A Thesis

entitled

Deglacial chronology and glacial stratigraphy of the western Thunder Bay lowland, northwest Ontario, Canada

by

Henry Munro Loope

Submitted as partial fulfillment of the requirements for the Master of Science degree in Geology with emphasis in Earth Surface Processes

Thesis Committee:

Dr. Timothy G. Fisher, Professor

Dr. James M. Martin-Hayden, Associate Professor

Dr. Thomas V. Lowell, Professor, University of Cincinnati

Advisor: Timothy G. Fisher

Graduate School

The University of Toledo

December 2006
An Abstract of

Deglacial chronology and glacial stratigraphy of the western Thunder Bay lowland, northwest Ontario, Canada

Henry Munro Loope

Submitted as partial fulfillment of the requirements for the Master of Science degree in Geology with emphasis in Earth Surface Processes

The University of Toledo

December 2006

Glacial stratigraphy from outcrop and lacustrine cores from the western Thunder Bay lowland, northwest Ontario, indicate glaciolacustrine deposition prior to and after a Superior Lobe advance (Marquette advance). Rhythmites (interpreted as varves) are present below (glacial Lake O’Connor) and above (glacial Lake Kaministiquia) Superior Lobe till. In Lake O’Connor, ~300 years of glaciolacustrine sedimentation is recorded prior to Superior Lobe advance into the western Thunder Bay lowland and at a minimum, 160 years of glaciolacustrine sedimentation occurred in Lake Kaministiquia when the Superior Lobe was near its maximum extent at the Marks Moraine. Glaciolacustrine sediments below Superior Lobe diamicton (glacial Lake O’Connor) coarsen upward and indicate a slowly advancing Superior Lobe. This varve chronology, coupled with new deglacial radiocarbon dates from
northwest Ontario, indicate: i) the Rainy Lobe stood at the Brule Creek Moraine at ~10,200 $^{14}$C yrs BP fronted by glacial Lake O’Connor in the western Thunder Bay lowland; ii) the Superior Lobe advanced (the Marquette advance) westward into Lake O’Connor between ~9.9 and ~9.7 $^{14}$C yrs BP; iii) glacial Lake Kaministiquia occupied the reentrant between the Marks and Dog Lake Moraines between ~9.7 and ~9.5 $^{14}$C yrs BP; and iv) the Superior Lobe retreated from the Marks Moraine after ~9.5 $^{14}$C yrs BP, fronted by a glacial lake (glacial Lake Kakabeka). This deglacial chronology is younger than previous reconstructions and indicates the possible hydrologic connection between glacial Lake Agassiz and the Superior basin may have occurred later than expected or did not occur at all (e.g., during the Younger Drays cold period). Additionally, this constraining ice margin chronology may indicate ice remained in the Superior basin prior to the Younger Dryas and that the Marquette advance may have been a short readvance.
Acknowledgements

Tim Fisher and Tom Lowell provided much guidance and discussion during the past two years. Comments by committee member Jamie Martin-Hayden improved the thesis. Josh Michaels and Colby Smith (University of Cincinnati) assisted in collection of cores from Mokomon Lake and Bruce Skubon (University of Toeldo) assisted at Echo Lake. Matt Boyd assisted in collection of organics from the Boyd Cut. Discussion with Brain Phillips guided field work. Catherine Yansa assisted in identifying material suitable for radiocarbon dating. The Comer Science and Education Foundation, Geological Society of America Graduate Student Grant, and Sigma Xi Grants-In-Aid of Research provided funding for this study.
# Table of Contents

Abstract ........................................................................................................... ii

Acknowledgements ......................................................................................... iv

Table of Contents ............................................................................................ v

List of Figures .................................................................................................. vii

List of Tables ................................................................................................... ix

Chapter 1: Introduction ................................................................................... 1
  1.1 Northwest Ontario deglaciation ................................................................. 1
  1.2 Lake Agassiz and northwest Ontario ......................................................... 2
  1.3 Thunder Bay lowland and Marks Moraine ................................................. 6
  1.4 Marquette advance .................................................................................... 11
  1.5 Superior basin lake levels ........................................................................ 14
  1.6 Objectives .............................................................................................. 15

Chapter 2: Methods ....................................................................................... 17

Chapter 3: Description of glacial sediments .................................................. 20
  3.1 Mokomon Lake ......................................................................................... 20
  3.2 Echo Lake .............................................................................................. 21
  3.3 Unnamed Lake ........................................................................................ 21
  3.4 Whitefish River cutbank exposures ......................................................... 24
  3.5 Boyd Cut ............................................................................................... 28
  3.6 Geochemistry ......................................................................................... 31

Chapter 4: Radiocarbon dates ...................................................................... 33

Chapter 5: Interpretation .............................................................................. 34
  5.1 Mokomon Lake ....................................................................................... 34
  5.2 Echo Lake ............................................................................................. 36
  5.3 Unnamed Lake ....................................................................................... 37
  5.4 Whitefish River cutbank exposures ......................................................... 37
  5.5 Proposed varve formation for glacial Lake O’Connor ................................ 39
  5.6 Boyd Cut .............................................................................................. 40
  5.7 Paleogeographic reconstructions ............................................................. 41

Chapter 6: Discussion ................................................................................... 45
  6.1 Glacial Lake O’Connor ............................................................................ 45
  6.2 Glacial Lake Kaministiquia ..................................................................... 47
List of Figures

1. Digital elevation model (3 arc second SRTM) of a section of northwest Ontario with mapped moraines (Zoltai, 1965) and Livingstone core sites from Lowell et al. (2005) and this study………………………………………………………………………………… 7

2. Digital elevation model (3 arc second SRTM) of the Thunder Bay, Ontario area. The Thunder Bay lowland extends westward from the city of Thunder Bay to the Marks Moraine. Moraine traces after Zoltai (1963), Burwasser (1977) [Intola Moraine], and author’s analysis of the DEM …………………………………………………………. 8

3. Total area covered by glacial Lake Agassiz (from Teller et al., 1983 and Teller and Leverington, 2004). Note that only parts of the lake basin were covered at any one time. Lake Agassiz merged with glacial Lake Ojibway once ice retreated north of the subcontinental divide just north of Lake Nipigon. Study area indicated by red rectangle. S = southern outlet; NW = northwest outlet; E1 = Kashabowie-Sieme eastern outlets; E2 = Nipigon eastern outlets ………………. 9

4. Bedrock geology of part of northwest Ontario (Ontario Geological Survey, 1993). Striae/drumlin/streamlined topography orientation from the Thunder Bay area from Burwasser (1977); Isle Royale from Huber (1973); area west and south of Lake Nipigon from Barnett (2004); area east of Lake Nipigon from Thorleifson and Kristjansson (1993); author added orientation data from visual inspection of DEM ……………………………………………………………………………. 12

5. Lake Superior basin lake level history (from Farrand and Drexler, 1985). Scale is in radiocarbon years BP. The large drop in lake level before 10 ka is inferred from Lake Agassiz chronology. Note Houghton low stage after Marquette advance (10 ka) ……………………………………………………………………………… 16

6. Mokomon (Mud) Lake stratigraphic log and laboratory data. Two side-by-side cores (Hole A and B, water depth = 3.2 m [10.5 ft]) were taken to recover overlapping stratigraphy. Cores were collected from Hole C closer to shore (water depth = 2.1 m [7 ft]). Note radiocarbon dates (9510±75 and 9345±75 14C yrs BP) are from Hole C closer to shore within massive silty clay just below the contact with sapropel …………………………………………………………………… 22

7. Mokomon (Mud) Lake rhythmite particle size distribution data. W = winter (non-melt season), S = summer (melt season). Photograph has been enhanced to show color contrast and microlaminations ………………………………………………………………………… 23

8. Echo Lake stratigraphic log and laboratory data. Water depth was 3.35 m (11 ft) at the coring location. Radiocarbon date (9360±90 14C yrs BP) is from sieved portion of 422-426 cm ………………………………………………………. 24
9. Unnamed Lake stratigraphic log and laboratory data. Water depth at coring site was 3.1 m (10 ft). Macrofossils were found at the sapropel – clayey silt contact (~870 cm), but not enough carbon was obtained for a radiocarbon date ………………… 25

10. Longitudinal transect of Whitefish River exposures showing stratigraphy and elevation of described outcrops. Location map is 3 arc second SRTM DEM. Elevations of exposures taken from DEM (Table 1). Dashed line indicates correlation of units between sections ………………………………………. 26

11. Harstone Cut rhythmites, A) 1.5 m above base, B) 7 m above base, C) 16 m above base, D) 18 m above base, E) particle size distribution for unit 1 (clay-very fine silt-fine silt), unit 2 (medium silt), and unit 3 (coarse silt-very fine sand). Each alternating vertical black and white bar represents one varve (annual deposit). Appendix B contains the Harstone Cut stratigraphic log for reference. Photographs have been enhanced to show color contrast ……………………… 29

12. Bioturbation from chironomid (Insecta: Diptera: Chironomidae) larvae (plan view), A) Mokomon (Mud) Lake [1167 cm depth], B) Harstone Cut [1.5 m above base], C) Hymers Cut 2 [4.5 m above base]. Trails are infilled with (lighter colored) sediment from overlying bed/laminae (usually coarser sediment) ……………….. 30

13. Boyd Cut stratigraphic log, paleocurrent data, and radiocarbon dates ………….. 32

14. Mokomon (Mud) Lake loss-on-ignition (LOI) values (Hole A: 1421-1440 cm). LOI 550°C is a proxy for organic matter percentage and LOI 950°C is a proxy for carbonate percentage. W = winter (non-melt season), S = summer (melt season). Photograph has been enhanced to show color contrast between winter and summer units ……………………………………………………………………………. 35

15. Paleogeographic reconstructions of the area west of Thunder Bay from ~10.2 ka 14C yrs BP to ~9.4 ka 14C yrs BP. Black boxes designate radiocarbon dates from each time period …………………………………………………………………………… 42
List of Tables

1. Outcrop and core locations and elevations ......................................... 13
2. Radiocarbon dates ............................................................................... 33
Chapter 1: Introduction

1.1 Northwest Ontario deglaciation

The deglacial chronology of the Laurentide Ice Sheet (LIS) within northwest Ontario is critical to understanding the possible hydrologic linkage between glacial Lake Agassiz and the upper Great Lakes between 11.0 and 8.0 ka \(^{14}\text{C}\) yrs BP (13.0 to 8.9 ka cal yrs BP). Proposed outburst floods from Lake Agassiz through the Great Lakes to the North Atlantic Ocean have been implicated in the reduction of North Atlantic thermohaline circulation and pegged as a cause for the Younger Dryas cold event (11 – 10 ka \(^{14}\text{C}\) yrs BP; 13 – 11.4 ka cal yrs BP) (Rooth, 1982; Broecker \textit{et al}., 1989). However, there is little geomorphic, sedimentologic, or stratigraphic evidence for large outburst floods from Lake Agassiz to the Superior basin during or prior to the Younger Dryas in northwest Ontario (Lowell \textit{et al}., 2005; Teller \textit{et al}., 2005). Additionally, ice margin positions of the Rainy Lobe during deglaciation in northwest Ontario have been inferred from bulk (gyttja) organic dates from lacustrine sediment (Björck, 1985; Teller \textit{et al}., 2005) which are subject to error due to incorporation of old carbon (making the dated horizon appear older) within samples, both from late Pleistocene organic matter and uptake of pre-Pleistocene carbon (ultimately derived from carbonate rocks) by aquatic vegetation (Karrow, 1992; Teller, 1989; Clayton and Moran, 1982; Björck \textit{et al}., 1998). For example, Björck \textit{et al}.
\textit{(1998)} report 100-500 yr (mean: ~300 yr) differences between AMS \(^{14}\text{C}\) dated bulk sediment (older) and terrestrial macrofossils (younger) of contemporaneous levels from Swedish lake sediments which span the Younger Dryas.
They also note that age differences are larger (500 yrs) during rapid climate shifts (i.e., at the Allerød-Younger Dryas and Younger Dryas-Preboreal boundaries). Macrofossil (wood) dated lacustrine cores are optimal, but until recently (Lowell et al., 2005; Teller et al., 2005) none have been reported that record deglacial chronology of northwest Ontario. Therefore, previous reconstructions of ice margin positions may be too old (Dyke et al., 2003) (i.e., deglaciation occurred later then previously thought). Ice margin positions of the Superior Lobe are not well constrained until 10 ka $^{14}$C yrs BP from the southern end of modern Lake Superior. Moraines in northwest Ontario were mapped initially by Zoltai (1961; 1963; 1965) and with new digital elevation models (Shuttle Radar Topographic Mission [SRTM]; van Zyl (2001); Rabus et al. (2003)), greater morphological and spatial detail of moraines are available (Fig. 1 and Fig. 2). This thesis will focus on the chronology and glacial stratigraphy of the lowland west of Thunder Bay, Ontario, rimmed by the Marks Moraine (Fig. 1 and Fig. 2). Deglacial chronology and glacial stratigraphy are important in determining the response of the LIS to the Younger Dryas cold period (i.e. regional readvance). To obtain this chronology, stratigraphy, varves, and radiocarbon dating were used.

