MIDDLE HOLOCENE ARCHAEOLOGY AND PALEOENVIRONMENTS OF THE THUNDER BAY REGION, LAKE SUPERIOR BASIN

By

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ABSTRACT

This thesis describes the analysis of a deeply buried peat deposit discovered in a cutbank along the Kaministiquia River near Thunder Bay, Ontario. This exposure yielded a large sample of well-preserved plant macrofossils that have been dated between 9,100-8,900 cal yr BP. This is the first study using macrofossil remains to reconstruct vegetation and the environment in the Thunder Bay region. These data contribute to the development of a high-resolution reconstruction of a specific plant community that developed approximately 1,500 years after final deglaciation of the region. The vegetation data is used in conjunction with the sedimentological record of the cutbank to reconstruct paleoenvironmental events that occurred within and around the Lake Superior basin during the early and middle Holocene, providing previously unknown data to the paleoenvironmental record. This reconstruction not only identifies and explains these events but also assigns an environmental context to the archaeological record of the area which is extremely important to understand since there are many gaps and inconsistencies in the record.

ACKNOWLEDGMENTS

During the writing of this thesis, many people have provided me with guidance and encouragement that helped keep me on track. First, and foremost, I would like to thank my advisor Dr. Matthew Boyd (Department of Anthropology, Lakehead University) who showed great patience with me and always made himself available whenever I needed assistance and guidance. I would also like to thank SSHRC (Social Sciences Humanities Research Council of Canada) for providing funding for this research. I would like to thank Dr. Scott Hamilton (Department of Anthropology, Lakehead University) who put up with my incessant questions and comments regarding the archaeology of the area and clarified the cultural history for me; Dr. Kamil Zaniewski (Department of Geography, Lakehead University) for his input, advice, and final editing of my thesis; Erika North (Curator of the Claude E. Garton Herbarium, Lakehead University) for lending me comparative plant macrobotanicals that aided in my identification of my ancient macrobotanical remains; Robert Von Bitter (Archaeological Data Co-Ordinator, Ministry of Culture) for providing me with archaeological site records that were needed to look at Plano site distribution patterns, and faculty and fellow graduate students in the Northern Cultures and Environment program. Finally I would like to thank my family and friends for their continued support, encouragement and even their criticism throughout this long process. Specifically, I would like to thank my sister Angie who called me whenever I needed to talk or calmed me down whenever my emotions got the best of me. I would also like to thank Sylvie Larochelle-Bouchard and Eugene Payette whom put up with my constant mood-swings and were always there when I needed a break.

DEDICATION

This thesis is dedicated to my family and friends, who without, this thesis would not have been written.

TABLE OF CONTENTS

APPENDICES

LIST OF FIGURES

LIST OF TABLES

CHAPTER 1

INTRODUCTION

1.1. INTRODUCTION

The purpose of this thesis is to refine and contribute new information to the paleoenvironmental history and archaeological record for the Thunder Bay region. The major goal of this research is to reconstruct early Middle Holocene vegetation in the Lower Kaministiquia River Valley through the analysis of organic remains recovered from a buried forest. Based on the floral and faunal taxa identified from the organic remains, a reconstruction of local paleovegetation will be made. These results will then be used in conjunction with local and regional pollen records, as well as known deglacial and paleohydrological events to create a paleoenvironmental reconstruction of the Lower Kaministiquia River valley. Finally, this reconstruction will be applied to the local archaeological record to determine if an underrepresentation or scarcity of Early Archaic sites during the early Middle Holocene, may be explained by understanding the paleoenvironment.

The paleovegetation record within Northwestern Ontario is incomplete due to poor spatial and temporal resolution because the few existing palynological records have poor dating control and provide only regional-scale vegetation data (Birks 2003). Climatic trends can be seen in changing pollen frequency values, but this approach is constrained by the fact that pollen can travel long distances and can be affected by depositional time lags (Birks 2003). This reduces the temporal precision and areal specificity of the environmental reconstruction. The Hypsithermal, a period of warming and drying conditions that occurred throughout much of the Early and Middle Holocene, has been effectively documented in pollen cores deriving from the Northern Plains (Williams et al.

2010; Williams et al. 2009). This time-transgressive event has also been identified in pollen records throughout much of Northwestern Ontario and Minnesota (Birks 2003; McAndrews 1982) however, it remains unclear how the Hypsithermal directly affected the paleoenvironment of Northwestern Ontario, specifically with regard to vegetation changes and hydrology.

Macrobotanical remains, however, provide vegetation data at a local scale. Macrobotanical remains from Early to Middle Holocene [11,500-8,900 cal (10,000-8,000 ^{14}C) yr BP] deposits in this area are rare. Only one other study involving macrobotanical remains dating to this time has been published for the north shore of Lake Superior (Bajc et al. 2000). Because macrobotanical remains are organic, they breakdown over time, and unless subjected to certain favourable taphonomic conditions following deposition, rarely are preserved (Pearsall 2008). However, if they are recovered, they can often be identified and AMS radiocarbon dated (Pearsall 2008). Finding macrobotanical remains dating to the Early and Middle Holocene, can more precisely date vegetation migrations in Northwestern Ontario, and can result in the production of a more accurate vegetation reconstruction for the area they are recovered from.

Early archaeological survey and artifact analysis within Northwestern Ontario on Paleoindian and Archaic material have provided good insights on the two cultures. However, it is important to find sites dating to these early cultural traditions so information gaps can be filled in, and a proper description of the traditions can be made specifically for Northwestern Ontario, including Thunder Bay. To locate and properly contextualize archaeological sites, knowledge of the paleoenvironment is essential. For example, it is known that water levels within the Lake Superior basin fluctuated throughout the Early and Middle Holocene; some of these lake level fluctuations have been identified in the cutbanks located within the study area. These fluctuations altered the local environment and may have also affected the livelihood of the people occupying the landscape at the time.

1.2. OVERVIEW OF STUDY AREA

1.2.1. Lower Kaministiquia River Valley

Geomorphology

The Lower Kaministiquia River Valley is located within and west of the city of Thunder Bay in Northwestern Ontario (Figure 1.1). The landscape is dominated by Precambrian bedrock outcrops, numerous lakes and glacial deposits including the Marks and Brule Creek moraines, as well as eskers and drumlins (Figure 1.2) (Zoltai 1963). Remnants of glacial Lakes Kaministiquia and O'Connor are also present and include beach ridges and lacustrine sediment (Zoltai 1963). The Nor'wester Mountains offer the greatest relief with general elevations around 488 m above sea level (asl) with the surrounding area gradually decreasing to Lake Superiors' water level of 183 m asl (Zoltai 1963).

The valley is underlain by a large delta that was deposited by the ancestral Kaministiquia River (Farrand 1960; Zoltai 1963). The sediment forming the delta is believed to have originated from eroded sections of the Marks or Brule Creek moraines, eroded beaches from glacial Lake Kaministiquia, sediment derived from Lake Kaministiquia, or from meltwater streams coming off the Laurentide Ice Sheet (LIS) (Zoltai 1963). The finer deltaic sediment is visible within the cutbanks along the Kaministiquia and Whitefish Rivers (Loop 2006; Boyd et al. 2010), with the coarse material which may represent the apex of the Kaministiquia delta, located by Rosslyn Village (Zoltai 1963).

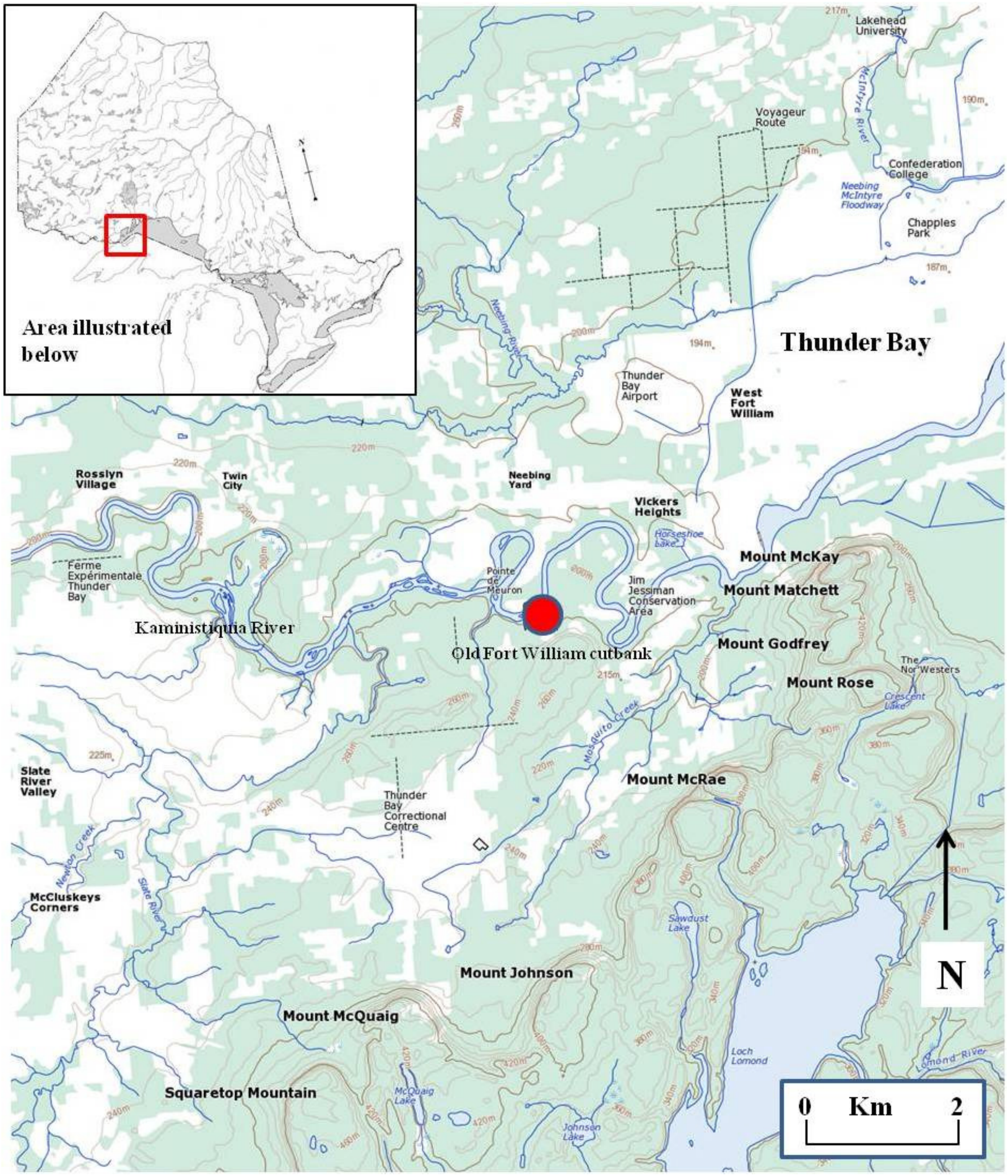


Figure 1.1. Location of the Lower Kaministiquia River Valley and the Old Fort William cutbank near Thunder Bay, Ontario (Based on NTS 52 A/6 (1:80,000 scale).

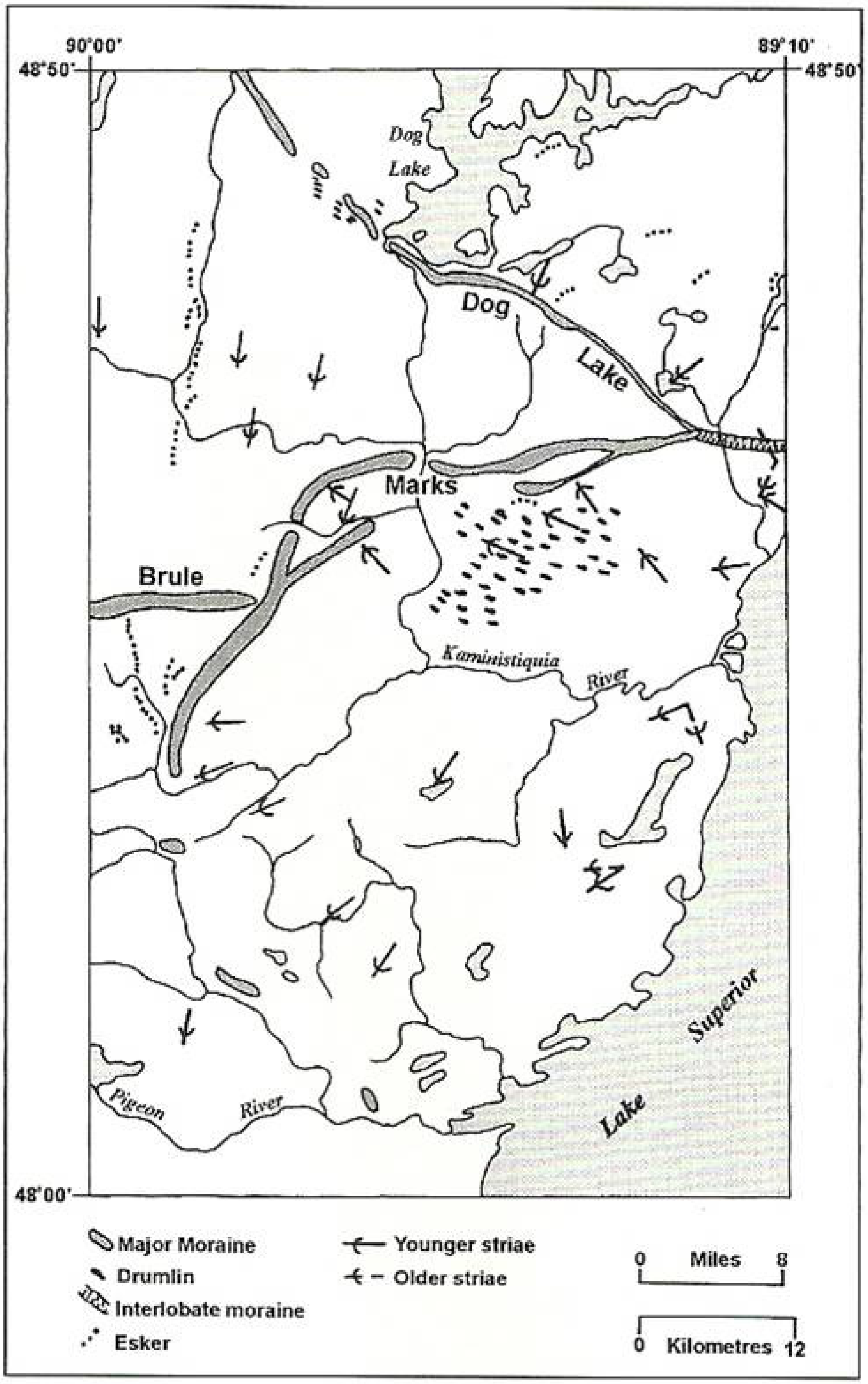


Figure 1.2. Location of glacial deposits within the Lower Kaministiquia River valley (from Zoltai 1963).

Many beach ridges are located within the river valley, most notably the Minong and Nipissing beaches, which represent phases within the Lake Superior basin when relative water levels were much higher than they are today in the Thunder Bay region. The river valley is also rich in archaeological sites, particularly in the vicinity of these relict beach ridges (Figure 2.10). One site (Cummins, DcJi-1), exhibits a long history of occupation dating from the Early Holocene ~10,700 cal (~9,500) yr BP to the Middle Holocene ~7,800 cal (~7,000) yr BP (Julig et al.1990; Dawson 1983a). With the fluctuating water levels, some archaeological sites in the valley may also be buried under sediment as evidenced by the Turning Basin site (DcJh-5) (Dawson 1973).

Vegetation

The Thunder Bay region borders two different forest regions (Figure 1.3). The first is the boreal forest. This forest is dominated by a mixture of coniferous and deciduous trees including black spruce (*Picea mariana*), and white spruce (*Picea glauca*), jack pine (*Pinus banksiana*), tamarack (*Larix laricina*), balsam fir (*Abies balsamea*), white birch (*Betula papyrifera*) and trembling aspen (*Populus tremuloides*) (Kemp 1991; Liu 1990). In the southern portions of this region, including the city of Thunder Bay, lies the Great Lakes-St. Lawrence forest (Figure 1.3) (Kemp 1991). This forest is dominated by deciduous hardwoods including maple (*Acer*), oak (*Quercus*), ash (*Fraxinus*), and elm (*Ulmus*), in addition to conifers such as *Pinus strobus* (white pine) (Kemp 1991). Because Thunder Bay is located at the northern extent of the Great Lakes-St. Lawrence forest region, and south of the boreal forest region, past changes in vegetation may be more easily identified through pollen and macrofossil analysis as this boundary may have shifted throughout the Holocene.

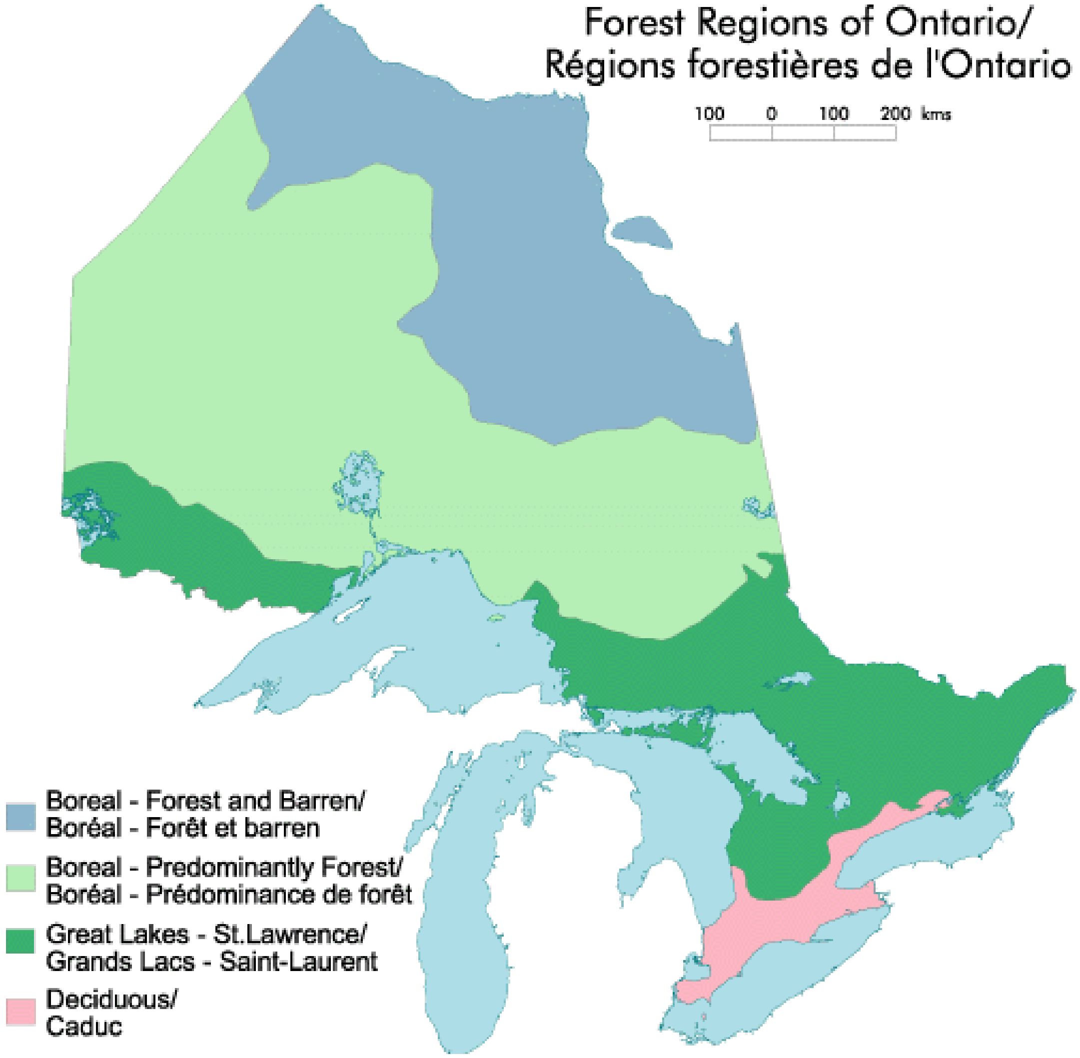


Figure 1.3. Forest regions of Ontario. Note the location of the boundary between Boreal forest and Great Lakes - St. Lawrence forest regions in Northwestern Ontario (from Environment Canada, http://wildspace.ec.gc.ca/maps-e.html 2010).

1.2.2. Old Fort William (OFW) Site

The paleoenvironmental reconstruction presented in this thesis was based on the identification of macrofossil remains recovered from a cutbank along the Kaministiquia River near Old Fort William Historical Park, roughly 10 km upstream from Lake Superior the OFW cutbank (Figure 1.1). Within a 1 km stretch along the river are a series of cutbanks with sediment representing deposition beginning in the Early Holocene (Boyd et al. 2010). Paleochannels, approximately 250 m wide, have been identified within the

Boyd and Old Fort William (OFW) cutbanks (Boyd et al. 2010: Loope 2006). These previously active channels cut into older sediment likely deposited as the LIS left the Marks moraine (Boyd et al. 2010). They were later filled in by silt and clay varves representing high water levels within the Lake Superior basin (Boyd et al. 2010).

The OFW cutbank consists of roughly a 10 m thick sequence of sediment varying from sand to clay (Figure 1.4). The base of the cutbank is approximately 187 m asl, 4 m above the relative lake level of Lake Superior, with the top at approximately 197 m asl. Many paleohydrological events that occurred throughout the Early and Middle Holocene are recorded in the sediment, providing better dating control of the Houghton phase, a period of low water levels within the Lake Superior basin. At the base of the cutbank, within the paleochannel, lies an organic layer representing a buried forest. Organic samples were extracted from this layer and provide the data for the paleovegetation reconstruction for the Lower Kaministiquia River Valley between 9,100-8,900 cal yr BP.

This information will be used to create a paleoenvironmental reconstruction that aims to identify and more accurately date Early and Middle Holocene deglacial and paleohydrological events, as well as the migration of vegetation into Northwestern Ontario. After the paleoenvironment is constructed, it can then be possible to understand how ancient peoples may have used it.

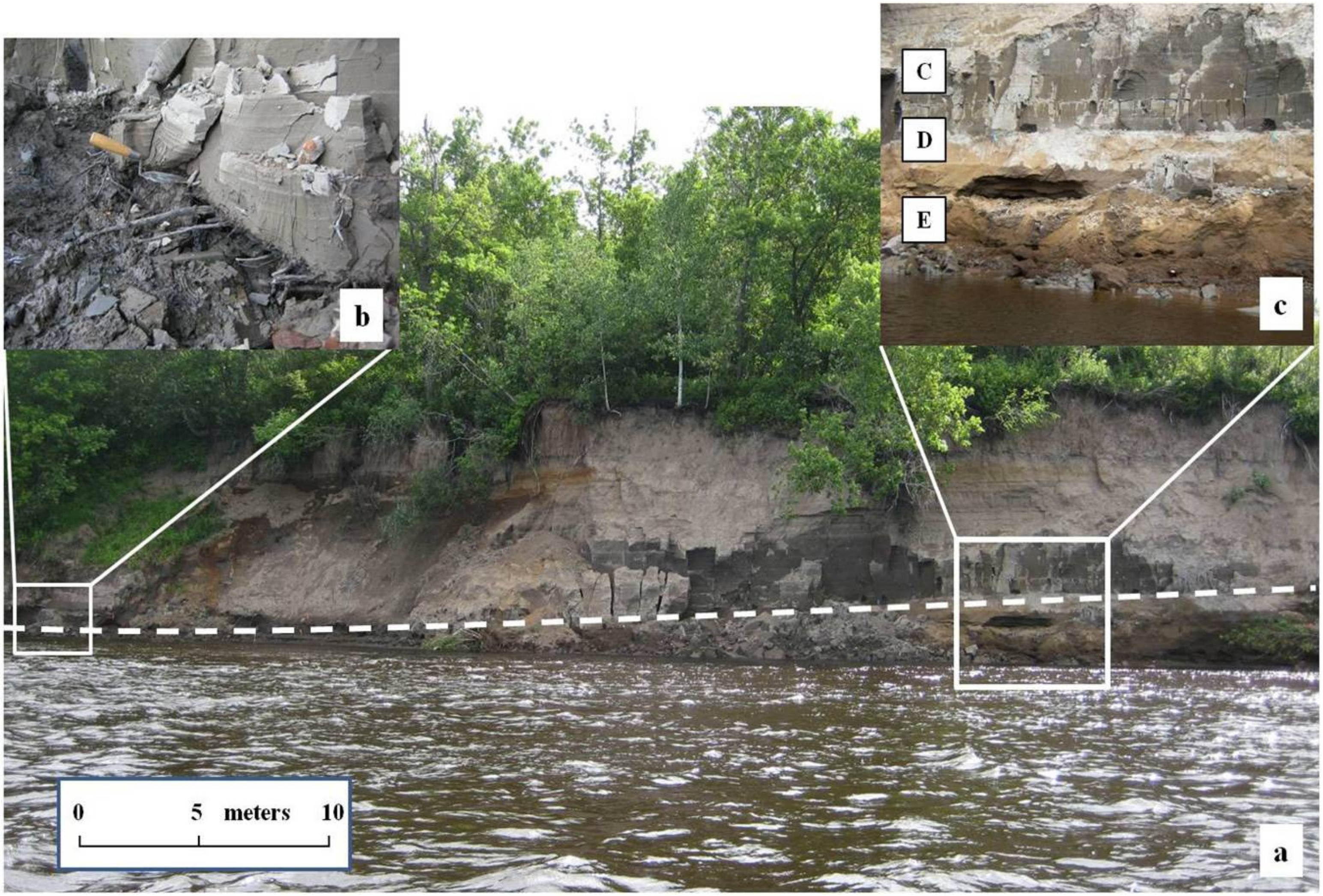


Figure 1.4. Photograph 'a' shows the Old Fort William cutbank. The dashed line indicates the location of Unit D (organic layer), which is enlarged in photo 'b'. Photo 'c' shows Units C, D, and E. Unit E = fluvial sand; D = organic layer/buried forest; C = silt and clay varves.

CHAPTER 2

DEGLACIAL AND PALEOHYDROLOGICAL HISTORY OF THE THUNDER BAY REGION

2.1. INTRODUCTION

The post-glacial hydrological history of the Thunder Bay basin is a complex result of climate change, differential isostatic rebound, outlet switching, outlet blocking, and the influx of glacial meltwater from other basins. By the early Holocene 10,700 cal (9,500 ^{14}C) yr BP, the Laurentide Ice Sheet (LIS) was almost absent from the Lake Superior basin however, it continued to contribute glacial meltwater into the basin until approximately 8,900 cal (8,000 ^{14}C) yr BP (Farrand and Drexler 1985; Saarnisto 1974). Although the history of Lake Superior has been of interest to many researchers over the last several decades (Agassiz 1850; Lawson 1893; Stanley 1932; Farrand 1960; Zoltai 1965; Saarnisto 1974, 1975), many aspects of this history are poorly understood. This is particularly the case with regard to the timing of deglaciation and associated water level fluctuations prior to the Houghton phase ~ 8,900 cal (~8,000 ^{14}C) yr BP.

For the past 10,000 years, Lake Superior’s water levels have fluctuated leaving evidence such as shorelines (e.g., Minong and Nipissing beaches) and remnants of forests buried by lake transgressions. In general, these lake level fluctuations are the result of isostatic rebound, changes in the geometry of the Laurentide Ice Sheet, and meltwater routing from upstream glacial lakes (Farrand and Drexler 1985). Although the general sequence of recent geological events in the Superior basin is understood, many events remain poorly dated. The goal of this chapter is to briefly summarize the history of deglaciation, isostatic rebound, and lake level fluctuations within the Lake Superior basin.

2.2. LAKE SUPERIOR

Lake Superior is the largest of the Great Lakes and is the largest freshwater water body in the world (Cohen 1998). It has a surface area of 82,260 km^2, holds 12,258 km^3 of water and has a mean depth of approximately 150 m (Farrand and Drexler 1985). The current water level is 183 m asl which is controlled by the St. Mary's River outlet at Sault Ste. Marie, Ontario (Farrand and Drexler 1985). The north shore of the basin was the last to be deglaciated and therefore the rate of isostatic rebound is greater there compared to the rest of the basin (Phillips 1993). The northeast section of the basin is currently being uplifted approximately 27 cm/100 years and the southwest (Duluth) basin is subsiding around 21 cm/100 years (Farrand and Drexler 1985).

Radiocarbon dating within and around the basin is problematic because organic remains dating to the Early and Middle Holocene are rare (Saarnisto 1975). Organics are not commonly found within paleoshorelines, resulting in difficulties dating when they were formed (Saarnisto 1975). To fill in voids in the record, correlations are made with the Huron and Michigan basins (Saarnisto 1975) however, as explained later in this chapter, event timelines do not correspond in all basins.

2.3. DEGLACIAL CHRONOLOGY OF NORTHWESTERN ONTARIO

Throughout the Quaternary period, the Lake Superior basin was repeatedly occupied by fluctuations in the Laurentide Ice Sheet (LIS) (Farrand and Drexler 1985), which advanced and retreated in response to climate change. After the post-Valderan advance around 13,700 cal (11,800 ^{14}C) yr BP, the Late Glacial Maximum came to an end and the LIS began a general retreat out of the Lake Superior basin (Figure 2.1) (Farrand and Drexler 1985).

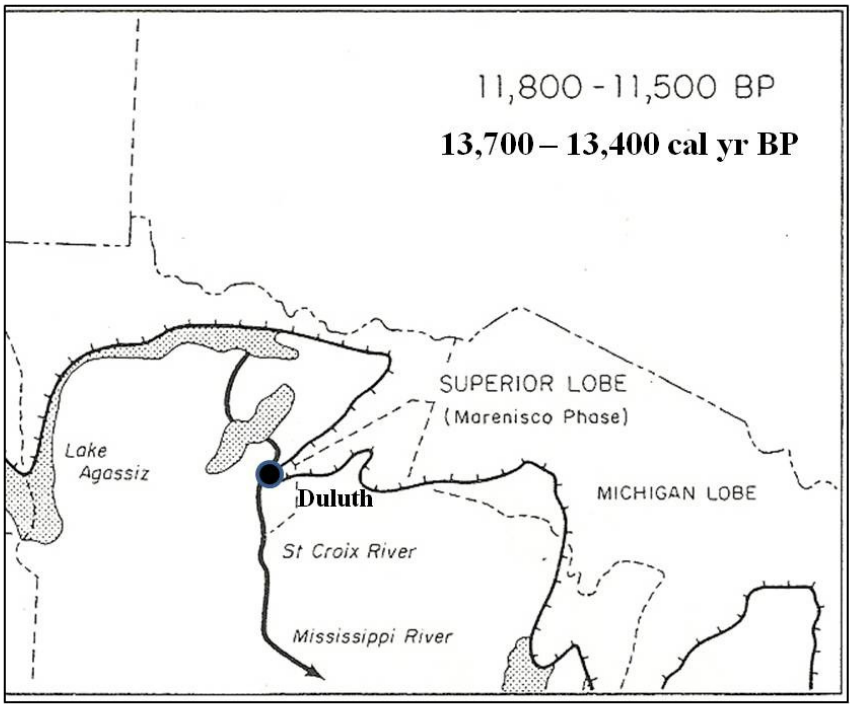


Figure 2.1. Location of the LIS at the end of the Late Glacial Maximum ~13,700 cal yr BP (modified from Farrand and Drexler, 1985).

2.3.1. Pre-Marquette Retreat (13,700-11,500 cal BP)

Dating the LIS retreat in Northwestern Ontario is difficult because there are few radiocarbon dates and limited datable organic material found in sediments from within and around the basin (Breckenridge et al. 2004; Loope 2006). It is unclear how far the LIS retreated into Northwestern Ontario following the Late Glacial Maximum, however, ice position throughout the Early and Middle Holocene is reflected in multiple recessional and end moraines located throughout Northwestern Ontario. Since organic remains are not generally found in these deposits, dating the formation of the moraines has come indirectly through the dating of organic remains obtained from lake cores located near the moraines (Lowell et al. 2009; Björck 1985).

Lowell et al. (2009) cored 26 lakes throughout Northeastern Minnesota and Northwestern Ontario and correlated radiocarbon dates obtained from organic remains within these lakes with other radiocarbon dates produced by lake cores within the area, including Björck (1985), and Teller et al., (2005). These dates were used to identify the sequence of glacial retreat from the study area (Figure 2.2). This study suggests that the Vermillion moraine, located south and west of Thunder Bay, was likely formed around 13,900 cal (12,100 ^{14}C) yr BP, the Eagle-Finlayson moraine around 12,300 cal (11,100 ^{14}C) yr BP, the Brule Creek moraine around 12,100 cal (10,300 ^{14}C) yr BP, the Hartman-Dog Lake moraine between 12,900-11,700 cal (11,000-10,100 ^{14}C) yr BP, and the Lac Seul and Kaiashk moraines likely between 11,500-11,300 cal (10,000-9,900 ^{14}C) yr BP (Lowell et al. 2009). Another prominent moraine, the Marks moraine, located in an arc around the city of Thunder Bay, represents the position of the LIS's Superior Lobe during the Marquette advance and is also dated around 11,500-11,300 cal (˜10,000 ^{14}C) yr BP (Loope 2006). If accurate, some of the moraine dates produced by Lowell et al., (2009), indicate that a large area to the north and west of Thunder Bay was deglaciated prior to the Marquette advance (11,500 cal yr BP).

2.3.2. The Marquette Advance (11,500 cal yr BP)

Prior to 11,500 cal (10,000 ^{14}C) yr BP, a re-advance of the LIS called the Marquette advance (also called the Gribben Interstadial, and Marquette Interstadial) took place (Teller and Thorleifson 1983). The advance extended into northern Minnesota's Duluth area, Wisconsin, and Michigan (Figure 2.3) (Farrand and Drexler 1985).

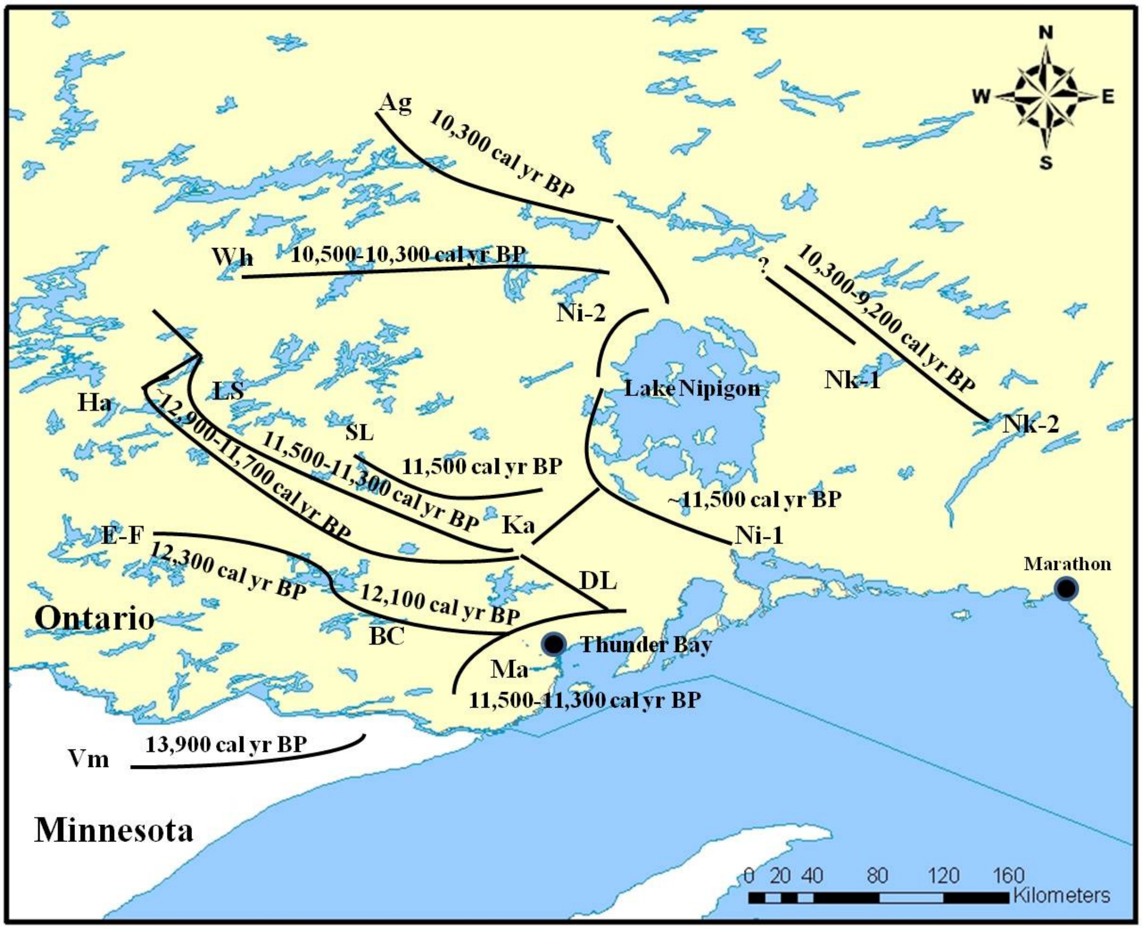


Figure 2.2. Location of recessional and end moraines within Northwestern Ontario. Vm=Vermillion, E-F=Eagle-Finlayson, BC=Brule Creek, Ma=Marks, Ha=Hartman, DL=Dog Lake, LS=Lac Seul, SL=Sioux Lookout, Ka=Kaiashk, Wh=Whitewater, Ag=Agutua, Ni-=Nipigon, Nk-=Nakina (after Lowell et al. 2010 and Björck 1985).

The general consensus is that this advance began prior to 11,500 cal (10,000 ^{14}C) yr BP, and is supported by evidence from the Gribben Lake Forest, a forest which was submerged by ice-marginal flooding and subsequently buried by glaciofluvial sediments in northern Michigan (Lowell et al. 1999). An average date of around 11,500 cal (10,025 ^{14}C) yr BP was obtained from nine wood samples from the buried trees (Lowell et al. 1999). This site also indicates that the ice margin during the maximum of the Marquette Interstadial was located no further south than the Gribben Forest, because the forest was not overrun by the glacier itself (Figure 2.3) (Lowell et al. 1999).

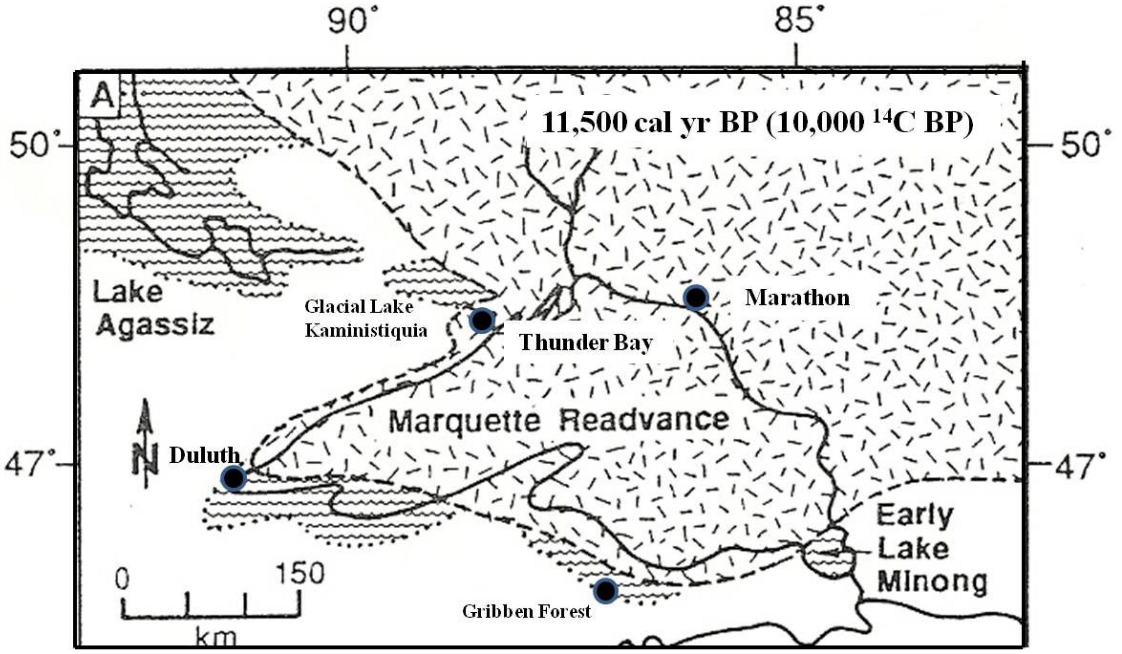


Figure 2.3. Maximum extent of the LIS during the Marquette advance 11,500 cal yr BP (modified from Phillips 1993).

The position of the LIS during the advance has been inferred from ground moraines deposited by the ice front (Farrand and Drexler 1985). The Grand Marais I moraine can be traced from Marquette, Michigan east through Wisconsin and north to near Sault Ste. Marie (Figure 2.4) (Farrand and Drexler 1985). This provides evidence for a widespread advance in the ice sheet and not just a localized advance of a single lobe (Lowell et al., 1999). This moraine is also correlated with the burial of the Gribben Forest by glaciofluvial sediments and therefore was deposited around 11,500 cal (10,025 ^{14}C) yr BP (Farrand and Drexler, 1985; Lowell et al., 1999).

The Marquette advance in Northwestern Ontario is associated with the Marks, Lac Seul and Kaiashk moraines (Figure 2.2) (Lowell et al. 2009). The location of the LIS at the Lac Seul and Kaiashk moraine defines the maximum extent of the ice sheet during the Marquette advance, as the Hartman moraine which is thought to pre-date them, was not over-run by ice (Lowell et al. 2010). The Marks moraine was formed by the Superior lobe as it advanced from the Lake Superior basin (Zoltai 1963). The Dog Lake moraine, which was formed prior to the advance, in combination with the newly formed Marks moraine, created a glacial lake that was dammed between them (Zoltai 1963). This lake, Kaministiquia, initially drained west into Lake Agassiz as the ice blocked lower outlets to the south, depositing distinctive red clays on top of the generally grey Lake Agassiz sediments (Zoltai 1963; Teller and Thorleifson 1983). This lasted for a short period of time, as Lake Kaministiquia subsequently drained south through the ancestral Kaministiquia River shortly after the Superior Lobe left the Marks moraine (Loope 2006).

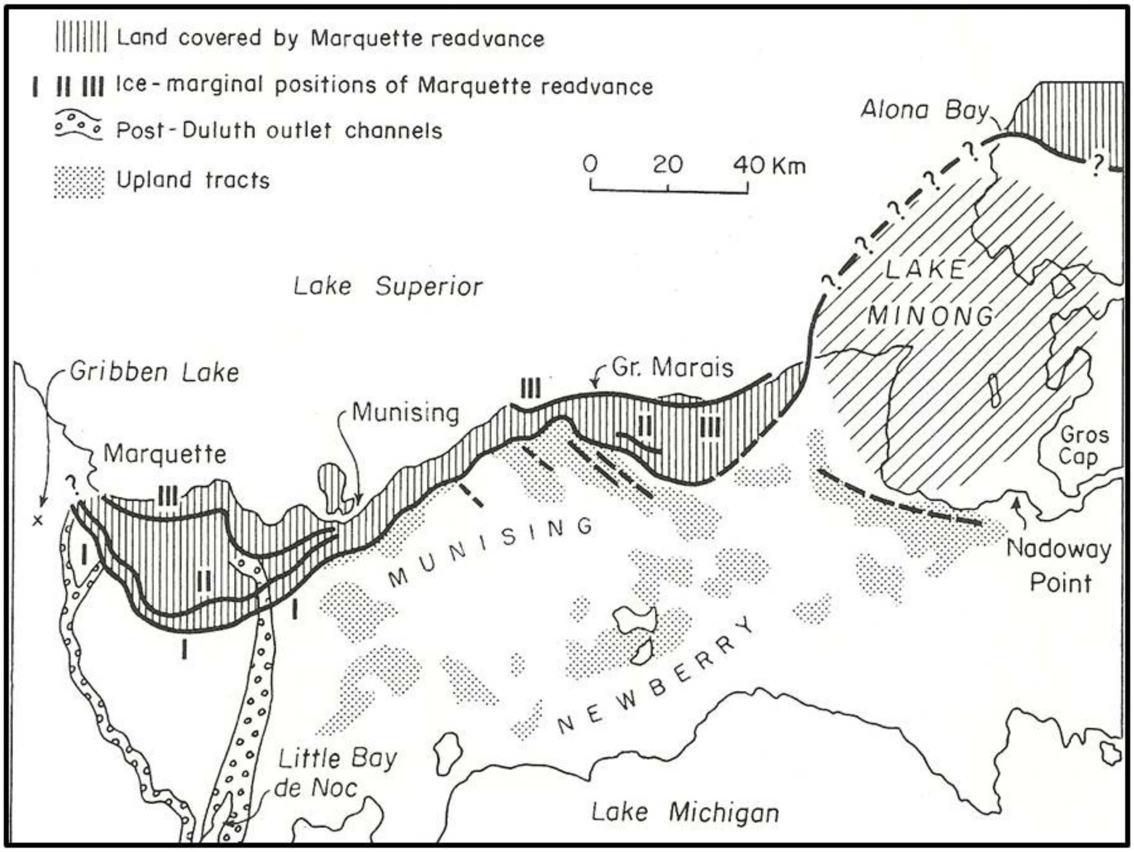


Figure 2.4. Location of the Grand Marais moraine, marking the extent of Marquette ice in Wisconsin, Michigan, and near Sault Ste. Marie, Ontario (from Farrand and Drexler 1985).

It is unknown if Thunder Bay was deglaciated prior to the Marquette advance because moraines or glacial lake shorelines that may have pre-dated it would likely have been destroyed by the readvance (Phillips 1993). Loope (2006) analyzed sediment from various cutbanks along the Whitefish and Kaministiquia Rivers as well as lake cores from Echo and Mokomon lakes west of Thunder Bay. This study revealed that the lowland area west of Thunder Bay may have been deglaciated for approximately 300 years before the Marquette advance and that the Superior Lobe was already present in the area before the advance. This suggests that the advance in the Thunder Bay area may have only been approximately 50 km (Loope 2006). Cutler et al.'s (2001) research on the effects of the Great Lakes on the movement of the LIS suggested that since Lake Superior is rather deep compared to the other Great Lake basins, the LIS would have moved at a slower pace and would have been more prone to calving, hindering its advancement. If this study is accurate, then it would support Loopes' data that the ice advance covered a short distance.

Retreat of the LIS following the advance is believed to have occurred rapidly (Slattery et al. 2007; Loope 2006) since the ice margin is thought to have been east of Lake Nipigon by 10,700-10,000 cal (9,500-9,000 ^{14}C) yr BP (Clayton 1983; Farrand and Drexler 1985). This leaves approximately 500-1500 cal years for the icesheet to retreat from the southern basin to the northern basin of Lake Superior (Farrand and Drexler 1985). Farrand and Drexler (1985), state that ice retreat may have been as fast as 200 m in 100-200 years. This may have been possible since ice wastes faster over water because it is warmer than the frozen land (Cutler et al. 2001). Although it is difficult to determine the actual rate of retreat of Marquette ice, the depth of the basin may have resulted in high rates of glacial calving and coupled with a warming climate, could account for the rapid retreat (Cutler et al. 2001).

2.3.3. Post-Marquette Retreat (11,500-9,000 cal yr BP)

Radiocarbon dates from wood associated with glacial Lake Minong beaches, suggest that the Thunder Bay area was ice-free no later than 10,700 cal (9,500 ^{14}C) yr BP, as glacial Lake Minong formed in the Lake Superior basin along the southern ice margin as the LIS retreated (the discussion of Lake Superior's lake phases will be discussed below in 2.4. Post-glacial paleohydrology of the Lake Superior basin) (Julig et al. 1990). As previously stated, the LIS is thought to have been around Lake Nipigon, specifically the Nakina moraines between 10,700-10,000 cal (9,500-9,000) yr BP (Figure 2.5) (Saarnisto 1975; Slattery et al. 2007), which is supported by a radiocarbon date of approximately 10,200 cal (9,060 ± 150 ^{14}C) yr BP obtained on wood from Jock Lake, west of Marathon, that indicates the area was deglaciated by that time. Slattery et al. (2007), question the absence of the LIS from the moraines at this time because it would mean that all of Lake Agassiz's Nipigon outlets would have been open at the same time. This is in contrast to research that indicates that the Nipigon outlets were opened and closed in succession as the LIS gradually retreated (Slattery et al. 2007). Slattery et al., (2007) then suggest that the LIS likely left the Nakina moraine sometime between 10,200-9,200 cal (9,000-8,200 ^{14}C) yr BP.

Sometime between 9,300-9,100 cal (8,300-8,100 ^{14}C) yr BP, the LIS had retreated out of Northwestern Ontario with water from the now- amalgamated glacial Lakes Agassiz and Barlow-Ojibway draining into the Ottawa River via the Kinojévis outlet (see Figure 2.6 for location of the Kinojévis outlet) (Lewis and Anderson 1989). Small magnitude advances and retreats in the LIS (Cochrane advances) are thought to have occurred between 9,200-8,900 cal (8,200-8,000) yr BP before the LIS melted away completely from this region as the Tyrell Sea formed in the Hudson Bay basin (Prest 1970).

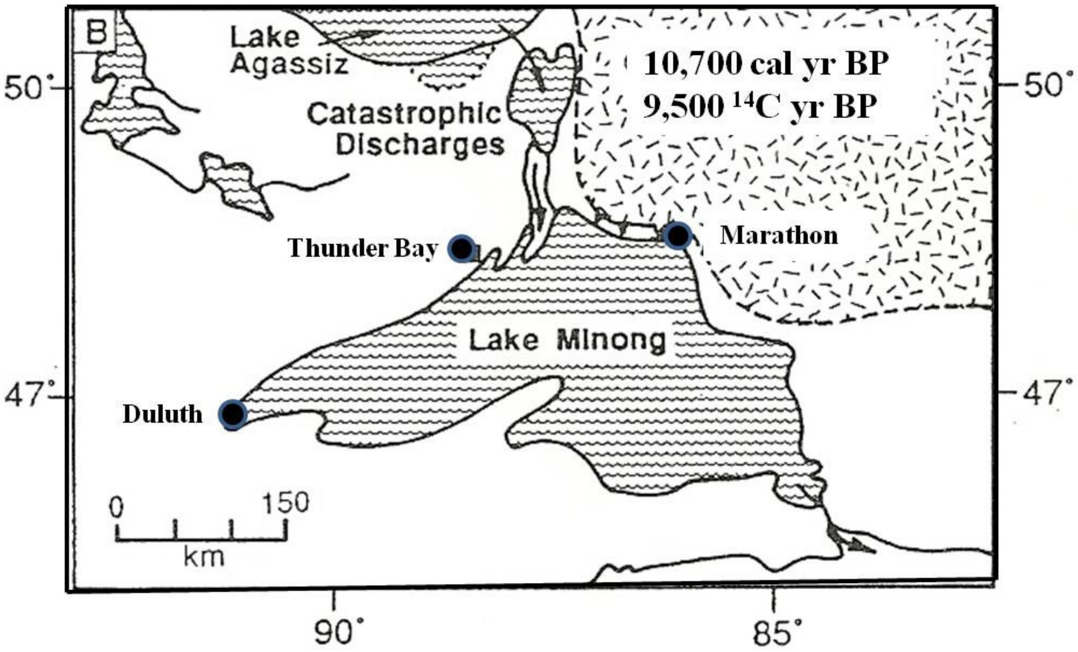


Figure 2.5. Location of the LIS, likely at the Nakina moraines around 10,700 cal yr BP. Glacial Lake Minong encompasses the entire Lake Superior basin and glacial Lake Agassiz discharges through eastern outlets into Lake Nipigon (from Phillips 1993).

2.4. POST-GLACIAL PALEOHYDROLOGY OF THE LAKE SUPERIOR BASIN

2.4.1. Glacial Lake Agassiz and the Agassiz-Superior Connection

Following deglaciation (10,700 cal yr BP), and until about 8,900 cal (8,000 ^{14}C) yr BP, the main contributor of water influx into the Lake Superior basin was from glacial Lake Agassiz (Figure 2.7). Glacial Lake Agassiz formed around 14,000 cal (12,000 ^{14}C) yr BP (Bajc et al. 2000) and had a lifespan of approximately 5,000 calibrated years

(Teller et al. 2005). It covered a vast area including northeastern Saskatchewan, most of Manitoba, and parts of Northwestern Ontario, as well as Northwestern Minnesota and eastern North Dakota (Boyd 2007). During its history it drained through various routes, including south through the Minnesota and Mississippi Rivers to the Gulf of Mexico, northwest through the Clearwater and Athabasca Rivers through the Mackenzie River to the Arctic Ocean, northeast to Lake Ojibway then through the Ottawa River to the north Atlantic Ocean, north into Hudson Bay, and east into the Superior Basin via the Nipigon Basin (Figure 2.6) (Teller et al. 2005).

During the Moorhead phase (12,900-11,300 cal yr BP), Teller and Thorleifson (1983) suggested that meltwater from Lake Agassiz was diverted into the Lake Superior basin either through the Thunder Bay area, or through Lake Nipigon. Recent studies, however, negate these claims. Teller et al. (2005) conclude that the morphology of spillway channels located west of Thunder Bay, do not indicate that catastrophic outbursts, that would have been produced by Lake Agassiz, travelled through them. Large bedload material, generally deposited by catastrophic outbursts, was also not located within these channels, further suggesting that Lake Agassiz did not flow through this area. Lowell et al. (2009) also dismiss eastern drainage of glacial Lake Agassiz during the Younger Dryas because the LIS's Rainy Lobe was positioned at the Brule Creek moraine no later than 12,100 ± 100 cal (10,300 ^{14}C), blocking these outlets. The Brule Creek moraine, which lies on what Teller and Thorleifson (1983) perceive to be a possible outlet, shows no evidence that it was breached or eroded by a spillway channel, meaning Lake Agassiz water likely did not discharge east through Thunder Bay (Lowell et al. 2009).

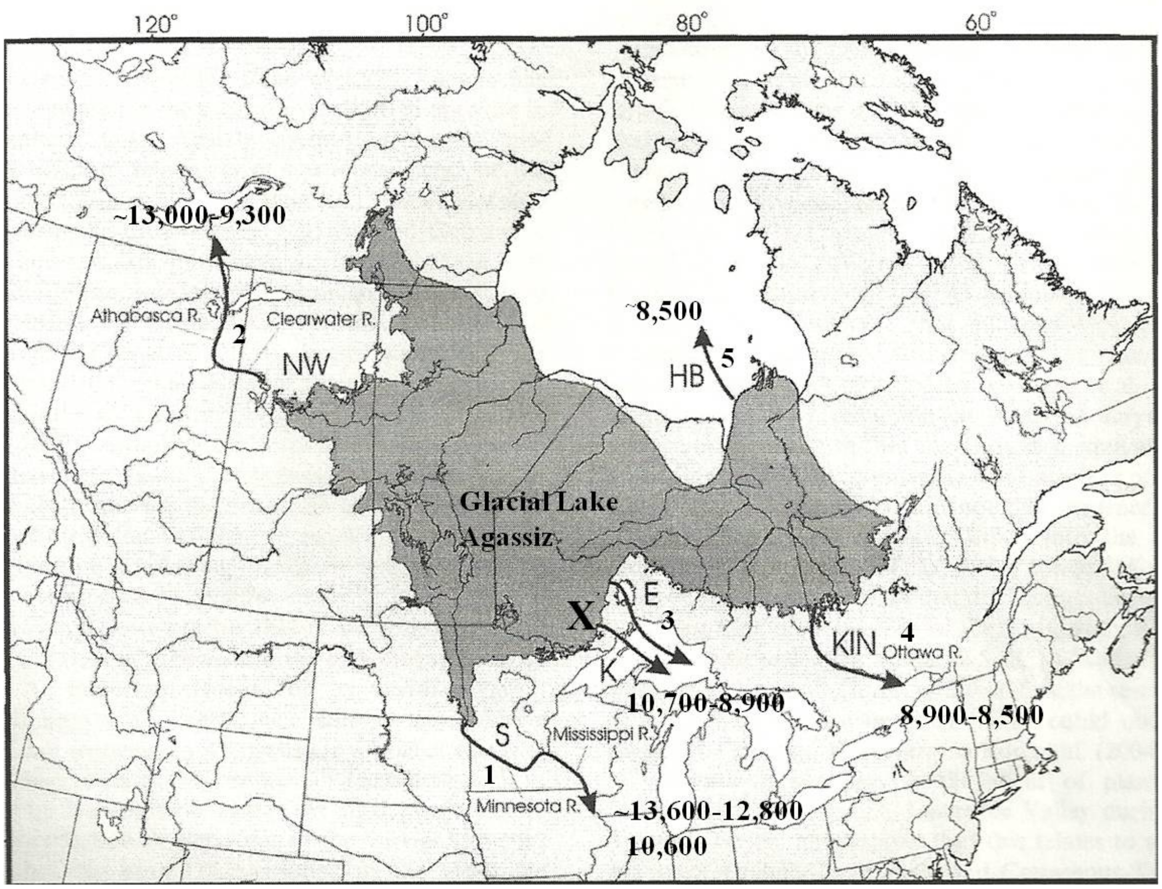


Figure 2.6. The extent of glacial Lake Agassiz and its outlets throughout its existence. All dates have been calibrated. Numbers represent order of spillway opening and utilization. NW= northwest outlet, S= southern outlet, K= eastern outlets through the Thunder Bay area, the X indicates that this spillway was never opened to Lake Agassiz, E= eastern outlets through the Nipigon basin, KIN= Kinojévis outlet, HB= Hudson Bay drainage (modified from Leverington and Teller 2003, additional dates are from Lewis et a*l*. 2010, Murton et al. 2010 and Teller et al. 2005)

Murton et al. (2010) instead, suggest that Lake Agassiz drained to the northwest through the Athabasca-Clearwater outlets to the Arctic Ocean at this time. At the mouth of the Mackenzie River, fluvial and aeolian sand located above coarse gravel thought to represent greater water influx has yielded luminescences dates that indicate possible Lake Agassiz drainage between 13,000 ± 200 and 11,700 ± 100 cal (11,200-10,100 ^{14}C) yr BP (Murton et al. 2010). This information, as well as the identification of a spillway located further upstream from the Mackenzie Delta, suggests that Lake Agassiz may have utilized this outlet during the Moorhead phase (Murton et al. 2010). This may also hold true because there is no other evidence supporting the opening of any other outlet during this time (Murton et al. 2010).

It was not until after final deglaciation of the basin occurred (10,700 cal yr BP) that the eastern outlets into the Lake Nipigon basin were open (Teller and Mahnic 1988). Lake Agassiz likely occupied these eastern outlets from 10,700-9,000 cal (9,500-8,000 ^{14}C) yr BP, as Lake Superior sediments dating between these times were deposited as rhythmites (varves), suggesting steady meltwater influx and sediment deposition (Breckenridge et al. 2004). However, after 9,000 cal yr BP, varve deposition in the basin ceased (Breckenridge et al. 2004). This occurred because Lake Agassiz amalgamated with glacial Lakes Barlow and/or Ojibway as the LIS retreated further out of northern Ontario, resulting in the meltwater bypass of Lake Superior (Breckenridge et al. 2004). Rhythmites in the Nipigon basin, however, continued to be deposited but represent more recent activity, possibly due to the drainage of glacial Lakes Kelvin and Nakina (Breckenridge et al. 2004).

2.4.2. Pre-Minong Lake Phases (11,500-10,700 cal yr BP)

During the retreat of the LIS following the Marquette advance, there were multiple successive lake phases within the Lake Superior basin. These include Duluth, Washburn and Beaver Bay among others (Phillips and Fralick 1994b; Phillips et al. 2001). Glacial Lake Duluth was the first proglacial lake to occupy the southern Superior basin after the initial retreat of Marquette ice around 11,500 cal (10,000 ^{14}C) yr BP (Figure 2.7a) (Farrand 1960; Farrand and Drexler 1985; Phillips 1993; Saarnisto 1974,1975). The LIS continued to block the St. Mary's River outlet, which drains the Lake Superior basin, at this time so Lake Duluth discharged through the Brule River and

Moose Lake outlets in the southwestern part of the basin by Duluth, Minnesota (Farrand 1960). Lake Duluth shorelines are present in the Duluth area between 331-314 m asl and are located above a well defined and much later Nipissing level shoreline (Farrand 1960). According to Farrand (1960), no Lake Duluth shoreline has been identified further north than Grande Portage and possibly Isle Royale. Phillips and Hill (2004) state that if portions of Northwestern Ontario near the international border were deglaciated when Lake Duluth took form, then these shorelines would be found around 442 m asl; however, no shoreline has been identified near Thunder Bay. As Lake Duluth covered the southern basin, early glacial Lake Minong was taking form in the eastern part of the basin, which remained blocked on the west by the retreating icesheet (Figure 2.7a, b) (Farrand and Drexler 1985).

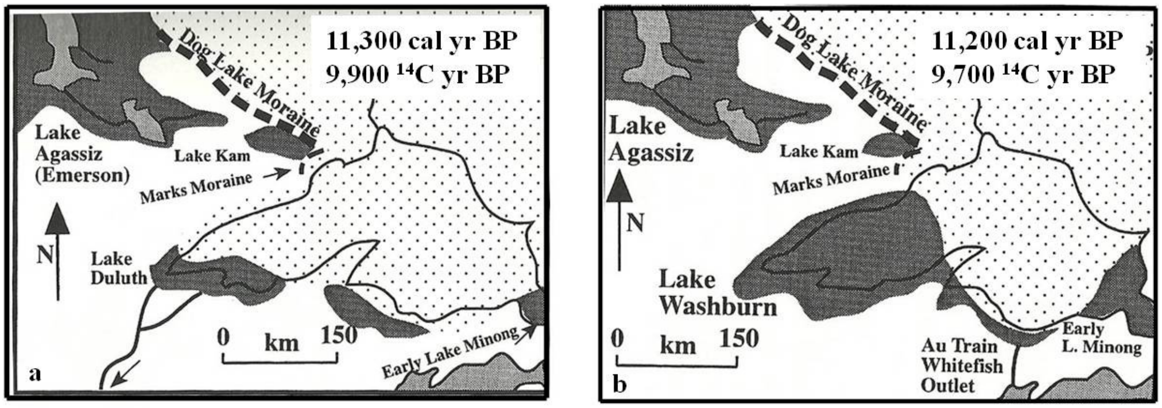


Figure 2.7. The formation of glacial Lakes Duluth (Figure 8a) and Washburn (Figure 8b) in the southern basin following retreat of the LIS after the Marquette advance. The formation of glacial Lake Kaministiquia between the Dog Lake and Marks moraines also occured at this time (from Farrand and Drexler 1985).

