the study site and the
active outlet during certain time periods. A long-term lake level lowering in a paleohydrograph indicates the ground surface at the active outlet is not rising as fast as the study site. Using the pattern of GIA of Mainville and Craymer (2005), the long-term lowering would suggest that the Chicago outlet could be the dominate outlet when part of the Ipperwash strandplain formed. The rate of GIA at the Chicago outlet of Lake Michigan-Huron is falling at a rate of ~12 cm/century relative to Ipperwash and the Port Huron/Sarnia outlet (Mainville and Craymer, 2005; Figure 2). This rate of GIA matches the long-term trend of the oldest millennium trend (10 ± 2 cm/century) which may indicate that the Chicago outlet regulated lake levels in Lake Michigan-Huron during the oldest millennium trend recorded in the Ipperwash paleohydrograph. Previous work also indicates the Chicago outlet of Lake Michigan-Huron was completely abandoned around 2400 cal BP (Chrzastowski and Thompson, 1992) which roughly corresponds with the end of the oldest millennial trend at Ipperwash, 2180 years ago.

The youngest millennium linear lowering trend in the Ipperwash paleohydrograph, lasting from 2020 to 710 years ago with a rate of 6 ± 4 cm/century, also falls within the possible range of the rate of GIA between the Chicago outlet and Ipperwash. It is therefore possible that Chicago was also active during the most recent millennium trend in the Ipperwash paleohydrograph. This would suggest that the Chicago outlet remained the active outlet for Lake Michigan-Huron until at least 710 years ago, the youngest possible date for the final abandonment of the Chicago outlet.

Millennium trends can be explained by natural climate change and the abandonment of the Chicago outlet. Based on the interpretation of the Ipperwash paleohydrograph, the abandonment of the Chicago outlet is the most likely scenario responsible for the linear millennium trends within the Ipperwash paleohydrograph. Variations about this trend would then be related to climate. The active Chicago outlet most easily explains the millennium trend within the Ipperwash paleohydrograph and fits with previous interpretations of an active Chicago outlet during the Algoma lake phase interpreted from Lake Michigan strandplains (Chrzastowski and Thompson, 1992). However, the Ipperwash strandplain suggests the Chicago outlet remained active longer than previously thought, up to 710 years ago and during 1700-high lake phase in Lake Michigan. More detailed geologic information at the Chicago and Port Huron/Sarnia outlets are needed to determine the activation and abandonment of these outlets during geologic time and interpreted with strandplain data of Lake Michigan (Baedke and Thompson, 2000) and unpublished data of Lake Huron strandplains.
6.4 Centennial patterns

Groups of 3 to 7 ridges were identified based on cross-strandplain changes in foreshore thickness, mean grain size, beach ridge width and ridge height (Table 4; Figure 20). These groups are similar in length to beach ridges groups found in Lake Michigan (Baedke and Thompson, 2000) and Lake Superior (Johnston et al., 2007). At Ipperwash, a similar length oscillation is observed as rises and falls in lake level (basal foreshore) above and below the millennium trend lines (Figure 30). This pattern is best represented as oscillations about the millennium pattern because both patterns are interpreted to related to climate.

Rises and falls in lake levels about the millennium trends marks a centennial pattern in the Ipperwash strandplain and consists of groups of 3 to 5 ridges (Figure 30). Centennial patterns last an average of 208 ± 114 years and have an amplitude of 0.8 ± 0.4 m (Figure 30). Fraser et al. (1990) studied several Lake Michigan-Huron shorelines and observed a similar length cycle (lasting 100 to 150 years) based on radiocarbon dating of several high lake levels proxies (beach erosion, stream aggradation, marsh formation, and soil formation) and attributed the lake level cycle to natural climate change. This cycle was also observed in strandplains in Lake Michigan as a 150 ± 30 year-long centennial lake level oscillations (Baedke and Thompson, 1997). A similarity in length to the centennial pattern observed in the Ipperwash paleohydrograph leads to the interpretation of the centennial pattern as a result of natural climate changes.