1.2 Lake Agassiz and northwest Ontario

The study area is located along the eastern edge of the former extent of glacial Lake Agassiz (Fig. 3). This area straddles the divide between the Superior basin to the east and the Agassiz basin to the west. Lake Agassiz developed as the Red River Lobe of the Laurentide Ice Sheet retreated north of the subcontinental drainage divide (Lake Agassiz/Ojibway drainage basin in Fig. 3) at the headwaters of the Red River near the
intersection of Minnesota, South Dakota, and North Dakota. As ice retreated north, the lake occupied a larger area within the Red River drainage basin and eventually expanded eastward and westward. Further retreat resulted in the opening of lower outlets (northwest (NW) and eastern outlets (E1 and E2) in Fig. 3) and subsequent changes in lake level. Four phases of Lake Agassiz have been defined as the ice sheet retreated north. These phases (Lockhart, Moorhead, Emerson, and Nipigon) represent major changes in lake level and outlet occupation. The oldest phase, the Lockhart Phase, is thought to have begun around 11,700 \(^{14}\text{C}\) yrs BP (Fenton et al., 1983) as ice retreated from the Big Stone Moraine. However, this age is based upon interpolation of radiocarbon dates between \(~12,300\) \(^{14}\text{C}\) yrs BP of the Des Moines Lobe and \(~9,900\) \(^{14}\text{C}\) yrs BP of the later Emerson Phase of Lake Agassiz. During the Lockhart Phase, Lake Agassiz water was routed through the southern outlet to the Minnesota and Mississippi Rivers and eventually to the Gulf of Mexico (Fenton et al., 1983; Fisher, 2003) (S on Fig. 3). As the Rainy Lobe retreated further north, isostatically depressed and topographically low areas became available for Lake Agassiz water to drain through lower outlets. Teller and Thorleifson (1983) describe flow through an eastern outlet (E1 in Fig. 3) once ice retreated north of the Thunder Bay, Ontario area. This proposed flow was routed directly from Lake Agassiz to Lake Superior and then through the lower Great Lakes to the St. Lawrence River and the North Atlantic Ocean. This input of freshwater from Lake Agassiz to the North Atlantic has been proposed as a mechanism for the shutdown of thermohaline circulation and cause of the Younger Dryas cold period (11.0 – 10.0 ka \(^{14}\text{C}\) yrs BP) (Broecker et al., 1989). However, recent field work (Lowell et al., 2005) has questioned the idea that Lake Agassiz water was routed through the Kashabowie-Siene
and/or Dog-Kaministiquia valleys (E1 on Fig. 3) at the start of the Younger Dryas (~11 ka \(^{14}\text{C} \text{ yrs BP}\)). Results from Lowell et al. (2005) indicate that ice had not retreated north of the Thunder Bay area until after the beginning of the Younger Dryas. Therefore, the hypothesis that a Lake Agassiz outburst flood was the cause of the Younger Dryas may be incorrect. The accepted hypothesis for the last 20 years describes catastrophic eastern outlet flow at ~10.8 ka \(^{14}\text{C} \text{ yrs BP}\), which caused the level of the lake to drop by ~100 m to the Moorhead Phase (low water phase) (Teller and Thorleifson, 1983). During the low water Moorhead Phase (~10.8 – 10.0 ka \(^{14}\text{C} \text{ yrs BP}\)), deltas developed and wetlands were abundant (Yansa and Ashworth, 2005). Ages from Yansa and Ashworth (2005) constrain the Moorhead low from 10.2 – 9.92 ka \(^{14}\text{C} \text{ yrs BP}\) in the southern Lake Agassiz basin near Fargo, North Dakota. Bajc et al. (2000) report ages from the Rainy River basin in northwestern Ontario (10.8 – 9.9 ka \(^{14}\text{C} \text{ yrs BP}\)) although the older dates are from detrital organic matter and cannot be used to date the start of the lake phase because they have likely been reworked. However, Fisher and Lowell (2006) question the timing of the Moorhead low phase based upon the presence of reworked wood in fluvial and littoral sediments associated with the low lake level. The authors argue that the oldest dates which constrain the Moorhead low are reworked and within stratigraphic units which contain younger wood, thereby making the (earliest) start of the Moorhead phase later (~10,675 \(^{14}\text{C} \text{ yrs BP}\)) than previously thought. The Emerson Phase (~10.0 – ~9.4 ka \(^{14}\text{C} \text{ yrs BP}\) followed the Moorhead Phase and is typified by transgressive deposits within the Lake Agassiz basin. The transgression is thought to be the result of the blocking of the eastern outlets (E1 in Fig. 3) by ice of the Marquette advance dated at ~10,025 \(^{14}\text{C} \text{ yrs BP}\) in Upper Michigan (Lowell et al., 1999). The ice dam prevented water from flowing east
and thus the level of Lake Agassiz rose until it reached the northwestern outlet at 9.9 ka $^{14}$C yrs BP (Smith and Fisher, 1994). Flow through the northwest outlet may have lasted a couple hundred years until water was routed to the lower southern outlet before 9.4 ka $^{14}$C yrs BP presumably due to isostatic rebound. The southern outlet was subsequently abandoned at 9.4 ka $^{14}$C yrs BP (Fisher, 2003). Additional dates constraining the Emerson Phase are from Bajc et al. (2000) (9,750 – 9,500 $^{14}$C yrs BP) and Teller et al. (2000) (10,040 – 9,330 $^{14}$C yrs BP). The Emerson Phase ended and Nipigon Phase began as Rainy Lobe ice retreated north of the Nipigon eastern outlets ~9.5 ka $^{14}$C yrs BP (E2 in Fig. 3) (Teller and Thorleifson, 1983), although this age is only an estimate and results within this thesis constrain the possible chronology of eastern outlet occupation (E2 in Fig. 3). Nipigon eastern outlet occupation is assumed to be from ~9.5 to ~8.5 ka $^{14}$C yrs BP (Teller and Thorleifson, 1983). After ~8.5 ka $^{14}$C yrs BP, Lake Agassiz merged with glacial Lake Ojibway (Teller and Thorleifson, 1983) and flow was through the Ottawa River valley to the St. Lawrence River (bypassing the Great Lakes).

Also of importance in relation to Lake Agassiz and northwest Ontario is glacial Lake Kaministiquia. The lake formed when ice was at its Marks and Dog Lake Moraine positions (Fig. 2), impounding water between the ice and the subcontinental drainage divide (Fig. 1). Zoltai (1963) mapped the distribution of glaciolacustrine sediments between the Dog Lake and Marks Moraines and proposed that the lake drained to the west via the Siene River valley. After retreat from the Marks Moraine, Zoltai (1963) and Burwasser (1977) proposed the lake drained south via the modern Kaministiquia River valley which cuts through the Marks Moraine. Determining the routing of proglacial water associated with the Marks and Dog Lake Moraines is important because red clays
within the Lake Agassiz basin further west are thought to have come from glacial Lake Kaministiquia. If the timing of the potential overflow from Lake Kaministiquia is known, then the ages of the moraines further west can be constrained from the varve chronology near Dryden, Ontario (Warman, 1991; Minning et al., 1994; Sharpe et al., 1992).

1.3 Thunder Bay lowland and Marks Moraine

Zoltai (1963) made the first detailed glacial geology map of the area west of Thunder Bay (Fig. 2) and Burwasser (1977) provided greater detail of the Quaternary geology around the city of Thunder Bay in the eastern end of the lowland. The lowland is situated along the boundary between Superior Province Archean granite-greenstone terrane and the Paleoproterozoic Amimikie Group (Gunflint and Rove Formations) (Burwasser, 1977; Ojakangas et al., 2001) (Fig. 4). The Gunflint Formation is composed of shallow water carbonates (including stromatolites), shale, volcanic beds (tuff), and iron formation while the Rove Formation is dominantly shale (Fralick et al., 2002). Multiple glacial cycles have exploited this contact, scouring out deep (50 m) bedrock troughs in places near Thunder Bay (Burwasser, 1977). However, the bedrock topography (and glacial sediment thickness) varies considerably within the Thunder Bay lowland. For example, in the Whitefish River valley (Whitefish 1 exposure; Table 1 and Fig. 2), glacial sediment thickness is >30 m, although bedrock occurs at the surface <3 km away. The bedrock troughs act as localized centers of accumulation, as individual diamicton units in
Figure 1. Digital elevation model (3 arc second SRTM) of a section of northwest Ontario with mapped moraines (Zoltai, 1965) and Livingstone core sites from Lowell et al. (2005) and this study.
Figure 2. Digital elevation model (3 arc second SRTM) of the Thunder Bay, Ontario area. The Thunder Bay lowland extends westward from the city of Thunder Bay to the Marks Moraine. Moraine traces after Zoltai (1963), Burwasser (1977) [Intola Moraine], and by author’s analysis of the DEM.
Figure 3. Total area covered by glacial Lake Agassiz (from Teller et al., 1983 and Teller and Leverington, 2004). Note that only parts of the lake basin were covered at any one time. Lake Agassiz merged with glacial Lake Ojibway once ice retreated north of the subcontinental divide just north of Lake Nipigon. Study area indicated by red rectangle. S = southern outlet; NW = northwest outlet; E1 = Kashabowie-Siene eastern outlets; E2 = Nipigon eastern outlets.
troughs may be 5 m thick, while the same diamicton may be <1 m in thickness where bedrock is near the surface. Diamicton units within the lowland are grey (Rainy Lobe sediments sourced from Shield lithologies) or red-brown (Superior Lobe sediments sourced from Sibley Group conglomerate-sandstone-shale, Animikie Group shale-carbonate-iron formation, and Keweenawan diabase) (see Fig. 4 for bedrock source areas). Many glacial units within the Thunder Bay lowland are brown-red in color (e.g., 7.5YR 3/2 (dark brown); 10YR 4/3 (brown); 2.5YR 4/1 (dark reddish grey)), being distinctive of Superior Lobe sediments derived from Proterozoic red sandstones-siltstones within the Superior Basin. A major source for the sediments within the Thunder Bay lowland is the Proterozoic Sibley Group, composed of red conglomerate, sandstone, and shale exposed east and north of Thunder Bay (Bajc, 2000) (Fig. 4). Clayton and Moran (1982) were the first to correlate the Marks (Moraine) ice margin with the Saxon margin of northern Wisconsin (Lake View margin and Douglas Member of the Miler Creek Formation of Clayton, 1984; Attig et al., 1985), the Sixmile margin in western Upper Michigan (Peterson, 1986), and the Marquette margin in central and eastern Upper Michigan (Hughes and Merry, 1978) (Grand Marais I margin of Drexler et al., 1983). Teller et al. (2005) recently provided the only chronological constraint on the correlation of the Marks margin with the well dated Lake View and Marquette margins (Lowell et al., 1999), providing dates of 9990±360 14C yrs BP (GX-11407) and 9640±450 14C yrs BP (GX-11406) from a cutbank exposure along the Kaministiquia River inside the Marks Moraine. All other radiocarbon dates in the area are too young to be associated with the deglacial chronology (Dyke et al., 2003). The oldest of these dates are associated with glacial Lake Minong (9380±150 14C yrs BP [GSC-287] (Zoltai, 1965), 9260±170 14C yrs
BP [TO-547] (Julig et al., 1990)), which formed in the Lake Superior basin as ice retreated northward. The presence of brown-red diamicton inside the Marks Moraine, the configuration of the moraine close to Lake Superior, the cross-cutting relationship with the Brule Creek Moraine (Fig. 2), and streamlined topography (including drumlins) (Fig. 4) lends support to westward Superior Lobe flow after retreat of Rainy Lobe ice to the north.