Following the Lake Duluth phase was a short phase called Lake Washburn around 11,200 cal (9,700 ^{14}C) yr BP (Figure 2.7b); however, this was quickly succeeded by the Lake Beaver Bay phase at 11,200-11,100 cal (9,700-9,600 ^{14}C) yr BP (Figure 2.8a, b) (Farrand 1960; Farrand and Drexler 1985; Phillips and Fralick 1994b). This lake may have been the first of the glacial lakes to reach Northwestern Ontario (Farrand and Drexler 1985). Evidence of Lake Beaver Bay may be found by Kakabeka where a paleoshoreline has been identified (Phillips and Fralick 1994b; Saarnisto 1974), as well as a Lower Beaver Bay shoreline in the north end of Thunder Bay along the 274 m asl contour (see Figure 2.8b) (Phillips and Fralick 1994b). The problem with identifying these shorelines as Beaver Bay is that there are no dates associated with them and until datable material is found this shoreline should not be accepted as belonging to Lake Beaver Bay.

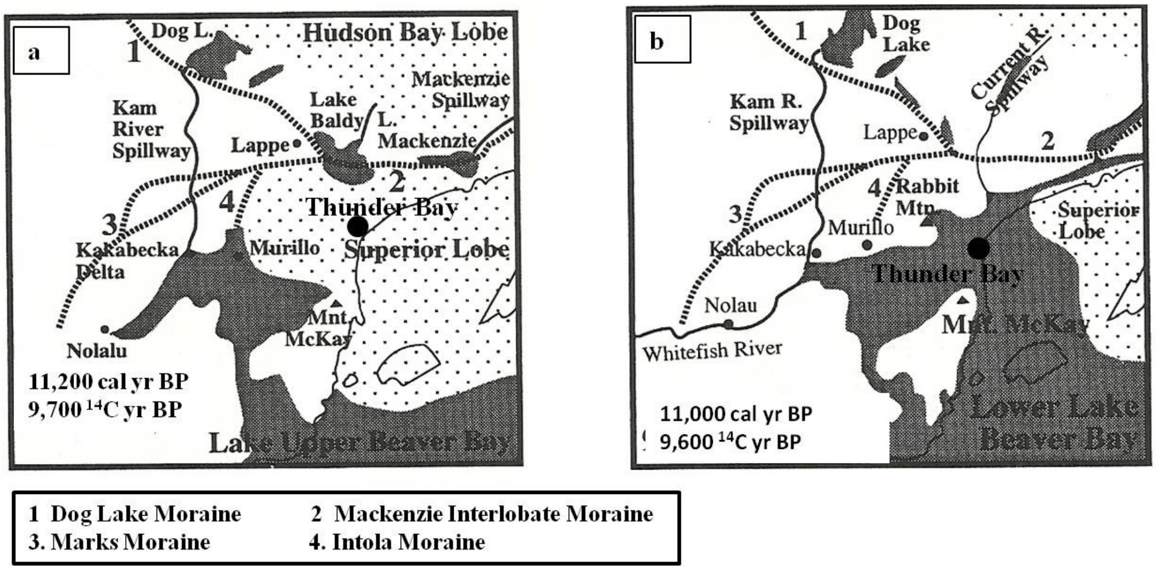


Figure 2.8. Possible location of Lakes Upper and Lower Beaver Bay if they were present in Northwestern Ontario (from Hamilton 1995).

2.4.3. Glacial Lake Minong (~10,700 cal yr BP)

Glacial Lake Minong was named by Stanley (1932) for beaches he identified on Isle Royale, a large island located in Northwestern Lake Superior. It was the first of the proglacial lakes to occupy the entire Lake Superior basin after the LIS retreated; it is associated with the earliest shoreline which there is definitive evidence of in

Northwestern Ontario (Farrand and Drexler 1985; Slattery et al. 2007). The Minong shoreline is approximately 40 m higher than the modern lake level (Booth et al. 2002), although Julig et al. (1990) place the Minong level between 37-57 m higher than the modern lake level due to fluctuations during its occupancy. Prominent shorelines such as the Minong beach where the Cummins site is located, lie between 225 to 240 m asl (Figure 2.9) (Julig et al. 1990). A radiocarbon date obtained from wood taken from the base of a Cummins Pond core, places the formation of the main Minong beach prior to 10,500 cal (9,260 ± 170 ^{14}C) yr BP (Julig et al. 1990). This value coincides with the date of 9,380 ± 150 ^{14}C ka (˜10,700 cal yr BP) from wood taken from the base of a Minong beach near Rosslyn Village (Zoltai 1965) and a date of 9,345 ± 240 ^{14}C ka (˜10,600 cal yr BP) from wood found in a Minong beach in Grand Marais, Minnesota (Drexler et al. 1983).

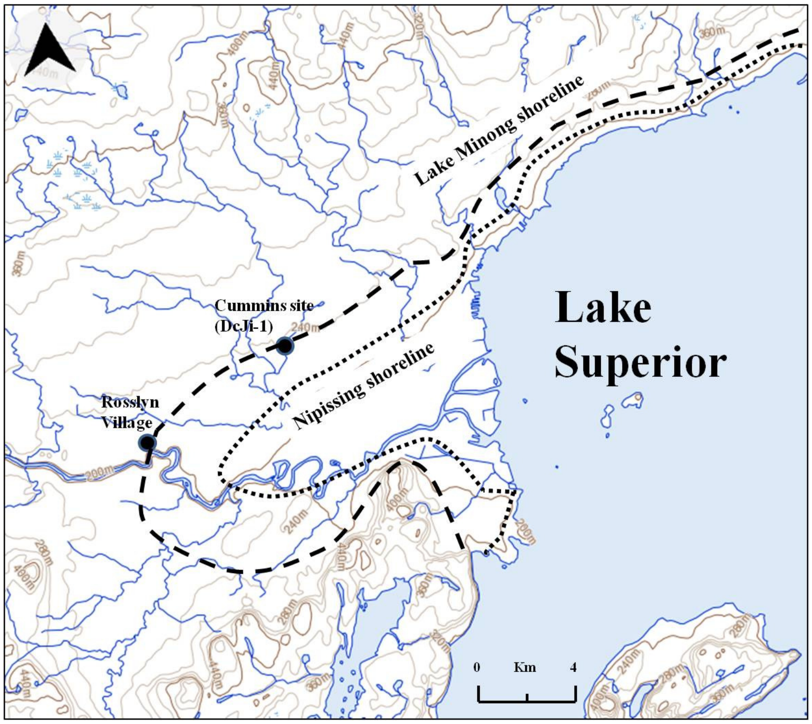


Figure 2.9. Approximate location of the Lake Minong and Nipissing shorelines around the city of Thunder Bay, Ontario. Based on NTS 52 A/6, (1:150,000 scale) (location of shorelines from Zoltai 1963 and Julig et al. 1990).

As previously mentioned, glacial Lake Minong water levels fluctuated significantly. There have been multiple shorelines associated with these various Minong levels but only a few radiocarbon dates are associated with these shorelines. Only one Minong shoreline is located in the Thunder Bay area, and it represents the highest level this lake attained (Farrand 1960); as previously stated, it dates around, or slightly after, 10,700 cal (9,500 ^{14}C) yr BP (Drexler et al. 1983; Julig et al. 1990; Zoltai 1965). Slattery et al. (2007) suggest a date of approximately 10,000 cal (9,000 ^{14}C) yr BP, for the Minong III level based on the deposition of the Nakina Moraine, which marks the location of the ice front at the time. The Minong II level falls between these two dates and elevations (Slattery et al. 2007). In Marathon, Ontario, on the north shore of Lake Superior, Minong III and IV shorelines have been identified (Bajc et al. 1997). A radiocarbon date of approximately 9,100 cal (8,200 ^{14}C) yr BP, was obtained from wood found in a deltaic deposit between these two shorelines, indicating that the Minong III shoreline was formed before this date and Minong IV after (Bajc et al. 1997). Differential isostatic rebound is likely responsible for Thunder Bay having only one Minong shoreline, as the north shore is rebounding faster than the rest of the basin (Farrand 1960; Farrand and Drexler 1985; Phillips 1993).

The high levels during the Minong phase may have been caused by water influx from glacial Lake Agassiz, which perhaps caused hydraulic damming the St. Mary’s River outlet, in combination with the presence of a morainal barrier at Nadoway Point, near Sault Ste. Marie (see Figure 2.10 for the location of the Upper Great Lake’s outlets) (Breckenridge et al. 2010; Farrand and Drexler 1985). During the Minong high phase, both the Michigan and Huron basins were in low stages, named Chippewa and Stanley respectively, which supports the idea that there was a barrier preventing overflow from

Lake Superior into these basins (Farrand and Drexler 1985). This barrier has been identified as the Nadoway Point sill, a moraine that was blocking Lake Minong water from overflowing through the Sault Ste. Marie outlet and into the Lake Huron basin (Farrand and Drexler 1985; Slattery et al. 2007). The cause of the gradual erosion of this sill has been attributed to the catastrophic outbursts from Lake Agassiz through the Nipigon outlets (Farrand and Drexler 1985; Fisher and Whitman 1999).

Figure 2.10. The location of the various outlets controlling water levels within the Upper Great Lake basins (Ontario basemap from Natural Resources Canada, http://atlas.gc.ca 2010; location of outlets after Lewis and Andserson 1989).

Lake Agassiz drained into glacial Lake Minong through multiple spillways between 10,700-8,900 cal (9,500-8,000 ^{14}C) (Teller and Mahnic 1988; Teller and Thorleifson 1983). After 10,700 cal yr BP, spillways were successively opened to the east of glacial Lake Agassiz into the Lake Nipigon basin (Teller and Mahnic 1988; Teller and Thorleifson, 1983). The Nipigon inlets and outlets consisted of five groups of channel complexes and seventeen smaller channels (Figure 2.11) (Farrand and Drexler 1985; Leverington and Teller 2003; Teller 1985; Teller and Mahnic, 1988). The five channel groups are named the Kaiashk, Kopka, Pillar, Armstrong, and Pikitigushi, all of which drained into the northern and western basin of Lake Nipigon (Teller and Mahnic, 1988). From the Nipigon basin the water travelled into Lake Superior via the Wolf, Wolfpup, Shillabeer, Black Sturgeon, Nipigon, Cash, and Pijitiwabik channel systems (Teller and Mahnic, 1988). Teller (1985), Teller and Mahnic (1988), and Teller and Thorleifson (1983) believe that meltwater outbursts from these channels may have been catastrophic, resulting in fluctuating water levels within the Lake Superior basin which Farrand and Drexler (1985) suggest may have been at a magnitude of 30-50 m.

Breckenridge (2007) has dated three possible outbursts based on varve formation within the western Lake Superior basin. His dates of 10,630, 10,600, and 10,570 cal (mean age ˜9,400 ^{14}C) yr BP also correspond to a period of higher water levels in the Lake Huron basin that has been dated between 11,000-10,500 cal (9,600-9,300 ^{14}C) yr BP based on organic remains obtained in lake cores (Lewis and Anderson 1989). Yu et al., (2010) suggest that the Nadoway sill was breached around 10,500 cal (9,300 ^{14}C) yr BP, causing the cold Lake Superior water to overflow through the Sault Ste. Marie outlet, into the Michigan and Huron basins, through the North Bay outlet, into the Ottawa and St. Lawrence Rivers and into the North Atlantic, possibly disrupting thermohaline circulation resulting in a period of climatic cooling. A negative shift in $\delta^{18}O$, indicative of an influx of cold glacial meltwater, is recorded in the Lake Huron basin around 10,500 cal (9,300 ^{14}C) yr BP (Breckenridge and Johnson 2009), which supports Yu et al.'s (2010) hypothesis of Lake Superior overflow at this time. It is, however, unclear if the abrupt breaching of the sill is tied to one of the outburst events identified by Breckenridge (2007), or to an unidentified outburst event.

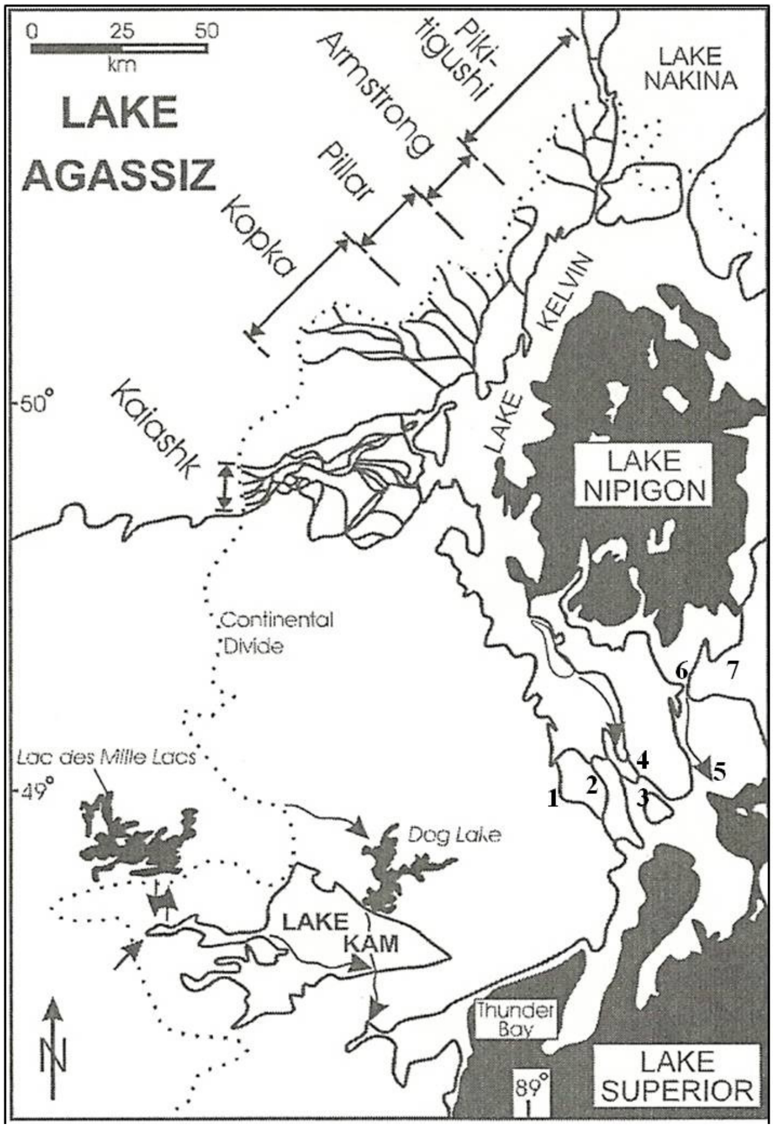


Figure 2.11. The multiple spillways and channel complexes draining glacial Lake Agassiz into the Lake Nipigon basin between 10,700-8,900 cal yr BP. Numbers represent spillways carrying meltwater into the Lake Superior basin. 1= Wolf, 2= Wolfpup, 3= Shillabeer, 4= Black Sturgeon, 5= Nipigon, 6= Cash, 7= Pijitiwabik (from Teller and Mahnic 1988).

A rise in water levels has been identified by Phillips and Fralick (1994a) by the Little Pic River near Marathon. Modification of an ice-contact delta that was likely formed during the highest Minong stage has been attributed to this rise. Phillips and Fralick (1994a) suggest that the water level may have risen approximately 18 m based on known elevations of the highest Minong beach (311 m asl) and stratigraphy of the delta. This transgression likely occurred sometime between 10,700-9,000 cal (9,500-8,100 ^{14}C) yr BP, when Lake Agassiz drained into the Lake Superior basin. However, because the delta was overridden and incised, the transgression likely occurred early in this timeframe (Phillips and Fralick 1994a) and may, in fact, record one of the events identified by Breckenridge (2007). A possible outburst has been recorded in lake cores obtained from Fenton Lake near Marathon (Breckenridge et al. 2010). This transgression took place between 9,500-9,000 cal (8,600-8,100 ^{14}C) yr BP and raised lake levels at least 12 m. The cause of the transgression is unknown, but may have been from an influx of meltwater from the LIS or Lake Agassiz, or it could have been as a result of hydraulic damming of the Lake Minong outlet (Breckenridge et al. 2010). Based on the timing of this transgression, Breckenridge et al., (2010) state that the transgression identified by Phillips and Fralick (1994a) likely occurred before 10,200 cal (9,000 ^{14}C) yr BP.

Additional evidence supporting periodic glacial meltwater outbursts comes from the Lake Huron basin, which experienced a completely different deglacial and paleohydrological history than Lake Superior (Lewis and Anderson 1989; Lewis et al. 2007). Unlike the Lake Superior basin, relative water levels within the Lake Huron basin were low at the start of the Early Holocene, as its highstand ended around 11,500 cal (10,000 ^{14}C BP). The cause of the lowstand, known as Lake Stanley, has been attributed to the opening of the North Bay outlet which drained glacial Lake Algonquin which occupied the basin at the time (Larsen 1985; Lewis et al. 2007). Climate change may have also kept lower water levels low as aridity increased during the Middle Holocene (Williams et al. 2010). After 11,500 cal yr BP, water levels were much lower than the preceding Lake Algonquin levels; however, they continuously fluctuated, with a general low phase separated by higher water phases known as the Mattawa highstands (Lewis et al. 2007). These highstands have been attributed to periods of discharge from glacial Lake Agassiz through the Sault Ste. Marie outlet (Lewis et al. 2007). In addition to the increase of water levels that occurred between 11,000-10,500 cal (9,600-9,300 ^{14}C) yr BP as previously explained, the relative water level remained high until around 9,500 cal (8,400 ^{14}C) yr BP, suggesting frequent overflow of Lake Superior water (Lewis and Anderson 1989). These dates correspond with the timeline of 10,700-8,900 cal (9,500-8,000 ^{14}C) yr BP proposed by Teller and Thorleifson (1983) for the initiation and terminus of Lake Agassiz meltwater influx into the Lake Superior basin, because around 9,500-8,900 cal (8,500-8,000 ^{14}C) yr BP, Lake Agassiz water re-routed along the retreating margin of the LIS and amalgamated with glacial Lake Barlow and/or Ojibway, ceasing all inflow into the Great Lake basins (Figure 2.12) (Teller and Thorleifson 1983).

There is evidence, however, that suggests that meltwater from Lakes Agassiz and/or Barlow-Ojibway continued to periodically enter the Lake Superior basin after it had initially ceased (Boyd et al. 2010). As discussed earlier in this chapter, the Cochrane advances are thought to have blocked the Kinojévis outlet, perhaps multiple times, between 9,200-8,900 cal (8,200-8,000 ^{14}C) yr BP (Prest 1970). As a result, drainage of the glacial lakes may have moved back to Lake Superior (Slattery et al. 2007). Evidence to support this comes from the Longlac area located between Nipigon and Marathon along the north shore of Lake Superior (Slattery et al. 2007). The Mullet outlet was identified by Slattery et al., (2007) as a glacial meltwater route that may have carried water from glacial Lakes Agassiz and/or Barlow-Ojibway into the Pic River and finally to Lake Superior. According to Slattery et al., (2007), this outlet would have been operational following the retreat of the LIS from the Nakina II moraine. Dating the abandonment of this moraine, however, has been problematic because there are no radiocarbon dates associated with it. However, Slattery et al. (2007) suggest that it occurred sometime between 10,200-9,200 cal (9,000-8,200 ^{14}C) yr BP based on pollen records and associated radiocarbon dates within lake cores produced by Saarnisto (1975) for the area, as well as deglaciation dates proposed by Zoltai (1965). Depending on the deglaciation of the Nakina II moraine, some of the transgressions identified may have resulted from meltwater outbursts from this outlet. However, more research needs to be completed to support this. Shortly after Lake Agassiz joined with Barlow-Ojibway (~9,000 cal yr BP), the Minong phase ended, as relative water levels in the basin hit their lowest levels (Figure 2.12) (Farrand 1960).

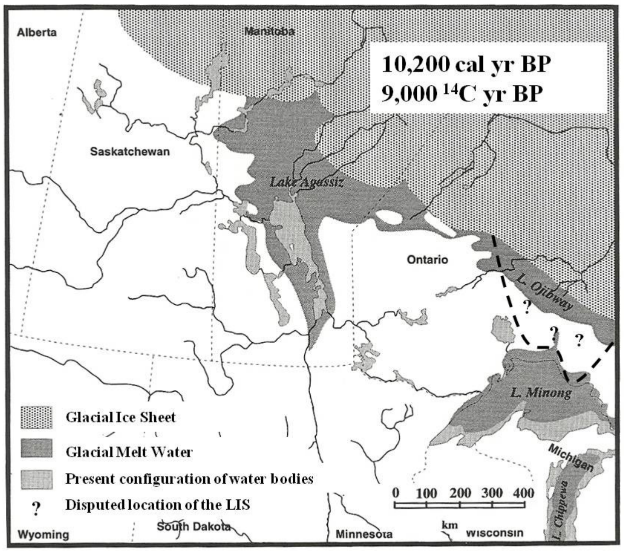


Figure 2.12. The amalgamation of glacial Lake Agaissiz and glacial Lake Ojibway after 10,200 cal yr BP. The dashed line indicates that the exact location of the ice front is not known at this time (after Hamilton 1996).

2.4.4. The Houghton Phase (˜8,900 cal yr BP)

The Houghton phase was a stage of low water levels resulting, in part, from the cessation of influx of Lake Agassiz water into the Lake Superior basin (Farrand and Drexler 1985) as well as perhaps a more arid climate (Figure 2.13) (Boyd et al. 2010). Dating the onset of this period is somewhat difficult as only a few radiocarbon dates exist. However most dates associated with the extremely low water levels within the basin are in agreement. Bulk samples from Crozier and Fenton Lakes on the eastern side of the Superior basin produced dates of ˜8,400 cal (7,590 ± 180 ^{14}C) yr BP and ˜9,000 cal (8,100 ± 180 ^{14}C) yr BP (Saarnisto 1975). Dates obtained from organics at the Boyd cut along the Kaministiquia River place the Houghton phase also around 9,000 cal (8,070 ±70 ^{14}C) yr BP and 8,900 cal (7,995 ± 65 ^{14}C) yr BP, (Loope 2006) and organic material from a deltaic deposit, representing low water levels near Marathon, dates around 9,200 cal (8,200 ^{14}C) yr BP (Bajc et al. 1997). Breckenridge (2007) records the decline of water levels within the western Lake Superior basin around 9,000 cal (8,100 ^{14}C) yr BP, as varve deposition ceases at this time.

As previously mentioned, Lakes Huron and Michigan were already in their low stages prior to the Houghton phase; however, there appears to be similarity between Lake Houghton and the Late Stanley phase of the Huron basin. Although the Stanley Low phase began around 11,500 cal (10,000 ^{14}C) yr BP, there were three separate extreme lows; the last, during the Late Stanley phase, coincides with the Houghton phase around 8,800 cal (7,900 ^{14}C) yr BP) (Lewis et al. 2007). These dates are based on radiocarbon-dated ostracod shells and seismic profiles of sediment cores (Lewis et al. 2007). Late Stanley water levels may have been 50 m below the modern level according to Lewis et al., (2007) however, Hunter et al., (2006) suggest that it was 100 m lower than the modern level. More data is needed to determine how much water levels declined during the Late Stanley lowstand, however it appears that the relative water level was below that of all its outlets, possibly resulting in a closed lake (Lewis et al. 2007). This also may have occurred within the Michigan basin but perhaps later, between 8,400-8,200 cal (7,400-7,580 ^{14}C) yr BP (Lewis et al. 2007).

Determining the relative water level during the Houghton Low has proven to be just as difficult. Fisher and Whitman (1999) suggest that water levels within the Lake Superior basin may have lowered to 170 m asl, meaning that the relative lake level was 13 metres lower than the modern water level of 183 m asl. Yu et al., (2010) suggest a more modest decline of 45 m from 226 m asl to 181 m asl, meaning the relative lake level would have been two meters below the outlet at Sault Ste. Marie. Boyd et al. (2010) also suggest that the Lake Superior basin was closed, at least between 9,100 and 8,900 cal yr BP, based on radiocarbon-dated macrobotanical remains from a deeply buried forest located west of Thunder Bay. This, however, will be discussed at greater length in Chapter 7 (Interpretations and Discussion). Submerged peat from the Apostle Islands, Wisconsin, indicate that the Houghton phase was 16.5 m below the modern sea level as it was located under 4 m of sand and 12 m of water. Radiocarbon dates from organic silt from Portage Lake on the Keweenaw Peninsula indicate that water levels were 60 m lower than present during this period, as this layer was found near the surface of the lake bottom (Farrand 1960). Isostatic rebound likely accounts for the appearance of differences in water level depth as the north shore rebounded at a faster rate, preserving the Houghton shoreline in this area, while they remain submerged along the south shore (Farrand 1960).

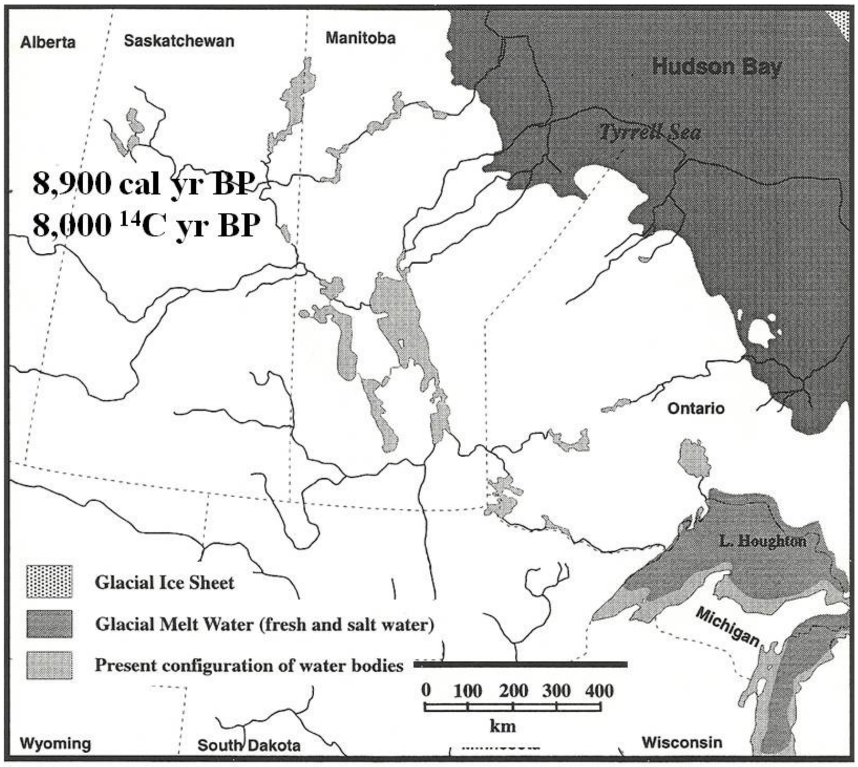


Figure 2.13. Estimation on the configuration of Lake Houghton 8,900 cal yr BP (from Hamilton 1995).

2.4.5. The Nipissing Transgression (6,800-4,500 cal yr BP)

Rebounding of the North Bay outlet caused water levels to rise, resulting in the Nipissing transgression, a high water phase experienced by all of the Upper Great Lakes (Fisher and Whitman 1999; Teller 1985). The North Bay outlet was originally deglaciated around 11,500 cal (10,000 ^{14}C) yr BP, and between this time and 5,700 cal (5,000 ^{14}C) yr BP, it was gradually uplifting due to isostatic rebound (Karrow et al. 2007). This uplift caused a backflooding of water and a rise in water levels within Lakes Huron, Michigan and Superior (Lewis and Anderson 1989). These high water levels persisted because this outlet had not yet rebounded to the height of the Port Huron outlet, where Lake Huron drains into Lake Erie (Farrand and Drexler 1985).

In the literature focusing on the Nipissing transgression, there has not been a consensus on dating the onset and peak of the rising water levels within the Upper Great Lake basins due to differential isostatic rebound (Johnston et al. 2004). Research has divided the Nipissing into two stages, Nipissing I and Nipissing II. Larsen (1985) identifies Nipissing I between 5,500-4,700 cal (4,740-4,185 ^{14}C) yr, where water levels were at approximately 183 m asl, while Booth et al. (2002) place it between 6,800-5,700 cal (6,000-5,000 ^{14}C) yr BP, and about 14 m higher than today near the Sault Ste. Marie outlet. Earlier suggested dates of the onset of the transgression come from Saarnisto (1975), who place it at 7,900 cal (7,000 ^{14}C) yr BP, and by Lewis and Anderson (1989), who suggest between 9,300-9,000 cal (8,300-8,100 ^{14}C) yr BP.

Water levels during the Nipissing II stage increased between 4,700-4,200 cal (4,185-3,800 ^{14}C) yr BP, but did not exceed 178 m asl (Larsen, 1985). These dates come from a marsh deposit in Lake Michigan, and Larsen used them to conclude that the transgression was a relatively short event. The latest date from offshore stumps in Lake Huron is approximately 7,800 cal (7,000 ^{14}C) yr BP and may coincide with the end of the Stanley Low and the beginning of the Nipissing phase, thus allotting a period of 2,000 years for the transgression in Lake Huron.

Johnston et al. (2004) analyzed data from various reports and concluded that although there are a number of competing dates for the Nipissing transgression, the general consensus is that the Nipissing II phase ended by 4,500 cal (4,000 ^{14}C) yr BP, after which water levels within the basins dropped. This drop in water level has been attributed to the erosion of the Port Huron outlet below that of the Chicago outlet along the southern shore of Lake Michigan, resulting in the diversion of water through the Port Huron outlet and the abandonment of the Chicago outlet (Johnston et al. 2004).

The Nipissing shoreline has been identified along the southern shore near Duluth where it has been identified just over 1 m above the present water level (Farrand 1960). In the Pigeon River and Marathon areas, two Nipissing beaches were found (Farrand 1960). The separation in the beaches has been attributed to uplift (7.5 m in Marathon) during the transgression and not to fluctuating Nipissing water levels (Farrand 1960). A prominent Nipissing bluff is located in Hurkett, Ontario between 213 and 216 m asl and, in the Thunder Bay area, the Nipissing shoreline is present at 193 m asl (Figure 2.10) (Farrand 1960).

2.4.6. Post-Nipissing Transgressions (3,400- ~2,200 cal yr BP)

Algoma (3,400-2,300 cal yr BP)

A pause in the downcutting of the Huron outlet resulted in the Algoma phase in Lakes Superior, Huron, and Michigan. Saarnisto (1975) agrees with the earlier work by Lewis (1969) that this phase occurred sometime between 3,400-2,600 cal (3,200-2,500 ^{14}C) yr BP in the Huron basin, which correlates with Farrand’s assessment of shoreline displacement in the Superior basin. More recently Baedke et al., (2004) used a hydrograph to determine level fluctuations in the Michigan and Huron basins and found that in accordance with previous studies, the Algoma phase occurred between 3,500-2,300 cal yr BP. Between the Nipissing and Algoma stages, there was a drop of approximately 4 m in the Superior, Michigan and Huron basins.

Evidence of an Algoma beach was found near Sault Ste. Marie at 189 m asl (Farrand 1960). Due to the tilting of the basin, Farrand (1960) estimates that the Algoma beach should be about 2 m below the present water level near Duluth.

Sault (2,200 cal yr BP)

The next small transgression was the Sault. The Sault level was reached as Lake Superior and Lake Huron became separated because of the rebounding Sault Ste. Marie outlet around 2,200 cal yr BP (Farrand 1960; Farrand and Drexler 1985; Johnston et al. 2004). The Sault shoreline can be visible below the Nipissing shoreline (Farrand and Drexler 1985), but it is only visible in a few places along the north shore of Lake Superior (Farrand 1960). Typically the Sault shoreline is visible from Thunder Bay to Marathon at 190 m asl, 190.5 m asl in Terrace Bay, and between 183-184 m asl in Sault Ste. Marie (Farrand, 1960). The differences in elevation have been attributed to differential isostatic rebound and not to fluctuations in water level during this time (Farrand 1960).

In some locations, below the main Sault shoreline, is a Sub-Sault shoreline (Farrand 1960). This shoreline has been identified by bluffs in Hurkett around 186 m asl, Terrace Bay at 190 m asl and Marathon at 190 m asl (Farrand 1960). A Sub-Sault beach has also been identified in Sault Ste. Marie at 183 m asl (Farrand, 1960).

2.5. SUMMARY

The histories of deglaciation and lake level fluctuations are closely entwined in the Lake Superior basin. The deglaciation of Northwestern Ontario and the Thunder Bay area is complicated by a lack of direct radiocarbon dates from moraines, and the possible erosion or burial of pre-Marquette landforms and landscapes by Marquette ice. If the Marquette advance was a small advance as suggested, then a large area of Northwestern

Ontario may have been deglaciated for a longer period of time. This could have important impacts on both the paleovegetation and archaeological records of this area. Accurate dating of the moraines, as attempted by Lowell et al., (2010) has provided a valuable resource to understanding glacial retreat within Northwestern Ontario however, much is still to be learned, particularly in regards to the location of the LIS after 10,700 cal yr BP.

Fluctuating water levels, mainly those of glacial Lake Minong, were largely controlled by movements in the LIS. Once the Nipigon outlets were open, Lake Agassiz was able to drain into the Lake Superior basin, producing high water levels in the basin until 9,000 cal yr BP. With the bypassing of the Lake Superior basin by glacial Lake Agassiz meltwater, in addition to the warming and drying climate that developed following glacial retreat, the low levels of the Houghton Low were reached. It was not until the uplifting of the North Bay outlet above the Sault Ste. Marie outlet, in addition to a wetter and more humid climate around 6,800 cal yr BP, that water levels once again rose within the basin.

Dating deglacial events and lake level fluctuations is difficult because there are limited organics associated with depositional features formed by them. Locating and correlating shorelines around the basin is also hindered by differing rates of glacial rebound, which makes producing basin-wide correlations difficult. A major key to understanding the deglacial and lake level history of the Lake Superior basin is to attach dates to the formation of paleoshorelines and moraines along the north shore of the basin such as the Minong shorelines and the Nakina moraines. Since the north shore of the basin rebounds at a faster rate, this area is the only area where a more complete record of Holocene lake level fluctuations is preserved. However, the areas on the same isobase as

Sault Ste. Marie, such as Thunder Bay, should not be ignored since these areas are likely to record actual changes in lake level and are not as affected by differential isostatic rebound.

CHAPTER 3

PALEOVEGETATION AND CLIMATE CHANGE IN NORTHWESTERN ONTARIO

3.1. INTRODUCTION

This chapter focuses on the relationship between climate and vegetation. It summarizes the paleovegetation and climatic histories of Northwestern Ontario using pollen, macrofossil, diatom and oxygen and carbon isotope records from Minnesota, northern Ontario, the Plains, and the other Great Lake basins with an emphasis on changes in vegetation due to migration, expansion and retraction, climate change in the form of warming and cooling trends and dating of these events.

The vegetation history of Northwestern Ontario has been affected by the presence of the Laurentide Ice Sheet, glacial Lake Agassiz and climate change. The cold climate that the LIS and glacial Lake Agassiz produced created short-lived tundra like conditions in Northwestern Ontario followed by boreal forest vegetation as the LIS retreated and the climate began to warm. This climate reversal to warm and dry conditions occurred around 11,500 (10,000 ^{14}C) yr BP and is known as the Hypsithermal. The Hypsitermal had a profound effect on vegetation as the prairie biome expanded north and west and the Great Lakes-St. Lawrence forest expanded further into northern Ontario. The change to a warm and more arid climate appears to have been time-transgressive as these conditions were originally felt on the Plains, and later in Northwestern Ontario. A delay in warming conditions however, was also likely due to the continued presence of the LIS and as a result of local and regional factors including the proximity of Lake Superior, which continues to create colder conditions along its margins today than in inland areas.

3.2. MAJOR CLIMATIC TRENDS OF THE LATE PLEISTOCENE AND HOLOCENE

Climate during the Late Pleistocene and Early and Middle Holocene fluctuated dramatically, with multiple major and minor events identified, mainly through the Greenland Ice Core and through isotope analysis of organics (i.e., shells, macrobotanical remains) in lakes. Many of these fluctuations correspond to a 1,500 year cycle however, the driving force(s) behind these changes are poorly understood (Broecker 2000). Alterations in thermohaline circulation (Broecker 2000), changes in atmospheric circulations (Bryson and Wendland 1967), glacial meltwater outbursts (Broecker et al. 1989) and dust (Broecker 2000) have all been proposed as causes for the climatic changes observed in the ice core and isotope records. None of these hypotheses, however, have explained how these climatic trends impacted areas at such a large scale (Broecker 2000). This section describes the major climate trends that likely affected Northwestern Ontario and can be seen in pollen records from the region.

As the Last Glacial Maximum was coming to an end, the late-glacial period experienced warming temperatures, beginning around 18,300 cal (15,000 ^{14}C) yr BP, (Jackson et al. 1997). This trend, commonly referred to as the Bølling-Allerød (BOA) was identified through isotopes in the Greenland ice core, and lakes throughout Europe and North America (Yu and Eicher 1998). Isotope records from Crawford Lake in southern Ontario place this warming trend between 14,400-12,900 cal (12,300-11,000 ^{14}C) yr BP (Yu and Eicher 1998). Climate conditions changed from cold and dry to relatively warm, much like the climate today (Broecker 2000).

In deglaciated parts of North America, it is within this time that *Picea* populations abruptly decline (Wright 1974) and non-analogous vegetation assemblages appear (Birks 2003). In early lake sediments from Minnesota as well as from the Cummins and Oliver Ponds in Thunder Bay and Rattle Lake in Northwestern Ontario, there are consistent, but low values of thermophilous deciduous tree pollen including *Fraxinus* (ash)*, Quercus* (oak) and *Ulmus* (elm) (Baker 1965; Birks 2003). It is believed that this pollen was transported over long distances because it is thought that the climate was not warm enough to sustain them, and no macrofossils from these species have been found dating to this time (Birks 2003). However, macrofossils from the thermophilous deciduous trees *Alnus glutinosa*, *Tilia cordata*, and *Betula pendula* have been found in Sweden mixed with coniferous species *Picea abies* and *Larix sibirica,* supporting the idea that thermophilous deciduous trees were part of the boreal environment in the early Holocene (Kullman 1998). Yu and Wright (2001) suggest that *Fraxinus, Quercus,* and *Ulmus* could have been in the area prior to the Younger Dryas due partly to the increase in summer insolation but also because the LIS may have blocked the dry and cold Arctic air mass. As a result, the warm and humid Caribbean air mass may have had more of an effect on the climate in the interior continent during this time, allowing thermophilous deciduous trees to grow close to the ice sheet (Yu and Wright 2001). Pollen from these species disappear from Northwestern Ontario at the onset of the Younger Dryas (Björck 1985).

Beginning around 12,900 cal (11,000 ^{14}C) yr BP there was a shift to much colder temperatures that was felt throughout much of the Northern Hemisphere (Hu et al. 1999). This cold period is known as the Younger Dryas, which is thought to have been responsible for the Marquette advance in the Laurentide Ice Sheet (LIS) (Farrand and Drexler 1985; Teller and Thorleifson 1983) as discussed in Chapter 2 (Deglaciation and

Paleohydrology). Oxygen isotopes can reveal changes in atmospheric precipitation and mean annual water surface temperatures and have been studied at Crawford Lake and Twiss Marl Pond in southern Ontario indicating hat temperatures may have dropped 3ºC throughout the Younger Dryas from the earlier warming period (Yu et al. 1998).

The cause of this cooling trend may have been a glacial meltwater outburst into the North Atlantic (Broecker et al. 1989). Broecker et al (1989) propose that a massive influx of glacial meltwater into the North Atlantic may have shut down the formation of North Atlantic Deep Water (NADW), causing a shift in thermohaline circulation. This would have resulted in the widespread cooling of the Northern Hemisphere as the warm tropical waters brought by the Gulf Stream would have been obstructed by the cold glacial meltwater. Broeker et al*.,* (1989) suggest that the source of this meltwater was glacial Lake Agassiz which would have drained through its eastern outlets into the Lake Superior basin, through the St. Lawrence River and into the North Atlantic. The problem with this theory is that Murton et al., (2010) found evidence that Lake Agassiz drained north, through the Athabasca-Clearwater route during this time and not east through the Nipigon outlets, which were likely still covered by the Laurentide Ice Sheet. Another cause may have been by an increase in precipitation as the LIS retreated (Eiseman et al. (2009). Eiseman et al., (2009), suggest that as the LIS retreated, the jet stream migrated north, bringing more wind and precipitation to the northern latitudes. This may have caused a cooling of North Atlantic Ocean surface waters, an extension of sea ice and a disruption in thermohaline circulations which could have also resulted in cooler atmospheric temperatures (Eiseman et al. 2009). The exact cause of the Younger Dryas remains debatable. However, it has been accepted that thermohaline circulation was disrupted by an influx of freshwater, although where this water came from (i.e., through the Mississippi into the Gulf of Mexico, through the St. Lawrence, through the Arctic Ocean into the North Atlantic, or from an increase in precipitation) remains unclear.

Although the LIS was present throughout most of Northwestern Ontario at this time, pollen obtained from Rattle Lake, north of Fort Frances, Ontario, has shown a period of climatic deterioration between 13,000-11,900 cal (11,100-10,200 ^{14}C) yr BP (Figure 3.2 for location of lakes mentioned in this chapter) (Björck 1985). High pollen values of *Salix,* Cyperaceae, Graminae, Compositae and *Picea* indicate that dwarf shrub tundra was the dominant vegetation during this time, and pollen of the thermophilous deciduous trees *Fraxinus* and *Ulmus* identified earlier between 13,300-13,000 cal (11,400-11,100 ^{14}C) yr BP disappear all together (see Figure 3.1) (Björck 1985). However, the pollen from these deciduous trees could be the result of long-distance transport and therefore their disappearance could simply be explained by an increase in local plant productivity and greater local pollen influx (Björck 1985).

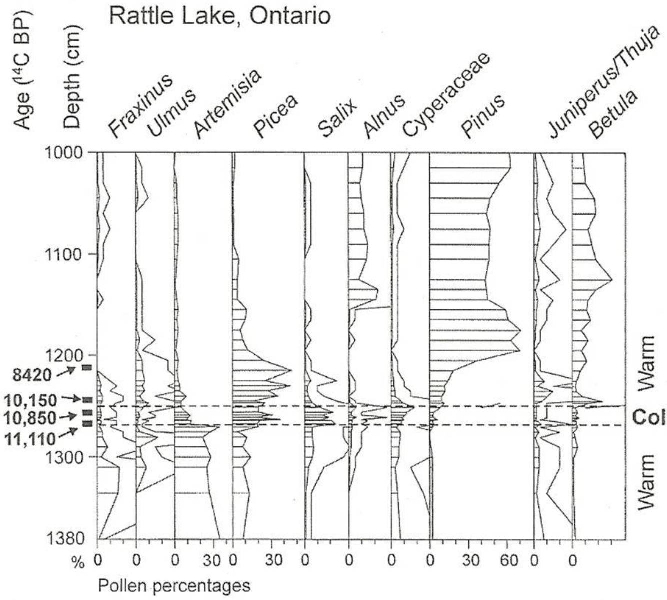


Figure 3.1. Pollen diagram from Rattle Lake showing the warm Bølling-Allerød period at the bottom of the figure, followed by the cold Younger Dryas, followed by Early Holocene warming (from Björck 1985).

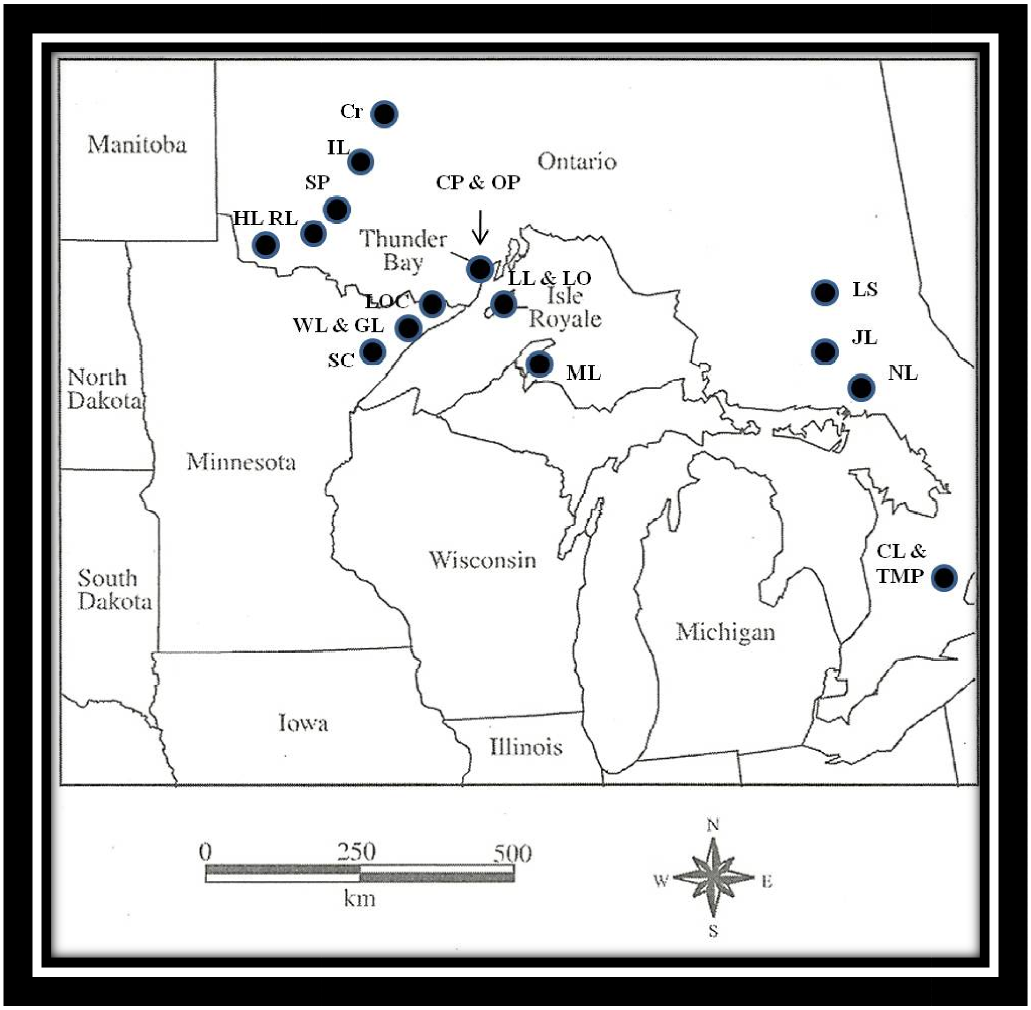


Figure 3.2. Location of some of the lakes mentioned in this chapter (modified from Flakne 2003). CL= Crawford Lake, TMP= Twiss Marl Pond (Yu et al. 1997, 1998), NL= Nina Lake, JL= Jack Lake, LS = Lake Six (Liu 1990), LL= Lily Lake, LO= Lake Ojibway (Flakne 2003), CP= Cummins Pond, OP= Oliver Pond (Julig et al. 1990), SC= Spider Creek (Baker 1965), Weber Lake (Fries 1962), GL= Gegoka Lake (Huber 1996), LOC= Lake of the Clouds (Craig 1962), RL= Rattle Lake, SP= Sioux Pond, IL= Indian Lake, Cr= Cristal Lake (Björck 1985), HL= Hayes Lake (McAndrews 1982), ML= Mud Lake (Booth et al. 2002).

Another major climatic trend that impacted vegetation in the Upper Great Lakes is the Hypsithermal, a period of unusually warm and arid conditions that occurred during the Early and Middle Holocene (Webb et al. 1983; Williams et al. 2009, 2010). Drying conditions began on the Plains around 17,200 cal (14,000 ^{14}C) yr BP as the LIS began to retreat, and lasted until 6,800 cal (6,000 ^{14}C) yr BP. However, most sites showed signs of drying between 11,500-8,900 cal (10,000-8,000 ^{14}C) yr BP (Williams et al. 2010). Not all locations experienced these conditions at the same time. Instead, sites from the interior portion of the Plains region showed signs of drying first with those on the periphery experiencing it later (Williams et al. 2010). In Minnesota the dry and warm conditions occurred between 11,500-7,900 cal (10,000-7,000 ^{14}C) yr BP (Clark et al. 2001), and in the Mid-West between 6,300-3,200 cal (5,500-3,000) ^{14}C yr BP (Webb III et al. 1983), making this climate trend time-transgressive from west to east (Baker et al. 1992). Oxygen isotope data from Twiss Marl Pond and Crawford Lake along the Niagara Escarpment in Southern Ontario, indicate that temperatures may have risen 6ºC from the colder Younger Dryas period (Yu et al. 1997). Temperatures, however, likely fluctuated during both the Younger Dryas and Hypsithermal (Yu et al. 1997).

The warmer and drier conditions have been attributed to a combination of glacial retreat and to an increase in summer insolation (Webb et al. 1993). Changes in atmospheric circulation also may have produced these conditions, as it is possible that the warm and dry Pacific air mass from the west dominated in the Canadian and American Plains and parts of the Mid-West for a longer period of the year (Figure 3.3) (Baker et al. 1992; Bryson and Wendland 1967). Drainage of Lake Agassiz may have also resulted in an increase in temperatures for areas around the glacial lake, as its presence is thought to have kept surrounding areas 1-2ºC cooler (Hostetler et al. 2000). Due to glacial retreat, the areas farther away from the LIS, such as the Great Plains, should have experienced warming and drying first, and this is, in fact, what appears to have occurred (Wright 1983). Areas closer to the ice front, such as Northwestern Ontario, would have experienced these conditions much later and also differently than the Great Plains or locations to the south, as the LIS would have continued to keep summer temperatures cooler (Wright 1983). Sediment obtained from a core in Mud Lake along the southern shore of Lake Superior in Wisconsin shows a decline in Lake Superior water levels (Houghton phase) beginning around 9,600 cal (8,700 ^{14}C) yr BP however, even drier conditions were recorded around 8,800 cal (7,900 ^{14}C) yr BP by a decline in submersed aquatic species and in increase in emergent taxa (Booth et al. 2002). This resulted in Mud Lake turning into a shallow water fen until 6,600 cal (5,800 ^{14}C) yr BP, when water levels within the Lake Superior basin and Mud Lake rose due to the Nipissing transgression (Booth et al. 2002).

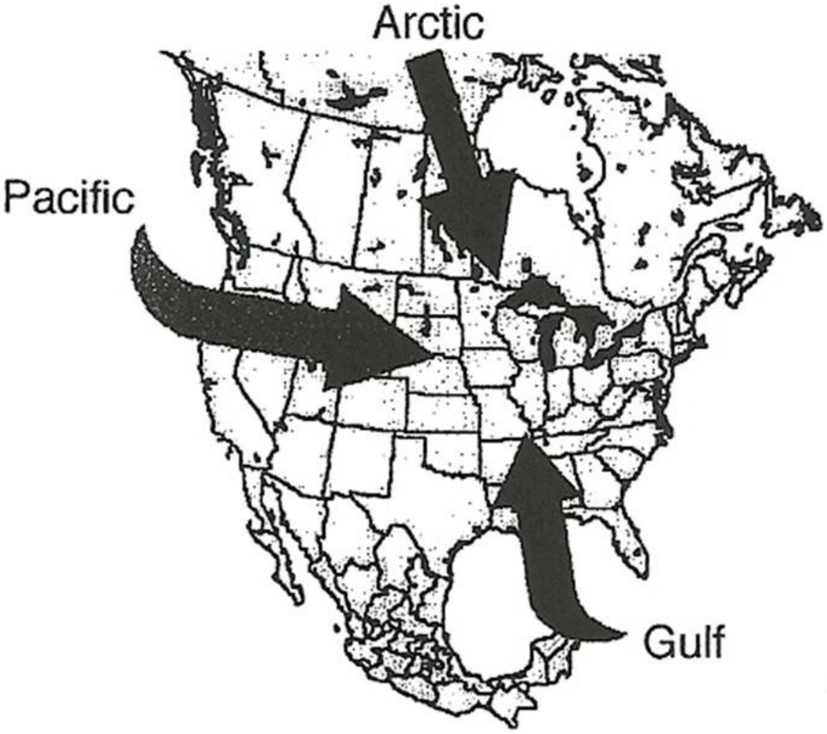


Figure 3.3. Origin and direction of the Arctic, Pacific and Gulf air masses which contribute to the climate of North America (from Clark et al. 2001)

Pollen data suggest that most species had reached their most northern extent during the Middle Holocene warming by 6,800 cal (6,000 ^{14}C) yr BP (Jackson et al. 1997; Webb et al. 1993). Prairie vegetation expanded both north and east during this time, migrating further north in the Canadian provinces and as far east as eastern Minnesota (Bryson and Wendland 1967), with high herb pollen values found in virtually every pollen record from Minnesota (Birks 2003). The hardwood dominating the Great Lakes-St. Lawrence forest also shifted north at this time (Liu 1990). In Northern Ontario the forest may have reached 140 km north of its present position between 7,800-3,200 cal (7,000-3,000 ^{14}C) yr BP (Liu 1990).

The Great Lakes themselves appear to have been affected by the Hypsithermal. As discussed briefly in Chapter 2 (Deglaciation and Paleohydrology), Lakes Huron and Michigan were in low phases beginning in the Early Holocene as the North Bay outlet opened, draining the lakes. Lake levels remained low except for possible periodic outbursts from glacial Lake Agassiz. However, around 8,900 cal (8,000 ^{14}C yr BP), the relative water level in the Lake Superior basin appears to have been below all possible outlets, suggesting that drier conditions and less precipitation and/or higher rates of evaporation also contributed to the low levels (Boyd et al. 2010; Lewis et al. 2007; Yu et al. 2010). Radiocarbon dated submerged stumps in Lake Huron record low water levels between 11,500-7,900 cal (10,000-7,000 ^{14}C) yr BP, as forests were able to grow on previously submerged lake bottom (Karrow et al. 2007). Submerged river channels through the Straits of Mackinac prove that water levels were much lower than today during this phase (Farrand 1960). A submerged forest at the Olson Site in Lake Michigan, roughly 62 m below the present water level, also testify to the extent and timing of this low phase, with stumps from this locality providing an average date of 9,300 cal (8,300 ^{14}C) yr BP (Chrzastowski et al. 1991).

Lastly a period of climatic cooling and more moist conditions occurred following the Hypsithermal, beginning around 7,800 cal (7,000 ^{14}C) yr BP and peaking between 4,500-3,200 cal (4,000-3,000 ^{14}C yr BP (Wright 1974). Summer temperatures decreased as the 20ºC isotherm (mean July isotherm) moved south of the Canadian border (Webb et al. 1993). The southward shift of this isotherm affected vegetation, as the Great Lakes-St. Lawrence forest that had migrated north during the warm and dry Hypsithermal retracted to its modern location and boreal forest vegetation migrated south near its present position (Liu 1990). *Picea* also migrated further south into northeastern Minnesota, replacing the pine population (Wright 1992). *Picea* populations are generally controlled by the 18ºC isotherm (mean July temperature), suggesting a shift to cooler temperatures at this time as this isotherm also moved south (Webb et al. 1993).

Carbon isotopes on *Sphagnum* from swales along the south shore of Lake Superior support an increase in precipitation and humidity between 3,400-2,300 cal yr BP (Sharma et al. 2005). Two periods of increased moisture were identified through the analysis of stable oxygen and carbon isotopes from calcite ostracod shells from Pickerel Lake, South Dakota during the Late Holocene (Schwalb and Dean 1998). The first episode was around 4,500 cal (4,000 ^{14}C) yr BP and the second around 2,200 cal (2,200 ^{14}C) yr BP. Schwalb and Dean (1998) suggest that the increased moisture may have been from changes in thermohaline circulation, and/or a dominance of the warm and moist air from the Gulf of Mexico, over the dry and cooler air from the Pacific and Arctic air masses. Booth and Jackson (2003) record increased moisture between 3,200-2,300 cal yr BP based on pollen, macrofossil, and testate amoeba from the southern shore of Lake Huron. They also attribute the increase in moisture to fluctuations in atmospheric circulation (Booth and Jackson 2003).

3.3. MODERN VEGETATION IN NORTHWESTERN ONTARIO

Two forest zones merge in northern Ontario, providing a mix of coniferous and deciduous species (Liu, 1990). The Thunder Bay area, in particular, borders both Boreal forest and Great Lakes-St. Lawrence forest (Figure 1.3) (Kemp 1991). The boreal forest is dominated by *Picea glauca* (white spruce), *Abies balsamea* (balsam fir), *Pinus banksiana* (jack pine), *Betula papyrifera* (white birch), *Populus tremuloides* (trembling aspen), *Populus balsamifera* (balsam poplar), *Picea mariana* (black spruce), and *Larix laricina* (tamarack) (Liu 1990). Great Lake-St. Lawrence forest consists of *Pinus strobus* (white pine), *Pinus resinosa* (red pine), *Tsuga canadensis* (hemlock), *Fagus grandifolia* (American beech), *Betula alleghaniensis* (yellow birch), and *Acer saccharum* (sugar maple) (Liu 1990).

Fourteen hardwood species reach their northern limits in northern Ontario (Liu 1990). These include *Fagus grandifolia, Betula alleghaniensis, Acer saccharum, Acer rubrum* (red maple), *Ostrya virginiana* (Hop-hornbeam), *Ulmus americana* (American elm), *Fraxinus nigra* (black ash), *Fraxinus pennsylvanica* (green ash), *Tilia americana* (American basswood), *Populus grandidentata* (larged-toothed aspen), *Pinus strobus, Pinus resinosa, Tsuga canadensis,* and *Thuja occidentalis* (eastern white cedar) (Liu 1990).

3.3.1. Vegetation and Climate

In Northwestern Ontario, the Arctic air mass brings dry and cool air between six and ten months out of the year, extending over the winter (Kemp 1991). The southward limit of the boreal forest coincides with the mean winter location of the Arctic air mass (Bryson and Wendland 1967), as trees such as white and black spruce, balsam fir, jack pine, and tamarack tolerate the colder climates characteristic of the Subarcitc environment (Kemp 1991). The Pacific air mass generally dominates the warmer summer months (Kemp 1991). Most of the precipitation received in this area comes in the form of snow in the winter from Arctic air mass or from periods where the warm and humid Gulf air mass reaches the area throughout the year (Kemp 1991).

Changes in vegetation are mainly controlled by climatic factors, notably precipitation, insolation and temperature, although soil conditions, human impact, topography and other local factors also play a role (Flakne 2003). It is possible to infer climatic conditions from vegetation because species are sensitive to particular environmental/climatic conditions and will respond if these conditions change (Birks 2003). The main response to climate/environmental change is the migration, expansion or retraction of species. In areas where two ecozones converge, such as in the southern region of Northwestern Ontario where the Great Lakes-St. Lawrence forest and boreal forest abut (Kemp 1991; Liu, 1990), it is easier to see vegetation fluctuations in response to changing environmental conditions such as those that occurred throughout the Holocene.

Webb et al.'s (1993) study of pollen types, including species found in this area, connect their presence to particular climatic preferences. For example, the presence of *Picea* seems to be more influenced by the mean July temperature than mean January temperature, whereas *Pinus* appears to be influenced more by mean January temperatures and *Quercus* appears to be influenced by both (Webb et al. 1993). It has also been found that *Pinus, Ulmus,* and *Quercus* are more sensitive to annual precipitation than other species (Webb et al. 1993). Therefore a prolonged increase in summer temperature may result in the decline of spruce, and an increase in precipitation may result in the migration or expansion of *Pinus, Ulmus*, and *Quercus* (Webb et al. 1993).

Local variables which impact vegetation include soil moisture, soil depth, soil nutrients, soil acidity, microclimate, fire, shade, and competition with other types of vegetation, to name a few (Barnes et al. 1998; Farrar 2007; Flakne 2003). A unique environment exists on Isle Royale. The southwest half of the island is northern hardwood forest while the northeast half is mixed boreal forest (Flakne 2003). The southwest half is also covered with glacial till whereas the northeast half of Isle Royale is not (Flakne 2003). Lake Ojibway, on the northeast part of the island, remained relatively stable with coniferous forest throughout much of the Holocene, but became mixed coniferous forest in the mid-Late Holocene (Flakne, 2003). Lily Lake, on the southwest part of the island, was less stable, with spruce and fir dominated forest in the Early Holocene, followed by a mixed forest with birch, pine and alder, and then by a pine dominated forest in the Middle Holocene. It returned back to mixed forest with birch, maple, oak, spruce and fir in the mid-Late Holocene then finally to a birch dominated forest in the Late Holocene (Flakne 2003). Reasons as to why these sites have two separate forest types remains unclear however, differences in microclimate, soil, founder effects (isolated population of trees which become genetically different than the original population) and disturbance have all been proposed (Flakne 2003).

3.4. PALEOVEGETATION RECORDS FOR NORTHWESTERN ONTARIO AND ADJACENT REGIONS

What is known about the paleovegetation record of Northwestern Ontario comes from either pollen records or, more rarely, through macrofossil analysis (see Chapter 5 [Methodology and Theory] for methodological information on macrofossil analysis and its use in paleovegetation reconstructions). There are few sites throughout this region which have been examined palynologically (Liu 1990), and fewer still using macrofossils. Terasamae (1967) recognized that immediately following glacial retreat, soil was slow to accumulate, which may have resulted in gaps in the pollen record as have been found in the early pollen record from Quadrangle Lake near Sault Ste. Marie (Terasamae 1967). This means that pollen representing initial vegetation may be absent. It is unknown if tundra vegetation was quickly succeeded by boreal forest as is commonly thought or if boreal forest species were actually the first to occupy particular areas, nor is it known if the ambiguous data on this situation reflect a major gap in the pollen record. These problems make it necessary to not only analyze pollen data from Northwestern Ontario, but to also look at vegetation histories of regions adjacent to this region such as Northeastern Minnesota and northern Ontario so more accurate paleovegetation reconstructions can be made.

