CHANGES IN WATER LEVEL, VERTICAL GROUND MOVEMENT, SHORELINE BEHAVIOR AND CLIMATE IN THE LAKE SUPERIOR BASIN DURING THE LAST 5,000 YEARS

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DEDICATION

This dissertation is dedicated in loving memory to my sister, Jackie. Her sudden and unexpected death paralyzed us all. She was survived by two young boys and a husband. Although saddened by our loss, we are thankful for all the fond memories. Her never-ending love and support for me, her drive for success, and her quest for knowledge helped me during my journey. She has changed the life of every person she touched and will forever be missed.

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PREFACE

Concern for climate change and how it will affect our water resources is at the forefront of the scientific community today. Climate change has the potential not only for adverse impacts on humans into the future but there is also the uncertainty of human influence on the Earth's climate system. Instrumental records have too short a time span into the past to examine the full range of natural climatic variability and few data sets exist to bridge between the historical and geological records. Ancient shorelines called beach ridges, which are typically preserved in embayments provide critical information about changes in water level, vertical ground movement, shoreline behavior, and climate ranging from many decades to five millennia in length during this important time period between the historical and geological records. The Great Lakes of North America provides an ideal location for such a study. It is also the largest fresh surface-water system in the world providing water for consumption, transportation, power, and recreation for two nations, Canada and the United States. It contains a large number of well-preserved shorelines that were minimally affected by tides and bioturbation. The Great Lakes are also situated in a midcontinental position where climatic patterns and lake level can be influenced by several sources, which make for a complex but complete study. About 300 beach ridges located in four beach-ridge strandplains adjacent to Lake Superior were studied to increase the resolution between the historical and geological records and to help identify past patterns of long-term change in water level, vertical ground movement, shoreline behavior, and climate. A better understanding of these unique features will help us properly identify and evaluate shorter-duration events in the historic records that are superimposed on longer-duration events to help us potentially prepare for the future.

This dissertation is divided into three chapters. Each chapter was an invited paper included in three different special volumes. They describe how this study has advanced the understanding of individual beach ridges, strandplains of beach ridges, and analysis of multiple strandplains within a common basin.

The first chapter explains the development of facies and age models used to create a relative lake-level hydrograph for a strandplain of beach ridges in the Tahquamenon Bay embayment along the coastline of Lake Superior in Michigan. The chapter includes a sedimentological analysis of beach ridges in Lake Superior at a study site that has experienced similar conditions to the current outlet for Lake Superior. This resulted in the creation of the most detailed and continuous relative lake-level hydrograph for Lake Superior to date for the late Holocene. Multidecadal and centennial fluctuation in lake level during the late Holocene was recognized and evidence of recent tectonism in an area with several faults and high glacial rebound rates was revealed. These faults may have played a role in the separation of Lake Superior from Lake Michigan and Huron. This work was presented at the 2002 Superior: State-of-the-Lake conference at Michigan Technological University in Houghton, Michigan and was subsequently invited for publication in a special volume on Lake Superior research in the Journal of Great Lakes Research. This manuscript is co-authored by Steven Baedke, Robert Booth, Todd Thompson, and Douglas Wilcox and is currently in press.

The second chapter presents evidence for a systematic pattern of beach-ridge development and preservation based on a previously developed conceptual model by

Thompson and Baedke (1995) and profiles through beach ridges collected by using ground penetrating radar (GPR). The internal architecture of beach ridges revealed by high-resolution GPR facilitated a refinement of the conceptual model to account for beach-ridge preservation, which, in turn, helped explain how variables interact to control beach-ridge development and preservation. A paper on this subject was presented at the 2002 Geological Society of America North-Central section regional conference in Lexington, Kentucky. It was invited for publication in a Geological Society of America Special Publication titled “Advances in Stratigraphic Analyses using GPR”. This manuscript is co-authored by Todd Thompson and Steven Baedke and currently in review.

The third chapter centers around the analysis of 294 cores collected from four strandplains of beach ridges along the coastline of Lake Superior aimed at interpreting the causes and timing of the separation of Lake Superior from Lake Michigan and Huron. A discontinuity in each strandplain sequence was defined on the basis of geomorphic and sedimentologic characteristics and was related to the separation of the lakes. The analysis indicated that the date of separation of the lakes occurred about 1,000 years closer to the present than had previously been reported. The results of this analysis were presented at the 2003 Geological Society of America annual conference in Seattle, Washington, the 2004 International Association for Great Lakes Research annual conference in Waterloo, Ontario in a special session titled “The Greater and Lesser Great Lakes”, and at the 2004 Geological Association of Canada annual conference in St. Catherines, Ontario in a special session titled “Paul F. Karrow Symposium – Reviews and reflections on

Quaternary Sciences”. This manuscript will be submitted for publication in a special volume pertaining to the session at the IAGLR conference.

ONGOING AND FUTURE RESEARCH

Ongoing research includes developing age models for each of the four study sites described in this dissertation (Grand Traverse Bay, Au Train Bay, Tahquamenon Bay, and Batchawana Bay) and an additional site near Sault Ste. Marie to create a chronological framework for each strandplain and assign an age to each beach ridge. This involves working with more than 100 radiocarbon ages and 70 optically stimulated luminescence ages. It is hoped that OSL will become a viable alternative to radiocarbon dating so that actual features being studied (sand in beach ridges) can be age dated and not an associated deposit with the feature (peat in intervening swales) as has been standard practice. This would allow strandplains with little or no preserved peat to be studied, which is the case for several strandplains in Lake Superior because high glacial rebound rates elevate peat deposits above water tables reducing peat production and allowing peat degradation. Age models will then be used to create five relative lake-level hydrographs for Lake Superior. Vertical ground movement will be removed at each site using an approach that minimizes residuals (difference between glacial rebound and water level) and a lake-level hydrograph relative to the active Lake Superior outlet will be created. Because the outlet changed for Lake Superior during the late Holocene, two hydrographs will need to be produced: one for the Port Huron/Sarnia outlet, and the other for the Sault outlet.

Several preliminary studies have been conducted investigating the role of climate in lake-level fluctuations using information collected from geologic records of strandplains. These include studies of pollen and macrofossils (Dr. Robert Booth), and isotopes (Drs. Shikha Sharma and German Mora) from wetland sediments, and synoptic conditions in the historical record characteristic of patterns found in the geologic record (Dr. Sara Pryor and Nathan Polderman). It is hoped that these studies will continue as Lake Michigan hydrographs are reevaluated, Lake Superior hydrographs are refined, and Lake Huron hydrographs are created as the role of climate to lake-level fluctuations in the upper Great Lakes is pursued.

Further investigation is needed to determine the cause of the four-meter fall in lake level from the Nipissing II high-water-level phase about 4,000 years ago, possible relationships between ancient faults in the Lake Superior basin, tectonic events, isostatic rebound, and the separation of Lake Superior from Lake Michigan and Huron. Ground penetrating radar needs to be utilized more and compared to cores and conceptual models to create continuous 2D and 3D frameworks of sedimentary deposits for architectural studies relating development versus preservation. This would help us better understand long-term processes related to shoreline behavior that is responsible for creating our coasts. There is also a need to better recognize and understand in the historical record long-term patterns of lake-level change identified in the geologic record so we can put short-term events in proper context for interpretation and use this to potentially predict future changes.

ABSTRACT

Establishing a long-term framework of water level, vertical ground movement, shoreline behavior, and climate is important to examine pre-regulation or ‘natural” conditions, the context of historical events, and help predict future changes critical for effective resource management in the Great Lakes region. Four large strandplains of beach ridges in Lake Superior embayments provide a multi-decadal to five-millennial geologic framework. Over 300 beach ridges were cored and surveyed and many were profiled using ground penetrating radar. Defined nearshore and onshore facies and subsurface relationships were compared to the modern shoreline.

Major findings of this research include: 1) A rapid lake-level drop (approximately 4 m) from the Nipissing II stage (approximately 4,000 cal. yrs. BP) was followed by superimposed approximately 30-year and 150-year quasi-periodic lake-level fluctuations in the Lake Superior basin. This extends similar findings from the Lake Michigan basin to the Lake Superior basin supporting commonality in water levels between basins. 2) Geomorphic (topography, drainage, ridge and swale continuity, orientation, relief and spacing) and sedimentologic (facies contact elevations and mean grain size) characteristics define a correlative break in the strandplain sequences at each site. The change in outlet position for Lake Superior from Port Huron/Sarnia to Sault Ste. Marie is shown by change in cross-strandplain elevations at all sites. Age information suggests the outlet change occurred about 1,200 cal. yrs. BP (after the post-Algoma phase), about 1,000 years younger than previously reported. After the outlet switch littoral transportation directions and/or sediment source changed at several sites and beach ridges relief and spacing decreased. 3) High vertical ground-movement rates, the presence of several faults near the outlet as well as missing record at two sites suggest a tectonic event at or near the Sault outlet may have been a factor in the outlet change. 4) A systematic preserved beach-ridge architecture containing concave lakeward-dipping ravinement surfaces, overlain by aggrading and lakeward prograding sigmoids were identified using high resolution ground penetrating radar. These data combined with a previously defined model of beach-ridge development supports a positive rate of sediment supply and multi-decadal fluctuation in lake level as the mechanism for beach-ridge formation.

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CHAPTER 1:

LATE HOLOCENE LAKE-LEVEL VARIATION IN SOUTHEASTERN LAKE SUPERIOR: TAHQUAMENON BAY, MICHIGAN

Abstract

Internal architecture and ages of 71 beach ridges in the Tahquamenon Bay embayment along the southeastern shore of Lake Superior on the Upper Peninsula of Michigan were studied to generate a late Holocene relative lake-level curve. Establishing a long-term framework is important to examine the context of historic events and help predict potential future changes critical for effective water resource management. Ridges in the embayment formed between about 4,200 and 2,100 calendar years before 1950 (cal. yrs. B.P.) and were created and preserved every 28 ± 4.8 years on average. Groups of three to six beach ridges coupled with inflections in the lake-level curve indicate a history of lake levels fluctuations and outlet changes. A rapid lake-level drop (approximately 4 m) from about 4,100 to 3,800 cal. yrs. B.P. was associated with a fall from the Nipissing II high-water-level phase. A change from a gradual fall to a slight rise was associated with an outlet change from Port Huron, Michigan/Sarnia, Ontario to Sault Ste. Marie, Michigan/Ontario. A complete outlet change occurred after the Algoma high-water-level phase (ca. 2,400 cal. yrs. B.P.). Preliminary rates of vertical ground movement calculated from the strandplain are much greater than rates calculated from historical and geologic data. High rates of vertical ground movement could have caused tectonism in the Whitefish Bay area, modifying the strandplain during the past 2,400 years. A tectonic event at or near the Sault outlet also may have been a factor in the outlet change from Port Huron/Sarnia to Sault Ste. Marie.

1.1 Introduction

Lake Superior is the largest by volume of the Great Lakes of North America; by area, it is the largest freshwater lake in the world and is a part of the largest fresh surface-water system in the world (Cohen 1998). It provides water for consumption, transportation, power, and recreation and borders two nations, Canada and the United States. Documenting the magnitude and frequency of past Great Lakes water-level fluctuations is critical to optimize water resource management, to help understand past climate change, and to predict potential future changes in lake levels. Although lake-level gauges provide short-term records of water-level fluctuations for the past 150 years, it is difficult to establish the importance of these fluctuations without a framework of long-term fluctuations. Records of lake-level fluctuations spanning the last several millennia can be extracted from the sedimentary deposits preserved in ancient shorelines. Researchers have studied late Holocene shoreline features preserved along the coastline of Lake Superior for more than a century (Lawson 1893), but no one has developed a detailed continuous chronology of events that can be used to establish the physical limits and timing of paleo lake-level fluctuations. Most previous research focused on three different water-level phases of ancestral Lake Superior: Nipissing, Algoma, and Sault. These high-water phases are best preserved in shoreline features found along the northern part of the Lake Superior basin where relatively large isostatic rebound rates have elevated the features far above lake level and separated different lake-phase features. Stanley (1932) documented these phases on the eastern side of Lake Superior, Farrand (1960) on the northern and western side of Lake Superior, and Cowan (1978, 1985) near the outlet for Lake Superior in the Sault Ste. Marie area. These studies infer past lake levels from geomorphic evidence, but lack detail between major lake phases. Work by Larsen (1994) at Whitefish Point, Michigan, partially adopted the sedimentological approach of Thompson (1992) to interpret lake level from a chronosequence of beach ridges, but Larsen focused his studies on isostatic rebound and used an age model different from that used here. Further work by Larsen (1999 a, b, c) and Larsen et al. (1999) filled in parts of the late Holocene water-level record by constraining the timing and elevation of past lake-level low-stands and attempting to determine the time when the outlet at Sault Ste. Marie began regulating water levels in the Lake Superior basin. Despite these recent efforts and studies by Johnston et al. (2000 a, b; 2001 a, b; 2002 a, b) and Thompson et al. (2002), more sedimentological studies are needed to accurately define long-term patterns of lake-level variability (e.g., Baedke and Thompson 2000). This report summarizes a study of beach ridges in the Tahquamenon Bay embayment along the southeastern coast of Lake Superior in the Upper Peninsula of Michigan updating and expanding upon previous work by Johnston et al. (2001a). We chose this site because it contains many well-defined ridges separated by well-developed wetlands and is easily accessible. Tahquamenon Bay is important because it is located near the outlet of Lake Superior and may have experienced a similar rate of vertical ground movement (isostatic uplift) as the outlet. Data from this site will help define past vertical ground movement, lake-level fluctuations, and outlet constraints crucial to interpreting the lake's history.

Our purpose is to produce a relative lake-level curve for the embayment. We used methods that were consistent with those outlined by Thompson et al. (1991) and Thompson (1992), where researchers used foreshore elevations to approximate the elevation of the lake at the time each beach ridge formed; they also used radiocarbon dates to develop an age model to approximate the age of each beach ridge. Similar data sets from around the lake are useful in removing vertical ground movement from each site's relative curve to produce a combined high-resolution curve for the entire lake (cf. Baedke and Thompson 2000) . The resulting record helps to determine the physical limits and timing of lake-level variation, long-term patterns of shoreline behavior, vertical ground movement, and paleoclimate change over the past several thousand years. We hope that this study will provide a geologic framework for other research in the Lake Superior basin, including ongoing ecological (e.g., Keough et al. 1999, Kowalski and Wilcox 1999), geomorphological (e.g., Loope and Arbogast 2000, Arbogast et al. 2002), paleoecological (e.g., Booth et al. 2002, Jackson and Booth 2002), pedological (e.g., Barrett 2001), regulatory (e.g., International Joint Commission 2002), and sedimentological (e.g., Fuks and Wilkinson 1998, Nichols 2002) studies.

1.2 Study site

The study area is located in northwest Chippewa County, Michigan, between 46°27'30" and 46°29'30" north latitude and 84°57'30" and 85°02'30" west longitude (Fig. 1.1A). It is approximately 35 km south of Whitefish Point, Michigan, 42 km northeast of Newberry, Michigan, and 50 km west of the outlet for Lake Superior at Sault Ste. Marie, Michigan. This northward-opening embayment extends approximately 6 km east-west and 2.5 km north-south (Fig. 1.1B). The limit of the embayment is marked by an elevated bedrock headland to the east and an elevated bedrock and/or till surface to the south and west. The approximate edge of the embayment follows a 198.1-m (650-ft)

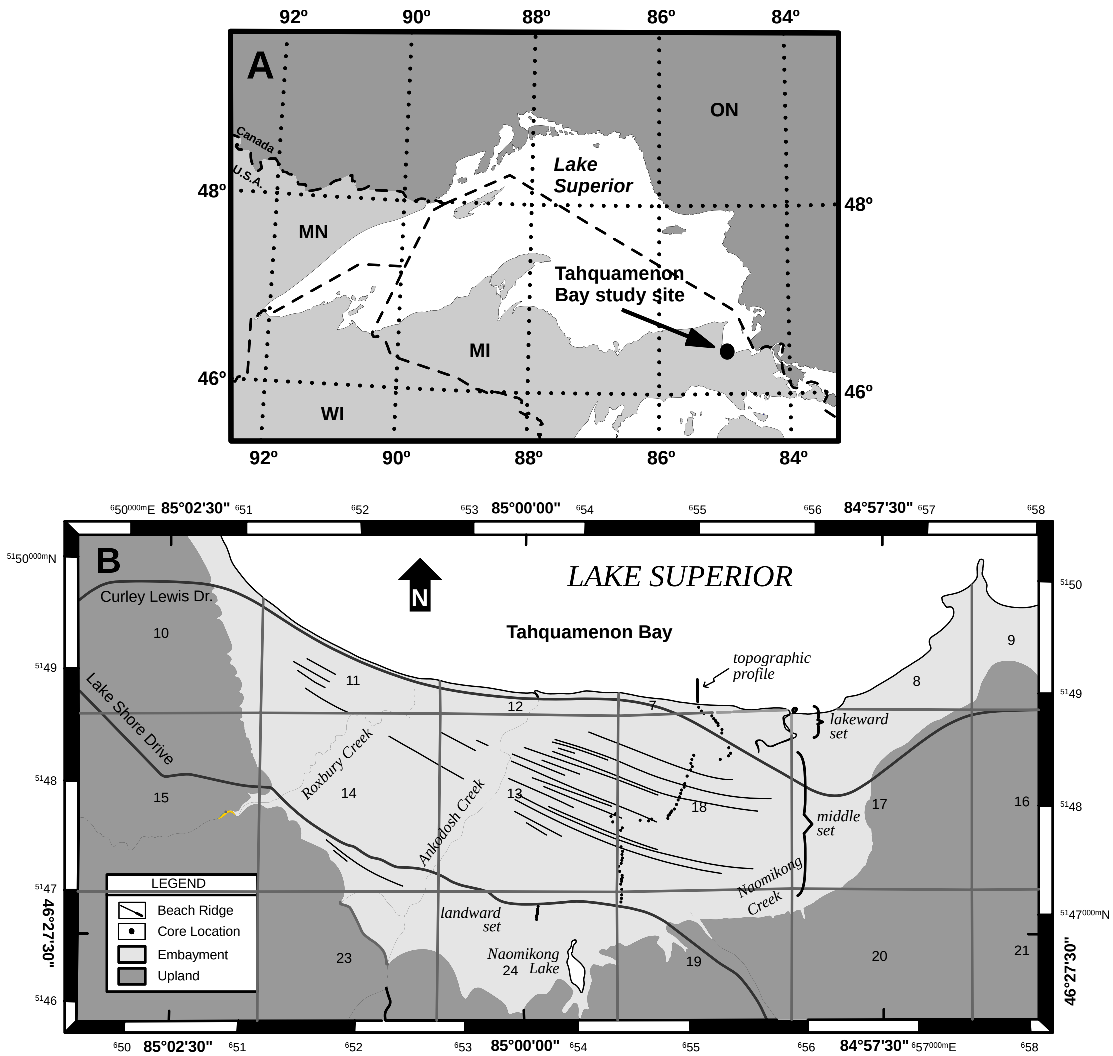

Fig. 1.1 Map showing the Tahquamenon Bay study area. (A) Map of Lake Superior showing the location of Tahquamenon Bay. (B) Map of the Tahquamenon Bay embayment. The lakeward side of 71 beach ridges were cored to determine the elevation of the lake when each beach ridge formed. Lines indicate beach-ridge crests and dots indicate core locations.

elevation contour on the U.S. Geological Survey McNearny Lake and Piatt Lake quadrangles and is about 14.6 m above the average elevation of Lake Superior (183.5 m, 602 ft). A bluff 10- to 15-m-high forms the southeastern and eastern margin of the embayment, and a 6-m bluff forms the western margin. The southwestern margin of the basin extends approximately 1 km further inland than the adjacent margins and forms an elevated platform that slopes to the northeast into the embayment.

Several nearshore bars, paralleling the modern coastline, are on a platform that extends 300 m offshore and lies under less than 2 m of water. Further offshore, the embayment is a part of a larger platform (< 9 m deep) that extends lakeward to a line near Paradise, Michigan, to Salt Point, Michigan. This platform opens up into Whitefish Bay (< 90 m deep). The Tahquamenon River flows into Lake Superior about 9 km north of the Tahquamenon Bay embayment, and a submerged spit at the mouth of the river extends southward to about 3 km north of the study site. The southward-projecting spit suggests that the Tahquamenon River is a source of sediment to the embayment and distributes it southward. The extension of a sand spit at the mouth of Naomikong Creek, which drains the embayment as observed in the 1975 USGS Emerson quadrangle and 1997 Michigan Department of Natural Resources (MDNR) air photographs (scale 1:15,840), indicates that littoral transport is currently toward the east along the coast in the study area.

The Tahquamenon Bay embayment contains approximately 80 beach ridges separated by wetland-filled swales. The most well-defined ridges are in the eastern part of the embayment (Fig. 1.1B). Dense vegetation in the most landward part of the strandplain limits access and makes it difficult to define ridges and swales from air photos. The 13 ridges adjacent to Lake Superior are oriented east-west, as is the modern shoreline (Fig. 1.1B). The relief between ridge crests and adjacent swales is less than 1 m, and most of the swales between ridges are dry or contain discontinuous wetlands. In the middle of the strandplain ridge-crest orientations are 15° from the modern shoreline and follow a ESE to WNW direction (Fig. 1.1B). These ridges generally have greater relief than the lakeward ridges, with an average of about 3 m. Most of the swales in this group are continuous across the central part of the embayment. Ridge-crest orientations of the ten most landward ridges vary from east-west to WSW-ENE (Fig. 1.1B). These ridges are on an elevated platform and have less than 1 m of relief. Swales in this last set commonly are ponded and contain little organic matter; standing water in some of these swales may be related to nearby beaver dams in Naomikong Creek.

Groups of three to six ridges separated by wider than average wetlands are evident in air photographs and in changes in relief. These groups are defined by a systematic rise and fall in elevation between wider-than-average wetlands, and they are less common in the most lakeward and landward ridges.

1.3 Methods

During the summer of 1999, we collected 71 cores vibracored through 80 beach ridges along four transects roughly perpendicular to the modern shoreline (Fig. 1.1B) using a land-based vibracorer (Thompson et al. 1991). The lakeward sides of all accessible beach ridges were cored to minimize the amount of recovered dune sand and to ensure that penetration was deep enough to recover basal foreshore sediments. We cored several of the vibracore holes twice to penetrate clay beneath the nearshore sequence. The clay acted as a plug in the end of the aluminum tubes and prevented loss of the basal sediments. Most cores contained the entire vertical nearshore sequence. We recorded core orientations before cores were removed from the ground so the orientation of the sedimentary structures were maintained. Core elevations were surveyed and calibrated to the International Great Lakes Datum 1985 (IGLD85) (Coordinating Committee on Great Lakes Basic Hydraulic and Hydrologic Data 1995) using the National Oceanic and Atmospheric Administration (NOAA) Point Iroquois, Michigan, gauging station. Core sites were located using differentially corrected Global Positioning System (GPS) and 1997 MDNR air photographs (scale 1:15,840). Using a stereo zoom transfer scope, we then transferred beach-ridge crests and core locations to the USGS McNearney Lake and Piatt Lake 7½ minute quadrangles to determine distance landward from the modern shoreline to each beach ridge.

We surveyed an onshore/offshore-oriented profile 250 m in length across the modern shoreline, and retrieved sediment samples from each recorded elevation (Fig. 1.1B). The profile extended from the top of the first dune ridge landward from the modern shoreline, into the lake, and across two nearshore bars and two troughs. Theses data allowed us to observe modern sediment and bedform distributions that should be reflected vertically in beach-ridges cores as changes in grain size and sedimentary structures.

In the laboratory, vibracores were split in half, described, sampled, photographed, and glued to strips of masonite using Rub-R-MoldTM latex. The latex strips enhance the visibility of sedimentary structures and create a permanent physical record of the core. We sieved 1,300 grain-size samples using ½phi intervals, and calculated statistical parameters (mean, standard deviation, and skewness) for each sample by the mathematical method of moments. The coarsest one-percentile (C1%) was determined from cumulative grain-size distributions. Statistical parameters were plotted per core to identify grain-size variations with depth. We compared visual descriptions, grain-size data, photographs, and latex strips to distinguish nearshore and onshore facies, specifically the upper and lower contacts of foreshore deposits.

To constrain the ages of the ridges, we used a combination of conventional and accelerator mass spectroscopy (AMS) to radiocarbon-date the basal wetland sediments (Table 1.1). Samples consisted of peat collected from the deepest point in the swales along the same transects as the beach-ridge cores. We collected samples for AMS radiocarbon dating using a wide-diameter (10.2 cm) piston corer (Wright et al. 1984) and for conventional dating by hand-augering 7.5-cm-diameter aluminum tubes through the peat profile. We sampled only wetlands that were laterally continuous, which contained at least 0.5 to 1.0 m of peat; about a third of the wetlands were cored. A single sample was collected from the lakeward set of swales and none from the landward set as those areas lacked a suitable thickness of peat. For conventional radiocarbon dating, we removed one basal peat sample, approximately 3 cm thick, from each core. For AMS radiocarbon dating, terrestrial plant macrofossils (needles, seeds) or macroscopic charcoal fragments spanning the lowest 1 cm of peat were used. Geochron Laboratories (Cambridge, Massachusetts) performed conventional radiocarbon dating of bulk sediment. AMS dating of macrofossils and macroscopic charcoal fragments was prepared at the Institute of Arctic and Alpine Research (INSTAAR) (Boulder, Colorado) and analyzed at the National Ocean Sciences Accelerator Mass Spectrometry Facility

Table 1.1 Radiocarbon dates collected from the base of swales between beach ridges in the Tahquamenon Bay embayment.

Ridge number [a]	Distance landward (m) [b]	Material dated	Lab number [c]	Date reported	^{14}C yr B.P.	Calibrated yr B.P. [d]
14	374.0	Peat	GX-25952	9/30/1999	2470 ± 80	2644
15	408.0	Peat	GX-25953	9/30/1999	2000 ± 120	1936
16	442.0	Peat	GX-25954	9/30/1999	2450 ± 70	2481
17	482.8	Peat	GX-25955	9/30/1999	2640 ± 120	2753
18	523.6	Peat	GX-25956	9/30/1999	2810 ± 70	2908
23	714.0	Peat	GX-25957	9/30/1999	2890 ± 120	2996
24	795.6	Peat	GX-25958	9/30/1999	3110 ± 80	3350
29	945.2	Peat	GX-25959	9/30/1999	2950 ± 80	3128
37	1135.6	Peat	GX-25960	9/30/1999	2770 ± 120	2853
38	1190.0	Peat	GX-25961	9/30/1999	3160 ± 130	3378
40	1244.4	Peat	GX-25962	9/30/1999	2990 ± 130	3184
41	1258.0	Peat	GX-25963	9/30/1999	3170 ± 80	3381
45	1380.4	Peat	GX-25964	9/30/1999	3090 ± 80	3279
46	1428.0	Peat	GX-25965	9/30/1999	2880 ± 80	2979
48	1468.8	Peat	GX-25966	9/30/1999	3730 ± 80	4088
51	1543.6	Peat	GX-25967	9/30/1999	2840 ± 100	2949
52	1577.6	Peat	GX-25968	9/30/1999	3160 ± 130	3378
54	1625.2	Peat	GX-25969	9/30/1999	3230 ± 130	3465
55	1659.2	Peat	GX-25970	9/30/1999	3090 ± 60	3279
56	1686.4	Peat	GX-25971	9/30/1999	3320 ± 130	3564
59	1774.8	Peat	GX-25972	9/30/1999	2890 ± 80	2996
61	1808.8	Peat	GX-25973	9/30/1999	3530 ± 100	3786
8	163.2	*Betula* sp. (seeds), *Larix laricina* (needles) and *Pinus strobus* (needles)	NSRL-11678	10/31/2000	1580 ± 35	1493
17	482.8	*Larix laricina* (needles) and *Pinus strobus* (needles)	NSRL-11556	10/16/2000	2180 ± 35	2273
23	714.0	*Larix laricina* (needles)	NSRL-11557	10/16/2000	3110 ± 35	3350
37	1135.6	*Larix laricina* (needles) and *Pinus strobus* (needles)	NSRL-11558	10/16/2000	2570 ± 50	2740
42	1292.0	*Larix laricina* (needles) and *Pinus strobus* (needles)	NSRL-11680	10/31/2000	2990 ± 40	3184
43	1326.0	Macroscopic charcoal fragments	NSRL-11559	10/16/2000	3570 ± 45	3844
51	1543.6	Macroscopic charcoal fragments	NSRL-11560	10/16/2000	3750 ± 35	4125
61	1808.8	*Larix laricina* (needles) and *Pinus strobus* (needles)	NSRL-11561	10/16/2000	3840 ± 35	4240

[a] Ridge number assigned to each beach ridge identified in the embayment, starting from the modern shoreline increasing in number landward.

[b] Distance landward of the beach ridge crest from the modern shoreline.