1.4 Marquette Advance

The discovery of the in-situ Lake Gribben buried forest near Marquette, MI and designation of the Marquette Stadial by Hughes and Merry (1978) (Marquette Phase of Karrow et al. (2000)) initiated a revised post-Great Lakean history of the LIS in the Superior Basin. Prior to the discovery of the Lake Gribben site, Hack (1965) and Black (1976) reported radiocarbon ages on wood within and above till between ~10.4 and 9.6 ka $^{14}$C yrs BP from western Upper Michigan and northern Wisconsin. More recently, Lowell et al. (1999) and Pregitzer et al. (2000) added additional age control to the Lake Gribben site. Incorporating all dates associated with the Marquette Stadial, a mean age of 9990±17 $^{14}$C yrs BP was calculated using CALIB 5.0 (Stuiver and Reimer, 1993). Based on the work reported by Hack (1965), Black (1976), and Hughes and Merry (1978), a number of publications in the early and middle 1980s (e.g. Clayton and Moran (1982), Teller and Thorleifson (1983), Clayton (1983), Drexler et al. (1983), Teller (1985)) put forth the concept that the LIS retreated from the Superior Basin prior to ~10.9 ka $^{14}$C yrs BP in order to allow Lake Agassiz to overflow into the Superior Basin, initiating the Moorhead low phase in the Agassiz basin. Additionally, these studies suggested
Figure 4 (from previous page). Bedrock geology of part of northwest Ontario (Ontario Geological Survey, 1993). Striae/drumlin/streamlined topography orientation from the Thunder Bay area from Burwasser (1977); Isle Royale from Huber (1973); area west and south of Lake Nipigon from Barnett (2004); area east of Lake Nipigon from Thorleifson and Kristjansson (1993); author added orientation data from visual inspection of DEM.

Table 1. Outcrop and core locations and elevations

<table>
<thead>
<tr>
<th>Outcrops</th>
<th>Latitude (N)(^a)</th>
<th>Longitude (W)(^a)</th>
<th>Elevation(^b,c)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Whitefish 1 Cut</td>
<td>48° 17.28’</td>
<td>89° 45.34’</td>
<td>293.5 m (963’)</td>
</tr>
<tr>
<td>Whitefish Adjacent 1</td>
<td>48° 17.37’</td>
<td>89° 45.68’</td>
<td>324.5 m (1065’)</td>
</tr>
<tr>
<td>Whitefish Adjacent 2</td>
<td>48° 17.33’</td>
<td>89° 45.58’</td>
<td>319 m (1047’)</td>
</tr>
<tr>
<td>Hymers Cut 1</td>
<td>48° 18.21’</td>
<td>89° 42.30’</td>
<td>260 m (853’)</td>
</tr>
<tr>
<td>Hymers Cut 2</td>
<td>48° 17.84’</td>
<td>89° 43.06’</td>
<td>267 m (876’)</td>
</tr>
<tr>
<td>Harstone Cut</td>
<td>48° 21.31’</td>
<td>89° 38.03’</td>
<td>222.5 m (730’)</td>
</tr>
<tr>
<td>Flint Cut</td>
<td>48° 21.01’</td>
<td>89° 40.06’</td>
<td>238 m (781’)</td>
</tr>
<tr>
<td>Boyd Cut</td>
<td>48° 20.43’</td>
<td>89° 21.53’</td>
<td>186 m (610’)</td>
</tr>
<tr>
<td>Lacustrine cores</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mokomon Lake</td>
<td>48° 31.07’</td>
<td>89° 30.79’</td>
<td>438 m (1437’)</td>
</tr>
<tr>
<td>Echo Lake</td>
<td>48° 19.53’</td>
<td>89° 53.38’</td>
<td>479 m (1572’)</td>
</tr>
<tr>
<td>Unnamed Lake</td>
<td>48° 17.70’</td>
<td>89° 35.83’</td>
<td>411.5 m (1350’)</td>
</tr>
</tbody>
</table>

\(^a\) NAD83 datum
\(^b\) Elevations taken from SRTM DEM
\(^c\) Outcrop elevation (a.s.l.) is from bottom of the exposure

the Marquette advance (10 ka \(^{14}\)C yrs BP) of the Superior Lobe blocked the eastern outlets and resulted in the transgressive Emerson phase in the Agassiz basin. This paper will describe sediments associated with the Marquette advance and try to test hypotheses regarding the chronology and distance of ice advance.
1.5 Superior basin lake levels

Subsequent to the Marquette advance (10 ka $^{14}$C yrs BP), a large glacial lake (glacial Lake Minong) fronted the LIS as it retreated north out of the basin. Lake Minong continued to receive meltwater from the LIS until about 8.1 ka $^{14}$C yrs BP (~9000 cal yrs BP) (Breckenridge et al., 2004). After this point, LIS water was diverted into the Arctic drainage (to Hudson Bay) and Superior basin water levels began to drop. The level became low because of lack of meltwater entering the basin and because the lake had an isostatically lowered outlet in the southeast part of the basin (Farrand and Drexler, 1985). The low stage was named the Houghton low stage by Farrand (1960) (Fig. 5). Farrand (1960) first named the Houghton low stage of Lake Superior based upon organic silt in a core found 60 feet below the modern level of Portage Lake near Houghton, Michigan. As Farrand notes, a sill at a depth of 20-25 feet connects Portage Lake with Lake Superior. The depth of the organic silt from Portage Lake, although not dated, indicates that Lake Superior was lower than modern levels in the past. Earlier work (Taylor, 1931) recognized lower water levels must have existed in the Superior basin, although the details of the pre-Nipissing and Nipissing Great Lakes have evolved since that time. Saarnisto (1975) reported the first age estimate based on radiocarbon chronology for the Houghton low stage. A bulk date of 7590±180 $^{14}$C yrs BP (Hel-396) from Crozier Lake and a bulk date of 8100±180 $^{14}$C yrs BP (Hel-397) from Fenton Lake on the east side of the Superior basin indicate that the Houghton level was reached between 8100 $^{14}$C yrs BP and 7590 $^{14}$C yrs BP. Additional dates for the Houghton low stage come from Fisher and Whitman (1999) from Beaver Lake, Michigan on the south side of the basin (8520±60 $^{14}$C yrs BP [WW-1363], 8360±70 $^{14}$C yrs BP [WW-1356], 8320±60 $^{14}$C yrs BP [WW-
1362], and 7340±50 $^{14}$C yrs BP [WW-1357]). However, the oldest three dates may be too old because they are peat or bulk dates. The youngest date (7340±50 $^{14}$C yrs BP) is on wood, so it may represent the age of the Houghton low. After the Houghton low stage, lake levels began to rise as the outlet which controlled lake level began to rebound (isostatic rebound). The resulting rise is named the Nipissing transgression and Lake Nipissing peaked in level just after 5 ka $^{14}$C yrs BP when all three upper Great Lakes were confluent. Outlet incision resulted in fall from the Nipissing high stand and isolation from Lake Michigan/Lake Huron occurred as the sill at Sault Sainte Marie became the outlet for Lake Superior due to differential isostatic rebound (Johnston et al., 2004).

1.6 Objectives

This research is part of a larger project involving the deglacial chronology of the LIS and history of Lake Agassiz and its outlets. The timing of outlet occupation and routing has been a major foci of this project because the routing of Lake Agassiz water has been implicated in abrupt climate change as described in the first part of this introduction. My side of the project involved determining the chronology of deglaciation of the Rainy and Superior Lobes and proglacial lake evolution west of Thunder Bay, Ontario with emphasis placed on the chronology of the Marks Moraine (Fig. 2). The Marks Moraine is important because it has been correlated to the well-dated ~10,000 $^{14}$C yrs BP margin of the south side of the Superior basin. However, the correlation is tenuous because a continuous (or quasi-continuous) moraine cannot be traced along the north shore of Minnesota and no radiocarbon dates constrain the timing of Superior Lobe
Figure 5. Lake Superior basin lake level history (from Farrand and Drexler, 1985). Scale is in radiocarbon years BP. The large drop in lake level before 10 ka is inferred from Lake Agassiz chronology. Note Houghton low stage after Marquette advance (10 ka).

advance to the Marks Moraine or immediate retreat (until ~9300 $^{14}$C yrs BP). I used radiocarbon dating, varves (floating chronology), and glacial stratigraphy (relative chronology) to deduce the timing of glacial events in the Thunder Bay lowland.
Chapter 2: Methods

Sediment cores were collected from three lakes (Echo Lake, Mokomon Lake, and Unnamed Lake in Table 1 and Fig. 2) on both sides of the Marks Moraine. Cores were obtained from lake ice using a modified (square-rod) Livingstone coring system (modified from Wright [1967]) to obtain glacial and post-glacial stratigraphy as well as radiocarbon dates from macrofossils within the lowermost post-glacial sediment. To minimize the ‘kettle’ problem (i.e., slow stagnant ice decay which may give a large range of radiocarbon ages and hence not reflect the true time of deglaciation [Florin and Wright, 1969]) lakes within small bedrock basins were cored. In order to recover the complete post-glacial sediment sequence, a hydraulically-assisted Livingstone coring system was used. In most cases, glacial sediment (till, glaciolacustrine) density is too great for recovery with a standard Livingstone corer. A hydraulic cylinder clamped to push rods allowed the coring tube to penetrate dense glacial sediments and enable recovery of the entire post-glacial sequence as well as part of the glacial sequence. Multiple one m long cores were taken at each site to increase the probability of encountering macrofossils and also to verify stratigraphy. The cores were extruded, wrapped, and labeled on site. In the laboratory, the cores were split, photographed, and described. Macrofossils for radiocarbon dating were identified by microscope after sieving the lowermost organic-inorganic sediment contact (i.e. the glaciolacustrine-post glacial sediment contact). Samples were also taken for particle size analysis. Loss-on-ignition was performed mostly at two cm intervals (five cm, four cm, and one cm
intervals were used where sediment texture, mineralogy, color, organic matter content and carbonate content were thought to be constant or, in the case of one cm sampling interval, a smaller interval was deemed appropriate in order to delineate a contact with greater confidence) to determine percent organic matter (550°C) and percent carbonate (950°C) following the methodology of Heiri et al. (2001).

Volume magnetic susceptibility (κ) (10⁻⁵ SI units) was measured at one cm intervals parallel to bedding using a Bartington MS2E surface scanning sensor. Magnetic susceptibility is used as a measurement of the concentration and grain size of magnetic minerals (e.g. detrital magnetite) (Verosub and Roberts, 1995). In this study, magnetic susceptibility is used as a proxy for glaciogenic clastic input within small bedrock catchments. Post-depositional formation or alteration (authigenetic/diagenetic) of magnetic minerals is not considered here because this study is only interested in identifying the contact between glaciolacustrine and post-glacial (organic rich) sediments.

Particle size analysis was completed on a Malvern Mastersizer 2000 laser diffractom using a Hydro 2000SM sample dispersion accessory. Sample preparation and optical settings for the Mastersizer 2000 generally followed Sperazza et al. (2004). ASTM standards were used for sodium hexametaphosphate (Calgon) concentration (40 g/L) (ASTM International, 1998) and because the Hydro 2000SM accessory was used, sonication was done prior to samples being introduced into the sample dispersion accessory. Samples were treated with dilute HCl, soaked in Calgon for >24 hours, and sonicated for >60 seconds prior to pipette introduction into the sample dispersion accessory. A refractive index (RI) value of 1.544 (silica) and an absorption value of 1 was used based on results of Sperazza et al. (2004) from fine grained (<50 µm)
sediments, similar to the particle size we examined. Rhythmite thickness and greyscale measurement was made using the National Institute of Health image processing program, ImageJ 1.35s. Eight cutbank exposures were stratigraphically logged within the Whitefish and Kaministiquia River valleys (Table 1 and Fig. 2). Parameters recorded at each exposure included: i) thickness of units, ii) color, iii) nature of contacts, iv) composition (mineralogy), and v) grain size. Additional details such as sedimentary structures, paleocurrent indicators, and sorting were described when applicable. At the Boyd Cut exposure, samples for radiocarbon dating were collected.