3.4.1. Northern Minnesota

The nature of vegetation from 12,900-11,500 cal (11,000-10,500 ^{14}C) yr BP in this region has been reconstructed based largely on pollen with some macrofossil remains and varies between sites: from tundra at Lake of the Clouds to prairie-tundra at Weber

Lake based on high values of Gramineae (grasses), Cyperaceae (sedges), Amaranthaceae (e.g., the goosefoot family), and Compositae (e.g., aster, daisy). At Spider Creek the pollen record indicates that late-glacial vegetation was dominated by non-arboreal pollen (NAP), including *Dryas integrifolia* (entireleaf mountain-avens) and *Salix herbacea* (snowbed willow), ambrosia (ragweed), *Artemisia* (e.g., sagebrush, wormwood), and Amaranthaceae with some arboreal pollen of *Fraxinus, Quercus* and *Ulmus* (Baker 1965). Based on this pollen assemblage, the environment was either tundra, park-tundra, or boreal-prairie (Baker 1965). However, macrofossils identified at this site represent only tundra vegetation (Baker 1965). After 12,500 cal (10,500 ^{14}C) yr BP, the pollen assemblages change with the addition of *Picea* and *Betula*, suggesting that the tundra or tundra-prairie dominated landscape was changing into shrub parkland or open conifer forest (Craig 1972; Fries 1962; Huber 1996). At Spider Creek, macrofossils of *Larix* have been found, indicating that larch was also a pioneering species in this region in early post-glacial times (Baker 1965).

After 11,500 cal (10,000 ^{14}C) yr BP, there was a change in vegetation, likely due to a warming climate resulting from the retreat of the LIS. This change is evident in an increase in thermophilous deciduous tree pollen (*Quercus, Fraxinus, Ulmus*) and *Pinus* pollen at Weber Lake between 11,500-10,200 cal (10,000-9,000 ^{14}C) yr BP (Fries 1962). A mixed conifer-hardwood forest was present at Gegoka Lake between 11,900-10,200 cal (10,200-9,000 ^{14}C) yr BP, and an increase in *Pinus* pollen (jack pine or red pine) may indicate the growing dominance of this genus in the forest (Huber 1996). By 10,300 cal (9,200 ^{14}C) yr BP., boreal forest had been established at Lake of the Clouds and persisted until 9,300 cal (8,300 ^{14}C) yr BP after which the pollen record shows an increase in prairie vegetation and warming conditions as indicated by high *Pinus* and NAP values that lasted until 7,400 cal (6,500 ^{14}C) yr BP (Craig 1972). Forest composition changed at Lake of the Clouds from spruce dominated boreal forest to a mixed conifer, *Pinus* dominated hardwood forest (Craig 1972). At the same time at this time, the forest at Gegoka Lake remained mixed conifer-hardwood, but pine and NAP vegetation may have played a bigger role, as their pollen values were increasing during this period (Huber 1996).

Around 7,800 cal (7,000 ^{14}C) yr BP, the pollen assemblage again changes, mainly due to the migration of *Pinus strobus* (white pine) into the region, which also indicates that the climate was becoming warmer and possibly moister (Craig 1972). Drier conditions may also have been present around 8,100 cal (7,300 ^{14}C) yr BP as implied by a decrease in *Picea* and thermophilous deciduous tree pollen at Weber Lake (Fries 1962). At Gegoka Lake, the migration of white pine and its replacement of jack/red pine is thought to have occurred around 7,800 cal (7,000 ^{14}C) yr BP however, this replacement may have been short-lived since *Pinus* values decrease and *Picea* and *Betula* pollen values begin to increase shortly thereafter (Huber 1996). White pine reaches Lake of the Clouds around 7,400 cal (6,500 ^{14}C) yr BP and thermophilous deciduous tree pollen values also increase around this time. This is indicative of the warmer and/or drier conditions that, according to the pollen assemblage, lasted until 3,200 cal (3,000 ^{14}C) yr BP (Craig 1972).

Climate conditions changed shortly after 3,200 cal (3,000 14C) yr BP at Lake of the Clouds from warm and dry to cool and moist (Craig 1972). This is reflected in the pollen assemblage by increasing values of *Picea* and decreasing values of *Pinus* (Craig 1972). This climate reversal is also noted at Weber Lake around 2,000 cal yr BP, where pollen values of *Picea, Larix,* and *Abies* increase, while those of thermophilous deciduous species are low, indicating a coniferous forest dominated landscape (Craig 1972). It may be, though, that these cooler and moister conditions only occurred at a local scale and may not reflect the climate over a larger area (Craig 1972).

3.4.2. North-Central Ontario

Liu (1990) cored three lakes (Nina, Jack, and Six) located between Lake Huron and James Bay. All three of the lakes showed *Picea* (mainly *Picea glauca*) as the dominant vegetation in the post-glacial period [~11,200 cal (9,800 ^{14}C) yr BP], indicating a boreal forest environment (Liu 1990). At Nina Lake between 10,100-8,200 cal (8,900-7,400 ^{14}C) yr BP, *Pinus banksiana/resinosa* and *Alnus* pollen values increase while *Picea* decreases, indicating a warming climate (Liu 1990). Around 8,200 cal (7,400 ^{14}C) yr BP, the boreal forest in North-Central Ontario was transformed into Great Lakes-St. Lawrence forest by the migration of *Pinus strobus*, *Fagus, Tsuga* and *Acer* into the area (Liu 1990). This change in forest composition coincides with the warming and drying conditions of the Hypsithermal (Liu 1990). At 4,500 cal (4,000 ^{14}C) yr BP, climatic cooling results in the decline of *Pinus strobus*, a retraction of Great Lakes-St. Lawrence forest species south, and the expansion of the boreal forest south (Liu 1990).

Ritchie (1987) uses various pollen records from northern Ontario to reconstruct the paleovegetation. Initial vegetation after deglaciation was shrub tundra with white spruce and poplar groves (Ritchie 1987). Around 10,700 cal (9,500 ^{14}C) yr BP, forest communities spread into mesic areas where *Picea glauca* dominated with *Populus*, but *Ulmus, Pinus,* and *Quercus* may also have been present (Ritchie 1987). At 10,300 cal (9,200 ^{14}C) yr BP, *Picea* decreases and *Larix, Betula papyrifera, Alnus* and *Pinus* *banksiana* expand (Ritchie 1987). The increase in the latter two may indicate a higher frequency of forest fires during this time (Ritchie 1987). At 8,200 cal (7,400 ^{14}C) yr BP there is a marked increase in *Pinus strobus, Tsuga,* and *Acer* pollen, indicating the migration of these trees into northern Ontario from Southern and Central Ontario (Ritchie, 1987). Around 6,800 cal (6,000 ^{14}C) yr BP, the addition of *Fagus* is noted in the pollen record as well as the presence of *Thuja* and *Betula alleghaniensis*, which may have replaced *Pinus strobus* on mesic sites (Ritchie 1987). Finally, from 4,500 cal (4,000 ^{14}C) yr BP to the present, there is an increase of *Picea, Pinus banksiana,* and *Betula papyrifera* as *Tsuga, Acer saccharum* and *Fagus* decline to very low values, indicating cooling temperatures (Ritchie 1987).

3.4.3. Northwestern Ontario

In Kenora, located near the Manitoba border, McAndrews (1982) cored Hayes Lake (McAndrews 1982). Initial vegetation following deglaciation was dominated by herbs and indicates a tundra environment however, the cold glacial waters of glacial Lake Agassiz may have kept this region colder than surrounding areas (McAndrews 1982). Prior to 11,500 cal (10,000 ^{14}C) yr BP, *Picea* pollen dominated the pollen assemblage with the addition of *Larix, Betula, Populus,* and *Ulmus*, which signifies warmer and possibly wetter conditions (McAndrews 1982). Between 11,500-3,900 cal (10,000-3,600 ^{14}C) yr BP warming continues, specifically with the Hypsithermal, which has been identified at this site as occurring between 10,300-3,900 cal (9,200-3,600 ^{14}C) yr BP, as indicated by the pollen record (McAndrews 1982). *Pinus banksiana/resinosa*, as well as *Betula*, dominate the early pollen assemblage of this period however, *Pinus* pollen starts to decline and herb pollen increases around 7,400 cal (6,400 ^{14}C) yr BP, indicating the eastern and northern migration of prairie vegetation (McAndrews 1982). During the Hypsithermal, McAndrews (1982) suggests that the site and surrounding area were an open, grassy jack pine and poplar woodland with poplar and white spruce inhabiting mesic sites. After 3,900 cal (3,600 ^{14}C) yr BP, a climatic cooling occurs as indicated by increasing *Picea, Larix,* and *Abies* pollen values, as well as a decrease in herb and *Populus* pollen. This cooling trend changed the open-mixed woodland into closed forest (McAndrews 1982).

Björck (1985) cored four lakes and ponds in Northwestern Ontario, Rattle Lake, Sioux Pond, Indian Lake, and Cristal Lake, to obtain information on deglacial events and paleovegetation. From 13,300-11,900 cal (11,400-10,200 ^{14}C) yr BP the pollen assemblage at Rattle Lake is dominated by Compositae, *Fraxinus* and Amaranthaceae, indicating an arctic environment with tundra vegetation (Björck 1985). The *Fraxinus* pollen most likely is as a result of long distance transport. Between 11,900-11,300 cal (10,200-9,900 ^{14}C) yr BP, the pollen assemblage is dominated by *Betula, Ulmus,* and *Picea*, suggesting that boreal forest vegetation quickly moved into the Rattle Lake area as the temperature began to warm during the Early Holocene (Björck 1985). At the same time however, the pollen assemblage from Sioux Pond is dominated by Compositae, *Ulmus, Juniperus*, Cyperaceae, *Picea* and *Salix* which indicates the landscape was either park-tundra, or tundra-forest with occasional stands of spruce (Björck 1985).

Between 11,300-9,500 cal (9,900-8,600 ^{14}C) yr BP, *Picea* dominated the pollen assemblage at both Rattle Lake and Sioux Pond, with *Larix, Populus, Betula, Ulmus* and *Thuja* also present however, *Picea* did not dominate the assemblage at Indian Lake until around 10,300 cal (9,200 ^{14}C) yr BP. At Indian Lake around 10,500 cal (9,300 ^{14}C) yr BP, Compositae pollen dominated, with the presence of *Pinus,* Cyperaceae, *Salix* and *Larix*, indicating more open vegetation, however it was quickly being replaced by the rapid migration of boreal forest species, most notably *Picea* and *Larix* (Björck 1985). The spruce dominated pollen assemblage persisted until 9,800 cal (8,800 ^{14}C) yr BP at Indian Lake, 9,500 cal (8,600 ^{14}C) yr BP at Rattle Lake, and 10,700 cal (9,500 ^{14}C) yr BP at Sioux Pond (Björck 1985).

At Sioux Pond the *Picea* dominated assemblage was replaced by one dominated by *Pinus* (jack and/or red pine) with *Betula, Alnus,* and *Picea* from 10,700-7,800 cal (9,500-7,000 ^{14}C) yr BP, indicating pine dominated boreal forest (Björck 1985). This *Pinus* assemblage occurred between 9,500-8,700 cal (8,600-7,900 ^{14}C) yr BP at Rattle Lake and 9,800-8,200 cal (8,800-7,400 ^{14}C) yr BP at Indian Lake (Björck 1985).

The pollen assemblage changes at Rattle Lake between 8,700-7,400 cal (7,900-6,400 ^{14}C) yr BP with the introduction of *Pinus strobus, Alnus rugosa* (speckled alder) and *Alnus crispa* (green alder) into the area (Björck 1985). *Pinus* remains the dominant vegetation however, white pine has replaced the existing jack/red pine populations (Björck 1985). Björck (1985) suggests that white pine may have migrated into the area either from Minnesota to the south, from Central Ontario to the east, or both, based on dates of the presence of white pine in those areas (Figure 3.4). In addition to white pine, higher pollen values of *Quercus,* and *Juniperus* indicate a warmer and drier climate at this time (Björck 1985), which coincides with dates obtained from Hayes Lake for the Hypsithermal interval. White pine appears at both Indian Lake and Sioux Pond around 7,800 cal (7,000 ^{14}C) yr BP, and higher values of Compositae at Indian Lake between 8,200-7,400 cal (7,400-6,500 ^{14}C) yr BP indicate that prairie was also expanding at this time (Björck 1985). The pollen record first records *Picea,* Compositae*, Salix,* Cyperaceae, and Cupressaceae (cypress family) at Cristal Lake between 7,800-7,700 cal (7,000-6,900 ^{14}C) yr BP, indicating a spruce-dominated boreal forest. However, this segment of the record also records the highest pollen values which likely reflects the expansion of prairie vegetation in Manitoba and Minnesota (Björck 1985).

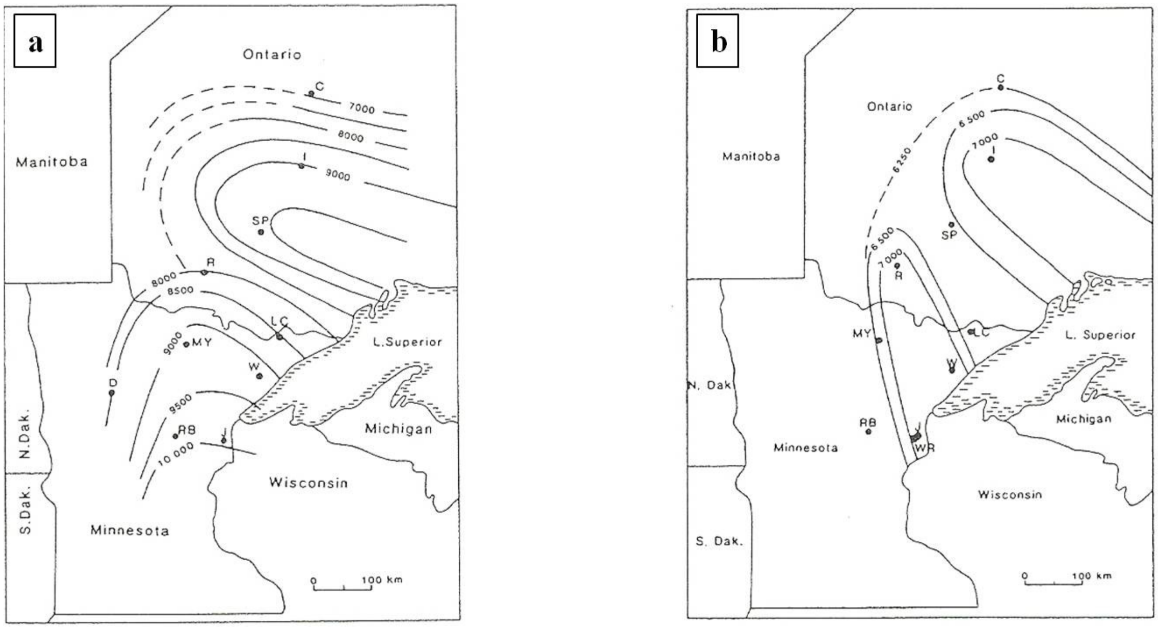


Figure 3.4. Diagram "a" represents the immigration of alder (*Alnus rugosa* and *crispa*) into Northwestern Ontario based on pollen records from northeastern Minnesota. LC= Lake of the Clouds (Craig, 1972); W= Weber Lake (Fries, 1962); MY= Myrtle Lake (Janssen, 1968); J=Jacobson Lake (Wright and Watts, 1969); RB= Rossburg Bog (Wright and Watts, 1969); and D= Bog D (McAndrews, 1966). Saarnisto`s (1974, 1975) sites on the northeastern shore of Lake Superior are also taken into consideration. Diagram "b" shows the immigration of white pine (*Pinus strobus*) into Northwestern Ontario between 7,800 cal (7,000 ^{14}C) yr BP and 7,200 cal (6,250 ^{14}C) yr BP. The dating is based on the four sites in Diagram "a", Willow River Pond (WR) (Jacobson, 1979), Jacobson`s (1979) mapping of white pine migration in Minnesota and Saarnisto`s (1974, 1975) studies on the northeast side of Lake Superior (from Björck 1975).

Climatic cooling occurs after the Hypsithermal period. At Sioux Pond this cooling occurs between 7,800-3,800 cal (7,000-3,500 ^{14}C) yr BP with the vegetation dominated by *Abies balsamea* while white pine and prairie species pollen decline (Björck 1985). Cooler and more humid conditions occur at Rattle Lake between 6,700-6,200 cal (5,900-5,400 ^{14}C) yr BP where *Ambrosia*, Amaranthaceae, *Pinus*, *Juniperus*, and *Quercus* pollen values decline while those of *Ostrya-carpinus, Picea, Larix* and *Abies* increase (Björck 1985). This climate reversal is noted at Indian Lake between 7,400-7,100 cal (6,500-6,100 ^{14}C) yr BP where *Picea* gradually replaces birch, white pine, and prairie pollen (Björck 1985). Cristal Lake experiences the cooling effects between 7,500-6,800 cal (6,600-6,000 ^{14}C) yr BP when white pine is replaced by both jack/red pine and birch, and *Picea* and *Abies* become more locally present (Björck 1985).

Two ponds were cored in association with archaeological site excavations at the Cummins site (DcJi-1), near Thunder Bay, to identify site stratigraphy and geomorphic process and to refine the paleovegetation record for this region (Julig et al. 1990). The Cummins site is located 11 km from the present day Lake Superior shoreline, and is associated with a glacial Lake Minong beach at 230-235 m asl (Julig et al. 1990). The Cummins Pond was a former lagoon or embayment of glacial Lake Minong. Oliver Pond, located 2 km northeast of Cummins Pond, is located at 250 m asl, which is above the highest elevation of Lake Minong (Julig et al. 1990).

Basal dates obtained from some of the cores appear too old based on the accepted deglaciation chronology (i.e., 13,892 ± 193 [11,950 ± 70 ^{14}C] yr BP and 13,012 ± 188 cal [11,080 ± 190 ^{14}C] yr BP) (Julig et al. 1990) and it remains debatable whether the Thunder Bay area was deglaciated prior to the Marquette advance, leaving the validity of these dates in question. As discussed in chapter 2 (Deglaciation and Paleohydrology), it is unlikely that this area was deglaciated prior to the Marquette advance since there was no evidence that Lake Agassiz drained into the Lake Superior basin (Teller et al., 2005), and Loope (2006) found that the Superior Lobe was present in the area prior to the advance. However, the following reconstruction has been made based on radiocarbon dates and pollen from these ponds. Initial vegetation following deglaciation was likely tundra however, it is unknown how long it lasted in the area (Julig et al. 1990) (see Figure 3.5 for Cummins and Oliver Ponds pollen diagrams). By 12,500 cal (10,500 ^{14}C) yr BP, spruce moved into the area and dominated the vegetation along with shrubs such as birch and soapberry (*Shepherdia canadensis*) (Julig et al. 1990). Values for herb pollen (sage and sedges) were still high at this time, suggesting that vegetation was probably sparse resulting in an open-tundra landscape with spruce in more protected areas (Julig et al. 1990). This changed around 11,500 cal (10,000 ^{14}C) yr BP, probably as a result of climatic warming (Julig et al. 1990). Increasing pollen values of *Pinus banksiana* and *Betula papyrifera* indicate that the spruce forest was giving way to jack pine and white birch however, it continued to dominate upland areas (Julig et al. 1990). *Salix*, *Populus balsamifera*, and grasses were likely located in glacio-lacustrine plains with *Ulmus*, *Fraxinus nigra*, and *Larix laricina* in wetland areas (Julig et al. 1990). However, as stated before, the pollen from the thermophilous deciduous species such as *Ulmus* and *Fraxinus nigra* may have been a result of long distance transport, with these species not actually present in the area at this time.

Around 8,900 cal (8,000 ^{14}C) yr BP, *Pinus* replaces *Picea* as the dominant vegetation cover and there is also an increase in *Betula* and *Alnus* pollen (Julig et al. 1990). Climatic warming continues during the Middle Holocene with the migration of white pine into the area around 7,200 cal (6,250 ^{14}C) yr BP, which is thought to have replaced the jack and/or red pine populations (Julig et al. 1990). Climatic cooling is observed around 4,200 ^{14}C BP with an increase in *Picea* pollen values and the replacement of white pine by jack and/or red pine. White birch, however, dominates the pollen assemblage (Julig et al. 1990).

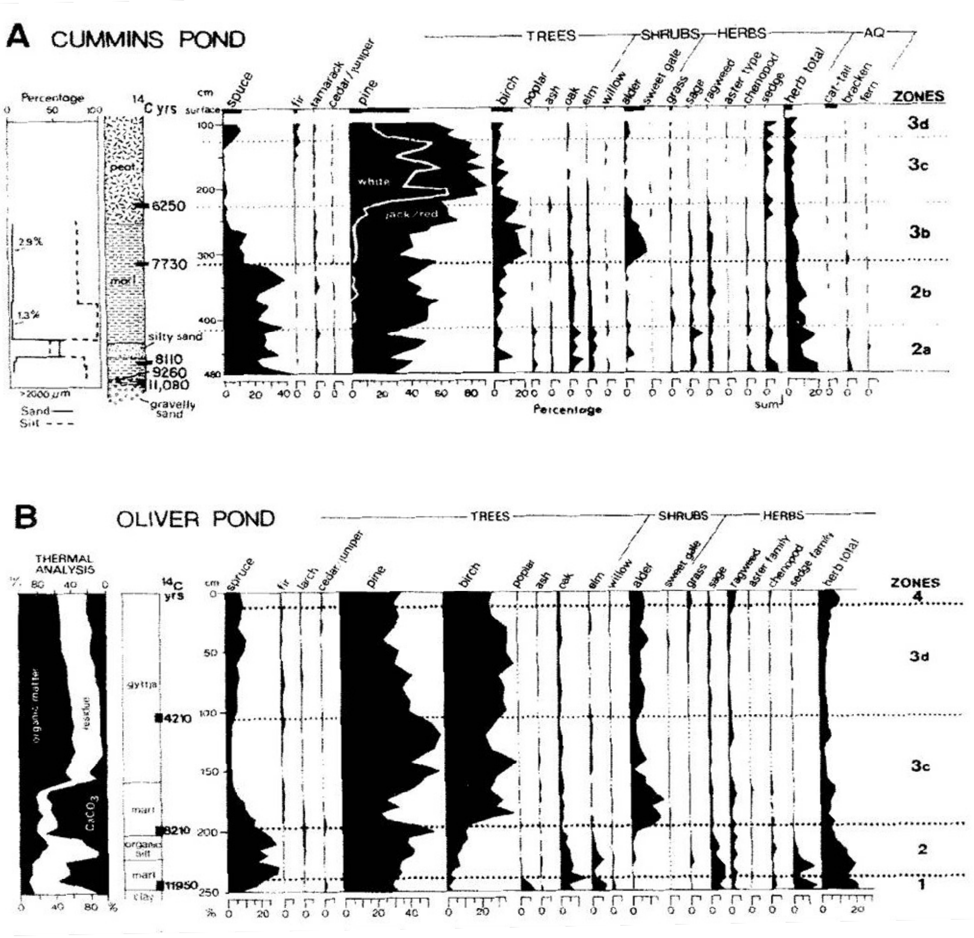


Figure 3.5. Pollen diagrams from the Cummins and Oliver Ponds. Early vegetation was herbs and shrubs indicating a tundra landscape (Zone 1), quickly followed by the migration of spruce (Zone 2). As the climate became drier and possibly warmer around 8,900 cal (8,000 ^{14}C) yr BP, jack pine and birch replaced spruce (Zone 3). Note that white pine moves into the Thunder Bay area ~7,200 cal (6,250 ^{14}C) yr BP. After 4,800 cal (4,200 ^{14}C) yr BP, spruce replaces pine as the climate becomes cooler and moister (from Julig et al. 1990).

Moos and Cumming (2011) analyzed the pollen records from multiple lake cores throughout Northwestern Ontario, including many previously discussed. They observed trends in pollen values and reconstructed vegetation and climate during the Early and Middle Holocene. They found that initial vegetation following the retreat of the LIS was spruce and deciduous trees. Pine, however, was quick to migrate into the area. The climate during this time was likely cool with low rates of evapotranspiration and precipitation. In the Early and Middle Holocene (8,600-4,500 cal yr BP), the climate became warmer, possibly by 1-2ºC relative to today, with an increase in precipitation. *Picea* pollen values remained low, but Ambrosia and Cupressaceae pollen values were high, indicating an open boreal forest. After 4,500 cal yr BP, the climate cooled. Pine pollen remained high however, *Picea* pollen values increased, possibly suggesting a closing of the forest canopy.

3.5. SUMMARY

Information gathered to reconstruct paleovegetaion is based on the relationship between vegetation and climate. Throughout the Holocene, climatic conditions changed from a late-glacial cold and dry period in the Early Holocene, to post-glacial warming of the Hypsithermal in the Early to Middle Holocene, and finally to cooler and moist/humid conditions in the Late Holocene. Climate fluctuations during the Holocene were mainly a result of the retreat of the Laurentide Ice Sheet, changes in atmospheric and thermohaline circulation, and the drainage of glacial Lake Agassiz. Although the climatic trends occurred over most of the northern hemisphere, it is important to note that the effects were not universal, as local factors also contributed to the fluctuations. These fluctuations in climate have been identified partly through changes in vegetation as revealed by pollen and macrofossils records, and also through carbon and oxygen isotope analysis.

The migration of vegetation into Northwestern Ontario was controlled by various factors, including deglaciation, presence of glacial lakes on the landscape, and climate. Initial vegetation in Northeastern Minnesota and Northwestern Ontario was tundra or forest tundra. In Northwestern Ontario, this tundra period was likely brief and was quickly succeeded by the migration of coniferous forest with some deciduous elements. After 8,900 cal (8,000 ^{14}C) yr BP, the warming and drying conditions of the Hypsithermal caused pollen assemblages to change as new species migrated north from Minnesota and west from central Ontario. Forest composition changed in most areas from coniferous forest to mixed conifer-hardwood. The migration of white pine around 7,800 cal (7,000 ^{14}C) yr BP into Northwestern Ontario and the expansion of prairie in Minnesota and Manitoba are also recorded. This warming lasted until approximately 6,800 cal (6,000 ^{14}C) yr BP when cooler and moister conditions prevailed. These cooling conditions began after 6,800 cal (6,000 ^{14}C) yr BP, but are most evident in palynological records around 3,200 cal (3,000 ^{14}C) yr BP in Northeastern Minnesota, 4,800 cal (4,200 ^{14}C) yr BP near Thunder Bay, 3,900 cal (3,600 ^{14}C) yr BP by Kenora, and around 4,500 cal (4,000 ^{14}C) yr BP in Northern Ontario. These dates, however, may also reflect localized environmental conditions and a lack of more refined radiocarbon dates from sediment cores.

Plant species migrated into Northwestern Ontario at different times throughout the Early and Middle Holocene. Tree species such as *Picea glauca, Larix laricina, Betula papyrifera* and *Pinus banksiana/resinosa* established themselves shortly after deglaciation, between about 11,500-10,700 cal (10,000-9,500 ^{14}C) yr BP, creating a boreal forest dominated landscape. *Abies balsamea* and *Thuja occidentalis* may have also been present in this early period, however, it is difficult to tell due to low pollen percentages for these species. As warming continued in the Middle Holocene (8,900-7,800 cal, 8,000-7,000 ^{14}C, yr BP) taxa including *Pinus strobus, Alnus crispa* and *Alnus rugosa* migrated into the area. The migration of vegetation from Northeastern Minnesota into Northwestern Ontario appears to be time transgressive. However, the migration of vegetation into the region west of Thunder Bay, including Fort Frances, Sioux Lookout and Kenora occurred earlier as this region was deglaciated much earlier than Thunder Bay, which was reoccupied by the LIS in the Early Holocene.

CHAPTER 4

EARLY AND MIDDLE HOLOCENE CULTURAL HISTORY OF NORTHWESTERN ONTARIO

4.1. INTRODUCTION

This chapter summarizes and critiques the archaeological literature pertaining to the Paleoindian and Archaic peoples who occupied Northwestern Ontario during the Early and Middle Holocene. The focus will be on timelines, migration patterns, material culture, subsistence patterns, and site distribution of these traditions. Discussion will also include the shortcomings of archaeological survey in this area that has likely led to biases and misinterpretations of that history.

Initial human occupation of the Thunder Bay region was by Paleoindian peoples after the LIS retreated from the area around 10,700 cal yr BP. Paleoindian peoples were highly mobile, using both local and regional resources to produce finely made tools. Numerous Paleoindian sites have been found in the Thunder Bay area, particularly near the glacial Lake Minong shoreline or taconite outcrops. Apart from lithic artifacts, limited archaeological material is recovered from these sites. By the Middle Holocene the Archaic tradition emerged with noticeable changes in the tool kit, particularly with the addition of native copper and woodworking tools. Not much is known about the local Archaic culture because there appears to be an underrepresentation of Archaic sites and materials dating to this timeframe. It is not clear whether this derives from biases in survey coverage, or significant climate change that profoundly affected the paleoenvironment (i.e., hydrology, vegetation). This may have resulted in sites being distributed in places that are not readily archaeologically accessible.

4.2. LIMITATIONS OF THE ARCHAEOLOGICAL RECORD IN NORTHWESTERN ONTARIO

Northwestern Ontario has been the subject of professional archaeological research since the 1960s (Wright 1963). Despite this, there are some serious gaps in our understanding of the ancient humans of this region. This is primarily due to four factors; 1) poor organic preservation, 2) shallow site stratigraphy (contributing to taphonomic problems), 3) survey biases, and 4) survey limitations. Firstly, very little organic material (e.g.,wood, bone, antler) dating to Paleoindian and Archaic times has survived in the archaeological record. This is due, in part, to the acidic soil in the boreal forest which destroys organic material. Lithic artifacts, such as projectile points, tend to be the only archaeological material recovered from sites (Hamilton 2007; Wright 1972). As a result, valuable cultural information related to subsistence resources, transportation technology and dwellings is consistently lost (Wright 1972). With regards to the subsistence economy, assumptions are made using environmental reconstructions coupled with extrapolations about what sorts of animals preferred those particular environments based on animal behavior today and analogies based upon subsistence strategies of historic aboriginal groups (Dawson 1983a; Wright, 1995). Using historic groups as models for early subsistence patterns may not be very informative as it assumes that similar cultural, technological, and environmental conditions occurred in the past, which is unlikely. For example, the climate changed throughout the Holocene, causing vegetation to migrate and likely animals also; this means different resources and resource densities may have been available in the past. Historic groups also possessed different technology, including pottery, and consumed plants, such as wild rice, which may not have been available to Paleoindian and Archaic groups. Additionally, it is also not known if ancient people consumed all of the available resources to them, nor is it always possible to know if they possessed the proper technology to obtain them.

Secondly, site stratigraphy in this region does not generally allow for clear separation of archaeological components (Dawson 1983a, 1983b). Sediment accumulation is very slow, resulting in thousands of years of deposition represented in relatively shallow deposits (Phillips 1993). These thin soil accumulations make separating cultural layers difficult (Dawson 1983a, 1983b). As a result, Paleoindian and Archaic components at a given site are often mixed. This makes it difficult to determine if a site was re-occupied later by Archaic people or if the site represents a transition period where both technologies were used. In addition to these issues, archaeological material is often heavily disturbed due to tree root penetration, tree throws, rodent holes, freeze-thaw cycles or destruction by forest fires (Phillips 1993). The lack of datable organic material also makes it difficult to establish cultural chronologies this issue even if stratigraphic separation exists (Phillips 1993).

Thirdly, the search for archaeological sites tends to suffer from systematic bias such as the focus on Minong paleoshorelines for the Paleoindian period (Figure 4.1) (Hinshelwood 2004). This trend is repeated in southern Ontario where there has also been an archaeological focus on glacial Lake Algonquin, Lake Ardtrea, and Nipissing strandlines (Ellis and Deller 1997; Jackson et al. 2007) in searches for early archaeological sites. This may be in part, due to the fact that ancient shorelines are often easily recognized and followed (Phillips 1993). In contrast, dense boreal forest restricts visibility and may make getting to sites in other geomorphic contexts much more difficult. These shoreline sites then contribute most of the archaeological record, producing a bias, as survey for inland sites, such as those around inland lakes, rivers, and wetlands, is minimal (Phillips 1988). However, some studies (Anderton et al. 2004; Boyd 2007; Phillips 1988), suggest that inland wetland areas may have been more productive than shorelines of cold glacial lakes, making these areas more attractive to Paleoindian and Archaic peoples and therefore optimal places to examine for archaeological sites.

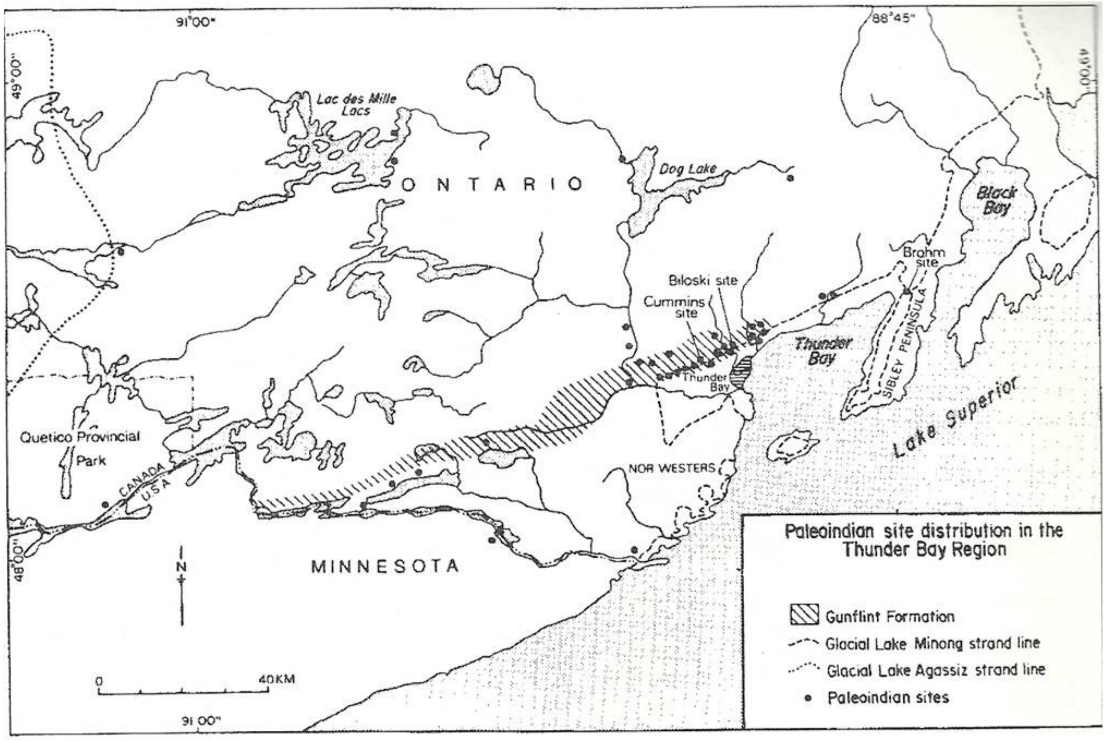


Figure 4.1. Location of known Paleoindian sites in proximity to the Lake Minong shoreline and the Gunflint Formation (from Julig et al. 1990).

Lastly, although there have been multiple Paleoindian and Archaic sites located, very few have been excavated (Hinshelwood 2004; Julig et al. 1990). Many of the projects are assessment or reconnaissance operations and it is rare that a site can be fully excavated (Hinshelwood 2004). In part, this reflects a lack of funding and problems with access to land, whether private or public, as well as time constraints that force a primary focus on surface collection (Julig et al. 1990; Phillips 1993). Many sites also show evidence of looting or are disturbed by road or building construction or from quarrying operations (Arthurs 1989).

4.3. EARLY PALEOINDIAN TRADITION: CLOVIS AND FOLSOM

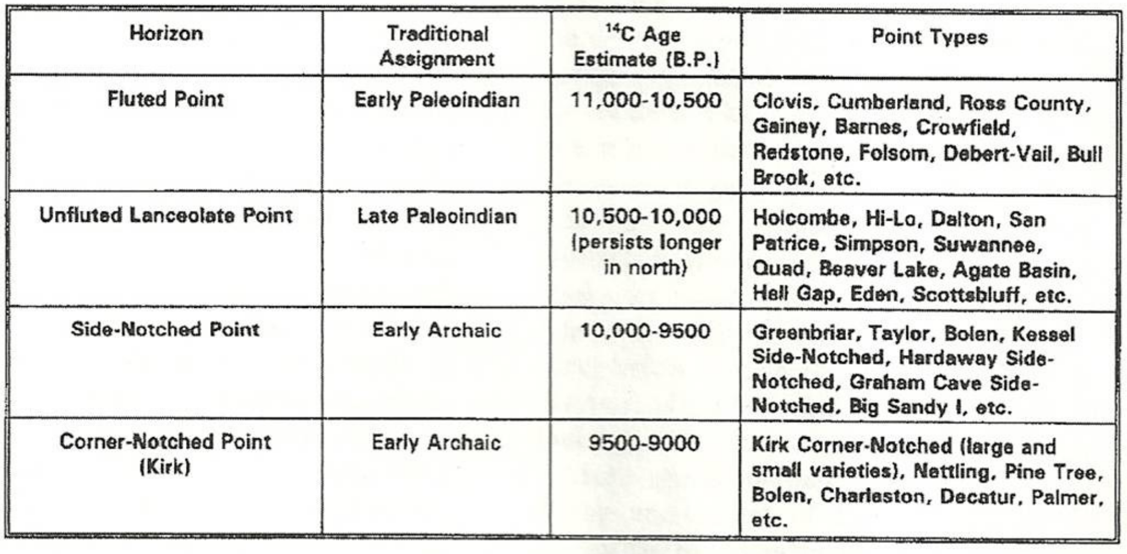
The timing of initial migration of people to the Americas is continuously being challenged. It was originally thought that the first people to occupy the Americas were associated with the Clovis culture (Meltzer 2009). As discussed below, the oldest dates from this culture are around 13,500 cal (11,200 ^{14}C) yr BP. However, archaeological excavations at the Meadowcroft shelter in Pennsylvania produced a date of ~17,200 cal (˜14,000 ^{14}C) yr BP for human occupation (Adovasio et al. 1990). The Monte Verde site in Chile also provides evidence of early human occupation of the South American continent as early as 14,800 cal (12,500 ^{14}C) yr BP (Dillehay 1997). These sites attest to the presence of a pre-Clovis culture in the Americas. However, since pre-Clovis studies are in their infancy and excavation of pre-Clovis sites is continuously scrutinized, not much is known about this culture (Meltzer 2009).

The Clovis culture has been identified throughout North America and dates between 13,500-12,800 cal (11,200-10,700 ^{14}C) yr BP (Holliday 2000). This culture is identified by their fluted projectile points (Figure 4.3). These points are relatively long and thick; they have thinning flakes chipped off which extend from the base and terminate prior to the mid-point (Meltzer 2009:243). The scar(s) left behind are called flutes however, the purpose of flutes is unknown. The Folsom culture is thought to have replaced Clovis in central North America, particularly the Great Plains. It was present between 12,800-11,500 cal (10,900-10,000 ^{14}C) yr BP (Holliday 2000). Folsom points are generally smaller and thinner than Clovis points however, fluting generally extends beyond the middle of the point (Meltzer 2009:295). Hypotheses as to why fluting may have changed include an adaptation to a changing environment and food sources, to facilitate hafting onto a spear, to allow for deeper penetration into prey, ceremonial purposes, or decoration (Ahler and Geib 2000). Ahler and Geib (2000) suggest that the likeliest scenario, at least for the later Folsom fluted points, would be purely for hunting purposes.

Early Paleoindian people have been traditionally thought of as big game hunters, hunting megafauna and large mammals such as caribou and bison, resulting in a specialized subsistence pattern (Wright 1972). This theory was based on the numerous Clovis and Folsom projectile points found in association with mammoth and bison remains (Cannon and Meltzer 2008). Those claims have since been dismissed as it has become apparent that not all early Paleoindians consumed or relied on the same resources (Cannon and Meltzer 2008). Faunal remains recovered from early Paleoindian sites include remains from both terrestrial and aquatic resources and incorporate various sized game including turtle, fox and fish (Meltzer 2009). Cannon and Meltzer (2008) examined the faunal remains from multiple early Paleoindian sites throughout the United States and one in Canada, and they found that Paleoindians in northeastern North America likely hunted caribou as their main food staple, while those in the Mid-west consumed larger animals such as the mammoth. They also found evidence to support that northeastern North American Paleoindians consumed a range of other fauna in addition to caribou, resulting in greater dietary diversity than areas to the west. They suggest that the broader diet of the eastern Paleoindians may be a result of the environment, as pollen records indicate a mosaic or patch-like environment that resulted in different vegetation types. These patches would likely have attracted different animals which the humans were able to hunt (Cannon and Meltzer 2008). Interestingly, the same may be said for northwestern Ontario as it is located near three ecozones: the boreal Forest, mixed conifer-hardwood forest, and prairie. Climatic fluctuations which caused the boundaries of these ecozones to shift, coupled with the high mobility of Paleoindian groups, suggest that early human occupants of this region also consumed a variety of resources. This will be addressed later in the chapter.

As mentioned in Chapter 3 (Paleovegetation and Climate), the climate deteriorated abruptly around 12,900 cal (11,000 ^{14}C) yr BP, lasting until approximately 11,500 cal (10,000 ^{14}C) yr BP (Hu et al. 1999). The end of this climatic event, the Younger Dryas, roughly coincides with the terminus of the early Paleoindian fluted point traditions in northeastern North America (Newby et al. 2005). An analysis of Paleoindian points and regional pollen records from New England and the Canadian Maritimes reveals that fluted points were present prior to the Younger Dryas but not after it (Newby et al. 2005). The pollen records indicate that during the Younger Dryas, the landscape was dominated by patches of conifer woodlands and open tundra-like vegetation, which Newby et al. (2005) suggest would have been the preferred environment of migratory animals such as caribou. After the Younger Dryas, the landscape changed to a mixed conifer-deciduous forest that would have favoured solitary animals such as deer and moose (Newby et al. 2005). These changes, including an increase in white pine, also indicate a warmer and drier climate, characteristic of the Early Holocene (Newby et al. 2005).

Newby et al. (2005) suggest that there is a direct relationship between the observed climate and vegetation changes and the abandonment of fluting on projectile points and the development of more regional projectile point variations (e.g., Holocombe, Hi-Lo, Dalton, Plainview, Agate Basin, Hell Gap, Scottsbluff and Eden) of the emerging late Paleoindians traditions (Figures 4.2 and 4.3).



Horizon	Traditional Assignment	^{14}C Age Estimate (B.P.)	Point Types
Fluted Point	Early Paleoindian	11,000-10,500	Clovis, Cumberland, Ross County, Gainey, Barnes, Crowfield, Redstone, Folsom, Debert-Vail, Bull Brook, etc.
Unfluted Lanceolate Point	Late Paleoindian	10,500-10,000 (persists longer in north)	Holcombe, Hi-Lo, Dalton, San Patrice, Simpson, Suwannee, Quad, Beaver Lake, Agate Basin, Hell Gap, Eden, Scottsbluff, etc.
Side-Notched Point	Early Archaic	10,000-9500	Greenbriar, Taylor, Bolen, Kessel Side-Notched, Hardaway Side-Notched, Graham Cave Side-Notched, Big Sandy I, etc.
Corner-Notched Point (Kirk)	Early Archaic	9500-9000	Kirk Corner-Notched (large and small varieties), Nettling, Pine Tree, Bolen, Charleston, Decatur, Palmer, etc.

Figure 4.2. Point style, associated tradition, age and point type of projectile points common in eastern North America (from Ellis et al., 1998).

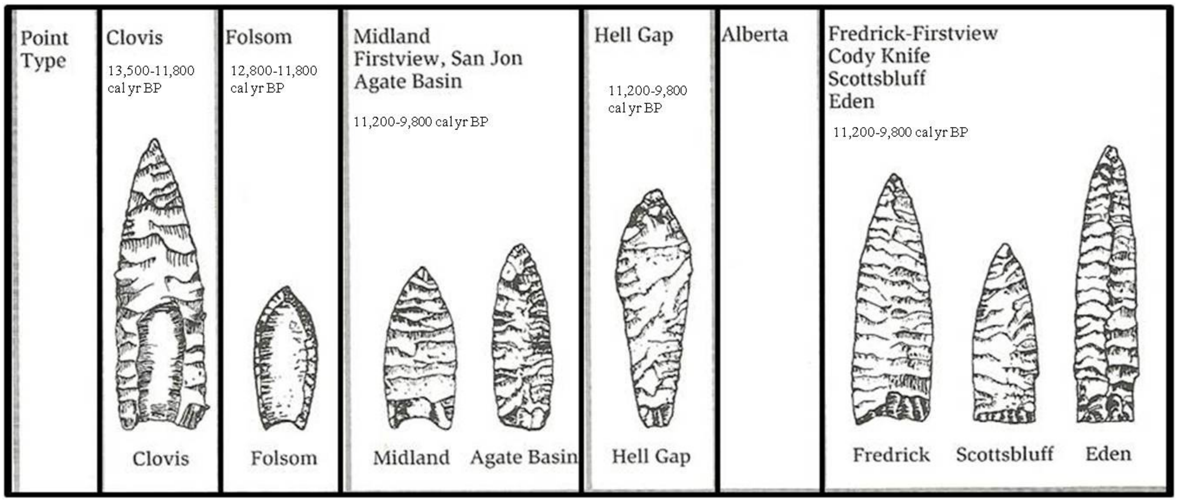


Figure 4.3. Projectile point types typically found on the Plains. Note the fluting on Clovis and Folsom points (from Fagan 2000:97, point type dates updated from Holliday 2000).

As of yet, no fluted points have been found in Northwestern Ontario (Phillips 1993). The relatively late deglaciation (~10,700 cal yr BP) of this area, particularly around Thunder Bay, may account for the lack of fluted points. It is possible that early Paleoindians may have been present in Northwestern Ontario based on the deglacial history, however, a lack of fluted points may indicate that the areas proximity to the retreating ice front and the presence of glacial lakes created an environment that was not suitable for substantial or extended human occupation. It could also be that archaeological survey and excavation have not focused on areas where early Paleoindians may have been.

The closest evidence of early Paleoindians has come from Minnesota. Mulholland et al. (1997) state that there have been two early Paleoindian points found in northern Minnesota and one in northern Wisconsin. A Clovis point was reportedly found near Duluth, and what is believed to be a Folsom point was recovered near Round Lake. These points were both made from gunflint silica, which comes from sources throughout northeastern Minnesota as well as in the Thunder Bay area (Mullholland et al. 1997). The point from northern Wisconsin is a Clovis point manufactured from Hixton silicified sandstone, which originates from a source in that state.

4.4. THE LATE PALEOINDIAN (PLANO) TRADITION IN NORTHWESTERN ONTARIO 10,700-7,800 cal (9,500-7,000 ^{14}C) yr BP

It is generally thought that Paleoindian peoples entered Northwestern Ontario from Minnesota and Wisconsin as the LIS retreated from the area around 11,500 cal (10,000 ^{14}C) yr BP (Phillips 1993). The timing of initial human occupation, however, has been a subject of debate. Phillips (1993) and Phillips and Hill (2004) suggest that early

Paleoindians using fluted points or late Paleoindians using unfluted points could have migrated into the area prior to the Marquette advance around 11,500 cal (10,000 ^{14}C) yr BP. However, others (Dawson 1983a; Wright 1995) suggest that occupation occurred sometime after 11,500 cal (10,000 ^{14}C) yr BP, with the final retreat of the Laurentide Ice Sheet. Phillips and Hill (2004) suggest that the most potential for archaeological evidence of early Paleoindians is in the Interlakes Corridor, an area located north of the Canadian/US border between Ontario and Minnesota, between Quetico and Thunder Bay (Figure 4.4) (Phillips and Hill, 2004).

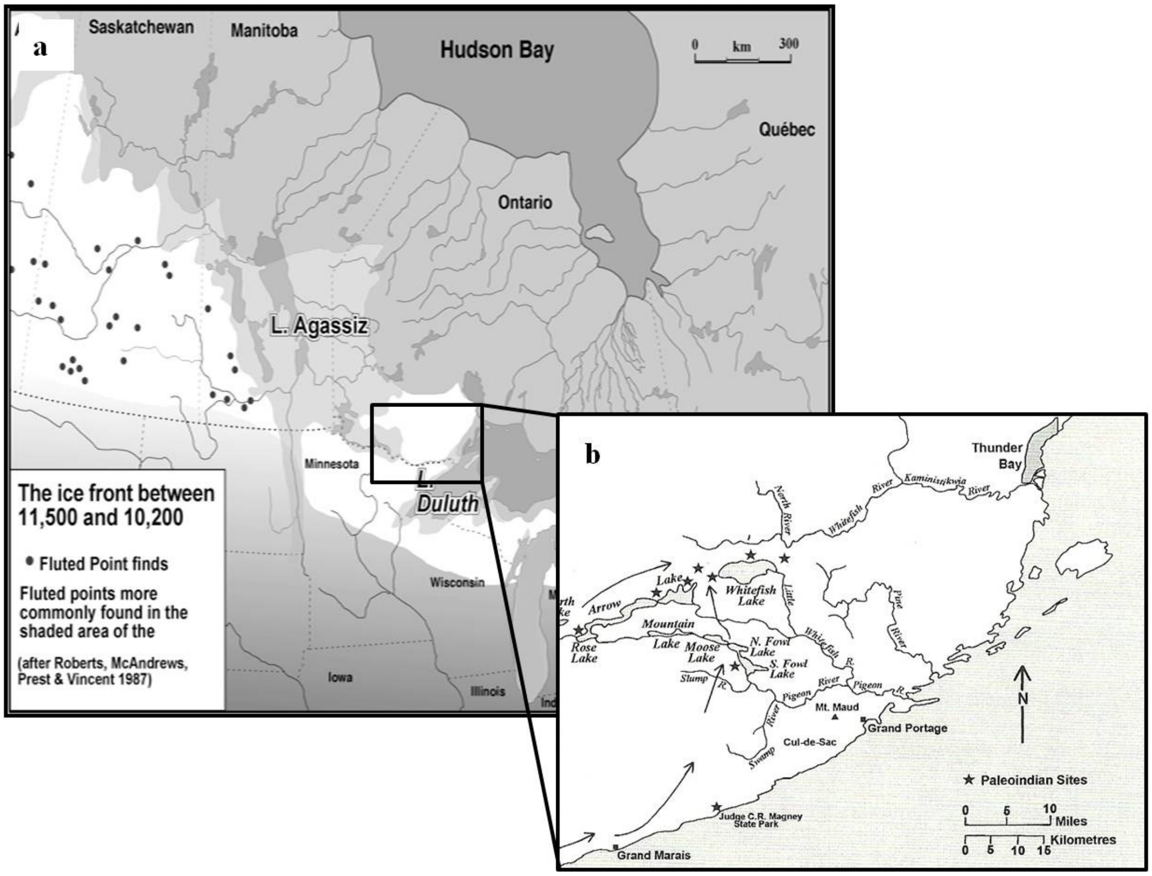


Figure 4.4. Location of the ice front and glacial lakes between 13,400-11,900 cal (11,500 and 10,200 ^{14}C) yr BP, revealing a "cul-de-sac" in the Gunflint-Arrow-Whitefish lakes area where evidence of early Paleoindians may be found (map "a" from Hamilton 2010, map "b" from Phillips and Hill 2004).

A problem with the “corridor” hypothesis is that there has not been any archaeological evidence to suggest that people were in the area at this time. Phillips and Hill (2004) suggest that evidence of their presence may be found on the highest shorelines, those associated with glacial Lakes Duluth or Beaver Bay, or further inland from them (Phillips and Hill 2004). A paleoshoreline has been found near Kakabeka Falls and is thought to represent the pre-Minong Lake Beaver Bay level within the Lake Superior basin, however, there are no dates associated with it, nor have early Paleoindian artifacts been found upon it (Phillips and Hill 2004).

Support for an early partial deglaciation west of Thunder Bay comes from the moraines located throughout this area. As discussed in Chapter 2 (Deglaciation and Paleohydrology), Lowell et al. (2009) used lake cores to date the formation of the recessional moraines deposited by the LIS. If accurate, they indicate that the LIS was located at the Lac Seul moraine during the Marquette advance, with the Superior Lobe located at the Marks moraine west of the city of Thunder Bay. Loope (2006) suggests that the Superior Lobe was likely located within the Thunder Bay area prior to the advance, meaning some parts of Northwestern Ontario may have been open before 11,500 cal (10,000 ^{14}C) yr BP. Although the LIS may have been absent, numerous glacial lakes may have been present as the ice sheet waned, and the corridor was likely between the Superior Lobe of the LIS and glacial Lake Agassiz, creating a cold environment that may not have been hospitable to humans. Regardless of the climatic conditions or the presence of glacial lakes, it remains possible that part of Northwestern Ontario was deglaciated prior to the Marquette advance, and the areas highlighted by Phillips (1993) and Phillips and Hill (2004) may have been available for early human occupation

The first evidence of human occupation in Northwestern Ontario is following the Marquette advance, as many Paleoindian sites have been found on glacial Lake Minong shorelines which formed before 10,700 cal (9,500 ^{14}C) yr BP (Julig et al. 1990). Not much is known about the late Paleoindians in Northwestern Ontario however, some diagnostic lithic artifacts, including projectile points, have been recovered from archaeological sites in this region. This late Paleoindian culture is is referred to as the Plano.

4.4.1. Plano Technology

Typical Plano projectile points from this area are lanceolate in shape, parallel or obliquely flaked (Dawson 1983b; Hamilton 1995), and exhibit grinding on the lower lateral edges (Hamilton 2007). Although taconite was the predominant choice, other lithic materials such as Knife River flint, Hixton silicified sandstone, quartzite, Hudson Bay Lowland chert, dolomite, mudstone and shale have all been reported from archaeological sites in the area (Hinshelwood 2004; Hinshelwood and Webber 1987; McLeod 1980; Wright 1972).

There appears to be a variety of different Plano point types that have been found in Northwestern Ontario, including Agate Basin, Hell Gap, Scottsbluff and Eden, Plainview and Minocqua (Figure 4.3) (Hinshelwood 2004). However, dating these points has proven difficult for this region. Since radiocarbon dates from Paleoindians sites are either non-existent or of questionable accuracy, the dates assigned to the points reflect those from Paleoindians sites in the Northern and Southern Plains, eastern North America, and the Great Basin. Ellis et al. (1998) suggest that these points are found in the above regions from around 12,500 cal (10,500 ^{14}C) yr BP until 11,500 cal (10,000 ^{14}C) yr BP however, they also suggest that some of them persisted longer, perhaps until 8,900 cal

(8,000 ^{14}C) yr BP in the upper and northern Great Lake regions, including Northwestern Ontario, Wisconsin and Minnesota.

In the Thunder Bay area, Plano sites were assigned to the "Lakehead Complex" (Fox 1975). A site was assigned to this complex if: 1) there were Plano artifacts found there 2) those artifacts were manufactured predominately from Gunflint Formation taconite and cherts and 3) they were associated with the glacial Lake Minong shoreline (Hinshelwood 2004). Similar complexes such as Reservoir Lakes, Rainy River, Quetico, and Flambeau and Minocqua phases have been identified along the eastern Manitoba border and around the southern shore of Lake Superior in Minnesota and Wisconsin (Ross 1995). Although these phases were originally grouped under their own complex designation, Ross (1995) suggests that the similarities between the archaeological assemblages should result in them being classified together under what he terms the "Interlakes Composite" (Figures 4.5, 4.6 and 4.7).

What is unique about the Interlakes Composite is that archaeological sites found within this large area exhibit more than one projectile point type and include a range of lithic material selection (Ross 1995). These point types have also been found in the same cultural level (Ross 1995). Due to the diversity of the point types, Hinshelwood (2004) suggests that it may reflect temporary occupation of this area by many different groups of people. The use of local taconite and exotic material such as Knife River Flint and Hixton silicified sandstone could indicate either trade with other Paleoindian groups or the migration of Paleoindian groups into Northwestern Ontario from those areas (Wright 1972). With the deglaciation of Northwestern Ontario, it is feasible that Paleoindian groups ultimately deriving from the Plains and areas of eastern North America such as Wisconsin, could have arrived in Northwestern Ontario at the same time, resulting in a mixture of point styles and lithic material. Unfortunately, with the limited radiocarbon dates from Paleoindian sites in this area, it has yet to be determined if the above scenario is true or if the varying point types reflect the diffusion of ideas and technology to the Plano peoples in Northwestern Ontario from the Plains and eastern North America.

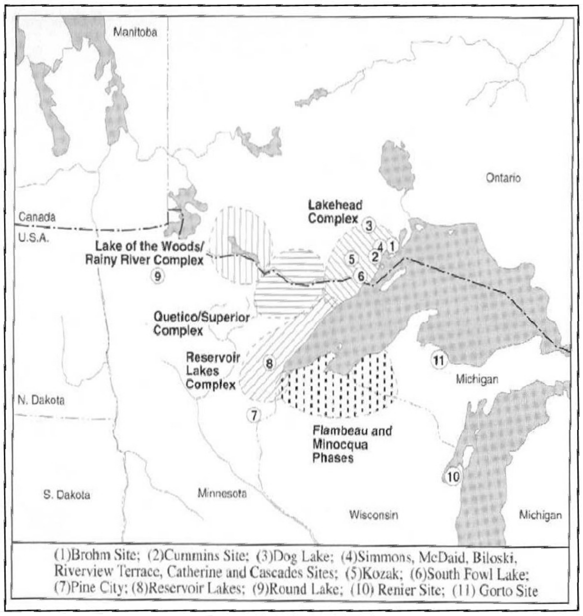


Figure 4.5. The complexes now amalgamated into the "Interlakes Composite" based on similarities in Late Paleoindian toolkits (from Ross 1995).

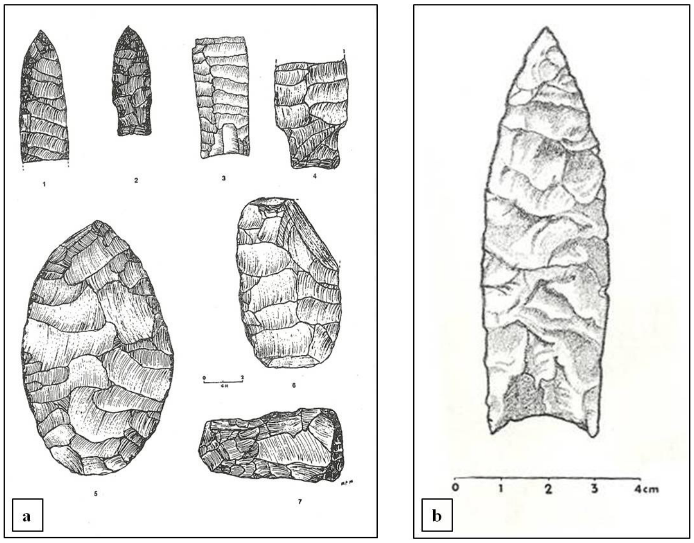


Figure 4.6. Late Paleoindian artifacts found within the Lakehead Complex. Figure 'a' 1= lanceolate point from Boulevard Lake (DcJh-4), 2= point from the Cummins site (DcJi-1), 3= Parallel flaked projectile from the Cummins site (DcJi-1), 4= Scottsbluff point from Arrow Lake (DaJn-7), 5 and 6= biface from the Cummins site,

7= end scraper from the Cummins site (from Dawson 1983a). Figure 'b' has been identified as a Plainview point from the Brohm site (DdJe-1) (from Steinbring 1976).

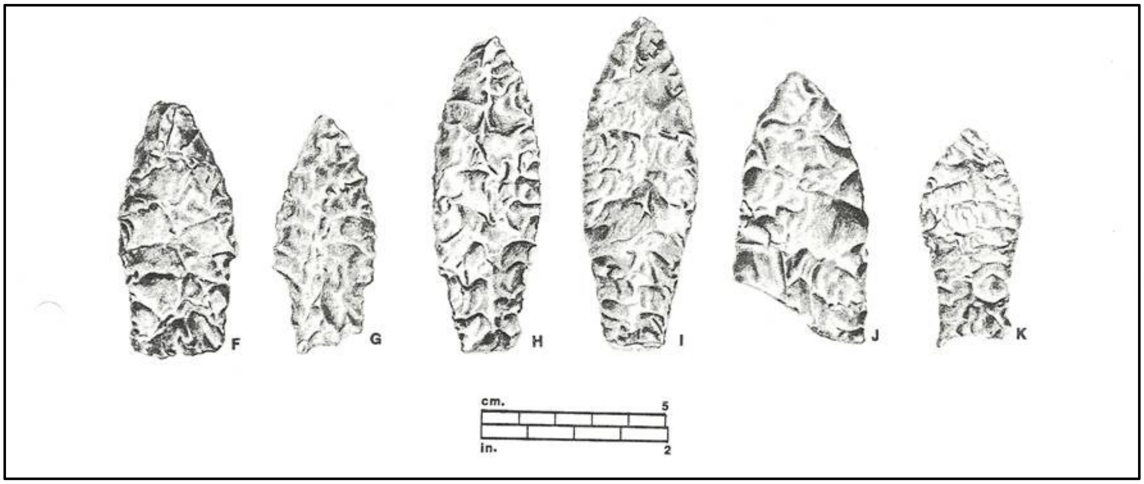


Figure 4.7. Paleoindian points from the Reservoir Lakes Complex, near Duluth, Minnesota. F and G= Scottsbluff and related forms, H,I,K= Hell Gap, J= jasper taconite biface (from Steinbring, 1976, Figure 2).

In addition to projectile points, scrapers, knives, burins, drills, hammerstones, choppers, and anvils can all be found within the Plano toolkit (Wright 1995) and have been found in archaeological sites in Northwestern Ontario (Dawson 1983a). Due to poor organic preservation, excavation techniques, and a focus on quarry and shoreline sites, not much is known about Plano housing (Dawson 1983a). It is also not known if they possessed watercraft technology or snow and ice transportation however, it is believed that watercraft were likely used since many sites have been located near waterways (Wright 1995).