[c] Lab numbers GX- correspond to conventional analysis reported from Geochron Laboratories (Cambridge, Massachusetts) and NSRL- correspond to accelerator mass spectroscopy (AMS) analysis reported from INSTAAR (Boulder, Colorado).

[d] Ages were calibrated to calendar years before 1950 using the University of Washington's Quaternary Isotope Lab Radiocarbon Calibration Program, CALIB version 4.3 (Stuiver and Reimer 1993, Stuiver *et al.* 1998).

(NOSAMS) (Woods Hole, Massachusetts). All radiocarbon dates were adjusted for variations in atmospheric 14C through time and we calibrated them to calendar years before 1950 using the University of Washington's Quaternary Isotope Lab Radiocarbon Calibration Program, CALIB version 4.3 (Stuiver and Reimer 1993, Stuiver et al. 1998) (Table 1.1).

1.4 Results and Discussion

1.4.1 Facies model

Horizontal trends in sediment facies across the modern shoreline compare well with vertical trends in beach-ridge cores and help to interpret the ancient record preserved in beach ridges. Grain size and sedimentary structures define three different facies (dune, foreshore, and upper shoreface) in both modern and ancient sediments. Dune, foreshore, and upper shoreface sediments all show certain characteristics that distinguish them from each other in both the modern and ancient sediments. The contact between the foreshore and upper shoreface is of great importance because it best approximates the average elevation of the lake (Thompson 1992). We describe modern and ancient characteristics in two separate sections to avoid confusion.

1.4.1.1 Modern facies

The modern topographic shore profile comprises three different sedimentary and geomorphic parts: 1) the dune in the onshore zone and in the nearshore zone, 2) the foreshore, and 3) the upper shoreface. The foreshore extends from where waves break to the maximum run up on the beach face. The dune (also called the foredune) extends landward from the foreshore, and the upper shoreface extends lakeward from the foreshore. The plunge point occurs at the still water line and is at the contact between the foreshore and upper shoreface.

Elevations decrease lakeward, with the largest fall occurring in the landward part of the profile between the foredune and the water line (plunge point). The profile in this area decreases about 1.2 m over 10 m and contains several scarps. No bedforms were present, and the foreshore had no well-defined cusps. From the plunge point lakeward, the profile decreases another 0.6 m across a 35-m-wide sand bar and into a trough in the upper shoreface. From there lakeward, the profile is relatively flat and varies only about 0.2 m over a second nearshore bar and trough. The surface of the upper shoreface was rippled along the entire profile at the time of the survey.

Most samples along the profile are medium-grained sand, with a slight increase in mean grain size across the foreshore (swash zone) (Fig. 1.2). The coarsest part of the foreshore occurs at the plunge point. Lakeward from the plunge point, grain size decreases ½-phi across the bar. In all, the mean size of foreshore sediments are coarser than adjacent dune and upper shoreface sediments, with the coarsest mean grain size occurring at the plunge point.

Coarsest one-percentile trends change from coarse-grained sand in the dune, medium-grained sand to granules in the foreshore, and coarse- to very coarse grained sand in the upper shoreface (Fig. 1.2). The coarsest C1% values occur at the plunge point and on the lakeward slope of the bar. Although the foreshore has the widest range in C1%, most of the foreshore sand was coarser than adjacent dune and upper shoreface sediment.

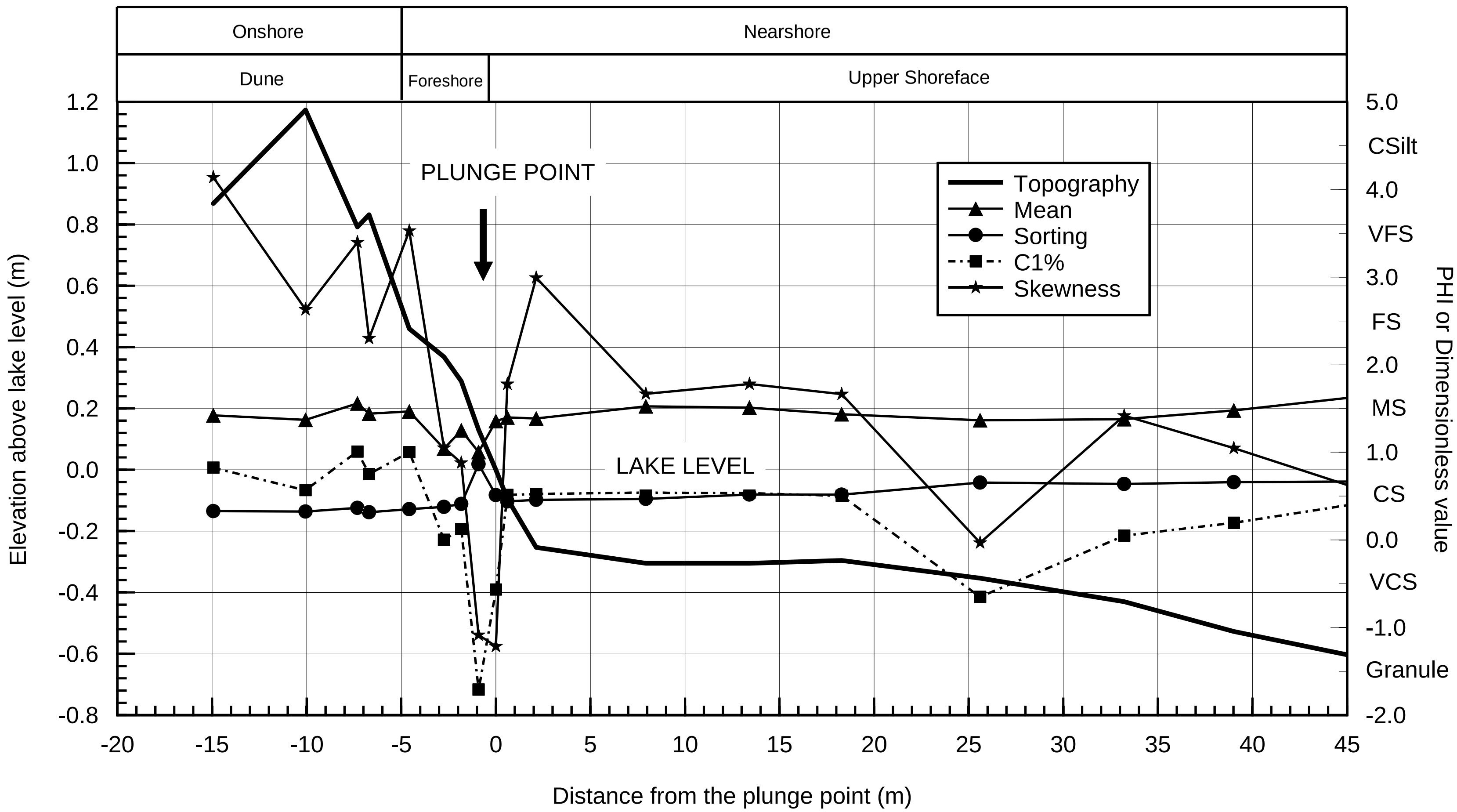

Fig. 1.2 Graph of onshore/nearshore topography and grain-size trends perpendicular to the modern shoreline at Tahquamenon Bay, Michigan. The profile extends across the modern beach and plunge point (land/water interface) and into the nearshore. The modern shoreline is eroding, producing a steep topographic profile on the landward part of the profile. Abbreviations along the second y-axis describe the grade of sediment (granular gravel to very coarse, coarse-, medium-, fine-, and very fine grained sands to coarse-grained silt).

Sorting trends change from very well sorted in the dune, well sorted in the foreshore (moderately sorted at the plunge point), and well sorted to moderately sorted in the upper shoreface (Fig. 1.2). Within the upper shoreface, C1% and mean grain size is finer and more poorly sorted with increasing distance offshore.

Skewness changes from more strongly fine skewed (more positively skewed) in the dune, to strongly fine skewed to strongly coarse skewed in the foreshore, to less strongly fine skewed (more negatively skewed) in the upper shoreface (Fig. 1.2). Dune and upper shoreface sediments contain more fine grains; foreshore sediments contain more coarse grains.

1.4.1.2 Ancient facies

Data from the modern topographic profile indicate that the most discriminating statistical parameters of the three facies are mean grain size, sorting, and C1% (Fig. 1.2). Therefore, we use these statistical parameters in conjunction with sedimentary structures to identify three sedimentary facies in cores (dune, foreshore, and upper shoreface) (Fig. 1.3).

In cores, foreshore sediments are coarser grained, more poorly sorted, more fine skewed, and have coarser C1% than dune and upper shoreface sediments. Dune deposits are more poorly sorted, more fine skewed, and have a coarser C1% than upper shoreface deposits. Dune sands are also finer grained in mean and C1% and more coarse skewed than foreshore deposits. Upper shoreface sediment is finer grained, more poorly sorted, more coarse skewed, and has a finer C1% than foreshore and upper shoreface sediments.

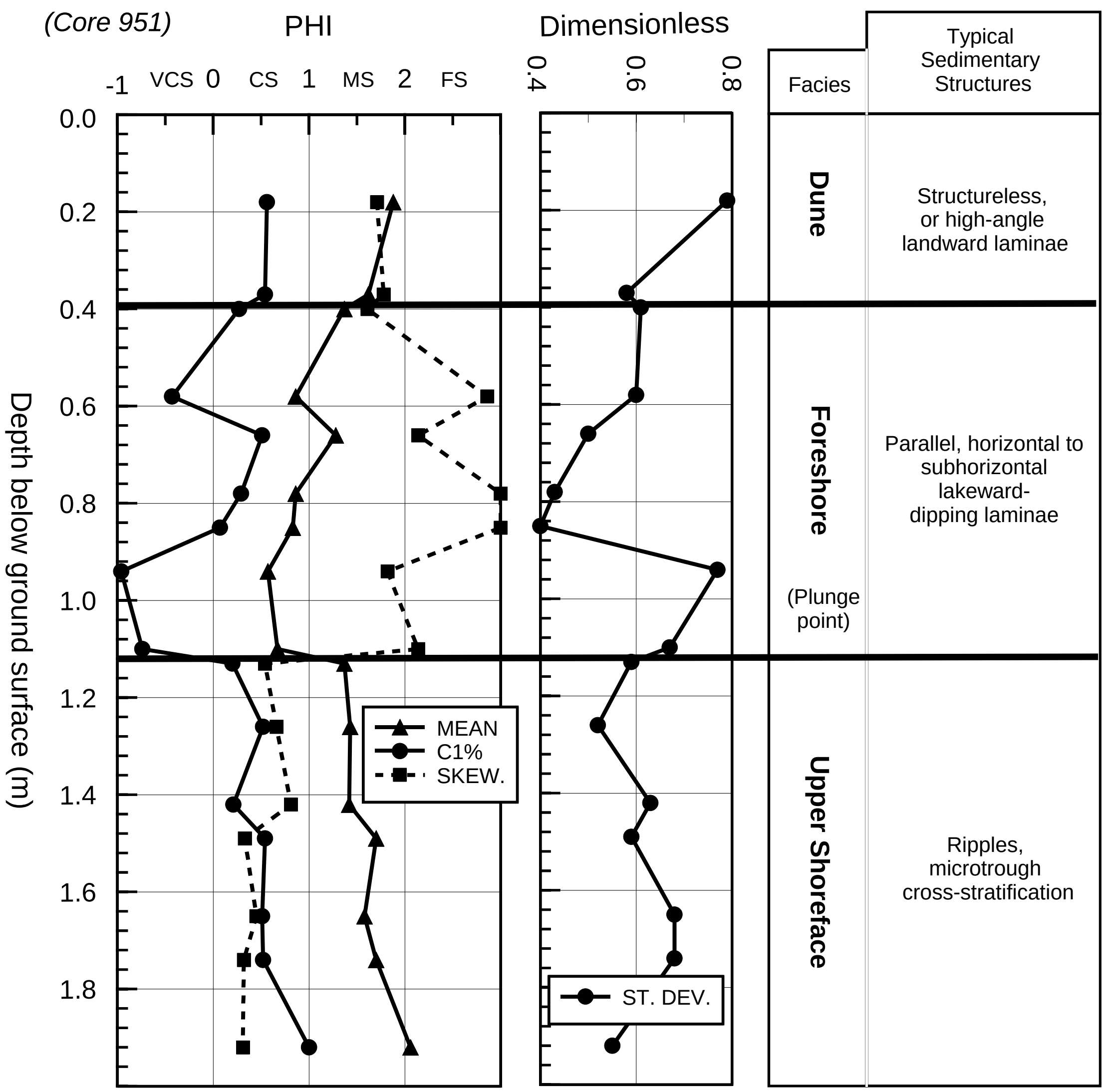

Fig. 1.3 Schematic diagram illustrating facies grain-size characteristics in core 951 with respect to depth and typical sedimentary structures in all cores. See Fig. 2 for grain-size abbreviations.

In cores, dune sediments are commonly structureless or contain high-angle landward-dipping laminae, whereas foreshore sediments typically contain parallel, horizontal to lakeward-dipping, subhorizontal laminae; upper shoreface sediments are rippled and contain micro-trough cross-stratification.

1.4.1.3 Facies synthesis

Facies relationships derived from grain size and sedimentary structures are similar in cores and across the modern beach. Dune sediments are commonly structureless, consist of more fine grains, and are better sorted than foreshore and upper shoreface sediments. This is mainly owing to wind transportation, because wind is efficient at transporting only a small range of fine grains and, therefore, structures are less apparent (Hunter 1977, Allen 1985, Komar 1998).

Foreshore sediments commonly contain parallel, horizontal to low-angle lakeward-dipping laminae, consist of abundant coarse grains, and are relatively well sorted to poorly sorted. These characteristics primarily result from oscillation and transportation by water in the nearshore zone; coarse grains are driven onshore and the few fine grains are winnowed out by offshore return flows (Fox et al. 1966, Fraser et al. 1991, Komar 1998). Wind also removes fine grains from the edge of the nearshore zone in the foreshore and transports them onshore into the dune (Komar 1998). The coarsest and most poorly sorted sand is deposited where the waves break at the plunge point in the foreshore, at the water line (Fraser et al. 1991). It is important to recognize this facies in core because it records the approximate elevation of the lake (Thompson 1992).

Upper shoreface sediments commonly contain ripples and micro-trough cross-stratification, organics, fine grains, and are relatively well sorted. This is mainly owing to nearshore processes where coarse and fine-grained sediments are driven onshore but only fine-grained sediment is returned offshore (Komar 1998). Ripples and micro-troughs are developed by onshore-offshore and alongshore currents in the nearshore zone (Greenwood and Davidson-Arnott 1979, Reineck and Singh 1980, Fraser et al. 1991).

1.4.2 Cross-strandplain facies relationships

A cross section of the entire strandplain shows that contact elevations for each pair of facies generally decrease lakeward (Fig. 1.4), except in the lakeward and landward sets of beach ridges (Fig. 1.1). A platform of foreshore elevations occurs in the most landward part of the strandplain above the rest. Constant to slightly rising foreshore elevations occur in the lakeward part of the strandplain closer than 250 m from shore. The top and bottom contacts of the foreshore do not parallel each other across the entire strandplain. Basal foreshore elevations decrease about 12 m from the landward to lakeward part of the strandplain, with the largest decrease occurring in the most landward part, dropping at least 4 m from 2.1 to 1.8 km from the shore (Fig. 1.4). Basal foreshore elevations indicate that relative lake levels ranged from a maximum elevation of 195.12 m to a minimum elevation of 182.86 m above sea level (IGLD85) in the Tahquamenon Bay area.

Foreshore elevations rise and fall in groups of three to six beach ridges (Fig. 1.4). This trend is similar to ground elevation changes described previously. As we will discuss later, such groupings are more readily apparent in the residual curve where

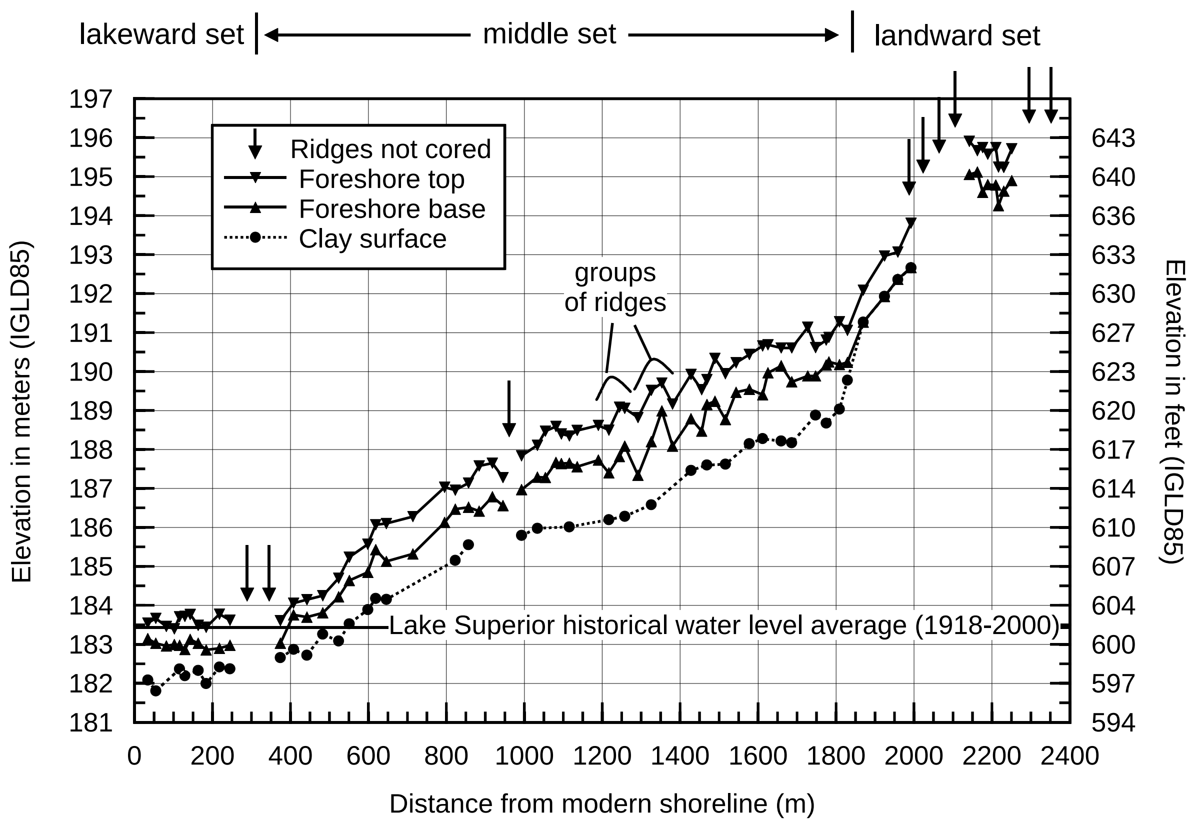

Fig. 1.4 Graph of the elevation of upper and lower limits of foreshore sediments versus distance landward from the modern shoreline. The middle line represents the boundary between foreshore and upper shoreface deposits and approximates the elevation of the lake during beach-ridge formation. The uppermost line represents the boundary between dune and foreshore deposits and is used in conjunction with the middle boundary to calculate foreshore thickness. The lowermost line corresponds to the contact between upper shoreface sediment and the predepositional surface composed of clay.

vertical ground movement is removed. Although the trends are similar between contacts, basal foreshore elevations best approximate the elevation of the lake when the beach ridges formed.

The sand/clay contact forms the pre-depositional surface underneath the strandplain. The slope of the clay surface decreases lakeward but is almost five times steeper between 2000 m and 1800 m than its slope farther lakeward (Fig. 1.4). The clay surface also forms a slightly lakeward-dipping platform between 900 m and 1300 m. The sharp sand/clay contact and coarse sand to gravel commonly found above this contact in core suggests that this surface is erosional. The age or thickness of the clay is unknown, but it was most likely deposited offshore during a previous lake stage and subsequently eroded before progradation and beach-ridge deposition. Between the basal foreshore contact and the clay surface are upper shoreface deposits that accumulated subaqueously. The lower limit of the shoreface corresponds to the depth of average fair weather wave base (Reineck and Singh 1980). Upper shoreface sediment thicknesses range from 0 along the steeply sloping clay surface to as much as 1.94 m. The upper clay surface trends are roughly parallel to basal foreshore trends, except in the most lakeward set where basal foreshore elevations rise slightly and clay upper surface elevations fall, and in the most landward part of the curve where they join.

1.4.3 Age model

We collected 30 peat samples from the base of 25 swales along the Tahquamenon Bay vibracore transects (Table 1.1). All 30 samples (22 conventional and 8 AMS radiocarbon) were used to estimate the ages of beach ridges (Fig. 1.5). Calibrated

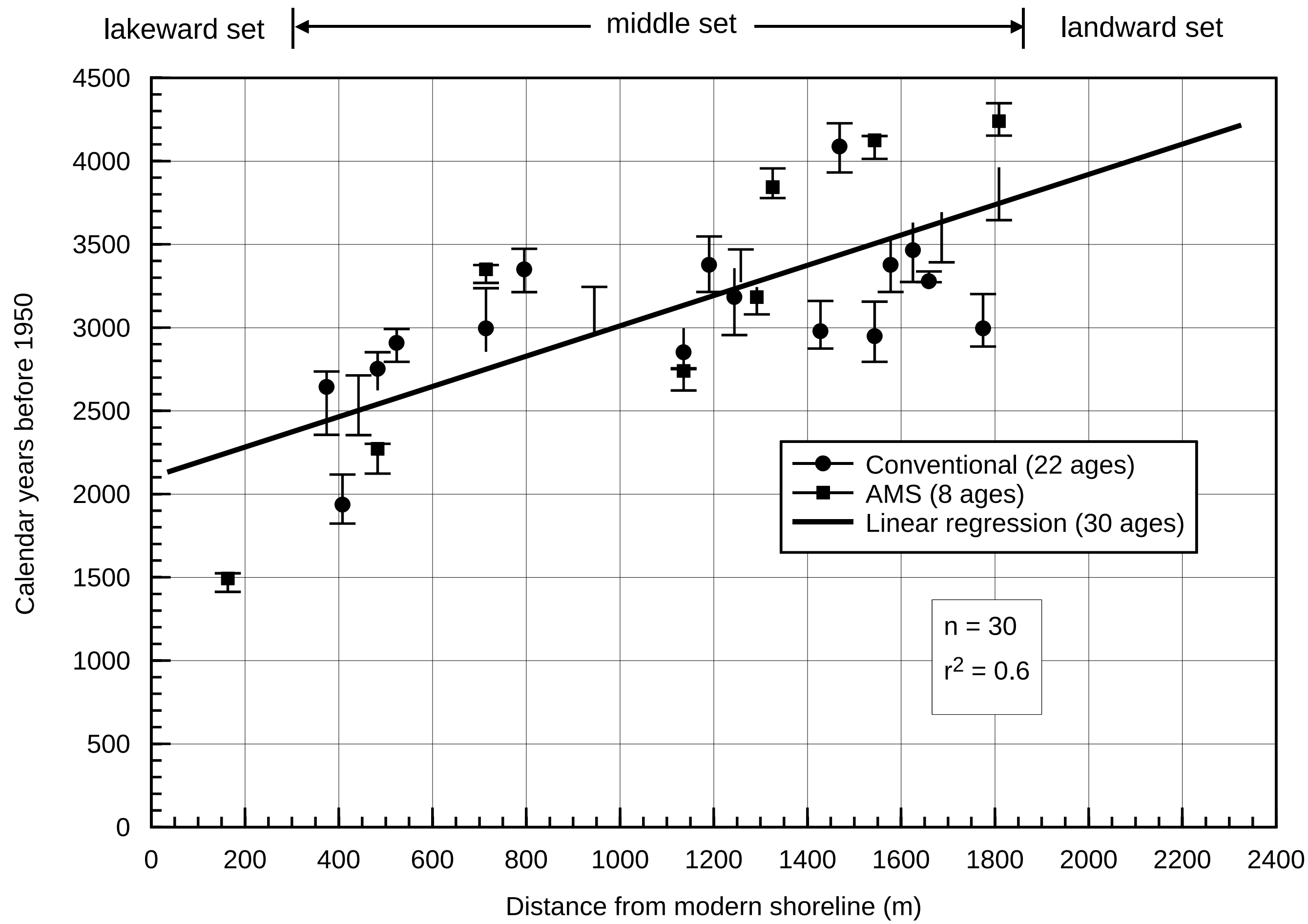

Fig 1.5 Graph of calibrated ages versus distance from the modern shoreline. The regression line was used to calculate approximate ages for each beach ridge in the Tahquamenon Bay embayment.

conventional and AMS ages generally show a similar trend, decreasing in age toward the modern shoreline. This lakeward decline illustrates the progradational nature of the shoreline. Although the conventional samples were not AMS dated, two basal peat samples from five common wetlands were analyzed using the two dating methods. We used two different collection techniques to sample basal peat at various locations within the same wetland. All of the conventional ages except one varied from the AMS ages by 100 to 500 calendar years (older and younger). One AMS age was about 1,200 calendar years older than a conventional age in samples from the same wetland. However, this may be a result of dating a macroscopic charcoal fragment that was transported from older deposits. Conventional ages should be younger than AMS ages; they are an average age representing a larger amount of sample and may be contaminated by modern rootlets. Although it is expected that the AMS ages more accurately reflect the date of peat inception in the swales, there does not appear to be a consistent relationship between AMS and conventional ages across the strandplain.

This disparity between ages poses a problem for manipulation and interpretation. Individual ages cannot be treated as "absolute" because of apparent age reversals that are inconsistent with strandplain development and a progradational system. Fitting complex functions to the data also creates apparent time reversals and apparent periods when no time has elapsed, which cannot be explained in this system. Further, these ages may reflect only the minimum ages of the deposits in the associated beach ridges. For this study, we used least-squares regression through all 30 calibrated ages versus distance from the modern shoreline to create an age model that statistically approximates the age of each beach ridge (Thompson 1992, Thompson and Baedke 1997). This best-fit line was used because of the variability between dates ($r2 = 0.60$) (Fig. 1.5) and it allows us to estimate ages for ridges that were not dated.

The ages of ridges in the landward set of beach ridges were extrapolated from the age model, as we did not collect data in this area. Age estimates correlated to a well-defined Nipissing II geomorphic feature (explained more fully in the next section). We also extrapolated the age of the lakeward set because it was within the range of variability for all ages collected; we have no other data to suggest otherwise. Janzen's (1968) archaeological work at the Naomikong Point site on a peninsula (about 1.5 km to the northeast of the Tahquamenon Bay embayment) uncovered material from a Middle Woodland culture (approximately 2,200 to 1,600 cal. yrs. B.P.). This age corresponds to our age model. We are currently investigating age dating with optically stimulated luminescence to help refine the age model.

In our model, the regression line intersects the modern shoreline (0 meters in figure 1.5) at about 2,100 cal. yrs. B.P. This suggests that beach ridges did not form or were not preserved during the last few millennia in the Tahquamenon Bay embayment, assuming continuous deposition between the middle and lakeward sets. Based on the regression and assuming continuous deposition, the oldest preserved shoreline is about 4,200 cal. yrs. B.P. The inverse of the slope of the regression line in the distance-versus-age plot indicates that the average long-term progradation rate was about 1.10 m/yr. To determine the average time it takes to create a beach ridge, we calculated a regression between the calibrated ages and numbers assigned sequentially to each beach ridge (cf. Thompson and Baedke 1997). The slope of the regression indicates that a beach ridge was created and preserved approximately every 28 ± 4.8 years, which leads us to believe that groups of beach ridges were created and preserved approximately every 140 years.