Select glacial sediment samples were sent for geochemical analysis to ALS Chemex Laboratories (Sparks, Nevada) to try to delineate glacial units based on elemental concentrations (i.e., provenance of individual units). Twenty samples were run in duplicate (40 samples total) with analysis by ICP-MS (HF-HNO$_3$-HClO$_4$ acid digestion) for 47 elements.
Chapter 3: Description of glacial sediments

Lacustrine cores

3.1 Mokomon Lake

Mokomon Lake is located within a small (<1 km$^2$) bedrock basin just south of the Marks Moraine as mapped by Zoltai (1963) and Burwasser and Ferguson (1980) (Fig. 2). The 12.5 m-long sediment sequence from Mokomon Lake consists of clayey silt-clayey silt rhythms (couplets), diamicton, massive silty clay, and sapropel (gyttja) (Fig. 6). The lower 7.5 m is composed of 5YR 4/1 (dark gray) clayey silt (average: 5.35 µm) and 5YR 5/2 (reddish gray) clayey silt (average: 5.14 µm) couplets (Fig. 7) and two zones of diamicton, with the rhythms accounting for the majority of the thickness (6 m). The rhythms show variable thickness, from <1 cm to ~12 cm. Many of the 5YR 4/1 clayey silt beds (laminae) contain laminae (sub-laminae), indicating multiple sedimentation events within the deposition of each unit. Less than ten of the rhythms contain graded sand laminae/beds. However, coarse sand, very coarse sand, and small granules are present within the rhythms as lonestones, being more common in the thicker (>5 cm) rhythms. Larger granules and pebbles are uncommon, with less than five found within the rhythms. The diamicton units truncate rhythms and some contain deformed or overturned sections of rhythms. Pebbles, granules and very coarse sand are found within the diamicton units, but the matrix particle size of the diamicton units is clayey silt, similar to that of the rhythms. A 0.5 m-thick massive silty clay overlies rhythms from a depth of 10 to 10.5 m and 5 m of sapropel overlies the massive silty clay.
Through visual inspection, 158 rhythmites were counted from Hole A and 169 rhythmites from Hole B within the lower 7.5 m (Fig. 6). The discrepancy between cores is possibly due to truncation of couplets by mass movements. Magnetic susceptibility, loss-on-ignition, and varve thickness measurements are shown in Fig. 6.

3.2 Echo Lake

Echo Lake is located within a narrow northeast-southwest trending bedrock basin (<1 km$^2$) just outside the Marks Moraine as mapped by Zoltai (1963) (Fig. 2 and Table 1). The 1.7 m-long core from Echo Lake contains 1.4 m of laminated sandy silt and silty sand and 0.3 m of peaty sapropel (Fig. 8). The lower 1.4 m is 5Y 5/1 (gray) and 2.5Y 4/1 (dark gray), contains abundant muscovite, and has a gradational upper contact with the overlying peaty sapropel.

3.3 Unnamed Lake

Unnamed Lake is located within a small (<1 km$^2$) bedrock basin among a set of diabase buttes above the Thunder Bay lowland south of the Kaministiquia River and east of the Whitefish River (Fig. 2 and Table 1). The basal 1.7 m of the nearly 4 m sequence consists of interbedded and interlaminated clayey silt, very fine to fine sand, and coarse sand to granules. The upper 2.2 m consists of sapropel (gyttja), peat, and peaty sapropel (Fig. 9).
Figure 6. Mokomon (Mud) Lake stratigraphic log and laboratory data. Two side-by-side cores (Hole A and B, water depth = 3.2 m [10.5 ft]) were taken to recover overlapping stratigraphy. Cores were collected from Hole C closer to shore (water depth = 2.1 m [7 ft]). Note radiocarbon dates (9510±75 and 9345±75 14C yrs BP) are from Hole C closer to shore within massive silty clay just below the contact with sapropel.
Figure 7. Mokomon (Mud) Lake rhythmite particle size distribution data. W = winter (non-melt season), S = summer (melt season). Photograph has been enhanced to show color contrast and microlaminations.

Avg. mass median diameter \([d(0.5)]\) for all samples

- Summer (melt season): 7.69 phi (5.35 \(\mu\)m) \(n=42\)
- Winter (non-melt season): 7.94 phi (5.14 \(\mu\)m) \(n=23\)
Figure 8. Echo Lake stratigraphic log and laboratory data. Water depth was 3.35 m (11 ft) at the coring location. Radiocarbon date (9360±90 14C yrs BP) is from sieved portion of 422-426 cm.

3.4 Whitefish River cutbank exposures

The stratigraphy of seven cutbank exposures along the Whitefish River is presented in Fig. 10 (see Fig. 2 and Table 1 for location). The stratigraphy and sediments of each of the seven exposures along the Whitefish River can be correlated in order to determine a relative chronology of glacial events. The basal unit (unit A) within the Whitefish River valley is a black bouldery diamicton. This unit was only seen at the Flint Cut, where it contained abundant striated Shield lithology boulders, as well as locally derived (<10 km to the north) Gunflint Formation boulders.

Unit B is a fine-grained (mainly silt with subordinate clay and sand) rhythmically interbedded and interlaminated unit. At the Flint Cut, unit B conformably drapes unit A
Figure 9. Unnamed Lake stratigraphic log and laboratory data. Water depth at coring site was 3.1 m (10 ft). Macrofossils were found at the sapropel – clayey silt contact (~870 cm), but not enough carbon was obtained for a radiocarbon date.
Figure 10. Longitudinal transect of Whitefish River exposures showing stratigraphy and elevation of described outcrops. Location map is 3 arc second SRTM DEM. Elevations of exposures taken from DEM (Table 1). Dashed line indicates correlation of units between sections.
and the contact between the two is undulatory. Unit B is present in all of the exposures except for Whitefish Adjacent 1 and 2, where only diamicton is present (unit C). These two exposures lie on the northern margin of a bedrock trough and hence unit B might not be preserved at those sites. Unit B is best represented and is the thickest at the Harstone Cut (Fig. 10), where 17 m of rhythmically interbedded and interlaminated very-dark gray, very-fine silt–to–fine silt, gray medium silt, olive-gray coarse silt–to–very fine sand, and reddish brown clay are exposed. Unit B is thinner in the other sections, being ~10 m thick at the Whitefish 1 Cut, Hymers Cut 1 and Hymers Cut 2. At the Flint Cut, the unit is ~7 m in thickness where it conformably overlies the undulating contact with unit A. Additionally, unit B does not show the same detailed characteristics at all exposures. At the Whitefish 1 Cut, unit B is dominated by silt, but no clear rhythmicity is seen. At the Hymers Cut 2 exposure, the lower 8 m of unit B is composed of rhythmic very fine sand, clayey silt, and silty clay, and the upper 3.5 m of unit B is composed of interbedded very fine–to–fine sand and clayey silt. The sandy beds are rippled and soft sediment deformation occurs where sand beds overlie clayey silt beds. The sandy beds are typically thicker in the upper 3.5 m of unit B than they are in the lower 8 m. At the Harstone Cut and Hymers Cut 2, unit B coarsens upward, and at the Harstone Cut, the thicker and coarser rhythmites at the top of unit B are overlain by diamicton (unit C). However, this transition is not sharp, with ~2 m of a sand-silt unit containing blocks of diamicton subsequently overlain by brown-red diamicton. The Harstone Cut gives the best example of the rhythmic nature of glacial Lake O’Connor sediments. Particle size data (Fig. 11E) indicate three distinct grain size populations, with units 2 and 3 being summer (melt season) deposits and unit 1 being winter (non-melt season) sediment.
Heavily bioturbated zones are limited to gray, medium silt beds and laminae, although zones of very-dark gray, very-fine silt–to–fine silt contain bioturbation traces, but the traces are thinner and the zones are less frequent. The traces are similar to chironomid (Insecta: Diptera: Chironomidae) larvae trails (Duck and McManus, 1984; Morrison, 1987) (Fig. 12). In general, the traces are ~1 mm in diameter and occur as curvilinear and straight segments. The traces are most visibly seen at the contact between gray medium silt and very-dark gray, very-fine silt–to–fine silt beds/laminae, where chironomid larvae burrowed downward and filled burrows with gray medium silt (Figs. 11B, 11C, and 12).

Unit C is found at each cut except for Hymers Cuts 1 and 2, where unit E overlies unit B. Unit C contains both brown-red and gray diamicton as seen at the Flint Cut, Whitefish 1 Cut, and Whitefish Adjacent Cuts 1 and 2. Unit D, only seen at the Flint Cut, is composed of silt couplets and has a lower sharp contact with diamicton (unit C). Unit E is present at Hymers Cut 1 and 2 contains cross-bedded sand and gravel and plane-bedded sand with interlaminated silt.

3.5 Boyd Cut

The Boyd Cut stratigraphic log and data is shown in Fig. 13. This 14 m high exposure along the Kaministiquia River (Table 1 and Fig. 2) exposes from bottom to top: i) a basal 4 m thick unit of very-fine sand which contains ripple-drift cross lamination and climbing ripples (unit 1); ii) an overlying 5 m thick package of interbedded very-fine sand, silt, and clayey silt with starved sand ripples and rhythmic beds of sand and clayey
Figure 11. Harstone Cut rhythmites, A) 1.5 m above base, B) 7 m above base, C) 16 m above base, D) 18 m above base, E) particle size distribution for ‘1’ (clay-very fine silt-fine silt), ‘2’ (medium silt), and ‘3’ (coarse silt to very fine sand) from far right side of A-D. Each alternating vertical black and white bar represents one varve (annual deposit). Appendix B contains the Harstone Cut stratigraphic log for reference. Photographs have been enhanced to show color contrast.

silt at the top of the unit (unit 2); iii) a 3 m thick iron-stained trough cross-bedded sand and gravel unit (unit 3); iv) a <2 cm thick organic bed containing abundant charcoal (unit 4); and v) a 1 m thick interbedded and interlaminated brownish-red clayey silt and sand unit (unit 5). The modern soil profile extends ~1 m from the top of the exposure.
Figure 12. Bioturbation from chironomid (Insecta: Diptera: Chironomidae) larvae (plan view), A) Mokomon (Mud) Lake [1167 cm depth], B) Harstone Cut [1.5 m above base], C) Hymers Cut 2 [4.5 m above base]. Trails are infilled with (lighter colored) sediment from overlying bed/ laminae (usually coarser sediment).

Paleocurrent measurements taken from ripples 3 to 6 m above the base of the section (Fig. 13) indicate transport to the northeast, while one measurement taken at 9.5 m indicates a northwest transport direction.
3.6 Geochemistry

Geochemistry data is presented in Appendix C. Sample names and sample depth at each exposure are presented along with the stratigraphic log of each exposure in Appendix B. The ability to delineate provenance and correlate glacial units from elemental values for diamicton and glaciolacustrine units between exposures (Whitefish River exposures) was unsuccessful. For example, the chemistry of unit B (Fig. 10) varied between exposures although the sediments are interpreted to have been deposited in the same lake. Based on these results, clast lithology appears to be a better indicator of provenance (e.g., Rainy Lobe versus Superior Lobe) within the Thunder Bay lowland.
Figure 13. Boyd Cut stratigraphic log, paleocurrent data, and radiocarbon dates.

interlaminated 10R 3/4 (dusky red) clayey silt (avg. d(0.5): 13μm [n=2]) and 7.5YR 4/2 (brown) coarse silt- v.fine sand
8070 ± 70 14C yrs BP  δ13C:-32.2  (ETH-31437)
7995 ± 65 14C yrs BP  δ13C:-31.6  (ETH-31438)

Fe-stained trough cross-bedded sand and gravel, west to east transport direction

Type B climbing ripples
N 40° W
rhythmic sedimentation (30-40 cm thick couplets)
10R 4/4 (weak red) more pronounced from ~5 m to 10 m
interbedded and interlaminated 2.5Y 4/1 (dark gray) to 2.5Y 7/2 (light gray) clayey silt
(d(0.5): 22μm) and 10R 4/4 (weak red) to 2.5Y 6/1(gray) rippled v.fine sand
starved ripples
N 68° E
N 82° E
N 90° E
N 31° E
n=6 (mean)

2.5Y 6/1 (gray) to 10R 4/4 (weak red) rippled v.fine sand (d(0.5): 77μm), abundant mica (muscovite and biotite)
Chapter 4: Radiocarbon dates

Radiocarbon ages are presented in Table 2. Ages have been corrected for fractionation from $\delta^{13}C$ values and calibrated using CALIB 5.0 (Stuiver and Reimer, 1993; Reimer et al., 2004). Radiocarbon ages have one sigma error, and the mean of the greatest probability distribution at one sigma was used as the calibrated age.