4.4.2. Plano Subsistence Economy

Evidence of Early and Middle Holocene food resources are scarce in Northwestern Ontario (Julig 1984), partly due to poor organic preservation in this region as discussed above. What organic remains (e.g., bone, protein, antler, residue) that are found are pieced together with environmental reconstructions to determine what faunal and floral resources may have been available at the time (Wright 1995). There have yet to be any positively identified faunal remains found in any Paleoindian site in Northwestern Ontario (Julig 1984), so information about subsistence has come from other areas and reflects speculation based on archaeological site location (i.e., near possible past caribou crossings).

Some burned bone fragments have been found at the Cummins site (DcJi-1). However, only one bone was tentatively identified as likely belonging to a caribou (Julig 1984). Blood protein residue obtained from lithic tools found at the Cummins site reveals the presence of bovid (musk oxen or bison), rodentia (muskrat, beaver etc.), and cervid (caribou, moose, deer, wapiti) (Newman and Julig 1990). These residues indicate that large, medium, and small mammals were being hunted and/or consumed, providing evidence supporting a generalized subsistence economy (Newman and Julig 1990).

Since multiple early Paleoindian sites elsewhere in North America indicate use of a variety of resources (Meltzer and Cannon 2008; Newby et al. 2005; Kuehn 1998), this may also hold true for late Paleoindians in Northwestern Ontario. Pollen records from Cummins and Oliver Ponds indicate mainly sparsely vegetated tundra with spruce in protected areas between 12,500-11,500 cal (10,500-10,000 ^{14}C) yr BP, gradually becoming a closed spruce forest around 8,900 cal (8,000 ^{14}C yr BP) (Julig et al.1990). Caribou may have frequented this area during this time however, with the closing of the forest, deer and moose may have replaced caribou as a primary staple food in the Plano diet, much like is suspected of occurring in eastern North America and the Canadian Maritimes following the Younger Dryas (Newby et al. 2005). Since Plano people are thought to have travelled long distances, it may be possible that the Plano peoples in

Northwestern Ontario travelled to Minnesota or Manitoba to hunt bison; this could account for the bovid blood residue found on lithics from the Cummins site.

Large game were likely not the only resources consumed by the Plano. Other resources such as fish and birds (Wright 1972), beaver and plants were also probably consumed (Julig 1984) however, traps or snares made out of organic material would likely not be preserved, so evidence that these resources were consumed would not survive in the archaeological record (Phillips 1993; Wright 1972).

4.4.3. Plano Site Distribution in Northwestern Ontario

Plano sites in Northwestern Ontario tend to be associated with certain geographic features including bedrock outcrops and paleoshorelines (Julig et al. 1990). As previously mentioned, this may be attributed to biases in site survey, with these locations being more recognizable and/or more accessible. Regardless of the biases, particular features have yielded multiple archaeological sites within close proximity to one another.

The Gunflint Formation is a prominent aspect of the geology in the Thunder Bay region, and is exposed in a linear configuration running south-west from the city of Thunder Bay as far as Gunflint Lake (Figure 4.1) (Hinshelwood 1987). This formation is taconite rich and appears to be the lithic material of choice for local Paleoindians, probably due to its abundance in the area (Hamilton 1996; Phillips et al. 2001). Taconite, however, is relatively difficult to work into flakes and tools because it is prone to shatter along fault planes hidden within it (Hamilton 1996). To combat this, it is thought that Paleoindians heated the taconite, rendering it easier to work (Wright 1995). Evidence to support this comes from ‘pot-lid’ scarring, a pattern of circular chips that occurs during the heating process and is found on some taconite artifacts (Ross 1979). Many sites including the Cummins Site are found within easy walking distances from outcrops, allowing for quarrying and easy transport to workshop sites (Julig et al. 1990; Phillips et al. 2000).

Due to glacial advances and retreats, climate change, and isostatic rebound, relative water levels of the Great Lake basins, including glacial Lake Agassiz, have fluctuated over time (Boyd et al. 2010; Farrand and Drexler 1985; Lewis and Anderson 1989; Lewis et al. 2007). These fluctuations created a series of beaches or other shoreline features such as deltas (Phillips 1988). These environments may have been ideal for Paleoindian occupation in part due to the availability of fresh water and the various plants and animals attracted to it (Julig et al. 1990). In the Thunder Bay region, glacial Lake Minong beaches have yielded a particularly large number of Paleoindian sites, including Brohm, Simmonds, McDaid, Biloski, Catherine, Rosslyn, and Cummins sites (Figure 4.10) (Phillips 1988, 1993).

Inland wetlands, riparian zones and lakes would also have been areas with considerable biodiversity, and may have been key locations for human occupation. However, with the exception of Dog Lake, these areas have been relatively neglected by archaeologists (Phillips 1993). Since the environment may have been different during the Paleoindian timeframe, paleoenvironmental reconstructions may be needed to locate past locations of such areas prior to archaeological survey.

Hinshelwood (2004) suggests that it was the combined presence of workable lithic material and areas of high relief that drew Paleoindians to certain locations. For example, the sites mentioned above are also located near outcrops of the Gunflint Formation, so it is difficult to deduce if it was access to lithic material, the shoreline, or both that enticed them to these locations. Although archaeological material is found on shorelines, there is also debate around whether people were there while the beach feature was active, or after the lake levels dropped. The waterworn artifacts from the Minong beach at the Cummins site provide supporting evidence that these beaches were in fact occupied while active (Julig et al. 1990).

According to Hinshelwood (2004), the majority of Paleoindian sites in the Thunder Bay area are clustered in four locations (Figure 4.8). The first grouping Hinshelwood (2004) identifies is the Neebing Series, which includes Cummins (DcJi-1), Cummins West (DcJi-16), Widar (DcJi-5), Quackenbush (DcJi-3), and Stetko (DcJi-7), among others. These sites are found within the Neebing River drainage basin and are on the Kaministiquia River delta but are more closely associated with beaches produced by glacial Lake Minong. The Cummins site (DcJi-1) is the most well known and has been one of the most extensively excavated and studied sites in the region.

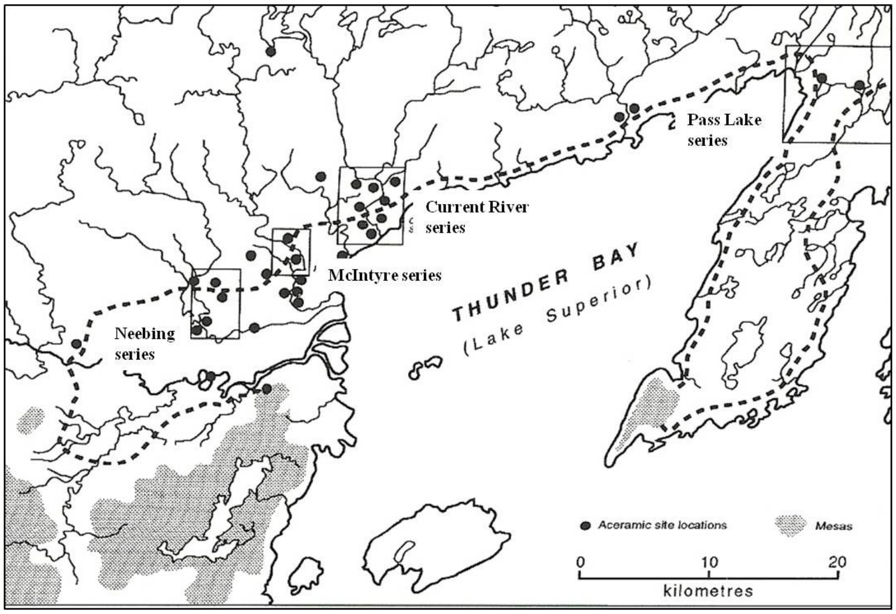


Figure 4.8. Location of the four Paleoindian site clusters in relation to their associated watersheds. Note the location of sites and their proximity to the Minong shoreline, which is shown by the dashed line (from Hinshelwood 2004).

The Cummins site was first excavated in 1963 by J.V. Wright and K.C.A. Dawson and again re-examined and excavated in 1983 by Patrick Julig (Julig 1984). The site is located on a Minong level beach and is also near a Gunflint Formation outcrop (Julig et al. 1990). An abundance of lithic artifacts attributed to both Paleoindian and Archaic traditions were excavated from the site and suggest it was used as a quarry, lithic workshop and encampment for a long period of time (Julig et al. 1990). What is unique about this site is a partial cremation burial radiocarbon dated between 10,012-8,993 cal (8,480 ± 390 ^{14}C) yr BP (Julig et al. 1990); this is the only absolute date from the site (Julig 1984). This date, however, provides a relatively wide temporal range and may not be that informative. Geologic examination of the site, however, establishes initial occupation around 10,700 cal (9,500 ^{14}C) yr BP, one thousand years before the burial, besed on waterworn artifacts located under beach gravel and under Archaic and Plano occupation levels (Julig 1984). The site, however, continued to be occupied by Paleoindian and Archaic people despite the fact that the lake levels declined after 10,600 cal (9,400 ^{14}C) yr BP (Julig 1984).

The second Plano site cluster is associated with the McIntyre River. Most material is reported from three sites: Biloski (DcJh-9), McIntyre (DcJh-12) and River Terraces South (DcJh-38) (Hinshelwood 2004). Like many Paleoindian sites in the Thunder Bay area, the Biloski site is situated on a Lake Minong shoreline (Hinshelwood and Webber 1987). Agate Basin-like and Minoqua points were found at this site, placing it within the same general occupation timeframe as the Cummins site (Hinshelwood and Webber 1987). River Terraces South yielded a projectile point base made from Hixton silicified sandstone, an exotic material from Wisconsin (Hinshelwood 2004).

The third cluster is located near the Current River (Hinshelwood, 2004), and includes the Simmonds (DcJc-4), McDaid (DcJh-16), Chairs Quarry (DcJh-40), Happy Days (DcJh-39), and Naomi sites (DcJh-42) (Hinshelwood 2004). These sites are either situated along the Current River or at the Minong phase location of the river mouth (Hinshelwood 2004). At the Simmonds site a Plainview projectile point made from taconite and a biface were found, and at the Naomi site, a Hixton silicified sandstone point base was recovered (Hinshelwood 2004). The McDaid site, located on the Minong shoreline (240-236 m asl), yielded Agate Basin-like bifaces (or preforms), some of which were parallel flaked with convex bases (Stewart et al. 1984).

The last cluster is located near Pass Lake, on or near the northern portion of the Sibley Peninsula. These sites are also located near Lake Minong shorelines (Hinshelwood 2004). There are only six sites included in this series, most notably Brohm (DdHe-1), Bak (DdHe-2), LeGarde (DdHe-3), and Andersen (DdHe-4) (Hinshelwood 2004). The Brohm site was the first Paleoindian site excavated in Northwestern Ontario (Hinshelwood 1988). It is suspected that the area was a roaming or calving ground for caribou, although there is no evidence to support this (Hinshelwood 1990). Quimby (1959) states that Paleoindians occupied the site shortly after the terraces and beaches it is situated on were deposited, and, based on projectile point typologies, he suggests it was likely occupied from 10,200 cal (9,000 ^{14}C) yr BP to 7,800 cal (7,000 ^{14}C) yr BP. The lithic assemblage from this site is dominated by taconite from the Gunflint Formation, though Hudson Bay lowland chert, Lake Superior agate and Hixton silicified sandstone are all noted within the debitage (Hinshelwood 1990). The assemblage consists of 86 bifaces, two Plainview blades, a small number of possible projectile points, a groundstone axe, a taconite spear point of unknown type, a millstone and numerous flakes and other debitage (Hinshelwood 1990).

These clusters are based on the present archaeological research for this area, and may reflect the biases previously discussed. Hinshelwood attempts to categorize these sites based on their location near various watersheds however, it can also be suggested, for example, that many of the sites could be grouped as Minong shoreline sites. It may be too quick to group sites into categories at this point in time due to poor dating control of sites and a lack of information regarding inland archaeological sites which are often underrepresented.

4.4.4. Other Plano Site Clusters in the Thunder Bay Region

The Dog Lake area, northwest of Thunder Bay, has also been examined for archaeological sites (Figure 4.9). Numerous sites were located, excavated and mapped in the 1960s and 1970s by K.C.A. Dawson, Bill Fox, Peter Englebert and Mike McLeod (McLeod 1980). This area is important because it is one of the few locations where archaeological survey has identified inland occupation sites. However, Dog Lake is now a hydro dam headpond (McLeod 1980). The damming of the lake raised the water level, which washed out the old shoreline however, many sites have been exposed when water levels have been low (McLeod 1980). This area is important because many multi-component sites have been identified, ranging from Paleoindian, Archaic, Woodland, and Historic times (McLeod 1980). These include Hakka (DeJi-1), Brown (DeJi-9), Hawkeye Lake (DeJi-11), Dynes (DeJi-13), Summer Snow (DeJi-3), Sandstrom (DeJj-21) and Kathy (DeJj-22) (McLeod 1980). All of these sites incorporate Paleoindian, Archaic, Woodland, and Historic components, suggesting that they were re-occupied throughout much of the Holocene. The Brown site produced multiple projectile points, including both lanceolate and side-notched points made from Hudson Bay Lowland chert, and a stemmed point possibly made from Swan River chert, a lithic material quarried from Manitoba (McLeod 1980). The presence of both Plano and Archaic traditions may represent a transition period however, this is notcertain (McLeod 1980). McLeod (1980) notes that there appears to be a gap in Archaic projectile point types and other artifacts around Dog Lake extending from the Plano/early Archaic occupation, until the later Archaic period, however he does not suggest a reason for why this is.

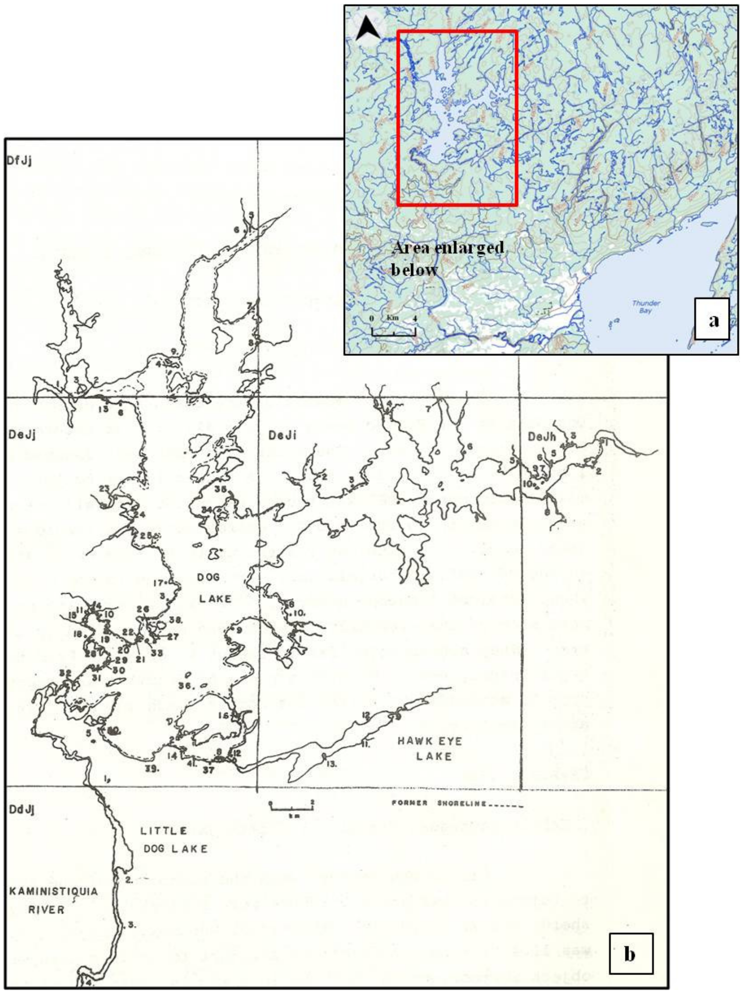


Figure 4.9. Map "a" shows the location of Dog Lake northwest of Thunder Bay (from NTS 52/A6, scale = 1:300,000. Figure "b" shows the location of archaeological sites and Borden System coordinates around Dog and Hawk Eye lakes (from McLeod 1980).

The Kaministiquia Delta located under and around the city of Thunder Bay has been found to be rich in Paleoindian and Archaic artifacts (Figure 4.10) (Hamilton 1996). In a survey of agricultural fields between 213 and 259 m asl in elevation, Hamilton (1996) found multiple areas with concentrations of flakes and debitage. The finds occurred over a large area, suggesting widespread use of the delta for a short duration of time (Hamilton 1996). The lithic assemblage consists predominately of taconite however, Knife Lake siltstone was also used (Hamilton 1996). Till-derived siliceous material was available on the delta but was not utilized by Paleoindian people, even though it was a much better material for manufacturing tools than taconite (Hamilton 1996). This supports previous statements that Paleoindian people preferred taconite derived from bedrock sources as opposed to more readily “flakable” material pebbles and cobbles deriving from gravel deposits. However, it has yet to be determined why this is.

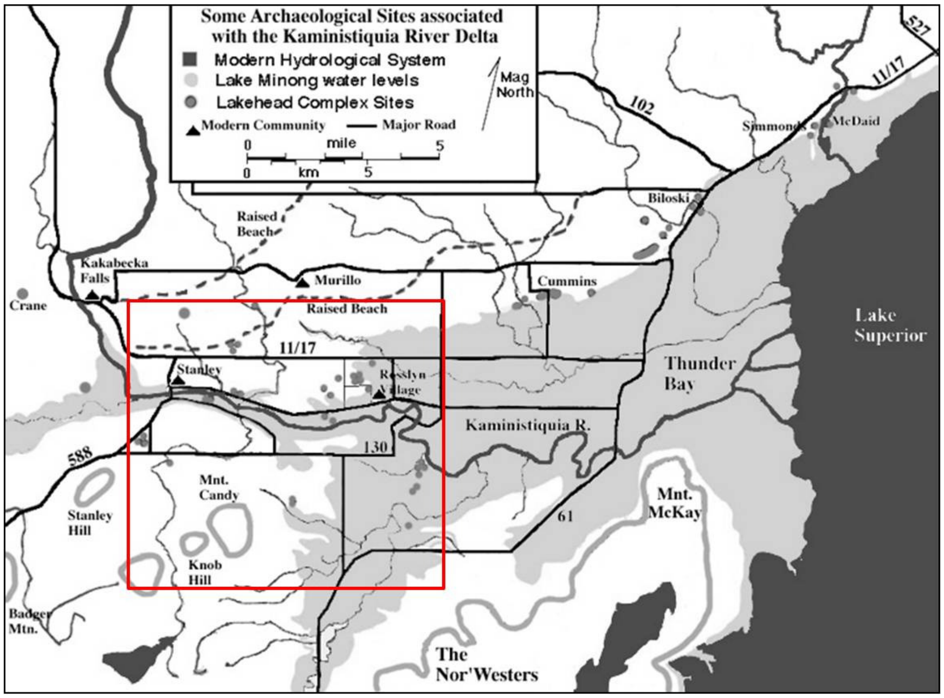
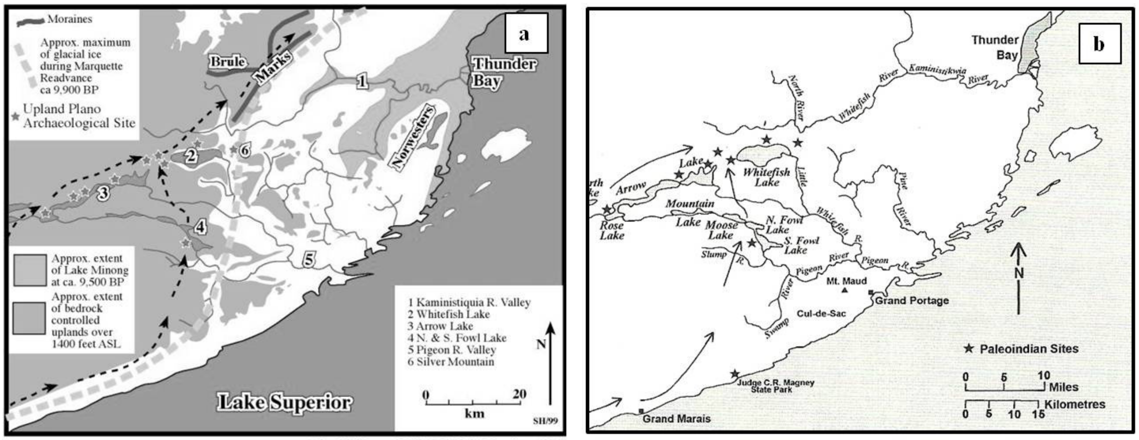


Figure 4.10. Part of the Kaministiquia Delta where archaeological survey revealed the presence of Paleoindian and Archaic sites (from Hamilton 1996).

Finally, the Gunflint-Arrow-Whitefish lakes area located north of the Minnesota-US border and southwest of Thunder Bay has also been recognized for Paleoindian occupation (Figure 4.11a, b) (Phillips and Hill 2004). As previously mentioned, there have been no Early Paleoindian sites found in this area, but late Paleoindian, Archaic, and Woodland sites have all been identified.



(After Phillips and Hill 1995:22-24, Hamilton 1996)

Figure 4.11. The upland area southwest of Thunder Bay. Figure "a" depicts major lakes in the area (from Hamilton 2010). Figure "b" shows the locations of Paleoindian sites found (Phillips and Hill 2004). Arrows on both figures indicate possible route of Paleoindian migration.

4.5. THE ARCHAIC TRADITION IN NORTHWESTERN ONTARIO 7,800-2,600 cal (7,000-2,500 ^{14}C) yr BP

The Archaic tradition of Northwestern Ontario is believed to have evolved from the existing Paleoindian culture (Dawson 1983a, 1983b). The term and definition of "Archaic" was first applied to the Eastern Woodlands of the United States and Southern Ontario to reflect changes observed in the archaeological record that first appear between 12,500-11,500 cal (10,500-10,000 ^{14}C) yr BP (Ellis et al. 1998). These changes include a shift to a more generalized subsistence economy with various animal and plant resources consumed, an increase in population, and less mobility as settlements appear to be repeatedly occupied (Ellis et al. 1998). In Northwestern Ontario, the term "Archaic" is used, as some changes in the archaeological record are visible, such as side and corner notching on projectile points, however, there is little to no evidence to suggest that populations increased or subsistence economy changed (S. Hamilton, personal communication 2010). These observed changes occurred approximately 3,000 years after the Archaic period first appears in the Eastern Woodlands (Dawson 1983a). Therefore, in Northwestern Ontario, the term "Archaic" does not imply the same cultural and technological transformations, nor timing that it is associated with anywhere else in Canada and the United States. However, the Eastern Woodland Archaic cultures do contribute to the appearance of the Archaic tradition in Northwestern Ontario, but the technological trends associated with this transition reflect adaptations to the boreal forest environment (Wright 1972).

Dawson (1983a) places the Archaic period in Northwestern Ontario between 7,800-2,600 cal (7,000-2,500 ^{14}C) yr BP. This is based on tool modifications from earlier Plano toolkits, the introduction of notching on projectile points, preferred use of stone nodules as opposed to bedrock sources, and the introduction of native copper into the toolkit (Dawson 1983a). Dawson (1983a) identifies two geographically and technologically defined Archaic complexes within this area; the Northern Shield Archaic and the Southern Shield Archaic (also identified as the Old Copper Archaic) (Figure 4.12). The major difference between these two groups is the latter's use of native copper, however, there appear to be few differences in other aspects of their toolkit (Dawson 1983a; Hamilton 2007) and due to this, this thesis will treat the Northern and Southern Shield Archaic as a single complex, as proposed by Wright (1972).

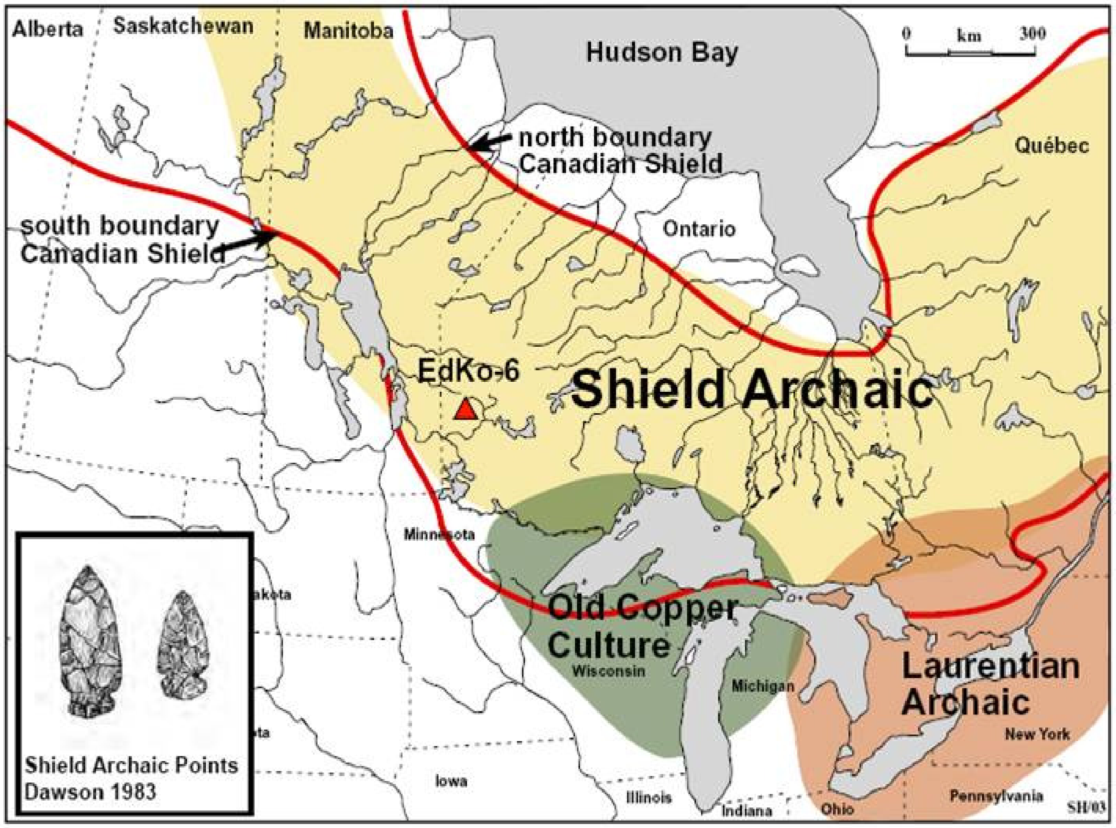


Figure 4.12. Distribution of Archaic traditions around the Great Lakes. The Old Copper culture likely reflects either Shield Archaic or Laurentian Archaic use of native copper, and does not necessarily indicate a completely separate culture (Hamilton 2007).

The transition from the Paleoindian to Shield Archaic traditions has been identified as an *in situ* cultural change as opposed to a migration-and-replacement scenario (Dawson 1983a). Dawson (1983a) suggests that it was climate change and a closing boreal forest that caused this cultural and technological change. Pollen records from Cummins and Oliver Ponds indicate that a major change in climate to warmer, drier and windier conditions occurred by 8,900 cal (8,000 ^{14}C) yr BP (Julig et al. 1990), approximately 1,000 years earlier than the date proposed by Dawson (1983a) for the emergence of the Shield Archaic tradition. This could either mean that the transition between Paleoindian and Shield Archaic could have occurred 8,900 cal yr BP if climate was the driving force behind the cultural change, or that there are other unidentified driving forces behind the changes observed within the archaeological assemblage. Since there is evidence that climatic change may have been responsible for distinctive changes between the fluted early Paleoindian Clovis and Folsom points to the unfluted late Paleoindian points, it may be logical that the warm and dry Hypsithermal climate resulted in the changes from the Plano to Shield Archaic cultures.

Wright (1972) divides the Shield Archaic chronologically into Early (7,900 -6,800 cal [7,000-6,000 ^{14}C] yr BP) and Middle (6,800-3,200 cal [6,000-3,000 ^{14}C] yr BP) periods, with most cultural and technological changes occurring in the Middle period, however, this division is not universally accepted and there is inadequate archaeological evidence to support it. This is partly because no typologies have been assigned to Archaic artifacts in this area and there is a lack of dated sites, particularly in the early Shield Archaic period (Hinshelwood 2004).

The under-representation of sites following the early Shield Archaic period may be attributed to limited archaeological identification of Archaic sites and fluctuating water levels within the Lake Superior basin throughout the Early and Middle Holocene, meaning some early sites may be submerged below water or buried under sediment (Hinshelwood 2004; see also Chapter 7 [Discussion and Interpretations]), or located inland from the present shoreline (Wright 1995). Wright (1995) argues that the scarcity of Shield Archaic sites along the northern shore of Lake Superior could be rectified by better dating paleoshorelines and extensively surveying raised beaches. This archaeological gap has also been identified around Dog Lake, which will be discussed later in this chapter.

4.5.1. Archaic Technology

Technologically there was a marked change between the Paleoindian and Shield Archaic toolkits. The major differences between the Plano and Shield Archaic toolkits is a decrease in bifaces, an increase in projectile points and scrapers, a decrease in lanceolate points, an increase in side-notched and triangular projectile points (Figure 4.13) and an increased use of nodular flint such as Hudson Bay Lowland chert, and an increase use in ground stone tools (Dawson 1983a, 1983b). The changes in projectile points may be explained by the introduction of new technology such as the bow and arrow (Dawson 1983a) however, since these were likely made from wood or bone, they probably would not have been preserved.

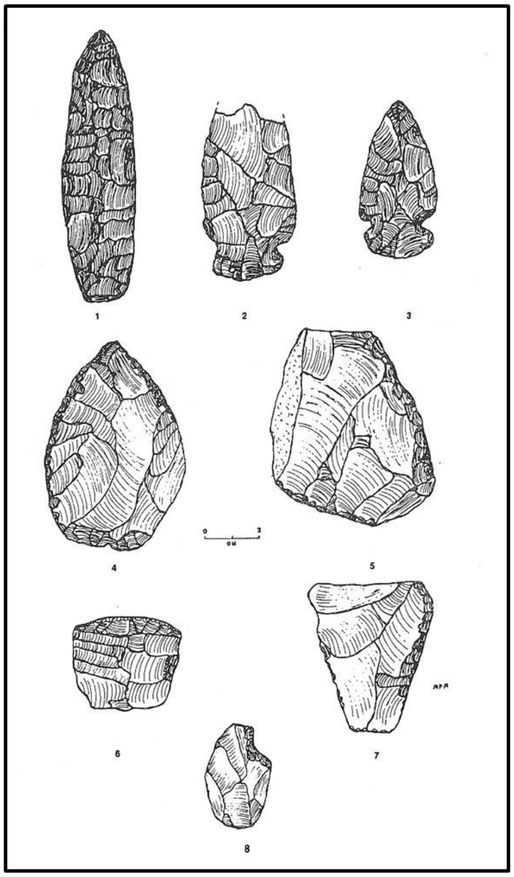


Figure 4.13. Shield Archaic artifacts recovered from Northwestern Ontario. “1” is a projectile point from Dog Lake DeJj-32, “2” is a projectile point from the Kam River DeJh-2, “3” is a projectile point from Dog Lake DeJi-1, “4” is a biface from Dog Lake DeJh-5, “5” is a uniface/knife/scraper from Dog Lake DeJj-22, “6” is an end scraper from Lake Superior (west) DaJi-3, “7” is a side scrape from Dog Lake DeJj-8 and “8” is a spokeshave from Dog Lake DeJh-9 (Dawson 1983a).

Dating when these changes occurred however, is problematic since organic material is rare in Shield Archaic sites. In particular, it is difficult to differentiate early Archaic from late Paleoindian archaeological components because projectile point size is similar and projectile points remain lanceolate or stemmed (Dawson 1983a). However, Shield Archaic projectile points are not as finely made as Plano points, with point bases that are more convex as opposed to the more concave bases observed on Paleoindian points (Hamilton 2007). Bifacially and unifacially flaked tools remain common within the toolkit however, they are generally culturally non-diagnostic (Hamilton 2007). It is not until later in the Shield Archaic period where projectile points become smaller and notched.

Sometime during the Shield Archaic period, woodworking tools such as adzes, axes, gouges, celts and chisels, made out of stone or native copper were introduced (Figure 4.14) (Dawson, 1983a). Other tools made of copper include socketed spear points, projectile points, knives, harpoons and pikes (Dawson 1983b). The manufacture of native copper implements is thought to have started in the Archaic period with material from around the Lake Superior basin, including Isle Royale, an island located in the northwestern part of the basin (Dawson 1983a, 1983b; Hinshelwood 2004) and the Keewinaw Peninsula located in Wisconsin (Martin 1999). Wright (1995) suggests that the peak in copper use was around 6,800 cal (6,000 ^{14}C) yr BP, although he acknowledges its initial use around 7,800 cal (7,000 ^{14}C) yr BP. A wooden shaft attached to a copper spearhead recovered near South Fowl Lake provided a radiocarbon date of 7,600 cal (6,700 ^{14}C) yr BP (Beukens et al. 1992). This date represents the earliest evidence of copper tool manufacturing around the Lake Superior basin (Beukens et al. 1992).

Figure 4.14. Archaic artifacts found in the Thunder Bay area. a= copper celt, b= plaster casts of copper spear points, c= chalcedony biface fragment thought to be of ceremonial or ritual purpose as its size would make it unsuitable to use as a tool (photos taken by author).

An important copper find is the Nipigon Cache found south of Lake Nipigon (Figure 4.15) (Wright 1966). Many copper artifacts were found in this cache, including open socketed projectile points and knives, bracelets, pendants, beads and an axe, as well as many pieces of nodular flint (Wright 1966). Knife River Flint stone tools were also found in the cache suggesting either trade from North Dakota or mobility of the Archaic people (Wright 1972). This cache holds decorative and/or ceremonial items which are thought to be grave offerings. No bones have been recovered however, there was evidence of red ochre on some of the artifacts which is indicative of a grave since it has been associated with Shield Archaic graves elsewhere (Wright 1972). Unfortunately there are no dates associated with the cache.

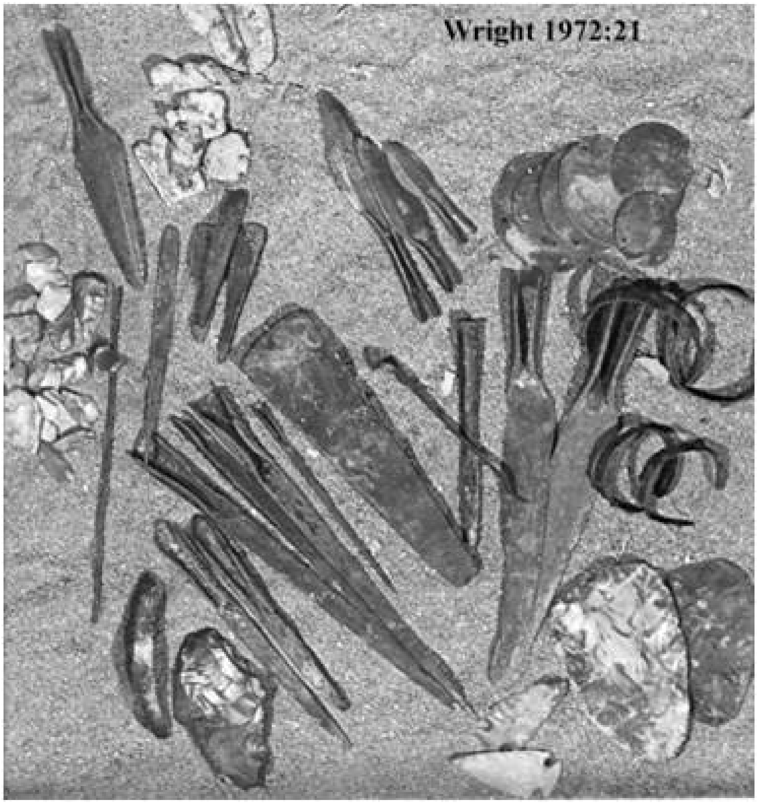


Figure 4.15. Photo of the copper, chalcedony and taconite artifacts found within the Nipigon cache (from Wright 1972).

4.5.2. Archaic Subsistence

There is no direct archaeological evidence in the way of animal bones to tell what foods Archaic peoples consumed. Although mammal bone such as beaver (*Castor canadensis*) have been identified at some sites, including the Hawk Bay site (DeJj-2) at Dog Lake (McLeod 1980), most of the remains are either too fragmented, burned, or disturbed for proper identification, radiocarbon dating, and determination of cultural affiliation. The Archaic diet was probably similar to that of the Plano, as they are believed to have hunted large animals such as moose and bear, as well as medium and smaller game including deer, hare, waterfowl and beaver, plants and berries were also likely consumed (Dawson 1983a). Many fishing implements including hooks and gaffs are part of the Archaic toolkit and based on these artifacts, plus the proximity of archaeological sites to lakes and rivers, fish were also likely a staple food in the Archaic diet (Dawson 1983a).

A new approach in Subarctic archaeology is to study paleo-diets using stable isotope analysis (Hamilton and Finch 2010). Basin on stable carbon and nitrogen isotopes found in bone collagen it is possible to determine what kinds of plants and animals people were consuming. Hamilton and Finch (2010) submitted bone samples from multiple individuals from throughout northern Ontario, including five samples from Wapekeka (FlJj-1), one from Bug River (FkJm-1) (Figure 4.16), and the remains from four undated but likely Late-Holocene-age individuals from northern Ontario. It was anticipated that ^{13}C values would have been high due to the dominance of C3 plants in the Subarctic however, the actual values from the bone collegen were lower than anticipated (Hamilton and Finch 2010). Hamilton and Finch (2010) attribute this to the possibility of a higher intake of aquatic resources, suggesting fish were an important staple food. Even though radiocarbon dates obtained from the individuals range from 7,900-4,500 cal (7,000-4,000 ^{14}C) yr BP, the percentages of ^{13}C and ^{15}N were fairly consistent (^{13}C= -18.24 to -20.3, ^{15}N= 10.46 to 12.83) (Hamilton and Finch 2010). Hamilton and Finch (2010) suggest that this could indicate that the diet of Archaic people, at least between 7,800-4,500 cal (7,000-4,000 ^{14}C) yr BP may have been the same or similar. They also, however, acknowledge that stable isotope analysis is in its infancy in the Subarctic and not much information is known about the plants of the Canadian Shield (Hamilton and Finch 2010). More data analysis and samples may be needed to reconstruct paleo-diets for this area (Hamilton and Finch 2010) however, this proxy is able to shed some light on subsistence strategies of northern populations where faunal and plant remains are scarce.

Figure 4.16. Location of Wapekeka and Kitchenuhmaykoosib Inninuwug First Nation where human remains have been found (from Hamilton and Finch 2010).

Diet is not the only subsistence information obtained from the skeletal remains from Wapekeka. Hypoplastic lines have been identified in tooth enamel on two of three individuals recovered. These lines indicate that the individuals suffered some form of physical stress in their lives. As the three individuals were all likely under the age of 18, based on tooth eruption patterns, this stress occurred at a young age. Due to the presence of multiple hypoplastic lines Hamilton (2004) suggests that the form of stress was likely chronic, perhaps reflecting periodic malnutrition. It is also suggested that the subsistence practices of these northern peoples should be examined in an attempt to identify possible issues which may have resulted in the stresses identified in the skeletal remains. Interestingly, the dates of the remains cluster around ~7,800 cal (7,000 ^{14}C) yr BP, coinciding with the Hypsithermal period, which is known to have caused severe drought on the Great Plains however, it is not known how it impacted the people of northern Ontario.

The Deadman Slough and Sucices sites in northern Wisconsin revealed the remains of white-tail deer, beaver, turtle, and bird in association with Paleoindian and early Archaic artifacts, suggesting a generalized foraging subsistence economy for both (Kuehn 1998). Archaic populations in Northwestern Ontario probably had access to a variety of resources, much like their counterparts in Wisconsin, as the LIS carved out many lakes in the region likely creating productive aquatic settings. This also supports the idea that Archaic peoples followed a generalized foraging subsistence economy. At the Itasca site, a bison kill site along the prairie-woodland border in northern Minnesota, Archaic people consumed bison during the Middle Holocene when the prairie expanded (Shay 1971). It is unknown if bison were present in northern Ontario at this time, however, blood residue from tools at the Cummins site indicate the presence of bovid (likely either musk oxen or bison) (Newman and Julig 1990) dating before 9,000 cal yr BP, as the artifacts were Paleoindian. Bison remains were also recovered near Kenora, Ontario and date to 5,575 ± 71 cal (4,850 ± 60 ^{14}C) yr BP, within the Late Archaic period, but not associated with any archaeological material (McAndrews 1982). The bison in Kenora however, could represent a rogue bison, located well out of its normal range of the adjacent prairies.

4.5.3. Archaic Site Distribution in Northwestern Ontario

Archaic artifacts and sites have been found scattered throughout northern Ontario, many associated with the Nipissing shoreline (Figure 4.17) (Phillips 1993). The Nipissing shoreline is located around 210 m asl in the Thunder Bay area and has been correlated with the Nipissing shoreline from the Sault Ste. Marie area that has been dated to around 6,300 cal (5,500 ^{14}C) yr BP, within the Late Archaic period (Phillips 1993). Archaic peoples also re-occupied Paleoindian sites, as Archaic artifacts have also been found at the Cummins and Brohm sites (McAndrews et al. 2004), as well as many sites around Dog Lake and on the Kaministiquia River Delta (Dawson 1972; Hamilton 1995).

In the northern part of this region, two Archaic sites have been identified and represent early occupation following deglaciation. The first site is the Allen Site (EcJs-1) in the Sioux Lookout district (Pilon and Dalla Bona 2004). Evidence suggests that people occupied the site approximately 400 years after the LIS retreated from the area around 8,853 ± 127 cal (8,000 ± 80 ^{14}C) yr BP, a date based on charcoal from a paleosol at the site (Pilon and Dalla Bona 2004). This paleosol however, is not directly associated with the artifacts recovered from the site. Artifacts found include unifaces, bifaces, trihedral adzes, lanceolate points and a possible tip from a native copper awl (Pilon and Dalla Bona 2004). Some of these artifacts such as the trihedral adzes, are similar to ones associated with the Caribou Lakes Complex, a late Paleoindian complex from southeastern Manitoba, but the presence of copper indicates Archaic technology (Pilon and Dalla Bona 2004).

The second site is the Wapekeka site (FlJj-1). What is unique about this site is that human remains were found and radiocarbon dated to between 6,630 ± 90 and 7,080 ± 90 ^{14}C BP (Hamilton 2004). The only artifacts found were a bifacially worked stone gouge and a possible chitho however, due to disturbance, it is unknown if these artifacts were associated with the human remains and therefore cannot be dated to the same age (Hamilton 2004). The gouge may have been a woodworking tool, a class of implement that is commonly associated with the Early Shield Archaic tradition, and the chitho, which makes its first appearance in the Archaic toolkit, may have been used as a hide scraper (Hamilton 2004).

Archaic sites have also been found west of Thunder Bay by Shabandowan Lake and the Kashabowie River. Dawson (1972) describes 10 projectile points, 24 scrapers, copper projectile points and a socketed copper projectile point that were collected by a local land owner from the mouth of the Kashabowie River, a locality that Dawson (1972) termed as the Boyes site (DdJo-1). Since these artifacts were collected by the owner, the exact context they were found in is unknown. This, however, is the case for many reported Archaic sites in Northwestern Ontario.

South of Thunder Bay, near the Minnesota border, a cache of bifaces has been found at Rose Lake (Dawson 1986). This cache consists of 17 bifaces and three projectile points all made of taconite (Dawson 1986). One projectile point is corner-notched, indicating Archaic origin, and two are stemmed and parallel flaked, which is typical of Paleoindian points (Dawson 1986). Although no direct radiocarbon dates were obtained from this cache, Dawson (1986), assigns a date of approximately 6,000 yr BP based on the presence of Paleoindian flintknapping traits on the Archaic points at this site. Obviously this age estimate should be viewed as speculative, because there is limited evidence to suggest when the Archaic timeframe began and no typologies have been created for Archaic artifacts in Northwestern Ontario.

Many Late Archaic sites have been identified in Northwestern Ontario based on the point attributes mentioned above however, only one will be mentioned here. The Renshaw site (DaJi-1) has been identified as a late Archaic copper tool manufacturing site (Arthurs 1989). It is located south of the city of Thunder Bay along the west coast of Lake Superior. This site was occupied during the higher Algoma phase of Lake Superior and was probably an island during this time (Arthurs 1989). This site was rich in lithic and copper artifacts and features including hearths and post moulds, indicating possible dwellings. A sample of charcoal was taken from a pit feature and radiocarbon dated to 3,700 cal (3,400 ^{14}C) yr BP (Arthurs 1989), putting it within the Late Archaic timeframe, however, initial occupation may have been earlier. The Renshaw site, if not disturbed, could have provided much more data on the Archaic archaeological record. Unfortunately, however, the site was heavily disturbed by gravel pit operations and is also suspected of having been looted prior to archaeological survey.

4.6. BURIED AND SUBMERGED ARCHAEOLOGICAL SITES AND ARTIFACTS

As previously mentioned, there appears to be a limited number of sites in the archaeological record following the late Paleoindian/early Archaic timeframe in many places across northern Ontario. The archaeological survey work at Dog Lake by McLeod (1980) found many archaeological sites dating from Paleoindian to Historic times. However, apart from some evidence of early and possibly late Archaic artifacts, there appears to be an underrepresentation of early Archaic sites and artifacts (McLeod 1980). Based on climatic reconstructions as well as historical data regarding the flooding of Dog

Lake as a reservoir, lake levels were originally lower and may have been still lower during the Middle Holocene as the climate was warmer and drier than today. It is possible that the under-representation of Early Archaic material at Dog Lake may be because sites from this timeframe remain submerged under present lake levels.

Closer to the city of Thunder Bay is the Turning Basin site (DcJh-5) (Figure 4.17). This site is located near the mouth of the Kaministiquia River and encompasses a series of accidental finds over a period of years (Dawson 1972). In 1913, sewer excavations approximately 200 m north of the turning basin revealed many copper artifacts approximately 12 m below the surface (Dawson 1972). In 1914 copper chisels and a copper adze were found in association with a flexed burial roughly 1,036 m north of the turning basin and less than 2 m below the surface (Dawson 1972). When sewers were being excavated years later in the same area, more copper artifacts were found, including a fish hook associated with cedar cribbing approximately 9 m below the surface (Dawson 1972). However, it is unknown if these artifacts are in their primary context (Dawson 1972). During a slip-way excavation in 1918 a mammal bone and a socketed copper spear point were found approximately 12 m below the surface (Dawson 1973), on top of clay and overlaid with sandy deposits, suggesting that they were pre-Nipissing in age and can be attributed to early Archaic peoples (Dawson 1972). These finds have been grouped together and labeled the Turning Basin site. The problem with these finds is that most of the artifacts, including the human remains have been lost and the only records that exist of them are from sewer excavation reports and an eyewitness account from one of the site workers (Dawson 1972).

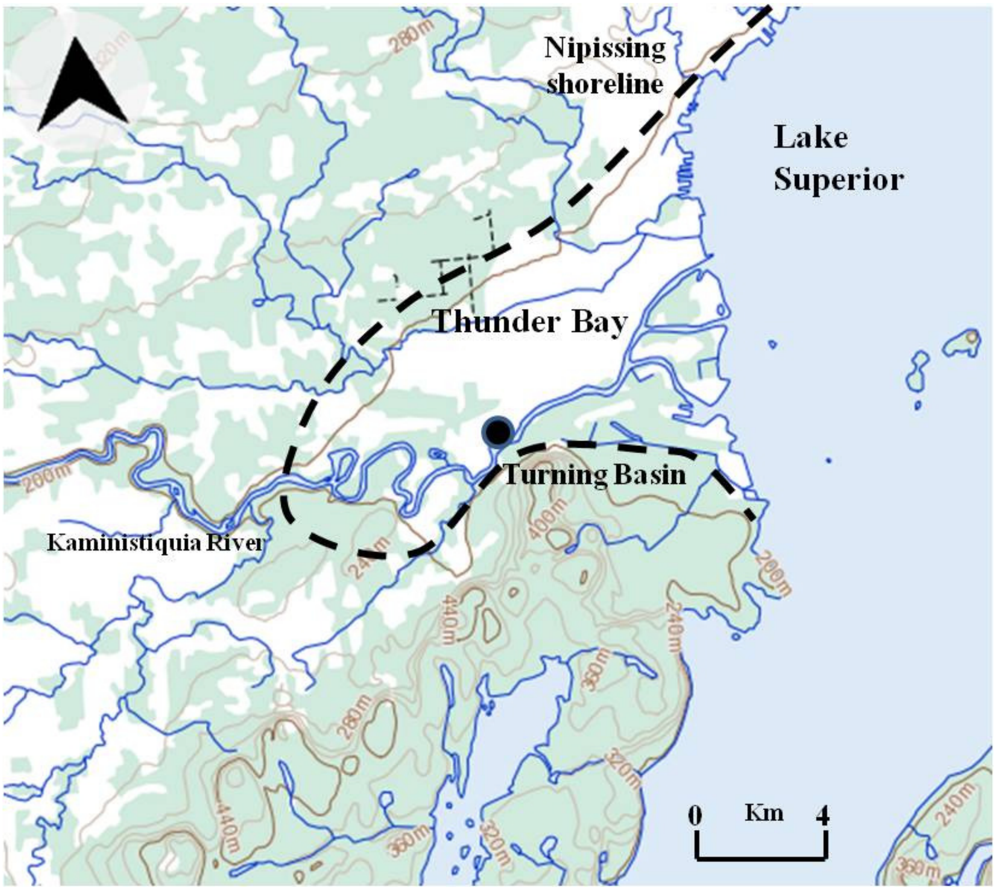


Figure 4.17. Location of the Turning Basin site and Nipissing shoreline (Based on NTS 52/A6, scale 1:150,000)

Many isolated Archaic copper finds have been found along the north shore of the Lake Superior basin, also under lacustrine sediment. At the Armstrong site, located along the northern shore of Lake Superior near the Pic River, a copper gaff was associated with a hearth approximately 30 m above the present lake level and under 7.5 m of what have been described as "stratified clays and silts" thought to have been deposited during the Nipissing transgression (Wright 1963, 1995). If all of these artifacts were located in their primary context, then their burial under lacustrine sediment indicates that these areas were occupied prior to a rise in water level within the Lake Superior basin. This is a key point which will be addressed at length later in this thesis, as it implies a connection between buried Archaic artifacts and the Houghton Low which occurred around 8,900 cal (8,000 ^{14}C) yr BP.

4.7. SUMMARY

The archaeological record for the Early and Middle Holocene is incomplete due to a variety of reasons ranging from poor preservation conditions to the burial of sites by lake sediment. The first people to occupy this area were Paleoindians of the Plano tradition, who likely migrated into the area shortly after deglaciation. Archaeological evidence supports the theory that many different Plano peoples occupied the Thunder Bay area on a seasonal basis or perhaps shared ideas and technology within a wide area, encompassing Northwestern Ontario and the adjacent states. Although multiple sites have been identified, particularly on Lake Minong shorelines, very little is known about the Plano tradition in Northwestern Ontario, largely due to poor preservation of organic/biological remains and archaeological survey biases.

Following the Plano culture, the Shield Archaic culture emeraged. Exactly when these changes/adaptations occurred is unknown, but there is a marked change in the toolkits between these two traditions. The manufacture of copper tools and ornaments, as well as notching on projectile points, are two of the main difference. Problems with the Archaic period stem from poor site visibility, lack of paleoenvironmental understanding, and few radiocarbon dates. To combat these problems, the following questions must be addressed when dealing with the Archaic timeframe: 1) Does the lack of Archaic sites reflect archaeological survey bias such as looking in the wrong locations? 2) Are Archaic sites buried under lacustrine sediment deriving from subsequent higher water levels and is that why there is limited evidence of them? 3) Does the fact that there are no typologies associated with Archaic artifacts skew the data? These questions may have to be answered before determining whether divisions can be assigned to this period and associated with particular date ranges. Only then may it be possible to understand Early and Middle Holocene peoples.

CHAPTER 5

METHODOLOGY AND THEORY

5.1. INTRODUCTION

This chapter summarizes the study and analysis of macrofossils, both floral and microfaunal, in paleoenvironmental reconstruction. Methods in field sampling, as well as laboratory methods and procedures for macrofossil processing, will also be summarized.

5.2. MACROFOSSILS AS PALEOENVIRONMENTAL INDICATORS

Macrofossils are remains that can be seen by the naked eye. However, low powered magnification may be needed to see morphological features which aid in their identification (Dincauze 2003; Pearsall 2008). For the purpose of this thesis, macrofossils include both floral and microfaunal remains. Plant macrofossils, also known as macrobotanical remains, can be split into two groups: reproductive and vegetative parts (Birks 2001). The former includes plant remains such as seeds and fruits and the latter comprises remains such as leaves, buds, and twigs (Birks 2001). Microfaunal remains include bones, shells, insect remains, and fish scales (Birks 2001).

5.2.1. Macrobotanical Remains

The use of macrofossils in the reconstruction of past vegetation has become increasingly important due to a growing need of more refined data in both the archaeological and paleovegetational records (Beaudoin 2007). In paleoenvironmental research, macrobotanical remains provide one of the best ways of reconstructing local and regional vegetation change. Archaeologically, macrobotanical remains are useful in reconstructing paleo-diet, studying plant domestication, identifying trade networks, and determining site use patterns (Pearsall 2008).

The analysis of plant macrofossils in paleoenvironmental studies began in the early nineteenth century and was the only proxy used in paleovegetation reconstructions until the 1920s when pollen analysis began to dominate (Birks 2001). It was not until the 1960s when the limitations of pollen analysis were realized and it became apparent that macrofossil analysis was an important proxy, especially when added to pollen analysis, for reconstructing paleoenvironments (Birks 2001).

Macrobotanical remains are useful paleovegetation proxies because the information obtained from them can lead to precise vegetation reconstructions (Pearsall 2008). Compared to other paleoenvironmental proxies such as pollen, macrobotanical remains better represent local vegetation (Dincause 2003). This is due to their larger size which inhibits their widespread dispersal, resulting in them remaining in close proximity to their source or parent plant (Dincauze 2003). It is however, possible that plant remains collected from lake sediments may have been transported via fluvial processes, waves, or currents, but it may be possible to identify the resulting mechanical damage from redeposition on the macroremains themselves. In terms of taxonomic identifications, macrobotanical remains can often be identified to the genus or species level (Birks 2001). This produces more accurate vegetation and environmental reconstructions because certain species are indicators of particular environmental or site conditions. Finally, macrobotanical remains can also be dated by AMS radiocarbon analysis (Birks 2001). Dates obtained from macrobotanical remains can provide ranges for deposition and perhaps burial if the plant material has not been re-worked (Birks 2001). More accuate dates can also aid in determining tree-line movements and other vegetation migrations through time (Birks 2001). These changes can be mapped and compared with climate data to reconstruct environmental conditions over time (Birks 2001).

Macrobotanical remains however, may be problematic when it comes to generating quantitative data due to their scarceness in some depositional environments (Dincause 2003). As a result, vegetation reconstructions should be made based on their presence and never on their absence (Dincauze 2003). It must also be acknowledged that macrobotanical remains recovered from a particular location may not represent the total number of taxa that was present because not all vegetation will leave macrobotanical remains behind (Dincause 2003). Due to this, vegetation reconstructions should only be made based on the presence of taxa and not on absence (Dincauze 2003).

5.2.2. Microfaunal Remains

Microfaunal remains provide information on paleoenvironmental conditions, mainly climate and habitat type, although what information can be derived depends on the microfaunal assemblage (Dincauze 2003). Certain animal species are sensitive to environmental factors like temperature and moisture levels, however, this can depend on the size of the animal (Dincauze 2003).

Small invertebrates tend to be better paleoenvironmental indicators, because they are more sensitive to changes in temperature and moisture, making it easier to track changing environmental conditions (Dincauze 2003). Many of these small invertebrates are aquatic animals. Unfortunately, certain classes and families (e.g., mollusks and insects) have been studied more rigorously than others (Dincauze 2003). At present, little environmental data exists for other invertebrates such as sponges (Porifera), and moss-like animals (Bryozoans).

The macrofossil analysis undertaken for this thesis does not focus on the microfaunal assemblage. The main goal of this thesis is to reconstruct the past vegetation community within the Old Fort William paleochannel. However, basic identification of microfaunal remains including aquatic invertebrates was made to aid in determining paleoenvironmental factors (e.g., moisture settings) within the Lower Kaministiquia River Valley.

5.3. TAPHONOMIC CONSIDERATIONS

Plant and animal remains may be subjected to multiple cultural and natural processes after death. The study and interpretation of these processes is called taphonomy (Dincauze 2003). These processes include transportation by processes like wind, water, and erosion, secondary deposition, burial, bioturbation by phenomena like rodent holes, tree roots and tree throws, and any other process which disturbs the remains from their primary context and ultimately impact if and how the remains get preserved (Dincauze 2003). It is therefore important to understand the depositional context in which macrofossils are found, in order to assess if they have been disturbed, redeposited or if they are *in situ* (Dincauze 2003). Identifying the taphonomic processes which may have affected the remains helps in dating and provides context and validity to the data analyzed (Dincauze 2003).

Macrobotanical remains are not nearly as abundant as pollen grains and are frequently subjected to taphonomic processes that destroy or degrade them (Dincauze 2003). There are, however, certain conditions in both terrestrial and lacustrine settings that are more likely to preserve them (Beaudoin 2007). Anaerobic conditions such as occurs in low-energy water saturated settings like peat bogs preserve organic remains because oxygen is too limited to encourage the breakdown of the organic material (Dincauze 2003). Anaerobic conditions can also occur as a result of rapid sedimentation and burial of organics (Retallack 1981), as is likely the case for macroremains recovered from Unit D at the Old Fort William (OFW) cutbank and examined for this thesis (see Chapter 6 [Results]). The burning of macrofossils, whether anthropogenically or naturally helps preserve them if the fire is not intense enough to burn the material completely (Dincauze 2003). Frozen environments may also preserve organic remains because cold temperatures prohibit bacteria from destroying them (Dincauze 2003). Arid regions also preserve macrobotanical remains because it is too dry for the remains to break down however, the remains are generally fragile and must be handled with great care (Dincauze 2003). Finally, macroremains may be identified in volcanic ash (Dincauze 2003). The ash, however, does not preserve the macrobotanical remains themselves. Instead the organic material leaches out and the ash hardens around the resulting void, leaving a mold or impression of the macroremains (Dincauze 2003).

5.4. MACROFOSSIL RECOVERY

In non-archaeological settings macrofossils are generally obtained through coring or sampling submerged or terrestrial sediments and soils (Birks 2001); such sampling may focus on paleosols (Dincauze 2003; Retallack 1981) or peat deposits (Barber 1993). When coring sequences of lake sediment, Birks (2001), suggests that coring the littoral zone of small and shallow lakes is best because more local vegetation will be represented due to the presence of aquatic, emergent, and terrestrial plants growing within or around the shore. Vegetation and climate history of an area can be reconstructed if macrobotanical remains are present throughout the entire core (Birks 2001).

Paleosols, on the other hand, are important contexts in paleovegetation reconstructions because the macrobotanical remains found in them may reflect what was growing both locally and regionally at the time the ancient soil formed (Dincauze 2003). In paleosols that have been buried quickly, with little to no erosion, both plant and animal remains tend to be well-preserved (Retallack 1981). In addition to preserving macroremains for paleoenvironmental analysis, paleosols also provide information on water table, soil chemistry and paleotopography (Retallack 1981). When examining a paleosol, it is beneficial to try to match it to a modern soil based on its physical and chemical characteristics because a match probably means that similar conditions of topography, drainage, vegetation, and climate occurred in the past, provided both soils formed on similar parent material (Retallack 1981). Depth of the water table can be determined by the nature of the B horizon, particularly its depth below the former surface (Retallack 1981), and repeatedly wet conditions can be determined if root systems are shallow or are positioned laterally instead of angled downward (Retallack 1981). The absence of a B horizon could indicate that the paleosol formed in too little time and therefore is immature (Retallack 1981).

Peat deposits are important sources of information on paleovegetation and other aspects of past environments because they keep an *in situ* and stratified record of past, often local-scale, vegetation in the form of both macrobotanical remains and pollen (Barber 1993). Macroremains found in peats tend to be excellently preserved because anoxic conditions prevent organic breakdown after burial (Barber 1993). From these two proxies it is possible to reconstruct a continuous, high resolution, paleovegetation history because the organic remains tend to be well preserved and may also be dated (Barber 1993).

5.5. MACROFOSSIL SAMPLING AND ANALYSIS

The analysis of macroremains can be divided into three stages: recovery, identification, and interpretation (Pearsall 2008). It is important to choose proper methods and techniques in all of these stages to ensure quality of data (Pearsall 2008). The following is an outline and description of the methods and procedures utilized in the sampling and processing of the sediment samples from the Old Fort William (OFW) cutbank.

5.5.1. Sampling Method

Sampling for this thesis was conducted at the OFW cutbank over a period of four days in the summer of 2008. Upon first arrival to the cutbank, the organic layer was identified, exposed by shoveling, and found to cover a distance of approximately 75 m horizontally. A primary datum location was chosen, labeled Datum 0W, and a stake was driven into the cutbank for a more secure marker. This datum also indicates where the organic layer is no longer visible because it decends below the modern water level of the Kaministiquia River. Sampling was conducted horizontally across the organic layer at roughly 5 m intervals in order to study small changes in the spatial distribution of vegetation within the sample area. East from Datum 0W, the organic layer is visible for approximately 30 m before it descends below the modern river level (Figure 5.1). A total of seven samples were taken from this location (0W-30E). The organic layer in this section ranges from 1 to 5 cm in thickness and is easily visible between the sand unit below and the silty/clay unit above. West of the datum, the layer is below the modern water level for roughly 10 m. At 15 m west of the datum the layer is above the water level and is continuous for 30 m more, reaching to 45 m west of the datum. This distance is represented by seven sample locations (15W-45W) (Figure 5.2). The organic layer west of the datum is thinner or less organically enriched than east of the datum. Reasons for this will be outlined in Chapter 7 (Interpretations and Discussion).