1.4.4 Relative lake-level variation

We combined upper and lower foreshore elevations with age estimates from the model to create a relative lake-level curve for Tahquamenon Bay (Fig. 1.6A). Beach ridges at Tahquamenon Bay provide information on relative lake-level fluctuations from about 4,200 to 2,100 cal. yrs. B.P., assuming continuous deposition between the lakeward, middle, and landward sets. Basal foreshore elevations indicate that relative lake-levels ranged from about 195.12 m to 182.86 m above sea level (IGLD85). The oldest part of the curve records a lake-level high about 4,100 cal. yrs. B.P. This high lake level corresponds to the Nipissing II high-water level phase when water in all three basins (Lake Superior, Lake Michigan, and Huron) were confluent and the lake drained through the Port Huron/Sarnia and Chicago outlets (Hough 1958; Farrand 1969; Lewis 1969, 1970; Larsen 1985, 1994). Hough (1958) reported the Nipissing II phase as ending about 3,700 cal. yrs. B.P., Lewis (1969) as occurring from about 5,400 to 4,000 cal. yrs. B.P. (4,700 to 3,700 14C yrs. B.P.), Larsen (1985, 1994) as ending about 4,200 cal. yrs. B.P. (3,800 14C yrs B.P.), and Baedke and Thompson (2000) as ending between 4,500 and 3,400 cal. yrs. B.P. And so the general consensus on the end of the Nipissing II phase is about 4,000 years ago, but its actual timing is still debated because of the variability associated with sample types, collection and laboratory methods, calibration, modeling and interpretation procedures, and the different interpretations of sediment types and vertical ground movement. Multiple sedimentological analyses of more complete records

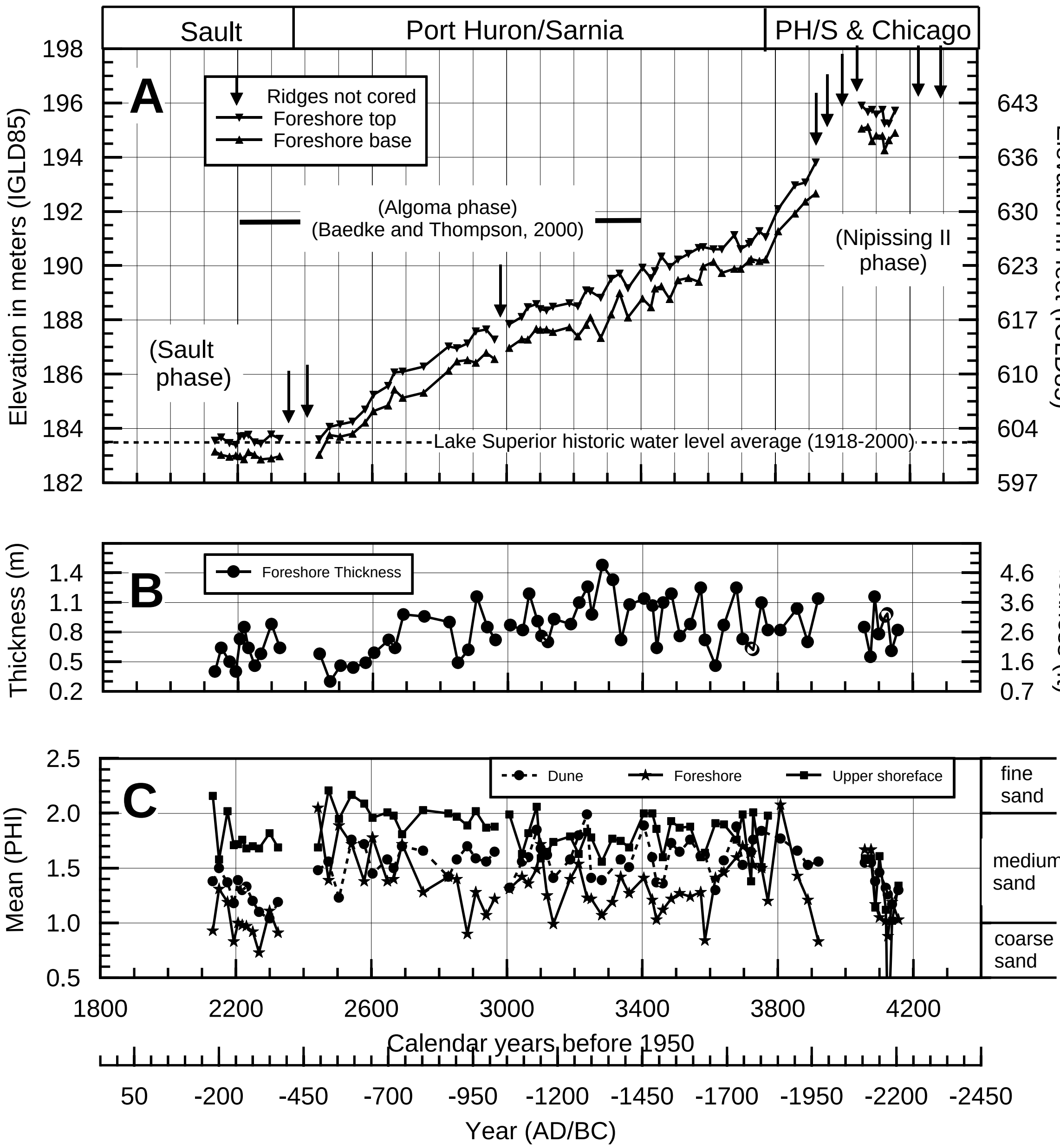

Fig. 1.6 Graph of (A) foreshore (top and base) elevation trends, (B) foreshore thickness, and (C) mean grain size of nearshore (foreshore and upper shoreface) and onshore (dune) facies per core versus calendar years before 1950 across the entire Tahquamenon Bay strandplain. Basal foreshore elevations in (A) indicate the relative elevation of the lake through time. (Note: Basal foreshore elevations in (A) are relative lake-level elevations only with respect to Tahquamenon Bay because vertical ground movement is not removed.)

will help to better define this time period. The end of the Nipissing II phase currently is best constrained by the record of Thompson and Baedke (1997) from a strandplain in Manistique, Michigan, because it, to date, contains the most complete data in a continuous sequence. The record from Tahquamenon Bay also encompasses this time period and helps to correlate between basins.

A rapid relative lake-level drop of at least 4 m, recorded in the most landward part of the Tahquamenon Bay strandplain, is similar in magnitude to the drop recorded by Thompson and Baedke (1997) and suggests commonality between basins. The Nipissing II phase is placed about 200 to 300 calendar years later in the Tahquamenon Bay curve than in Thompson and Baedke's (1997) Manistique curve. The rapid drop in lake level is defined by more ridges at Manistique than at Tahquamenon Bay. Adding additional ridges at Tahquamenon Bay to account for those not formed or preserved or age model errors may account for differences between sites. At some time during this rapid relative lake-level drop, the Nipissing II phase ended and the Chicago outlet closed (Baedke and Thompson 2000). The end of the Nipissing high water-level phase and the closing of the Chicago outlet may be related to erosion at the Port Huron/Sarnia outlet and/or large loss of water from the lake (Baedke and Thompson 2000).

A gradual relative lake-level lowering of approximately 7 m from about 3,800 to 2,400 cal. yrs. B.P. occurs in the middle part of the Tahquamenon Bay strandplain (Fig. 1.6A). This gradual lowering is similar to strandplain trends recorded by Thompson and Baedke (1997) in Lake Michigan. A comparison of the Tahquamenon Bay data to both the Lake Michigan record and the calculated record for the Port Huron/Sarnia outlet (Baedke and Thompson 2000) indicates that the Tahquamenon Bay embayment experienced more vertical ground movement than the outlet regulating water level in the Lake Superior basin. During this time, lake levels were common in all three basins (Lake Superior, Lake Michigan, and Huron basins), and the lake drained through the Port Huron/Sarnia outlet (Farrand and Drexler 1985). Long-term deviations of up to 1 m during the gradual lowering in relative lake level defines the Algoma high-water-level phase. Water-level fluctuations should become more apparent when multiple data sets from around Lake Superior are completed and compared with the Tahquamenon Bay record so that ground movement can be removed (e.g., Baedke and Thompson 2000). Between about 2,300 and 2,100 cal. yrs. B.P., gradual relative lake-level rise of about 0.3 m occurs in the most lakeward part of the Tahquamenon Bay strandplain (Fig. 1.6A). This slight rise indicates that the Tahquamenon Bay embayment experienced similar or slightly less vertical ground movement than the outlet regulating lake levels in the Lake Superior basin at that time. The change in relative water levels from a fall to a rise suggests that the location of the outlet regulating water levels in the Lake Superior basin changed from Port Huron/Sarnia to Sault Ste. Marie. The Tahquamenon Bay record suggests that the outlet change occurred after the Algoma high-water-level lake phase after about 2,400 cal. yrs. B.P. Farrand (1960) called the phase after the Algoma the Sault phase and corresponds to the time when Lake Superior stood at a separate level than Lakes Michigan and Huron.

Farrand (1962) intersected an exponential uplift curve for the Sault outlet and a linear curve representing downcutting at the Port Huron/Sarnia outlet on an age-versus-elevation plot. The intersection, he reasoned, was the time when the Sault outlet rebounded above Lakes Michigan, and Huron. He calculated that the change in outlets occurred at about 2,200 radiocarbon yrs. B.P. Larsen (1994) worked on a strandplain on Whitefish Point, Michigan and reported that the change in outlets occurred at about 2,100 cal. yrs. B.P. Incomplete records, different types of data, varying methods, and errors associated with age models may account for the various age estimates for outlet change. The most likely locations for a time gap in the Tahquamenon Bay record are between sets (lakeward, middle, landward) of beach ridges, where beach ridges did not form or were created and subsequently eroded. Reorientation beach-ridge crests observed in air photographs, abrupt grain-size changes in cores, and reduced accumulation rates or a hiatus recognized in peat cores support the presence of a time gap. However, the lack of ages in lakeward and landward sets do not allow for the confirmation of time gaps in the Tahquamenon strandplain.

Larsen (1999a,b,c) and Larsen et al. (1999) researched submerged features below current lake level along the southern shore of Lake Superior and interpreted lake-level fluctuations after the outlets changed; this work may be useful in understanding the youngest part of the Tahquamenon Bay record after vertical ground movement is removed.

Our data record short-term relative lake-level fluctuations in the Tahquamenon Bay strandplain on the order of decades to centuries. The age model indicates that water levels rose and fell about every 28 years to form an individual beach ridge, and groups of beach ridges suggest that longer-term fluctuations occurred about every 140 years. Similar quasi-periodic fluctuations in Lake Michigan (Thompson and Baedke 1997) suggest that water levels fluctuated with a similar periodicity in both the Lake Superior and Lake Michigan basins. Similar water-level fluctuations in both basins are expected if they experienced similar rates of preservation, because the lakes were confluent during most of the period recorded in the Tahquamenon Bay strandplain.

1.4.5 Cross-strandplain variations in foreshore thickness and mean grain size

Cross-strandplain variations in foreshore thickness and mean grain-size provide insight into past changes in wave and wind climates, and sediment supply and transportation. Foreshore thicknesses range from 0.3 to 1.5 m, averaging 0.8 m at Tahquamenon Bay (Fig. 1.6B). A direct correlation between foreshore thickness and wave climate has not been formulated, but Howard and Reineck (1981) suggest that increased foreshore thickness is related to increased wave energy or average wave height. Because wind duration, speed, and fetch governs wave generation (Komar 1998), foreshore thickness may reveal past predominant wind characteristics. Foreshore thicknesses generally are greater during the Nipissing and Algoma phases than during the Sault phase in the Tahquamenon Bay record (Fig. 1.6A and 1.6B). We expect this relationship because the size of the water bodies was much greater during the Nipissing and Algoma phases (Hough 1958). This increased size would have increased the available fetch and the nearshore water depth, increasing wave height; however, a predominant wind direction, velocity, duration change or a combination of these may also have played an important role in affecting foreshore thickness. A slight rise in foreshore thickness about 3,800 to 3,300 cal. yrs. B.P., an abrupt rise around 3,300 cal. yrs. B.P., and a decrease about 3,300 to 2,400 cal. yrs. B.P. indicate variations in paleo-wave and wind climates or both during the Algoma phase. As water levels rose or stabilized, foreshore thickness generally increased. Short-term variations in foreshore thickness are similar in duration to short-term variations in relative lake level. This suggests that wave and wind climates may have fluctuated every 140 years. The range in magnitude of short-term variations in foreshore thickness and relative lake level decreased after the Sault outlet began to regulate water levels in the Lake Superior basin. This dampening may be related to a decreased size of the lake and changes in the predominant wind direction.

Mean grain size ranges from 0.7 phi (coarse sand) to 2.2 phi (fine sand) in the Tahquamenon Bay strandplain (Fig. 1.6C). This relatively small range in grain size may reflect a single source or numerous homogeneous sources of sediment. The most likely source of sand is the medium to fine-grained friable sandstone of the Munising Formation, which outcrops up-drift along the coastline and along the Tahquamenon River (Hamblin 1958). The only long-term mean grain-size change in all three facies across the strandplain is between the middle and lakeward set. Results from one of the very few grain-size studies across the nearshore in a nearly tideless setting were reported by Fox et al. (1966) in Lake Michigan, and suggest that mean grain size closely reflects the energy level of the wave processes. The coarse grain-size shift after 2,400 cal. yrs. B.P. at Tahquamenon Bay, therefore, may be related to an increase in the energy level of the wave processes; however, foreshore thickness, an indicator of wave climate, contradicts this and remains fairly constant and relatively low during this period. Alternatively, the abrupt change corresponds to the relative water-level change from a fall to a slight rise the middle to lakeward set and may be related to changes in the source of sediment and the distance/direction of transport after the change in outlets regulating water levels in the Lake Superior basin. The only major long-term change in mean grain size within facies occurs between 2,900 and 2,400 cal. yrs. B.P., where the foreshore sediment gradually fines while the dune and upper shoreface sediment remains relatively constant. This change corresponds to a decreasing foreshore thickness and a relative lake-level lowering near the end of the Algoma phase and may suggest an association between long-term mean grain size, foreshore thickness, and water level. Short-term changes in mean grain size are similar in magnitude to short-term changes in foreshore thickness and water-level elevations.

1.4.6 Vertical ground movement

A linear best-fit line was run through the middle set of the age-versus-basal-foreshore-elevation plot (Figure 1.6) to determine a preliminary rate of vertical ground movement for the Tahquamenon Bay embayment. The regression was run only through the middle set because it is the longest part of the record where only one outlet regulated water levels in the Lake Superior basin. The best-fit line slope suggests that the Tahquamenon Bay strandplain rebounded at a rate of 51 ± 1.3 cm/century between 3,800 and 2,400 cal. yrs. B.P. This is almost double the rates calculated from historical and geologic data. The rate of linear isostatic rebound calculated from historic water-level-gauge data is between about 18 and 30 cm/century relative to the Port Huron/Sarnia outlet (Coordinating Committee on Great Lakes Basic Hydraulic and Hydrologic Data [CCBHD] 1977, 2001). Linear rates calculated from late Holocene geologic data reported by Baedke and Thompson (2000) at several sites in the Lake Michigan basin, by Larsen (1994) at a site in the Lake Superior basin, and by Lewis (1970) at a site in the Huron basin compare well with the range in rebound reported from historical gauge data.

All of these studies used a linear age model because a best-fit line of each of their data sets suggests a linear relationship of rebound through time, not an exponential relationship.

The over-steepening at Tahquamenon Bay may be the result of age-model errors or incorrect elevation data. Elevation data were checked on topographic maps and in the field and we found no errors. There is the possibility of age-model error because of the variability observed in the radiocarbon dates. Although we collected a large number of dates, it was difficult to fully ascertain age-model errors because of the numerous variables and uncertainties involved. We are investigating using optically stimulated luminescence for age dating to resolve these problems. We are also studying other strandplains of beach ridges along Lake Superior to compare to the Tahquamenon Bay data set.

Tectonics near the Sault Ste. Marie area may be an alternative explanation for the over-steepening. There are many faults in this area extending from the province of Ontario into Lake Superior (Giblin et al. 1976, Ontario Geological Survey 1991, Manson and Halls 1994, 1997, Manson 1996), but they are not mapped in eastern upper Michigan near the Tahquamenon Bay embayment. Several factors support fairly recent tectonism in the Sault Ste. Marie area: isostatic rebound, earth adjustments after loss of large volumes of water in the Lake Superior basin after the Nipissing and/or Algoma high water-level phases, Michigan structural basin subsidence, Superior province (Canadian Shield) adjustment, and midcontinent rift-related adjustments. More research is needed to resolve these issues.

1.4.7 Future of the Tahquamenon Bay embayment

The large supply of sand from the north (Tahquamenon River and coastal reworking) and the shallow-water platform defining Tahquamenon Bay are ideal for beach-ridge development and preservation. Because the Tahquamenon Bay embayment (as defined in this study) lacks sediment accommodation space, it is likely that future beach ridges will form along the shore of Tahquamenon Bay, which encompasses the Tahquamenon Bay embayment. The presence of scarps on the modern beach, the possibility that Whitefish Point and beach structures (jetties and groins) may capture littoral sediment, and a long-term rising lake level because of vertical ground movement at the lake outlet, however, all are conditions unfavorable to beach-ridge development and preservation. As long-term relative water levels rise in the future, the Tahquamenon Bay embayment will again be inundated and the youngest part of the strandplain record may eventually erode. Human activities as well as tectonic events could modify this long-term trend.

1.5 Conclusion

We have created a relative lake-level curve for Tahquamenon Bay, Michigan, by systematically vibracoring the lakeward margin of beach ridges; using data from the cores, we obtained paleo lake-level elevations and radiocarbon dates of basal wetland sediments between beach ridges to determine ages. The resulting curve indicates that beach ridges in the Tahquamenon Bay embayment recorded lake levels from about 4,200 to 2,100 cal. yrs. B.P., assuming continuous deposition across the entire strandplain.

During this time, relative lake levels dropped rapidly (approximately 4 m) from 4,100 to 3,800 cal. yrs. B.P., lowered gradually (approximately 7 m) from 3,800 to 2,400 cal. yrs. B.P., and remained fairly constant from 2,300 to 2,100 cal. yrs. B.P. The rapid drop from 4,100 to 3,800 cal. yrs. B.P. is associated with a drop in water level at the end of the Nipissing II high-water-level phase; the change from a gradual fall in the middle set to a fairly constant slope in the lakeward set is associated with an outlet change from Port Huron/Sarnia to Sault Ste. Marie. Data from the Tahquamenon Bay embayment strandplain suggest that this outlet change occurred after about 2,400 cal. yrs. B.P. Mean grain-size coarsening after the outlet change suggests that the source of sediment or distance/direction of transport changed as relative water levels started rising.

A line of best fit through the Tahquamenon relative lake-level curve shows that the strandplain is over-steepened with respect to estimated rates of vertical ground movement from historical gauge and geologic data. This over-steepening may indicate an error in the age model or elevation data. More strandplains around Lake Superior are being studied and age-dating sand within beach ridges is being investigated to check for errors. One mechanism for over-steepening of the curve involves tectonism. Tectonism may have modified the Tahquamenon Bay strandplain after about 2,400 cal. yrs. B.P. Tectonism in the Sault area could be related to isostatic rebound, adjustments after the Nipissing and/or Algoma high-water-level phases, Michigan structural basin subsidence, Superior Province (Canadian Shield) adjustment, and/or midcontinent rift-related adjustments. A tectonic event in the Sault area may have also been a factor in the outlet change from Port Huron/Sarnia to Sault Ste. Marie, important because the outlet regulating water levels in Lake Superior. This possibility in the past few millennia raises concern about future events in the area.

In the past, sand eroded from the Munising Formation and transported by the Tahquamenon River and littoral currents around Whitefish Point provided a positive rate of sediment supply to Tahquamenon Bay and it continues today. However, a long-term rising lake level, caused by vertical ground movement at the lake's outlet at Sault Ste. Marie, has not favored beach-ridge development and preservation from about 2,400 cal. yrs. B.P. to the present.

A short-term quasi-periodic lake-level fluctuation with a period of about 28 years was instrumental in the formation of beach ridges in the Tahquamenon Bay embayment. Foreshore elevations rise and fall in groups of three to six beach ridges in each set of ridges observed; we interpret these to represent quasi-periodic fluctuations of longer duration (ca. 140 yrs.). Changes in mean grain size and foreshore thickness follow these longer duration fluctuations and are related to paleo wave and wind climates. Superimposed on these shorter-duration fluctuations are differential vertical ground movements, outflow location changes or restrictions, and Lake Superior hydrodynamics. Long-term records of Great Lakes water-level variability are critical to understanding the potential future magnitude and frequency of water-level fluctuations. Long-term records of water-level variability and vertical ground movement also provide a geologic and climatic framework for paleoecological studies, so that past wetland and terrestrial responses to these changes can be investigated (e.g., Booth et al. 2002, Jackson and Booth 2002). The compilation of multiple sedimentological records of past lake-level variability (Thompson and Baedke 1997, Baedke and Thompson 2000), and the direct comparison of these records to independent paleoclimate records from the region (e.g., Booth and Jackson, in press) should provide insight into the relative importance of the mechanisms driving lake-level variability (e.g., differential ground movement, outlet switching and erosion, climate variability) at centennial to a millennial timescales. Understanding these mechanisms is critical to future management of the Great Lakes water resource.

1.6 Acknowledgements

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CHAPTER 2:

SYSTEMATIC PATTERN OF BEACH-RIDGE DEVELOPMENT AND PRESERVATION: CONCEPTUAL MODEL AND EVIDENCE FROM GROUND PENETRATING RADAR

Abstract

A systematic pattern of beach ridges forming strandplains commonly fill embayments in the Great Lakes of North America. Ground penetrating radar and vibracore results define a common preserved architecture inside beach ridges. Comparing the preserved architecture with a conceptual model of beach-ridge development explains the conditions responsible for their development and preservation. Great Lakes beach ridges are a product of a positive rate of sediment supply and a multidecadal fluctuation in lake level. Many shoreline behaviors are experienced throughout the development of a beach ridge, but not all sequences originally formed by these behaviors are preserved. Beach ridges are stratigraphically separated by concave lakeward-dipping ravinement surfaces, extending at depth below beach-ridge crests to the surface in adjacent landward swales. These surfaces are formed during rapid rises in water-level, where previously laid deposits erode, forming a base for the beach-ridge core. As the rate of rise decreases and the water-level elevation approaches a highstand, the core of the ridge is built by vertical aggradation. Subsequent deposits build lakeward during progradation when water levels become stable, protecting the core from being eroded during future rapid rises in water level. Dune sand deposits on beach-ridge cores are stabilized by vegetation and swales are commonly filled with organic material.

2.1 Introduction

Beach ridges are common features along coastlines of the Great Lakes. Most occur as parallel to sub-parallel ridges of sand separated by intervening swales that often contain organic-forming wetlands. Multiple beach ridges and swales fill embayments from the coasts and form what are called strandplains. Beach ridges are more prevalent in Lakes Superior, Michigan, and Huron because these lakes have abundant littoral sediment supplies, numerous embayments to trap littoral drift, and isostatic rebound patterns that uplift beach ridges away from modern shore processes after they form. Strandplains of beach ridges in the Great Lakes have been used to reconstruct past lake-level changes (Larsen, 1985; Thompson, 1992; Dott and Michelson, 1995; Lichter, 1995; Thompson and Baedke, 1997; Johnston and others, in press), isostatic uplift patterns (Larsen, 1994; Baedke and Thompson, 2000), and long-term shoreline behavior (Chrzastowski and Thompson, 1992; Chrzastowski and others, 1994; Thompson and Baedke, 1995). Most of these studies stress the importance of understanding the subsurface stratigraphy for sound interpretation, especially in determining the origin and processes that formed these beach ridges.

Reviews of beach-ridge development in marine settings (Tanner, 1995; Taylor and Stone, 1996; Otvos, 2000) have emphasized two requirements for the creation of beach ridges: (1) a high rate of sediment supply, and (2) a low offshore gradient. General consensus is that beach ridges consist of water-lain sediment, wind-lain sediment, or both, and become a beach ridge when they are separated from the active shoreline by progradation. The development of multiple beach ridges to form a strandplain requires a complex interplay between hydrodynamic effects and sediment supply. Several researchers debate the factors dominant in initiation, development, and isolation of a ridge and whether the hydrodynamic effect involves changes in water level or wave conditions or both. Thompson and Baedke (1995) combined vibracore data with Curray’s (1964) model of shoreline behavior to create a conceptual model of beach-ridge development for the Lake Michigan coastline. In their model, beach ridges are initiated in the final stages of lake-level rise with dune-cap growth and lakeward translation of the shoreline occurring during the subsequent lake-level fall. They stress that rates of change in water level and sediment supply are important in beach-ridge initiation and development and that shorelines experience several different types of behavior to construct individual beach ridges. Types of shoreline behavior range from progradation to aggradation, depositional transgression and regression, and forced regression (Thompson and Baedke, 1995). However, in such an active environment it is not likely that sequences representing all of these shoreline behaviors are preserved within beach ridges. Understanding the stratigraphy inside a beach ridge is important and necessary.

Beach-ridge stratigraphy in Great Lakes strandplains is difficult to ascertain because of dense vegetation, standing water, limited access, and infrequent exposures. Cores can provide a glimpse into the subsurface at isolated points, but only if the water table is shallow enough to avoid friction of dry sand causing limited penetration. Few studies in the Great Lakes have examined beach-ridge architecture. Fraser and Hester (1977) used exposures and cores to define subsurface facies in a beach-ridge sequence in Lake Michigan for environmental interpretation, and Thompson (1992) used sediment data from vibracores to define facies to infer past lake-level elevations. Multiple vibracores collected through beach ridges provided information at isolated points of interest but do not provide a full stratigraphic framework through the ridges. Ground penetrating radar (GPR) provides continuous views into the subsurface and has shown great potential for use in coastal settings since Leatherman (1987). A number of applications, mainly in marine settings studying barriers, spits, and deltas, have helped initiate only a few studies in the Great Lakes specifically studying strandplains. Dott and Mickelson (1995) were the first to incorporate ground penetrating radar (using 80 MHz antennae) with vibracore data to better interpret Great Lakes beach-ridge stratigraphy in continuous transects along the western shore of Lake Michigan. Johnston (1999) also used GPR (using 100 MHz antennae) to study beach-ridge development in southern Lake Huron. Both studies, however, lacked sufficient resolution (namely, 250 MHz antennae used in this paper) for detailed analysis and integration with a conceptual model to explain beach-ridge formation and preservation.

This study investigates beach-ridge architecture, development, and preservation in embayments along the Great Lakes coastline. The most complete sequence was collected at Au Train Bay, Michigan, along the southern coast of Lake Superior (fig. 2.1), and is the focus of this paper. A model of beach-ridge development presented by Thompson and Baedke (1995) is expanded and used to explain a persistent preserved sequence recorded in beach ridges with information from vibracores and GPR.

2.2 Methods

As a part of a long-term project to understand late Holocene lake-level variations in the upper Great Lakes (Thompson and Baedke, 1997; Baedke and Thompson, 2000; Johnston and others 2000, 2001, 2002, and in press), more than 500 beach ridges have been studied in ten embayments along Lake Michigan and Lake Superior shorelines (fig. 2.1). Most beach ridges were vibracored along the lakeward margin of accessible beach ridges to recover basal foreshore (swash zone) sediments at their highest attainable elevation within beach ridges. These vibracores were cut open, described, photographed,

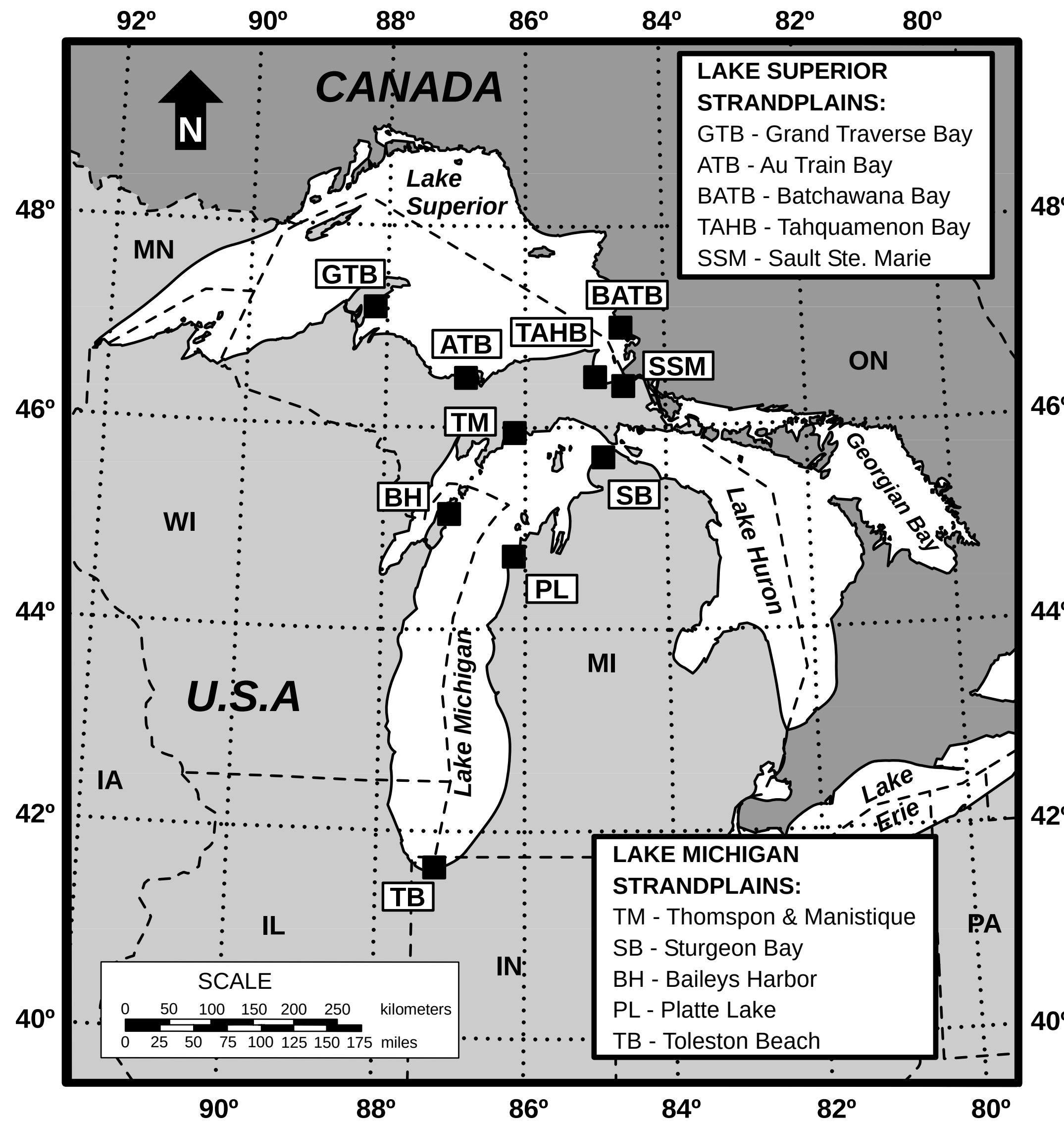

Figure 2.1 Map of the upper Great Lakes showing locations of strandplains studied by the authors (five each in Lake Michigan and Lake Superior basins). Approximately 30 to 100 beach ridges are separated by swales at each site.

and sampled for grain-size analysis. A peel of each core was created using latex and masonite to preserve each core and enhance sedimentary structures. Descriptions were updated to include data from the latex peels. Almost 10,000 grain-size analyses of sand samples from vibracores were completed using 1/2 phi sieves. We computed mean grain-size, sorting, coarsest one-percentile, skewness, and kurtosis values for each sand sample by the mathematical method of moments (Krumbein and Pettijohn, 1938). Results from vibracore data were used to determine three sedimentary facies: dune, foreshore, and upper shoreface. The elevation of the contact between the foreshore and upper shoreface facies provided the lake-level elevation when each beach ridge formed (Thompson, 1992). Sedimentologic data from vibracores was used to help interpret GPR profiles. Nine GPR profiles were collected from four study sites in Lake Superior. Each profile provides a continuous view inside beach ridges of different character (size, shape, sediment grain-size, and age). Although variations exist in each profile, a sequence persists at each study site and between beach ridges. A profile collected from Au Train Bay, Michigan, across five beach ridges is presented in this paper because it contains the most complete preserved sequence.