Table 2. Radiocarbon dates

<table>
<thead>
<tr>
<th>Lab # (ETH)</th>
<th>Site</th>
<th>Depth / Elevation</th>
<th>Material</th>
<th>$^{14}C$ age$^a$</th>
<th>$\delta^{13}C$ (%)</th>
<th>Calibrated age range (BP)$^c$</th>
<th>Probability</th>
<th>Calibrated age (BP)$^d$</th>
</tr>
</thead>
<tbody>
<tr>
<td>31437</td>
<td>Boyd Cut</td>
<td>653' (199 m)</td>
<td>charcoal</td>
<td>8070±70</td>
<td>-32.2</td>
<td>8832-8780</td>
<td>0.19</td>
<td>8832-8780</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>8919-8861</td>
<td></td>
<td>8963-8952</td>
<td>0.03</td>
<td>8963-8952</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>9092-8968</td>
<td></td>
<td>9117-9106</td>
<td>0.56</td>
<td>9117-9106</td>
</tr>
<tr>
<td>31438</td>
<td>Boyd Cut</td>
<td>653' (199 m)</td>
<td>charcoal</td>
<td>7995±65</td>
<td>-31.6</td>
<td>8996-8774</td>
<td>1</td>
<td>8885</td>
</tr>
<tr>
<td>31653</td>
<td>Echo Lake</td>
<td>422-426 cm</td>
<td>wood frags.</td>
<td>9360±90</td>
<td>-21.9</td>
<td>10462-10431</td>
<td>0.09</td>
<td>10462-10431</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>10707-10485</td>
<td></td>
<td></td>
<td>0.91</td>
<td>10707-10485</td>
</tr>
<tr>
<td>32171</td>
<td>Mokomon Lake</td>
<td>591 cm</td>
<td>wood</td>
<td>9510±75</td>
<td>-23.6</td>
<td>10830-10681</td>
<td>0.52</td>
<td>10756</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>10868-10840</td>
<td></td>
<td></td>
<td>0.08</td>
<td>10756</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>11070-10951</td>
<td></td>
<td></td>
<td>0.40</td>
<td>10756</td>
</tr>
<tr>
<td>32172</td>
<td>Mokomon Lake</td>
<td>591 cm</td>
<td>wood</td>
<td>9345±75</td>
<td>-23.2</td>
<td>10485-10436</td>
<td>0.08</td>
<td>10583</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>10678-10488</td>
<td></td>
<td></td>
<td>0.92</td>
<td>10583</td>
</tr>
</tbody>
</table>

$^a$ 1σ error  
$^b$ 1σ range  
$^c$ calibrated using IntCal04 (Stuiver and Reimer, 1993; Reimer et al., 2004)  
$^d$ mean of greatest probability distribution
Chapter 5: Interpretation

Lacustrine cores

5.1 Mokomon Lake

Results from magnetic susceptibility (Fig. 6) and loss-on-ignition (Fig. 14) indicate that the rhythmites are annual (varves). Magnetic susceptibility (MS) values are higher within summer beds and lower within winter beds, indicative of the mineralogy and density of melt season sediment (increased magnetic grain concentration) (Fig. 6). Suspension setting of mainly quartz (silt) and clay minerals dominates non-melt season sedimentation and produce lower MS values. Large fluctuations of MS from 10.5 to 13.4 m and 16 to 17.4 m are due to rhythmites. Loss-on-ignition data (Fig. 14) show increased carbonate values in summer beds and increased organic matter in winter beds. Dell (1973) found similar results for carbonate values in varves from Lake Superior, with summer beds having higher carbonate values as a result of rapid sedimentation and burial of carbonate-rich sediments before partial dissolution by cold, undersaturated (with respect to carbonate) lake water. Loss-on-ignition 550°C data and magnetic susceptibility delineate the mineral/organic (glacial/non-glacial) contact (Fig. 6). Particle size analysis in general shows little variation between melt (summer) and non-melt season (winter) beds, not typical for varves (Smith and Ashley, 1975). However, there are several couplets which show distinct particle size differences between winter and summer layers (~1 µm winter beds and ~10 µm summer beds). Average mass median diam \([d(0.5)]\) for winter beds was 7.94 phi, while summer beds were slightly
Figure 14. Mokomon (Mud) Lake loss-on-ignition (LOI) values (Hole A: 1421-1440 cm). LOI 550°C is a proxy for organic matter percentage and LOI 950°C is a proxy for carbonate percentage. W = winter (non-melt season), S = summer (melt season). Photograph has been enhanced to show color contrast between winter and summer units.
coarser at 7.69 phi (Fig. 7). The diamicton units within the Mokomon Lake cores (Fig. 6) are potentially a result of: i) iceberg dumps and/or ii) subaqueous debris flows. The diamicton units do not exhibit character of basal till (\emph{i.e.}, no structure and low density). Some rhythmites (varves) near diamicton units are disturbed and show evidence of soft sediment deformation.

Radiocarbon dates (Table 2 and Fig. 6) of \(9345 \pm 75\) \(^{14}\)C yrs BP (ETH-32172) and \(9510 \pm 75\) \(^{14}\)C yrs BP (ETH-32171) within massive silty clay (Fig. 6) above \(\sim 160\) varves (correlated from Holes A and B) indicate the drainage of glacial Lake Kaministiquia just prior to that time. The massive silty clay above varves indicates dominantly clastic deposition within Mokomon Lake, possibly resulting from debris flows/sediment gravity flows entering the lake. However, vegetation must have encroached and established fairly quickly once Lake Kaministiquia drained. Vegetation may have occupied high bedrock knobs adjacent to Lake Kaministiquia when the Superior Lobe was at the Marks Moraine or had just retreated from it.

5.2 \textit{Echo Lake}

The laminated sandy silt–to–silty sand from the lower 1.5 m of the core records glaciolacustrine sedimentation within the small bedrock basin containing Echo Lake prior to shallow lacustrine deposition of peaty sapropel (Fig. 8). The radiocarbon date (\(9360 \pm 90\) \(^{14}\)C yrs BP [ETH-31653]) from the sieved portion of 422 to 426 cm is just above the gradational contact between organic-rich and non-organic rich sediments. This date agrees with ages from Mokomon Lake (\(9345 \pm 75\) and \(9510 \pm 75\) \(^{14}\)C yrs BP) which indicate the drainage of Lake Kaministiquia. However, Echo Lake was very likely
deglaciated prior to the drainage of Lake Kaministiquia, as ice within the western Thunder Bay lowland would have prevented drainage of the lake. Additionally, it is possible that the age is too young to represent deglaciation because the sample was obtained above the contact between organic-rich and non-organic rich sediments, although the contact was gradational. Furthermore, meltwater from the Superior Lobe while at the Marks Moraine may have never reached Echo Lake (indicated by a dominance of gray silt, sand, and muscovite within the lower 1.5 m of the core) and hence the lake may not record the influence of the Marquette advance. If this is the case, a deglacial date should be similar to that of Sunbow or Crawfish Lake (i.e., ~10.2 ka $^{14}$C yrs BP) (Fig. 1 and Lowell et al., 2005).

5.3 Unnamed Lake

The lower 1.7 m of the 3.9 m core from Unnamed Lake (Fig. 9) indicate glaciolacustrine sedimentation within a small bedrock basin prior to accumulation of organic-rich sediment. A small glacial lake developed as the Superior Lobe retreated eastward out of the Thunder Bay lowland, likely perched among diabase buttes. The lake was likely short-lived and drained once the Superior Lobe retreated far enough east for a lower outlet to be uncovered.

5.4 Whitefish River cutbank exposures

Unit A represents lodgement till deposited by the Rainy Lobe based on its color (very dark gray to black), high degree of cementation, and clast lithology, size, and number. Unit B is composed of rhythmically interbedded and interlaminated
glaciolacustrine sediments deposited by an advancing Superior Lobe correlative to sediments of glacial Lake O’Connor described by Zoltai (1963). The sediments of unit B are interpreted to be varved (annual) based upon particle size and intervals of bioturbation (Figs. 11 and 12). Teller and Mahnic (1988) and Lemoine and Teller (1995) report bioturbation within silt laminae from glacial Lake Minong and glacial Lake Kelvin rhythmites (varves) northeast of Thunder Bay which appear similar to bioturbated beds/laminae within glacial Lake O’Connor. Warman (1991) reports similar bioturbation from glacial Lake Agassiz sediments (varves) near Dryden, Ontario.

The rhythmic nature of heavily bioturbated coarser (medium silt) beds/laminae overlain by finer (clay–to–fine silt) beds/laminae with minor bioturbation indicate annual deposition. In the lower 13 m of unit B at the Harstone Cut, winter (non-melt season) beds/laminae are thicker than summer (melt season) beds/laminae (Fig. 11). Additionally, 5Y 5/2 (olive gray) coarse silt-very fine sand stringers are found within winter beds. These coarser units within winter beds may be a result of slide/slump-generated turbidity currents as a result of lake level fall during winter (Shaw, 1977; Shaw et al., 1978). Olive gray, coarse silt–to–very fine sand are found within summer beds (gray medium silt) from 13 to 17 m at the Harstone Cut, possibly indicative of an increasingly proximal source. The fact that the distinct olive gray beds/laminae occur both in the summer and winter beds is difficult to explain. Perhaps the olive gray beds/laminae represent sediment derived from non-glacial fluvial and deltaic sources. In the summer, turbidity flows caused by increased deltaic sedimentation may account for the sediment, while in the winter, delta sediment instability caused by lake level drop may account for the presence within winter beds. From 13 to 17 m, very-dark gray, very-
fine silt–to–fine silt laminae are present within summer beds, most likely due to flocculation (O’Brien and Pietraszek-Mattner, 1998). Fecal pellets were not observed (cf. Smith and Syvitski, 1982). Very-dark gray, very-fine silt–to–fine silt laminae within the summer beds are less common in the lower 13 m of the Harstone Cut. In general, varve thickness increases upsection at the Harstone Cut, from <5 cm below 13 m to ~15 cm from 13 to 17 m to ~80 cm from 17 to 19 m (Fig. 11A-D). The increase in varve thickness and presence of till at the top of the section indicate an advancing Superior Lobe from east to west within the Thunder Bay lowland. Based on average varve thickness (5 cm below 13 m; 15 cm from 13 to 17 m; 80 cm from 17 to 19 m), ~300 years of sedimentation are represented below Superior Lobe brown-red till.

5.5 Proposed varve formation for glacial Lake O’Connor

Gray medium silt (15-30 µm) is deposited from suspension during the melt season. Sedimentation occurs throughout the melt season, allowing chironomid larvae to heavily bioturbate the unit. Increased lake productivity during the melt season also likely contributes to the increased bioturbation. Very-dark gray, very-fine silt–to–fine silt and reddish-brown clay is confined to the epilimnion until the lake overturns near the beginning of the non-melt season. However, flocculation of very-fine silt–to–fine silt during the melt season produces finer (<10 µm) laminae within summer beds (15-30 µm). Subsequent to overturning, suspension settling of clay–to–fine silt occurs, periodically interrupted by turbidity currents producing olive gray coarse silt-very fine sand laminae, sometimes cross-laminated. Multiple mm-scale graded laminations occur within the non-melt season beds (Fig. 11B), potentially indicating subaqueous sediment-gravity flows.
from unstable slopes (potentially deltas) during lake level lowering during winter (Shaw, 1977; Shaw et al., 1978).

5.6 Boyd Cut

Teller et al. (2005) reported new ages (9640±450 \(^{14}\)C yrs BP [GX-11406] and 9990±360 \(^{14}\)C yrs BP [GX-11407]) from a 10 m cutbank exposure along the Kaministiquia River, ~30 km east of the Marks Moraine. The cut (here named the Boyd Cut) was described and sampled in the hope of further refining the chronology and stratigraphy reported by Teller et al. (2005). Our dates (8070±70 \(^{14}\)C yrs BP [ETH-31437] and 7995±65 \(^{14}\)C yrs BP [ETH-31438]) (Table 2) are significantly younger although from the same organic horizon (same stratigraphic setting) as dates reported by Teller et al. (2005). Although not associated with an identifiable fluvial terrace (due to Holocene cutbank erosion) (Fig. 2), the elevation of the 3 m thick trough cross-bedded sand and gravel unit and the <2 cm thick organic bed spans 196 to 199 m a.s.l. (643 to 653 ft.), which corresponds well to the projected elevation of the Houghton level shoreline at Thunder Bay (Farrand, 1960). The dated horizon lies at the contact between channel trough cross-bedded sand and gravel and floodplain clayey silt and sand. The organics just below the clayey silt and sand are interpreted to be flotsam that was deposited in a floodplain environment (possibly in an oxbow lake). The older dates of Teller et al. (2005) are interpreted here to be reworked.
5.7 Paleogeographic reconstructions

Figure 15 depicts the interpreted locations of the LIS and proglacial lakes of the area west of Thunder Bay at ~10.2 ka $^{14}$C yrs BP, ~9.9 ka $^{14}$C yrs BP, ~9.7 ka $^{14}$C yrs BP, ~9.5 ka $^{14}$C yrs BP, and ~9.4 ka $^{14}$C yrs BP. The reconstructions are based upon radiocarbon dates (five), varves (below and above Superior Lobe till), and stratigraphy from outcrops along the Whitefish River valley.
Fig. 15 (con’t)
Fig. 15 (con’t)
Figure 15. Paleogeographic reconstructions of the area west of Thunder Bay from ~10.2 ka $^{14}$C yrs BP to ~9.4 ka $^{14}$C yrs BP. Black boxes designate radiocarbon dates from each time period.
Chapter 6: Discussion