After the samples were taken from the cutbank, they were stored and labeled in Whirl-Pak bags and refrigerated until processed. A total of 14 samples were taken from the cutbank, however only 13 samples were analyzed for macrofossils, a decision which will be discussed later. Due to the thinness of the organic deposit sampled (1 to 5 cm), it was not practical to sample stratigraphically (i.e., through time).

A description of the OFW cutbank sediments was conducted in the field by Dr. Matthew Boyd and is discussed in more detail is Boyd et al (2010) and in Chapter 6 (Results) of this theis. The interpretation of these sediments provided essential data needed for the paleoenvironmental reconstruction discussed in Chapter 7 (Interpretations and Discussion).

5.5.2. Macrofossil Processing

Macrofossil processing followed Beaudoin (2007), with some modifications. To quantify macrofossil data it is best to use a known volume during the initial stages of processing (Birks 2001). Beaudoin (2007) determined that a sample volume of 100 ml, measured by water displacement, was enough to allow for a representative data set in

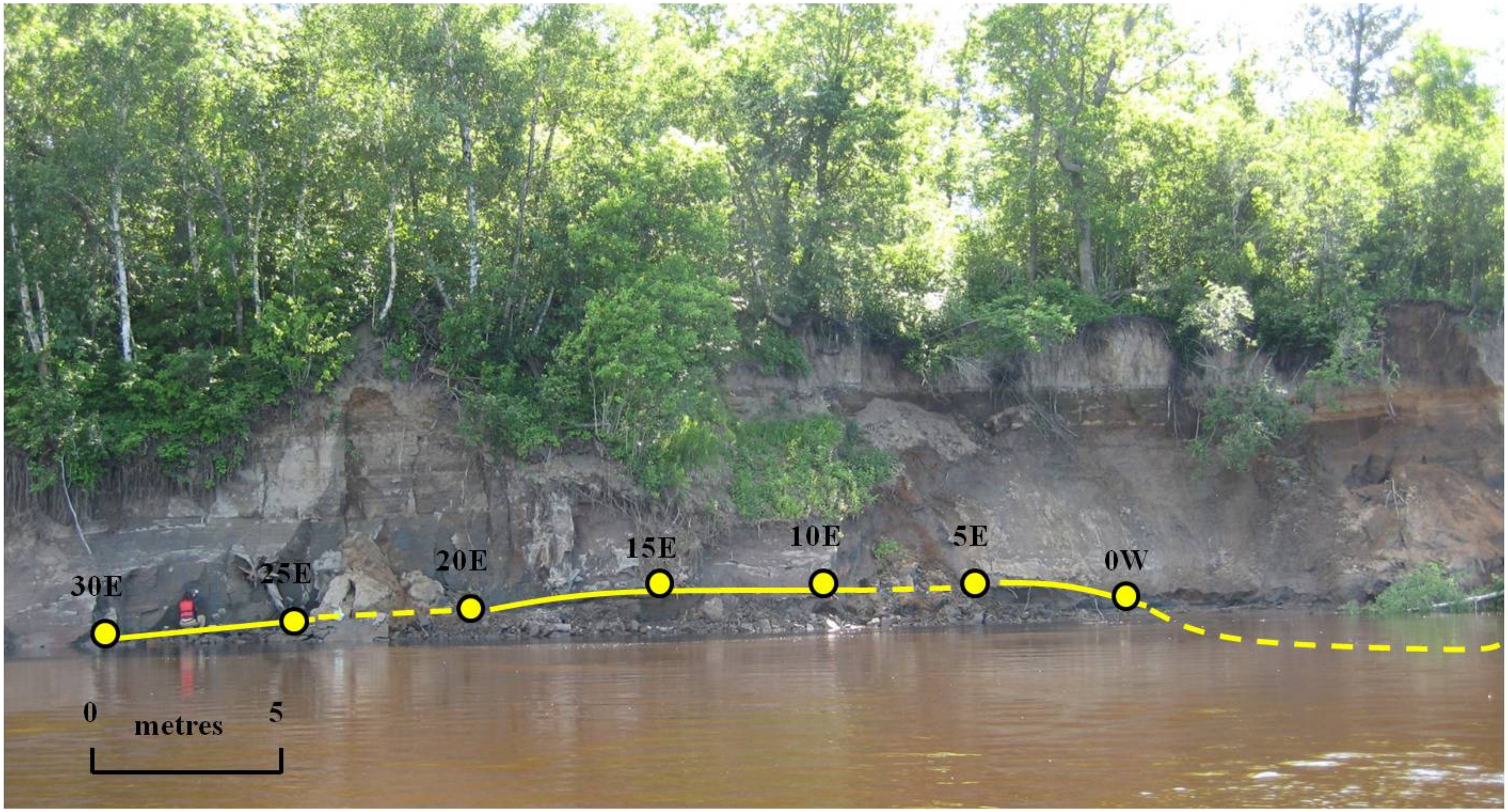


Figure 5.1. Distribution of sample locations (yellow dots) located east of the 0Wdatum within the OFW cutbank. The organic layer (unit D) (yellow line) was sampled at roughly 5 metres intervals. The dashed yellow line between 25E and 20E indicates slumping that has covered the organic layer and the dashed line after 0W indicates where the organic layer dips below the modern water level of the Kaministiquia River.

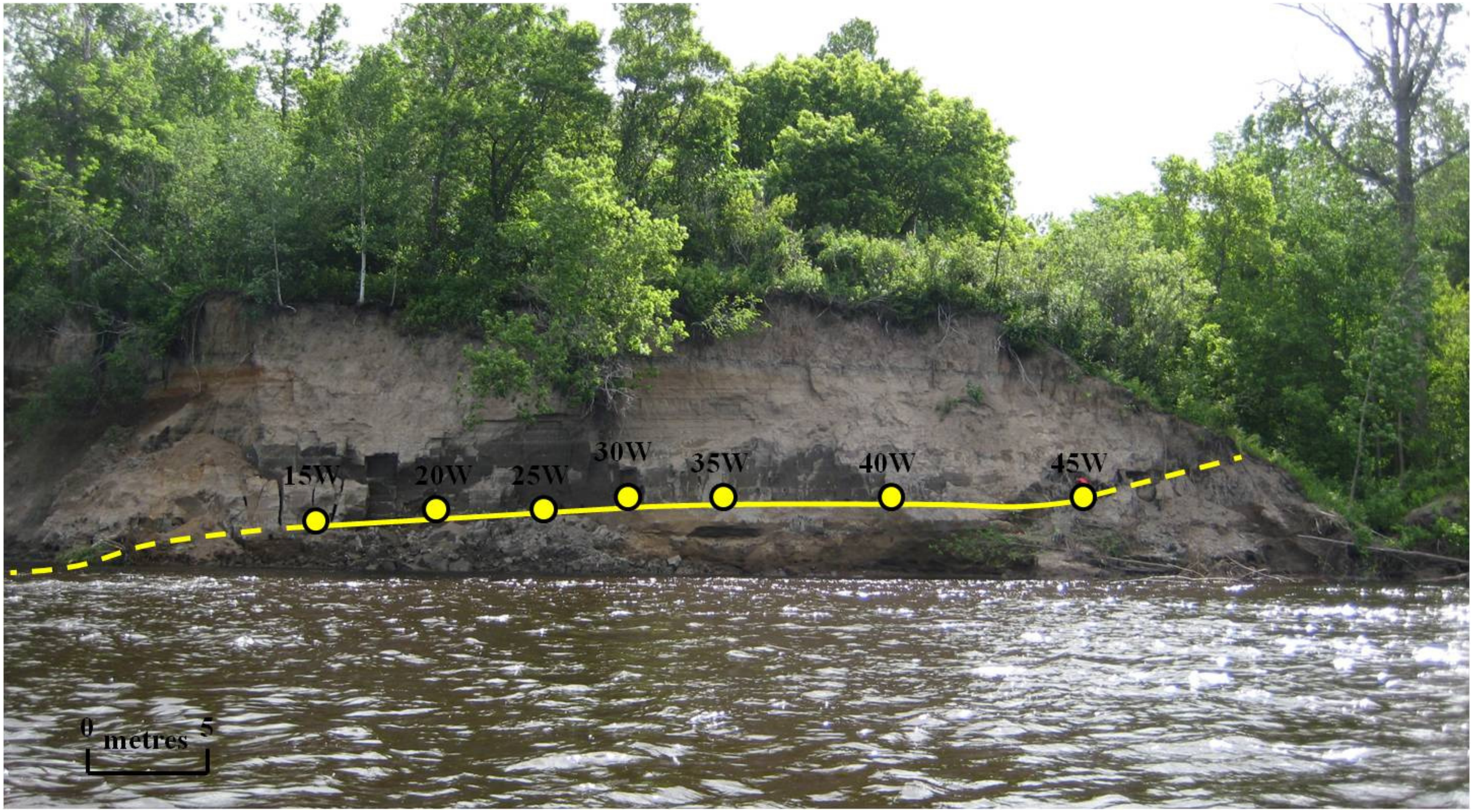


Figure 5.2. Distribution of sample locations (yellow dots) located west of the 0Wdatum within the OFW cutbank. The organic layer (unit D) (yellow line) was sampled at roughly 5 metres intervals. The dashed yellow line east of 15W indicates where the organic layer is not visible and eventually dips below the modern water level of the Kaministiquia River. The dashed line west of 45W indicates where slumping has covered the organic layer.

most depositional environments. Due to high clay content, samples were deflocculated in 5% sodium hexametaphosphate for approximately 24 hours.

Following deflocculation, the samples were wet screened. Mesh sizes were chosen based on Beaudoin (2007), bearing in mind the potential range of sizes of macrofossils screened such as larger seeds, conifer cones, small insect remains and small seeds, however, there were a few alterations. Availability of certain mesh sizes prompted the use of a 1 mm mesh instead of 1.18 mm, and 75 micron µm mesh represented the smallest size fraction instead of the suggested 90 µm. Therefore, screen mesh sizes originally utilized were 2 mm, 1 mm, 250 µm, and 75 µm. After the first sample was screened, it became apparent that the 75 µm mesh size was too small since the finer sediment particles did not pass through it. Instead a 125 µm mesh size was used which fixed this problem and still adhered to accepted screening methods. Samples from 40W and 45W were screened twice because too much sediment was adhering to the organics after the initial screening. After the samples were washed, they were laid out to dry for at least 24 hours, then stored in vials and labeled.

5.5.3. Macrofossil Identification

The macroremains were examined using a dissecting microscope at between 10 and 40 times magnification. The size fractions were each sorted and organics picked out for identification. They were sorted according to macrofossil type (i.e., needles, seeds, twigs, cones etc.) as stated by Beaudoin (2007), and identified with the use of manuals including the *Seed Identification Manual* (Martin and Barkley 2000), *Guide to the identification of plant macrofossils in Canadian peatlands* (Lévesque et al. 1988), *Seeds and fruits of eastern Canada and Northeastern United States* (Montgomery 1977), *Trees* *in Canada* (Farrar 2007) and *Ecology and Classification of North American Freshwater Invertebrates* (Thorp and Covich 1991). Plant specimens from the Lakehead University Herbarium, as well as comparative samples collected from the Thunder Bay area and stored in the Anthropology Department, were essential in the identification of many of the macrofossils. Online resources including the United States Department of Agriculture Plant Database and the Colorado State University seed image database were also utilized for seed pictures and descriptions.

To facilitate the counting process and to produce more useful quantitative data, only certain parts of macrofossils or size categories of some macrofossil types were counted. For example, only conifer needle tips were counted instead of total needle fragments. This was to obtain a minimum number of needles, which otherwise could be inflated due to fragmentation. Only twigs larger than 1 cm long were counted for this same reason. Moss fragments were only counted if they had both the stem and leaves intact. Counting individual leaves would have been too time-consuming and, because of the nature of this study, it would not have been of any benefit to count them all separately. Finally the data was organized and presented using "C2" biostratigraphy software (http://www.staff.ncl.ac.uk/staff/stephen.juggins/software/code/C2Description.pdf).

5.6. SUMMARY

The utility of macrobotanical remains as a proxy in paleovegetation studies cannot be understated. Macrobotanical remains represent vegetation at the local scale which can result in more accurate paleovegetation reconstructions. The recovery of macrobotanical remains is generally dependant on various taphonomic processes that act upon them once they are deposited. However, for accurate vegetation reconstructions, the identification of these processes should be accounted for to ensure validity of data and conclusions.

Sampling and processing methods chosen for macrofossil studies are dependent on the research questions being asked. However, particular steps should be followed throughout the sampling, processing, and identification process. If properly followed, an accurate paleovegetation and/or paleoenvironmental reconstruction is possible.

CHAPTER 6

RESULTS

6.1. INTRODUCTION

This chapter summarizes the macrofossil results from the samples collected from the organic layer (Unit D) located within the Old Fort William (OFW) paleochannel (Figures 6.1 and 6.2). The stratigraphy of the OFW (Boyd et al. 2010) and Boyd (Loope 2006) cutbanks is discussed to provide a sedimentological context to the macrofossil data, and again to provide paleoenvironmental data discussed in Chapter 7 (Interpretations and Discussion).

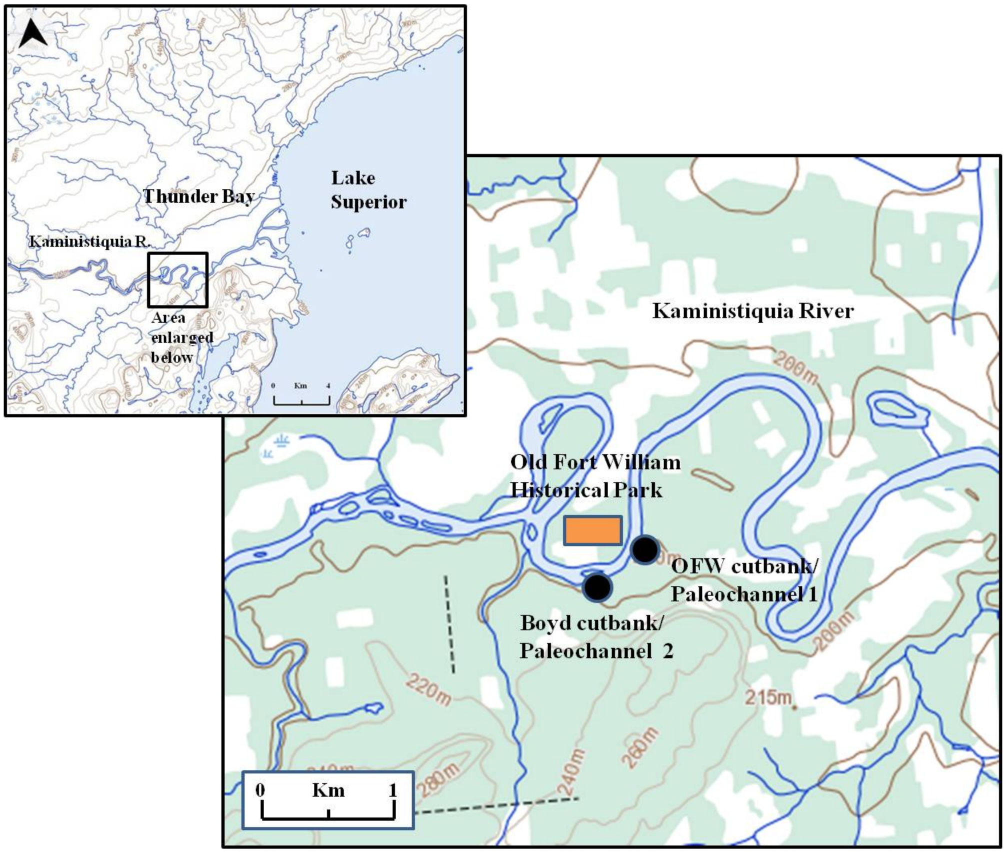


Figure 6.1. Location of the Old Fort William (OFW) and Boyd cutbanks along the lower Kaministiquia River, across from the Old Fort William Historical Park (from NTS 52/A6, scale 1:40,000). The inset map shows the location of the cutbanks west of the city of Thunder Bay (from NTS 52/A6, scale 1:150,000).

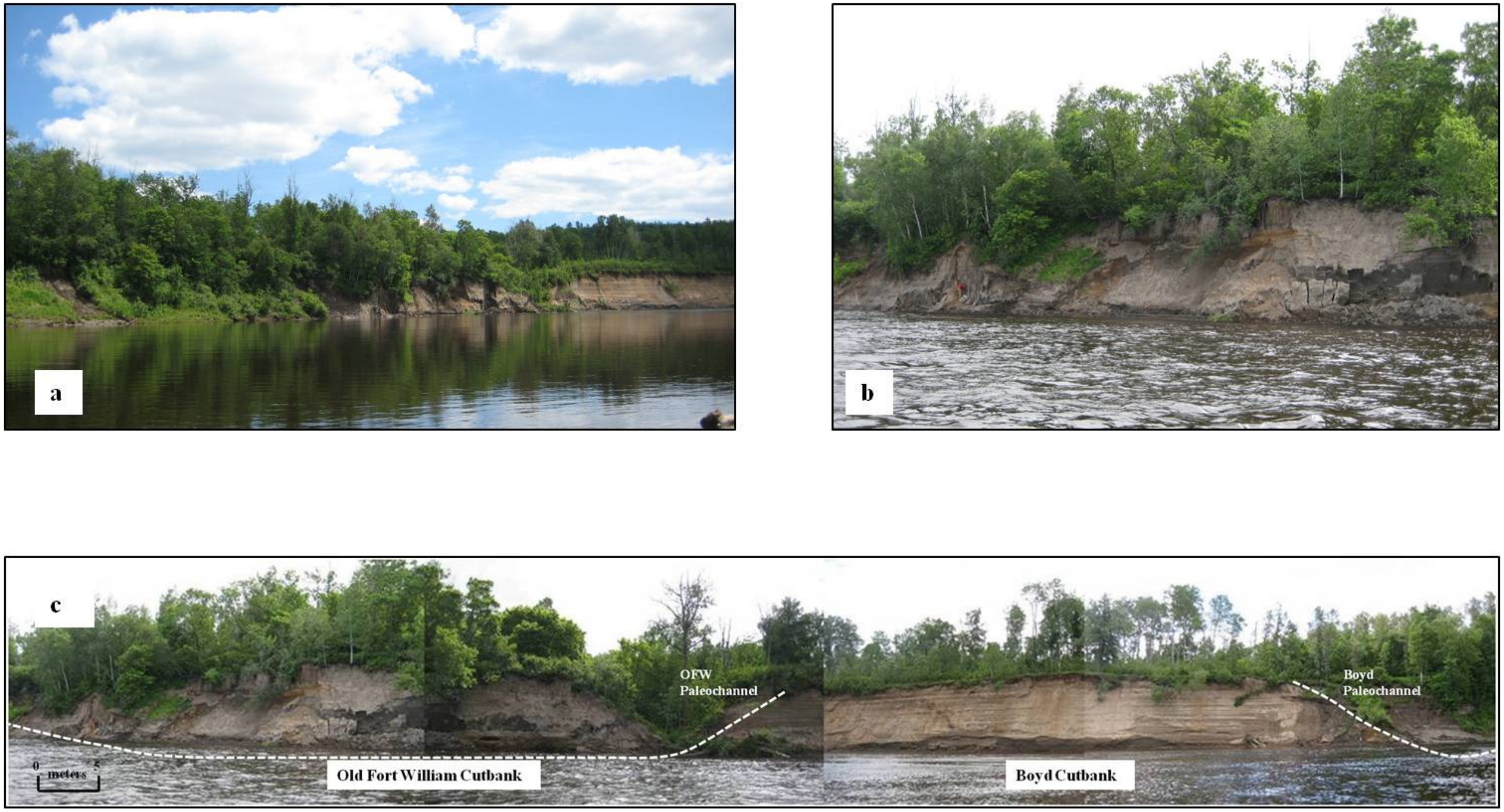


Figure 6.2. Photographs of the cutbanks along the lower Kaministiquia River. Photo "a" is an upstream view of the paleochannel across from the historical park. Photo "b" is a downstream view of the OFW cutbank and photo "c" shows the OFW and Boyd cutbanks with the paleochannel margins highlighted by the dashed line (notes that the paleochannels dip below the modern water level [187 m asl] of the Kaministiquia River).

6.2. CUTBANK STRATIGRAPHY

A series of cutbanks are exposed along a 1 km stretch of the Kaministiquia River (Figures 6.1 and 6.2), with stratigraphic descriptions collected by M. Boyd in 2007 and 2008; these are published in Boyd et al. (2010). The stratigraphic description produced by Boyd et al. (2010) includes the OFW cutbank, as well as the Boyd cutbank and smaller cutbank exposures located further upstream. Within the OFW and Boyd cutbanks, paleochannels likely of the ancestral Kaministiquia River were identified. The sediments from these exposures provide details about past lake level fluctuations and paleoenvironmental settings within the Lower Kaministiquia River Valley throughout much of the Early and Middle Holocene (Boyd et al. 2010). The organics analyzed for this thesis were obtained from the OFW paleochannel. This paleochannel was favoured over the Boyd paleochannel because the whole sedimentary sequence was visible and the organic layer, Unit D, was mostly above the present water level of the Kaministiquia River.

In total, six units (A-F) have been identified and described within the OFW and Boyd cutbanks by Boyd et al. (2010) and Loope (2006). These units are separated based on sediment structure and grain size (Boyd et al. 2010), and will be described from the oldest to youngest deposited sediments (F-A) (Figure 6.3). These deposits will be discussed in terms of geomorphic processes in Chapter 7 (Interpretations and Discussion), and how they relate to geomorphic events discussed in Chapter 2 (Deglaciation and Paleohydrology).

Unit F is best exposed at the Boyd cutbank, located immediately upstream of the OFW exposure (Figure 6.4). This cutbank consists of roughly 12 m of massive to horizontal-or cross-laminated fine sand and silt (Boyd et al. 2010). The laminated units

Figure 6.3. The left side of this figure is a stratigraphic diagram showing sediment structure and division into stratigraphic units (from Boyd et al. 2010). The adjacent photo shows the location of these units within the OFW cutbank. The paleochannel margin is located at the right side of the photo. The yellow represents Unit D that was deposited by 8,900 cal yr BP, based on calibrated radiocarbon dates obtained on plant macrofossils from this unit. All plant macrofossil data reported in this thesis were obtained from Unit D (from Boyd et al. 2010).

are approximately 10-25 cm thick, occasionally ripple-laminated, with climbing ripples common. Approximately 1.2 km upstream from the Boyd Cut, Unit F is underlain by 50+ cm of coarse sand and gravel (Unit G) (Figure 6.5) (Boyd et al. 2010).

Figure 6.4. Unit F (~12 m) at the Boyd cut located between the OFW and Boyd paleochannels.

Figure 6.5. Sedimentary structure of Unit G. Notice the coarse sand and gravel. This unit is stratigraphically below the coarse and fine sands of Unit F.

Units E-A have been described from sediments within the OFW cutbank. Unit E is a < 0.5 m – 2.0 m thick unit consisting of fine to medium sand that is occasionally laminated but mainly massive (Figure 6.3 and 6.6) (Boyd et al. 2010). The laminated sections are mainly cross-bedded however, horizontal laminae are also present (Boyd et al. 2010). Along the margins of the paleochannel, laminae were deposited parallel to the contact between Units E and F, and the presence of fine to coarse sand that dip towards the deepest part of the river channel in some places suggests a point bar deposit. Inside the paleochannels, the upper contact between Unit E and the overlying Unit D, a well-preserved organic layer, is abrupt. Most noticeably at the Boyd cut, Unit E forms an abrupt contact with Unit F as sediments from Unit E were deposited in the active paleochannel as the river incised into the older Unit F sediments (Figure 6.6) Further upstream, Unit E is a coarse sand and gravel unit that is massive to weakly bedded (Boyd et al. 2010).

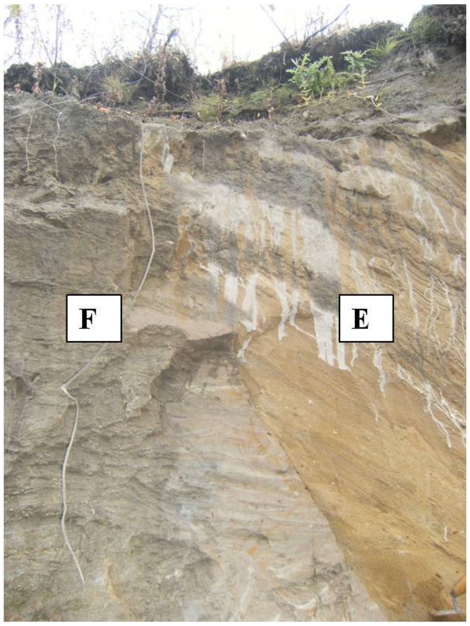


Figure 6.6. Photo showing abrupt contact between Units E and F at the Boyd cutbank, formed when the river channel incised into older sediments of Unit F.

Unit D is a relatively thin unit (0.1 – 9cm), composed of organic remains between the fine silty clay varved Unit C and the sandy Unit E (Figure 6.3 and 6.9) (Boyd et al. 2010). The thickness of this unit varies; it is thicker (5 - 9 cm) near the channel bottom and thins to millimetres along the paleochannel slope. Most organic remains with this unit were concentrated within two laminae overlying Unit E (Boyd et al. 2010). Many tree roots extend horizontally from this unit in places, and the trunk portions of several trees extend vertically into Unit C (Figure 6.7b). For example, one *in situ* tree measuring 3 cm in diameter, was observed to extend roughly 1 m into Unit C (Figure 6.7a) (Boyd et al. 2010). The presence of these *in situ* trees provide evidence that little disturbance took place following the deposition of Unit D however, in one zone of the OFW cutbank the organic deposit appears to have been disturbed on a small scale of <1 m horizontally (Figure 6.7d) (Boyd et al. 2010).

Two radiocarbon dates were obtained from organics located within this unit. A date of 9,065 cal (8,135 ± 25 ^{14}C) yr BP (UCIAMS-26800) was obtained from an 18 cm diameter *in situ* tree located by the 30E sample location (Figure 6.8a) and a date of 8,900 cal (8,010 ± 25 ^{14}C) yr BP (UCIAMS-26801) was obtained from a *Picea* sp. (spruce) cone also located within Unit D (Figure 6.8b) (Boyd et al. 2010). In addition to these dates, Loope (2006) reported two ages obtained on charcoal from Unit D at the Boyd Cut; 8,950 cal (8,070 ± 70 ^{14}C) yr BP (ETH-31437) and 8,900 cal (7,995 ± 65 ^{14}C) yr BP (ETH-31438), as well as a date of 8,900 cal (7,990 ± 20 ^{14}C) yr BP (UCIAMS-61732) from a *Pinus banksiana* (jack pine) cone from the organic deposit further upstream at the Boyd Cut (Boyd et al. 2010).

The organic deposit is stratigraphically in the same unit within both paleochannels, indicating that the accumulation of organics was contemporaneous in both paleochannels (Boyd et al. 2010). The date from the *in situ* tree of 8,135 ± 25 ^{14}C (9,065 cal) yr BP, with the allowance of time for tree growth (36 rings were counted from its truck), suggests that organic accumulation of Unit D began no later than ~9,100 cal yr BP (see Table 6.1. for radiocarbon dates and other relevant data) (Boyd et al. 2010).

Figure 6.7. Photos showing the appearance of the Unit D (organic deposit) throughout the OFW cutbank. Photo "a" is a 3-cm diameter *in situ* tree located between 25E-30E that extends 1 m into the varved sediment of Unit C. Photo "b" shows horizontally extending roots from Unit D by 20E, photo "c" is Unit D by 10E and15E, photo "d" is a disturbed zone of the deposit by 20E and 25E, and photot "e" shows the thin appearance of the deposit at 35E, between Unit E and Unit C. The thinness of the organic layer throughout many of the sample locations west of the datum (15W-45W) looks similar to the layer shown in photo"e".

Figure 6.8. Photos showing examples of macrobotanical remains exposed in Unit D. Photo "a" is an *in situ* tree stump (~18 cm diameter). A similar *in situ* tree in the same deposit was radiocarbon dated to 8,135 ± 25 ^{14}C (9,065 cal) yr BP. Photo "b" is a white spruce cone located in Unit D between Unit E and C dated and to 8,010 ± 25 ^{14}C (8,900 cal).

Unit C is roughly a 2.5 m-thick unit predominately consisting of silty clay laminae (Figure 6.9). The contact with the underlying unit Unit D is characterized by small-scale convolutions measuring no more than millimetres, and rip-ups within the organic laminae of Unit D (Boyd et al. 2010). The silty clay laminae of Unit C are generally indistinct alternating light-dark couplets, or rhythmites, measuring ~1-2 mm in thickness each however, larger couplets around 2 cm are also occasionally present (Figures 6.10a and 6.10b). Each of these couplets consists of a darker, thin (<1 mm) clay lamina, and one or more overlying slightly coarser, lighter coloured silt lamina and then another lamina with silt mixed in a clay matrix (Boyd et al. 2010). Based on their uniformity and repetitiveness, these rhythmites are interpreted as varves (Boyd et al. 2010; Shultis 2009).

The finer clay laminae may represent winter deposition as the sediment particles settled to the lake bottom, with the coarser silt laminae possibly representing sediment deposited by density underflow currents during the summer (Boyd et al. 2010; Murck and Skinner 1999). Based on the varve counts in the parts of Unit C where they are visible, the sedimentation rate was approximately 0.74cm/yr, meaning that Unit C at the OFW site represents at least 200 years of deposition (Boyd et al. 2010). This is supported by the average of 192 varves (range = 182-203) counted from the same unit at the Boyd paleochannel located upstream (Boyd et al. 2010).

Figure 6.9. Sediment differences between the varved Unit C (silty clay), peaty Unit D, which is 1 mm thick near 40W, the sample location in this photo, and the basal sandy sediments of Unit E.

Table 6.1. Radiocarbon ages and other related data associated with early post-glacial sites in the Thunder Bay area. Radiocarbon years were calibrated using the IntCal 04 terrestrial dataset (Reimer et al. 2004) with the 2σ range and means shown (from Boyd et al. 2010).

Site Name	Lab Number	Radiocarbon Date	Calibrated 2σ range and mean (cal BP)	Material Dated	Context	Modern Elevation (m asl)	Latitude (N)	Longitude (W)	Reference
OFW Paleochannel (Unit A)	UCIAMS-61733	6,420 ± 20	7,421-7,291 (7,356)	*Scirpus* sp. seeds	In upper organic deposit	195	48°20.7’	89°21.1’	Boyd *et al*, (2010)
OFW Paleochannel (Unit C)	UCIAMS-26802	7,970 ± 30	8,993-8,659 (8,826)	Wood	Log in silty clay ~2.5 m above base of Unit C	189	48°20.7’	89°21.1’	Boyd *et al*, (2010)
OFW Paleochannel (Unit D)	UCIAMS-26801	8,010 ± 25	9,005-8,776 (8,891)	*Picea* sp. cone	Located at base of silty clay in organic laminae	186	48°20.7’	89°21.1’	Boyd *et al*, (2010)
OFW Paleochannel (Unit D)	UCIAMS-26800	8,135 ± 25	9,127-9,009 (9,070)	*In situ* sapling	Tree in life position extending into silty clay	186	48°20.7’	89°21.1’	Boyd *et al*, (2010)
Boyd Paleochannel (Organic layer)	UCIAMS-61732	7,990 ± 20	8,994-8,774 (8,884)	*Pinus banksiana* cone	Located at base of silty clay in organic laminae	186	48°20.5’	89°21.2’	Boyd *et al*, (2010)
Boyd Cutbank	ETH-31437	8,070 ± 70	9,046-8,658 (8,850)	Charcoal	Base of interlaminated clayey silt and sand	199	48°20.4’	89°21.5’	Loope (2006)
Boyd Cutbank	ETH-31438	7,995 ± 65	9,014-8,641 (8,828)	Charcoal	Base of interlaminated clayey silt and sand	199	48°20.4’	89°21.5’	Loope (2006)
Rosslyn Pit	GSC-287	9,380 ± 150	11,094-10,248 (10,671)	Wood	Base of Minong beach	227	48°21.8’	89°27.8’	Dyck *et al*,. (1966)
Cummins Pond	TO-547	9,260 ± 170	11,092-9,948 (10,500)	Conifer wood	Basal sand (Minong beach)	230	48°24.3’	89°20.9’	Julig *et al*., (1990)
Surprise Lake	BETA-230960	8,170 ± 40	9,010-9,260 (9,135)	*Betula* leaves and scales	Lake isolation	187	48°20.1’	89°49.3’	Yu *et al*, (2010)
Little Harbor	ETH-32328	8,365 ± 100	9,090-9,540 (9,315)	*Betula* leaves and twigs	Lake isolation	183	46°49.2’	85°21.7’	Yu *et al*, (2010)

Throughout Unit C, shells of bivalves and gastropods are found, the largest of which was a 7 cm diameter bivalve shell. Large, horizontal, re-deposited logs are also found within this unit. One log, located approximately 2.5 m from the base of the unit, was radiocarbon dated to 7,970 ± 30 ^{14}C (8,800 cal) yr BP (UCIAMS-26802) (Figure 6.10c). Within the bottom 1.2 m of the unit, occasional *in situ* trees extend into the unit from Unit D (Figure 6.7a) (Boyd et al. 2010).

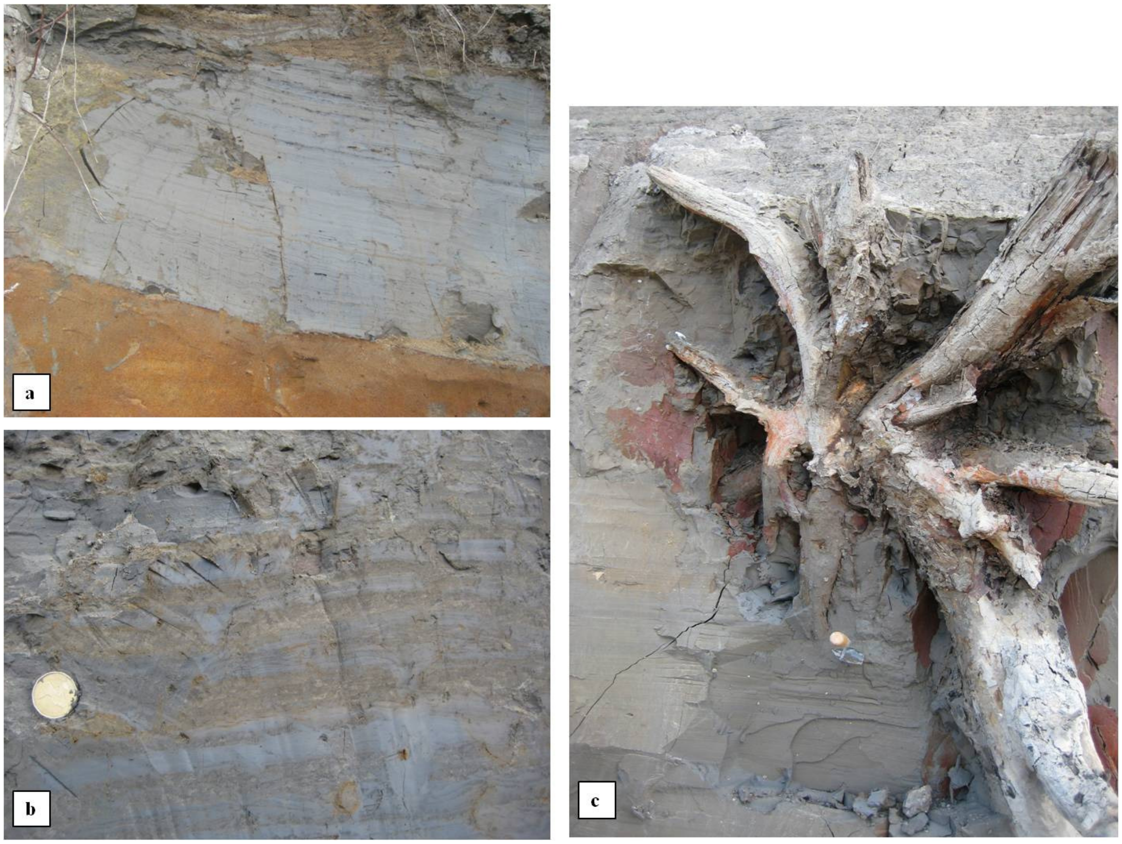


Figure 6.10. Photos showing the silty clay sediments of Unit C. Photos "a" and "b" show the rhythmites (varves) with occasional sandy laminae. Photo "c" shows the root portion of a redeposited horizontal log radiocarbon dated to 8,800 cal (7,970 ± 30 ^{14}C) yr BP.

Unit B is dominated by massive to faintly laminated silty clay (Figure 6.8) (Boyd et al. 2010). The silty clay laminae, which form couplets in some locations throughout this unit, are generally subhorizontal to horizontal and form abrupt boundaries with other laminae and the sandy beds located throughout the unit. The sandy beds range from 13-

100 cm in thickness and contain ripples, including climbing ripples, which indicate downstream directional water flow in the same direction as the modern Kaministiquia River (Boyd et al. 2010). Three of these sandy beds have fibrous/woody organic material present within them. Iron oxide mottling is common throughout this unit however, it does not follow the laminar planes in this unit and therefore is postdepositional (Boyd et al. 2010).

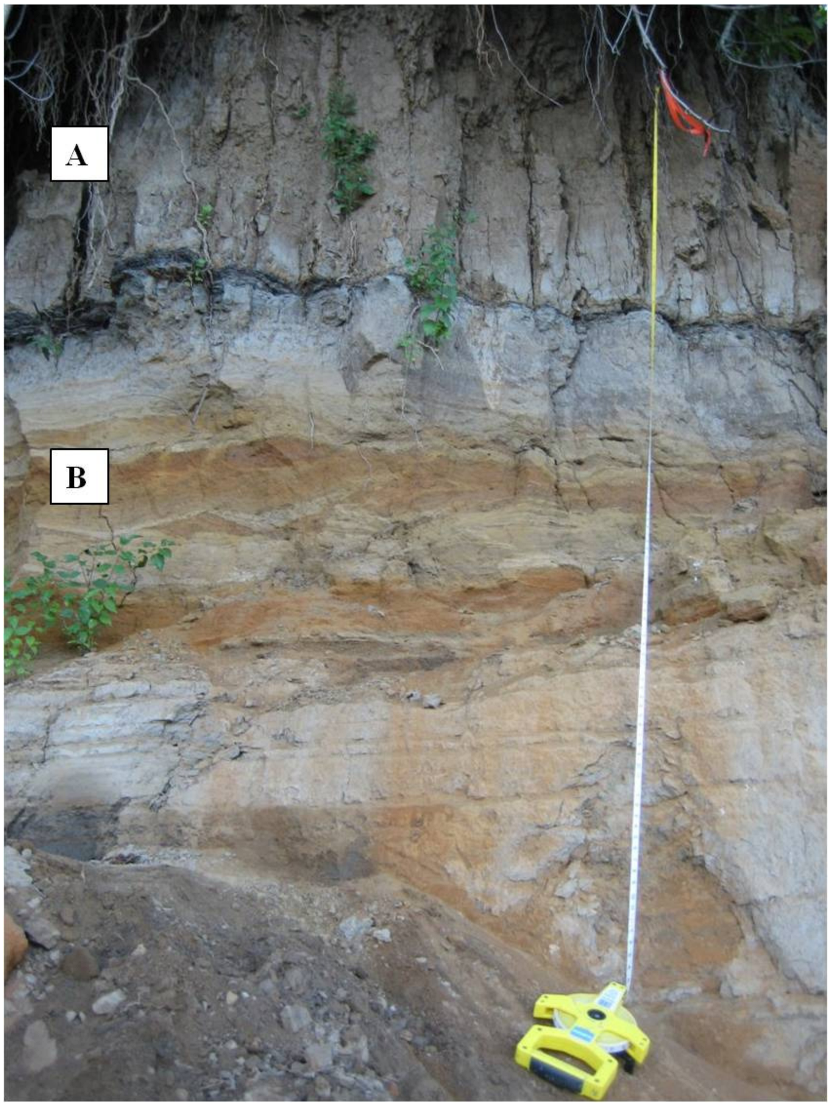


Figure 6.11. Unit A consists of the modern top soil as well as a small zone of organic laminae. Within these laminae *Scirpus* sp., achenes yielded a calibrated radiocarbon date of around 7,400 cal yr BP, indicating when these laminae were deposited. Beneath these laminae is Unit B, which is comprised of multiple laminated sand beds within predominately silty clay.

Unit A is an approximately 2.5 m thick massive to poorly laminated sandy silty unit (Figure 6.11). At the top of this unit is the modern topsoil overlying sandy silt sediment. This pedogenic unit contains prismatic peds, and iron oxide mottling is found throughout. Approximately 2 m below the topsoil lies an organic zone with roughly 13 alternating organic/silty laminae, each measuring approximately 1.5 mm in thickness. Plant macrofossils collected from this layer include bulrush (*Scirpus* sp.) and sedge (*Carex* sp.), the former providing a radiocarbon age of 6,420 ± 20 ^{14}C (7,400 cal) yr BP (UCIAMS-61733) for this zone (Boyd et al. 2010). These emergent plant taxa suggest a shallow pond environment at the time of deposition. This organic zone overlies a massive silty zone that forms an abrupt boundary with Unit B.

6.3. MACROFOSSIL DATA

6.3.1. Summary of Macrofossil Results

Macrofossil remains were collected from 14 sample locations at 5 m intervals (30E-45W) within the OFW paleochannel (Figure 6.12). Approximately 100 ml of sediment and organics were separated from each sample with just the macrofossil remains analyzed. These samples were processed following Beaudoin (2007) and were sieved using 2 mm, 1 mm, 250 µm and 125 µm mesh sizes, sorted, and then identified. In total, 17,186 macrobotanical and macrofaunal remains were counted.

Tree and shrub macrobotanical remains (n = 9,138), including leaves, seeds, twigs, cones, cone scales, new shoots and fruit scales from *Abies balsamea* (n = 4,152), *Picea* sp. (1,515), *Larix laricina* (n = 1,922), *Pinus* (n = 206) *Betula papyrifera* (n = 96), *Alnus viridis* (n = 10), *Arctostaphylos uva-ursi* (n = 31), and *Rubus idaeus* (n = 10), dominate the assemblage, totaling 53% of the total macrofossil assemblage (see Figure

6.14. for arboreal macrofossil numbers and distribution). Aquatic invertebrate remains (n = 5,488) account for 32% of the assemblage and include ephippia from the Cladoceran (water flea) order (n = 3,937), including the genus *Daphnia;* statoblasts from two bryozoans (moss animals), *Cristatella mucedo* (n = 28) and a species from *Plumatella* (n = 1,035); gemmules from the phylum Porifera (sponges) (n = 484); and bivalves from the class Ostracoda (n = 4). Bryophyte (moss) remains (n = 321) are represented by two species, *Dicranum polysetum* (n = 47) and *Pleurozium schreberi* (n = 274), which together account for roughly 2% of the total macrofossil assemblage. Terrestrial herb remains (n = 23) make up <1% of the total macrofossil assemblage and include a seed from *Lycopus americanus* (n =1) and possible seeds from the Poaceae (grass) family (n = 22). Aquatic herb macrofossils (n = 161) account for approximately 1% of the assemblage, with achenes from unidentified members of the Cyperaceae family (n =50), as well as achenes (n = 57) and perigynia (n = 2) identifiable to the Cyperaceae genus *Carex* (sedge), and achenes (n = 47) identifiable to the Cyperaceae genus *Scirpus* (bulrush). Possible seeds from the Nymphaeaceae family (n = 4) and part of a node from a possible *Equisetum fluviatile* (horsetail from the Pteridophytes Division) stem have also been identified. Fungi macrofossils are represented by sclerotia (n = 1,152) from *Cenococcum geophilum* and make up roughly 7% of the macrofossil assemblage. Insect remains (n = 635) contribute almost 4% of the total macrofossil assemblage. Lastly, taxonomically unidentified seeds (n = 146) and macrofossils (n = 120) together account for almost 2% of the remaining macrofossils assemblage (see Figure 6.13. for macrofossil percentage diagram). These remains could not be identified either from a lack of distinctive features or damage they sustained after deposition or through the sieving process after they were extracted from the cutbank.

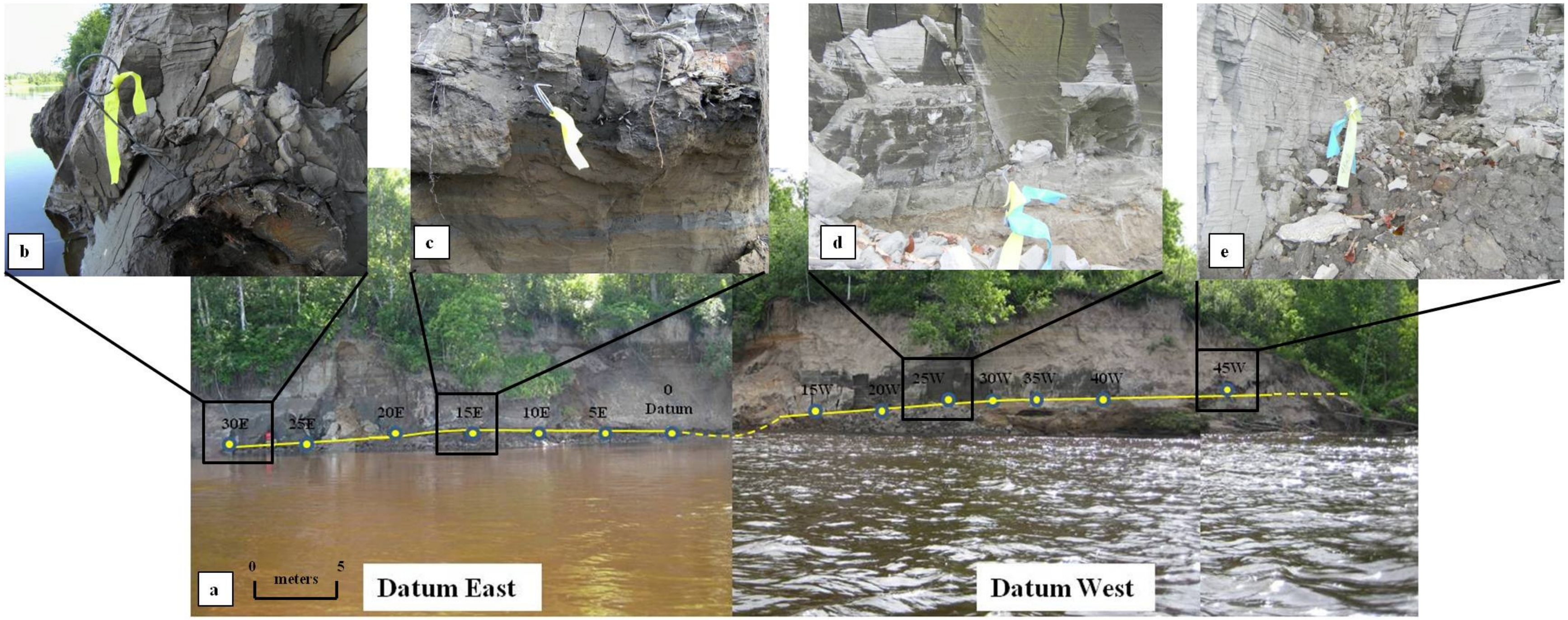


Figure 6.12. Photo "a" shows the 14 sample locations (yellow dots) obtained from Unit D (yellow line) throughout the OFW paleochannel. Photo "b" is the sample location 30E, photo "c" is the 15E sample location, photo "d" is the 25W sample location and photo "e" is is the 45W sample location.

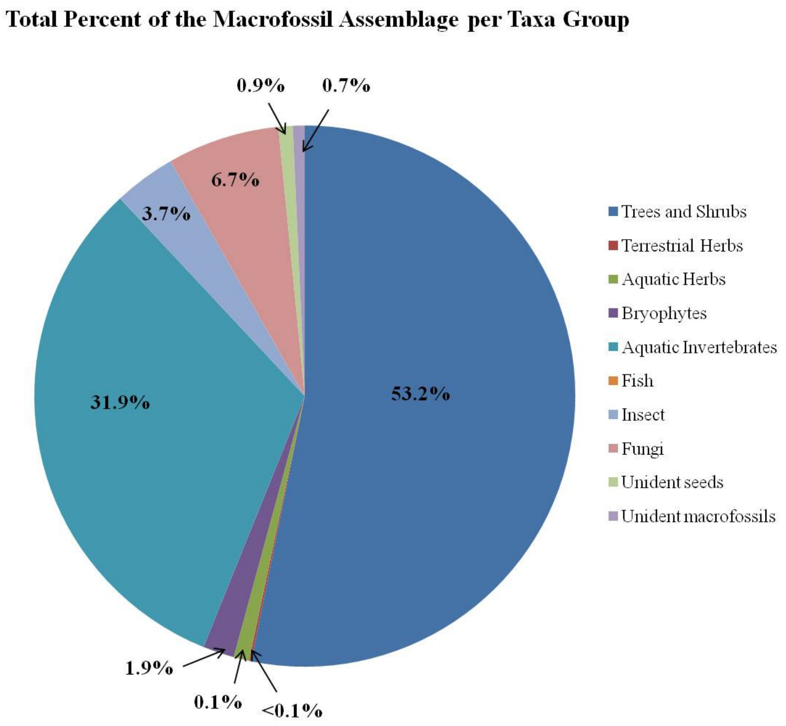


Figure 6.13. Pie chart showing percentage of total macrofossils from specific taxa groups identified within the Old Fort William cutbank.

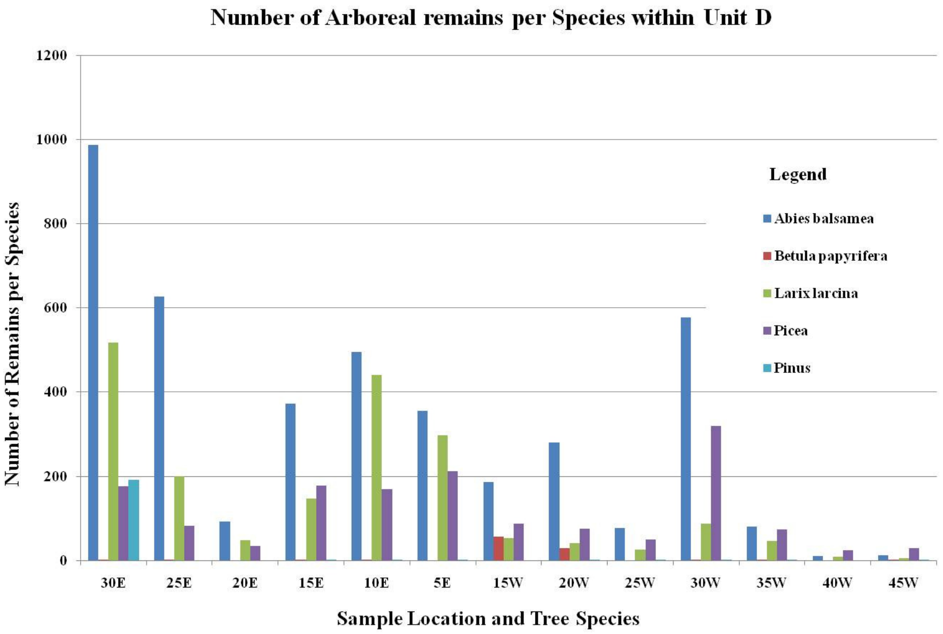


Figure 6.14. Number and distribution of tree remains in all sample locations. *Abies balsamea* generally dominates in all of the sample locations, followed by *Larix laricina*, and *Picea*. At 30E *Pinus* remains out-number *Picea* remains. *Betula papyrifera* is only a significant presence at 15W and 20W.

A total of 15 plant species are represented by the macrofossil assemblage: *Abies balsamea, Betula papyrifera, Larix laricina, Picea glauca, Pinus* sp., *Alnus viridis, Arctostaphylos uva-ursi, Rubus idaeus, Lycopus americanus,* at least two *Carex* spp., cf. *Equisetum fluviatile, Scirpus americanus, Dicranum polysetum,* and *Pleurozium schreberi*. This is a minimum number and does not include plant remains identified to the family level only, such as Poaceae and Nymphaeaceae. Of the 15 species, eight are trees and shrubs, four are aquatic herbs, one is a terrestrial herb, and two are mosses. Freshwater invertebrates are represented by at least five species including two moss animals (bryozoans), at least one species of water flea (Cladocera), one ostracod species, and at least one species of freshwater sponges (Porifera). One vertebra from an unidentified species of fish was found, as well as remains from one species of fungus. Although insect remains were numerous, they were not a central focus in this thesis and therefore all remains were counted under the title of "insect," with no distinction between species. In total, at least 22 floral and faunal species have been identified (Table 6.2). See Appendix A for photographs of macrofossils).

Table 6.2. List of taxa (scientific and common names) identified within Unit D

Taxa	**Commom Name**
Trees	
Abies balsamea	Balsam fir
Betulaceae	Birch
Betula papyrifera	White birch
Larix laricina	Tamarack/ American Larch
Picea	Spruce
Picea glauca	White spruce
Pinus	Pine
Shrubs	

Taxa	**Common Name**
Alnus viridis (= A. crispa)	Green alder
Arctostaphylos uva-ursi	Bearberry/Kinnikinnick
Rubus idaeus	Wild red raspberry
Terrestrial Herbs	
Lycopus americanus	American water-horehound
Poaceae	Grass Family
Aquatic Herbs	
Cyperaceae	Sedge Family
Carex spp.	Sedge genus
cf. *Equisetum fluviatile*	Swamp horestail
cf. Nymphaeaceae	Pond lily
Scirpus spp.	Bulrush genus
Scirpus americanus	American bulrush
Bryophytes	**Mosses**
Dicranum polysetum	Wavy moss
Pleurozium schreberi	Red-stemmed feathermoss
Cladocera	**Water flea (crustacean)**
Unidentified spp.	
Fish	
Unidentified sp.	
Fungi	
Cenococcum geophilum	
Insecta	
Unidentified	
Ostracoda	**Seed Shrimp (Crustacean)**
Unidentified sp.	
Porifera	**Freshwater Sponges**
Unidentified spp.	

The above summary of species only involves those identified within the organic layer. It should not be assumed that these are the only species that were present within the paleochannel (Birks 2001). However, macrobotanical sampling of Unit D at the OFW cutbank was comprehensive, both in terms of space, with samples taken every 5 m horizontally, and in terms of volume, with 100 ml samples, a volume considered to yield an adequate number of macrofossils for this type of study. Therefore the macroremains that were identified probably represent the dominant plant species living at the site during the time represented by Unit D, barring under-representation due to differential preservation. For example, most leaf litter in these samples was badly fragmented and could not be identified; this would mean that deciduous plants which reproduce asexually much of the time (e.g., *Populus*) would be under-represented or invisible in the macrobotanical record. This will be discussed further in Chapter 7 (Interpretations and Discussion).

6.3.2. Summary of Macrofossil Results by Sample Location

The following are the results from the 14 sample locations. This includes Datum 0W, where no macrofossils in adequate condition for identification and counting were found, a situation which eliminated this sample location from further consideration in this thesis. The order is from east to west across the OFW cutbank (Figure 6.9). Figures 6.15, 6.16, and Figure 6.17 provide complete macrofossil number and percentage results. Appendix B reports total macrofossil recoveries divided by sample location and sieve size. Macrofossil counts are presented in both percentages and numbers, with numbers in parentheses. Macrofossil numbers from each sample location are based on 100 ml of sediment taken from each sample.

30E

A total of 3,407 macrofossil remains were counted from this location. This represents 19.8% of the total macrofossil remains recovered from Unit D. Tree and shrub remains (n = 2,101) account for 61.7% of the assemblage from this sample location. Dominant arboreal remains are those from *Abies balsamea* (n = 988) and *Larix laricina* (n = 517) however, remains from *Pinus* (n = 192), *Picea* (n = 176) and *Betula papyrifera* (n = 1) are also present. Unidentified tree remains including new shoots, twigs, seeds, and seed wings, number 226. *Pinus* is best represented at this sample location, accounting for 5.6% of the total macrofossil assemblage. Remains from the shrub *Alnus viridis* (n = 1) are also present. Terrestrial herbs are represented by Poaceae (n = 1) making up < 1% of the assemblage. Bryophyte remains from *Pleurozium schreberi* (n = 3) make up < 1% of the assemblage. Aquatic herb remains (n = 15) also account for < 1% of the assemblage, with the Cyperaceae family represented by the genus *Carex* with lenticular (n =7) and trigonous (n = 1) achenes; a *Scirpus americanus* achene; and 4 achenes that could not be identified beyond the family level. Possible seeds (n = 2) from the Nymphaeaceae family were also identified.

Freshwater invertebrate remains (n = 1,126) account for 33.0% of the assemblage. Bryozoans are represented by *Cristatella mucedo* (n = 1) and the highest number of statoblasts from *Plumatella* sp. (n = 270) from all sample locations. Cladoceran ephippia (n = 825) make up most of the aquatic invertebrate remains from this sample and represent the highest number of ephippia from all sample locations. Porifera gemmules (n = 30) are also numerous, with the third highest total of all sample locations. Sclerotia from *Cenococcum geophilum* (n = 66) and insect remains (n = 76) each account for roughly 2% of the assemblage, with unidentified macrofossils (n = 14) and seeds (n = 5) accounting for < 1% of the macrofossil total.

25E

A total of 1,781 macrofossils were counted from this sample location, accounting for 10% of the total macrofossil assemblage from the OFW cutbank. Tree and shrub remains (n = 973) dominate the assemblage, accounting for 54.5% of the total macrofossils. As with 30E, arboreal remains are dominated by *Abies balsamea* (n = 627) and *Larix laricina* (n = 200), with remains from *Picea* (n = 83) and *Betula papyrifera* (n = 1) also present however, *Pinus* remains are notably absent. A stone from *Arctostaphylos uv-ursi* (n = 1) is the only shrub remain from this location. Terrestrial herbs, which are only represented by seeds from Poaceae in this sample, account for < 1% (n = 11). However, this is the highest number of grass seeds recovered from any sample location within Unit D. Aquatic herb remains account for 1.0% of the assemblage, and include achenes from unidentifiable members of the Cyperaceae family (n = 8), lenticular achenes (n = 5) from the genus *Carex,* achenes (n = 2) from the genus *Scirpus* sp., and achenes (n = 2) identified as *Scirpus americanus,* in addition to a possible seed (n = 1) from the Nymphaeaceae family. Bryophyte remains account for 7.1% of the assemblage and include both *Pleurozium schreberi* (n = 124) and *Dicranum polysetum* (n = 2); 25E is one of three sample locations where both species have been identified.

Aquatic invertebrate remains (n = 564) account for 31.6% of the assemblage, with bryozoans represented by *Plumatella* sp. statoblasts (n = 20). This sample also yielded the second highest number of Cladocera ephippia among all sample locations (n = 515). Porifera gemmules (n = 29) were also recovered. *Cenococcum geophilum* sclerotia (n =

45) make up 2.5%, insect remains (n =24) make up 1.3% , and unidentified macrofossils (n = 3) and seeds (n = 17) account for 1.1% of the total macrofossil assemblage.

20E

The total number of macrofossils counted from this location is 747. Compared to other samples, there appear to be fewer macrofossils per unit volume counted, as this sample accounts for only 4.3% of the total macrofossil assemblage. In total, macrofossils from trees and shrubs (n = 254) account for 34.0% of the assemblage. Arboreal vegetation is dominated by *Abies balsamea* (n = 93)*, Larix laricina* (n = 49)*,* and *Picea* (n = 35), with no *Pinus* remains identified. *Arctostaphylos uva-ursi* (n = 1) is the only shrub identified from this sample location, however, the Betulaceae (n = 1) family is also present. There is an absence of bryophyte and terrestrial herb macrofossils. However, aquatic herb remains include unidentifiable achenes from the Cyperaceae family (n = 3), as well as achenes from *Carex* spp., (lenticular [n = 2])*,* and *Scirpus* sp., (n = 1), making up <1% of the macrofossil assemblage.

Aquatic invertebrate macrofossils are numerous (n = 275) and make up 36.8% of this sample location assemblage. These remains include statoblasts from the bryozoans, *Plumatella* sp., (n = 10) and *Cristatella mucedo* (n = 6); the latter represents the second highest count for this species. Porifera gemmule numbers (n = 214) are the highest among all sample locations, and Cladocera ephippia (n = 45) are also present. Sclerotia from the fungi *Cenococcum geophilum* (n = 131) comprise 17.5% of the macrofossil total, while insect remains (n = 56) account for 7.5%. Unidentified macrofossils (n = 2) and seeds (n = 23) make up 3.3% of the total macrofossil assemblage for this sample location.

15E

Macrofossils counted for this sample location total 1,333. This sample location accounts for 7.6% of the total macrofossil assemblage from Unit D. Together, trees and shrubs (n = 827) account for 62.0% of the macrofossil assemblage at this location. Arboreal remains are dominated by *Abies balsamea* (n = 373), *Picea* (n = 178, including a *Picea glauca* cone scale) and *Larix laricina* (n = 148), with lower numbers of *Betula* papyrifera (n = 1) and *Pinus* (n = 1) also present. Shrub taxa include *Alnus viridis* (n = 1) and remains from the Betulaceae family (n = 2), which likely derive from either *Alnus viridis* or *Betula* papyrifera. Aquatic herb remains (n = 19), including unidentifiable achenes from the Cyperaceae family (n = 6), a *Carex* sp., perigynium, and *Carex* spp., lenticular achenes (n = 12), account for 1.4% of the total macrofossil assemblage from this sample location. Bryophyte and terrestrial herb macrofossils are notably absent.

Aquatic invertebrate macrofossils are numerous (n = 218), accounting for 16.4% of the macrofossil assemblage. Bryozoans are represented only by *Plumatella* sp. statoblasts (n = 25), as no *Cristatella mucedo* statoblasts have been identified. Porifera gemmules (n = 15) and Cladocera ephippia (n = 178) are also present at this location. *Cenococcum geophilum* sclerotia (n = 200) make up 15.0% of the remains from this sample location, the second highest number of sclerotia among all sample locations. Insect remains are abundant (n = 48) and make up 3.6% of the macrofossil assemblage in this sample. Unidentified remains account for 1.6% of the assemblage, including macrofossils (n = 3) and seeds (n = 18).