GPR profiles were collected running reflection surveys using a Sensors and Software Noggin 250 MHz Smart Cart System. Profiles were collected parallel to the depositional dip to record the widest range in variation in subsurface stratigraphy. Profiles were run in step-mode collecting traces every 5 cm, triggered by an odometer wheel, with a stack of 64 measurements at each point. The GPR antennae were parallel to each other, separated by 0.3 m and oriented perpendicular to the direction of travel. A center frequency of 250 MHz was used because it provided the depth range required (3 to

8 m in this material) and offered the resolution necessary to identify stratigraphic boundaries and internal reflection configurations. Profiles were collected across the natural ground surface and away from roads to record the most complete and undisturbed sedimentary sequence. We cleared surface material as much as possible to avoid diffraction effects, and designed the profiles to attain the straightest possible line with the fewest obstacles (i.e. trees). Topographic surveys were recorded along the GPR profiles to correct for topographic variations. The relatively small topographic relief and angle of the slopes along the GPR profile were in the range deemed for topographic compensation and did not seem large enough to create artifacts. Topographic correction can normally account for differences in surface elevation less than the depth of penetration (Annan, 2003). Conversion from travel time to depth was calculated using known elevations in vibracores, hyperbolic matching of subsurface objects, and measured elevations of the water table. Calculated velocities ranged from 0.06 to 0.15 m/ns and compares well with typical values of saturated and unsaturated sand (cf. Annan, 2003). A value of 0.06 m/ns was used in all the GPR profile plots to help in interpretations because the majority of the sequence was below the water table. Data were processed using Sensors and Software WIN_EKKO software to apply a temporal (high-pass) filter (DEWOW) to remove very low frequency components, a time-gain adjustment to increase signal amplification, and a topographic correction to reduce reflection distortions from surface elevation changes.

2.3 Model of beach-ridge development

The only current existing conceptual model of beach-ridge development is the model described by Thompson and Baedke (1995). They used the Curray (1964) model of shoreline behavior relating sediment supply versus rate of water-level change and the Swift (1975) equilibrium profiles to create a conceptual model of beach-ridge formation in Lake Michigan. The model was used to explain the interaction of quasi-periodic lake-level variations and changes in sediment supply for strandplain development to create a regular depositional sequence and irregular erosional discontinuities in Lake Michigan. After studying about 350 additional beach ridges in another basin, the Lake Superior basin, and integrating information from vibracore and GPR data we have reformulated the model of Thompson and Baedke (1995). This section expands upon work by Thompson and Baedke (1995) by clarifying differences between “absolute” and “rates” of change in water level, relating types of shoreline behavior with the development of different sedimentary deposits, and using the model to explore the development and preservation of a persistent sedimentary sequence found in individual beach ridges.

Water level is normally described using absolute elevations (namely, high or low water level). Thompson and Baedke (1995) introduced the importance of understanding rates of change in water level to explain beach-ridge development because of the significance of another competing factor, sediment supply, with water-level change over time. If the competing rates of change keep pace with each other, then the type of shoreline behavior remains; however, but if the rates are out of sink, then the shoreline behaves differently. An explanation is needed to clarify the two and describe how each is depicted in the conceptual model phase-diagram. A simple rise and fall in water level can be illustrated by a sinusoidal curve (fig. 2.2a). Calculating the tangent about the curve produces rates of change. This is graphically represented by the slope of selected lines in the elevation versus time curve (fig. 2.2a). Rates of change are used to define shoreline

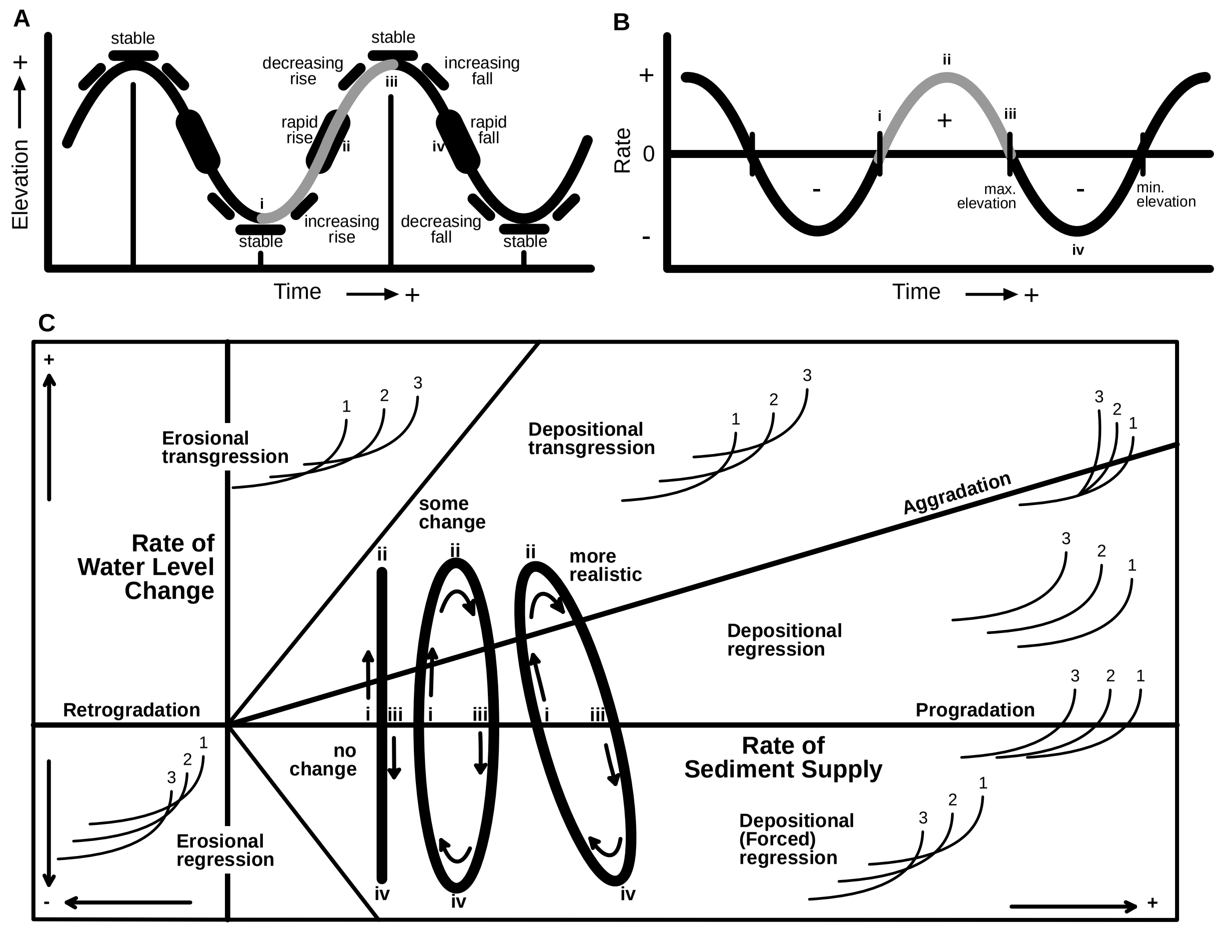

Figure 2.2 Simple rise and fall in water-level (A) and corresponding rates of water-level change (B). Thompson and Baedke (1995) model of beach-ridge development in Lake Michigan, showing types of shoreline behavior in response to rates of water level change and rates of sediment supply (C). Three scenarios of changing sediment supply with a rise and fall in water level (no change in sediment supply, some change in sediment supply, and more realistic relationship between water level change and sediment supply). Roman numerals were added to A, B, and C to help correlate between diagrams.

behavior types in the conceptual model (fig. 2.2b). Comparing common points within each of the curves (labels i through iv in figs. 2.2a and b) indicate the curves are out of phase by one-quarter wavelength. This shows that at high and low absolute water-level elevations the rate of change is zero and the maximum rate of change occurs between high and low absolute water-level elevations. Understanding this concept is crucial for comprehending the conceptual model diagram and how shoreline behavior relates to water-level changes. A simple rise and fall in water level with a constant rate of sediment supply would be represented by a vertical line on the conceptual model diagram (fig. 2.2c). This line would extend upward and downward through the fields depending upon the magnitude of rate in change of water level. Accounting for a rate in change of sediment supply, the vertical line becomes an oval (fig. 2.2c). Depending upon the relationship between rates of sediment supply and water level change, the oval may be more circular or become inclined or both. Thompson and Baedke (1995) inclined the oval toward the sediment supply origin to represent a rapid loss of sediment to the coastal system as lake level rises and river mouths are flooded. Correspondingly, a water-level fall produces an influx of sediment. As one translates around the canted oval, several different shoreline behaviors are experienced.

Thompson and Baedke (1995) point out that at least five different phases of shoreline behavior occur during beach-ridge formation, the most important being the aggradation phase for beach-ridge development. In the aggradation phase, corresponding positive rates of water-level rise and sediment supply produce a stillstand. That is, for every rate of lake-level rise there is a corresponding rate of sediment supply that keeps the shoreline in place, forcing it to vertically aggrade and over-steepen the equilibrium profile. Rates of water-level rise for a given rate of sediment supply above the aggradation line produce landward translation of the shoreline (depositional and erosional transgression). Rates of water-level rise below the aggradation line and rates of water-level fall produce an offshore translation of the shoreline (depositional and erosional regression, and progradation). In the depositional-transgression phase, the shoreline erodes landward but steps upward, preserving sediment between the predepositional surface and the base of the equilibrium profile. Greater preservation occurs closer to the aggradation line where thicker sequences are preserved at lower rates of water-level rise and higher rates of sediment supply. At extremely high rates of water-level rise or low to negative rates of sediment supply, the equilibrium profile cuts into the predepositional surface, stepping landward at an angle less than the slope of the predepositional surface. Depositional transgression and erosional transgression may produce ravinement surfaces that rise toward the basin margin (Swift, 1975; Nummedal and Swift, 1987). In the progradation phase, the rate of water level change is zero, or very small, and the equilibrium profile translates offshore. The progradation phase separates two depositional regression phases. This indicates that regressive sequences can be created during a water-level rise and a water-level fall (forced regression of Posamentier and others 1992). One would expect that regressive sequences that form during a rise will be thicker overall than during a water-level fall. During a water-level fall, the equilibrium profile steps downward and deeper-water deposits are eroded. Although each of the phases described above can be explained and illustrated using equilibrium profiles, the order of events is important for sequence development and preservation.

Thompson and Baedke (1995) describe a sequence of shoreline behaviors responsible for creating beach ridges in Lake Michigan as one translates clockwise around the path of the canted oval (Fig. 2.3a). They also describe how different scales of quasi-periodic variation in lake level are depicted on the conceptual-model diagram to explain strandplain evolution at several sites. They do not address the differences between beach-ridge development and preservation. One would not expect all shoreline behaviors to be preserved inside each beach ridge. A complete sequence of equilibrium profiles representing all of the possible shoreline behaviors that occur during a rise and fall in water level are shown in Figure 2.3b. The sequence of profiles is vertically exaggerated to decipher the individual phases of shoreline behavior. All phases, as described previously, would be experienced, however, three would repeat (aggradation, depositional regression, and progradation) within one rotation around the oval. The model suggests that the complete sequence of profiles would be similar to what is depicted in Figure 2.3b. Starting at a low lake level (i in fig. 2.2a), the shoreline is initially prograding (i in fig. 2.3b). As lake level starts to raise, the shoreline moves from progradation into the depositional regression field (fig. 2.3a). The shoreline translates lakeward but slightly climbs (fig. 2.3b). When the rate of rise is large enough to reach the aggradation line, the shoreline undergoes a stillstand and briefly aggrades. At rates of rise above the aggradation line, the shoreline moves into the depositional transgression field (and possibly the erosional transgression field, depending on the rate of sediment supply) and begins to translate landward (ii in fig. 2.3b). This landward translation removes the aggradational beach and erodes into nearshore deposits of the previous beach ridge, possibly producing a ravinement at the base of the equilibrium profile. The longer the

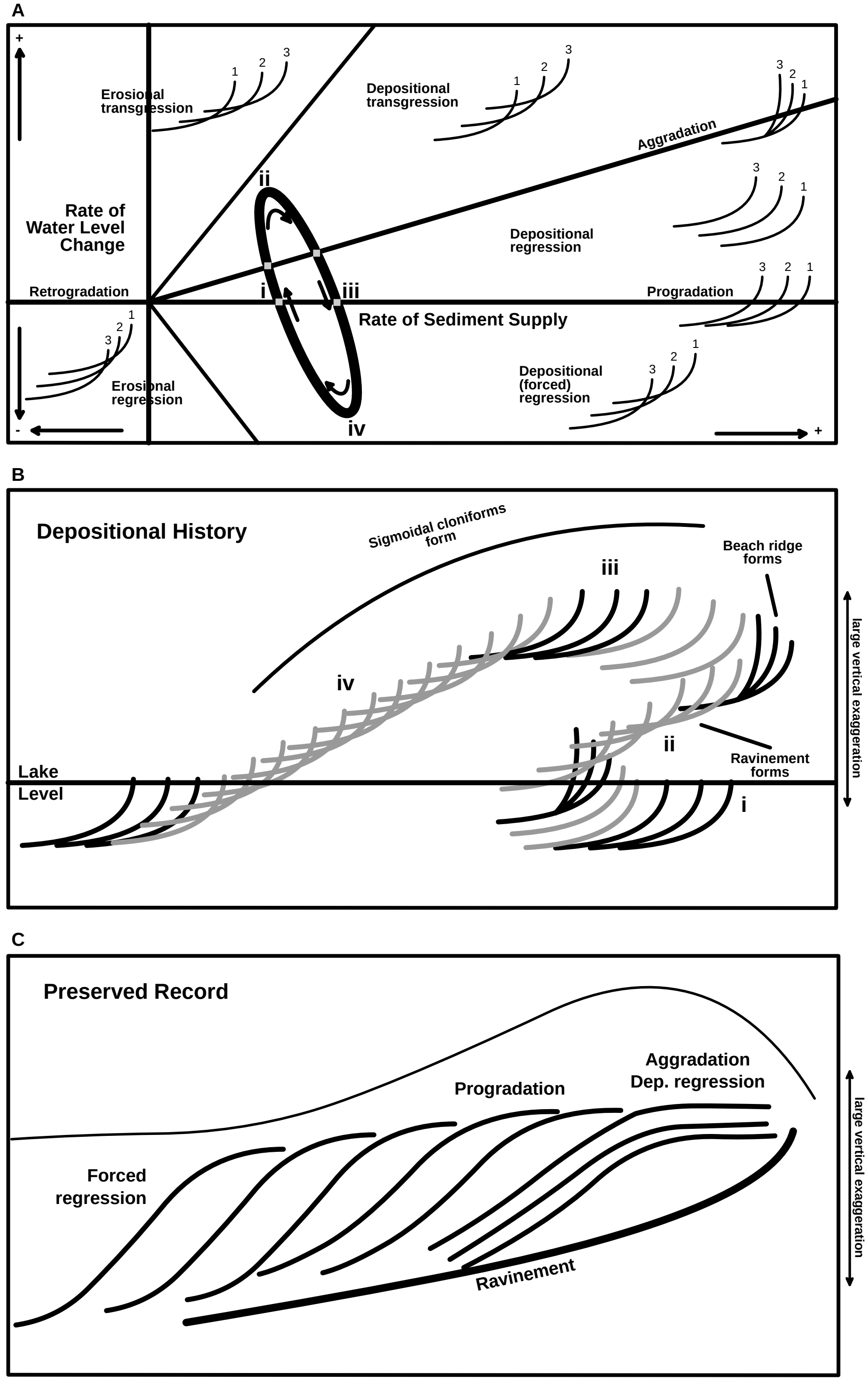

Figure 2.3 Model of (A) beach-ridge development, (B) depositional history of equilibrium profiles during a rise and fall in water level, and (C) potential sequence developed during the creation of a single beach ridge.

shoreline remains in the depositional transgression field, the greater the amount of sediment is moved offshore from the previous ridge and stored in the nearshore as the profile translates landward. As water level reaches its maximum elevation and the rate of rise begins to slow, the shoreline once again contacts the aggradation line. Here, the shoreline aggrades, producing the core of the beach ridge. Continued slowing of the rate of rise forces the shoreline back into the depositional regression field (iii in fig. 2.3a), starting the shoreline to translate back into the lake (iii in fig. 2.3b). At some point during the time the shoreline aggrades and regresses, a berm is colonized with grasses and a dune cap is established on the beach ridge. The subsequent fall in water level moves the shoreline through the progradation line and into the forced (depositional) regression field (iv in fig. 2.3a). Sediment stored in the nearshore during the lake-level rise is moved onshore and the shoreline rapidly translates lakeward. Some of this sediment is also moved onshore, adding to the foredune (Lichter, 1995). For shorelines with a positive rate of sediment supply, this lakeward translation creates distance between beach ridges, possibly insulating the newly formed beach ridge from erosion during the next long-term lake-level rise and fall.

The resultant preserved record not only depends upon what is developed but also on the order of shoreline behaviors that the coast experiences. Accounting for the order and degree of erosion and deposition in the conceptual model, one may expect that the early sequences are affected by the later shoreline behaviors. The three sequences formed during the beginning of rising water levels (during progradation, depositional regression, and aggradation) could be replaced by later sequences formed during a rapid rise in water level (during depositional transgressive or even erosional transgression). This suggests that the core of the preserved beach ridge attained relief during the second aggradational and depositional regression phases during the final stages of a lake-level rise. And successive lakeward-propagating sequences (namely, formed during progradation) protects the beach ridge from being eroded during forced regression or future depositional transgressive or erosional transgressive phases. Therefore, the majority of the preserved record would represent the later events during a rise and subsequent fall in lake level (fig. 2.3c).

2.4 Ground penetrating radar

A 94-m-long profile containing a complete and clear depositional sequence inside beach ridges was collected in the Au Train Bay embayment in Michigan, crossing five beach ridges and intervening swales (figs. 2.4 and 2.5). This profile shows the large-scale architectural features found in several beach ridges. A 35-meter portion of the profile was plotted to show the details within one beach ridge (fig. 2.6). Several reflectors emerge from the GPR profile. The uppermost reflectors, following topography, represent the air and ground waves, respectively. The uppermost horizontal reflector or position where there is a large decrease in amplitude occurs at the water table. Near-surface hyperbolic reflectors opening into the subsurface are responses to near-surface objects such as roots. The majority of the other reflectors are either horizontal or lakeward-dipping to a depth of about 5 meters below the ground surface (v = 0.06 m/ns).

GPR profiles were interpreted using the seismic stratigraphic methods of Mitchum and others (1977) and radar stratigraphic methods of Beres and Haeni (1991) and Van Heteren and others (1994). GPR facies units are identified by groups of GPR

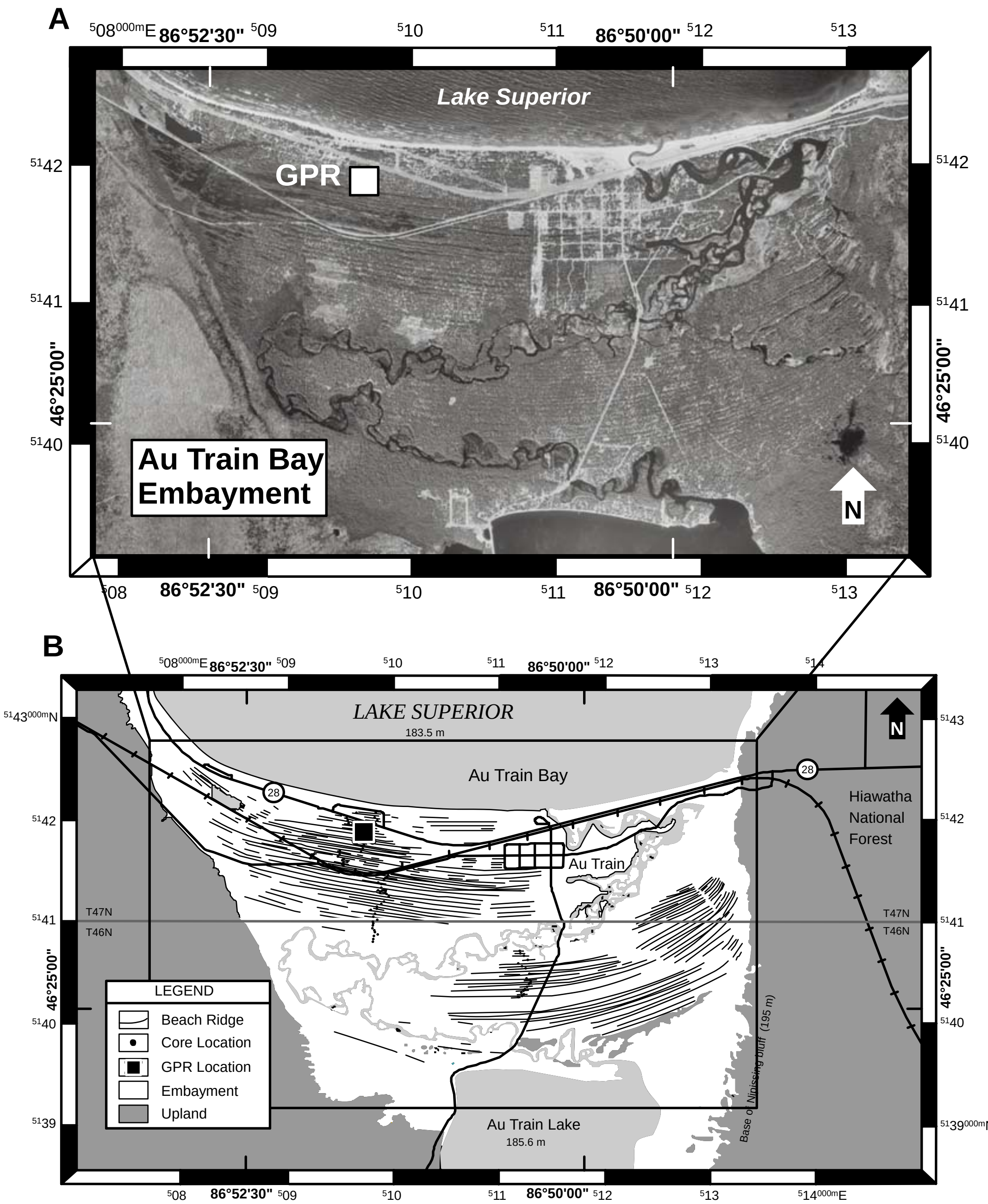

Figure 2.4 The Au Train Bay strandplain, adjacent to Lake Superior. This embayment is filled with beach ridges and intervening swales. Beach-ridge crests are tree-covered, forming arcs in the aerial photograph (A) and are drawn in (B). Swales are dark in color if they have standing water in them or light in color if they are dry (A). Location of GPR profile is labeled. Aerial photograph courtesy of USGS via http://terraserver.microsoft.com (image photographed 04/26/1998).

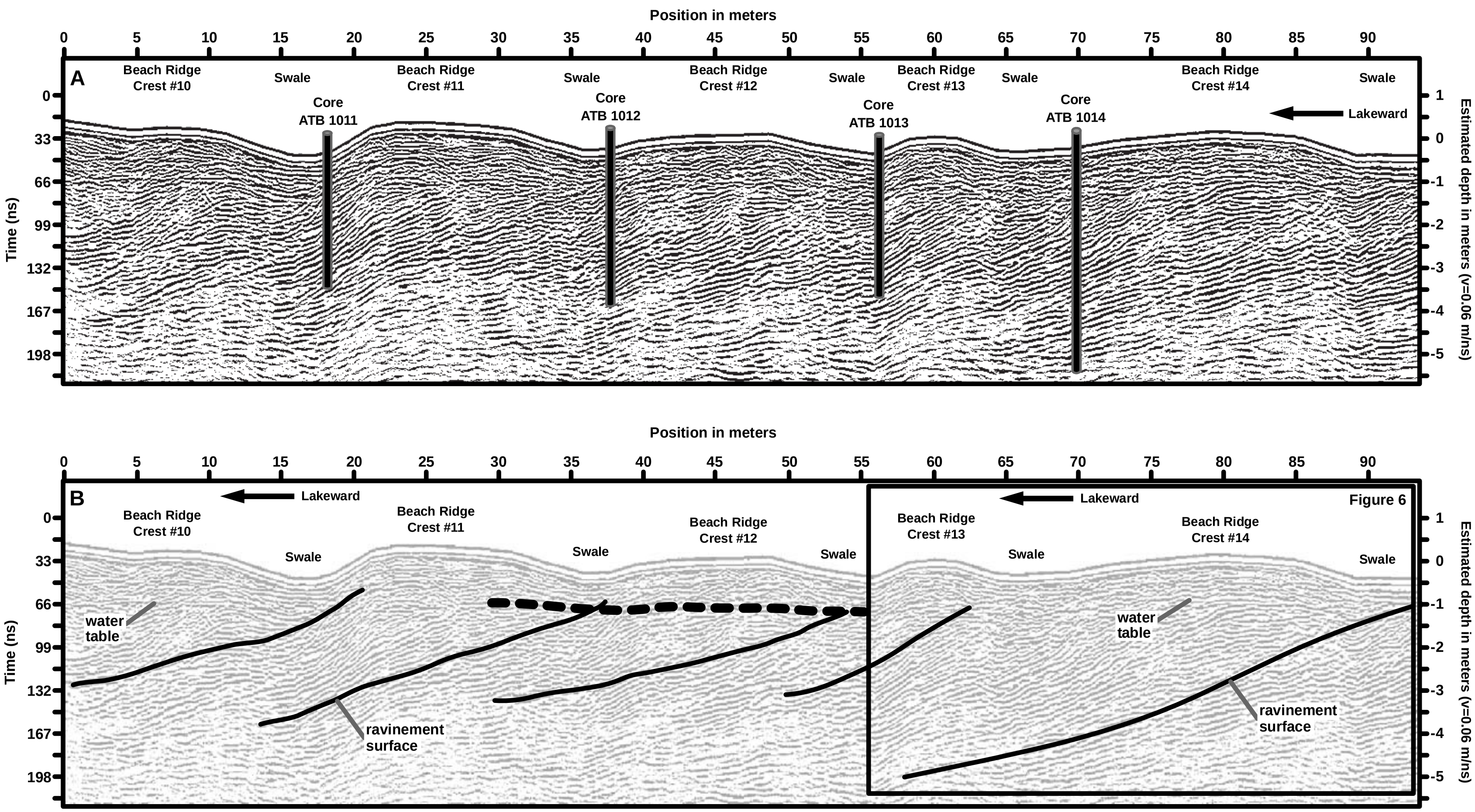

Figure 2.5 Processed (A) and interpreted (B) ground penetrating radar profile collected over five beach ridges and intervening swales at Au Train Bay, Michigan. Vertical exaggeration is 3x. We used 250 MHz antennae and applied a topographic correction, automatic gain control, and a velocity of 0.06 m/ns. Interpretation shows a ravinement surface associated with individual beach ridges. Further interpretation of inset shown in Figure 6.