6.1 Glacial Lake O’Connor

Zoltai (1963) initially described glacial Lake O’Connor sediments in the western Thunder Bay lowland (partly in the Whitefish River valley), evidenced by lake sediment below brown-red diamiction of the Murillo Lobe (Superior Lobe). He hypothesized that glacial Lake O’Connor was bounded to the north by ice at the Brule Creek Moraine and by ice to the east within the Thunder Bay lowland. Data from this study agrees with Zoltai’s hypothesis because glaciolacustrine sediments below brown-red diamiction at the Harstone Cut contain ~300 varves and Harstone varves contain reddish brown clay (~2 μm) (Fig. 11) and Sibley Group clasts from Superior Lobe sources. With radiocarbon dates from Lowell et al. (2005) and Fisher et al. (in prep) of 10,250±75 (ETH-31430) and 10,190±40 (Beta-195959) from Crawfish Lake and dates of 10,400±120 (ETH-31429) and 10,120±80 (ETH-30188) from Sunbow Lake (Fig. 1), an age of ~10.2 ka $^{14}$C yrs BP is assigned to the Brule Creek Moraine. Analysis of sediments at the Harstone Cut indicate glacial Lake O’Connor was in existence for at least ~300 years, likely fronted by the Brule Creek Moraine for part of that time, and that the Superior Lobe was present in the Thunder Bay lowland at least ~300 years before Superior Lobe ice advanced and reached its maximum position at the Marks Moraine in the western end of the lowland (Fig. 15). Shorelines of glacial Lake O’Connor have not been identified by the author within the study area (Fig. 2). Lake O’Connor sediments range in elevation from 222.5 m (770 ft) to 305.5 m (1002 ft) in the exposures studied (Table 1), although this range does
not capture the entire elevation range of glaciolacustrine sediment. Zoltai (1963) reports a maximum elevation for Lake O’Connor of 436 m (1,430 ft), although he only reports sediments up to 418 m (1,370 ft) within the Arrow and Stump River valleys south of the Thunder Bay lowland. These elevations are significantly above the exposures along the Whitefish River valley, although Lake O’Connor likely became larger and covered a larger elevation range as the Rainy Lobe retreated northward, uncovering lower topography. Regardless, if Lake O’Connor existed at or below ~440 m, the lake would have been contained by higher topography to the west and south (Fig. 4 of Zoltai (1963) and Fig. 1 of this thesis) and by ice to the east. As mentioned, Zoltai (1963) suggests the lake was bounded on the north by the Brule Creek Moraine, which was later overrun by the Superior Lobe. Zoltai (1963) infers that the lake drained to the south based on the presence of a short spillway (433 m [1420 ft]) and delta (428 m [1405 ft]) built southward between the Arrow and Pigeon River drainages. This research does not add any new data regarding a specific outlet location or drainage direction, but drainage to the south and/or southeast into the Pigeon and/or Swamp drainage basins and eventually into a proglacial lake(s) dammed against the Superior Lobe along the north shore of Minnesota as suggested by Phillips and Hill (2004) appears very likely. Glacial lakes impounded by the Superior Lobe along the north shore of Minnesota may have used the Brule or Portage (Moose Lake) outlets in northern Wisconsin and Minnesota, but that is speculative. Phillips and Hill (2004) point out even higher elevation (503 m [1650 ft]) glacial features (ice-contact deltas) above the elevation for Lake O’Connor west of Whitefish Lake and in the Arrow Lake drainage. Phillips and Hill (2004) suggest that these features were deposited into Lake Agassiz (upper Herman stage) based on their
elevation after correction for differential isostatic rebound. However, the age of the Herman stage is not known and deglacial dates from Lowell et al. (2005) indicate northern Minnesota and south of the Thunder Bay lowland may not have been deglaciated until ~10.4 ka $^{14}$C yrs BP. Additional detailed mapping of glacial features and a detailed chronology (radiocarbon and varve) within northern Minnesota and northwest Ontario will resolve these questions.

6.2 Glacial Lake Kaministiquia

Glacial Lake Kaministiquia formed between the Marks and Dog Lake Moraines (Figs. 2 and 15) and stood relatively high (469 m [1540 ft] from Zoltai (1963); 457 m [1500 ft] from Phillips and Fralick (1994)) in altitude because ice blocked a lower outlet. Cores from Mokomon Lake (Fig. 6) are here ascribed to being deposited in Lake Kaministiquia. Although Burwasser and Ferguson (1980) map Mokomon Lake as being inside the Marks Moraine, it is likely that Mokomon Lake was a embayment of Lake Kaministiquia, as a low (<4 m) bedrock sill seperates Mokomon Lake from the main Lake Kaministiquia basin. The varve record recovered and radiocarbon dates (Fig. 6 and Table 2) indicate Lake Kaministiquia was in existence for at least 160 years prior to ~9.5 ka $^{14}$C yrs BP (Fig. 15). As ice retreated from the north edge of the Marks Moraine, Lake Kaministiquia was retained at a high level (469 m [1540 ft] from Zoltai (1963); 457 m [1500 ft] from Phillips and Fralick (1994)) until ice retreated far enough east to allow drainage through the bedrock gap now occupied by the Kamininstiquia River near the town of Kaministiquia north of Kakabeka Falls (Fig. 2) (Burwasser, 1977; Phillips and Fralick, 1994). The final drainage of Lake Kaministiquia southward into the Thunder
Bay lowland occurred between ~9.5 and ~9.4 ka $^{14}$C yrs BP based on the reconstructions presented here. The details of the drainage, likely catastrophic, are not known and no direct field evidence of this event was found during this work. Teller and Thorleifson (1983) postulated that Lake Kaministiquia drained westward (with a sill at the sub-continental drainage divide) into the Lake Agassiz basin based in the presence of ~25 red laminae within the normally sequence of gray clays found near Dryden, Ontario (Warman, 1991; Minning et al., 1994; Sharpe et al., 1992). Based on topography, the western outlet of Lake Kaministiquia proposed by Teller and Thorleifson (1983) is likely correct, although it is unclear why only 25 red laminae are found in the Lake Agassiz basin when data from this study shows that Lake Kaministiquia was in existence for at least 160 years, presumably with an outlet to the west. If the red laminae from the Lake Agassiz basin do indicate flow from Lake Kaministiquia, then this ‘marker bed’ may be younger (~9.7 to ~9.5 ka $^{14}$C yrs BP) than previous reconstructions (~10 ka $^{14}$C yrs BP).

6.3 Deglaciation of Thunder Bay lowland

Subsequent to formation of the Marks Moraine, the Superior Lobe retreated eastward out of the Thunder Bay lowland. A proglacial lake (here named glacial Lake Kakabeka) fronted the ice margin as it retreated, evidenced by unit D at the Flint Cut (highest sediment elevation: 834 feet (254.5 m) (Fig. 10 and Fig. 15). Lake Kaministiquia likely drained into this lake (Lake Kakabeka) after ~9.5 ka $^{14}$C yrs BP. Farrand (1960), Burwasser (1977), and more recently, Phillips and Fralick (1994) have correlated shorelines within the Thunder Bay lowland to those initially described by Farrand (1960) along the north shore of Minnesota (e.g., Manitou and Beaver Bay).
introduce Lake Kakabeka only because correlation to lake levels (Manitou, Beaver Bay, or another isolated separate lake) outside or inside the Thunder Bay lowland is tenuous with the data obtained during this study. The Flint Cut (Fig. 10) is the only described exposure where glaciolacustrine sediments (Lake Kakabeka) overlie brown-red Superior Lobe till. Therefore, the paleogeographic reconstructions (Fig. 15) of the Thunder Bay lowland only reflect the exposures described in this study and correlation to other published lake levels would be tenuous.

Wood dates of 9380±150 $^{14}$C yrs BP [GSC-287] (Zoltai, 1965) from the Rosslyn site and 9260±170 $^{14}$C yrs BP [TO-547] (Julig et al., 1990) from the Cummins site further to the east were both obtained from sediments associated with glacial Lake Minong, which developed as the LIS retreated from the Superior basin. These dates very close to those obtained from Echo Lake and Mokomon Lake. The retreat from the Marks Moraine may have been rapid due to calving into a deepening lake (glacial Lake Kakabeka) although the Superior Lobe paused as it retreated eastward, forming the Intola Moraine (Burwasser, 1977).

6.4 Ice margin fluctuations

Rapid and lengthy ice margin fluctuations of the LIS during deglaciation are often employed due to constraining radiocarbon chronology and till stratigraphy (e.g. Maher and Mickelson (1996); Clayton et al. (1985); Fenton et al. (1983)). Additionally, low ice-surface profiles of lobes of the LIS have been inferred from the slope of moraines (Matthews, 1974; Clark, 1992; Wright, 1973). Such low ice-surface profiles led to the thinking that the LIS supported ice streams and surging was a major mechanism of
glacier motion (Clayton et al., 1985). Wright (1973) measured very low slopes of lateral moraines on the northwestern edge of the Superior Lobe and used surging to explain low ice-surface profiles and an apparent unidirectional vegetation change from pollen in Minnesota (i.e., no vegetation changes during large fluctuations in climate associated with ice advance). However, Lowell et al. (1999) concluded that surging is unlikely to have caused a nearly 1,000 km long ice margin (along the southern shore of Lake Superior and eastward into Ontario) when advancing both in a terrestrial and lacustrine setting.

Numerous studies have investigated the strong effects of calving and moraine shoal formation for ice advance into water (Alley, 1991; Trabant et al., 2003; Post, 1975; Oerlemans and Nick, 2006). Also well established is the positive linear relationship between water depth and calving (Brown et al., 1982; Warren et al., 1995; Skvarca et al., 2002; Warren and Kirkbride, 2003). During advance, moraine shoal development reduces water depth and consequently reduces calving, but is much slower than retreat due to calving because of the time necessary to deposit and continuously ‘recycle’ the moraine shoal as ice advances, sometimes into deep water (>100 m). Few studies have been able to calculate absolute advance rates, as high-resolution dating, continuous remote sensing data, and/or continuous observations are needed. However, records of historic and Holocene glacier fluctuations of the tidewater Hubbard Glacier in Alaska show that average advance rates were less than 50 m/a (average: ~30 m/a) into as much as ~300 m of water (Trabant et al., 2003; Barclay et al., 2001). Model results from Cutler et al. (2001) indicate the Great Lakes, especially Lake Superior, had a large impact on Wisconsinan glaciation due to calving losses during ice advance. The depth of Lake
Superior (maximum depth of >400 m) inhibited ice advance greatly in model runs, as the Langlade Lobe (which traversed the deepest sections of Lake Superior) took nearly 20 ka (45 ka to 25 ka BP) to cross the lake with calving and moraine shoal formation (120 m in height) included as parameters. Results also indicate that ice advance across Lake Superior was sediment flux limited with respect to development of a moraine shoal (i.e., moraine shoal may not have been subaerial). If a greater sediment flux is used, model runs indicate glaciation further south than actual maximum ice extent (LGM). These results are important because moraine shoals of most contemporary tidewater glaciers in Alaska are not sediment flux limited (i.e., moraine shoal is subaerial) and advance is less than 50 m/a. Quantitatively, it is not known how much reduced sediment flux to the moraine shoal influences advance rates, but it certainly would mean greater effective water depth at the ice margin (moraine shoal would be thinner), which would promote calving and reduce advance rates (Brown et al., 1982; Post, 1975). A complicating factor in this discussion is the relative rate of calving in fresh versus salt water, with calving in fresh water an order of magnitude less than in salt water for a given water depth (Warren et al., 1995; Warren and Kirkbride, 2003). However, the reduced calving in fresh water may limit sediment supply to the moraine shoal from iceberg dump, making the calving rate difference relationship with moraine shoal development more complicated. Additionally, Cutler et al. (2001) note that lake ice and/or high iceberg density would not have reduced calving and promoted ice advance across the basin because a positive energy balance would melt icebergs and lake ice.