10E

The total of macrofossils counted from this sample location, number 1,879. This sample location accounts for 10.9% of the macrofossil remains from Unit D. Tree and shrub remains (n = 1,234) account for 65.7% of this sample location's macrofossil assemblage. Arboreal vegetation macrofossils are dominated by *Abies balsamea* (n = 495), *Larix laricina* (n = 441) and *Picea* (n = 169), with *Pinus* (n = 3) and *Betula papyrifera* (n = 1) also represented. Terrestrial herbs constitute < 1% of the assemblage, as Poaceae (n = 1) is the only macrofossil identified. Aquatic herb macrofossils are abundant, with unidentifiable achenes from the Cyperaceae family (n = 2), a *Carex* sp., perigynium (n = 1), and *Carex* spp., lenticular achenes (n = 18) totaling 1.1% of the macrofossil assemblage from this location. No bryophyte remains were recovered from this sample location.

Aquatic invertebrate remains (n = 476) account for 25.3% of the total macrofossil total for this location. Cladocera ephippia (n = 417) dominate the invertebrate remains, with statoblasts from *Cristatella mucedo* (n = 1) and *Plumatella* sp., (n = 32) and gemmules from freshwater Sponges (n = 26) also contributing to the assemblage. Sclerotia (n = 61) from the fungi *Cenococcum geophilum* represent 3.2% of the assemblage, with insect remains (n = 57) totaling 3.0% of the assemblage. Unidentified remains (n = 29) comprising various macrofossils (n = 21) and seeds (n = 8) account for 1.5% of the assemblage.

5E

The total number of macrofossils from this sample is 1,748. This sample location accounts for 10.2% of the total macrofossil assemblage from Unit D. These tree and shrub remains (n = 988) account for 56.5% of the macrofossil assemblage from this sample location. Arboreal vegetation is dominated by *Abies balsamea* (n = 356), *Larix laricina* (n = 293), and *Picea* (n = 213), with *Pinus* (n = 2) also represented. Shrubs are represented by stones from *Arctostaphylos uva-ursi* (n = 5) and seeds from the *Betulaceae* family (n = 7), likely from either *Betula papyrifera* or *Alnus viridis*. Bryophyte remains are represented only by *Pleurozium schreberi* (n = 68), accounting for 3.9% of the macrofossil assemblage. This number is the second highest total of red-stem feathermoss remains identified within all sample locations. Terrestrial herbs represent <1% of the assemblage, with only Poaceae remains (n = 4) identified. Aquatic herbs account for 1.4% of the assemblage, with unidentifiable achenes from the Cyperaceae family (n = 10), the genus *Carex* (lenticular [n = 10]) and the genus *Scirpus* (n = 5).

Aquatic invertebrate remains are well represented in this sample (n = 563) and account for 32.2% of the macrofossil assemblage from this sample location. Bryozoans are represented by both *Cristatella mucedo* (n = 1) and *Plumatella* sp. (n = 260) species, with *Plumatella* sp*.,* statoblasts numbering the third highest among all sample locations. Cladoceran ephippia (n = 281), Porifera gemmules (n = 20) and Ostracoda remains (n = 1) make up the rest of the invertebrate remains. Insect remains (n = 40) and sclerotia from *Cenococcum geophilum* (n = 40) each account for 2.3% of the assemblage. Unidentified macrofossils (n = 6) and seeds (n = 11), including a millimetre-scale piece of silver, represent < 1% of the total assemblage.

0W Datum

This datum point was located just above the modern river level when sampling began in summer 2008. The only macrofossils found at this location were stems from bryophytes however, the criteria used for counting bryophytes required the presence of both leaves and stem together. Therefore none of these stems were counted and this sample location was omitted from further analysis.

15W

The total number of macrofossils counted from Unit D at this sample location is 1,247 which accounts for 7.3% of the total macrofossil assemblage. In total, tree and shrub remains (n = 477) account for 38.3% of the macrofossil count from this sample location. Arboreal vegetation is dominated by remains from *Abies balsamea* (n = 186), *Picea* (n = 88), the highest number of *Betula papyrifera* (n = 57) remains among all sample locations, and *Larix laricina* (n = 54). Shrub species include *Alnus viridis* (n =5), *Arctostaphylos uva-ursi* (n = 8) and *Rubus idaeus* (n = 2). *Rubus idaeus* is only found at this sample location and at 30W. Seeds from the Betulaceae family (n = 41) are numerous however, they could not be identified to genus or species (e.g., *Betula papyrifera* or *Alnus viridis*). Terrestrial herbs include seeds from Poaceae (n = 3) which make up <1% of the assemblage. Aquatic herb remains account for 1.8% and include unidentifiable achenes from the Cyperaceae family (n = 8), with achenes from the genus *Scirpus* sp., (n = 11) and from *Scirpus americanus* (n = 3). Bryophyte remains are absent from this location.

Aquatic invertebrate remains (n = 484) comprise 38.8% of the macrofossil assemblage from this sample location, numbering just slightly above the combined tree and shrub remains total. The bryozoans are represented by statoblasts from both *Cristatella mucedo* (n = 1) and *Plumatella* sp., (n = 63). Cladocera ephippia (n = 367) are also present, and there is a relatively high number of Porifera gemmules (n = 53) compared with other sample locations, as only 25E and 45W have a higher gemmule count. Sclerotia from the fungus *Cenococcum geophilum* sclerotia (n = 201) constitute 16.1% of the assemblage, resulting in the second highest number for sclerotia within all the samples. Insect remains (n = 39) represent 3.1% of the assemblage, and unidentified remains (n = 21) comprising macrofossils (n = 8) and seeds (n = 13) account for 1.7% of the total assemblage.

20W

Total macrofossils from this sample location number 1,464, accounting for 8.5% of the total macrofossil assemblage from Unit D. Together, tree and shrub remains (n = 533) account for 36.4% of the macrofossil assemblage. Arboreal vegetation is dominated by remains from *Abies balsamea* (n = 281) and, to a lesser extent, *Picea* (n = 76, including a *Picea glauca* wing fragment), *Larix laricina* (n = 41)*, Betula papyrifera* (n = 29) and *Pinus* (n = 1). Shrubs are represented by *Alnus viridis* (n = 3), and remains from the Betulaceae family (n = 48) were also identified. Terrestrial herbs make up < 1% of the assemblage, with only remains from Poaceae (n = 1) identified. Aquatic herb remains make up just over 1% of the assemblage and include unidentifiable achenes from the Cyperaceae family (n =6), achenes from the genus *Carex* (lenticular [n = 1]), and achenes from the genus *Scirpus* (n = 12). Bryophyte remains are numerous and represent 4.2% of the assemblage. This is also the only sample location where both *Dicranum polysetum* (n = 35) and *Pleurozium schreberi* (n = 26) are present in comparatively high numbers.

Aquatic invertebrate remains (n = 472) make up 32.2% of the macrofossil assemblage. *Plumatella* sp.*,* statoblasts (n = 15) are the only bryozoan remains identified, and the number of Porifera gemmules (n = 1) is the lowest number within all sample locations, with the exception of 30W, which also only has one gemmule. Cladocera ephippia (n = 456) make up 97.0% of the aquatic invertebrate percentage. *Cenococcum geophilum* sclerotia (n = 244) account for 16.7% of the macrofossil total at this sample location, the highest number of sclerotia counted within all samples. This sample location also has the highest number of insect remains (n = 91), accounting for 6.2% of the assemblage. Unidentified remains (n = 43) comprising macrofossils (n = 25) and seeds (n = 18) make up 2.9% of the total macrofossil assemblage.

25W

The total number of macrofossils counted from this sample location is 428, and accounts for only 2.5% of the total macrofossil assemblage. Tree and shrub remains (n = 205) account for 47.9% of the macrofossil assemblage from this sample location. Arboreal vegetation from this sample location is dominated by *Abies balsamea* remains (n = 78)*, Picea* (n = 50, including *Picea glauca* cones scales), *Larix laricina* (n = 26) and *Pinus* (n = 2). Seeds from the Betulaceae family drop significantly from the previous two sample locations (15W and 20W), with only two seeds identified. There is an absence of terrestrial herb and bryophyte remains at this location, and aquatic herb remains (n = 4) account for < 1% of the assemblage with unidentifiable achenes from the Cyperaceae family (n = 1), *Carex* sp., achenes (lenticular [n = 1]), and *Scirpus* sp., achenes (n = 2).

Aquatic invertebrate remains (n = 154) constitute 36.0% of the assemblage. Statoblasts from *Cristatella mucedo* (n = 1) and *Plumatella* sp., (n = 22) are present, and Cladocera ephippia (n = 105) have also been recovered however, in a comparatively lower number than most of the other sample locations. Porifera gemmule numbers are relatively high (n = 25), and an Ostracoda shell was recovered from this sample location. *Cenococcum geophilum* sclerotia numbers are low (n = 16), accounting for 3.7% of the macrofossil assemblage. Insect remains (n = 33) make up 7.7% of the assemblage however, this is the fourth lowest number among all sample locations. Finally, unidentified remains (n = 16) including seeds (n = 10) and macrofossils (n = 6) account for 3.7% of the assemblage.

30W

The total number of macrofossils counted is 1,886 of which 1,709 have been identified taxonomically. This sample location represents 11.0% of the total macrofossil assemblage of all sample locations. Tree and shrub remains (n = 1,174) account for 62.2% of the macrofossil assemblage from this sample location. The dominant arboreal species include remains from *Abies balsamea* (n = 578), *Picea* (n = 319, including a *Picea glauca* cone), and to a lesser extent, *Larix laricina* (n = 87)*, Betula papyrifera* (n = 2) and *Pinus* (n = 2). Shrub taxa are well represented by remains of *Rubus idaeus* (n = 8)*,* and *Arctostaphylos uva-ursi* (n = 16), and remains from the Betulaceae family (n = 11) were also identified. Aquatic herbs make up < 1%, with unidentifiable achenes from the Cyperaceae family (n = 2), *Scirpus* sp.*,* achenes (n = 4), as well as a seed that possibly comes from a plant in the Nymphaeaceae family. Bryophyte remains are low at this sample location, accounting for <1% of the macrofossil assemblage, with only *Dicranum polysetum* specimens (n = 3) identified.

Aquatic invertebrate remains (n = 431) constitute 22.9% of the macrofossil assemblage from this sample location. *Plumatella* sp. (n = 49) is the only bryozoan species present at this location, with Cladoceran ephippia (n = 381) accounting for most of the invertebrate remains. Porifera gemmule numbers (n = 1) are the lowest among all sample locations, except 20W, which also has only one gemmule. This location has a high number of sclerotia (n = 140) from *Cenococcum geophilum*, making up 7.4% of the assemblage. Insect remains (n = 104) accounting for 5.5% of the total macrofossil number. This is also the only sample location where fish remains were identified, with one fish vertebrae from a small, unidentified species making up < 1% of the assemblage. Together, unidentified seeds (n = 17) and macrofossils (n = 9) account for 1.4% of the macrofossil assemblage.

35W

The total macrofossil count from this location is 526, which accounts for 3.1% of the total macrofossil assemblage from Unit D. Tree and shrub remains (n = 257) dominate the assemblage, accounting for 48.9% of the macrofossil total at this location. Arboreal vegetation remains from this sample location are dominated by *Abies balsamea* (n = 80), *Picea* (n = 74, including *Picea glauca* cone scales), and *Larix laricina* (n = 47). *Betula papyrifera* has been identified through the presence of fruit scales (n = 3), and *Pinus* is represented by one leaf tip. Terrestrial herbs are absent, and aquatic herbs account for < 1% of the macrofossil assemblage, with *Scirpus* spp., achenes (n = 4) and a possible part of a node from cf. *Equisetum fluviatile* (swamp horsetail). This is the only location where remains from this species were identified. Bryophytes represent 11.0% of the total macrofossil assemblage, with both *Dicranum polysetum* (n = 7) and *Pleurozium schreberi* (n = 51) identified.

Aquatic invertebrate remains (n = 166) make up 31.6% of the macrofossil total. *Plumatella* sp. statoblasts (n = 2) are the only bryozoans remains identified, with Porifera gemmules (n = 6) among the lowest for all sample locations. Cladocera ephippia (n = 158) account for most of the invertebrate remains. Sclerotia from *Cenococcum geophilum* (n = 4) account for < 1% of the macrofossil assemblage, while insect remains (n = 27) are comparatively low compared to other sample locations, making up 5.1% of the macrofossil assemblage from this sample location. Unidentified seeds and macrofossils together (n = 9) represent just 1.7% of the macrofossil assemblage.

40W

The total macrofossil count at this sample location is 548. This sample location accounts for 3.2% of the total macrofossil assemblage recovered from Unit D. Tree remains (n = 58) make up 10.6% of the assemblage at this location; shrubs remains are notably absent. Dominating arboreal remains are those of *Picea* (n = 24), *Abies balsamea* (n = 10), and *Larix laricina* (n = 9). There is an absence of terrestrial herbs, as in the previous sample (35W), and there are also no aquatic herbs or bryophyte species present.

Aquatic invertebrate remains (n = 445) account for 81.2% of the macrofossil assemblage, the highest percentage from a sample location. Statoblasts from *Cristatella mucedo* (n = 3) are the third highest among samples, and those from *Plumatella* sp. (n = 267) represent the second highest number among all sample locations. Cladocera ephippia (n = 165), ostracod remains (n = 1), and Porifera gemmules (n = 9) make up the remaining aquatic invertebrate numbers. Sclerotia from *Cenococcum geophilum* (n = 3) account for < 1% of the assemblage, as do unidentified macrofossils (n = 3) and seeds (n = 2). Lastly, insect remains (n = 37) make up 6.8% of the macrofossil assemblage.

45W

The total macrofossil count from this sample location is 192, which makes up 1.1% of the total macrofossil assemblage from Unit D. Shrub remains are absent from this location, but tree remains account for 30% of its macrofossil assemblage. Arboreal remains (n = 57) are dominated by *Picea* (n = 30), and *Abies balsamea* (n = 12) however, remains from *Larix laricina* (n = 5), *Pinus* (n = 2) and *Betula papyrifera* (n = 1) are also present. The only terrestrial herb identified is *Lycopus americanus* (n = 1); this is the only sample location where this species was found, making up < 1% of the macrofossil total. Bryophytes are represented by *Pleurozium schreberi* remains (n = 2), which make up 1.0% of the macrofossil assemblage. Aquatic herb remains and *Cenococcum geophilum* sclerotia are absent from this location.

Aquatic invertebrate remains (n = 114) account for 59.4% of the assemblage. This location produced the highest number of statoblasts from *Cristatella mucedo* (n = 14), while *Plumatella* sp., statoblasts were noticeably absent. Cladocera ephippia (n = 44) counts are at their lowest however, Porifera gemmules (n = 55) are the highest of all sample locations. Insect remains (n = 2) are also the lowest among all sample locations, making up 1.0% of the assemblage. Unidentified macrofossils (n = 16) account for 8.3% of the assemblage.

Table 6.3. Macrofossil type totals for each sample location within the OFW paleochannel. The number of macroremains in each cell represents the total number of macroremains counted per type (e.g., seed, leaf, cone scale) from 100 ml of sediment and organics from each sample location.

Sample Number	**30E**	**25E**	**20E**	**15E**	**10E**	**5E**	**15W**	**20W**	**25W**	**30W**	**35W**	**40W**	**45W**
Macrofossil Type													
Leaves													
Abies balsamea	983	623	93	371	490	349	186	280	76	578	79	10	12
Larix laricina	485	191	47	140	438	295	53	39	24	87	46	9	5
Picea	149	74	29	143	126	195	81	68	43	305	64	21	27
Pinus	176	0	0	0	2	2	0	1	2	2	1	0	2
Unidentified	85	38	10	69	99	60	28	47	28	136	28	8	5
Cones													
A. balsamea	0	0	0	0	0	0	0	0	0	0	0	0	0
L. laricina	5	2	0	0	1	1	0	0	1	0	0	0	0
Picea glauca	0	0	0	0	0	0	0	0	0	1	0	0	0
Picea	0	0	0	0	0	0	0	0	0	0	0	0	0
Pinus	0	0	0	0	0	0	0	0	0	0	0	0	0
Unident. conelets	0	0	0	0	2	0	0	0	0	0	0	0	0
Cone /Fruit Scales													
A. balsamea	1	2	0	0	1	0	0	0	0	0	0	0	0
B. papyrifera	0	1	0	0	0	0	20	19	0	2	0	0	0
L. laricina	5	4	0	4	0	1	0	0	1	0	1	0	0
P. glauca	0	0	0	1	0	0	0	0	3	1	2	0	0
Picea	0	0	0	0	0	0	0	0	0	0	0	0	1
Pinus	0	0	0	0	0	0	0	0	0	0	0	0	0
Unident. cone scales	10	3	5	12	11	8	0	0	3	6	2	0	0

Sample Number	**30E**	**25E**	**20E**	**15E**	**10E**	**5E**	**15W**	**20W**	**25W**	**30W**	**35W**	**40W**	**45W**
Macrofossil Type													
Twigs													
A. balsamea	3	1	0	1	2	2	0	0	0	0	0	0	0
B. papyrifera	0	0	0	0	0	0	0	0	0	0	0	0	0
L. laricina	21	1	2	3	1	1	0	1	0	0	0	0	0
P. glauca	0	0	0	0	0	0	0	0	0	0	0	0	0
Picea	22	6	4	29	41	10	6	5	2	7	4	2	0
Pinus	16	0	0	0	0	0	0	0	0	0	0	0	0
Unident.	7	0	0	2	2	0	4	5	2	0	4	4	2
New Shoots	114	13	56	15	0	29	0	0	1	6	0	1	0
Seeds													
Trees and Shrubs													
cf. A. balsamea	0	1	0	1	0	3	0	1	2	0	1	0	0
A.balsamea seed wing	1	0	0	0	2	2	0	0	0	0	1	0	0
L. laricina	0	1	0	0	1	0	1	0	0	0	0	0	0
L. laricina wing	1	1	0	1	0	0	0	1	0	0	0	0	0
Picea	3	3	2	2	1	5	1	1	2	5	3	1	1
P. glauca seed wing	0	0	0	0	0	0	0	1	0	0	0	0	0
Picea sp. seed wing	2	0	0	3	1	3	0	1	0	0	1	0	1
Pinus	0	0	0	0	0	0	0	0	0	0	0	0	0
Pinus sp. seed wing	0	0	0	1	1	0	0	0	0	0	0	0	0
Unident. conifer seeds	4	4	4	19	3	0	0	1	5	1	1	0	0
Unident. conifer seed wings	6	2	0	6	4	10	4	1	8	2	13	2	0
Alnus viridis (crispa) samaras	1	0	0	1	0	0	5	3	0	0	0	0	0
Arctostaphylos uva-ursi stones	0	1	1	0	0	5	8	0	0	16	0	0	0
Betulaceae samaras	0	1	1	2	4	7	41	48	2	11	3	0	0
B. papyrifera samaras	1	0	0	1	1	0	37	10	0	0	3	0	1
R. idaeus	0	0	0	0	0	0	2	0	0	8	0	0	0
Terrestrial Herbs													
L. americanus	0	0	0	0	0	0	0	0	0	0	0	0	1
cf. Poaceae	1	11	0	0	1	2	3	1	0	0	0	0	0
cf. Poaceae fragment	0	0	0	0	0	3	0	0	0	0	0	0	0

Sample Number	**30E**	**25E**	**20E**	**15E**	**10E**	**5E**	**15W**	**20W**	**25W**	**30W**	**35W**	**40W**	**45W**
Macrofossil Type													
Aquatic Herbs													
Cyperaceae	4	8	3	6	2	10	8	6	1	2	0	0	0
Carex perigynia	0	0	0	1	1	0	0	0	0	0	0	0	0
Carex (lenticular)	7	5	2	12	18	10	0	1	1	0	0	0	0
Carex (trigonous)	1	0	0	0	0	0	0	0	0	0	0	0	0
cf. *Equisetum fluviatile*	0	0	0	0	0	0	0	0	0	1	0	0	0
cf. Nymphaeaceae	2	1	0	0	0	0	0	0	0	1	0	0	0
Scirpus	0	2	1	0	0	5	11	12	2	4	4	0	0
Scirpus americanus (pungens)	1	2	0	0	0	0	3	0	0	0	0	0	0
Bryophytes													
Dicranum polysetum	0	2	0	0	0	0	0	35	0	3	7	0	0
Pleurozium schreberi	3	124	0	0	0	68	0	26	0	0	51	0	2
Bryozoans													
Cristatella mucedo statoblasts	1	0	6	0	1	1	1	0	1	0	0	3	14
Plumatella spp., statoblasts	270	20	10	25	32	260	63	15	22	49	2	267	0
Cladocera													
Unidentified spp. ephippia	825	515	45	178	417	281	367	456	105	381	158	165	44
Fish													
Unident. sp., vertebrae	0	0	0	0	0	0	0	0	0	1	0	0	0
Fungi													
Cenococcum Geophilum sclerotia	66	45	131	200	61	41	201	244	16	140	4	3	0
Insect													
Various spp.	76	24	56	48	57	41	39	91	33	104	27	37	2
Ostracoda													
Unident. sp.	0	0	0	0	0	1	0	0	1	0	0	1	1
Porifera													
Unident. spp. gemmules	30	29	214	15	26	20	53	1	25	1	6	9	55

Sample Number	**30E**	**25E**	**20E**	**15E**	**10E**	**5E**	**15W**	**20W**	**25W**	**30W**	**35W**	**40W**	**45W**
Macrofossil Type													
Various Macrofossils													
Unident. seeds	5	17	23	18	8	11	13	18	10	17	4	2	0
Unident. macros	14	3	2	3	21	5	8	25	6	9	6	3	16
Silver piece	0	0	0	0	0	1	0	0	0	0	0	0	0
Sample Totals	**3407**	**1781**	**747**	**1333**	**1879**	**1748**	**1247**	**1464**	**428**	**1886**	**526**	**548**	**192**
Total 17,186													

6.4. MACROFOSSIL TRENDS

Many trends are visible in the macrofossil data. These will be discussed on an east to west basis. Based on macrofossil numbers, three “zones” of change have been identified; 30E-30W, 15W-35W and 35W-45W. These “zones” are shown in figure 6.18.

6.4.1. Vegetation Trends

30E-30W: This zone is dominated mainly by conifer macrofossils. Although present throughout most of the sample locations, conifer remains including those of balsam fir, tamarack, spruce, and pine are more numerous within this zone (n = 8,548), accounting for 95.9% of the total conifer remains (n = 8,913). Conifer macrofossils begin to decrease around 15W but are still a large part of the macrofossil assemblage before their numbers drastically decline after 30W. Within this zone, aquatic herbs are also abundant (n = 156), making up 96.9% of the total aquatic herb assemblage. Remains of aquatic herbs found between 30E and 30W comprise unidentifiable Cyperaceae (n = 50), *Carex* spp. (n = 59), *Scirpus* spp. (n = 37), and *Scirpus americanus* (n = 6) and Nymphaeaceae (n = 4). However the number of macrofossils drops significantly after 15W, with relatively few of them appearing at any sampling location further west. Poaceae remains (n = 22) within the zone 30E to 30W account for 95.7% of the terrestrial herb total (n = 23), with only one terrestrial herb specimen, a *Lycopus americanus* seed, found in the 35W-to-45W zone. *Cenococcum geophilum* sclerotia (n = 1,145) are consistent throughout all sample locations within the 30E to 30W zone and account for 99.4% of the total number of sclerotia from all sample locations, falling off west of 30W. Insect remains are also consistently found in all sample locations within the 30E-to-30W zone, but are also numerous in the 35W-to-45W zone, which has been assigned as a separate zone due to its low yield of other macrofossil remains. Although there is no apparent pattern in the distribution of red-stemmed feathermoss remains (n = 221), the 30E-to-30W zone yielded 80.7% of the total macrofossil remains of this species.

15W-30W

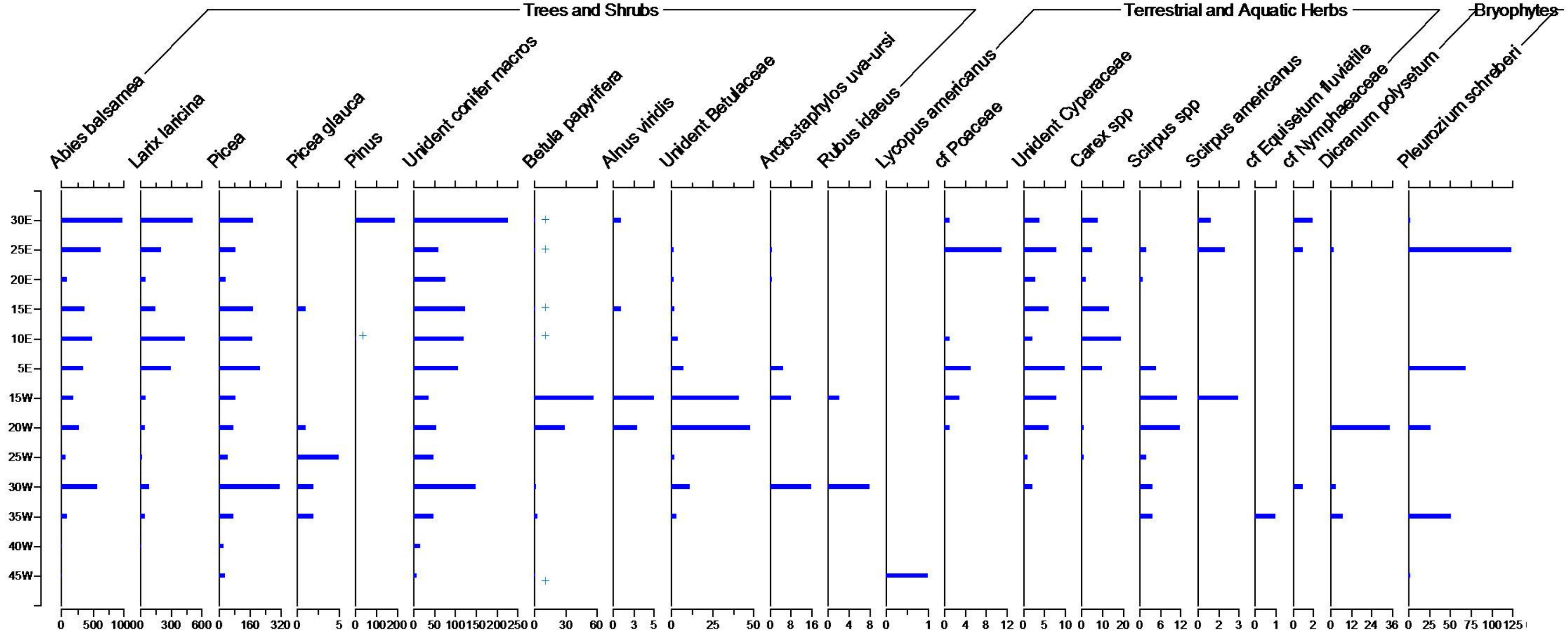
The major vegetation changes observed in this zone are an increase in white birch and shrub remains. At 15W there is a slight decline in conifer remains, which up to this point, comprised the majority of the arboreal species represented in the remains from the sample locations. This decrease coincides with an increase in seeds and fruit scales from white birch (n = 47). These remains account for 87.0% of the macrofossils from this species, as only 7 other white birch macrofossils have been identified in the other sample locations. In the sample locations east of 15W, macrofossils from shrub taxa are few (n = 9) however, between 15W-30W the number dramatically increases (n = 42). The number of shrub remains does not include Betulaceae seeds, which could not be identified to species level. Other shrub macrofossils from the 15W-to30W zone comprise green alder (n = 8), bearberry (n = 24), and raspberry (n = 10), accounting for 82.4% of the total shrub macrofossil assemblage. Although wavy moss numbers (n = 47) are low throughout the entire paleochannel sampled, the 15W-to30W zone represents 80.9% of the macrofossils identified to that species (n= 38).

35W-45W

The trend observed in this zone is a lack of macrofossil remains compared to all other sample locations. Conifer remains (n = 365) from the 35W-to-45W zone make up only 4.1% of the conifer macrofossil assemblage, with balsam fir, tamarack, and spruce all identified. Aquatic herb remains (n = 5) from this zone account for only 3.1% of the group's total assemblage, with low numbers of *Scirpus* spp. achenes (n=4), and the only macrofossil of what could be swamp horsetail. As previously mentioned, sclerotia from *Cenococcum geophilum* drop significantly within this zone and account for just 0.6% of the total number of remains for this species.

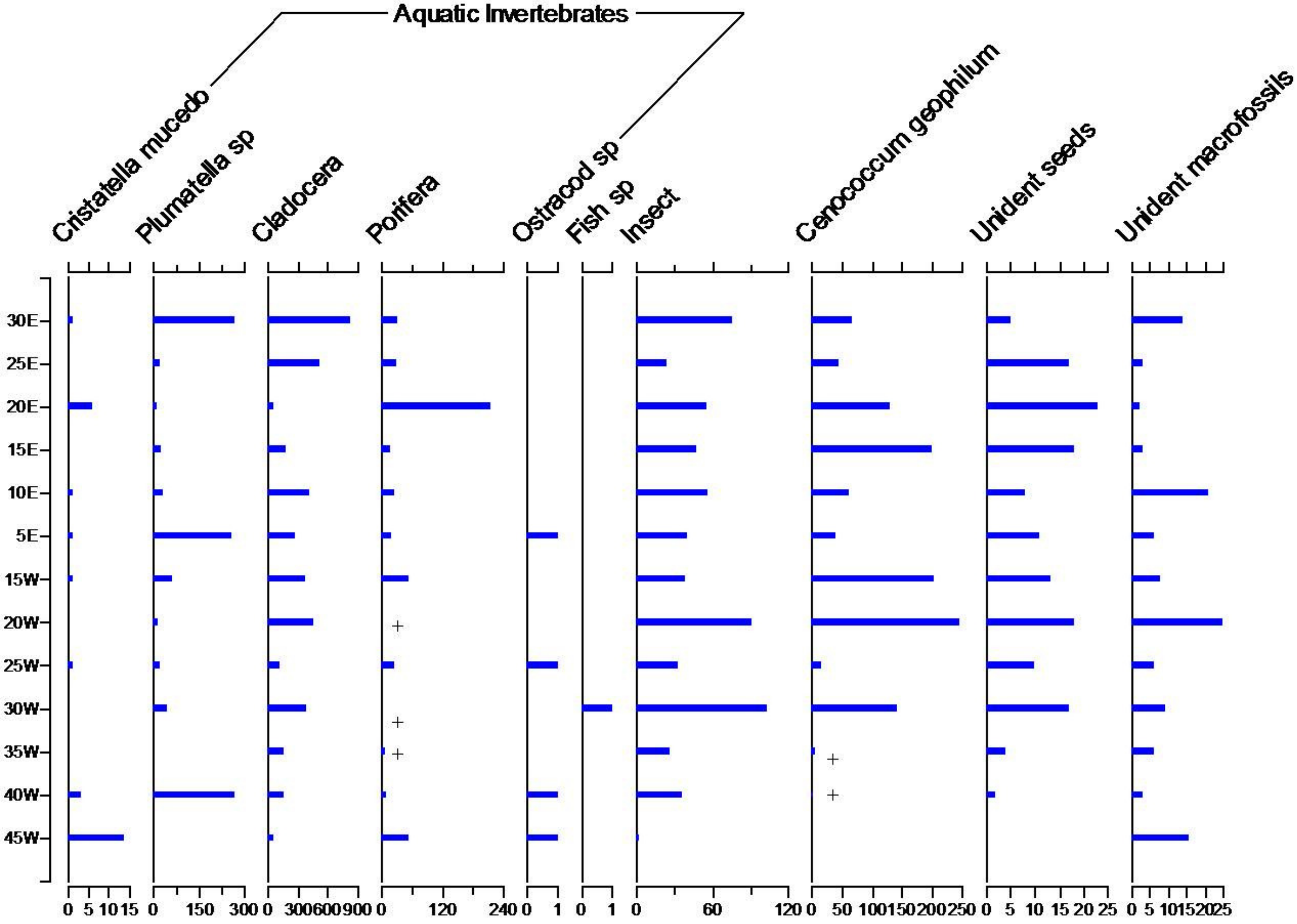
6.4.2. Microfaunal Trends

No major trends are observed in the aquatic invertebrate data. Macrofossils from the freshwater invertebrates (bryozoans, poriferans, ostracods, and cladocerans) are fairly consistent throughout all of the sample locations. Ostracod remains are limited (n = 4), with 3 bivalve shells found west of the datum. Only one fish remain was identified within the entire paleochannel, which makes identifying trends difficult.



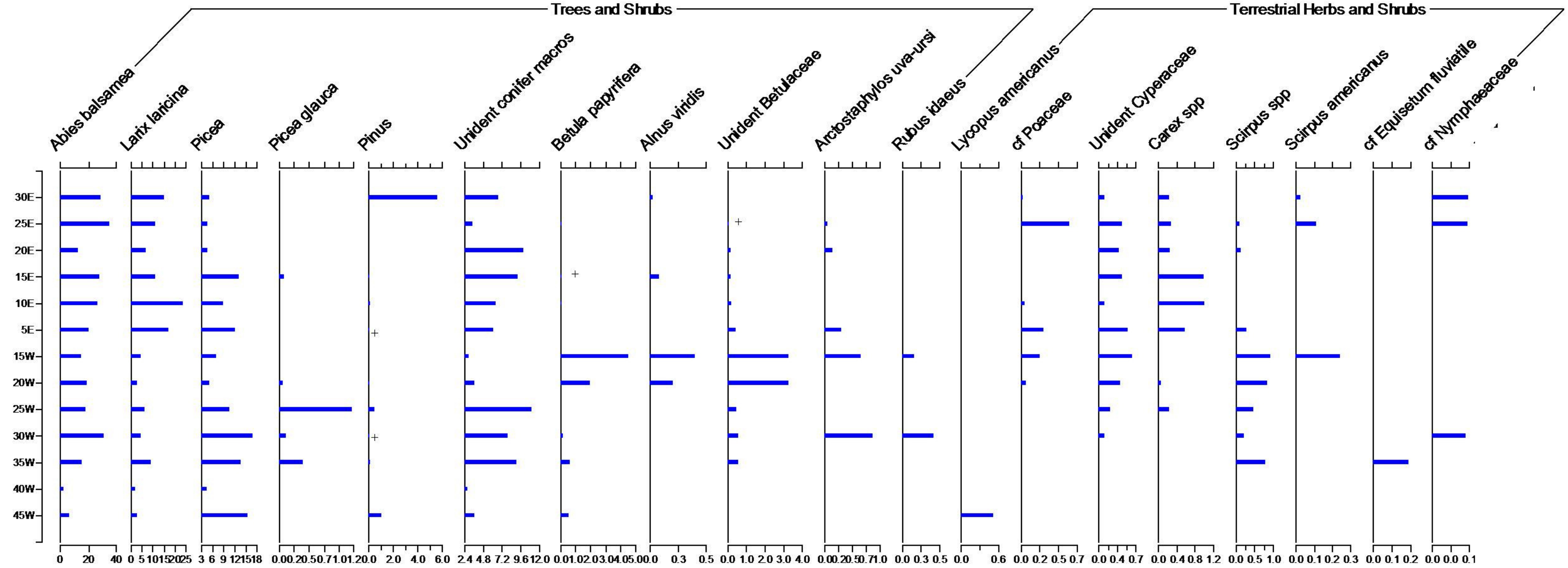
The '+' symbol indicates the presence of a macrofossil type, however the number is minimal (n = <3) compared to other sample locations

Figure 6.15. Macrofossil concentration diagram from the OFW paleochannel with all identified taxa and unidentified macrofossils represented (bars). Macrofossil are represented by number per 100 ml of sediment and organics.



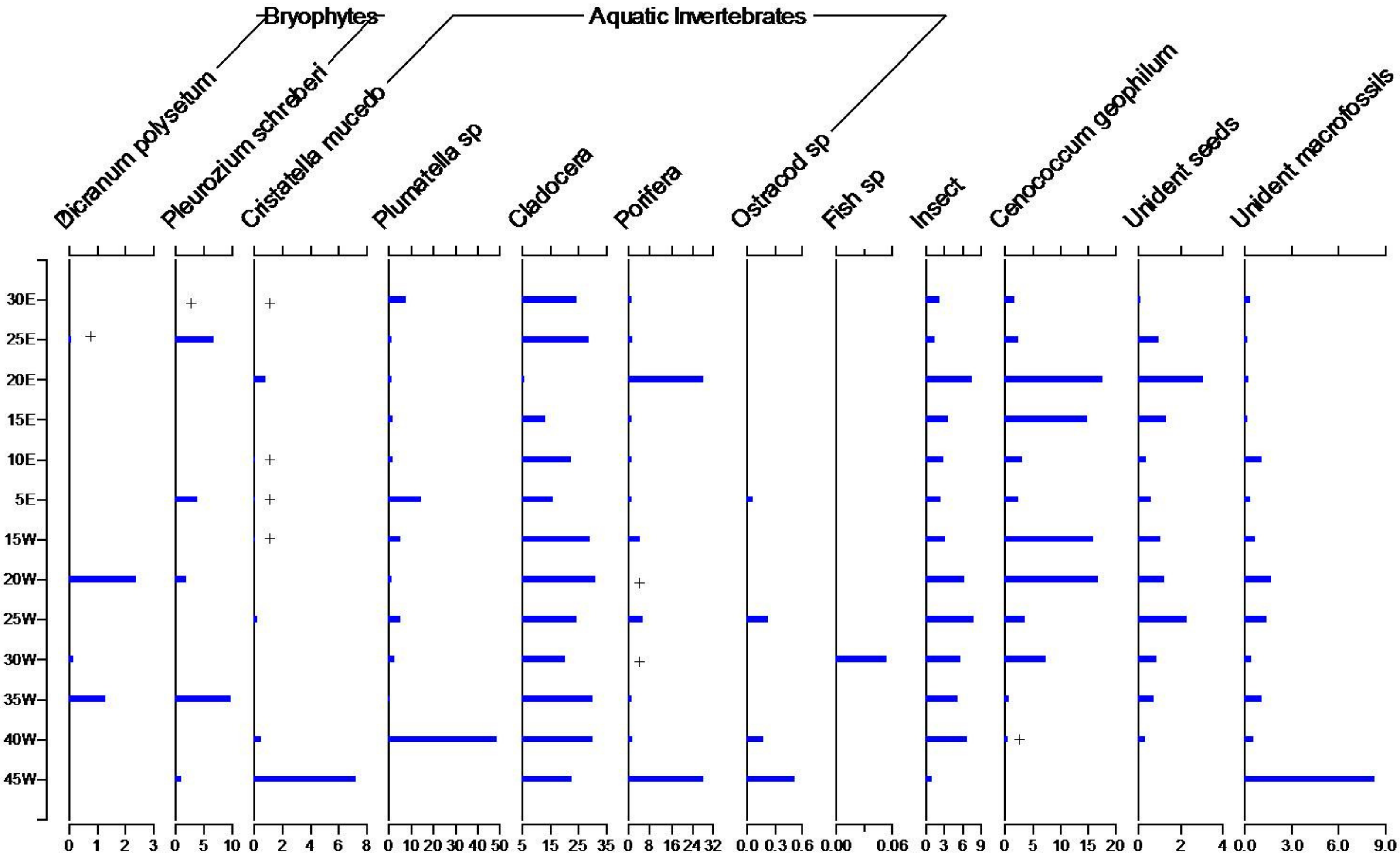
The '+' symbol indicates the presence of a macrofossil type, however the number is minimal (n=<4) compared to other sample locations

Figure 6.15 (continued). Macrofossil concentration diagram from the OFW paleochannel with all identified taxa and unidentified macrofossils represented (bars). Macrofossils are represented by number per 100 ml of sediment and organics.



The '+' symbol indicates the presence of remains from the particular taxa, however it represents a small percentage of the sample location 's macrofossil assemblage

Figure 6.16. Macrofossil percentage diagram from the OFW paleochannel with all identified taxa and unidentified macrofossils represented (bars). Percentages are based on number per 100 ml of sediment and organics.



The '+' symbol indicates the presence of remains from the particular taxa, however it represents a small percentage of the sample location 's macrofossil assemblage

Figure 6.16 (continued). Macrofossil percentage diagram from the OFW paleochannel with all identified taxa and unidentified macrofossils represented (bars). Percentages are based on number per 100 ml of sediment and organics.

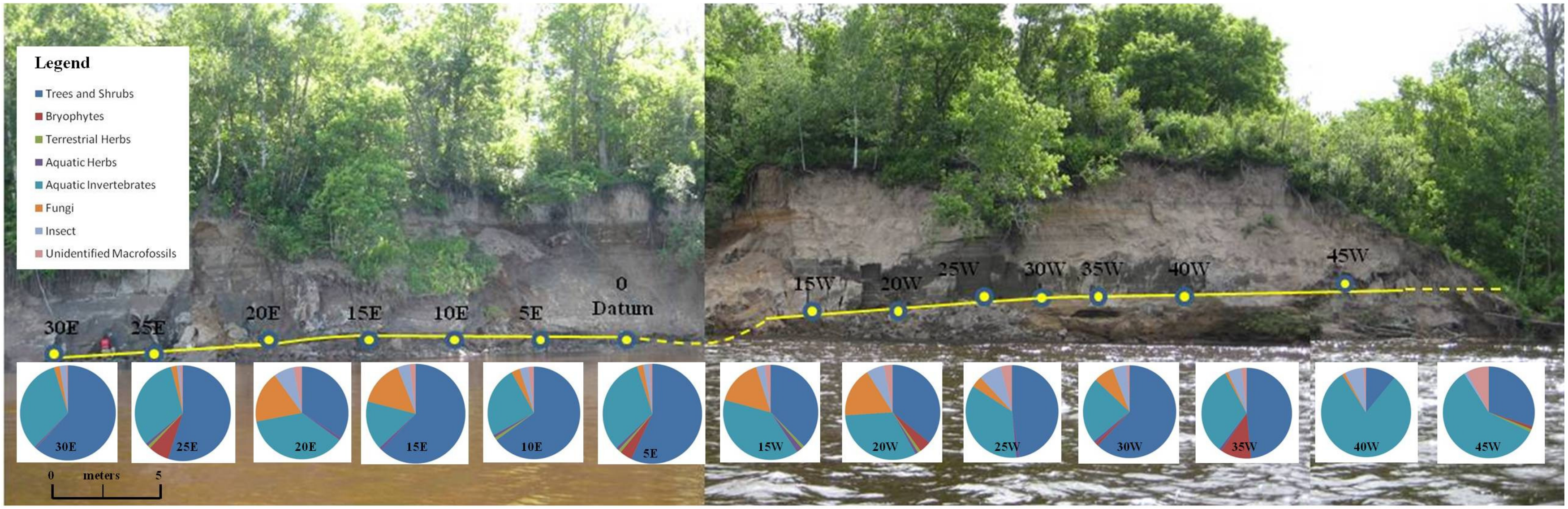
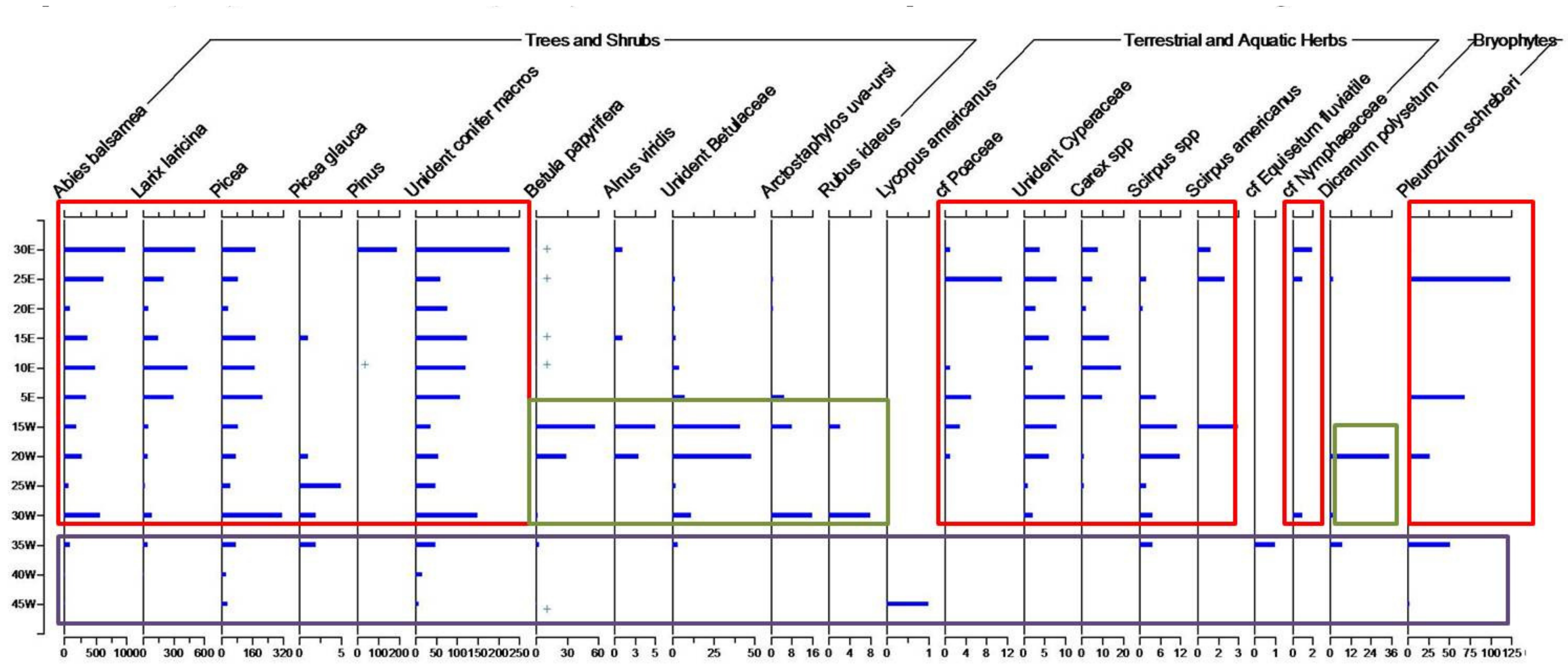


Figure 6.17. Distribution of sample locations within the OFW paleochannel with taxa group percentages represented in pie charts for each sample location. Sample locations are approximately 5 m apart



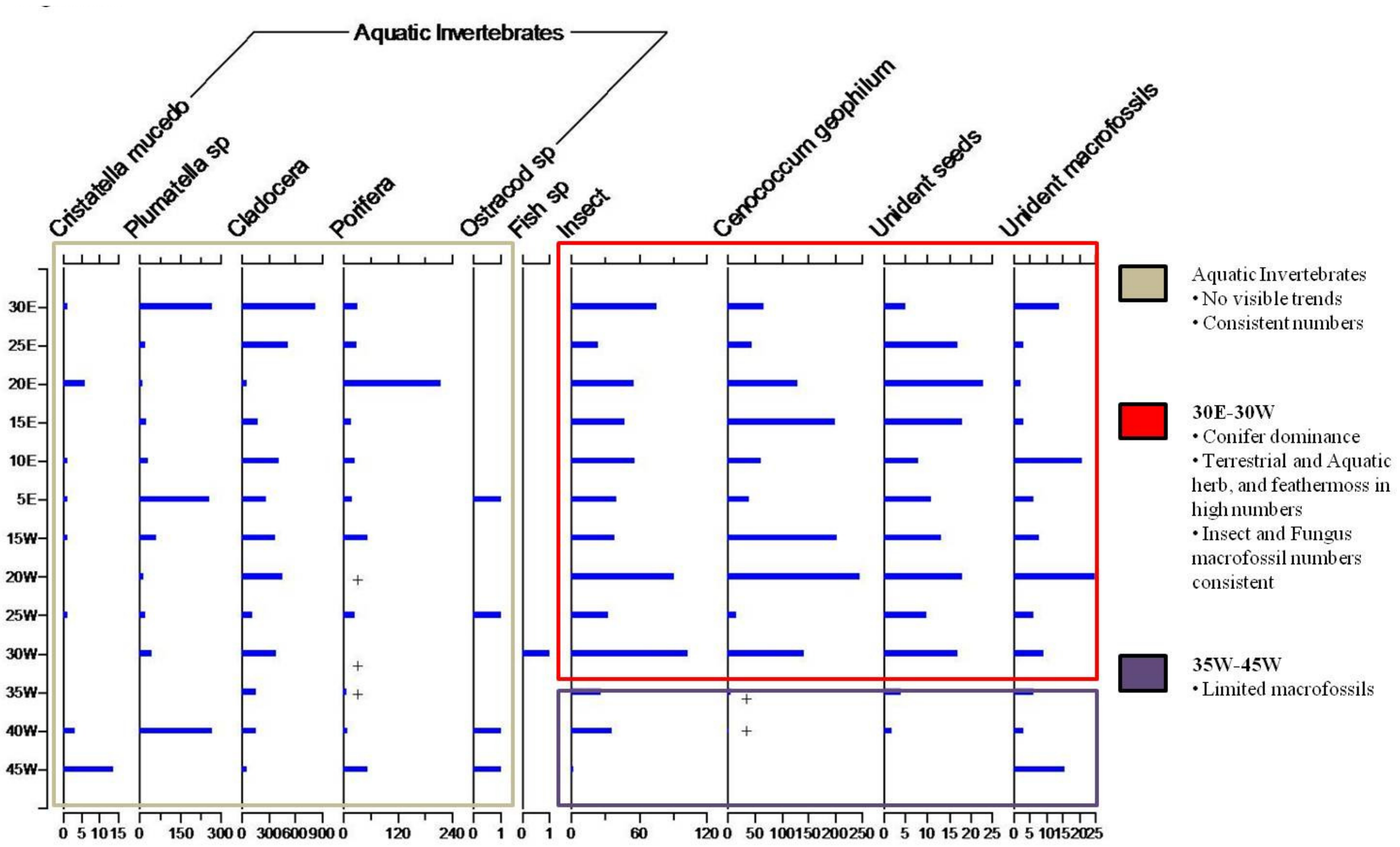
The '+' symbol indicates the presence of a macrofossil type, however the number is minimal (n = <3) compared to other sample locations

🟥 **30E-30W**
- Conifer dominance
- Terrestrial and Aquatic herb, and feathermoss in high numbers
- Insect and Fungus macrofossil numbers consistent

🟩 **15W-30W**
- Increase in white birch macrofossils
- High macrofossil count of shrubs
- Wavy moss count higher

🟪 **35W- 45W**
- Limited macrofossils

Figure 6.18. Macrofossil concentration trend diagram from the OFW paleochannel with all identified taxa and unidentified macrofossil remains represented (bars). Trends are outlined by boxes. Macrofossil numbers are based on 100 ml of sediment and organics.



The '+' symbol indicates the presence of a macrofossil type, however the number is minimal (n = <4) compared to other sample locations

Figure 6.18 (continued). Macrofossil concentration trend diagram from the OFW paleochannel with all identified taxa and unidentified macrofossil remains represented (bars). Trends are outlined by boxes. Macrofossil numbers are based on 100 ml of sediment and organics.

6.5. SUMMARY

The macrobotanical assemblage consists of a variety of vegetation types, including arboreal species, shrubs, herbs, and mosses. The macrofossil assemblage is dominated by conifer remains, with a heavy presence of *Abies balsamea, Larix laricina*, and *Picea*, with some *Pinus* remains. Deciduous trees are represented by *Betula papyrifera* however, macrofossils from this species are low compared to the conifer remains. Stones from *Arctostaphylos uva-usri* and *Rubus idaeus,* and seeds from *Alnus viridis* are also small in number but provide important information on understory vegetation, as does the presence of mosses, grasses, sedges and bulrush.

Although not a central focus of this thesis, aquatic invertebrate remains are the second most numerous macrofossil group within the assemblage. Remains from the various aquatic invertebrate taxa identified mainly consist of asexually generated reproductive bodies like sclerotia, gemmules and statoblasts, which serve to protect these species' genetic material so the organism can hatch when environmental conditions are favourable. These taxa may be useful in reconstructing the paleoenvironment as they may indicate the presence of aquatic settings.

The 22 plant and animal taxa identified offer enough information to allow for a detailed paleoenvironmental reconstruction, which will be discussed in the following chapter. With the addition of the stratigraphic analysis as well as controlled radiocarbon dates from within the cutbank series, as well as other locations around the Thunder Bay area, this reconstruction can be further refined and placed within a sequence of deglaciation, paleohydrological and climatic events.

CHAPTER 7

INTERPRETATIONS AND DISCUSSION

7.1. PALEOVEGETATION RECONSTRUCTION OF THE OLD FORT WILLIAM (OFW) PALEOCHANNEL (9,100-8,900 cal yr BP)

7.1.1. Paleochannel Vegetation ~8,900 cal yr BP

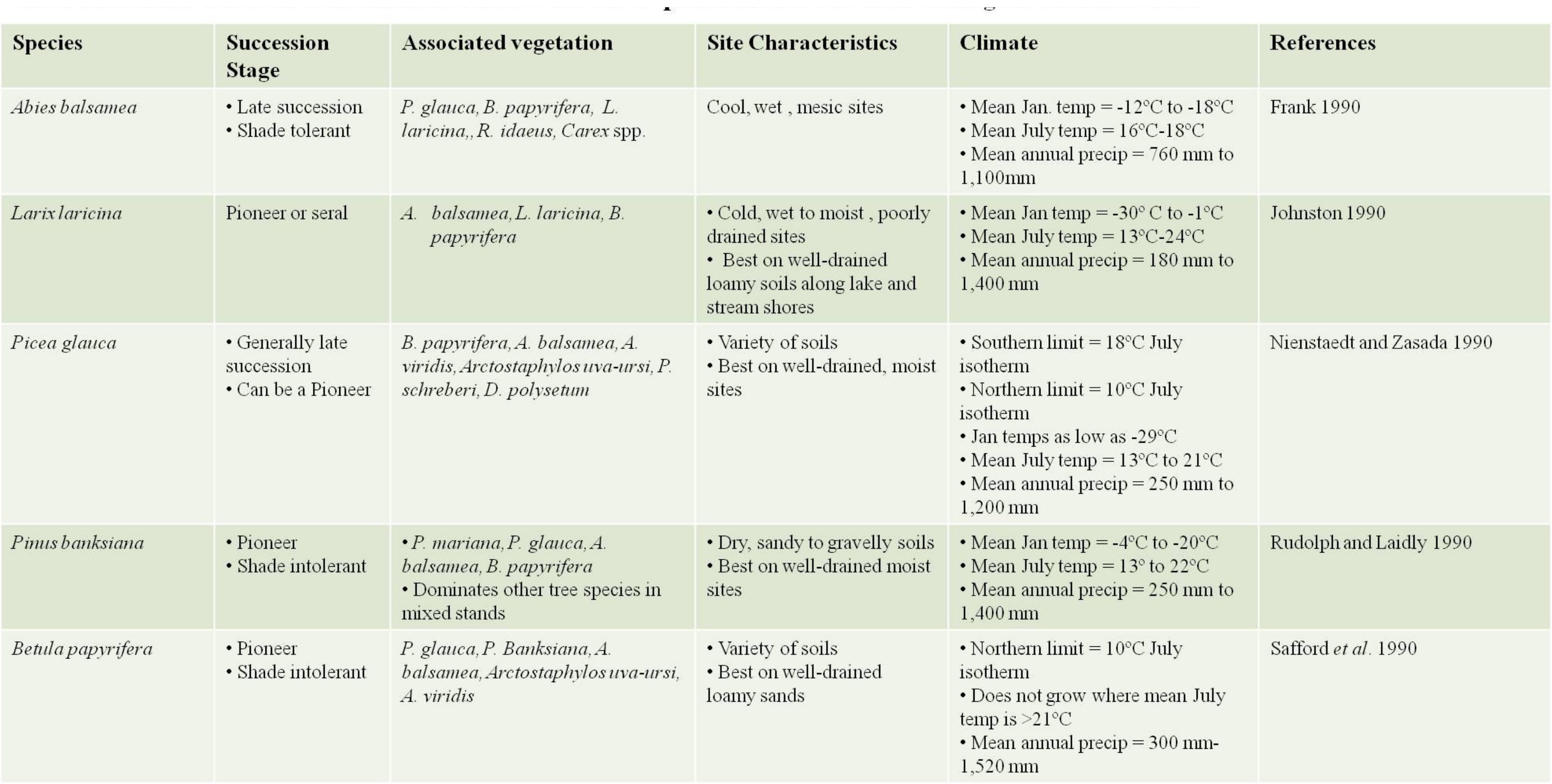
The macrobotanical remains recovered from Unit D represent vegetation that was growing in the OFW paleochannel at least by 8,900 cal yr BP. When plotted spatially, trends in vegetation are visible across the cutbank which may reflect small-scale changes in vegetation within the paleochannel. Apart from a small area (< 1 m) where there was a minor disturbance in Unit D, there appears to be no major disturbance or unconformity within the unit. The possibility of re-deposition of macroremains by wind or water from regional vegetation sources is minimal because: 1) no eolian or fluvial sediments have been identified within the Unit D organic deposit; 2) the presence of *in situ* trees indicates local presence of arboreal vegetation and 3) delicate macrobotanical remains that would likely have been destroyed if transported to the site have been recovered (e.g., Cyperaceae perigynia). Therefore, the species identified within Unit D most likely reflect vegetation that was growing in or near the paleochannel prior to 8,900 cal yr BP.

Based on the macrobotanical remains, the arboreal vegetation growing in the paleochannel likely represents either a conifer or mixed conifer-hardwood boreal assemblage (Figure 7.1). This forest (OFW forest) was likely dominated by white spruce, balsam fir, and to a lesser extent, tamarack.

Identifying all of the spruce remains recovered from Unit D to the species level was not possible because leaves could not be distinguished between white spruce (*Picea glauca*) and black spruce (*P. mariana*). However, the presence of white spruce cones, cone scales and seeds, as well as hairless twigs (black spruce twigs have hairs), indicate that white spruce was growing in the sample zone. Although the presence of black spruce cannot be ruled out, this species was likely not present in the paleochannel because the nutrient poor and well-drained sands were probably not suitable for black spruce which typically grows in peat (Farrar 2007). Liu (1990) also states that black spruce probablybecame established after white spruce following deglaciation because the sands and gravels deposited by the LIS and meltwater rivers would have been similarily well drained, a situation that is more suitable for the growth of white spruce and that was probably key in its pioneering presence following deglaciation (Liu 1990).

Within the OFW forest, white spruce may have been a dominant arboreal species because it generally comprises a significant part of arboreal vegetation when present in mixed stands (Table 7.1) (Nienstaedt and Zasada 1990). However, it likely co-dominated alongside balsam fir based on the large number of macrobotanical remains from this latter species in Unit D. Tamarack (*Larix laricina*) was probably only a minor component of the OFW forest because it is not numerous when present in mixed stands due to competition for resources, including sunlight (Johnston 1990). This is supported by the relatively low number of tamarack leaves recovered from Unit D compared to the other conifer leaves. Since tamarack leaves are shed every year, it is expected that they would be over-represented if it was a major arboreal species (Farrar 2007; Johnston 1990).

Table 7.1. List of arboreal taxa identified within the OFW paleochannel and their ecological characteristics.



Species	Succession Stage	Associated vegetation	Site Characteristics	Climate	References
Abies balsamea	• Late succession • Shade tolerant	*P. glauca, B. papyrifera, L. laricina,, R. idaeus, Carex* spp.	Cool, wet , mesic sites	• Mean Jan. temp = -12°C to -18°C • Mean July temp = 16°C-18°C • Mean annual precip = 760 mm to 1,100mm	Frank 1990
Larix laricina	Pioneer or seral	*A. balsamea, L. laricina, B. papyrifera*	• Cold, wet to moist , poorly drained sites • Best on well-drained loamy soils along lake and stream shores	• Mean Jan temp = -30° C to -1°C • Mean July temp = 13°C-24°C • Mean annual precip = 180 mm to 1,400 mm	Johnston 1990
Picea glauca	• Generally late succession • Can be a Pioneer	*B. papyrifera, A. balsamea, A. viridis, Arctostaphylos uva-ursi, P. schreberi, D. polysetum*	• Variety of soils • Best on well-drained, moist sites	• Southern limit = 18°C July isotherm • Northern limit = 10°C July isotherm • Jan temps as low as -29°C • Mean July temp = 13°C to 21°C • Mean annual precip = 250 mm to 1,200 mm	Nienstaedt and Zasada 1990
Pinus banksiana	• Pioneer • Shade intolerant	• *P. mariana, P. glauca, A. balsamea, B. papyrifera* • Dominates other tree species in mixed stands	• Dry, sandy to gravelly soils • Best on well-drained moist sites	• Mean Jan temp = -4°C to -20°C • Mean July temp = 13° to 22°C • Mean annual precip = 250 mm to 1,400 mm	Rudolph and Laidly 1990
Betula papyrifera	• Pioneer • Shade intolerant	*P. glauca, P. Banksiana, A. balsamea, Arctostaphylos uva-ursi, A. viridis*	• Variety of soils • Best on well-drained loamy sands	• Northern limit = 10°C July isotherm • Does not grow where mean July temp is >21°C • Mean annual precip = 300 mm-1,520 mm	Safford *et al.* 1990

The recovery of a jack pine cone from the Boyd paleochannel indicates the presence of at least this species of pine within the local area. Regional pollen records also indicate the presence of red pine during this time since jack pine and red pine pollen are indistinguishable from one another, resulting in the possibility that one or both species were regionally present. No macrobotanical remains from red pine have been recovered from Unit D indicating that this species was at least, not growing in the paleochannel. Macrobotanical remains from pine are limited throughout the entire sample area, and those that have been recovered are severely fragmented. Since conifer trees produce many needles which are easily preserved in anaerobic conditions, it is expected that if pine was growing in the paleochannel, then the number of macrobotanical remains would be more numerous and less fragmented. The ecology of both jack and red pine also argue against the presence of this species in the OFW forest. In mixed stands within the boreal forest, jack pine is generally associated with trembling aspen, white birch, balsam fir, and black spruce, and is not generally present in stands dominated by white spruce (Rudolph and Laidly 1990). Red pine can also be found in mixed forest stands with jack pine, white pine or trembling aspen, but is not generally found with white spruce or tamarack (Rudolph and Laidly 1990). Both jack pine and red pine are shade intolerant, suggesting that the jack pine identified in the macrofossil assemblage and any associated red pine may not have been in the immediate vicinity of the paleochannel, but instead grew nearby in a more exposed upland setting (Rudolph and Laidly 1990).