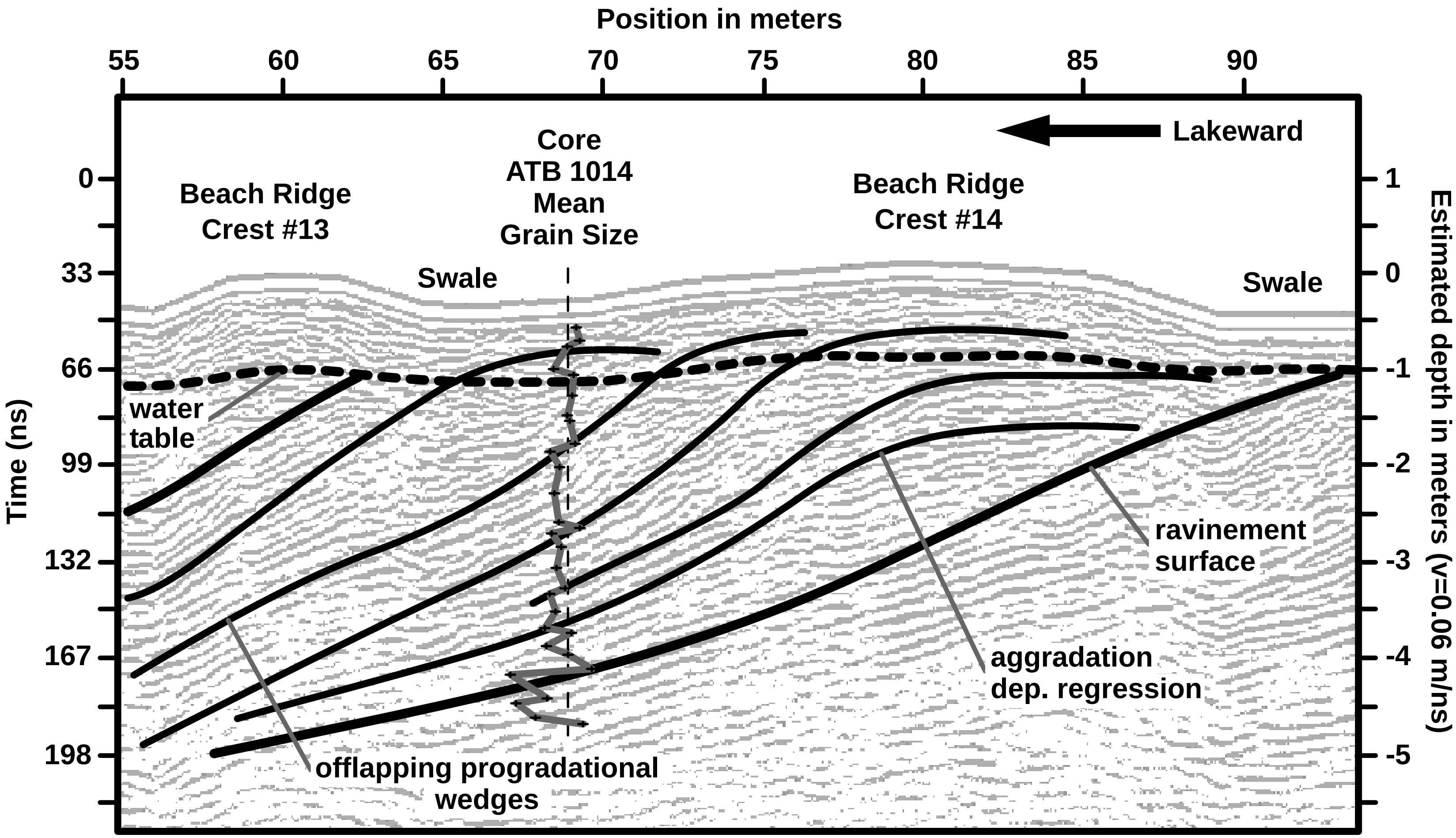

Figure 2.6 Interpreted sequence inside a beach ridge. Vertical exaggeration is 3x. A ravinement surface at depth below the beach ridge crest extends to the surface into the next landward adjacent swale. Sigmoids grow vertically, forming the core of the beach ridge and then translate lakeward. Grain-size changes in core correspond to interpreted sigmoid reflectors. Maximum grain-size range shown is from about one phi (medium sand) to 2.5 phi (fine sand).

reflections whose parameters, such as configuration, continuity, amplitude, and terminations of the reflections, differ from adjacent groups. The most obvious feature of GPR facies is reflection configuration. The main type of reflection configuration identified in the beach-ridge profile is a complex layered or clinoform configuration consisting of mainly low-angle (apparent angle <20 degrees) lakeward-dipping reflectors. Lateral and vertical outbuilding of reflections is prevalent in the Au Train profile. This pattern is expected in an overall prograding depositional system. Large variations in signal amplitude were not observed, except between saturated and unsaturated media, because the sediment consists of a fairly uniform lithology (medium-grained sand). No subsurface reflections are continuous throughout the entire profile, but several shorter reflections forming clinoforms are commonly observed. Discontinuous reflectors were common and may represent variations along surfaces, internal variations, or reflections from near-surface objects. Because of the location of the water table, soil heterogeneities, and near-surface objects reflection terminations of clinoform surfaces were often difficult to ascertain, especially in the upper part. Clinoform surfaces are steepest in the middle and largely terminate at low angles near the bottom and become horizontal near the top. Dip angles are small enough (<20 degrees) that the difference between apparent and actual dips (actual = inverse sin [tan of apparent]) is minimal (within a degree or two) and should not affect interpretations (Jol and Bristow, 2003).

Several clinoform surfaces occur in the Au Train Bay profile (figs. 2.5b and 2.6). A single concave upward reflector per beach ridge extends from near the base of the profile, below the beach-ridge crest to the ground surface in the next landward adjacent swale. At this reflector, cores show a sharp contact overlain by a pebble- to granule-sized lag. This reflector appears to create a boundary between individual ridge sequences. Additional sigmoidal-shaped clinoform reflectors stratigraphically occur above and lakeward of the basal concave upward reflector. These reflectors are truncated by the basal concave upward reflector of the next ridge. The sigmoidal reflectors are associated with grain-size variations in core, becoming either finer- or coarser-grained above and below the contact. Most ridges show an upward and lakeward building of the clinoforms to a position below the foot slope between the ridge crest and the lakeward adjacent swale. Sigmoidal reflectors build lakeward and stratigraphically lower from this point. Several horizontal to lakeward-dipping reflectors terminate between sigmoidal reflectors. Most ridges show a similar reflection pattern. Typical of all ridges is the upward-building under the crest of the ridge. The number of lakeward-building reflectors varies depending on distance between ridges.

2.5 Discussion

Several authors have used GPR in coastal sequences ranging from barriers, to spits, strandplains, and deltas (cf. Neal and Roberts, 2000). A common theme is the recognition of offshore-dipping reflectors. We see similar offshore-dipping reflectors in beach-ridge strandplains but have separated them into two types: concave and sigmoidal. One concave upward reflector occurs per beach ridge (figs. 2.3c and 2.5b). They extend at depth below the beach-ridge crest and extend into the next landward adjacent swale. Because these reflectors truncate adjacent reflectors, they are interpreted to form by landward erosion of the nearshore area. A sharp contact overlain by a pebble- to granule-sized lag observed in cores, correlating to the concave reflectors; support the interpretation that this surface forms from erosion. Dott and Mickelson (1995) identified similar reflectors in a beach-ridge complex along the Lake Michigan shoreline in Wisconsin. They also interpret the reflectors as erosional contacts between successive progradational units. In the conceptual model of beach-ridge development, erosion associated with landward translation of the shoreline occurs during the transgressive phase when the shoreline is experiencing the most rapid rate of water-level rise. We interpret that the concave upward surface formed at this time.

A pattern of sigmoidal reflectors occurs per beach ridge (figs. 2.3c and 2.6). They build predominantly upward and then lakeward. Sigmoidal surfaces truncate internal reflections, suggesting they are also erosional in nature. Changes in grain size observed in core occur at sigmoidal surfaces. Sedimentary deposits between sigmoidal surfaces vary in internal character. In core, sedimentary facies change from more offshore facies to nearshore facies upwards. This pattern corresponds to a lakeward-translating shoreline, where sigmoids predominantly propagate offshore. Facies changes correspond to certain locations within the translating sigmoids. Lateral changes in sedimentary deposits across sigmoidal surfaces separate facies and are isochronous surfaces coincident with bedding interfaces. In the conceptual model of beach-ridge development, vertical growth occurs during a stillstand and lakeward growth occurs during stable water levels. However, two scenarios are possible for the development of the core of the beach ridge in the conceptual model (fig. 2.3a). Vertical growth occurs during the early stages of a water-level rise, as well as the later stages. But close evaluation of the conceptual depositional history (fig. 2.3b) and preserved record using GPR (fig. 2.6) specify the core of the beach-ridge builds during the final stages of a lake-level rise and lakeward propagating sequences build during the subsequent fall in lake level.

The sequence of events can be explained using the conceptual model and GPR profiles but they cannot provide an age for each event or deposit. Preliminary age-model results using radiocarbon ages across the Au Train Bay strandplain suggest beach ridges develop, on average about every 30 years (Johnston and others, 2002). This corresponds to similar timings found in other strandplains in Lake Superior (Johnston and others, 2000, in press) and in Lake Michigan (Thompson, 1992; Thompson and Baedke, 1997). However, preliminary age-model results using optically stimulated luminescence (not at Au Train Bay) suggests a longer timing of beach-ridge development, after Lake Superior separated from Lake Michigan/Huron (Argyilan and others, in preparation). Age models are still preliminary and determination of when the beach ridges formed, described in this manuscript, has not been established. Whatever the case, a maximum limiting age of several decades has been established as the time scale for deposition of a single beach ridge. In other words, the time span between successive concave erosional surfaces and the duration of a lake-level fluctuation responsible for creating Great Lakes beach ridges is many decades. Difficulty in refining this time span today is controlled by limitations of age-dating. The time it takes to create individual sigmoidal surfaces and the sediment in between has to be less than several decades. The best-known variations in lake level in the Great Lakes are annual fluctuations. Winter beaches are generally erosional and may relate to the sigmoidal surfaces; summer beaches are generally depositional and may relate to the deposits in between. If this were the case, we would then expect to have about 30 sigmoidal packages in a perfectly preserved record per beach ridge. We see, however, only a fraction of this in the preserved record per beach ridge. If the sigmoids are annual, only the early history of each beach ridge is preserved. This is because the later part of the sequence would be eroded as the next concave upward surface forms. Storms may also contribute to creating the preserved sequences. Internal structures in core are commonly low-angle and lakeward-dipping that varies in grain size and often has isolated coarser grains in a finer matrix. These preserved sequences may represent past high-energy wave conditions. Fraser and others (1991) also found a predominance of high-energy deposits preserved in Lake Michigan coastal sedimentary sequence. They attribute these conditions to storm-driven processes. Few landward-dipping features resembling washover events or bar migration were observed.

Cores have been collected consistently on the lakeward side of beach ridges, by the authors of this manuscript for late Holocene lake-level analysis in Lake Michigan and Lake Superior (Thompson, 1992; Thompson and Baedke, 1997; Johnston and others 2000, 2001, 2002, and in press). Beach ridges are cored on the lakeward side because of the difficulty in coring on top of beach-ridge crests, as well as trying to collect basal foreshore (swash zone) sediment at their highest attainable elevation. Basal foreshore elevations are chosen to represent the elevation of past lake-levels because of the direct relationship between these sediments and lake level on the modern coastline. Evaluation of the GPR data and in light of our conceptual-model indicate that the lakeward position is adequate for recording the greatest vertical relief in beach-ridge development. It also suggests that this elevation represents the final stages of a multi-decadal water-level rise where the rate is decreasing (approaching a high and stable water-level elevation) and the shoreline is experiencing aggradation or possibly depositional regression. At this time, high-resolution GPR will not replace vibracoring methods to determine the elevation of past lake levels, because velocity determination errors in GPR are greater than coring and laboratory errors and reflectors are not facies specific. High-resolution GPR is best used to extrapolate between cores and define beach-ridge architecture.

Progradation rates are often calculated for shorelines around the world; however, the integration of high-resolution GPR results with information from vibracores and a conceptual model allows us to identify several different shoreline behaviors, other than just progradation, and relate these to different preserved sedimentary deposits. This increases the resolution, giving a better context for what has been commonly measured and classified as progradation, and permits the evaluation of lateral, vertical, and combinations of lateral and vertical variations in shoreline development and preservation.

2.6 Conclusion

Great Lakes beach ridges are a product of a positive rate of sediment supply and a multidecadal fluctuation in lake level. They normally form in embayments that have gentle offshore gradients. Ground penetrating radar results calibrated to vibracore data and combined with a conceptual model define a pattern of beach-ridge development and the resulting preserved record. A systematic pattern of beach-ridge architecture consistent with the model of beach-ridge development was identified using high-resolution GPR. Concave upward lakeward-dipping ravinement surfaces, extending from beneath beach-ridge crests to the next landward adjacent swale form the base for each beach-ridge sequence. Sigmoidal-shaped packages of sediments extend upward above the ravinement surface, giving them topographic relief, forming the core of the beach ridge. Several sigmoidal-shaped packages develop lakeward from the core. Architectural patterns observed in preserved beach ridges were compared to a conceptual model of beach-ridge development by Thompson and Baedke (1995). Not all shoreline behaviors experienced throughout their depositional history were represented by deposits in the preserved record. Ravinement surfaces form during depositional and erosional transgression, specifically during a rapid rate of water-level rise. Vertical relief is attained during aggradation, on the final stages of a water-level rise, as the rate of rise decreased. Lakeward translation forms during progradation and depositional regression, as water levels fall.

The preserved sequence found inside beach ridges records only part of the shorelines long-term behavior. Large parts of the sequence may have been developed but were not preserved. A simplistic model of beach-ridge development has been used to explain the developmental history and preserved sequence. However, changes in the predepositional surface (namely, slope and makeup) and type, size, sorting, and availability of sediment would all change beach ridge and strandplain character. The simplified explanation serves as a start and hopefully a catalyst for future models with increasing complexity.

2.7 Acknowledgments

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CHAPTER 3:

GEOMORPHIC AND SEDIMENTOLOGIC EVIDENCE FOR THE SEPARATION OF LAKE SUPERIOR FROM LAKE MICHIGAN AND HURON

Abstract

A common discontinuity was recognized in four Lake Superior strandplain sequences using geomorphic and sedimentologic characteristics. Cross-strandplain elevation trend changes from a lowering in the landward set to a rise or shallowing in the lakeward set indicates the discontinuity is associated with an outlet change for Lake Superior. Correlation of this discontinuity between study sites and age model results for the strandplain sequences suggest the outlet change occurred closer to the present than previously thought; about 1,200 years ago after the post-Algoma phase instead of about 2,200 years ago after the Algoma phase. The landward part of the strandplain was deposited when water levels were common in all three upper Great Lakes basins (Superior, Huron, and Michigan) and drained through the Port Huron/Sarnia outlet. The lakeward part was deposited after the Sault outlet started to control water levels in the Lake Superior basin. The landward set of beach ridges are commonly better defined and continuous across the embayments, more numerous, larger in relief, wider, have a higher vegetation density, and intervening swales contain more standing water and peat than the lakeward set. A change in direction and increased channel width of streams that drain through the strandplains into Lake Superior help to identify the separation in strandplain sequences. Coarsening and foreshore thickness increases at several sites indicate that the sediment source changed possibly in response to changes in littoral transport directions and wave climate increased after the lakes separated.

3.1 Introduction

The youngest ancient shorelines adjacent to Lake Superior have previously been classified as belonging to one of three lake phases: Nipissing, Algoma, or Sault (Farrand 1960). Reconstructed water planes, based on the elevation of coastal geomorphic features, indicate that the three upper Great Lakes (Superior, Huron, and Michigan) were joined during the Nipissing and Algoma phases (Leverett and Taylor 1915; Hough 1958; Farrand and Drexler 1985). Current interpretation places the end of the Nipissing II phase at about 4,000 years ago (Hough 1958; Farrand 1969; Lewis 1969, 1970; Larsen 1985, 1994; Baedke and Thompson 2000; Johnston et al. in press) when lake level fell four meters. The cause of the end of the Nipissing II phase is unknown but it corresponds to the closing of the Chicago outlet and may be related to erosion at the Port Huron/Sarnia outlet (Leverett and Taylor 1915; Hough 1958) or large loss of water from the lake related to climate or both (Baedke and Thompson 2000). The Sault phase is defined by Farrand (1960) as the time period when the water body in the Lake Superior basin stood separate from that in the Lake Huron because of a sill in the St. Marys River (near Sault Ste. Marie) that is topographically above the downstream lakes. The mechanism for the separation is attributed to isostatic rebound (Farrand 1960) but may also be related to faulting at the sill (Johnston et al., in press). Farrand (1962) calculated the age of the separation by intersecting an exponential uplift curve for the Sault outlet with a linear curve representing downcutting at the Port Huron/Sarnia outlet on an age-versus-elevation plot. His age estimate is 2,200 radiocarbon years before present. Larsen (1994) working on a strandplain on Whitefish Point, Michigan, reported a similar timing of separation. Data collected by Johnston et al. (2000) from a strandplain at Grand Traverse

Bay, Michigan indicate that the separation of the lakes actually occurred closer to 1,200 calendar years ago. This age is much different than previous findings and suggests that the Sault phase should start about 1,000 years later than proposed by Farrand (1960). Johnston et al. (in press) found a reorientation of beach-ridge crests in aerial photographs and an abrupt grain-size change in core in the strandplain sequence at Tahquamenon Bay with an age similar to those reported by Larsen's (1994) and Farrand's (1962). However, Johnston et al. (in press) reported that the separation of the lakes occurred after about 2,400 years ago because of a time gap indicated by missing ridges in the Tahquamenon Bay strandplain. The exact timing of separation is currently unknown. Refining the timing of the separation of the lakes has been partially limited because there is a lack of continuous data sets (missing ridges in strandplain sequences) that cross this important time period.

Strandplains of beach ridges provide some of the most continuous sedimentary records during the late Holocene. Only three Lake Superior strandplains have been studied to address the separation of the lakes (Larsen 1994; Johnston et al. 2000, in press). The separation was identified in the strandplain sequences by a change in trend of cross-strandplain topographic and foreshore contact elevations. Shorelines deposited before the separation of the lakes sequentially decrease in elevation toward Lake Superior. This pattern occurs because the shorelines isostatically rebounded faster than the active outlet at Port Huron/Sarnia. Shorelines deposited after the separation show no topographic change if the site is near the Sault outlet or sequentially increase in elevation for sites west of the Sault outlet. For the sites west of the outlet, the Sault outlet is rising more rapidly than the site. Although these general trends were recognized within strandplains two different approaches for determining long-term water level elevations were formulated. The sedimentological approach of Thompson (1992) employed by Johnston et al. (2000, in press) uses basal foreshore elevations and the geomorphic approach (Larsen 1994) uses beach-ridge topography. The geomorphic approach is used to provide a fast and reasonably accurate estimate of the elevation of past lake level and isostatic rebound (Larsen 1994). However, the sedimentological approach provides more accurate results because lake level at the time of beach-ridge development can be more closely determined from basal foreshore elevations (Thompson 1992; Thompson and Baedke 1997). Changes in topography have been shown to not necessarily coincide with changes in basal foreshore elevations (Thompson 1992). Regardless of their accuracy in determining past lake-level elevations, both methods provide data that is instrumental in establishing the position of the separation of the lakes in the strandplain sequence and information on changing patterns of shoreline behavior in response to new lake level scenarios.

This paper presents geomorphic and sedimentological evidence for a common discontinuity in the Lake Superior strandplain sequences associated with the separation of Lake Superior from Lake Michigan and Huron. Such evidence includes beach ridge topography, relief, and spacing; and facies elevations, thickness, and grain-size properties. Although several characteristics help target the separation of the lakes a subsurface sedimentary contact that has a direct correlation with the elevation of the past lake level is argued most accurate for isostatic rebound and water level calculations.

3.2 Study Area and Methods

Four embayments were studied along the Lake Superior shoreline, Batchawana Bay in Ontario and Tahquamenon Bay, Grand Traverse Bay, and Au Train Bay in the upper peninsula of Michigan (Fig. 3.1). These study sites were chosen because they have a large number of preserved beach ridges (> 70) and, therefore, potentially contain records of long duration. Beach ridges were traced from aerial photographs of the embayments to determine the number, orientation, and spatial extent. A total of 294 beach ridges were vibracored at these four sites following the methods described by Thompson et al. (1991). Ground-surface elevations at each core site were surveyed and corrected to the International Great Lakes Datum of 1985 (IGLD85) using the closest water level gauging stations. The elevation of beach ridge crests and swales were surveyed at two study sites (Batchawana Bay and Au Train Bay). Distance from the modern shoreline was calculated from maps created by tracing beach ridges from aerial photographs and global positioning system measurements recorded at core sites. Vibracores were transported to a laboratory for study. Each core was split open and visually described for grain size, lithology, color, structures, bedding, and any other distinguishing characteristics. One half of the core was photographed, latex peeled, and stored for future reference; and the other half was sampled at selected contact boundaries for grain-size analysis. Approximately 5,000 grain-size samples, averaging about 1,200 samples per study site were sieved using a ½ phi interval from gravel to sand. The following statistical parameters were calculated for each sample: mean, standard deviation, coarsest one-percentile, skewness, and kurtosis. Visual descriptions, photographs, and grain-size results were integrated to define three facies (dune,

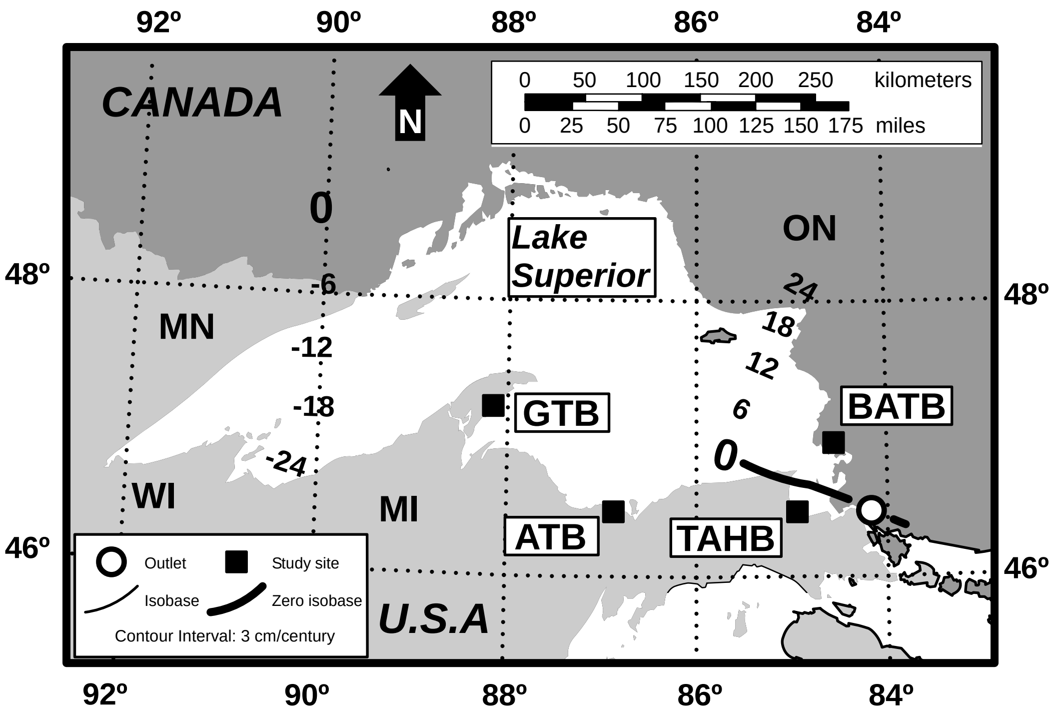

Figure 3.1 Map of study sites in Lake Superior showing location of strandplains studied. They are in the Batchawna Bay (BATB) embayment in Ontario and the Tahquamenon Bay (TAHB), Grand Traverse Bay (GTB), and Au Train Bay (ATB) embayments in Michigan. Contours of isostatic rebound are overlain with rates relative to the outlet at Sault Ste. Marie (modified from the Coordinating Committee on Great Lakes Basic Hydraulic and Hydrologic Data, 2001).

foreshore, and upper shoreface). Grain-size and sedimentary structure trends across the modern shoreline and their relationship to lake level at each study site were used to help define facies relationships per site (cf. Thompson and Baedke 1997 and Johnston et al., in press). The most consistently useful properties to determine facies were sedimentary structures and grain-size parameters. Establishing facies contacts in cores from Lake Superior relied on the combination of more parameters and was more complex than those examined by the authors in Lake Michigan (c.f. Thompson et al. 1991; Thompson 1992; Thompson and Baedke 1997) or Lake Huron (c.f. Johnston, 1999).

3.3 Results

Each strandplain were divided into landward and lakeward sets because of a discontinuity in the strandplain sequence at each study site. Geomorphic characteristics that change between sets at a majority of sites include cross-strandplain topography, drainage patterns, vegetation density; ridge and swale lateral continuity, average relief and width; and presence of standing water and peat in the intervening swales (Table 3.1). The landward set of beach ridges are commonly more laterally continuous across the embayments, more numerous, larger in relief and wider with a higher vegetation density and swales contain more standing water and peat than the lakeward set. The position between landward and lakeward sets is often associated with a bend in a drainage, a drainage width increase, and a cross-strandplain change in topography. Beach ridge crest and swale surface elevations lower in the landward set and rise or become shallower in the lakeward set towards the lake (Fig. 3.4, 3.5). The only geomorphic characteristic that is not consistent between most sites is ridge and swale orientation. An obvious orientation

Study Site	Grand Traverse Bay		Tahquamenon Bay		Au Train Bay		Batchawana Bay	
	Lakeward	Landward	Lakeward	Landward	Lakeward	Landward	Lakeward	Landward
GEOMORPHOLOGY								
Cross-standplain trend in topography (towards Lake Superior)	rises (~1 m)	lowers (~5 m)	slight rise (~horizontal) (~1 m)	lowers (~13 m)	rises (~2.5 m)	lowers (~3 m)	lowers (shallower) (~1 m)	lowers (steeper) (~20 m)
Number of ridges	14	56	13	67	14	69	9	72
Distance in meters	350	1825	300	2026	370	1402	200	2316
Drainage characteristics between landward and lakeward sets	Deer Lake creek bends & width increases; Slough		Naomikong creek bends & width increases		Au Train river bends & width increases		Carp creek bends & width increases	
Estimated vegitation density	lower	higher	lower	higher	lower	higher	lower	higher
Ridge and Swale:								
a) Lateral continuity	continuous to discontinuous	more continuous	continuous to discontinuous	more continuous	continuous to discontinuous	more continuous	continuous to discontinuous	more continuous
b) Orientation	~(N-S)	~(N-S)	~(E-W)	~(ESE-WNW)	~(E-W)	~(E-W)	~(E-W)	~(E-W)
c) Average relief	lower	higher	lower	higher	lower (0.81 m)	higher (1.12 m)	lower (0.37 m)	higher (0.61 m)
d) Average width	narrower (26 m)	wider (33 m)	narrower (26 m)	wider (30 m)	similar (20 m)	similar (20 m)	narrower (24 m)	wider (32 m)
Swale character:								
a) Standing water	no	yes	no	yes	no	yes	no	yes
b) Peat present	no	yes	yes (one swale)	yes	no	yes	no	yes
SEDIMENTOLOGY								
Cross-standplain trends: (towards Lake Superior)								
a) Foreshore top	rises	lowers	rises	lowers	rises	lowers	lowers	lowers
b) Foreshore base	rises	lowers	rises	lowers	rises	lowers	lowers	lowers
c) Average foreshore thickness in meters (variability)	1.33 (lesser)	1.29 (greater)	0.61 (lesser)	0.86 (greater)	1.87 (lesser)	1.66 (greater)	1.10 (lesser)	0.73 (greater)
d) Average dune facies grain size (variability)	medium sand 1.52 phi 0.35 mm (lower)	medium sand 1.77 phi 0.29 mm (higher)	medium sand 1.27 phi 0.41 mm (lower)	medium sand 1.60 phi 0.33 mm (higher)	medium sand 1.91 phi 0.27 mm (lower)	medium to fine sand 2.00 phi 0.25 mm (higher)	medium sand 1.91 phi 0.27 mm (lower)	fine sand 2.15 phi 0.23 mm (higher)
e) Average foreshore facies grain size (variability)	medium sand 1.09 phi 0.47 mm (lower)	medium sand 1.45 phi 0.37 mm (higher)	coasre sand 0.99 phi 0.50 mm (lower)	medium sand 1.37 phi 0.39 mm (higher)	medium sand 1.85 phi 0.28 mm (lower)	medium sand 1.85 phi 0.28 mm (higher)	medium sand 1.30 phi 0.41 mm (lower)	medium sand 1.87 phi 0.27 mm (higher)
f) Average upper shoreface facies grain size (variability)	medium sand 1.51 phi 0.35 mm (lower)	medium sand 1.80 phi 0.29 mm (higher)	medium sand 1.77 phi 0.29 mm (lower)	medium sand 1.85 phi 0.28 mm (higher)	medium sand 1.86 phi 0.28 mm (lower)	medium sand 1.78 phi 0.29 mm (higher)	fine sand 2.95 phi 0.13 mm (lower)	fine sand 2.99 phi 0.13 mm (higher)
g) Overall average grain size trend	coarser	finer	coarser	finer	similar	similar	coarser	finer

Table 3.1 Geomorphic and sedimentologic characteristics of landward and lakeward sets in four strandplains studied along the Lake Superior coastline.