Figure 29: Ipperwash paleohydrograph divided into centennial oscillations as defined by lake level rises and falls about the millennium pattern lines. Centennial oscillations are represented by groups of 3 to 5 ridges. Length and amplitude of each oscillation is also labelled.
6.5 Multi-decadal patterns

The multi-decadal pattern is represented by the average amount of time it takes for a single beach ridge to develop at Ipperwash for 2810 years (from 3520 to 710 years ago). The age model used to create the Ipperwash paleohydrograph assumes a constant rate of progradation (Figure 21). Using the linear model extending across 36 Ipperwash beach ridges indicates the average timing of beach ridge development at Ipperwash is 73 ± 35 years. Since the Ipperwash strandplain beach ridges are variably spaced across the strandplain (Figure 18), the amount of time it takes for a single beach ridge to develop may vary.

Hanrahan et al., (2009) report an intermodulation of two near-decadal climate cycles linked with quasi-periodic beach ridge formation in Lake Michigan based on analysis of the historic water gauge record. Because Lake Michigan and Lake Huron occupy the same basin, the beach ridges at Ipperwash are interpreted to result from climate cycles.

Multi-decadal lake level oscillations in the Ipperwash paleohydrograph occur every 73 ± 35 years and represent the average time it takes to form a single beach ridge. The interval of beach ridge development at Ipperwash overlaps with the range of beach ridge development in Lake Michigan (29-38 years; Beadke and Thompson, 2000) and Lake Superior (17-45 years; Johnston et al., 2012). Therefore, the multi-decadal oscillation in the Ipperwash paleohydrograph, Lake Michigan paleohydrograph, and Lake Superior paleohydrograph are interpreted to represent the natural rhythm in climate that has affected lake levels in Lakes Superior, Michigan and Huron over the late Holocene.
Chapter 7
Conclusion

The natural history of the LGL has been studied for over a century (Leverett and Taylor, 1915). Beach ridges record prehistoric lake levels (Thompson, 1992), and strandplains of beach ridges are used to reconstruct natural lake level fluctuations in the LGL (Johnston et al., 2014).

During the late Holocene, lake levels have naturally fluctuated in the LGL due to three dominate processes: GIA, outlet conveyance, and climate (Johnston et al., 2014). GIA is the rate of vertical ground movement in response to glacial advances and retreats. GIA occurs at different rates throughout the LGL as a result of the retreat of the Laurentide Ice Sheet causing lake levels to rise or fall at different relative rates within the same lake basin. Outlet conveyance is a change in the water carrying capacity of a lake's outflow channel(s) and causes lake levels to rise or fall at a universal rate within a single basin. Climate is a change in predominant regional weather patterns and cause lake levels to rise or fall within LGL basins.

The Ipperwash strandplain, southern Lake Huron, was studied to reconstruct past natural lake level changes in the Lake Huron basin and the Port Huron/Sarnia outlet during the late Holocene. The Port Huron/Sarnia outlet is the only remaining unregulated outlet (lacking any locks or dams) in the LGL and is therefore particularly susceptible to natural lake level changes. The Ipperwash strandplain offers a natural lake level record which mimics the lake level record of the Port Huron Sarnia outlet, because Ipperwash is the closest strandplain with the most beach ridges to the Port Huron/Sarnia outlet.

Elevation and age data were used to create the Ipperwash paleohydrograph. In order to obtain these data the lakeward margin of beach ridges were vibracored and analyzed to obtain prehistoric lake level elevations, and OSL ages were collected from beneath the crests of 10 beach ridges. OSL ages were then modelled using a linear regression to determine the age of individual beach ridges for the entire Ipperwash strandplain. The resultant Ipperwash paleohydrograph reconstructs a general cross-strandplain, linear, relative lake level lowering from a maximum elevation of 181.0 m to a minimum elevation of 177.8 m over a time period lasting from 3520 to 710 years ago.
Analysis of the Ipperwash paleohydrograph shows multi-millennium, millennium, centennial and multi-decadal patterns. GIA, outlet conveyance, and climate are the dominate drivers of lake level change in the LGL over historic (International Upper Great Lakes Study Board, 2009; 2012) and prehistoric times (Johnston et al., 2014) and used to interpret drivers of lake level patterns preserved within the Ipperwash paleohydrograph.