The OFW forest would be considered a mixed conifer-hardwood stand if white birch was growing within the paleochannel. Although the total number macrobotanical remains from this species is relatively low compared to the other arboreal species, Birks (1973) states that when found in sediments, seeds and fruit scales from white birch most likely represent a local presence. However, if this species represented a significant part of the arboreal vegetation within the paleochannel, the number of remains would likely be higher since this species produces a significant number of seeds per tree (Birks 2003). Therefore, if white birch was growing in the paleochannel, it would likely have only been a minor component of the arboreal vegetation. If it was not growing directly in the paleochannel, it could have been located nearby, perhaps with jack pine, in the surrounding upland areas.

The understory of the OFW forest likely consisted of sedges and rushes, grasses, *Lycopus americanus* (water hore-hound), *Equisetum fluviatile* (swamp horsetail), *Arctostaphylos uva-ursi* (bearberry) and *Rubus idaeus* (red raspberry) (Table 7.2). Macrobotanical remains from these taxa are low throughout the entire paleochannel, indicating that the OFW understory may have been open (Martin and Gower 1996). However, macrobotanical remains such as leaves from deciduous vegetation may be under-represented since they are not as numerous and are more fragile than conifer leaves. Mosses, including *Pleurozium schreberi* (red-stemmed feathermoss) and wavy moss (*Dicranum polysetum*), were likely growing on the forest floor. Macrobotanical remains from red-stemmed feathermoss were uncommon throughout the sampled area, suggesting that it had not formed the kind of mat which often occurs within mature, closed forest stands (Johnson et al. 1995). Wavy moss remains were also uncommon, suggesting that this species may not have been abundant on the forest floor.

Spatial trends of vegetation throughout the paleochannel were more visible with the understory vegetation. Taxa including sedges and rushes, grasses, and red-stemmed feathermoss were predominately in the lower portion of the paleochannel, within the 30E-

30W zone, but particularly more concentrated between 30E and the 0W datum. Shrub species such as *Arctostaphylos uva-ursi* (bearberry) and *Rubus idaeus* (red raspberry) appear to be more prevalent on the paleochannel slope, within the 15W-to-30W zone.

Sedges are mostly associated with moist or wet sites (Newmaster et al. 1997), and red-stemmed feathermoss are found on dry or moist sites in sheltered areas (Johnson et al. 1995). Bearberry is shade intolerant and is mainly found in dry, open areas (Soper and Heimburger 1994), and raspberry is also found in open areas where the soil is dry and nutrient poor (Johnson et al. 1995). Although macrobotanical remains from wavy moss were only recovered in four sample locations, three of these locations were on the western paleochannel slope/margin, suggesting this species may have been more abundant there than in the lower portion of the paleochannel. Wavy moss is most commonly found in open forests on dry or moist sites (Johnson et al. 1995; Viereck and Schandelmeier 1980). These trends at least suggest that the lower portion of the paleochannel may have been more densely vegetated, with a closed or semi-open forest canopy, as opposed to the paleochannel slope, which may have been sparsely vegetated or occupied by open canopy forest. Since no *in situ* trees were located on the paleochannel slope, it is possible that the slope was occupied by more understory vegetations, mainly shrubs, but with few or no trees. These trends may also reflect differences in moisture throughout the paleochannel, which will be discussed later in this chapter.

Since the OFW forest has both shade tolerant (e.g., balsam fir, white spruce, red-stemmed feathermoss) and shade intolerant species (e.g., tamarack, raspberry), the forest was likely mature (Martin and Gower 1996). The presence of an understory that was open or semi-open, as suggested by the limited macrobotanical remains from understory

Table 7.2. List of shrub and moss species identified within the OFW paleochannel and their ecological characteristics.

Species	Succession Stage	Associated Vegetation	Site Characteristics	References
Alnus viridis	• Pioneer or seral	• Open conifer forests	• Open areas • Moist or dry soils • Edges of wetlands or streams	Ringius and Simms 1997
Arctostaphylos uva-ursi	• Seral • Shade intolerant	• Pinus, Picea, and Betula forests • Open to semi-open forests	• Dry, nutrient poor soils • Along lake or river shorelines	Ringius and Simms 1997 Johnson et al. 1995 Soper and Heimburger 1994
Rubus idaeus	• Seral	• Understory in P. Banksiana, P. glauca, P, mariana, and A. balsamea dominated forests	• Open woods and thickets • Infertile, slightly acidic soils • Best on well-drained, nutrient rich soils	Johnson et al. 1995
Pleurozium schreberi	• Abundant in late succession forests • Fire intolerant	P. glauca, P. mariana, Arctostaphylos uva-ursi	• Mesic forests • Tolerates dry and nutrient poor soils • Prefers closed forest stands	Johnson et al. 1995
Dicranum polysetum		• Pinus forests • Thuja (cedar) or deciduous swamps	• Humus or decaying wood • Open, dry or moist forests • Mineral soils	Johnson et al. 1995 Viereck and Schandelmeier 1980

species, supports that the forest was mature, because mature forests tend to have open understories and closed canopies (Martin and Gower 1996). However, as previously stated, understory species such as deciduous shrubs may be under-represented due to limited production of macrobotanical remains such as seeds or limited preservation of remains such as leaves. The presence of red-stemmed feathermoss also supports that the forest was mature, since it is a late succession species, most commonly found in mature, close-canopied forest stands (Johnson et al. 1995; Soper and Heimburger 1994).

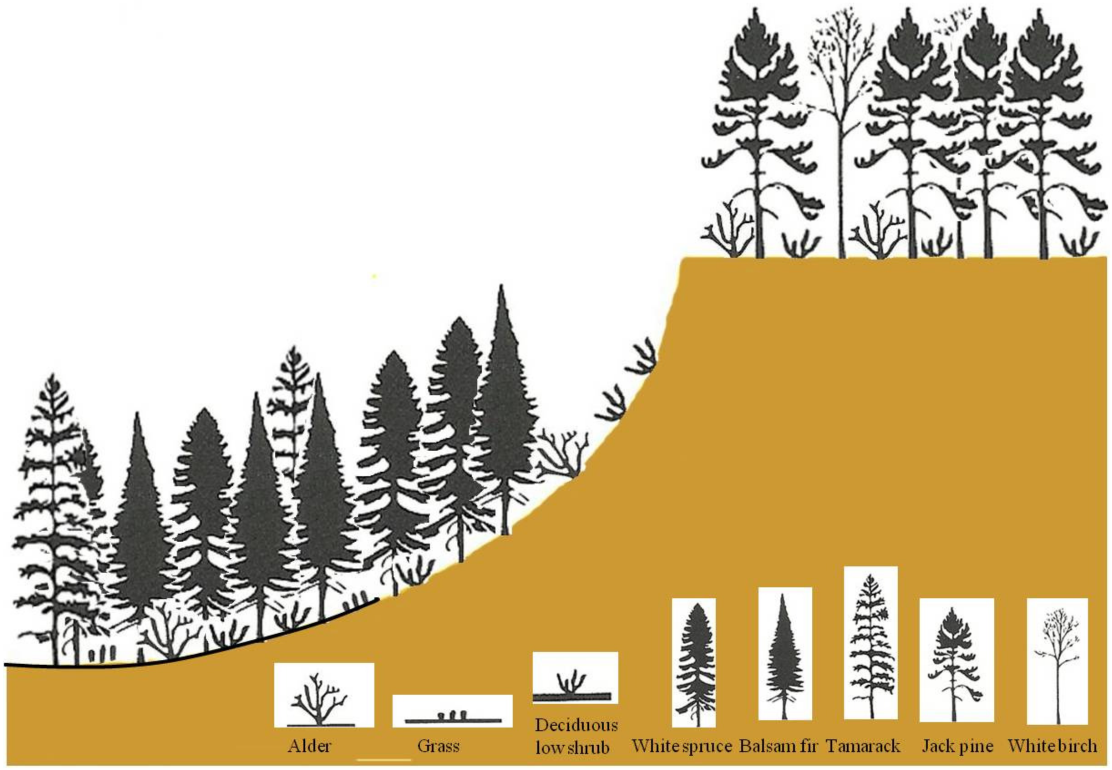


Figure 7.1. Generalized reconstruction of the vegetation growing within the OFW paleochannel around 8,900 cal yr BP. Dominant tree species are white spruce and balsam fir, with lesser quantities of tamarack. Deciduous low shrubs include bearberry and raspberry; the alder icon represents green alder. Red-stemmed feathermoss and wavy moss is represented by the black line along the paleochannel bottom. Note that this visual representation only reflects presence of taxa identified within the paleochannel and possible abundance and location of species based on spatial trends and species ecology; it is not an exact replica of the forest.

The moisture and organic content of the soils within the paleochannel around 8,900 cal yr BP may have varied due to topographic differences between the paleochannel slope and bottom. The differences in topography may also have affected the distribution of plant species, which may be identified though spatial trends in the macrobotanical data. Macrobotanical remains from white spruce, balsam fir, tamarack, and most of the aquatic herb remains were recovered from sample locations mainly within the lower portion of the paleochannel, indicating that it was a moist site. Vegetation on the paleochannel slope suggests that the soil there was dry. The presence of bearberry, an indicator species of dry and nutrient poor soils (Ringius and Simms 1997), on the slope, in addition to the decreasing number of aquatic herb remains in this area, support this inference. The soil on the slope may not have been suitable for tree species, including tamarack, balsam fir or white spruce, which prefer moist soils (Farrar 2007); this may account for the decrease in macrobotanical remains from these species in this area. Raspberry macroremains, which were only identified from sample locations on the slope, support the idea that the slope was either under an open forest canopy or wass less vegetated by arboreal vegetation since this species prefers more open areas (Johnson et al 1995).

The macrofossil assemblage may not only represent vegetation that was growing within the paleochannel immediately prior to its burial, but may also reflect changes in vegetation growing within the paleochannel through time. This possibility is discussed below.

7.1.2. Vegetation Change within the Paleochannel between ~9,100-8,900 cal yr BP

Based on the mean of the maximum and minimum radiocarbon ages from Unit D (Table 6.1 for a complete list of dates), there was an approximate interval of at least 200 years (9,100-8,900 cal yr BP) for organic accumulation within the paleochannel. It is unclear if organic remains began to accumulate immediately following the abandonment of the paleochannel however, enough time elapsed for the paleochannel to change from non-vegetated to forested between the time it ceased to carry stream flow and its subsequent burial.

In the modern boreal forest, it has been noted that young forests of approximately 50 years of age already have late succession species present (e.g., balsam fir) (Bergeron and Dubuc 1989). This is, however, if the process of succession is initiated by burning of the previous climax community. Seeds from tree and shrub species are able to survive forests fires if they are not severe, which allows for more rapid forest regeneration (Bergeron and Dubuc 1989). Soils, in addition to the organic matter may also still be present after forest fire, resulting in more rapid vegetation regeneration which does not exist on sites that are initially bare of soil. Studies on recently deglaciated land in Glacier Bay, Alaska, revealed that glacial till became forested within 100 years after exposure (Chapin et al. 1994). Within 20 years after the surface was ice-free, lichens, liverworts and forbs, colonized the area followed by the dwarf shrub *Dryas drummondii* after 30 years. Alder was the dominant vegetation by 50 years after exposure and persisted until 100 years, after which it was replaced by a Sitka spruce forest (Chapin et al. 1994). This suggests that previously non-vegetated and nutrient poor soils in a northern setting can become forested within 100 years of exposure after deglaciation.

Although the sediment within the paleochannel was previously unvegetated and likely nutrient poor, it is possible that it was able to support a mature forest community within at least 50 years and likely no longer than 100 years after it ceased carrying stream flow, since vegetation was present in the surrounding area prior to its exposure.

The reconstruction of the OFW forest is based on the presence of species in the paleochannel around 8,900 cal yr BP however, the presence of these species and their abundance within the forest may have changed within the 200 year period (Table 7.1). As previously mentioned, white spruce was likely a pioneering species in the paleochannel and may have remained a major forest component throughout the entire lifespan of the OFW forest, possibly only changing in abundance as later successional species moved in. Tamarack may have also been a pioneering species. However this likely depends on moisture levels at the site since it prefers moist settings (Farrar 2007). If the OFW site was initially moist around 9,100 cal yr BP, then it could have been a pioneer species. If the site was was too dry for tamarack then it would probably have occupied the site when site conditions became more favourable, at least by 8,900 cal yr BP.

White birch may have also been a pioneering species within the paleochannel. This species may have initially occupied the paleochannel alongside white spruce however, as competition for resources including sunlight increased, white birch may have decreased (Farrar 2007; Safford et al. 1990). White birch can persist in spruce-fir dominated stands as the stand matures, but their numbers tend to decline (Safford et al. 1990). If white birch was growing in the OFW forest, it was probably more abundant initially but rarer by 8,900 cal yr BP. The limited number of fruit scales and seeds identified within Unit D may also support this scenario as these remains could be from the few remaining white birch trees in the forest.

Balsam fir likely became part of the OFW forest after spruce and possibly birch were established. Balsam fir is a late succession species and will move into sites where arboreal species are already present (Farrar 2007). This is because it is a shade tolerant species (Frank 1990).

Soils within the paleochannel may have also varied through time in response to varying environmental or site characteristics that changed throughout the period that it was occupied by vegetation. The paleochannel was abandoned during a time of a more arid and possibly warmer climate (i.e., the Hypsithermal), which may have been, in part, responsible for the decline in water level within the Lake Superior basin during the Houghton phase (Boyd et al. 2010). After the paleochannel was abandoned by the river, pioneer vegetation would have grown directly on top of the nutrient poor sands of Unit E. The presence of sclerotia from *Cenococcum geophilum* suggest that at some point when the channel was vegetated, the soil was likely nutrient poor and possibly dry. When present in dry or nutrient poor soils, *Cenococcum geophilum* and arboreal vegetation share resources (Pigott 1982). The fungi provides the tree with nutrients including phosphorus, nitrogen and water, in return, the fungi gets carbon from the tree (Pigott 1982). These resources are shared through sclerotia which bore into tree roots (Pigott 1982). It is possible that the sclerotia were blown into the site however, as previously discussed, there was little evidence of eolian sediment in Unit D. The distribution of the sclerotia throughout the sample area coincides with the distribution of arboreal remains (30E-30W). This could indicate that the macrobotanical remains from the fungi and arboreal species were contemporaneous. If this is the case, it could indicate that the soil within the paleochannel was dry, as least for a brief period prior to 8,900 cal yr BP.

With the presence of aquatic vegetation (e.g., pond lilies, sedges, horsetail, water-horehound) and tamarack, at least part of the paleochannel was moist. The recovery of poriferan, ostracod, bryozoan, and cladoceran remains also suggests the nearby presence of freshwater and a moist habitat between 9,100 and 8,900 cal yr BP. Moist conditions likely occurred within the paleochannel at least by 8,900 cal yr BP, as water levels in the adjacent Lake Superior basin began to rise at this time. However, given the presence of taxa that prefer dry and moist habitats, it is possible that a range of moisture conditions existed in the channel at any given time.

As mentioned in the previous chapter, the organic layer between 30E and 0W Datum is relatively thick, measuring 5-9 cm as compared to sample locations west of the 0W datum (15W-45W) where the layer is between 0.1 - 5 cm thick. The thinning of the organic layer corresponds to an increase in the paleochannel slope; with a thicker layer in what would have been the lower portion of the paleochannel and a thinner layer on what would have been the paleochannel slope. The irregularity in the thickness of the organic layer may represent differences in preservation, rate of organic accumulation due to varying biotic productivity, and/or degree of erosion from one location to another. It is unlikely that there was post-depositional disturbance, as the paleochannel was rapidly buried by sediment and inundated by rising water levels within the Lake Superior basin. Although preservation of macroremains was excellent across the entire sample area, it is possible that there were differences in preservation and possibly even rates of accumulation throughout time.

Since the number of organic remains recovered from sample locations on the paleochannel slope are low in number compared to those in the lower portion of the paleochannel, it can be argued that the slope may not have been as biologically productive compared to the paleochannel bottom, a situation that may have been important in creating the much thinner organic layer in this part of the exposure. However, if the slope remained drier and anaerobic conditions were not present to allow for the accumulation of organics throughout the approximately 200 year period, then macrofossil remains that were preserved on the slope may only represent vegetation that was growing immediately prior ~8,900 cal yr BP, when burial of the site may have significantly improved its ability to preserve organic remains. This is supported by the excellent preservation of macroremains that were recovered from these sample locations. Without anaerobic conditions, organic remains would break down more quickly (Dincauze 2003), meaning that the slope may not have been less biologically productive, only the macrobotanical remains would not have been preserved compared to those in the lower portion of the paleochannel where anaerobic conditions may have been present, allowing accumulation of macrobotanical remains in a thick deposit that developed over many years. Still, even if the macrofossil remains only represent the final vegetation, the slope may have been more sparsely vegetated, vegetated by plants that did not produce many macrobotanical remains, or the slope had few arboreal species as previously discussed.

The OFW forest indicates that boreal forest was established in the area by at least 9,100 cal yr BP. This forest was comparatively similar to the modern boreal forest in Northwestern Ontario, incorporating spruce, balsam fir, tamarack, various understory species and moss covering the floor (Henry 2002). Regional pollen records indicate that a non-analogous spruce-oak-ash forest was present in the late-glacial period in Northern Ontario, prior to the growth of the OFW forest (Björck 1985; Julig et al. 1990; Liu 1990). As discussed in Chapter 3 (Paleovegetation and Climate), the pollen from oak and ash may have been transported into this area and may not represent local populations. However, if this forest type existed in the Thunder Bay area, it was likely replaced by boreal forest vegetation by at least 9,100 cal yr BP.

7.1.3. Paleoclimate of the Thunder Bay region during the early Middle Holocene

In Northwestern Ontario, the Middle Holocene is identified as a time of climatic and vegetative transition (Julig et al.1990; Björck 1985). The climate during the Middle Holocene was warmer and drier compared to the cold and moist Early Holocene (Julig et al. 1990). This warm and dry period is known as the Hypsithermal (Web III et al. 1983). Although initially beginning with the retreat of the LIS and manifesting first on the Plains around 17,200 cal yr BP, the most severe effects occurred between 11,000-8,900 cal yr BP and ended around 6,800 cal yr BP (Williams et al. 2010). The increasing aridity was time-transgressive from west to east (Bartlein et al. 1998), with warmer and drier climate conditions in Minnesota between 11,500-7,900 cal yr BP (Clark et al. 2001). The later occurrence in the east may be attributed to the continued presence of the LIS and/or proximity to glacial lakes, including Lakes Agassiz and Minong, which may have kept Minnesota and Northwestern Ontario cooler (Wright 1983). Pollen records from Cummins and Oliver Ponds (Julig et al. 1990) and Isle Royale (Flakne 2003) indicate a decline in spruce and an increase in pine (likely jack pine) and white birch at this time (Figure 7.2). Increasing pine and herb pollen values were also found in Hayes Lake near

Kenora however, the herb pollen likely records the migration of prairie vegetation as it expanded into Manitoba and Minnesota (McAndrews 1982).

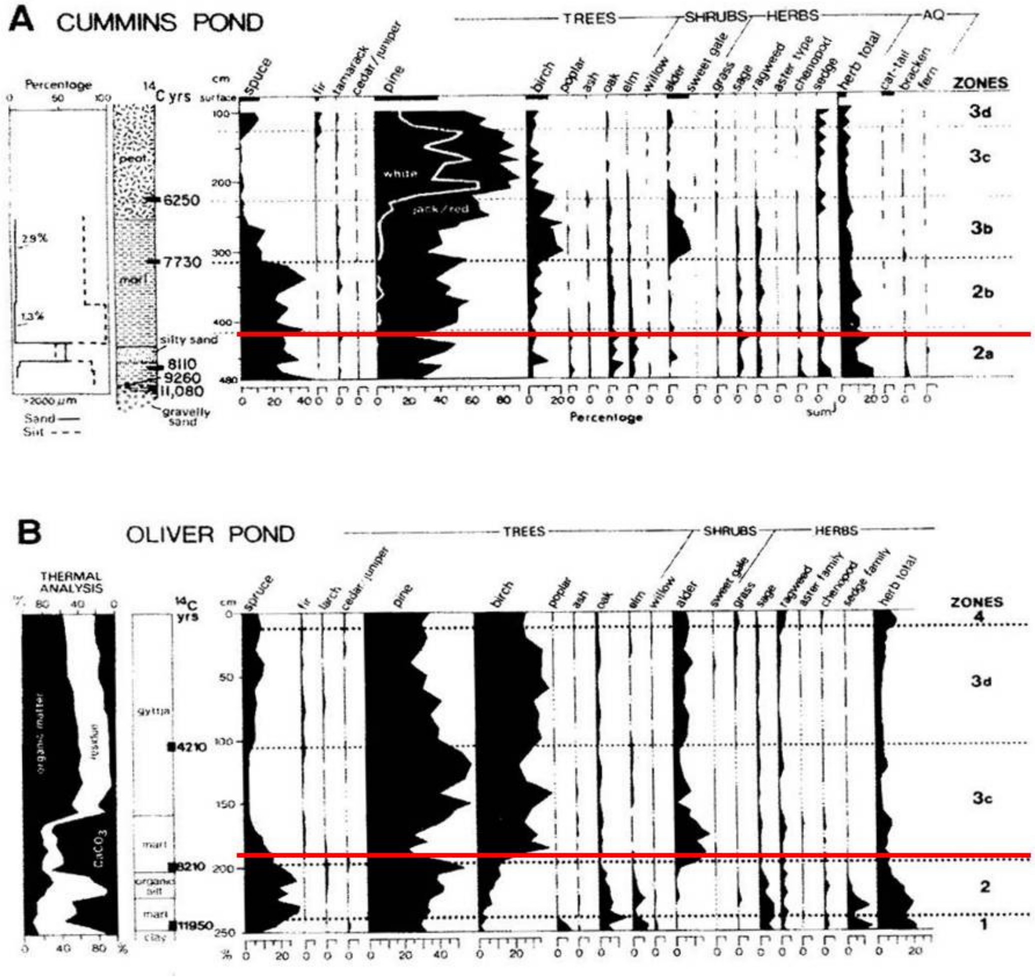


Figure 7.2. Pollen diagrams from the Cummins and Oliver Ponds. Early vegetation was herbs and shrubs indicating a tundra landscape (Zone 1), quickly followed by the migration of spruce (Zone 2). As the climate became drier and possibly warmer around 8,900 cal (8,000 ^{14}C) yr BP, jack pine and birch replaced spruce (Zone 3). Note that white pine moves into the Thunder Bay area ~7,200 cal (6,250 ^{14}C) yr BP. After 4,800 cal (4,200 ^{14}C) yr BP, spruce replaces some pine as the climate becomes cooler and more moist (from Julig et al. 1990). The red line shows the date of the growth of the OFW forest.

Pollen records, however, indicate regional changes in vegetation and may not represent vegetation at a local scale (Dincauze 2003). The macrobotanical remains from the OFW paleochannel provide better insight into local (i.e., Lower Kaministiquia River valley) vegetation and climate during this period. The macrobotanical remains indicate the presence of white spruce, balsam fir, tamarack, green alder, possibly white birch, bearberry, raspberry, water hore-hound, sedges, grasses, and red-stemmed and wavy feathermosses in the lowland areas by 8,900 cal yr BP. Uplands were likely dominated by jack pine and white birch, as suggested by Julig et al., (1990) and the limited remains recovered from the paleochannel. Balsam fir may have also been located in the upland areas however, because it is a poor pollen producer with poor pollen dispersal, it is under-represented in lake sediments unless it grew near the water-body that was cored (Birks 2003). This may also be said for tamarack however, this species is more common in the wetter lowland areas (Farrar 2007; Johnston 1990), mitigating the potential for it to be under-represented in lacustrine pollen assemblages.

Based on the temperature ranges of balsam fir, tamarack, white spruce, jack pine, and white birch, mean January temperatures around 8,900 cal yr BP at the OFW location were likely between -4ºC and -18ºC, with mean July temperatures between 13ºC and 18ºC (Frank 1990; Johnston 1990; Nienstaedt and Zasada 1990; Rudolph and Laidly 1990; Safford et al. 1990). Mean annual precipitation was likely between 180 mm and 1,100 mm (Frank 1990; Johnston 1990; Nienstaedt and Zasada 1990; Rudolph and Laidly 1990; Safford et al. 1990). However, since these species were growing in a low, wet setting, it is difficult to accurately reconstruct precipitation levels in the OFW location because its vegetation would have benefitted from the elevated groundwater levels here, compared to upland areas. These mean temperatures and precipitation rates fall within current temperature and precipitation rates calculated for the Thunder Bay area, which is not surprising since the same boreal forest species are present. The current mean January temperature is around -16ºC, with mean July temperature around 17ºC - 18ºC and mean annual precipitation is around 700 mm (Kemp 1991). The main difference, however, is that Thunder Bay is considered to fall within the Great Lakes-St. Lawrence vegetation zone, with the boreal forest zone located north of the city (Figure 1.3) (Kemp 1991).

No Great Lakes-St. Lawrence forest species were identified within the OFW forest. Many Great Lakes-St. Lawrence species, including *Ulmus americanus* and *Fraxinus nigra*, *Acer saccharum,* and *Quercus macrocarpa* are thermophilious, requiring warmer temperatures and/or moister conditions (Bey 1990; Godman et al. 1990; Johnson 1990; Sander 1990; Wright and Rausher 1990). Pollen records from this area (Cummins and Oliver Ponds) show little to no pollen from these species at this time, which may suggest that they had yet to migrate north into the Thunder Bay area from Northern Minnesota or west from North-Central Ontario. This may have been because the climate in the Thunder Bay area was not warm or moist enough to support them or because they had further to migrate following deglaciation, resulting in a time lag prior to their colonization. These species are also not as tolerant of nutrient poor soils as conifer trees, which may also explain their absence at this time (Johnson et al. 1995; Kemp 1991). As some of the Great Lakes-St. Lawrence forest taxa (e.g., *Ulmus americanus* and *Fraxinus nigra*) are commonly found in wet lowland areas, their absence in the OFW forest may indicate that they were not in the Thunder Bay area; providing supporting evidence for their absence in regional pollen diagrams.

A pollen record from Lily Lake on Isle Royale indicates the appearance of maple and oak beginning around 5,400 cal yr BP (Flakne 2003), well after the growth of the OFW forest. The climate in this area then likely remained cool but was more arid than in the Early Holocene. The proximity to the cold waters of Lake Superior may have kept the Thunder Bay area cooler than areas further inland, in much the same way as is thought to have occurred with the vegetation on Isle Royale (Flakne 2003) and near Marathon (Bajc et al. 1997) around this time. Arid conditions may have persisted in the Thunder Bay area until around 7,800 cal yr BP, when *Pinus strobus* migrated into Northwestern Ontario (Björck 1985). This species generally prefers more humid and warmer conditions, needing at least ~500 mm annual precipitation and mean July temperatures between 18ºC -23ºC (Wendel and Smith 1990). Mean July temperatures at this time may not have been within the optimal range for this species, and would specifically not tolerate arid conditions.

With an increase in aridity, there is also a suspected increase in forest fires in this area (Barnes et al. 1998). However, there is little evidence in the form of charred plant remains within Unit D at OFW to suggest that forest fires occurred in the local area between ~9,100-8,900 cal yr BP. Balsam fir, tamarack and red-stemmed feathermoss are all fire-intolerant species, suggesting that forest fire frequency was low within the paleochannel or other lowland areas within the Kaministiquia River valley. The lack of evidence of fire in the OFW paleochannel is likely due to the protective effect of topography, as well as higher moisture levels in this lowland setting, making it a strictly local-scale phenomenon. Regionally, however, if arid conditions persisted, there would have been the potential for severe forest fires, provided there was fuel to burn (Barnes et al. 1998). These fires would likely have occured in the drier upland areas (Julig et al. 1990). The decline in spruce populations observed in regional pollen profiles may have occurred due to forest fires, rather than just a warming climate. With severe forest fires, much of the white spruce population may have been destroyed, as it is a fire-intolerant species (Nienstaedt and Zasada 1990). Within the boreal forest, jack pine and white birch are common pioneering species on sites which have been recently burned (Rudolf and Laidly

1990; Bergeron and Dubuc 1989), because jack pine is a fire-tolerant species and able to survive less severe fires, and white birch quickly moves onto burned sites because of low resource competition (Safford et al. 1990).

Although pollen records from this region suggest that the climate was warmer and more arid during the Hypsithermal, the OFW forest suggests that the climate in the Thunder Bay area, particulary the OFW site, may have been relatively cool and moist. These conditions could be due to the lower elevation of the area, close proximity to the water table, or due to unrecognizable factors. The cooler and moist conditions that the OFW forest presents, may have only been a local scale phenomena, and likely did not occur over a wide area since the effects of the Hypsithermal are recognized over much of Canada and the United States. However, another possible explanation is that the OFW forest represents a brief interval within the Hypsithermal where the climate was cooler and moister. Unfortunately, without more precise paleovegetation data and radiocarbon dates spanning more of the Early and Middle Holocene, this is very difficult to tell. Still, since the geomorphic and topographic evidence outlined in Chapter 2 (Deglaciation and Paleohydrology) suggests that Lakes Superior, Huron and Michigan were probably closed lakes at this time, caused in part by a dry climate (Boyd et al. 2010, Lewis et al. 2007, Croley and Lewis 2006), it is unlikely that the growth of OFW forest occurred within an extended period of increased precipitation and decreased temperature.

In summary, the vegetation growing in the OFW study area ~8,900 cal yr BP can be described as a conifer dominated boreal forest stand. The growth of the OFW forest borders a time of climatic change from glacial to Hypsithermal conditions in an area where dramatic fluctuations in water levels within the Lake Superior basin occurred.

Based on sediment from the OFW paleochannel, as well as radiocarbon dated macrobotanical remains from the OFW forest, it is possible to reconstruct the paleoenvironment for Lower Kaministiquia River valley during the Early and Middle Holocene.

7.2. PALEOENVIRONMENTAL RECONSTRUCTION OF THE LOWER KAMINISTIQUIA RIVER VALLEY DURING THE EARLY AND MIDDLE HOLOCENE

7.2.1. Regional Paleoenvironmental Reconstruction: Lower Kaministiquia River Valley and the Lake Superior basin

Following the retreat of the LIS from the Thunder Bay area after 11,500 cal yr BP, the Lower Kaministiquia River valley was inundated by waters from glacial Lake Minong (Boyd et al. 2010). The high Lake Minong water levels persisted in the basin, in part due to the presence of the Nadoway sill, which blocked drainage through the St. Mary's River into the Lake Huron basin. The highest Lake Minong water level was approximately 40 m higher than the modern water level of Lake Superior (Booth et al. 2002), with a shoreline still visible between 240 and 225 m asl (Julig et al. 1990). This shoreline has been identified at the Cummins site, where it is approximately 11 km inland from the present shoreline (Julig et al. 1990), and near Rosslyn Village, where it is approximately 17 km west from the present shoreline (Figure 7.3) (Zoltai 1963). During this high stand, the ancestral Kaministiquia River was depositing a large delta within the Lower Kaministiquia River valley, as the river carried meltwater and sediment from glacial Lake Kaministiquia and possibly Lake O'Connor into the Lake Superior basin (Zoltai 1963).

At least part of the subaqueous fan sediments from the Boyd cutbank were likely deposited during this event, forming the ~13 m crossbedded sands of Unit F (Boyd et al. 2010; Loope 2006). During the Minong phase, much of the Lower Kaministiquia River valley was inundated by water, and the OFW site was submerged.

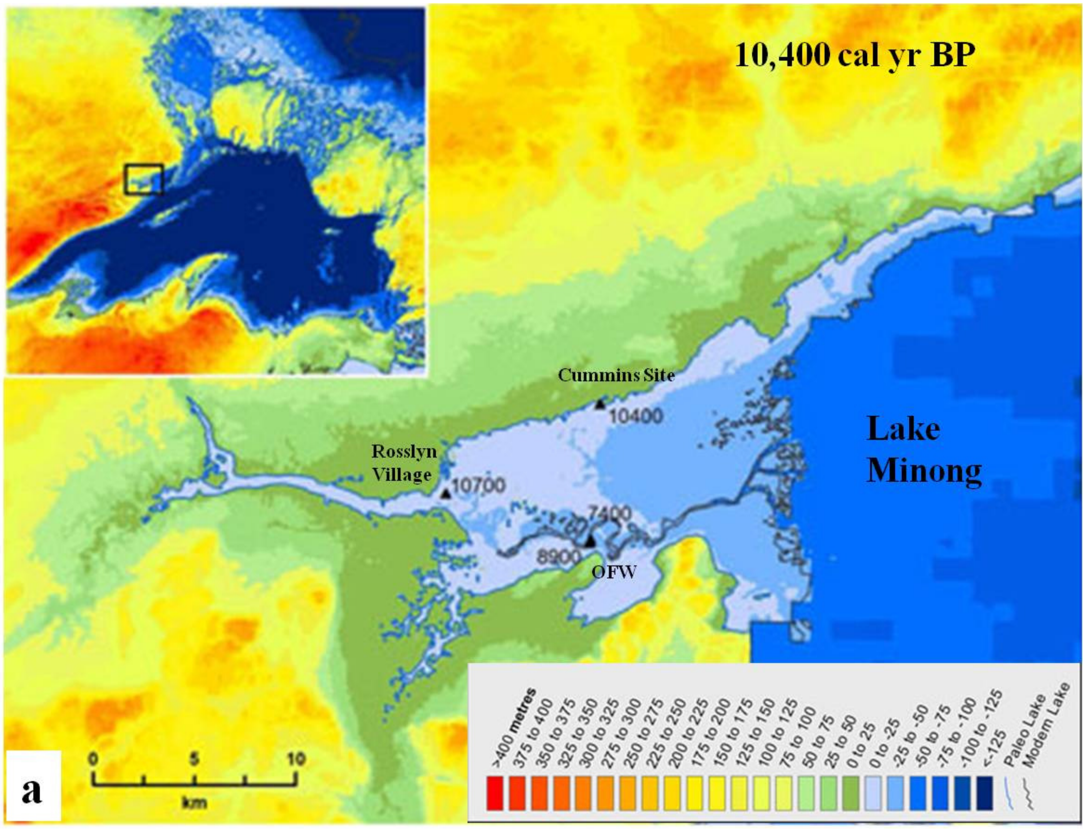


Figure 7.3. Paleotopographic map of the Thunder Bay area ~10,400 cal yr BP showing the extent of the glacial Lake Minong water level within the Lower Kaministiquia River valley. The oldest date for a Minong beach, ~10,700 cal yr BP, has come from a gravel pit near Rosslyn Village. Notice that the OFW site is 20 to 50 metres below lake level (from Boyd et al. 2010).

After 10,700 cal yr BP, Lake Minong water levels began to decline, resulting in the Houghton Low phase within the Lake Superior basin (Farrand 1960; Farrand and Drexler 1985). The lowering of water levels has been attributed to the rapid erosion of the Nadoway sill, possibly by a glacial meltwater influx from glacial Lake Agassiz (Farrand and Drexler 1985; Lewis and Anderson 1989; Yu et al. 2010). Once this sill was breached, it is believed water levels within the basin dropped ˜45 m (Yu et al. 2010).

Declining water levels within the basin are first indicated at the OFW and Boyd cutbanks by the incision of the paleochannels into the subaqueous fan sediment of Unit F. Provided that Yu et al., (2010) are correct in inferring that the sill was breached around 9,300 cal yr BP, the incision of the OFW channel into the fan sediments may have occurred between 9,300 and ~9,100 cal yr BP. The channel may have been active prior to 9,100 cal yr BP as could be interpreted by the deposition of the sandy Unit E, below the Unit D organic layer. Boyd et al. (2010)'s reconstruction of lake levels based on the rate of isostatic rebound around the basin, has revealed that Sault Ste. Marie and Thunder Bay have similar rates of rebound which should result in similar lake level records. Since the paleochannel at the OFW cutbank is below the present water plane, the water level during the Houghton phase was likely below the Sault Ste. Marie outlet. Since the relative water level was below this outlet, Lake Superior would have been a closed basin (Figure 7.4) (Boyd et al. 2010).

Although the local paleohydrological history is recorded in the cutbank sediments, it is unclear whether the paleochannel was abandoned between 9,100-8,900 cal yr BP, when Unit D was deposited, due to drier climatic conditions or if the ancestral Kaministiquia River simply switched to a different channel after this time. When water levels once again rose shortly after 8,900 cal yr BP, the OFW site was inundated and the paleochannel filled in with varved lacustrine sediment represented by Unit C. However, sometime after 7,400 cal yr BP, the Kaministiquia River downcut again within the valley, due to water levels once again declining and the Thunder Bay area possibly rebounding compared to the Sault Ste. Marie outlet (Boyd et al. 2010).

The declining waters of Lake Minong would have gradually exposed more of the Lower Kaministiquia River valley until the end of the Houghton Low ~8,900 cal yr BP. With water levels within the basin lower than present day, currently submerged landscapes in the Thunder Bay area may have been exposed, as the Houghton shoreline would most likely have been located at a lower elevation. This may have allowed for people, plants, and animals to occupy these exposed areas, including the OFW forest.

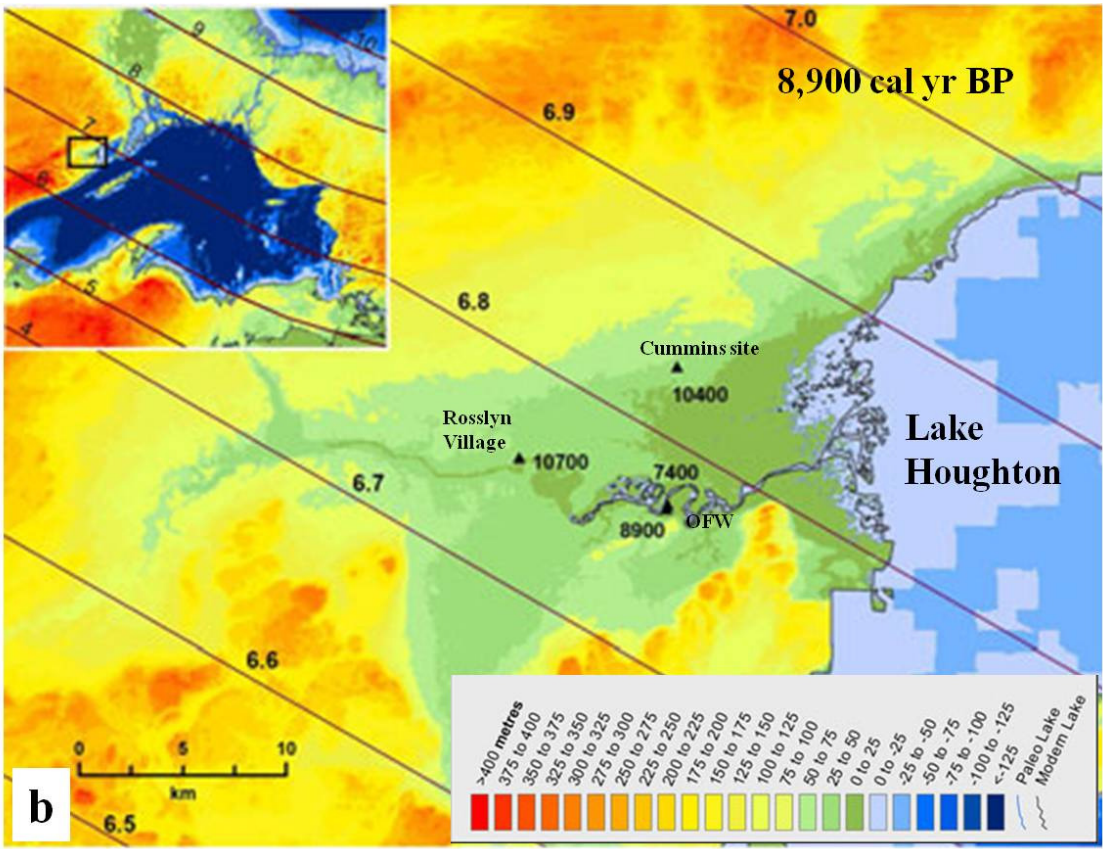


Figure 7.4. Paleotopographic map of the Thunder Bay area ~8,900 cal yr BP showing the extent of the Houghton Low phase within the Lake Superior basin. The lines represent isobases and rate of isostatic rebound. Note that much of the Thunder Bay area is above the lake level (From Boyd et al. 2010).

After 8,900 cal yr BP, part of the lower Kaministiquia River Valley was once again inundated by increasing water levels within the Lake Superior basin. This increase, at first, was rapid, as water inundated the site and the OFW forest was quickly buried (Boyd et al. 2010). Unit C of the OFW cutbank records this transgression, as many *in situ* trees and tree roots extended up from Uint D are enclosed by the varved sediment of Unit C (Boyd et al. 2010). This indicates a deep water environment with a high sedimentation rate for at least 200 years, as indicated by varve counts. The maximum elevation that this high-water event reached likely did not surpass 210 m asl, as this elevation represents the location of the later Nipissing shoreline. Therefore it can be stated that the highest water levels from this transgression were likely between 197 m asl (top of the cutbank) and 210 m asl (Nipissing shoreline). Unfortunately, there is no evidence such as a shoreline representing this pre-Nipissing water level rise. This could be because the duration of the event was not long enough and water levels not stable enough to form a shoreline, or because the rising water levels of the subsequent Nipissing transgression may have destroyed it.

Boyd et al. (2010) suggest that this pre-Nipissing water level rise may have been a result of proglacial lake runoff directed southwards from the Hudson Bay lowlands. Due to the instability of the ice front during this time, fluctuations of the ice margin could have resulted in the periodic blocking the of Kinojevis outlet, which carried meltwater to the St. Lawrence River (Prest 1970). With this outlet cut off, the next lowest available outlet(s) may have been along the north shore of Lake Superior. The discovery of the Mullet outlet north of Marathon by Slattery et al*.*, (2007) supports the hypothesis that there was a connection between Lake Superior and proglacial lakes (e.g., Lake Agassiz and Barlow-Ojibway) in the Hudson Bay lowlands, possibly resulting in meltwater entering the Lake Superior basin via the Pic River. This event may have also raised the water level within the basin back up to the Sault Ste. Marie outlet, resulting in overflow into the Lake Huron basin (Boyd et al. 2010).

As discussed in Chapter 2 (Deglaciation and Paleohydrology), after ~11,500 cal yr BP, the North Bay outlet began to gradually uplift. Thus uplift resulted in the backflooding of water in the Great Lake basins as the outlet rebounded to the Chicago and Port Huron outlet elevations (Farrand and Drexler 1985; Lewis and Anderson 1989). Within the Lake Huron basin, this transgression is initially recorded following the Late Stanley low phase at ~8,500 cal yr BP, indicating that it occurred sometime later within the Lake Superior basin, as the North Bay outlet continued to rebound to the Sault Ste. Marie outlet (Lewis et al. 2007). The peak of the transgression within the Great Lake basins is associated with the formation of the Nipissing paleoshorelines and is thought to have occurred between 6,800-5,700 cal yr BP (Lewis 1969; Booth et al. 2002), a date attributed to the formation of a wave cut bluff in the Thunder Bay area at 210 m asl (Farrand 1960).

The Nipissing transgression may or may not be recorded in the OFW cutbank sediments. Although fluctuating water levels are indicated in the interbedded fluvial and lacustrine sediments of Units A and B, the water level remained above the top of the cutbank until at least 7,400 cal yr BP (Figure 7.5). This is known because a a radiocarbon date from *Scirpus* sp., achenes recovered from Unit A indicates that sediment was still being deposited in the paleochannel at this time. Sometime later, the modern Kaministiquia River downcut until the modern channel level was reached. If the Nipissing transgression occurred prior to 7,400 cal yr BP, which is possible depending on the rate of rebound between the North Bay and Sault Ste. Marie outlets, then it could be recorded in Units B and A. Based on the lake level history of the Lake Superior basin, water levels dropped after the Nipissing transgression (~5,700 cal yr BP), with small-scale periodic rises (i.e., the Algoma and Sault transgressions) in the Late Holocene before the lake reached its modern level after 2,200 cal yr BP (see Table 7.3 for comparison of paleohydrological events between the OFW site and the Lake Superior basin) (Johnson et al. 2004).

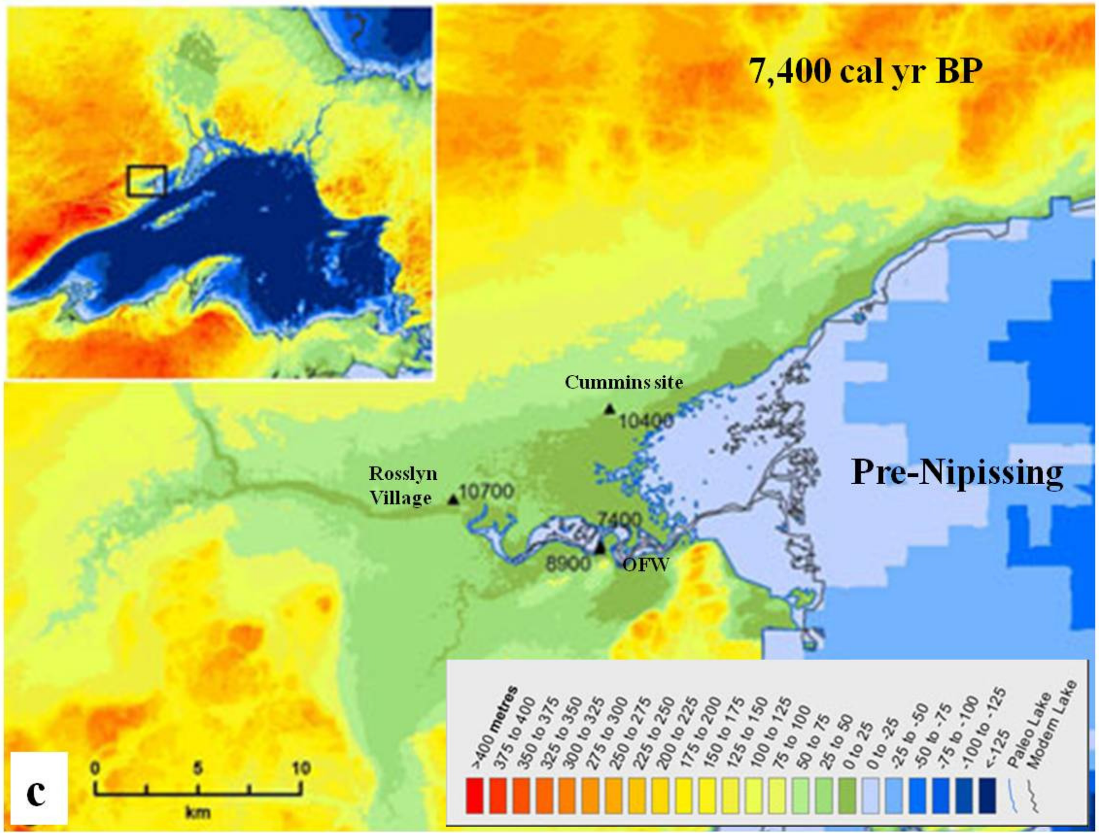
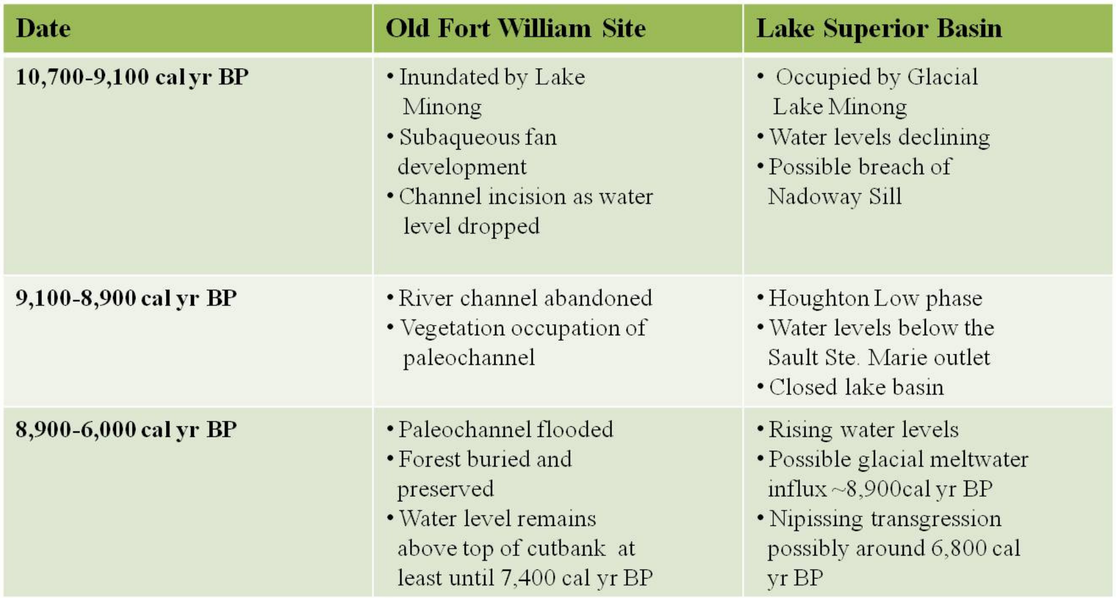


Figure 7.5. Paleotopographic map of the Thunder Bay area ~7,400 cal yr BP showing the high water levels within the Lake Superior basin following the Houghton phase. The OFW site is once again inundated (from Boyd et al. 2010.)

Table 7.3. Paleohydrological events recorded from the OFW cutbank and corresponding events within the Lake Superior basin during the Early and Middle Holocene. List of events is from oldest to youngest. Bullet points in the same timeframe are sequential events.



Date	Old Fort William Site	Lake Superior Basin
10,700-9,100 cal yr BP	• Inundated by Lake Minong • Subaqueous fan development • Channel incision as water level dropped	• Occupied by Glacial Lake Minong • Water levels declining • Possible breach of Nadoway Sill
9,100-8,900 cal yr BP	• River channel abandoned • Vegetation occupation of paleochannel	• Houghton Low phase • Water levels below the Sault Ste. Marie outlet • Closed lake basin
8,900-6,000 cal yr BP	• Paleochannel flooded • Forest buried and preserved • Water level remains above top of cutbank at least until 7,400 cal yr BP	• Rising water levels • Possible glacial meltwater influx ~8,900cal yr BP • Nipissing transgression possibly around 6,800 cal yr BP

7.3. ARCHAEOLOGICAL IMPLICATIONS OF THE PALEOENVIRONMENTAL RECORD

7.3.1. Archaeological Record of the Lower Kaministiquia River Valley

The paleoenvironmental reconstruction of lake levels, vegetation, and climate of the Thunder Bay area may also be utilized to better understand this are’s human history. Since humans relied on their environment for resources (e.g., food, shelter, tools, transportation), any significant changes in the environment may have altered settlement distribution and availability of resources. It is important to first understand what the environment was like to grasp how humans may have used it.

As mentioned in Chapter 4 (Early and Middle Holocene Culture History of Northwestern Ontario), the first evidence of human occupation of the Thunder Bay area is associated with Plano cultures dating ~10,700 cal yr BP (Julig et al. 1990). Hamilton's (1995) archaeological survey of cultivated fields located in the Lower Kaministiquia River valley near Stanley and Rosslyn Village revealed the presence of Paleoindian and possibly Archaic peoples based on artifacts recovered (e.g., parallel flaking on tools). Sites identified within this survey were located between 213 and 259 m asl, all at higher elevations than the Nipissing shoreline, which is located at roughly 210 m asl (Figure 7.6). Although the context of some of these sites is questionable (e.g., Pawlick Field sites [DcJj-24, 25, 26, 27]) the widespread distribution of artifacts in the survey area may indicate that people utilized a large portion of the Lower Kaministiquia River valley in the Early and possibly Middle Holocene. This is important because it places people in the OFW area prior to the decline in the Lake Minong water level.

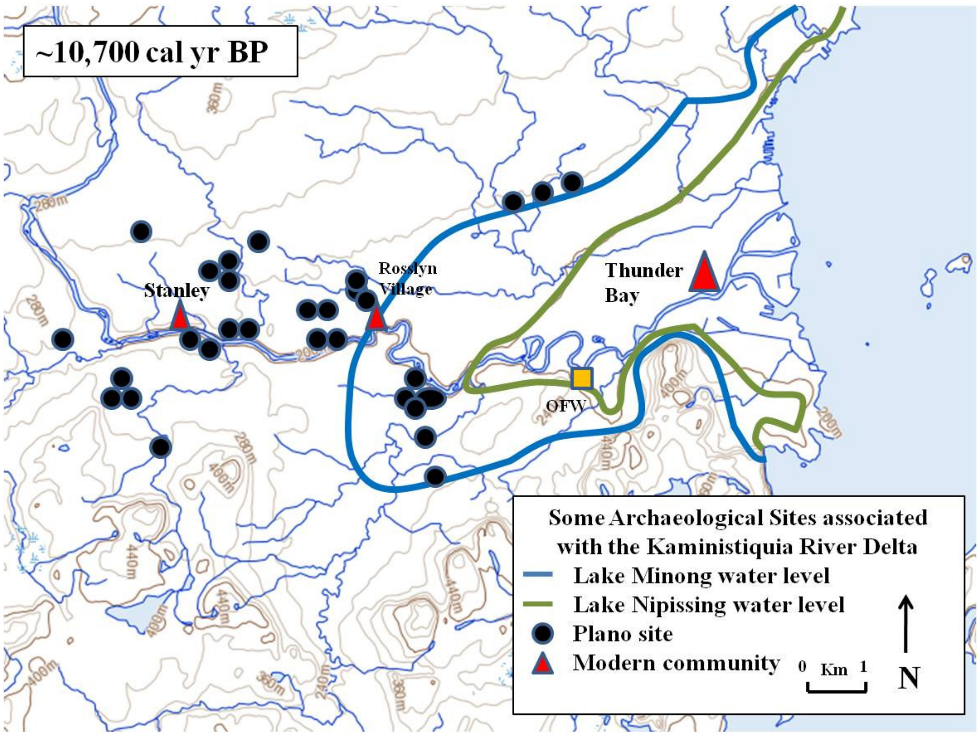


Figure 7.6. Distribution of Plano sites on the Kaministiquia River delta. The blue line represents the approximate location of the Lake Minong shoreline ~10,700 cal yr BP, with the green line representing the Nipissing shoreline ~6,500 cal yr BP. Paleoindian and/or Archaic sites may be deeply buried in the area outlined by the Nipissing shoreline as well as under the present water level of Lake Superior (after Hamilton 1995).

7.3.2. Archaeological Sites Buried by Post-Minong Water Level Fluctuations within the Lake Superior Basin

Because Early and Middle Holocene human occupation of the Lower Kaministiquia River valley has been established, it is possible that people living in these timeframes also utilized the lowland areas below 259 m asl, including the OFW paleochannel, when water levels within the Lake Superior basin declined following the Minong high phase. Since Plano sites have been identified below the Minong shoreline in the Kaministiquia River valley, it is possible that a large area extending from the Minong shoreline to currently submerged areas within the lake basin may have archaeological sites that have yet to be found.

Many Paleoindian sites in the Thunder Bay area are located on or near the Lake Minong shoreline (e.g., Cummins, Simmonds, Biloski) (Hinshelwood 2004). The Nipissing shoreline, which is visible throughout the city of Thunder Bay, was also occupied by Archaic peoples (Phillips 1993). Since these shorelines have proven to be archaeologically productive areas, it can be argued that a Houghton phase shoreline may have also been occupied by Paleoindian and/or Archaic peoples. Because evidence suggests that water levels during the Houghton phase were below the St. Mary's River outlet, at least between 9,100 and 8,900 cal yr BP, the shoreline would likely be located below Lake Superior's present water level of 183 m asl. Both Farrand and Drexler (1985) and Yu et al., (2010) suggest that water levels fell approximately 45 m between the Minong and Houghton phases, suggesting the Houghton shoreline may be around five m below (~178 m asl) the present water level. Unfortunately, no Houghton shoreline has been identified, and it is unknown if water levels during this lake phase were stable enough to form one. Rising water levels within the basin around 8,900 cal yr BP may have eroded and destroyed any evidence of a Houghton shoreline, and therefore no evidence of it may exist.

As of yet, no offshore archaeological sites have been identified in the Thunder Bay area. Wright (1963), however, notes that some Archaic artifacts of suspected Pre-Nipissing date have been picked up by fish nets near Thunder Bay and along the north shore of the lake. Unfortunately, a description of these artifacts is not given nor is the context they were found in known (i.e., location, water depth). It is also not widely accepted what Archaic artifacts can be designated "Pre-Nipissing" as a poor understanding of the Archaic timeframe in general (Hamiltin 2007) means that typologies have yet to be assigned to Archaic materials (Hinshelwood 2004).

Archaeological survey of offshore areas within the Great Lakes is in its infancy. However, O'Shea and Meadows (2009) show that submerged paleolandscapes may be discernable using sonar equipment and underwater remote operated vehicles. They discovered what they believe to be caribou hunting structures and camp sites on a landbridge in Lake Huron that was exposed during a lake level low stand between 11,900-7,800 cal (10,200-7,000 ^{14}C) yr BP, thereby dating these structures to the Paleoindian/Archaic timeframe. Although the archaeological and paleohydrological histories are different between the Great Lakes' basins, the same methods can still be applied to look for submerged sites within Lake Superior.

Archaeological material including copper artifacts and faunal remains have also been recovered in the Thunder Bay area and along the north shore of the basin buried under many meters of lacustrine sediment (Dawson 1972, 1973; Wright 1963, 1995). Along the north shore a hearth feature and a copper gaff from the Armstrong site were located approximately 30 m inland from the modern lake level and under roughly 7.5 m of “stratified clays and silts” These sediments are thought to represent the Nipissing transgression (Wright 1963, 1995). Although there is debate about when this transgression began, Booth et al., (2002) suggests that it could have been around 6,800 cal yr BP, meaning the gaff and hearth would pre-date this time.

In the Lower Kaministiquia River valley, artifacts from the Turning Basin site (DcJh-5), in the West Fort area of the Kam delta, include mammal bones and a copper spearhead buried beneath 12 m of crossbedded sands identified as lacustrine sediment (Tanton 1931); copper chisels, a copper adze and a flexed human burial approximately 1.5 m below the surface; and copper artifacts including fish hooks with cedar cribbing approximately 9 m below the surface (Tanton 1931). The mammal bone and socketed spear point were located on top of a clay surface and were overlain by 12 m of sand (Tanton 1931). Tanton describes the sedimentary sequence in the area as beginning with bedrock approximately 30 m below the surface, followed by 1-2 m of “conglomerate”, which is then overlain by 18-27 m of stratified blue clay, and 1-12 m of crossbedded sands (Figure 7.7). In some places, the clay has been eroded, forming channels that were filled by the crossbedded sands (Tanton 1931). Most of the artifacts from the Turning Basin site were recovered below the crossbedded sand and on top of the blue clay. Tanton (1931) reports that the lacustrine sands above the artifacts was a mix of sand and gravel that showed evidence of current action. He used this geomorphic context to suggest that the archaeological and faunal material were deposited during a period of relatively low water levels within the Lake Superior basin. His interpretation fits with the theory that the archaeological material was deposited prior to the Nipissing transgression, as it was preceded by low lake levels of the Houghton phase (Dawson 1972). However, the lack of chronological control for this sequence means that this interpretation is speculative.

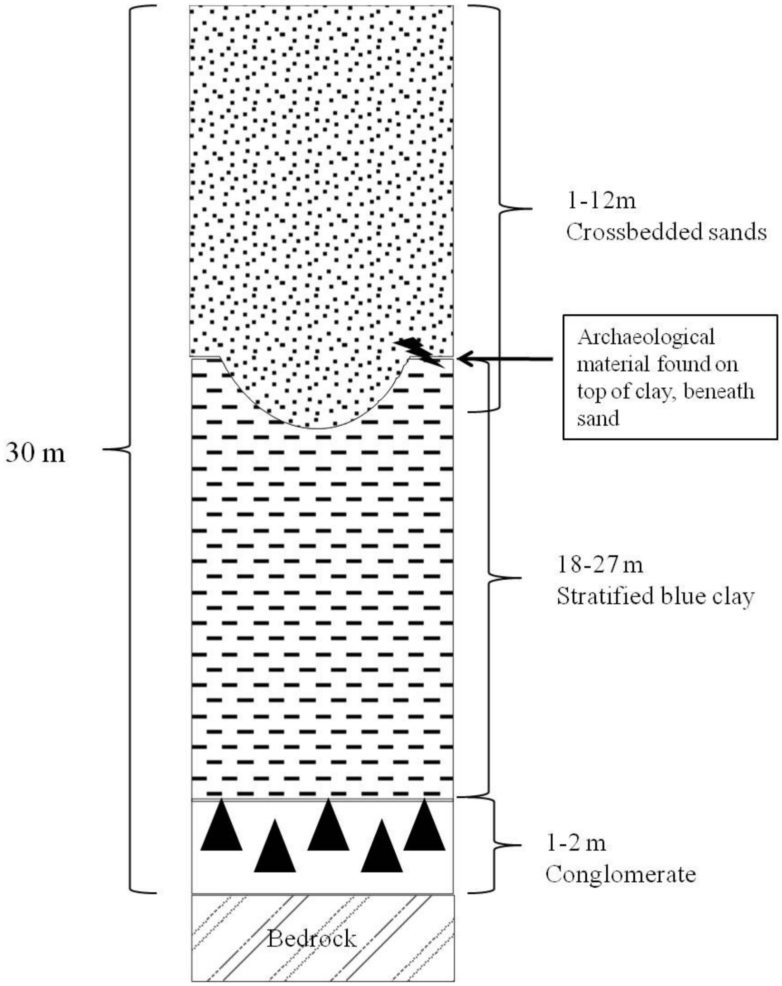


Figure 7.7. Diagram showing sediment sequence in the West Fort area of Thunder Bay near the Turning Basin site based on description in Tanton (1931). At the base is bedrock, which is overlain by conglomerate. On top of the conglomerate is stratified blue clay which has been incised by channels and filled with crossbedded sand. A copper spearpoint, mammal bone, copper fish hooks and cedar cribbing were reported from the contact between the clay and sand.