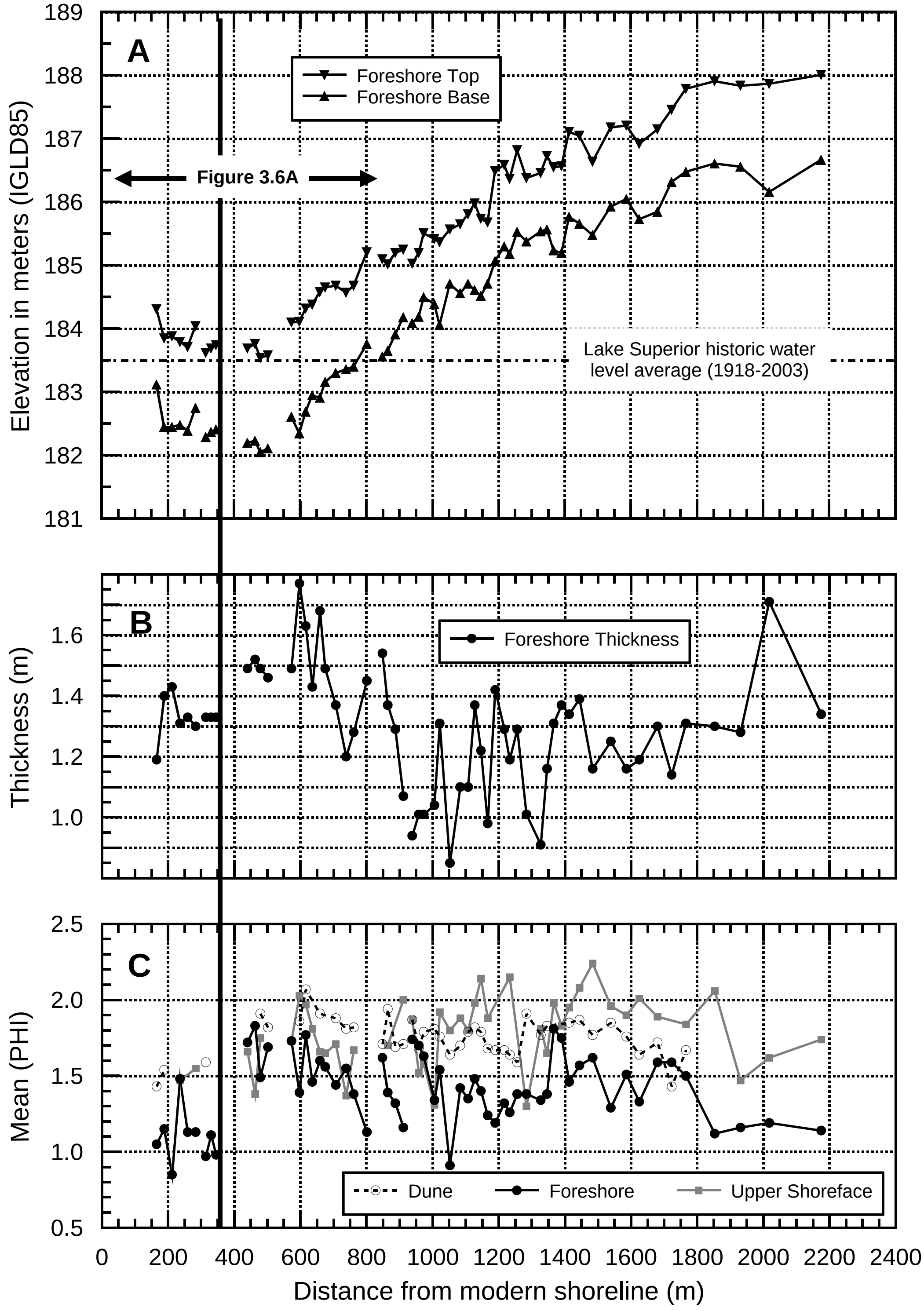

Figure 3.2 Graphs of A) facies contact elevations, B) foreshore thickness, and C) mean grain size per facies from the Grand Traverse Bay strandplain in Michigan. Gaps in data sets are locations where beach ridges were not cored.

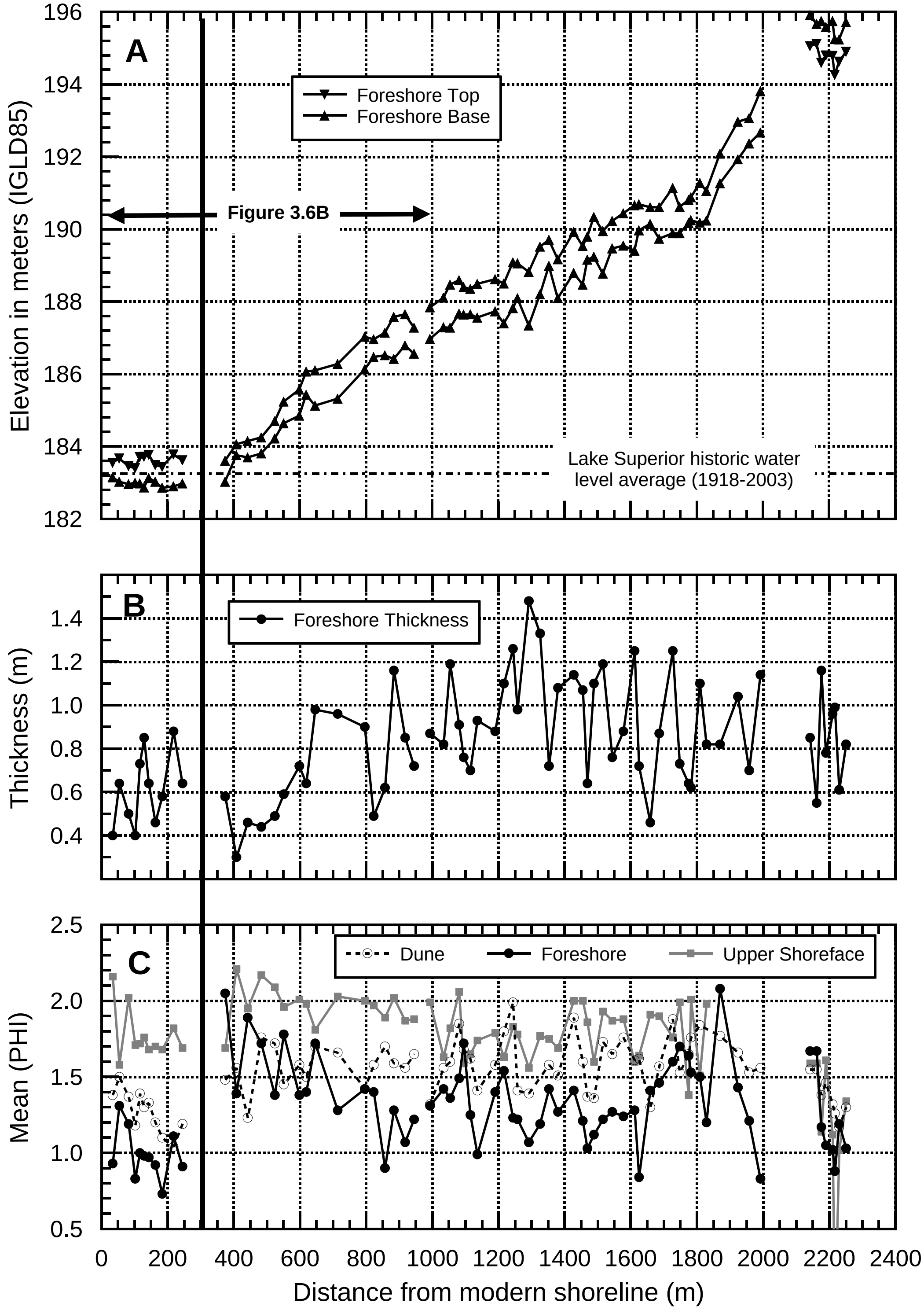

Figure 3.3 Graphs of A) facies contact elevations, B) foreshore thickness, and C) mean grain size per facies from the Tahquamenon Bay strandplain in Michigan. Gaps in data sets are locations where beach ridges were not cored.

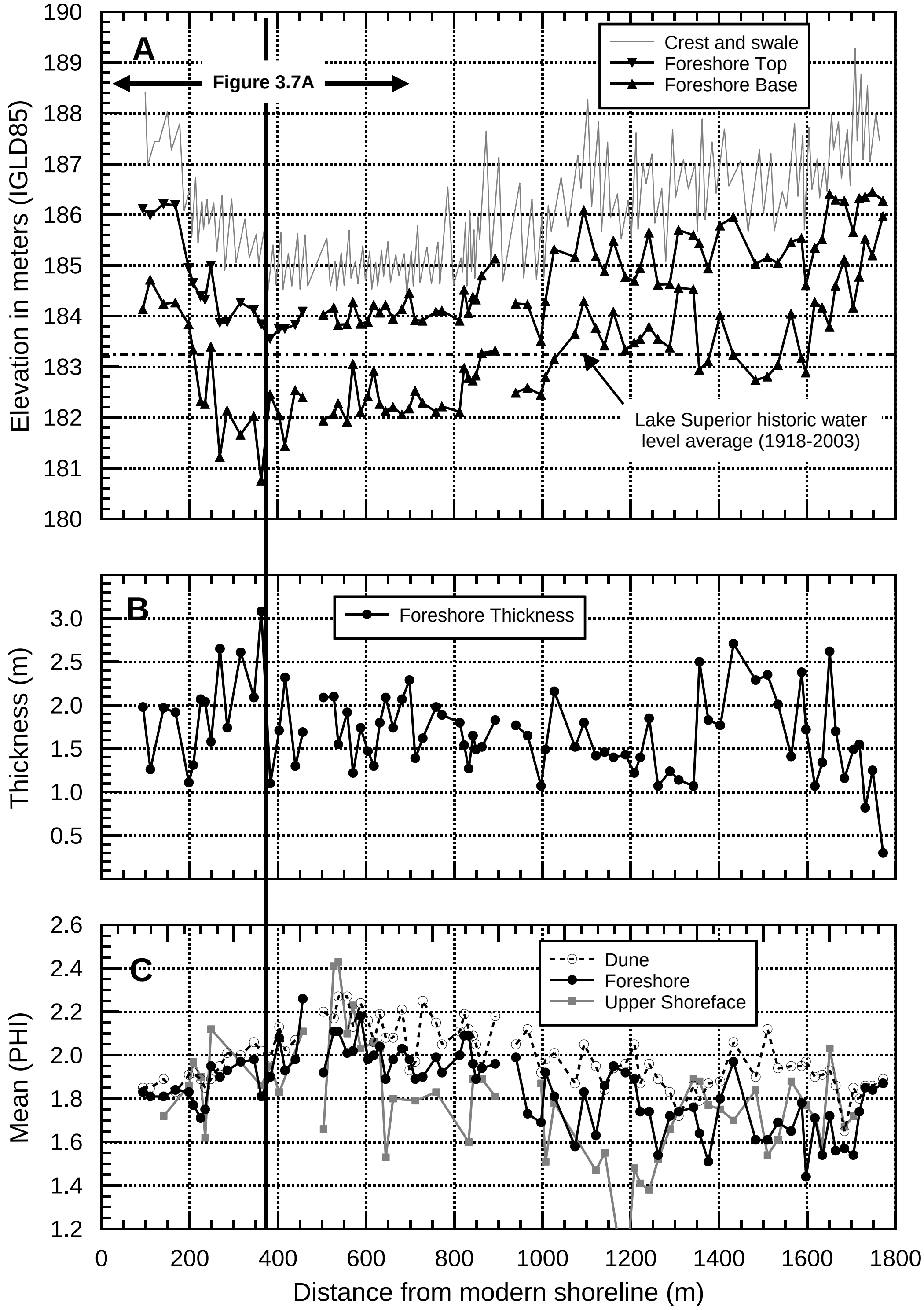

Figure 3.4 Graphs of A) crest, swale and facies contact elevations, B) foreshore thickness, and C) mean grain size per facies from the Au Train Bay strandplain in Michigan. Gaps in data sets are locations where beach ridges were not cored.

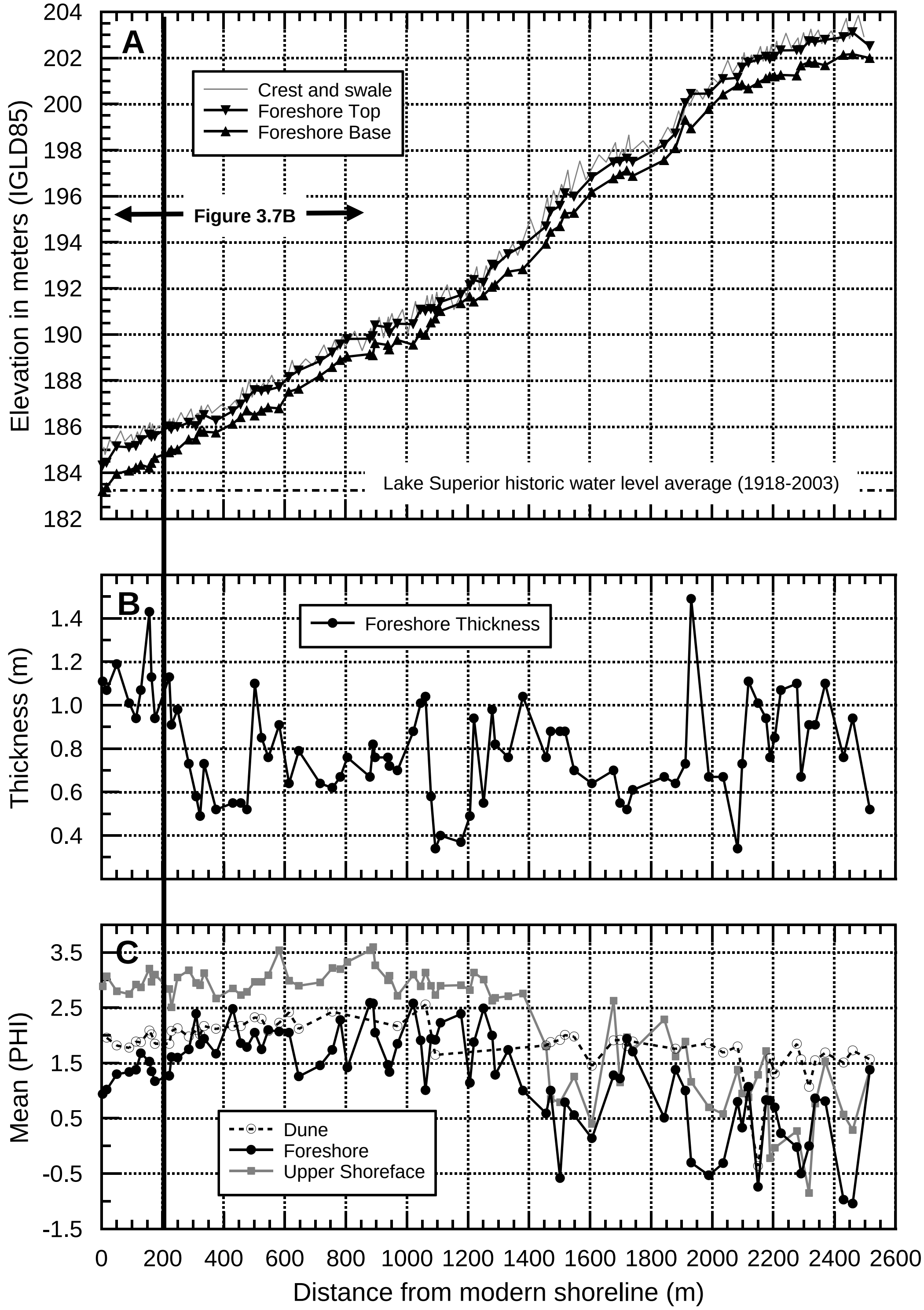

Figure 3.5 Graphs of A) crest, swale and facies contact elevations, B) foreshore thickness, and C) mean grain size per facies from the Batchawana Bay strandplain in Ontario. Gaps in data sets are locations where beach ridges were not cored.

change only occurs in the Tahquamenon Bay embayment where the lakeward set is isolated in a small portion of the eastern part of the embayment. This configuration is unlike the other sites where ridges and swales in the lakeward sets extend near the edge of embayments with no change in orientation.

Sedimentologic characteristics that change between sets at a majority of sites include cross-strandplain variations in facies contact elevations, average foreshore thicknesses, and facies mean grain sizes. The landward set is commonly finer-grained and has a thinner foreshore on average than in the lakeward set (Fig. 3.2 to 3.5). A coarsening from the landward to lakeward sets towards Lake Superior occurs at three of the four sites (Fig. 3.6A, B and 3.7B). Although facies contact elevations do not strictly parallel each other or beach crest and swale elevations, they all follow a similar cross-strandplain trend of falling (Fig. 3.2 to 3.5) in the landward set and rising (Fig. 3.2 to 3.4) or shallowing (Fig. 3.5) in the lakeward set. No trend change is observable between sets for average upper shoreface-facies mean-grain-size. Average grain sizes were too similar between sets to make a clear distinction in these deposits (Table 3.1).

Each study site is unique in its combination of characteristics that change between landward and lakeward sets. In aerial photographs drainage, vegetation, or crest and swale orientation changes are observed between sets. At Grand Traverse Bay, there is a 1.4 km-long slough that parallels the modern Lake Superior shoreline (Fig. 3.8). At Tahquamenon Bay, ridge and swale orientations change from about 15 degrees from the modern shoreline (ESE-WNW) in the landward set to roughly parallel to the modern shoreline (E-W) in the lakeward set (Fig. 3.9). At Au Train Bay the Au Train River changes from an anastomosing channel flowing to the northeast in the landward set to a

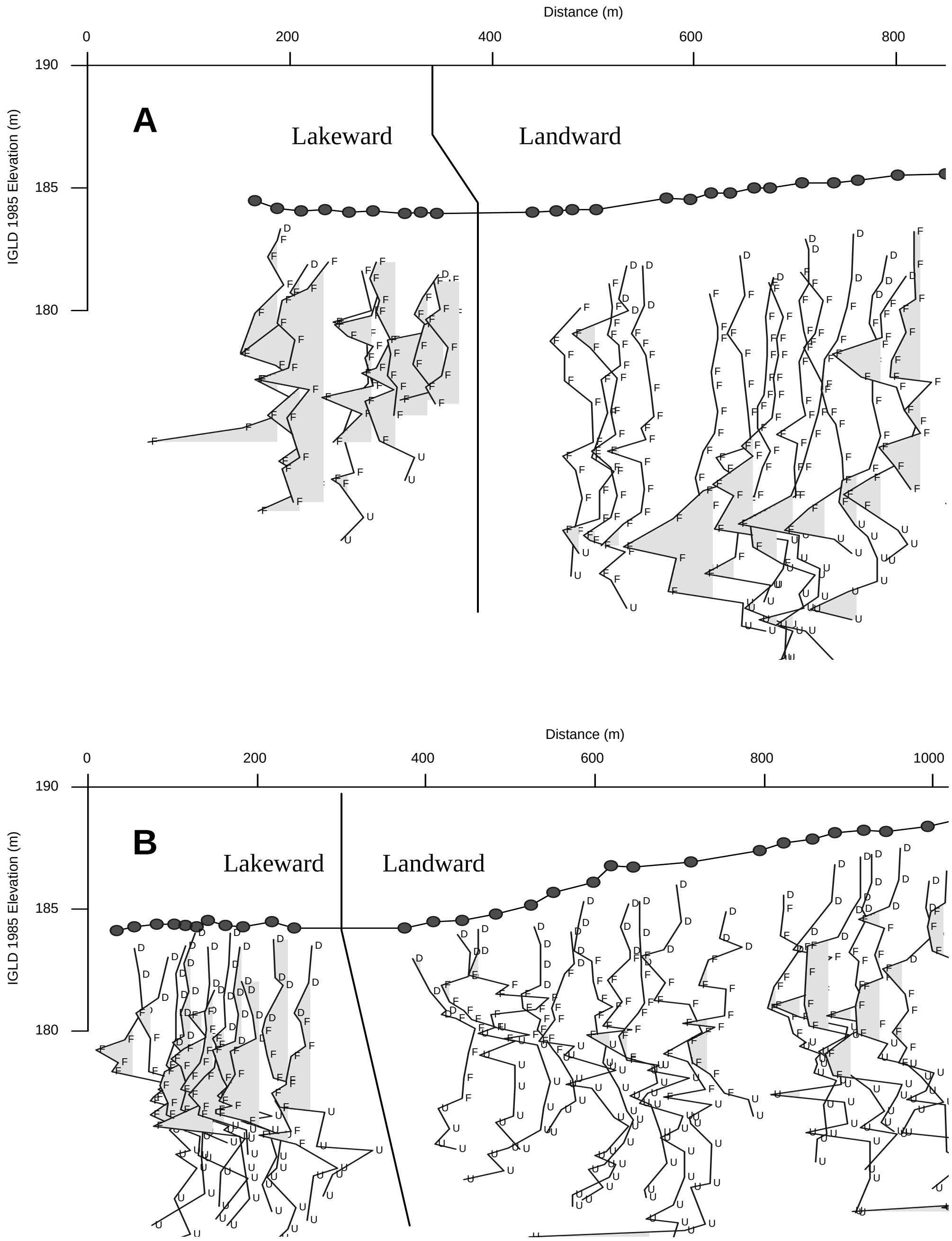

Figure 3.6 Plots of mean grain size per core for the first approximately 30 cores collected adjacent to Lake Superior for A) Grand Traverse Bay strandplain and B) Tahquamenon Bay strandplain. Circles represent the land surface where vibracores were retrieved and are plotted according to distance from the modern shoreline and IGLD85 elevation. Mean grain size for every sample collected in the cores is plotted beneath and labeled in reference to facies (D-dune, F-foreshore, and U-upper shoreface). Shaded areas represent samples that are coarser than the average mean grain size for the foreshore facies for the entire study site. Line represents discontinuity between lakeward and landward sets.

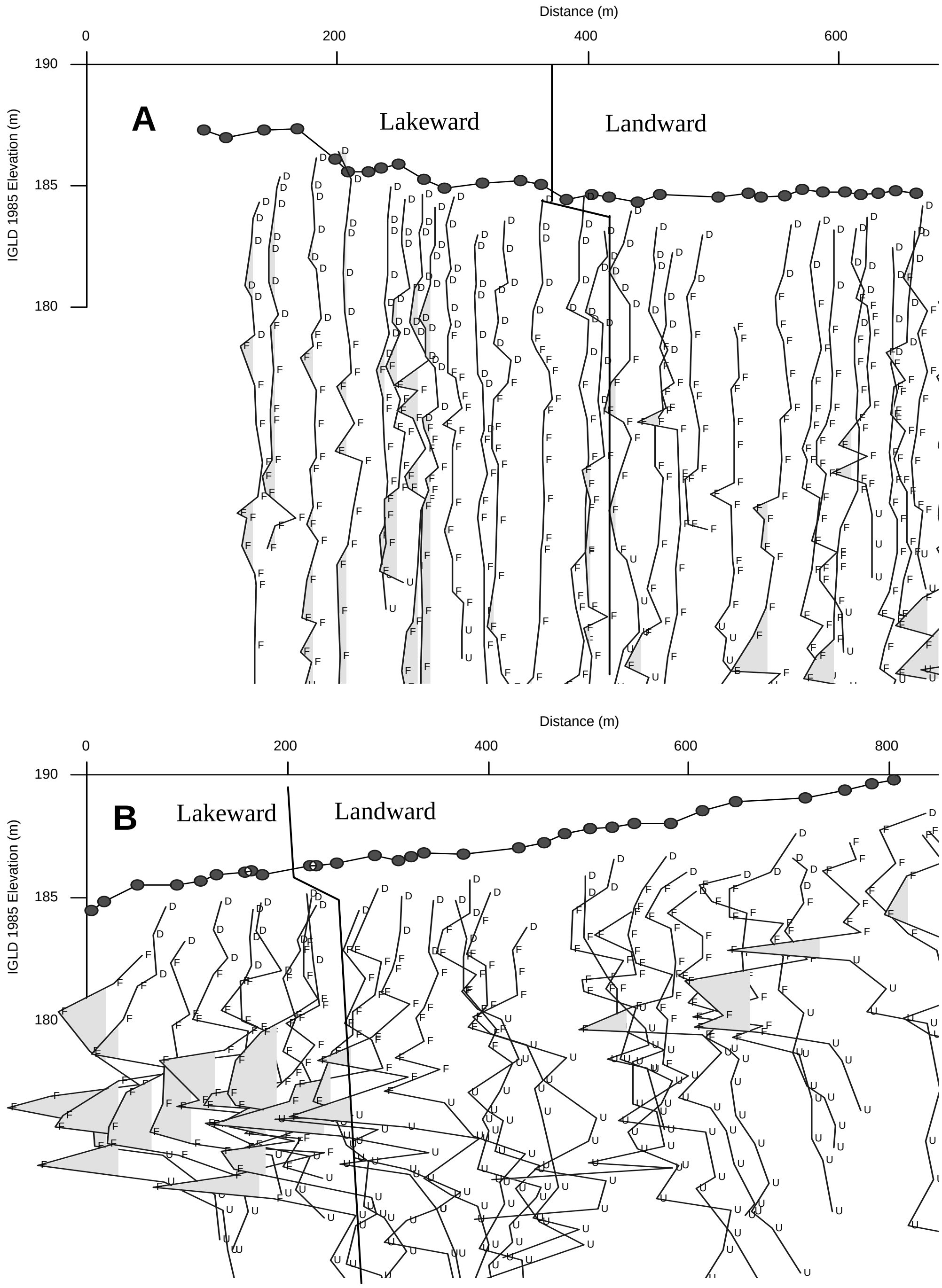

Figure 3.7 Plots of mean grain size per core for the first approximately 30 cores collected adjacent to Lake Superior for A) Au Train Bay strandplain and B) Batchawana Bay strandplain. Circles represent the land surface where vibracores were retrieved and are plotted according to distance from the modern shoreline and IGLD85 elevation. Mean grain size for every sample collected in the cores is plotted beneath and labeled in reference to facies (D-dune, F-foreshore, and U-upper shoreface). Shaded areas represent samples that are coarser than the average mean grain size for the foreshore facies for the entire study site. Line represents discontinuity between lakeward and landward sets.