A majority of Lake Superior is >150 m in depth (maximum depth of >400 m) and the Marquette readvance would need to cover >250 km to traverse from the north shore.
to the south shore of Superior (to the Gribben site and sites in western upper Michigan
and northern Wisconsin). Given known historical advance rates of Alaskan tidewater
glaciers (<50 m/a), the Marquette readvance would require 5000 calendar years to cross
the Superior basin from north to south. Paleogeographic reconstructions depicting the
Marquette advance have allowed ~800-900 $^{14}$C yrs (10,900/10,800 to 10,025 $^{14}$C yrs BP)
for ice advance across Lake Superior (Clayton, 1983). The concept of a long distance
readvance across the entire Superior basin is rooted in the chronology from the adjacent
Lake Agassiz basin to the west. The timing of the Marquette readvance has been pinned
from radiocarbon dates from the Moorhead low phase (10,960 to 9920 $^{14}$C yrs BP; see
Fisher and Lowell, 2006 for all dates), the duration of which is now being questioned
(Fisher and Lowell, 2006), and the in-situ Gribben site (9545 to 10,300 $^{14}$C yrs BP)
(Hughes and Merry, 1978; Lowell et al., 1999; Pregitzer et al., 2000) and wood within till
in western upper Michigan and northern Wisconsin (9600 to 10,250 $^{14}$C yrs BP) (Lowell
et al., 1999). The drop of Lake Agassiz to its Moorhead low level required that Lake
Agassiz spill into Lake Superior, and hence the Superior basin had to be free of ice. Most
readvances are recognized by a till of differing character lying over previously deposited
glacial sediments (till, glaciofluvial, or glaciolacustrine). However, only one till
(overlying bedrock) has been recognized in central and eastern Lake Superior (Farrand,
1969a; Farrand, 1969b; Dell, 1974; Dell, 1976). In the western Superior basin, till
alternates with lacustrine sediment, and is up to 500 m (1640 ft) thick in a bedrock trough
extending from Thunder Bay to Duluth (Wold et al., 1982; Farrand, 1969a; Farrand,
1969b). Need and Johnson (1984) also describe multiple tills in northern Wisconsin
along the Lake Superior coast, although interbedded lacustrine sediment is absent. Till
stratigraphy indicates that multiple glacier fluctuations occurred in the western Superior basin, but the eastern and central Superior basin may not have experienced glacier fluctuations beyond that of initial mid-Wisconsinan advance into the basin. Additionally, the Cambro-Ordovician escarpment which trends along the south shore of the Superior basin (the north edge of the Michigan structural basin) may have acted as a pinning point for retreating ice.

6.5 Houghton low stage

Some of the current radiocarbon dates constraining the Houghton low may be too old because of the hard-water effect (dates of Saarnisto (1975)), but most dates generally agree with the two charcoal dates from alluvium at the Boyd Cut west of Thunder Bay (8070±70 14C yrs BP [ETH-31437] and 7995±65 [ETH-31438]) (Table 2 and Fig. 13). However, the Houghton level of Farrand (1960) is truncated at Thunder Bay by the younger Nipissing shoreline (i.e., the projected shoreline drops below the level of Lake Nipissing due to differential isostatic rebound of the Superior basin). The elevation of the dated organics is 199 m (653 ft) at the Boyd Cut (Fig. 8 and Table 2), and the trough cross-bedded sand and gravel unit extends from 196 to 199 m (643 to 653 ft), which is below the elevation of the younger Nipissing shoreline (201 m [~660 ft]) in the area (Farrand, 1960). The dated horizon and sand and gravel unit does correspond to the extrapolated elevation of the Houghton level (198 m [~650 ft]). Multiple scenarios are possible: i) the cross-bedded sand and gravel unit (interpreted as fluvial) was deposited during the Houghton low stage and was not eroded away during the subsequent Nipissing transgression, ii) the fluvial unit was deposited as Lake Superior fell from the Nipissing...
highstand to its modern level. In the latter case, the radiocarbon dates from Teller et al. (2005) and this study represent reworked organic matter and do not represent the true age of the deposit. Additional radiocarbon dates and paleoecological analysis (pollen) would improve understanding of the chronology associated with the Boyd Cut and Houghton low phase.

6.6  *Northwest Ontario deglaciation*

The results from this study indicate deglaciation of the Thunder Bay lowland later than previously hypothesized (Teller and Thorleifson, 1983; Teller, 1985; Phillips and Fralick, 1994). This data puts into question the notion that the Rainy and Superior Lobes retreated to the north end of the Superior basin prior to or near the beginning of the Younger Dryas in order to allow Lake Agassiz to drain into upper Great Lakes (Clayton and Moran, 1982; Teller and Thorleifson, 1983).
Chapter 7: Conclusions

Glacial stratigraphy from outcrop and lacustrine cores from the western Thunder Bay lowland, northwest Ontario, indicate glaciolacustrine deposition prior to and after a Superior Lobe advance (Marquette advance). Rhythmites, interpreted as varves based on particle size analysis, magnetic susceptibility, loss-on-ignition, and intervals of bioturbation, are present below (glacial Lake O’Connor) and above (glacial Lake Kaministiquia) Superior Lobe till. In Lake O’Connor, ~300 years of glaciolacustrine sedimentation is recorded prior to Superior Lobe advance into the western Thunder Bay lowland and at a minimum, 160 years of glaciolacustrine sedimentation occurred in Lake Kaministiquia when the Superior Lobe was near its maximum extent at the Marks Moraine. Glaciolacustrine sediments below Superior Lobe diamicton (glacial Lake O’Connor) coarsen upward and indicate a slowly advancing Superior Lobe. This varve chronology, coupled with new deglacial radiocarbon dates from northwest Ontario (Lowell et al., 2005; Fisher et al., in prep.), indicate: i) the Rainy Lobe stood at the Brule Creek Moraine at ~10,200 $^{14}$C yrs BP fronted by glacial Lake O’Connor in the western Thunder Bay lowland; ii) the Superior Lobe advanced (the Marquette advance) westward into Lake O’Connor between ~9.9 and ~9.7 ka $^{14}$C yrs BP; iii) glacial Lake Kaministiquia occupied the reentrant between the Marks and Dog Lake Moraines between ~9.7 and ~9.5 ka $^{14}$C yrs BP; and iv) the Superior Lobe retreated from the Marks Moraine after ~9.5 ka $^{14}$C yrs BP, fronted by a glacial lake (glacial Lake Kakabeka). This deglacial chronology is younger than previous reconstructions (Dyke et al., 2003) and indicates the possible hydrologic connection between glacial Lake Agassiz and the Superior basin.
during the Younger Dryas cold period (11 – 10 ka $^{14}$C yrs BP) may have occurred later than expected or did not occur at all. Additionally, this constraining ice margin chronology may indicate ice remained in the Superior basin prior to the Younger Dryas and that the Marquette advance may have been a short readvance.
References


deglaciation events in the Green Bay Lobe, Wisconsin. *Quaternary Research* **46**, 251-
259.


geology and drift composition, Lake of the Woods region, northwestern Ontario.

318-321.

Wisconsin’s Lake Superior shoreline—Wisconsin Point to Bark Point. *Geoscience


development and sedimentation in the Lake Superior region, North America.

Ontario Geological Survey (1993). Bedrock geology, seamless coverage of the province

Peterson, W.L. (1986). Late Wisconsinan glacial history of northeastern Wisconsin and

morphology of perched glaciolacustrine deltas on the flanks of the Lake Superior

character of the region between the Agassiz and Superior basins, associated with the
‘Interlakes composite’ of Minnesota and Ontario. In “The Late Palaeo-Indian Great
Lakes: Geological and Archaeological investigations of late Pleistocene and early
Holocene environments” (L.J. Jackson, and A. Hinshelwood, Eds.), 275-301.


Appendix A: Aerial Photographs

Aerial photograph of location of Boyd Cut. Image obtained from Google Earth.
Aerial photograph showing location of Flint and Harstone Cuts.
Aerial photograph showing location of Hymers Cut 1 and Hymers Cut 2.
Aerial photograph showing location of Whitefish 1, Whitefish Adjacent 1, and Whitefish Adjacent 2 Cuts
Appendix B: Exposure stratigraphic logs

Boyd Cut (48° 20.43’ N, 89° 21.53’ W)

- Interlaminated 10R 3/4 (dusky red) clayey silt (avg. d(0.5): 13μm [n=2]) and 7.5YR 4/2 (brown) coarse silt- v. fine sand
- 8070 ± 70 14C yrs BP
- 7995 ± 65 14C yrs BP

Fe-stained trough cross-bedded sand and gravel, west to east transport direction

- Type B climbing ripples
- N 40° W
- Rhythmic sedimentation (30-40 cm thick couplets)
- 10R 4/4 (weak red) more pronounced from ~5 m to 10 m
- Interbedded and interlaminated 2.5Y 4/1 (dark gray) to 2.5Y 7/2 (light gray) clayey silt (d(0.5): 22μm) and 10R 4/4 (weak red) to 2.5Y 6/1 (gray) rippled v. fine sand

- Starved ripples
- N 68° E
- N 82° E
- N 90° E
- N 31° E
- n=6 (mean)

- 2.5Y 6/1 (gray) to 10R 4/4 (weak red) rippled v. fine sand (d(0.5): 77μm), abundant mica (muscovite and biotite)
Harstone Cut \(48^\circ 21'31"\ N, 89^\circ 38'.03"\ W\)

2.5yr 4/1 (dark reddish grey) silty diamicton, avg. d(0.5): 47μm \(n = 6\)

\[ S \] (Harstone D-1 and Harstone D-2)

rafts of grey and reddish brown/grey sandy diamicton/lacustrine sediment in lower 50 cm

SY 6/1 (grey) deformed silty/sandy matrix
blocks of silty/clayey reddish brown/grey diamicton and/or lacustrine sediment

Six 40 cm thick rhythmites (30-35 cm thick sandy silt/silty sand capped by 5 cm thick clayey silt)
clayey silt capping each rhythmite (both 10YR 3/1 (very dark grey) and 2.5YR 4/4 (reddish brown))
climbing ripples

\[ S \] (Harstone A2-1 and Harstone A2-2)

SY 6/1 silt unit thickness increases upward in section (up to 5 cm thick at 15 meters - total rhythrite thickness: 15 cm)

rippled SY 5/2 sandy silt stringers

SY 6/1 silt laminations/beds are heavily bioturbated

interbedded and interlaminated 10YR 3/1 (very dark gray) clayey silt
(avg. d(0.5): 7μm \(n = 14\)), SY 6/1 (gray) silt (avg. d(0.5): 20μm \(n = 6\))
SY 5/2 (olive gray) sandy v. fine silt (avg. d(0.5): 60μm \(n = 2\)), and
2.5YR 4/4 (reddish brown) clay (avg. d(0.5): 2μm \(n = 2\))

SY 6/1 silt laminations/beds are heavily bioturbated

rippled SY 5/2 sandy silt stringers

very dark grey clayey silt bed thickness and occurrence is greater in lower part of section

\[ S \] (Harstone A1-1 and Harstone A1-2)
Flint Cut (48° 21' 01" N, 89° 40' 06" W)

- Brown bed frequency and thickness increases upward
  - 7.5 YR 3/2 (dark brown), 7.5 YR 4/3 (brown), and 7.5 YR 2.5/1 (black) silt (avg. d(0.5): 11 µm [n=4]) — clay (avg. d(0.5): 4 µm [n=4]) couplets (rhythmites) — interpreted as varves (approx. 80 years present if use avg. varve thickness of 5 cm)

- 10 YR 4/3 (brown) diamicton, avg. d(0.5): 37 µm [n=2]
  - S (Flint E-1 and Flint E-2)
  - Gradational contact

- 5Y 3/1 (very dark grey) diamicton, avg. d(0.5): 27 µm [n=2]
  - S (Flint D-1 and Flint D-2)
  - Broad folds and microfaults

- Minor sand laminae

- 5Y 5/1 silt with reddish and dark grey laminae, abundant muscovite
  - S (Flint C-1 and Flint C-2)
  - Gradational contact
  - Syndepositional deformation
  - Interlaminated rhythmic silty clay and clayey silt (light gray-reddish dark gray) with dropstones (granules and pebbles) — Sibley Gp. clasts and Shield clasts

- Gradational contact

- Syndepositional deformation

- Irregular lower boundary — silt drapes diamicton
  - S (Flint A-1 and Flint A-2)

- 5Y 2.5/1 (black) bouldery diamicton

- Very well cemented, straited clasts in situ

238 m (781') a.s.l.
0 meters
Hymers Cut 1 (48° 18.21' N, 89° 42.30' W)

15

interbedded and interlaminated silt (d(0.5): 25μm) and fine sand

Fe-stained coarse sand
pebble-cobble lag (erosional contact)