The multi-millennium pattern of the Ipperwash paleohydrograph shows a net lake level fall from 3520 to 710 years ago. This relative lake level fall is considered a record of GIA with a rate of 7 cm/century between Ipperwash and the ground surface at the Port Huron/Sarnia outlet, which currently regulates the water level in Lake Michigan-Huron.

Millennium patterns show two vertically offset periods of net lake level fall from 3520 to 2180 and 2020 to 710 years ago. These patterns correspond with the Algoma phase in lakes Superior, Michigan and Huron, the Sault phase in Lake Superior, or the 1700-high phase in Lake Michigan. The rate of lake level lowering from 3520 to 2180 in the Ipperwash paleohydrograph corresponds with the published rate of GIA between the Chicago outlet and Ipperwash. Therefore, the millennium pattern may be related to the dominance of the Chicago outlet from 3520 to 2180, and/or from 3520 to 710 years ago. However, regional climate variability also contributes to the lake level pattern preserved in the Ipperwash strandplain.

Centennial oscillations (lasting an average of 208 years) occur as groups of beach ridges at Ipperwash. Multi-decadal oscillations (lasting an average of 73 years) are represented as a single beach ridge at Ipperwash. Both centennial and multi-decadal oscillations are interpreted as a product of natural climate variability.

Interpretation of multi-millennium pattern within the Ipperwash paleohydrograph estimates the rate of GIA at Ipperwash to be 7 cm/century. However, estimates of GIA based on water gauge data suggest the rate of GIA at Ipperwash is 0 cm/century. This difference suggests an underestimation of GIA at Ipperwash based on water level gauge data due to insufficient nearby water level gauge data, or an outlet subsiding relative to the Ipperwash strandplain being dominate during the deposition of the Ipperwash strandplain. Interpretation of the millennium pattern of the Ipperwash paleohydrograph may suggest the Chicago outlet as the dominate outlet for Lake Michigan-Huron.
from 3520 to 2180 years ago. However, variations in the millennium pattern may also be produced by climate variations. Based on interpretation of the Ipperwash paleohydrograph the Chicago outlet is presented as the dominate outlet from 3520 to 2180 years ago and may have remained the active outlet until as recently as 710 years ago. The dominance of the Chicago outlet affected the long-term rate of relative GIA across the entire Ipperwash paleohydrograph causing the rate of GIA to be higher than the hypothesized rate of GIA.

The Ipperwash paleohydrograph provides the natural record of lake level fluctuations at Ipperwash. The Ipperwash paleohydrograph shows a long-term relative lake level lowering driven by GIA. However, future lake level projections show lake levels in Lake Michigan-Huron will continue to fall over the next hundred years due to human caused climate change (Lofgren et al., 2002; 2011; Angel and Kunkel, 2010; Hayhoe et al., 2010; MacKay and Seglenieks, 2013). The long-term lowering driven by GIA will therefore likely be exacerbated by a lake level lowering driven by climate change. However, the Ipperwash paleohydrograph also shows centennial and multi-decadal lake level oscillations which will likely continue to affect lake levels at Ipperwash into the future. Researchers must be aware and account for the natural patterns shown in LGL paleohydrographs to more accurately predict future lake level changes.

Interpretation of the Ipperwash paleohydrograph suggests the Chicago outlet was the active outlet for Lake Michigan-Huron during the deposition of the Ipperwash strandplain, and potentially remained active until as recently as 710 years ago. By providing an alternative scenario for geologically recent outlet conveyance, this thesis provides new insights and theories into the history of outlet conveyance, GIA and climate in Lake Michigan-Huron. This reassessment of the natural history of Lake Michigan-Huron also shows the need to account for natural changes in outlet conveyance, GIA and climate when examining and interpreting instrumental records such as water level gauges. The Ipperwash paleohydrograph also shows coastal erosion and deposition at Ipperwash as a result of multi-decadal lake level rises and falls. This shows that the Ipperwash strandplain will continue to form new beach ridges into the future assuming sediment supply remains positive.
Chapter 8

Recommendations

Several recommendations for future studies, based on the results of this thesis, are presented that will further the understanding of the Ipperwash strandplain and the LGL.

- Compare and combine the Ipperwash paleohydrograph to four other unpublished Lake Huron strandplain paleohydrographs to create a single, Port Huron/Sarnia outlet reference paleohydrograph.