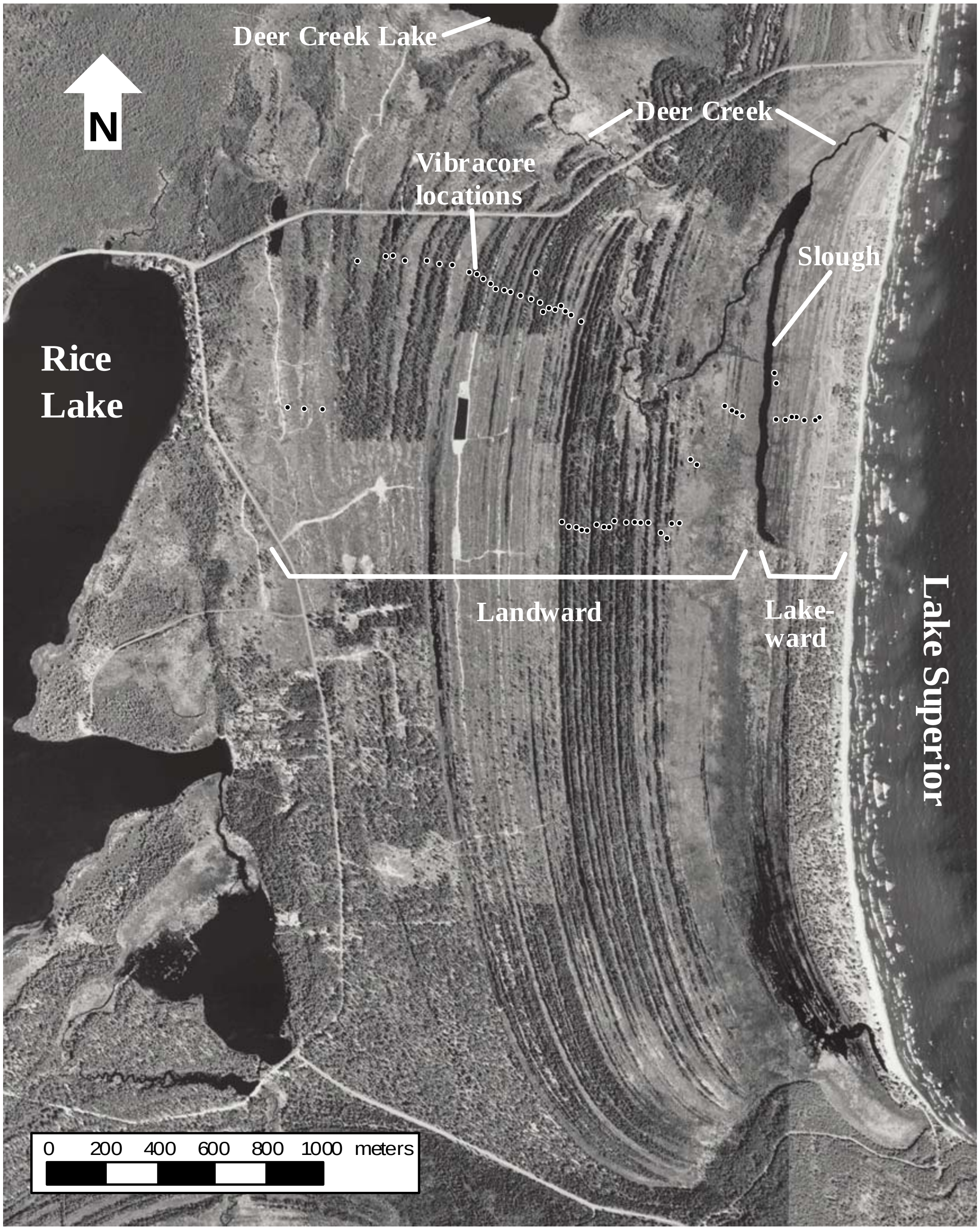

Figure 3.8 Aerial photograph of the Grand Traverse Bay strandplain illustrating the landward and lakeward sets and geomorphic features that occur at the common strandplain discontinuity. See table 3.1 for summary of geomorphic characteristics between sets. Vibracore locations are shown by circles. Aerial photograph from Terraserver courtesy of U.S. Geological Survey.

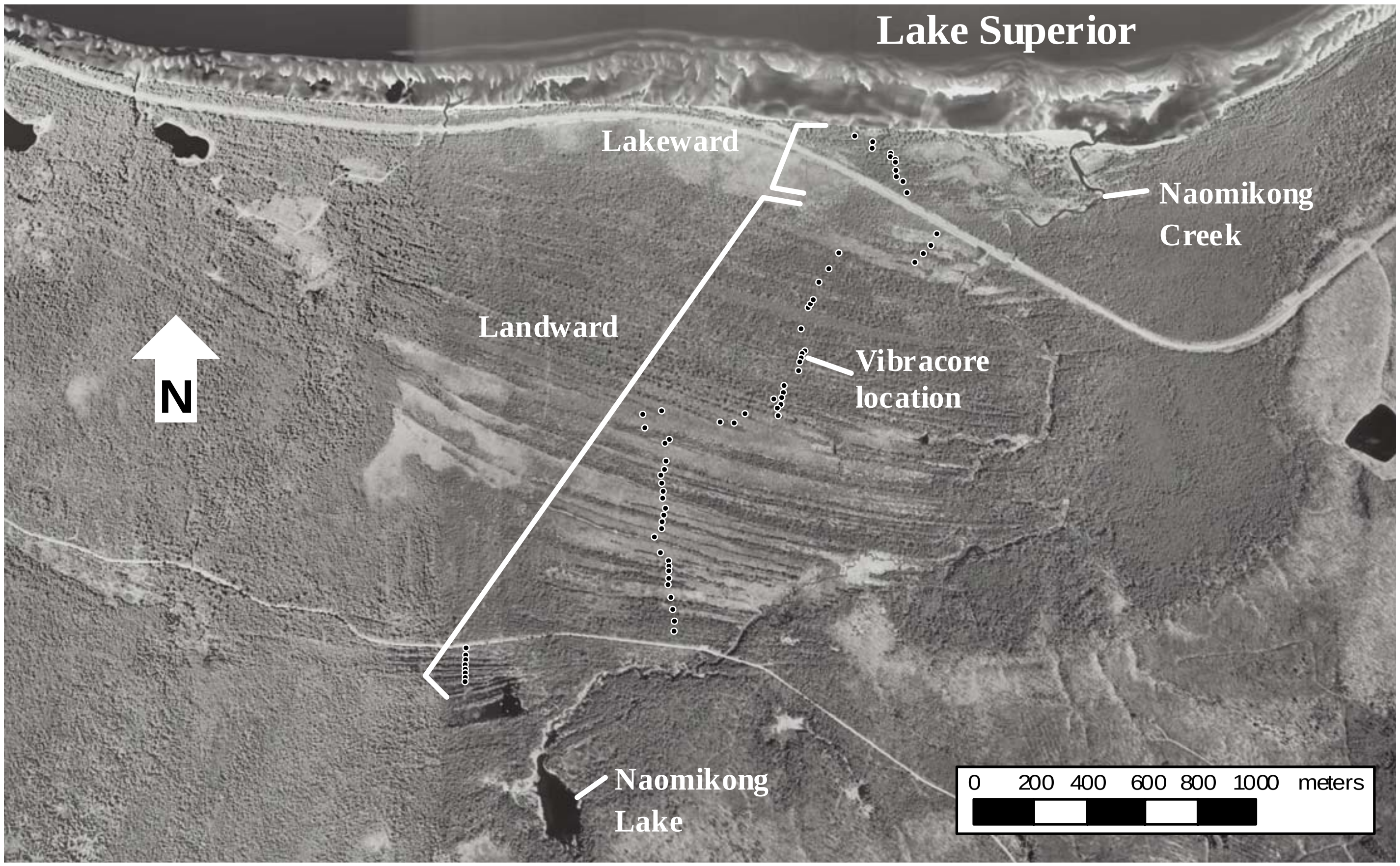

Figure 3.9 Aerial photograph of the Tahquameonon Bay strandplain illustrating the landward and lakeward sets and geomorphic features that occur at the common strandplain discontinuity. See table 3.1 for a complete list of geomorphic characteristics between sets. Vibracore locations are shown by circles. Aerial photograph from Terraserver courtesy of U.S. Geological Survey.

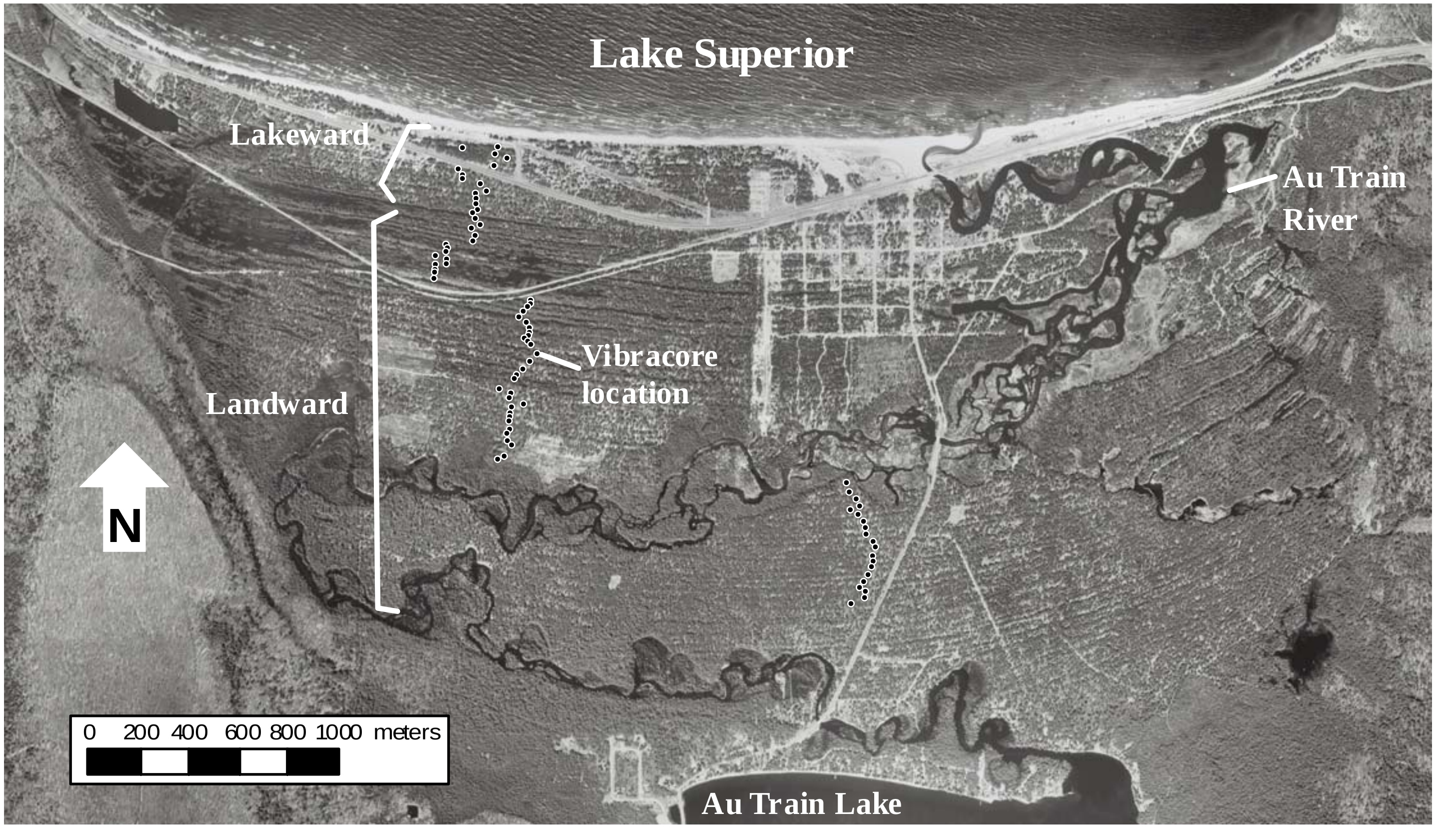

Figure 3.10 Aerial photograph of the Au Train Bay strandplain illustrating the landward and lakeward sets and geomorphic features that occur at the common strandplain discontinuity. See table 3.1 for a complete list of geomorphic characteristics bewteen landward and lakeward sets. Vibracore locations are shown by cirlces. Aerial photograph from Terraserver and courtesy of U.S. Geological Survey.

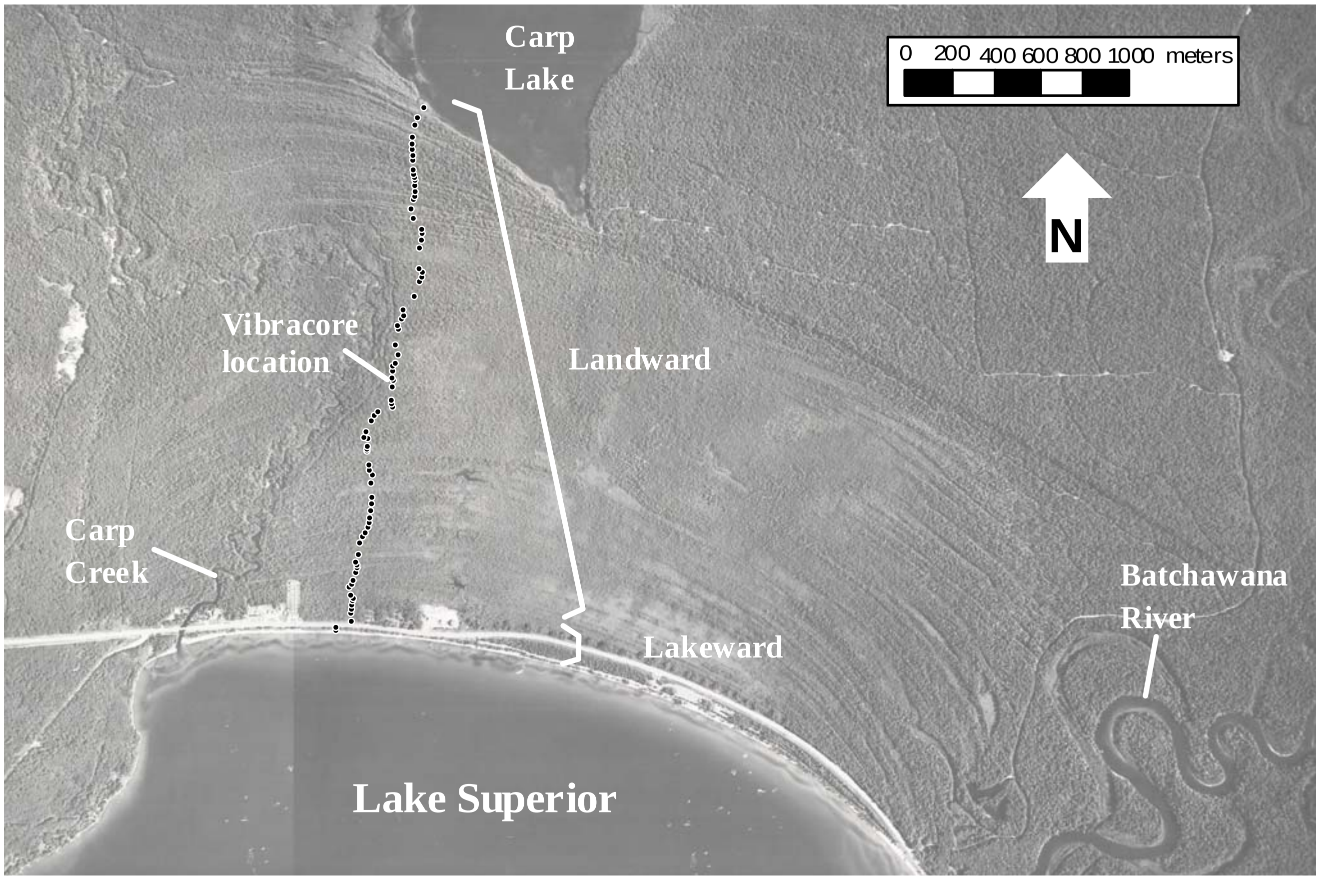

Figure 3.11 Aerial photograph of the Batchawana Bay strandplain illustrating the landward and lakeward sets and geomorphic features that occur at the common strandplain discontinuity. See table 3.1 for a complete list of geomorphic characteristics bewteen sets. Vibracore locations shown as circles. Aerial photograph courtesy of Ontario Ministry of Natural Resources.

meandering channel flowing westward in the lakeward set (Fig. 3.10). A vegetation change due to standing water occurs between sets at Batchawana Bay and is observed in aerial photographs (Fig. 3.11). Drainage pattern variations are noted in topographic maps at all sites but a change in topography between sets is commonly not observed because of relatively high contour intervals (3 to 10 m) that does not always intersect low relief beach ridges. A cross-strandplain elevation change between sets is observed at all sites in plots of topographic surveys of beach ridge crest and swale elevations (i.e. shallowing of slope at Batchawana Bay, Fig. 3.5). Other topographic changes occur across the strandplains but do not accompany subsurface changes. (e.g. about 800 m at Au Train Bay, Fig. 3.4). In subsurface data the characteristics that change abruptly between sets are a foreshore facies coarsening at Grand Traverse Bay (Fig. 3.2) and Tahquamenon Bay (Fig. 3.3), and a foreshore facies contact elevation trend inflection at Tahquamenon Bay (Fig. 3.3) and Au Train Bay (Fig. 3.4). At Batchawana Bay an abrupt foreshore thickening occurs between sets (Fig. 3.5).

Each of the characteristics does not change simultaneously at each of the four sites (Table 3.1). Although general elevation trends across the entire strandplain are similar between geomorphic and sedimentologic characteristics, they do not strictly parallel each other across the strandplains. Topographic elevations differ as much as 2.3 m at Batchawana Bay (Fig. 3.5) and 5.2 m at Au Train Bay (Fig. 3.4) where some of the largest differences occur in the lakeward sets. Identifying and interpreting discontinuities requires careful analysis of what processes deposited the different surface and subsurface sediments.

3.4 Discussion

A common discontinuity in four Lake Superior strandplains was identified using multiple geomorphic and sedimentologic characteristics. The most recognizable feature common to all sites is a cross-strandplain elevation change in either beach-ridge crest elevations or facies contact elevations (Fig. 3.2 to 3.5). A cross-strandplain change in elevations can only be explained by a change in the outlet that controls the elevation of water levels in the Lake Superior basin. Declining elevations indicate that a study site is isostatically rebounding faster than the outlet. Whereas, rising elevations indicate that the outlet is rising more rapidly than the site. If elevations are horizontal, the site is rebounding at the same rate as the outlet. The slope of the cross-strandplain trend provides an estimate of the differential rate between the site and the active outlet. Shallower slopes experience rates of rebound more similar to the outlet and visa versa. A spatial context of the pattern of isostatic rebound is shown in contoured rates of isostatic rebound in the Great Lakes from historical gauge data (Coordinating Committee on Great Lakes Basic Hydraulic and Hydrologic Data 2001). Overall, isostatic rebound rates increase to the northeast across the Great Lakes. Although these rates may have differed during the late Holocene, the contoured pattern is similar between studies of historical and geologic records (c.f. Gilbert 1898; Clark and Persoage 1970; Larsen 1994; Coordinating Committee on Great Lakes Basic Hydraulic and Hydrologic Data 1977, 2001). Comparing trend slopes in the landward set and patterns of contoured rebound in the upper Great Lakes suggests the controlling outlet during the landward lowering was either the Port Huron/Sarnia or Chicago outlets. Both outlets occur south of the study sites and underwent less rebound than the northern sites. Age models created for Grand

Traverse Bay (Johnston et al. 2000) and Tahquamenon Bay (Johnston et al., in press) specify part of the lowering trend at each site was during the Algoma phase. Algoma phase shorelines were identified and correlated between the Superior, Huron, and Michigan basins as early as Leverett and Taylor (1915), suggesting that the Algoma level was common to all three basins and the active outlet was at Port Huron/Sarnia. Our general lowering trends support the idea that the active outlet was at Port Huron/Sarnia when the landward beach ridges and swales were deposited.

The lakeward sets at the four sites show long-term rises and falls that can be used to locate the active outlet during the lakeward set development. The rise at Grand Traverse Bay (Fig. 3.2), Tahquamenon Bay (Fig. 3.3), and Au Train Bay (Fig. 3.4) indicates that the controlling outlet is rebounding faster than the study site, and the lowering at Batchawana Bay (Fig. 3.5) indicates that the study site is rebounding faster than the controlling outlet. Therefore, the active outlet must be between sites of opposite elevation trends following isobases (Fig. 3.2 to 3.5 and Coordinating Committee on Great Lakes Basic Hydraulic and Hydrologic Data 2001). Contoured patterns of historical isostatic rebound between sites establish the Sault outlet as the active outlet when beach-ridges formed in the lakeward set (Fig. 3.1).

A change in beach-ridge crest and swale surface elevations can be used to locate the separation in the strandplain sequences and interpret the cause of the discontinuity, but basal foreshore contact elevations more accurately determine the location within a strandplain sequence and provide a more accurate estimate of past lake-level elevations. Basal foreshore deposits accumulate at lake level and their elevation directly established the elevation of the lake. This relationship was established on the modern shoreline along

Lake Michigan (Fox et al. 1966; Fraser et al. 1991; Thompson 1992) and Lake Superior (Johnston et al. 2000, in press). Because lake level is the only beach-ridge forming factor that is common between strandplains (Thompson and Baedke 1995, Johnston et al., in review), basal foreshore elevations for time equivalent beach ridges at distant strandplains should be the same if isostatic rebound has not warped the basin (cf. Baedke and Thompson 2000). A geomorphic approach of using beach ridge crest elevation to reconstruct past lake level is not as reliable an indicator of lake-level change because this approach measures the elevation of the dune cap on top of the beach ridge core. Sediment within the dune cap is deposited after the core of the beach ridge formed and the thickness of dune sediment does not have a direct relationship to water level elevation because aeolian transport processes are not dependent on lake level. Crest elevations vary by as much as 2.3 m at Batchawana Bay (Fig. 3.5) and 5.2 m (Fig. 3.4) at Au Train Bay from basal foreshore elevations. They also do not parallel each other across the strandplain. For example, at Au Train Bay a beach-ridge crest and swale elevation trend change occurs at around 800 m landward from the modern shoreline (Fig. 3.4). Positioning the separation at this elevation change and using topographic elevations would alter water level and rebound interpretations.

As was demonstrated above, not all geomorphic and sedimentologic characteristics change concurrently because each is related to different processes. Directly underneath the dune facies is the foreshore facies. Unlike the dune sediment, the foreshore sediment is deposited by water. The base of the foreshore approximates past lake-level elevation, whereas the foreshore thickness approximates paleo-wave climate. Howard and Reineck (1981) showed that increased foreshore thickness is related to increased wave energy, or average wave height. Because wave generation is governed by wind duration, speed, and fetch length (Komar 1998), foreshore thickness may reveal past predominant wind characteristics. Although a consistent cross-strandplain trend is not obvious between sites, there is an abrupt change within sites between the landward and lakeward sets (Fig. 3.2 to 3.5). The average foreshore thickness from the landward to lakeward sets slightly increases and the range of variability slightly decreases at several sites (Table 3.1). This may suggest that wave and wind climates have increased after the outlet switch. However, comparison between sets is difficult without age control because a part of the record may be absent (either created and destroyed or not created at all) and not time equivalent. Identification of a cross strandplain discontinuity and preliminary age results, additional to Johnston et al. (in press) suggest that approximately 1,000 years of missing time exists within the strandplain sequence at Tahquamenon Bay and Batchawana Bay. In other words, a foreshore thickening at Batchawana Bay (Fig. 3.5) and thinning at Tahquamenon Bay (Fig. 3.3) may indicate a reorientation in the predominant wind from a more southerly to northerly trend but only when comparing the Algoma (landward) to the Sub-Sault (lakeward) phase. Preliminary age results including those from Johnston et al. (2000) suggest the record between sets at Grand Traverse Bay and Au Train Bay are more complete and contain records between the Algoma and Sault phases. With this in mind, an increase in foreshore thickness at Grand Traverse Bay from the landward to lakeward sets (Fig. 3.2) may correspond to a reorientation of predominant winds from a more southerly to northerly trend. If lakeward sets are time equivalent, decreasing foreshore thickness within the lakeward set at most sites (Table

3.1) suggests a decreasing wave and wind climate while the Sault outlet was controlling water levels in the Lake Superior basin.

An abrupt change in mean grain size per facies occurs between the landward and lakeward sets at most sites (Table 3.1, Fig. 3.2 to 3.5). The largest change is in the foreshore facies where grain size coarsens. This coarsening may be related to a sedimentary source change or littoral transportation direction change or both after Lake Superior became its own lake. Major reorientations of several drainages in embayments occur between sets that may correspond to changes in littoral transport. An obvious change in the direction of the Au Train River (Fig. 3.10) suggests the littoral transport direction changed from more eastward to westward after the lakes separated. This potential sediment source change is not reflected in the mean grain size because the Au Train Bay embayment has a uniform sediment source from the Munising Formation along both margins of the embayment (Hamblin 1958). No clear correspondence appears to occur between foreshore thickness and mean grain size, but a few instances suggest mean grain-size fines during decreased foreshore thickness and visa versa (Fig. 3.2 to 3.5). A decreasing wave climate would be expected to transport a finer size fraction because of reduced available energy for transportation.

Interpretation and differentiation of the lakeward set is much more difficult than that of the landward set because the lakeward set has less than 15 ridges and the landward set has commonly more than 50 ridges. This is especially critical for Batchawana Bay that rebounded more rapidly than the outlet during both sets. Here, a change in the trend of basal foreshore elevations is not always apparent as with an inflection in elevations. Basal foreshore elevation trends from the landward to lakeward sets seem similar (Fig.

3.5). However, a trend through the short record in the lakeward set may not be representative of a long-term trend for isostatic rebound. The trend in the lakeward set may portray part of a decreasing water-level trend as is depicted in figure 4C of Baedke and Thompson (2000). Other characteristics such as foreshore thickness are needed to help recognize the common discontinuity for sites north of the zero isobase at the active outlet. Isostatic rebound trends are more easily identified from long records at this stage but are more accurately determined after establishing age models and creating relative lake-level curves.

Strandplains of beach ridges in the Great Lakes are often been interpreted as continuous prograding sequences revealing long-term variations in past lake level and isostatic rebound. Breaks in these records can change how the sequence is interpreted if the break is not identified and handled. Larsen (1994) recognized a topographic change in beach ridge crests in a strandplain on the Whitefish Point promontory along the southern shore of Lake Superior. Declining crest elevations followed by horizontal elevations toward Lake Superior led Larsen (1994) to relate this change to the separation of the lakes. He accounted for the change by analyzing his record in two separate parts but only in isostatic rebound calculations and not in his age model. Comparison of ages on either side of the elevation change at Whitefish Point (Larsen 1994; fig. 15) suggests there are two populations of ages, and there is a time gap where ridges are missing. This corresponds to a suggested time gap in the Tahquamenon Bay strandplain by Johnston et al. (in press) and corresponds to the common discontinuity identified in this paper. Equating this discontinuity to the Whitefish Point strandplain and using age model results for Grand Traverse Bay (Johnston et al. 2000) would change Larsen's (1994) interpretation for when the lakes separated from about 2,200 years ago after the Algoma phase to about 1,200 years ago after the Sault phase. This timing better equates Farrand's (1960) Sub-Sault shorelines with the lakeward set defined in this paper. However, the Sault phase is defined as when the lakes separated (Farrand 1960). To keep this definition and account for the later timing of separation the Sub-Sault phase should be discarded and replaced with the Sault phase. This would move the Sault phase to include the present and place the separation of the lakes between a proposed "post-Algoma" phase and redefined Sault phase of Lake Superior. Comparison of multiple characteristics and collection of basal foreshore elevations from every beach ridge at Whitefish Point would more accurately identify the location of the discontinuity in the strandplain and would produce best results for water level and isostatic rebound.

Changing the time period for when the lakes separated to after the post-Algoma phase of Lake Superior (Farrand 1960) warrants further investigation of similar irregularities of other strandplains in other lake basins. Thompson and Baedke (1997) addressed missing ridges within the Manistique strandplain on the northern shore of Lake Michigan in their age and rebound models. These missing ridges occur around the proposed later time period for the separation of the lakes after the post-Algoma phase. This suggests that the post-Algoma phase defined above for Lake Superior and the unnamed phase (Thompson and Baedke 1997) of Lake Michigan could be correlated and the lakes separated after this phase. An inflection was identified in cross-strandplain topographic (Larsen 1994) and basal foreshore (Thompson and Baedke 1997) elevations in the Toleston Beach strandplain in southern Lake Michigan. The rising trend in basal foreshore elevations and calculated rates of isostatic rebound after the unnamed phase compare to historical patterns and rates of isostatic rebound relative to the Port Huron/Sarnia outlet (Baedke and Thompson 2000). Prior to this time period declining elevations have been explained by erosion at the Port Huron/Sarnia outlet (Larsen 1994) or related to a peripheral bulge near southern Lake Michigan (Tushingham 1992). Additional data sets from Lake Michigan and Lake Huron strandplains need to be investigated to evaluate the impact of the separation of Lake Superior from Lake Michigan and Lake Huron on downstream strandplains. Also, existing strandplain data records need to be revisited because of the change in the timing of the separation of the lakes to after the post-Algoma phase. Geomorphic and sedimentologic properties also need to be examined around this time period, especially at sites north of the zero isobase because changes in basal foreshore elevation trends may not be apparent. This may alter outlet, isostatic rebound, or water level results and interpretations.

3.5 Conclusion

More accurate identification of common strandplain discontinuities and using elevations that are directly related to lake level leads to better estimations of active outlets, past long-term lake-level, and isostatic rebound. Multiple characteristics can be used to identify or refine their location within strandplain sequences. They can also be used to interpret conditions such as wind and wave climate and littoral transport leading up to the discontinuity and afterwards. Comparison of data from many sites helps identify common discontinuities in the strandplain sequences that can be used to determine the cause.