2.5 Y 4/1 (dark grey) silty diamicton - laminated in places, contains clasts of lacustrine red-grey rhythmites
§ (Hymers 1 B-1 and Hymers B-2)

major syndepositional deformation (some laminae are vertical)
reddish laminae present (does not appear below this elevation)
above ~8 m gradual change to GLEY 1 5/N (grey)
horizontal beds/laminations
rhythmic sedimentation (4 cm thick packages [couplets]) - interpreted as varves
synsedimentary deformation
GLEY 1 3/N (very dark grey) and GLEY 1 2.5/N (black) silty clay

rhythmic sedimentation (couplets: 3 cm thick) - interpreted as varves

dropstone (boulder)
interbedded GLEY 1 3/N (very dark grey) and SY 2.5/1 (black) silty clay (avg. d(0.5): 3.5μm [n=5]) with minor silt laminae, muscovite present

rhythmic sedimentation (30-40 cm thick packages)

§ (Hymers 1 A-1 and Hymers 1 A-2)
Hymers Cut 2 (48° 17.84' N, 89° 43.06' W)

- Plane-bedded gravel and sand
- Rip-up clast of glaciolacustrine rhythmites
- Flame structures and small (cm-scale) ball and pillow structures
- Undulating lower contact of coarse silt-v. fine sand units (syndepositional deformation)
- Cross-bedded and cross-laminated coarse silt-v. fine sand (d(0.5): 52μm)
- Draped lamination and Type B climbing ripples
- $ (Hymers2 A1-1)
- Starved sand ripples with 2.5 YR 4/1 (dark reddish grey) and 2.5 YR 4/2 (weak red) clay drapes
- Draped lamination and Type B climbing ripples
- Localized contorted laminae and beds (syndepositional deformation)
- Scattered coarse sand and granules (interpreted as IRD)
- Interbedded and interlaminated GLEY 1 3/N (very dark gray)
  Clayey silt (avg. d(0.5): 9μm [n=2]), 2.5 YR 4/1 (dark reddish gray)
  to 2.5 YR 4/2 (weak red) silt clay, and coarse silt-v. fine sand (dominated by clayey silt)
- Dropstone and localized convolute bedding, no reddish clay drapes below ~4.5 meters
- Scattered coarse sand and granules (interpreted as IRD)
- Localized contorted laminae and beds (syndepositional deformation)
- $ (Hymers2 A2-1)
- River level

Silty clay
Clayey silt
Coarse silt-v. fine sand
Cross-laminations
$ = geochemistry sample
Whitefish 1 (48° 17.28′ N, 89° 45.34′ W)

75

7.5 YR 3/2 (dark brown) diamicton, mottled in some areas
avg. d(0.5): 6μm (n=5)

§ (Whitefish K-1 and Whitefish K-2)

10 YR 3/2 (very dark grayish brown) diamicton, mottling common
avg. d(0.5): 3μm (n=5)

7.5 YR 3/2 (dark brown) diamicton
avg. d(0.5): 3.5μm (n=5)

5 Y 3/1 (very dark gray) diamicton, red siltstone pebble (Sibley Gp.) present, no motting
§ (Whitefish H-1 and Whitefish H-2)
undulating (defomed) contact between silty clay and diamicton

7.5 YR 4/3 (brown) massive silty clay
SYR 4/4 (reddish brown) and 10YR 6/1 (gray) silt-clay couplets (approx. 10-15 cm)
interbedded sand and 10YR 5/6 (yellowish brown) silt with minor clay laminae
cross-bedded pebbly sand
defomed interbedded v. fine sand-silt-clay rhythmites, lower 30 cm contain angular clasts
interbedded laminated silty clay and 5Y 2.5/1 (black) diamicton
§ (Whitefish B-1 and Whitefish B-2)
contact based on presence of clasts

§ (Whitefish A-1 and Whitefish A-2)

avg. d(0.5): 12.5μm (n=3)

5Y 2.5/1 (black) clayey sandy silt, abundant muscovite, no distinct laminations

silty clay
rhythmic silt-clay
sand
cross-bedded sand
diamicton
silt
§ = geochemistry sample

293.5 m (963′) a.s.l. 0 meters
Whitefish Adjacent 1 (48° 17.37' N, 89° 45.68' W)

7.5YR 4/3 (brown) to 7.5YR 3/4 (dark brown) diamicton, numerous Sibley Gp. clasts, columnar structure

avg. d(0.5): 8μm (n=4)

$ (Whitefish1 C-1 and Whitefish1 C-2)

5YR 4/4 (reddish brown) clay, contains pebbles, granules, and coarse sand (IRD), avg. d(0.5): 2.5μm (n=3)

avg. d(0.5): 7μm (n=5)

5Y 3/1 (very dark grey) diamicton, numerous Sibley Gp. clasts, columnar structure

$ (Whitefish1 A-1 and Whitefish1 A-2)
Whitefish Adjacent 2 (48° 17.33' N, 89° 45.58' W)

7.5YR 4/3 (brown) laminated clay, reddish (Sibley Gp.) granules
gradational contact

Sibley Group clasts present

2.5YR 4/1 (dark grey) to 2.5YR 3/1 (very dark grey) diamicton, blocky structure

Sibley Group clasts present

$ (Whitefish2 A2-1)

$ (Whitefish2 C2-1)
deformation at contact

$ (Whitefish2 C1-1)

10YR 4/3 (brown) diamicton, blocky structure

$ = geochemistry sample

319 m (1047') a.s.l.
meters
Appendix C: Exposure geochemistry

**CERTIFICATE RE05099801**

Project: Thunder Bay Geocemetry

P.O. No.:

This report is for 40 Pulp samples submitted to our lab in Reno, NV, USA on 15-NOV-2005.

The following have access to data associated with this certificate:

<table>
<thead>
<tr>
<th>Name</th>
<th>Position</th>
</tr>
</thead>
<tbody>
<tr>
<td>TIMOTHY FISHER</td>
<td></td>
</tr>
<tr>
<td>HENRY LOOPE</td>
<td></td>
</tr>
</tbody>
</table>

**SAMPLE PREPARATION**

<table>
<thead>
<tr>
<th>ALS CODE</th>
<th>DESCRIPTION</th>
</tr>
</thead>
<tbody>
<tr>
<td>WEI-21</td>
<td>Received Sample Weight</td>
</tr>
<tr>
<td>LOG-24</td>
<td>Pulp Login - Rcd w/o Barcode</td>
</tr>
</tbody>
</table>

**ANALYTICAL PROCEDURES**

<table>
<thead>
<tr>
<th>ALS CODE</th>
<th>DESCRIPTION</th>
</tr>
</thead>
<tbody>
<tr>
<td>ME-M061</td>
<td>47 element four acid ICP-MS</td>
</tr>
</tbody>
</table>

This is the Final Report and supersedes any preliminary report with this certificate number. Results apply to samples as submitted. All pages of this report have been checked and approved for release.

Signature: [Signature]

---

To: UNIVERSITY OF TOLEDO
ATTN: HENRY LOOPE
EEES DEPT. MAIL STOP 604
2801 W. BANCROFT ST.
TOLEDO OH 43606

To: UNIVERSITY OF TOLEDO
EEES DEPT. MAIL STOP 604
2801 W. BANCROFT ST.
TOLEDO OH 43606

Finalized Date: 24-NOV-2005
Account: UNITOL
<table>
<thead>
<tr>
<th>Sample Description</th>
<th>Method</th>
<th>Analyte</th>
<th>Units</th>
<th>LOD</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>ME-MS61</td>
<td>ppm</td>
<td>Y</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ME-MS61</td>
<td>ppm</td>
<td>Zn</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ME-MS61</td>
<td>ppm</td>
<td>Zr</td>
</tr>
<tr>
<td>Flint A-1</td>
<td>23.0</td>
<td>196</td>
<td>123.0</td>
<td></td>
</tr>
<tr>
<td>Flint A-2</td>
<td>21.5</td>
<td>196</td>
<td>119.5</td>
<td></td>
</tr>
<tr>
<td>Flint C-1</td>
<td>19.0</td>
<td>66</td>
<td>124.5</td>
<td></td>
</tr>
<tr>
<td>Flint C-2</td>
<td>19.2</td>
<td>69</td>
<td>125.5</td>
<td></td>
</tr>
<tr>
<td>Flint D-1</td>
<td>19.1</td>
<td>118</td>
<td>128.5</td>
<td></td>
</tr>
<tr>
<td>Flint D-2</td>
<td>19.6</td>
<td>126</td>
<td>127.0</td>
<td></td>
</tr>
<tr>
<td>Flint E-1</td>
<td>18.8</td>
<td>125</td>
<td>125.0</td>
<td></td>
</tr>
<tr>
<td>Flint E-2</td>
<td>21.0</td>
<td>136</td>
<td>137.0</td>
<td></td>
</tr>
<tr>
<td>Harstone A-1-1</td>
<td>17.4</td>
<td>116</td>
<td>159.0</td>
<td></td>
</tr>
<tr>
<td>Harstone A-1-2</td>
<td>17.4</td>
<td>121</td>
<td>159.5</td>
<td></td>
</tr>
<tr>
<td>Harstone A2-1</td>
<td>20.1</td>
<td>152</td>
<td>112.5</td>
<td></td>
</tr>
<tr>
<td>Harstone A2-2</td>
<td>20.7</td>
<td>155</td>
<td>114.8</td>
<td></td>
</tr>
<tr>
<td>Harstone D-1</td>
<td>20.4</td>
<td>114</td>
<td>127.0</td>
<td></td>
</tr>
<tr>
<td>Harstone D-2</td>
<td>21.3</td>
<td>120</td>
<td>133.5</td>
<td></td>
</tr>
<tr>
<td>Hymer A-1</td>
<td>25.1</td>
<td>187</td>
<td>120.0</td>
<td></td>
</tr>
<tr>
<td>Hymer A-2</td>
<td>24.5</td>
<td>180</td>
<td>117.5</td>
<td></td>
</tr>
<tr>
<td>Hymer B-1</td>
<td>20.0</td>
<td>131</td>
<td>121.0</td>
<td></td>
</tr>
<tr>
<td>Hymer B-2</td>
<td>20.3</td>
<td>132</td>
<td>122.5</td>
<td></td>
</tr>
<tr>
<td>Hymer A-1-1</td>
<td>17.7</td>
<td>128</td>
<td>124.9</td>
<td></td>
</tr>
<tr>
<td>Hymer A2-1</td>
<td>21.9</td>
<td>269</td>
<td>132.5</td>
<td></td>
</tr>
<tr>
<td>Whitefish A-1</td>
<td>22.3</td>
<td>300</td>
<td>140.0</td>
<td></td>
</tr>
<tr>
<td>Whitefish A-2</td>
<td>21.9</td>
<td>293</td>
<td>137.5</td>
<td></td>
</tr>
<tr>
<td>Whitefish B-1</td>
<td>23.3</td>
<td>189</td>
<td>124.5</td>
<td></td>
</tr>
<tr>
<td>Whitefish B-2</td>
<td>22.7</td>
<td>182</td>
<td>121.5</td>
<td></td>
</tr>
<tr>
<td>Whitefish H-1</td>
<td>21.9</td>
<td>198</td>
<td>138.5</td>
<td></td>
</tr>
<tr>
<td>Whitefish H-2</td>
<td>21.8</td>
<td>198</td>
<td>133.0</td>
<td></td>
</tr>
<tr>
<td>Whitefish K-1</td>
<td>22.6</td>
<td>182</td>
<td>132.5</td>
<td></td>
</tr>
<tr>
<td>Whitefish K-2</td>
<td>22.2</td>
<td>182</td>
<td>136.5</td>
<td></td>
</tr>
<tr>
<td>Whitefish A-1-1</td>
<td>21.7</td>
<td>205</td>
<td>133.0</td>
<td></td>
</tr>
<tr>
<td>Whitefish A-2-1</td>
<td>21.6</td>
<td>209</td>
<td>131.5</td>
<td></td>
</tr>
<tr>
<td>Whitefish C-1</td>
<td>22.3</td>
<td>164</td>
<td>133.5</td>
<td></td>
</tr>
<tr>
<td>Whitefish C-2</td>
<td>23.2</td>
<td>169</td>
<td>144.5</td>
<td></td>
</tr>
<tr>
<td>Whitefish A-1</td>
<td>21.1</td>
<td>159</td>
<td>147.0</td>
<td></td>
</tr>
<tr>
<td>Whitefish A2-1</td>
<td>20.5</td>
<td>164</td>
<td>136.0</td>
<td></td>
</tr>
<tr>
<td>Whitefish C1-1</td>
<td>21.4</td>
<td>170</td>
<td>144.0</td>
<td></td>
</tr>
<tr>
<td>Whitefish C2-1</td>
<td>20.1</td>
<td>118</td>
<td>138.0</td>
<td></td>
</tr>
</tbody>
</table>

Comments: REE's may not be totally soluble in MS61 method.