- Examine strandplain lake level records near the Chicago outlet and compare to other strandplain paleohydrographs in Lake Michigan-Huron to determine the timing of activation and abandonment of the Chicago outlet.

- Determine ages of the three spits located in the Port Huron/Sarnia outlet (Campbell, 2016), either by directly sampling sediment in each spit or by bracketing ages of each spit using archeology to determine the timing of spit formation and providing direct evidence for potential outlet conveyance due to sedimentation at the Port Huron/Sarnia outlet.

- Extract subsurface and age data from the shorelines located within the Thedford embayment associated with either Lake Nipissing and/or Algonquin (Cooper, 1974) and compare with peak Nipissing elevation data collected at the Port Huron/Sarnia outlet (Thompson et al., 2014).
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Appendix A Ground Penetrating Radar Profiles

GPR profiles location

Radar facie description and interpretation

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<th>Description</th>
<th>Interpretation</th>
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<td>Dune</td>
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<tr>
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<td>Lakeward dipping to landward dipping to horizontal continuous reflections</td>
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<td>0-9 m</td>
<td>Sigmoidal reflections which truncate other reflections</td>
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<td>5-9 m</td>
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### Appendix B Modern Beach Survey and Elevation Calibration

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Water elevation at Goderich 176.905
## Appendix C Cross Strandplain Topographic Survey

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### Appendix D Modern Beach Sample Statistics

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<th>Mean (Φ)</th>
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<th>Sorting</th>
<th>Skewness (Φ)</th>
<th>Kurtosis (Φ)</th>
<th>Kurtosis D10 (Φ)</th>
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**Appendix E Core Statistics**
### Core Sediment Sample Statistics

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<th>Skewness (Φ)</th>
<th>Skewness</th>
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Mean (Φ): Average size of grains in phi units.  
Mean Sorting (Φ): Degree of sorting.  
Sorting: Classification of sorting.  
Skewness (Φ): Degree of skewness.  
Kurtosis (Φ): Degree of kurtosis.
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Visual laminae → no laminae
Photo laminae → no laminae, light brown → brown
Peel laminae → no laminae

2005 S-F Contact
Explanation
F-D Contact
Explanation

X-Pale in Sorting vs Mean, Skewness vs Mean graph

Visual laminae → no laminae
Photo laminae → no laminae, light brown → brown
Peel laminae → no laminae

2006 S-F Contact
Explanation
F-D Contact
Explanation

X-Pale in Sorting vs Mean, Skewness vs Mean graph

Visual laminae → no laminae
Photo laminae → no laminae, light brown → brown
Peel laminae → no laminae

2007 S-F Contact
Explanation
F-D Contact
Explanation

X-Pale in Sorting vs Mean, Skewness vs Mean graph
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**Visual**

- Abrupt layer of shell fragments above contact, consistently spaced horizontal laminae → inconsistently spaced, lakeward dipping laminae, grey → brown
- Laminae → no laminae, brown → light brown

**Peel**

- Abrupt layer of shell fragments above contact, consistently spaced horizontal laminae → inconsistently spaced, lakeward dipping laminae, grey → brown
- Laminae → no laminae, brown → light brown
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### Explanation

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### 2025 S-F Contact

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### X-Plots

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Visual:
- abrupt layer of shell fragments above contact, consistently spaced laminae
- lakeward dipping laminae
- brown
- light brown
- no laminae
- laminae

Photo:
- abrupt layer of shell fragments above contact, consistently spaced laminae
- lakeward dipping laminae
- brown
- light brown
- no laminae
- laminae

Peel:
- abrupt layer of shell fragments above contact, consistently spaced laminae
- lakeward dipping laminae
- brown
- light brown
- no laminae
- laminae

2031 S-F Contact Explanation F-D Contact Explanation
Mean x fine → normally graded x normally graded → fine
Skewness x fine → normally graded x normally graded → fine
Sorting x well sorted → moderately well sorted to well sorted
X-Plots x grouping in Sorting, Skewness and D(99) vs Mean
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<th>Description</th>
<th>S-F Contact</th>
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2037

S-F Contact

Explanation

F-D Contact

Description

2038

S-F Contact

Explanation

F-D Contact

Description

2039

S-F Contact

Explanation

F-D Contact

Description
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