Few studies have been conducted south of the zero isobase relative to the active outlet because shorelines normally coalesce, erode, or are submerged underwater during rising long-term water levels. Embayments are advantageous locations for study south of the zero isobase because of ample sediment supply and accommodation space that help preserve relict shorelines. Continuous records in the range of many decades to millennia can be created and preserved in embayments. At these sites it is sometimes easier to recognize past outlet changes because an inflection in cross-strandplain beach-ridge crest and basal foreshore elevations is created. Other characteristics associated with the elevation inflections can be used to help interpret sites north of the zero isobase where elevation changes are less apparent. Detailed shoreline research should not only focus on sites north of the zero isobase where isostatic rates are advantageous for preservation but also in embayments south of the zero isobase where sediment accumulation and accommodation space are advantageous for preservation.

Current outlet conditions and isostatic rebound patterns across the Lake Superior basin suggest the Sault outlet will potentially rebound faster than three of the study sites (Tahquamenon Bay, Grand Traverse Bay, and Au Train Bay) into the impending future. This will cause long term lake-level to potentially rise at each of these sites and cause problems for erosion if ample sediment is not supplied to the shoreline for construction or buffering. An increase in basal foreshore elevations in the lakeward sets and presence of erosional scarps on the modern beach at a few of the study sites (i.e. Tahquamenon Bay) are related to this long-term trend from the outlet change to the present and is a constant reminder for the need to reevaluate past lake level and isostatic rebound trends and prepare for the future.

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APPENDICIES

APPENDIX 1

Data per core collected from the four study sites, Batchawana Bay, Tahquamenon Bay, Grand Traverse Bay, and Au Train Bay. Data includes beach ridge number, distance from the modern shoreline, beach ridge crest elevation, swale elevation, facies contact elevations, foreshore facies thickness, and mean grain size for each facies (dune, foreshore and upper shoreface).

Batchawana Bay

Ridge #	Distance Lanward (m)	Crest Elev. IGLD85 (m)	Swale Elev. IGLD85 (m)	Foreshore Top IGLD85 (m)	Foreshore Base IGLD85 (m)	Foreshore Thickness (m)	Mean Dune (phi)	Mean Fore-shore (phi)	Mean Upper Shoreface (phi)
1	4	185.11	184.80	184.32	183.21	1.11		0.94	2.89
2	17	185.45	185.19	184.44	183.37	1.07	1.96	1.02	3.07
3	50	185.82	185.37	185.15	183.96	1.19	1.82	1.30	2.80
4	91	185.67	185.33	185.11	184.10	1.01	1.78	1.34	2.75
5	113	185.78	185.56	185.17	184.23	0.94	1.89	1.38	2.92
6	129	186.01	185.82	185.42	184.35	1.07	1.88	1.68	2.87
7	157	186.17	185.75	185.66	184.23	1.43	2.09	1.53	3.21
8	164	186.12	185.78	185.56	184.43	1.13	2.02	1.35	2.97
9	175	186.05	185.84	185.60	184.65	0.94	1.85	1.17	3.10
10	222	186.34	185.94	186.01	184.89	1.13	1.85	1.27	2.84
11	229	186.38	186.10	185.90	184.99	0.91	2.08	1.61	2.51
12	249	186.61	186.28	185.99	185.01	0.98	2.13	1.60	3.05
13	286	186.77	186.29	186.18	185.45	0.73	1.98	1.75	3.18
14	310	186.58	186.52	186.03	185.45	0.58	2.10	2.39	2.95
15	323	186.90	186.52	186.30	185.81	0.49	2.02	1.84	2.91
16	336	186.95	186.60	186.51	185.78	0.73	2.17	1.94	3.13
17	375	186.95	186.75	186.27	185.75	0.52	2.12	1.67	2.67
18	430	187.14	186.88	186.67	186.13	0.55	2.17	2.48	2.85
19	456	187.70	187.24	186.97	186.42	0.55	2.17	1.86	2.73
20	476	188.05	187.29	187.23	186.71	0.52		1.79	2.79
21	502	187.87	187.62	187.59	186.49	1.10	2.33	2.05	2.97
22	524	187.90	187.70	187.55	186.70	0.85	2.30	1.75	2.97
23	546	188.24	187.84	187.60	186.84	0.76	2.08	2.10	3.09
24	582	188.06	187.91	187.72	186.80	0.91	2.23	2.07	3.54
25	614	188.88	188.41	188.16	187.52	0.64	2.42	2.05	2.99
26	646	188.94	188.65	188.44	187.65	0.79	2.12	1.26	2.90
27	716	189.55	189.05	188.85	188.21	0.64		1.46	2.96
28	756	189.77	189.32	189.22	188.59	0.62	2.43	1.74	3.22
29	782	189.95	189.45	189.57	188.90	0.67		2.28	3.20
30	805	190.15	189.30	189.81	189.04	0.76		1.42	3.33
31	879	190.27	189.97	189.82	189.15	0.67		2.59	3.54
32	889	190.44	190.17	189.93	189.10	0.82		2.58	3.60
33	896	190.75	189.86	190.40	189.63	0.76		2.05	3.27
34	938	190.76	190.50	190.31	189.55	0.76		1.47	3.00
35	943	190.91	190.43	190.06	189.35	0.72		1.34	3.08
36	969	191.09	190.05	190.47	189.77	0.70	2.17	1.85	2.72
37	1021	191.43	190.74	190.45	189.56	0.88		2.58	3.10
38	1045	191.27	190.97	191.08	190.07	1.01		1.91	2.89
39	1061	191.68	191.11	191.02	189.99	1.04	2.56	1.01	3.14
40	1079	191.73	191.18	191.11	190.53	0.58		1.94	2.90
41	1093	191.84	191.37	191.02	190.69	0.34	1.65	1.92	2.73
42	1110	192.16	191.09	191.41	191.02	0.40		2.23	2.90
43	1177	192.01	191.61	191.72	191.36	0.37		2.39	2.91
44	1206	192.61	192.25	192.14	191.65	0.49		1.14	2.82

Ridge #	Distance Lanward (m)	Crest Elev. IGLD85 (m)	Swale Elev. IGLD85 (m)	Foreshore Top IGLD85 (m)	Foreshore Base IGLD85 (m)	Foreshore Thickness (m)	Mean Dune (phi)	Mean Fore-shore (phi)	Mean Upper Shoreface (phi)
45	1219	192.94	191.88	192.37	191.43	0.94		1.88	3.14
46	1251	192.98	192.57	192.25	191.70	0.55		2.49	3.01
47	1279	193.18	192.81	193.04	192.07	0.98		2.00	2.63
48	1289	193.63	193.24	192.98	192.16	0.82		1.29	2.68
49	1332	193.95	193.45	193.49	192.73	0.76		1.74	2.71
50	1380	195.02	194.09	193.86	192.83	1.04		1.00	2.76
51	1456	195.97	195.35	194.70	193.94	0.76	1.82	0.59	1.80
52	1471	196.26	195.79	195.34	194.46	0.88	1.88	1.00	0.85
53	1502	196.52	196.32	195.59	194.71	0.88	1.92	-0.58	0.79
54	1518	197.14	196.13	196.14	195.26	0.88	2.01	0.79	
55	1548	197.53	196.71	195.99	195.29	0.70	1.98	0.56	1.26
56	1606	197.80	197.49	196.83	196.19	0.64	1.45	0.14	0.40
57	1677	198.34	197.55	197.48	196.78	0.70	1.91	1.28	2.63
58	1698	198.05	197.79	197.50	196.96	0.55	1.92	1.22	1.15
59	1721	198.65	197.93	197.64	197.12	0.52	1.81	1.94	1.97
60	1740	198.41	197.83	197.50	196.89	0.61	1.89	1.71	1.73
61	1843	198.99	198.74	198.24	197.57	0.67		0.51	2.29
62	1880	199.71	199.24	198.73	198.09	0.64	1.76	1.38	1.62
63	1912	200.27	199.92	200.05	199.32	0.73		1.00	1.89
64	1931	200.61	200.22	200.45	198.95	1.49		-0.30	1.16
65	1989	201.11	200.91	200.46	199.79	0.67	1.86	-0.53	0.70
66	2036	201.91	201.36	201.09	200.42	0.67	1.69	-0.31	0.58
67	2083	201.82	201.53	201.14	200.81	0.34	1.80	0.80	1.38
68	2098	202.23	201.69	201.59	200.85	0.73		0.33	0.95
69	2119	202.13	201.81	201.80	200.68	1.11		1.07	0.88
70	2150	202.50	201.86	201.93	200.92	1.01	-0.36	-0.74	1.29
71	2176	202.50	202.01	202.06	201.12	0.94		0.83	1.72
72	2189	202.58	201.99	201.96	201.20	0.76	1.59	0.82	
73	2205	202.73	202.24	202.06	201.20	0.85	1.31	0.70	
74	2225	203.08	202.43	202.33	201.27	1.07		0.23	
75	2277	202.87	202.46	202.34	201.24	1.10	1.85	-0.02	
76	2291	203.10	202.60	202.35	201.68	0.67	1.57	-0.50	
77	2317	203.26	202.85	202.73	201.82	0.91	1.07	0.00	
78	2337	203.22	202.63	202.70	201.79	0.91	1.55	0.86	0.77
79	2370	203.17	202.68	202.79	201.69	1.10	1.69	0.81	1.53
80	2430	203.72	202.85	202.91	202.14	0.76	1.51	-0.97	0.57
81	2460	203.85	202.93	203.12	202.17	0.94	1.73	-1.04	0.29
82	2516	203.18	202.46	202.52	202.00	0.52	1.57	1.38	1.39

Tahquamenon Bay

Ridge #	Distance Lanward (m)	Crest Elev. IGLD85 (m)	Swale Elev. IGLD85 (m)	Foreshore Top IGLD85 (m)	Foreshore Base IGLD85 (m)	Foreshore Thickness (m)	Mean Dune (phi)	Mean Fore-shore (phi)	Mean Upper Shoreface (phi)
1	34			183.55	183.15	0.40	1.38	0.93	2.16
2	54			183.67	183.03	0.64	1.50	1.31	1.58
3	82			183.46	182.96	0.50	1.37	1.19	2.02
4	102			183.40	183.00	0.40	1.18	0.83	1.71
5	116			183.71	182.98	0.73	1.39	1.00	1.72
6	129			183.72	182.87	0.85	1.30	0.98	1.76
7	143			183.77	183.13	0.64	1.33	0.97	1.68
8	163			183.49	183.03	0.46	1.20	0.92	1.70
9	184			183.44	182.86	0.58	1.10	0.73	1.68
10	218			183.78	182.90	0.88	1.04	1.11	1.82
11	245			183.62	182.98	0.64	1.19	0.91	1.69
12	265								
13	299								
14	374			183.61	183.03	0.58	1.48	2.05	1.69
15	408			184.06	183.76	0.30	1.56	1.39	2.21
16	442			184.15	183.70	0.46	1.23	1.89	1.95
17	483			184.25	183.81	0.44	1.76	1.72	2.17
18	524			184.70	184.22	0.49	1.72	1.38	2.09
19	551			185.24	184.64	0.59	1.45	1.78	1.96
20	598			185.57	184.85	0.72	1.58	1.38	2.01
21	619			186.07	185.43	0.64	1.50	1.40	1.98
22	646			186.10	185.13	0.98	1.70	1.72	1.81
23	714			186.28	185.32	0.96	1.66	1.28	2.03
24	796			187.03	186.13	0.90	1.42	1.42	2.00
25	823			186.96	186.47	0.49	1.58	1.40	1.97
26	857			187.14	186.52	0.62	1.70	0.90	1.89
27	884			187.58	186.42	1.16	1.59	1.28	2.02
28	918			187.65	186.79	0.85	1.56	1.07	1.87
29	945			187.28	186.56	0.72	1.65	1.22	1.88
30	966								
31	993			187.84	186.97	0.87	1.32	1.31	1.99
32	1034			188.11	187.29	0.82	1.56	1.42	1.63
33	1054			188.47	187.28	1.19	1.60	1.36	1.82
34	1081			188.59	187.67	0.91	1.85	1.49	2.06
35	1095			188.40	187.64	0.76	1.68	1.72	1.59
36	1115			188.35	187.65	0.70	1.62	1.25	1.65
37	1136			188.49	187.56	0.93	1.41	0.99	1.74
38	1190			188.62	187.73	0.88	1.58	1.40	1.79
39	1217			188.50	187.40	1.10	1.80	1.54	1.63
40	1244			189.09	187.82	1.26	1.99	1.23	1.83
41	1258			189.06	188.09	0.98	1.41	1.22	1.78
42	1292			188.82	187.34	1.48	1.39	1.07	1.56
43	1326			189.52	188.20	1.33		1.19	1.77
44	1353			189.71	188.99	0.72	1.58	1.42	1.75

Ridge #	Distance Lanward (m)	Crest Elev. IGLD85 (m)	Swale Elev. IGLD85 (m)	Foreshore Top IGLD85 (m)	Foreshore Base IGLD85 (m)	Foreshore Thickness (m)	Mean Dune (phi)	Mean Fore-shore (phi)	Mean Upper Shoreface (phi)
45	1380			189.17	188.09	1.08	1.51	1.27	1.69
46	1428			189.93	188.79	1.14	1.89	1.41	2.00
47	1455			189.54	188.47	1.07	1.60	1.21	2.00
48	1469			189.80	189.16	0.64	1.37	1.03	1.86
49	1489			190.34	189.24	1.10	1.36	1.12	1.60
50	1516			189.95	188.77	1.19	1.73	1.22	1.93
51	1544			190.23	189.47	0.76	1.65	1.27	1.87
52	1578			190.44	189.55	0.88	1.76	1.24	1.88
53	1612			190.66	189.41	1.25	1.61	1.28	1.60
54	1625			190.69	189.97	0.72	1.62	0.84	1.64
55	1659			190.61	190.15	0.46	1.30	1.41	1.91
56	1686			190.61	189.74	0.87	1.57	1.46	1.90
57	1727			191.14	189.89	1.25	1.88	1.60	1.76
58	1748			190.62	189.89	0.73	1.53	1.70	1.99
59	1775			190.81	190.17	0.64	1.65	1.64	1.38
60	1782			190.88	190.25	0.62	1.76	1.53	2.01
61	1809			191.28	190.18	1.10	1.84	1.50	1.51
62	1829			191.06	190.24	0.82		1.20	1.98
63	1870			192.09	191.27	0.82	1.77	2.08	
64	1924			192.97	191.93	1.04	1.66	1.43	
65	1958			193.07	192.37	0.70	1.53	1.21	
66	1992			193.81	192.67	1.14	1.56	0.83	
67	2033								
68	2060								
69	2094								
70	2115								
71	2142			195.91	195.06	0.85	1.55	1.67	1.59
72	2162			195.67	195.12	0.55	1.55	1.67	1.59
73	2176			195.75	194.60	1.16	1.38	1.17	1.14
74	2190			195.58	194.80	0.78	1.46	1.05	1.61
75	2210			195.75	194.79	0.96	1.32	1.02	1.12
76	2217			195.25	194.26	0.99	1.26	0.88	-0.13
77	2230			195.24	194.63	0.61	1.17	1.19	1.02
78	2251			195.72	194.90	0.82	1.30	1.03	1.34
79	2278								
80	2326								

Grand Traverse Bay

Ridge #	Distance Lanward (m)	Crest Elev. IGLD85 (m)	Swale Elev. IGLD85 (m)	Foreshore Top IGLD85 (m)	Foreshore Base IGLD85 (m)	Foreshore Thickness (m)	Mean Dune (phi)	Mean Fore-shore (phi)	Mean Upper Shoreface (phi)
1	31								
2	55								
3	79								
4	110								
5	165			184.31	183.12	1.19	1.43	1.05	
6	188			183.85	182.45	1.40	1.54	1.15	
7	212			183.88	182.45	1.43		0.85	
8	236			183.79	182.48	1.31		1.48	1.47
9	259			183.71	182.39	1.33		1.13	
10	283			184.04	182.75	1.30		1.13	1.55
11	290								
12	314			183.62	182.29	1.33	1.59	0.97	
13	330			183.69	182.37	1.33		1.11	
14	345			183.74	182.41	1.33		0.98	
15	393								
16	416								
17	440			183.69	182.20	1.49		1.72	1.66
18	463			183.76	182.23	1.52		1.83	1.38
19	479			183.54	182.05	1.49	1.91	1.49	1.75
20	502			183.58	182.11	1.46	1.82	1.69	
21	542								
22	573			184.10	182.61	1.49		1.73	1.73
23	597			184.11	182.35	1.77	1.86	1.39	2.03
24	616			184.32	182.69	1.63	2.07	1.77	1.97
25	636			184.38	182.95	1.43		1.46	1.81
26	659			184.58	182.91	1.68	1.91	1.60	1.66
27	675			184.65	183.16	1.49		1.56	1.65
28	707			184.68	183.30	1.37	1.88	1.44	1.71
29	738			184.57	183.36	1.20	1.81	1.55	1.37
30	761			184.68	183.40	1.28	1.82	1.38	1.67
31	801			185.21	183.76	1.45		1.13	
32	816								
33	848			185.10	183.56	1.54	1.71	1.62	
34	864			185.02	183.65	1.37	1.94	1.39	1.70
35	887			185.20	183.91	1.29	1.69	1.32	
36	911			185.25	184.18	1.07	1.71	1.16	2.00
37	919								
38	938			185.03	184.09	0.94	1.87	1.74	1.87
39	958			185.20	184.19	1.01	1.68	1.70	1.52
40	973			185.51	184.50	1.01	1.79	1.63	1.58
41	1005			185.42	184.39	1.04	1.81	1.34	1.31
42	1021			185.37	184.06	1.31	1.76	1.54	1.92
43	1052			185.57	184.71	0.85	1.64	0.91	1.80
44	1083			185.65	184.56	1.10	1.70	1.42	1.88

Ridge #	Distance Lanward (m)	Crest Elev. IGLD85 (m)	Swale Elev. IGLD85 (m)	Foreshore Top IGLD85 (m)	Foreshore Base IGLD85 (m)	Foreshore Thickness (m)	Mean Dune (phi)	Mean Fore-shore (phi)	Mean Upper Shoreface (phi)
45	1107			185.81	184.71	1.10	1.79	1.35	1.80
46	1127			185.98	184.61	1.37	1.82	1.48	1.98
47	1146			185.74	184.52	1.22	1.79	1.40	2.14
48	1166			185.68	184.71	0.98	1.68	1.24	1.88
49	1189			186.49	185.07	1.42	1.67	1.19	
50	1217			186.59	185.30	1.29	1.67	1.32	
51	1233			186.37	185.18	1.19	1.64	1.26	2.15
52	1256			186.82	185.53	1.29	1.59	1.38	
53	1284			186.38	185.38	1.01	1.91	1.38	1.30
54	1327			186.46	185.54	0.91	1.77	1.34	1.81
55	1346			186.73	185.57	1.16	1.83	1.38	1.65
56	1366			186.55	185.24	1.31		1.81	1.98
57	1390			186.57	185.20	1.37		1.75	1.83
58	1413			187.11	185.77	1.34	1.85	1.46	1.95
59	1444			187.05	185.66	1.39	1.87	1.57	2.08
60	1484			186.64	185.48	1.16	1.77	1.62	2.24
61	1539			187.18	185.93	1.25	1.85	1.29	1.96
62	1586			187.21	186.05	1.16	1.76	1.51	1.90
63	1625			186.92	185.73	1.19	1.64	1.33	2.01
64	1680			187.15	185.85	1.30	1.72	1.59	1.89
65	1723			187.46	186.32	1.14	1.43	1.59	
66	1766			187.79	186.48	1.31	1.67	1.50	1.84
67	1853			187.91	186.61	1.30		1.12	2.06
68	1931			187.84	186.56	1.28		1.16	1.47
69	2018			187.87	186.16	1.71		1.19	1.62
70	2175			188.01	186.67	1.34		1.14	1.74

Au Train Bay

Ridge #	Distance Lanward (m)	Crest Elev. IGLD85 (m)	Swale Elev. IGLD85 (m)	Foreshore Top IGLD85 (m)	Foreshore Base IGLD85 (m)	Foreshore Thickness (m)	Mean Dune (phi)	Mean Fore-shore (phi)	Mean Upper Shoreface (phi)
1	94	188.42	186.98	186.12	184.14	1.98	1.85	1.83	
2	111	187.45	187.45	185.99	184.72	1.26	1.85	1.81	
3	141	188.03	187.28	186.21	184.24	1.97	1.89	1.81	1.72
4	168	187.79	186.09	186.19	184.27	1.92	1.83	1.84	
5	198	186.54	185.55	184.95	183.84	1.11	1.91	1.83	1.86
6	208	186.74	185.45	184.65	183.34	1.31	1.92	1.77	1.97
7	225	186.26	185.71	184.39	182.32	2.07	1.89	1.71	1.90
8	235	186.31	185.81	184.32	182.27	2.04	1.85	1.75	1.62
9	248	186.23	185.27	184.99	183.40	1.58	1.89	1.95	2.12
10	268	186.39	184.90	183.87	181.22	2.65	1.95	1.90	
11	285	186.32	185.06	183.88	182.14	1.74	2.01	1.93	
12	315	185.92	185.15	184.27	181.66	2.61	2.00	1.97	
13	346	185.62	185.04	184.12	182.03	2.09	2.06	1.98	
14	362	185.69	184.35	183.84	180.76	3.08	2.02	1.81	1.86
15	383	185.40	184.63	183.55	182.46	1.10	1.99	1.90	1.95
16	403	185.65	184.52	183.74	182.04	1.71	2.13	2.08	1.83
17	416	185.24	184.59	183.75	181.44	2.32	2.02	1.93	
18	440	185.63	184.53	183.83	182.54	1.30	2.07	1.98	
19	456	185.60	184.60	184.09	182.40	1.69		2.26	2.11
20	473								
21	503	185.54	184.60	184.03	181.94	2.09	2.20	1.92	1.66
22	527	185.08	184.51	184.17	182.07	2.10	2.17	2.11	2.41
23	537	185.25	184.61	183.83	182.28	1.55	2.27	2.11	2.43
24	557	185.70	184.75	183.84	181.92	1.92	2.27	2.01	2.10
25	570	185.09	184.64	184.28	183.06	1.22	2.13	2.02	2.23
26	587	185.39	184.70	183.85	182.11	1.74	2.24	2.18	2.03
27	604	185.28	184.53	183.89	182.42	1.47	2.16	1.98	
28	617	185.06	184.60	184.22	182.92	1.30	2.06	2.00	2.06
29	631	185.30	184.78	184.07	182.27	1.80	2.19	2.04	2.03
30	644	185.47	184.66	184.22	182.13	2.09	2.08	1.89	1.53
31	661	185.21	184.80	183.95	182.21	1.74	2.08	1.98	1.80
32	681	185.23	184.45	184.14	182.06	2.07	2.21	2.03	
33	698	185.28	184.59	184.46	182.18	2.29	1.93	1.98	
34	711	185.79	184.67	183.92	182.53	1.39	1.97	1.89	1.79
35	728	185.37	184.64	183.91	182.29	1.62	2.25	1.90	
36	758	185.46	184.63	184.09	182.11	1.98	2.15	1.99	1.83
37	772	186.55	184.61	184.11	182.22	1.89	2.05	1.92	
38	812	185.15	184.87	183.91	182.11	1.80	2.11	2.00	
39	822	185.85	184.59	184.52	182.98	1.54	2.19	2.09	
40	832	186.08	184.88	184.06	182.79	1.27	2.12	2.09	1.60
41	842	185.70	184.75	184.38	182.73	1.65	2.09	1.96	1.89
42	849	185.96	185.51	184.33	182.83	1.49	2.05	1.89	
43	862	187.65	185.05	184.80	183.27	1.52	1.94	1.94	1.89
44	893	187.13	184.69	185.14	183.32	1.83	2.18	1.96	1.81

Ridge #	Distance Lanward (m)	Crest Elev. IGLD85 (m)	Swale Elev. IGLD85 (m)	Foreshore Top IGLD85 (m)	Foreshore Base IGLD85 (m)	Foreshore Thickness (m)	Mean Dune (phi)	Mean Fore-shore (phi)	Mean Upper Shoreface (phi)
45	919								
46	940	186.63	184.75	184.25	182.49	1.77	2.05	1.99	
47	966	186.31	184.73	184.23	182.59	1.65	2.12	1.73	
48	997	186.01	184.77	183.51	182.45	1.07	1.92	1.69	1.87
49	1007	186.18	185.68	184.29	182.80	1.49	1.98	1.92	1.51
50	1027	186.73	185.76	185.32	183.15	2.16	2.01	1.81	1.78
51	1074	187.17	186.52	185.17	183.65	1.52	1.87	1.58	
52	1094	188.27	186.16	186.09	184.29	1.80	2.05	1.83	
53	1121	187.84	185.67	185.18	183.77	1.42	1.95	1.63	1.47
54	1141	187.43	185.95	184.88	183.42	1.46	1.84	1.86	1.55
55	1161	186.41	185.53	185.49	184.09	1.40	1.94	1.95	
56	1188	186.28	185.52	184.77	183.33	1.43	1.96	1.92	0.94
57	1208	187.61	185.71	184.70	183.48	1.22	2.05	1.89	1.48
58	1221	187.01	186.61	184.95	183.55	1.40	1.87	1.74	1.41
59	1242	187.20	185.84	185.65	183.79	1.85	1.96	1.74	1.38
60	1262	186.52	185.08	184.62	183.55	1.07	1.89	1.54	1.52
61	1289	187.68	186.34	184.63	183.38	1.24	1.83	1.72	1.66
62	1309	187.09	186.51	185.70	184.56	1.14	1.72	1.74	
63	1342	187.01	185.68	185.60	184.53	1.07	1.86	1.76	1.89
64	1356	187.89	185.90	185.44	182.94	2.50	1.79	1.64	1.88
65	1376	187.44	186.43	184.94	183.11	1.83	1.87	1.51	1.77
66	1403	187.69	186.57	185.79	184.02	1.77	1.88	1.80	1.75
67	1433	187.06	185.68	185.96	183.24	2.71	2.06	1.97	1.70
68	1483	187.28	186.02	185.03	182.74	2.29	1.90	1.61	1.84
69	1510	187.21	185.69	185.16	182.81	2.35	2.12	1.61	1.54
70	1534	186.45	186.13	185.05	183.04	2.01	1.94	1.69	1.61
71	1564	187.80	186.37	185.46	184.05	1.41	1.95	1.65	1.88
72	1587	187.57	185.71	185.54	183.17	2.38	1.95	1.78	
73	1597	187.72	186.50	184.61	182.89	1.72	1.97	1.44	1.77
74	1617	187.10	186.34	185.35	184.28	1.07	1.90	1.71	1.71
75	1634	187.00	186.46	185.52	184.17	1.34	1.91	1.54	1.59
76	1651	187.97	187.28	186.41	183.79	2.62	1.93	1.72	2.03
77	1664	187.84	186.72	186.30	184.60	1.70	1.86	1.56	
78	1685	187.67	186.58	186.28	185.12	1.16	1.65	1.57	1.67
79	1705	189.29	187.46	185.66	184.17	1.49	1.85	1.54	1.72
80	1718	188.77	187.09	186.33	184.78	1.55	1.80	1.74	
81	1732	188.55	187.05	186.36	185.53	0.82	1.86	1.85	
82	1748	188.01	187.46	186.45	185.20	1.25	1.86	1.84	
83	1772	188.15	187.09	186.28	185.97	0.30	1.89	1.87	

JOHN W. JOHNSTON

Curriculum Vitae

EDUCATION

2001 to Present	Ph.D. Geology, Indiana University, Bloomington, Indiana. Major: sedimentology, Minor: climatology
1999	M.Sc. Geology, University of Waterloo, Waterloo, Ontario.
1995	B.Sc. Physical Science, Minor in Geology, University of Guelph, ON.

PROFESSIONAL EXPERIENCE

1998 to Present	Research Associate, Indiana Geological Survey, Indiana University
1995-1998	Teaching Assistant, University of Waterloo, Waterloo, Ontario.
Summer 1995	Hydrodynamic Geologist, Chevron Canada Resources, Calgary, AB.

HONOURS/AWARDS/GRANTS

- IU Department of Geological Sciences Award for Academic Achievement 2003-04
- Geological Society of America - Limnogeology Div. Distinguished Service Award 2003
- International Association for Great Lakes Research Scholarship 2003
- IU Department of Geological Sciences Estwing Award 2003
- Best Graduate Research Presentation Award at D.O.G.S. Research Day 2003
- University Graduate Fellowship 2001-03
- IU Bloomington Professional Council and the Office of the Chancellor Award 2002
- Best Graduate Research Presentation Award at D.O.G.S. Research Day 2002
- AASG/NSF/USGS Field Mentoring Experience Program Grant 2002
- North-Central Geological Society of America Travel Grant
- University of Waterloo Graduate Scholarships 1997
- University of Guelph Geology representative for CSPG - SIFT 1995
- Canadian Society of Petroleum Geologists 1993-94 Undergraduate Award
- Geological Association of Canada 1993-94 Undergraduate Award
- Esse Quam Videri Ontario Academic Credit Award 1991

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