Based on the deglacial and paleohydrological histories of the Thunder Bay region, the startigraphy from the Turning Basin site could be interpreted in multiple ways. It is possible that the conglomerate was deposited by the LIS after the Marquette advance or, less likely, was deposited by alluvial action by the ancestral Kaministiquia River. The stratified clay unit indicates a deep water environment and could represent sediments deposited by glacial Lake Minong, which occupied the Lake Superior basin after the LIS retreated, or could represent both Lake Minong and Lake Nipissing sediments. The crossbedded sands could represent Lake Nipissing nearshore sediments but could also have been deposited alluvially as the Kam delta prograded into the lake; these sediments may not indicate a lacustrine setting as suggested by Tanton (1931).

If the stratified blue clay represents the Nipissing phase or both the Lake Minong and Nipissing phases, with the overlying crossbedded sands being post-Nipissing in nature, the archaeological remains would then have had to been deposited during a low phase that occurred after the Nipissing transgression. Based on the current knowledge of the Nipissing transgression, water levels likely began to rise in the basin around 6,800 cal yr BP (Booth et al 2002), and did not decline until around 4,500 cal yr BP (Johnson et al, 2004). The artifacts would then have to post-date 4,500 cal yr BP. The 12 m of crossbedded sands overlying the archaeological artifacts would then have been either deposited by the modern Kaministiquia River or from the Algoma and Sault phases within the Lake Superior basin. These transgressions, however, were minor compared to the Lake Minong and Nipissing water levels and would therefore be unlikely to deposit ~12 m of crossbedded sands.

If the clay represents Lake Minong sediments and the crossbedded sands Lake Nipissing sediments, then the archaeological artifacts may be older than previously thought. These artifacts would have been buried when water levels rose in the basin around 8,900 cal yr BP and could possibly date to at least this time. This could imply that copper tools were being manufactured much earlier than previously assumed, given that the earliest dated evidence of copper use has come from copper spearhead hafted to a wooden shaft recovered from South Fowl Lake, Minnesota and dated to 7,600 cal (6,700 ^{14}C) yr BP (Beukens et al. 1992). It is, however, impossible to date the remains from the

Turning Basin site or compare the copper artifacts to others found around the Lake Superior basin, because the artifacts have since been lost (Dawson 1972). If the artifacts date to the post-Nipissing timeframe then they are much more recent. However, no matter the date, the location and context of these artifacts attest to the presence of deeply buried archaeological materials in the Thunder Bay area.

It is important to note the presence, depth and location of these buried materials. With fluctuating water levels within the basin during the Early and Middle Holocene, the potential for many buried archaeological sites is great. At least between 9,100 and 8,900 cal yr BP, declining water levels within the Lake Superior basin exposed previously sub-areal surfaces that may have been inhabited by vegetation, animals, and humans. Locating these sites may have a significant impact on the archaeological record for this area since sites dating to the time period bridging the Early and Middle Holocene are scarce.

Fluctuating water levels within the Early and Middle Holocene may explain part of the low density of archaeological sites that date to between 10,500-6,800 cal yr BP in the Thunder Bay area. Since paleoshorelines appear to have been frequented by Paleoindian and Archaic peoples, it is likely that some shoreline sites may not have been used repeatedly or may have been disturbed, destroyed or buried by later fluctuations. The may also be said for river sites. River systems are sensitive to fluctuations in water levels of the lakes they drain into, and respond by downcutting when water levels decline, or bury and erode older surfaces or sediments, such as occurred at the OFW site. These processes could inpact human occupation of these areas as well as site preservation. For example, the OFW site appears to have been a biologically productive area and at least for some time was probably a wetland and although there is no evience indicating human occupation at the site, this type of environment may have attracted various animals and people. It is within this lowland setting where archaeological materials may be found. Unfortunately, locating these areas is difficult because of their deeply buried nature.

Areas such as the OFW site are unique due to the high level of preservation. Finding archaeological sites in this setting would be optimal because faunal and other biological remains may be preserved and even dated which would provide some dating control to archaeological remains which does not yet exist. Although it is difficult to find buried and inundated sites, understanding the paleoenvironment may contribute to the archaeological record even though archaeological material may not be found.

CHAPTER 8

CONCLUSIONS

The sediment from the OFW cutbank reveals a sequence of paleoenvironmental events that occurred within the Lake Superior basin during the Early and Middle Holocene. Many lake level fluctuations occurred throughout this time, most importantly a drop in water level to or below the elevation of the St. Mary's River outlet between at least 9,100 and 8,900 cal yr BP. This drop in water level, associated with the Houghton phase, may have been due to the erosion of the Nadoway sill, as suggested by Yu et al. 2010. During the Houghton phase, vegetation was able to occupy the OFW paleochannel for at least 200 years from 9,100-8,900 cal yr BP before water levels rapidly rose and buried the vegetation.

The OFW forest represents a remarkably well-preserved "snapshot" of a pioneering community of plants during what may have been the most arid period of the Holocene in the Upper Great Lakes. Macrobotanical remains from the OFW site represent mesic, fire-intolerant and late succession species, indicating a mature, closed or semi-open conifer or mixed conifer-hardwood forest. These results suggest that a boreal forest community dominated by spruce and fir was present in the Thunder Bay area by at least 9,100 cal yr BP. The OFW vegetation data probably indicate that regional pollen records feature an over-representation of upland vegetation, including pine and birch, and an under-representation of lowland vegetation. For example, species including tamarack and balsam fir were well represented in the macrobotanical remains recovered from the study site, despite their near absence in the regional pollen record.

Trends that were visible in the macrobotanical assemblage reveal that small-scale spatial differences in vegetation occurred within the OFW paleochannel. Wetland taxa were predominant in the paleochannel bottom and species that prefer dry areas were most commonly recovered from sample locations on the paleochannel slope. This likely reflects either a changing moisture gradient within the paleochannel itself, or it may also indicate changing moisture conditions within the Lower Kaministiquia River valley between 9,100-8,900 cal yr BP.

The paleovegetation reconstruction reveals that the Lower Kaministiquia River valley was generally moist around 8,900 cal yr BP, probably due to the low topographic setting and the proximity of the water table. Lack of evidence of forest fires indicates that fire-related disturbance was minimal, at least between 9,100 and 8,900 cal yr BP. The Lower Kaministiquia River valley may not have been as susceptible to Hypsithermal conditions, at least between 9,100-8,900 cal yr BP due to its protected lowland setting. The OFW macrobotanical remains and regional pollen records suggest that a variety of habitats may have existed in the Thunder Bay area during the early Middle Holocene, with dry upland areas dominated by pine and birch and moist lowlands dominated by spruce and balsam fir. It would be beneficial if future paleoenvironmental studies focused on identifying the possible effects the Hypsithermal had throughout Northwestern Ontario, specifically in relation to past water levels and vegetation.

Fluctuating water levels during the Early and Middle Holocene may have also affected human settlement patterns in the Lower Kaministiquia River valley. During the Houghton phase, water levels were low, exposing parts of the Lower Kaministiquia River valley that were previously inundated by glacial Lake Minong. Areas that are currently submerged by Lake Superior may also have been exposed as water levels appear to have been below the present water level, at least between 9,100-8,900 cal yr BP. However, evidence supporting the occupation and/or utilization of these exposed areas is limited, apart from the Turning Basin site and scattered buried archaeological material recovered along the north shore of the lake. This could be, in part, due to lake level fluctuations which could have buried or destroyed archaeological sites dating to at least the early Middle Holocene. These water level fluctuations and the erosion and/or deep burial of former terrestrial surfaces they might have caused may, in part, be responsible for low numbers of early Archaic archaeological sites that have been reported in the Thunder Bay area

The potential for deeply buried archaeological sites is important when considering the archaeological record for this area. The Turning Basin site and the discovery of the OFW forest indicate the presence of deeply buried archaeological material and paleo-landscapes at least 10 – 12 m below the ground surface in the Lower Kaministiquia River valley. Future archaeological work in the Lower Kaministiquia River valley should recognize that material dating to at least the early Middle Holocene is likely to be deeply buried in lowland settings. Specifically, elevations below 210 m asl should be a focus when looking for buried archaeological material dating between 9,500 and 6,800 cal yr BP. This elevation represents the Nipissing shoreline, and it is believed that water levels did not rise above this elevation after approximately 6,800 cal yr BP.

Accessing the buried material, however, would be difficult. Many of the artifacts from the Turning Basin site were recovered mainly due to sewer construction which required significant digging. The OFW site is unique because the modern river channel cut into the older sediment, revealing the buried forest. Given that the macrobotanical remains from the OFW site were exceptionally well-preserved and faunal remains and copper were recovered from the Turning Basin site, it is possible that other buried archaeological remains may be in the same condition. Organic remains like bone, wood and antler do not generally preserve in ancient archaeological sites in this area's podzolic soil conditions, but they may be found well-preserved in these buried contexts. Thus, although deeply buried archaeological sites may be difficult to find and excavate in the Thunder Bay area, they may yield information unattainable from sites dating to the same time period in upland settings. The information obtained on biological remains dating to the early Middle Archaic may add significant data to the archaeological record for this area in regards to technology and subsistence patterns.

This thesis highlights the importance of macrobotanical remains in the reconstruction of the paleoenvironment. The macrobotanical remains recovered from the OFW cutbank provide a local scale reconstruction of vegetation that is generally unattainable by other paleovegetation analyses. However, when used in conjunction with pollen and sediment analysis, it is possible to use the local scale data and apply it to a broader scale. This enables a far more accurate paleoenvironmental reconstruction. This thesis also stresses the importance of knowing what the paleoenvironment was like when examining the archaeological record. Rebuilding the paleoenvironment at both the local and regional scale, allows for a better understanding of human settlement patterns. This better quality information can then be used to help predict where other sites of similar age might be found.

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Appendices

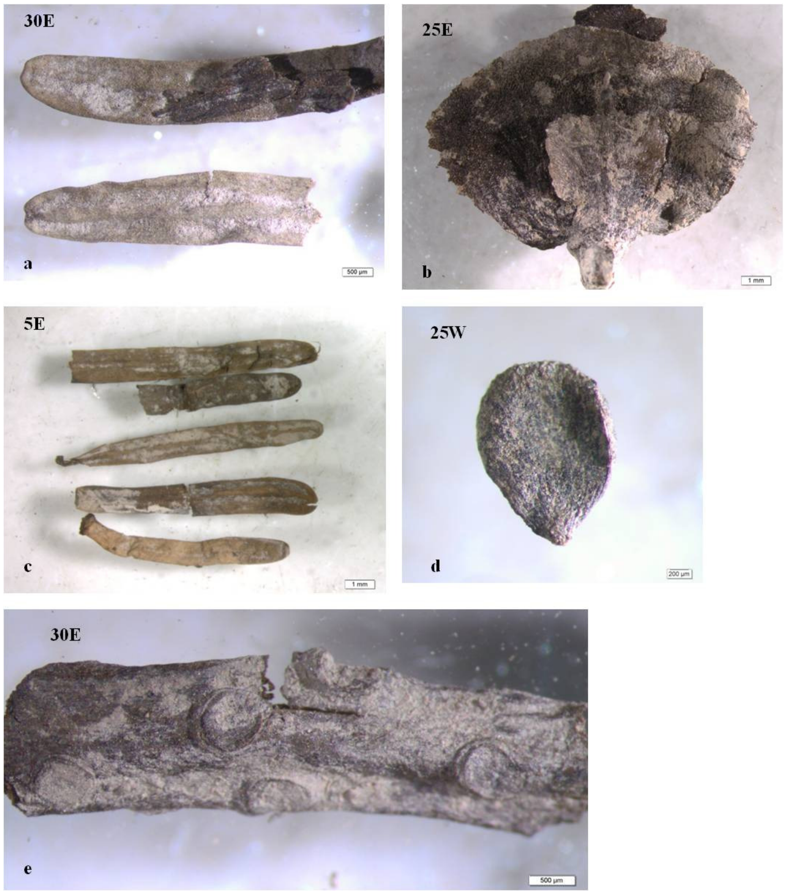
Appendix A

Visual Macrofossil Guide

The following pictures are macrofossils identified from Unit D of the OFW cutbank. Each photograph includes the sample location where the macrofossil was located and size of the macrofossil. The caption describes what type of macrofossil is depicted and a description from cited literature on morphological characteristics that were used to identify it.

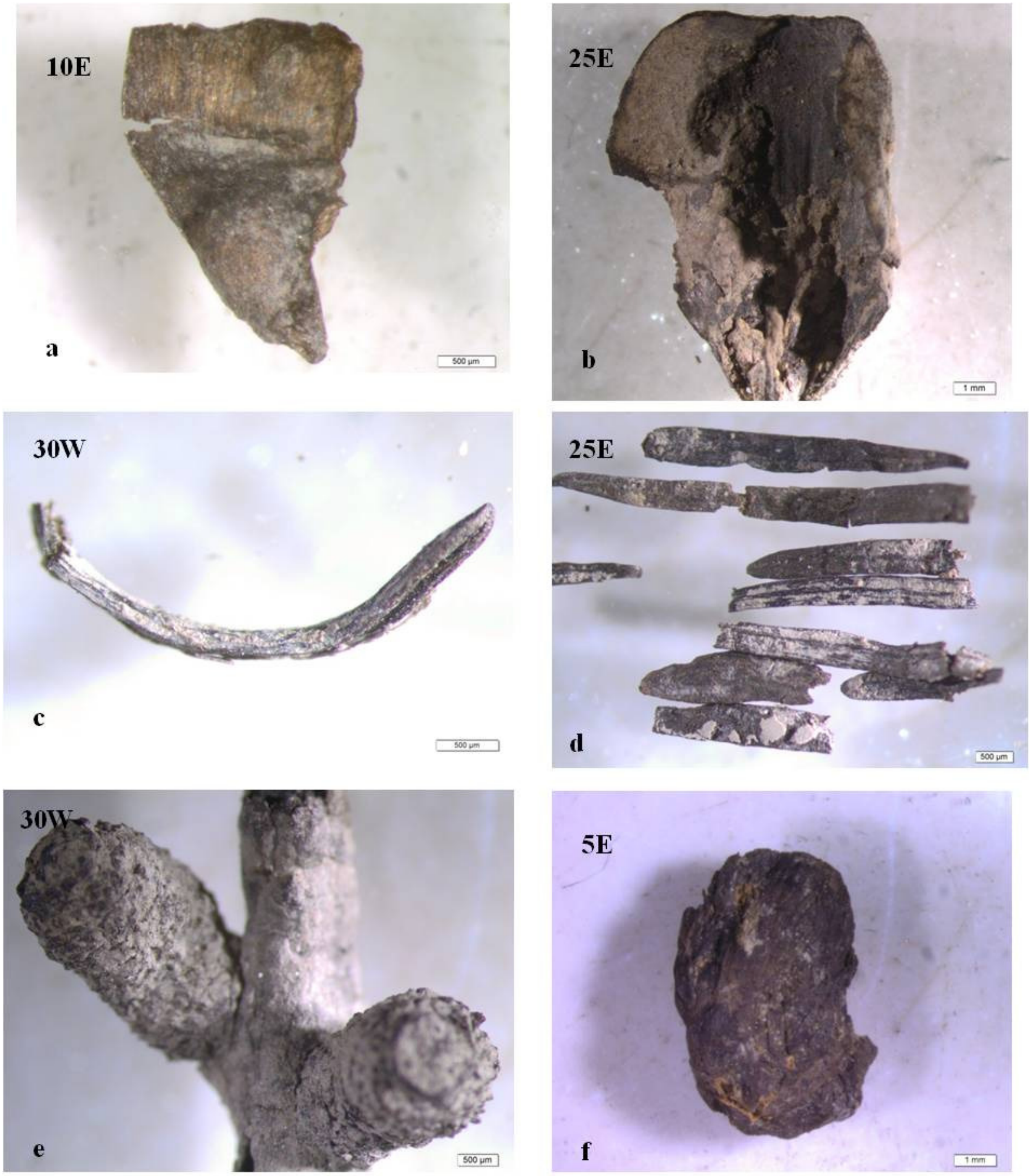
<u>Trees and Shrubs</u>

Abies balsamea



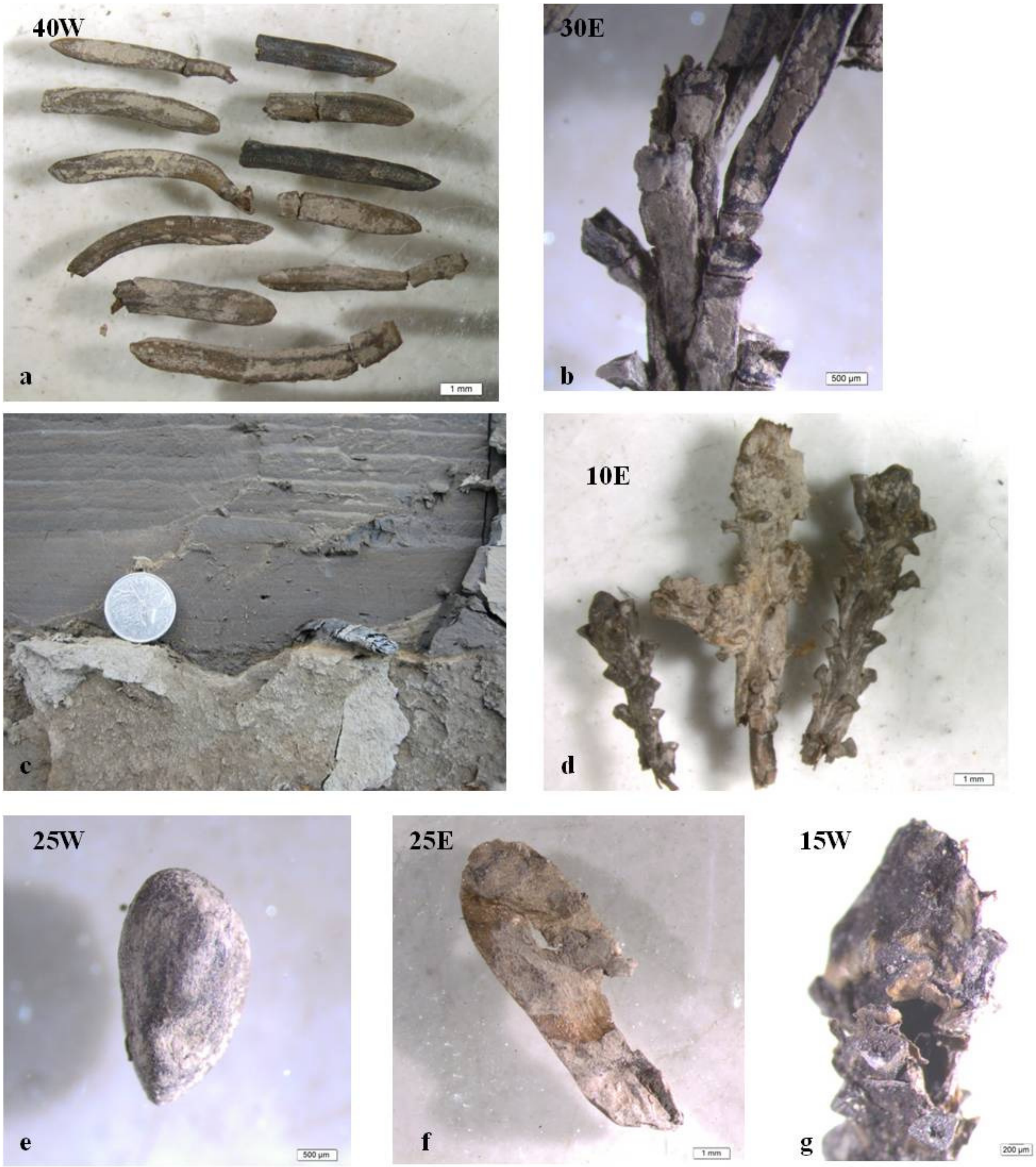
Macrobotanical remains from *Abies balsamea* (balsam fir). Photos ‘a’ and ‘c’ = leaves, ‘b’ = cone scale with bract, ‘d’ = possible seed, and ‘e’ = twig.

Larix laricina



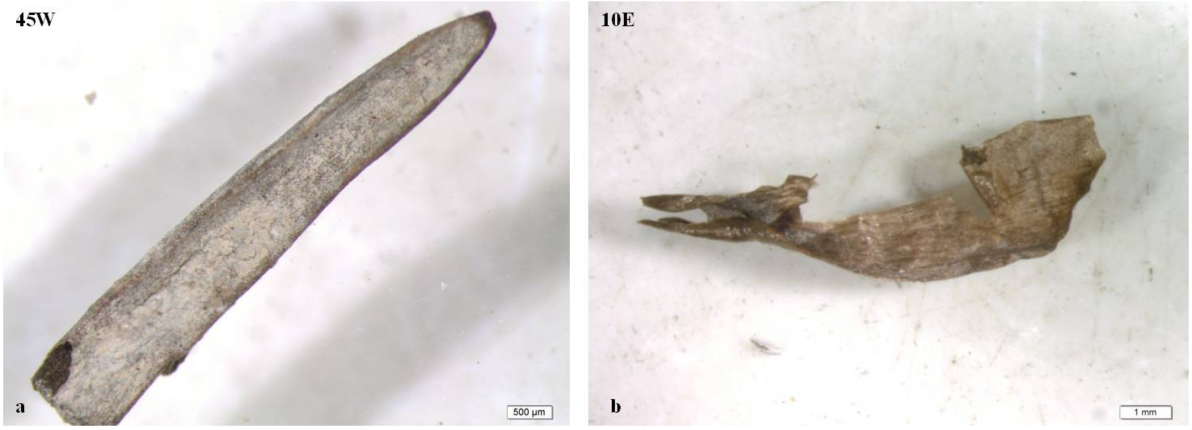
Macrobotanical remains from *Larix laricina* (tamarack, larch). Photo ‘a’ = seed with wing, ‘b’ = cone scale, ‘c’ and ‘d’ = leaves (leaves), ‘e’ = twig with new shoot growth and ‘f’ = cone.

Picea



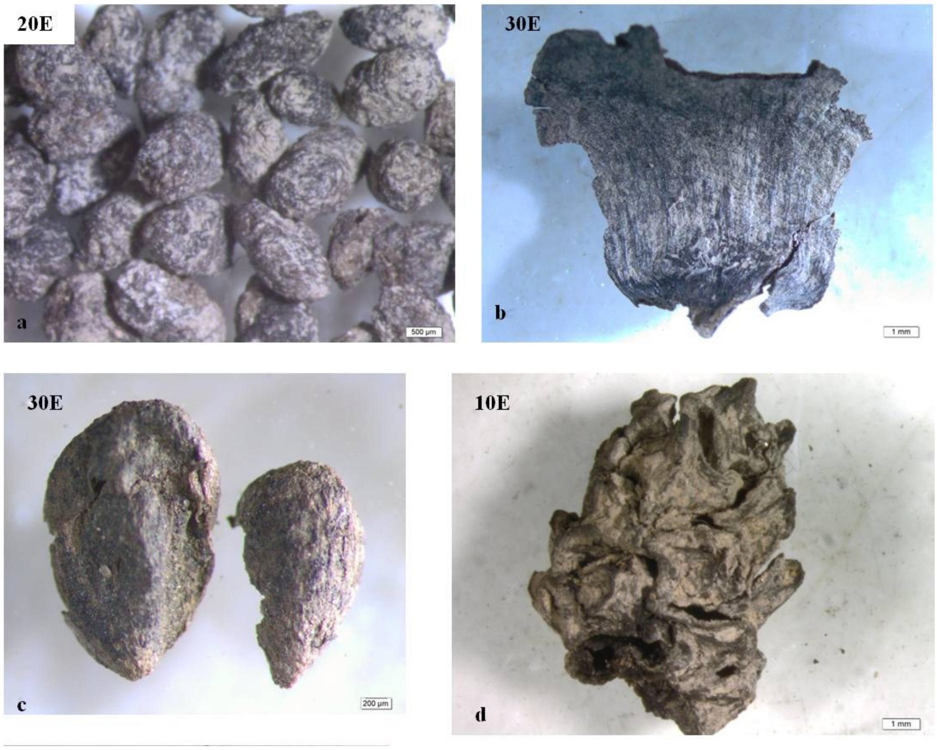
Macrobotanical remains from *Picea*. Photo 'a' = leaves, 'b' = twig with attached leaves, 'c' = *Picea glauca* (white spruce) cone, 'd' = twigs with terminal buds, 'e' = seed, 'f' = seed wing from *Picea glauca*, and 'g' = terminal bud.

Pinus



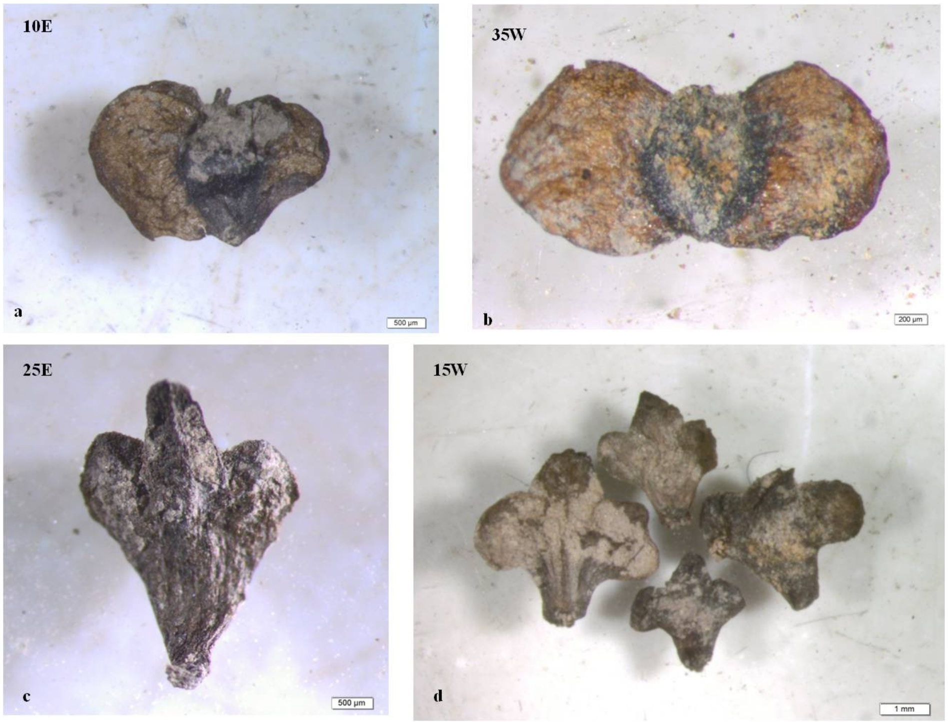
Macrobotanical remains from *Pinus*. Photos ‘a’ and ‘b’ = leaves, and ‘c’ = seed wing.

Conifer macrofossils not identified to species



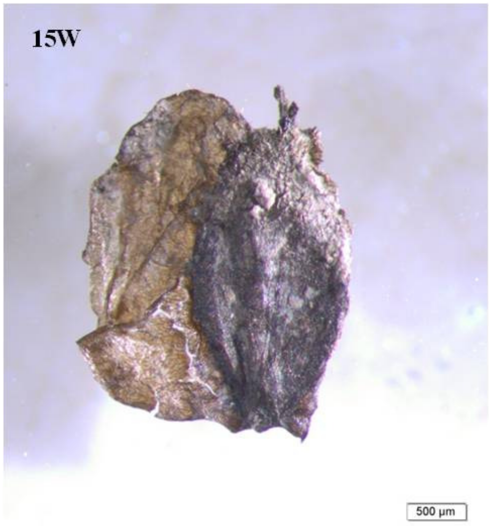
Various conifer remains which have not been identified to species. Photo 'a' = new shoots, 'b '= cone scale, 'c' = seeds and 'd' = conelet.

Betula papyrifera



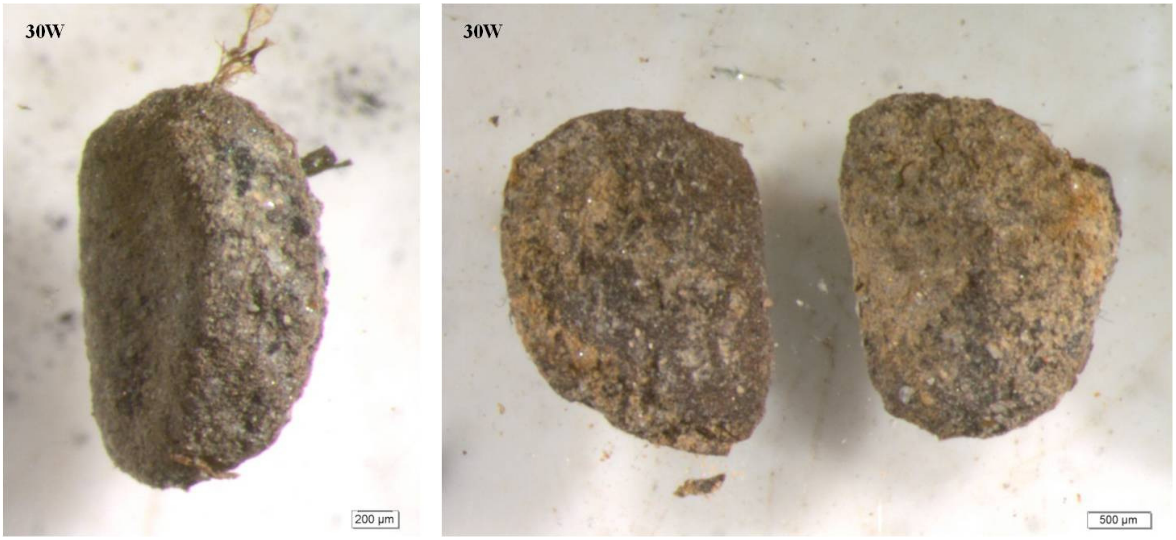
Macrobotanical remains from *Betula papyrifera* (white birch). Photo 'a' and 'b' = samaras (nutlet and seed wing), and 'c' and 'd' = fruit scales.

Alnus viridis



A samaras from *Alnus viridis (crispa)*.

Arctostaphylos uva-ursi



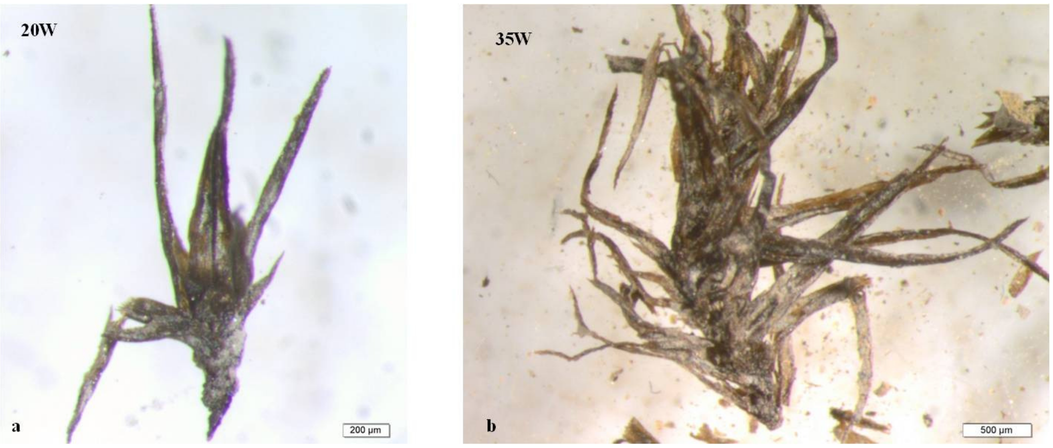
Stones from *Arctostaphylos uva-ursi* (bearberry).

Rubus idaeus

Stones from *Rubus idaeus* (wild red raspberry)

Bryophytes

Dicranum polysetum



Leaves and stem from *Dicranum polysetum* (wavy moss).

Pleurozium schreberi

Leaves and stem from *Pleuroziumm schreberi* (red-stemmed feather moss).

Terrestrial Herbs

Cf. Poaceae

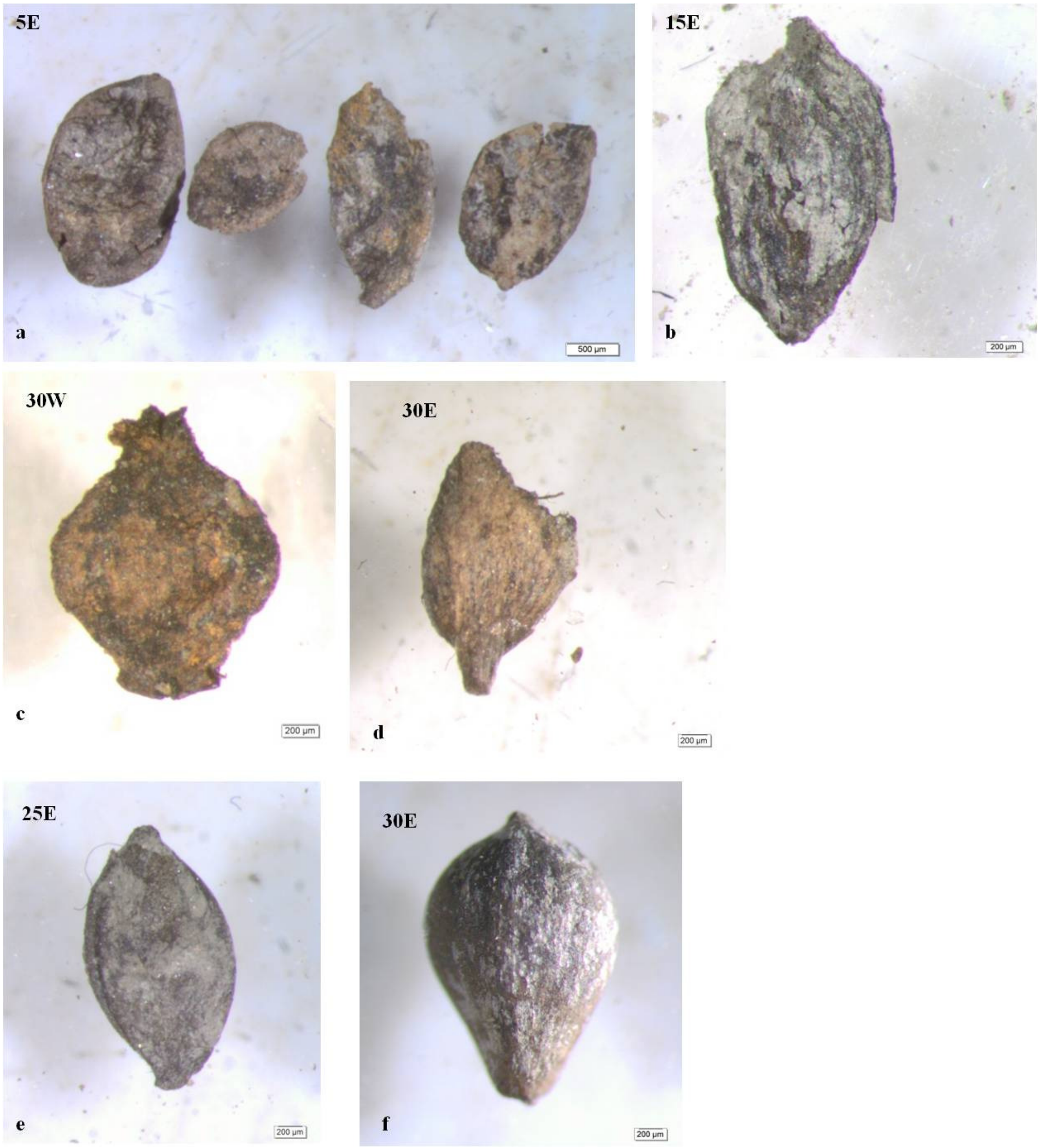
Possible seeds from the Poaceae family.

Lycopus americanus

Seed from *Lycopus americanus* (American water horehound).

Aquatic Herbs

Cyperaceae family

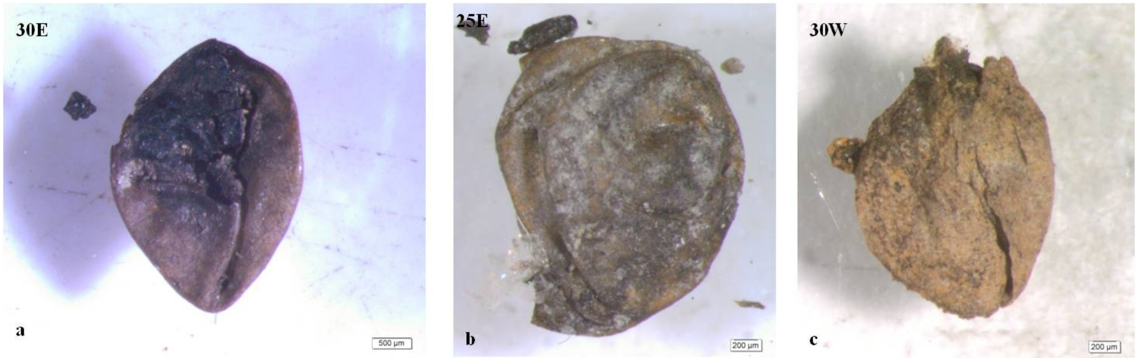


Macroboanical remains from the Cyperaceae family. Photo 'a' = *Carex* spp., (sedge) achenes, 'b' = unidentified *Carex* sp., perigynim, 'c' = achene, 'd' = trigonous achene, 'e' = *Scirpus* spp. (bulrush), achene, and 'f' = *Scirpus americanus* (American bulrush) achene.

Cf. Equisetum fluviatile

Part of a node on the stem

Cf. Nymphaeaceae

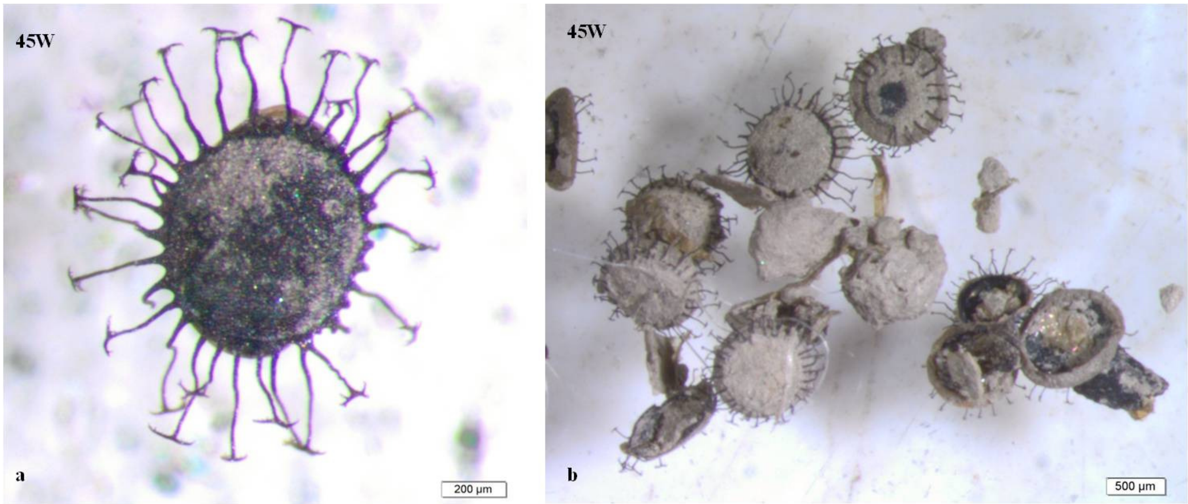


Possible seeds from the Nymphaeaceae family, specifically a species of water lily.

Aquatic Invertebrates

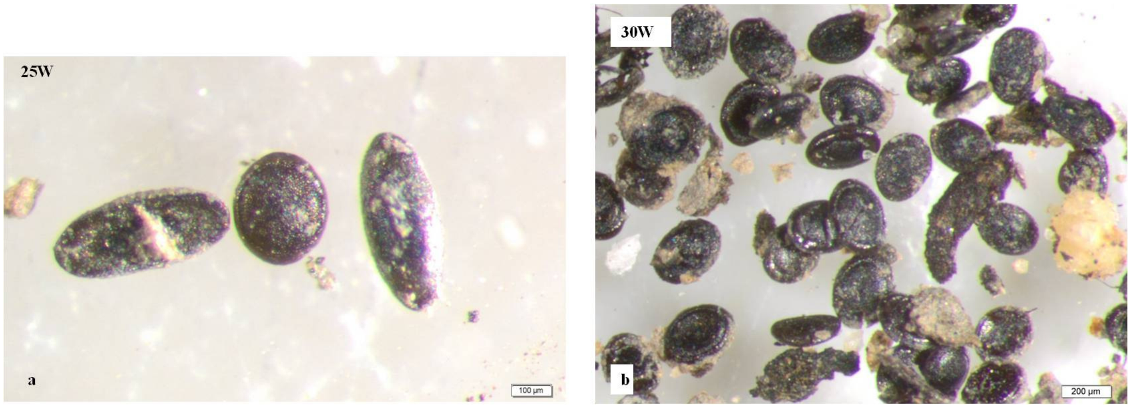
Bryozoa

Cristatella Mucedo



Statoblasts from *Cristatella mucedo*. Statoblasts are capsules that house reproductive material until the bryozoan (moss animal) hatches (Wood 1991).

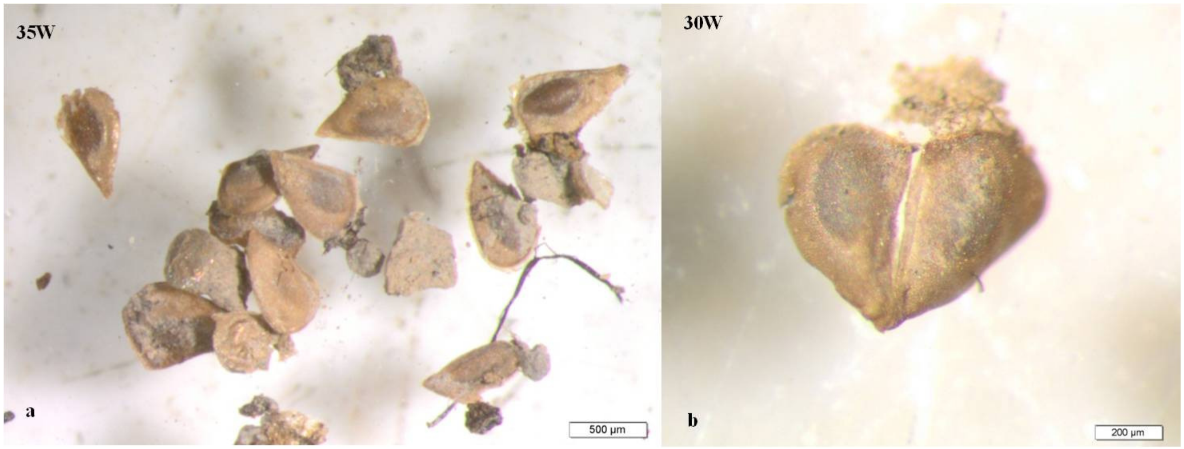
Plumatella sp.



Statoblasts from the genus *Plumatella* (Wood 1991).

Cladocera

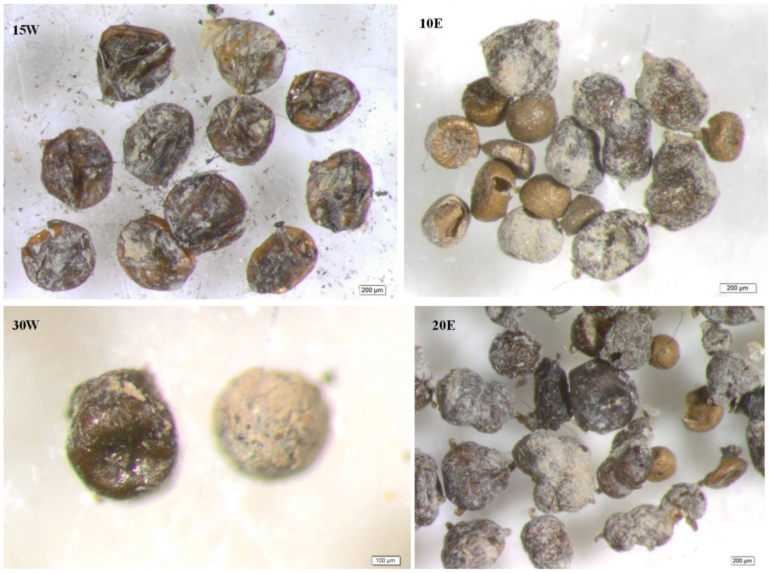
Unidentified species



Ephippia from a member(s) of the Cladocera order (water flea). Ephippia are chitinous shells which house the egg of the water flea to protect it through harsh winter conditions until it hatches when optimal conditions are reached (Dodson and Frey, 1991:731).

Porifera

Unidentified species



Gemmules from possibly two species of Porifera (freshwater sponges). Gemmules are masses of cells which carry genetic material by way of asexual reproduction and develop into freshwater sponges (Frost 1991).

Ostracoda

Unidentified species

Shell from a species of ostracod (seed shrimp) (Delorme 1991).

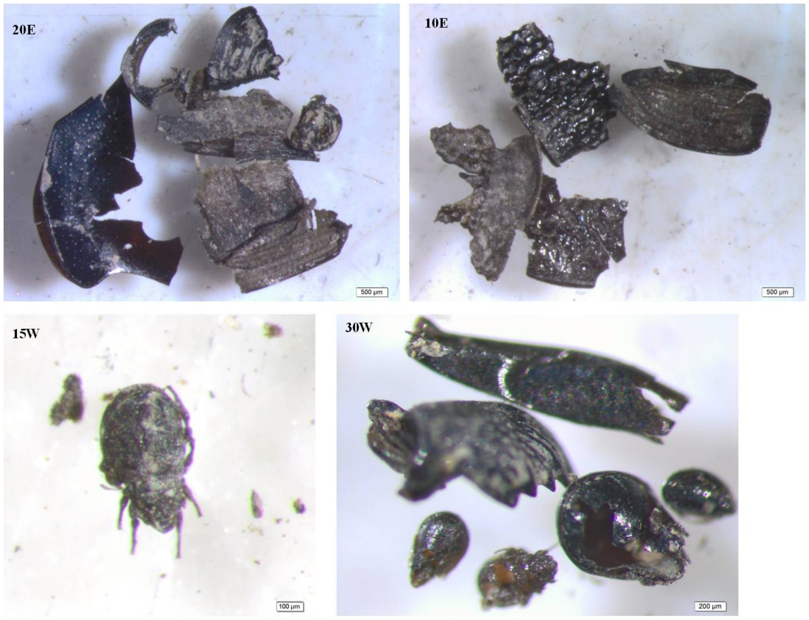
Fungi

Cenococcum geophilum

Sclerotia from the fungi *Cenococcum geophilum*. A sclerotium is a cluster of hyphae which harden to produce a hard body which is able to survive in harsh conditions such as and germinates when environmental conditions become favourable (Pigott 1982).

Insect

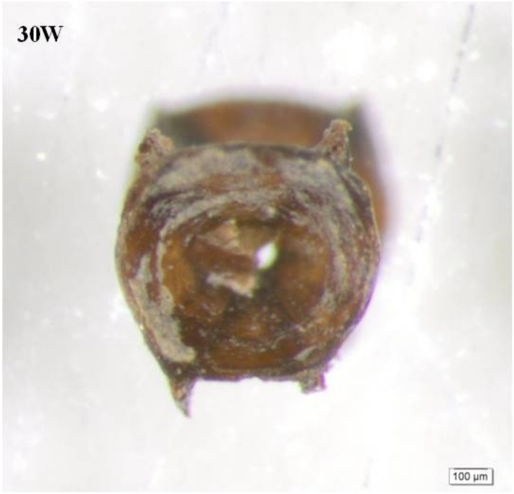
Unidentified species



Various insect remains

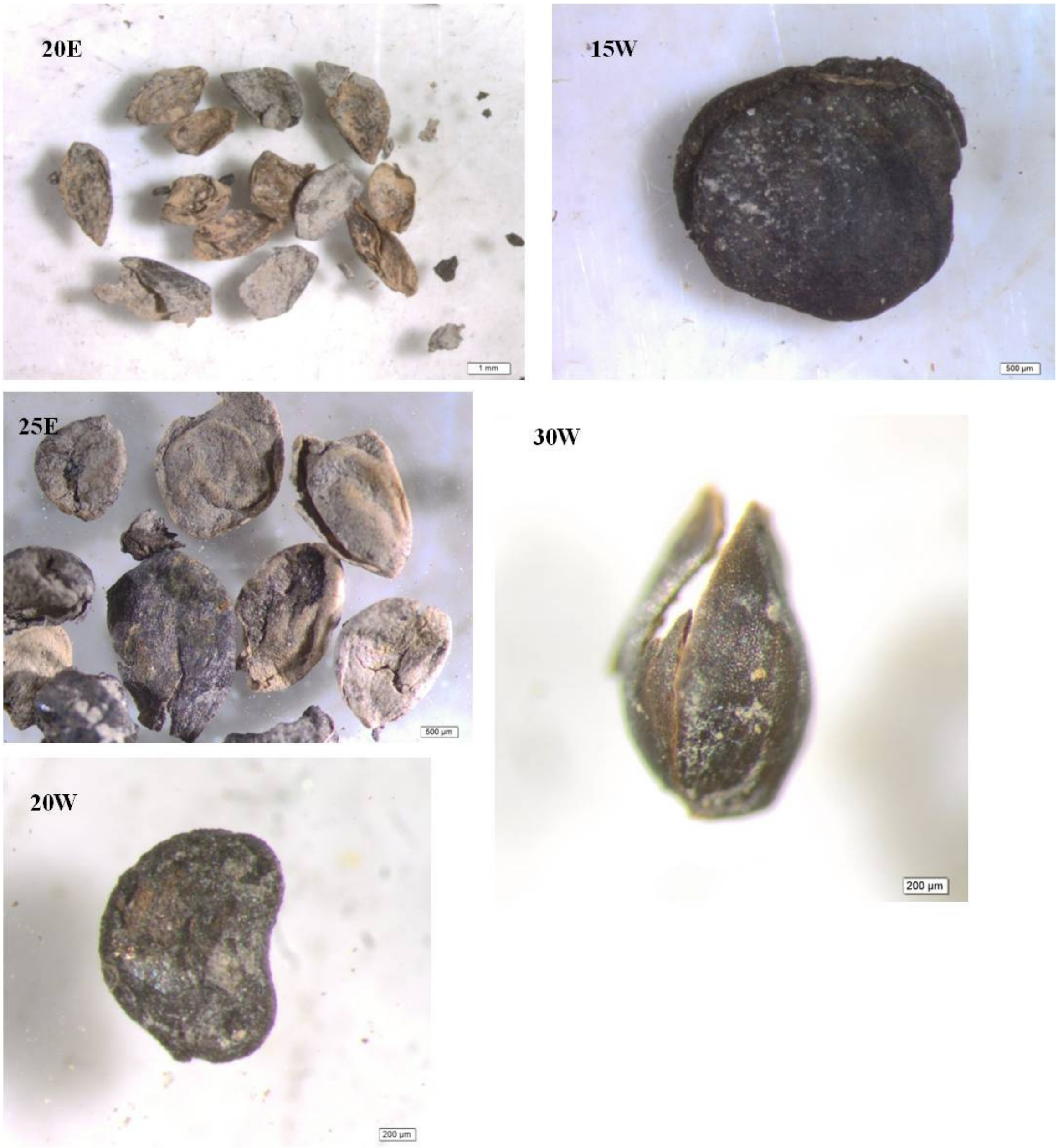
Fish

Unidentified species



Vertebrate from an unidentified species of fish.

Unidentified Seeds



Unidentified seeds found throughout Unit D.

Unidentified Macrofossils

Unidentified macrofossils found within Unit D.

Appendix B

Total Identified Macrofossil Totals divided by sample location and sieve size

Sample Location	**30E**				**25E**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**	**2mm**	**1mm**	**250µm**	**125µm**
Taxa								
Tree/Shrub								
Abies balsamea								
Needles	720	188	75	0	397	158	68	N/A
Twigs	3	0	0	0	1	0	0	N/A
Cf. seeds	0	0	0	0	0	1	0	N/A
Seed wings	1	0	0	0	0	0	0	N/A
Cones	0	0	0	0	0	0	0	N/A
Cone scales	1	0	0	0	2	0	0	N/A
Alnus viridis								
Samaras	0	1	0	0	0	0	0	N/A
Arctostaphylos uva-ursi								
Stones	0	0	0	0	1	0	0	N/A
Betulaceae								
Samaras	0	0	0	0	0	1	0	N/A
Betula papyrifera								
Twigs	0	0	0	0	0	0	0	N/A
Samaras	0	0	1	0	0	0	0	N/A
Fruit scales	0	0	0	0	1	0	0	N/A
Larix laricina								
Needles	167	104	214	0	52	30	109	N/A
Twigs	21	0	0	0	1	0	0	N/A
Seeds	0	0	0	0	1	0	0	N/A
Seed wings	1	0	0	0	1	0	0	N/A
Cones	5	0	0	0	2	0	0	N/A
Cone scales	5	0	0	0	4	0	0	N/A
Picea sp.								
Needles	112	33	4	0	56	14	4	N/A
Twigs	22	0	0	0	6	0	0	N/A
Seeds	0	3	0	0	2	1	0	N/A
Seed wings	0	2	0	0	0	0	0	N/A
Cones	0	0	0	0	0	0	0	N/A
Cone scales	0	0	0	0	0	0	0	N/A
Picea glauca								
Cf. seed wing	0	0	0	0	0	0	0	0
Cone	0	0	0	0	0	0	0	0
Cone scales	0	0	0	0	0	0	0	0
Pinus sp.								
Needles	60	48	68	0	0	0	0	N/A
Twigs	16	0	0	0	0	0	0	N/A
Seeds	0	0	0	0	0	0	0	N/A
Seed wings	0	0	0	0	0	0	0	N/A

Sample Location	**30E**				**25E**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**	**2mm**	**1mm**	**250µm**	**125µm**
Taxa (continued)								
Cones	0	0	0	0	0	0	0	N/A
Cone scales	0	0	0	0	0	0	0	N/A
Rubus idaeus								
Stones	0	0	0	0	0	0	0	N/A
Terrestrial Herbs								
Lycopus americanus								
Nutlets	0	0	0	0	0	0	0	N/A
***cf. Poaceae sp.* Fragment**	0	0	0	0	0	0	0	N/A
cf. Poaceae sp. Seed	0	1	0	0	0	0	11	N/A
Aquatic Herbs								
Cyperaceae								
Achenes	1	3	0	0	0	4	4	N/A
Perigynia	0	0	0	0	0	0	0	N/A
Carex sp.								
Lenticular achenes	2	5	0	0	2	2	1	N/A
Trigonous achenes	1	0	0	0	0	0	0	N/A
cf.Equisetum fluviatile								
Part of node	0	0	0	0	0	0	0	N/A
cf. Nymphaeceae								
Seeds	2	0	0	0	1	0	0	N/A
Scirpus sp.								
Achenes	0	0	0	0	1	1	0	N/A
Scirpus americanus								
Achenes	1	0	0	0	1	0	1	N/A
Bryophytes								
Dicranum polysetum								
Leaves and stems	0	0	0	0	2	0	0	N/A
Pleurozium schreberi								
Leaves and stems	3	0	0	0	86	38	0	N/A
Bryozoans								
Cristatella mucedo								
Statoblasts	0	0	1	0	0	0	0	N/A
***Plumatella** sp.*								
Statoblasts	1	0	18	251	0	0	20	N/A
Cladocera								
Daphnia sp.								
Ephippium	0	0	599	226	0	0	515	N/A
Insect								
Various body parts	1	2	23	50	0	6	18	N/A

Sample Location	**30E**				**25E**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**	**2mm**	**1mm**	**250µm**	**125µm**
Taxa (continued)								
Fungi								
Cenococcum geophilum sclerotia	12	10	44	0	2	1	42	N/A
Fish								
Vertebrae	0	0	0	0	0	0	0	N/A
Ostracod								
Unidentified bivalve	0	0	0	0	0	0	0	N/A
Porifera								
Unidentified gemmules	0	0	30	0	0	0	29	N/A
Totals	**1158**	**400**	**1077**	**527**	**622**	**257**	**822**	**0**

Sample Location	**20E**				**15E**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**	**2mm**	**1mm**	**250µm**	**125µm**
Taxa								
Tree/Shrub								
Abies balsamea								
Needles	53	29	11	N/A	238	114	19	N/A
Twigs	0	0	0	N/A	1	0	0	N/A
Cf. seeds	0	0	0	N/A	1	0	0	N/A
Seed wings	0	0	0	N/A	0	0	0	N/A
Cones	0	0	0	N/A	0	0	0	N/A
Cone scales	0	0	0	N/A	0	0	0	N/A
Alnus viridis								
Samaras	0	0	0	N/A	1	0	0	N/A
Arctostaphylos uva-ursi								
Stones	1	0	0	N/A	0	0	0	N/A
Betulaceae								
Samaras	0	1	0	N/A	0	2	0	N/A
Betula papyrifera								
Twigs	0	0	0	N/A	0	0	0	N/A
Samaras	0	0	0	N/A	1	0	0	N/A
Fruit scales	0	0	0	N/A	0	0	0	N/A
Larix laricina								
Needles	6	6	35	N/A	68	29	43	N/A
Twigs	2	0	0	N/A	3	0	0	N/A
Seeds	0	0	0	N/A	0	0	0	N/A
Seed wings	0	0	0	N/A	1	0	0	N/A
Cones	0	0	0	N/A	0	0	0	N/A
Cone Scales	0	0	0	N/A	4	0	0	N/A
Picea sp.								
Needles	19	8	2	N/A	104	33	6	N/A
Twigs	4	0	0	N/A	29	0	0	N/A
Seeds	1	1	0	N/A	2	0	0	N/A
Seed wings	0	0	0	N/A	3	0	0	N/A
Cones	0	0	0	N/A	0	0	0	N/A
Cone scales	0	0	0	N/A	0	0	0	N/A
Picea glauca								
Cf. seed wing	0	0	0	N/A	0	0	0	N/A
Cone	0	0	0	N/A	0	0	0	N/A
Cone scales	0	0	0	N/A	1	0	0	N/A
Pinus sp.								
Needles	0	0	0	N/A	0	0	0	N/A
Twigs	0	0	0	N/A	0	0	0	N/A
Seeds	0	0	0	N/A	0	0	0	N/A
Seed wings	0	0	0	N/A	1	0	0	N/A
Cones	0	0	0	N/A	0	0	0	N/A
Cone scales	0	0	0	N/A	0	0	0	N/A
Rubus idaeus								
Stones	0	0	0	N/A	0	0	0	N/A

Sample Location	**20E**				**15E**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**	**2mm**	**1mm**	**250µm**	**125µm**
Taxa								
Terrestrial Herbs								
Lycopus americanus								
Nutlets	0	0	0	N/A	0	0	0	N/A
***cf. Poaceae sp.* Fragment**	0	0	0	N/A	0	0	0	N/A
cf. Poaceae sp. Seed	0	0	0	N/A	0	0	0	N/A
Aquatic Herbs								
Cyperaceae								
Achenes	0	3	0	N/A	4	0	2	N/A
Perigynia	0	0	0	N/A	0	1	0	N/A
Carex sp.								
Lenticular achenes	0	2	0	N/A	9	3	0	N/A
Trigonous achenes	0	0	0	N/A	0	0	0	N/A
cf.Equisetum fluviatile								
Part of node	0	0	0	N/A	0	0	0	N/A
cf. Nymphaeceae								
Seeds	0	0	0	N/A	0	0	0	N/A
Scirpus sp.								
Achenes	1	0	0	N/A	0	0	0	N/A
Scirpus americanus								
Achenes	0	0	0	N/A	0	0	0	N/A
Bryophytes								
Dicranum polysetum								
Leaves and stems	0	0	0	N/A	0	0	0	N/A
Pleurozium schreberi								
Leaves and stems	0	0	0	N/A	0	0	0	N/A
Bryozoans								
Cristatella mucedo								
Statoblasts	0	1	5	N/A	0	0	0	N/A
***Plumatella sp*.**								
Statoblasts	0	0	10	N/A	0	0	25	N/A
Cladocera								
Daphnia sp.								
Ephippium	0	0	45	N/A	0	0	178	N/A
Insect								
Various body parts	0	8	48	N/A	11	0	37	N/A
Fungi								
Cenococcum geophilum sclerotia	1	8	122	N/A	4	5	191	N/A
Fish								
Vertebrae	0	0	0	N/A	0	0	0	N/A

Sample Location	**20E**				**15E**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**	**2mm**	**1mm**	**250µm**	**125µm**
Taxa (continued)								
Ostracod								
Unidentified bivalve	0	0	0	N/A	0	0	0	N/A
Porifera								
Unidentified gemmules	0	0	214	N/A	0	0	214	N/A
Totals	**88**	**67**	**492**	**0**	**88**	**67**	**492**	**0**

Sample Location	**10E**				**5E**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**	**2mm**	**1mm**	**250µm**	**125µm**
Taxa								
Tree/Shrub								
Abies balsamea								
Needles	345	90	55	N/A	275	38	36	0
Twigs	2	0	0	N/A	2	0	0	0
Cf. seeds	0	0	0	N/A	2	1	0	0
Seed wings	1	1	0	N/A	2	0	0	0
Cones	0	0	0	N/A	0	0	0	0
Cone scales	1	0	0	N/A	0	0	0	0
Alnus viridis								
Samaras	0	0	0	N/A	0	0	0	0
Arctostaphylos uva-ursi								
Stones	0	0	0	N/A	5	0	0	0
Betulaceae								
Samaras	0	3	1	N/A	0	4	3	0
Betula papyrifera								
Twigs	0	0	0	N/A	0	0	0	0
Samaras	1	0	0	N/A	0	0	0	0
Fruit scales	0	0	0	N/A	0	0	0	0
Larix laricina								
Needles	280	62	96	N/A	126	39	130	0
Twigs	1	0	0	N/A	1	0	0	0
Seeds	1	0	0	N/A	0	0	0	0
Seed wings	0	0	0	N/A	0	0	0	0
Cones	1	0	0	N/A	1	0	0	0
Cone Scales	0	0	0	N/A	1	0	0	0
Picea sp.								
Needles	112	11	3	N/A	172	19	4	0
Twigs	41	0	0	N/A	10	0	0	0
Seeds	0	1	0	N/A	4	1	0	0
Seed wings	0	1	0	N/A	3	0	0	0
Cones	0	0	0	N/A	0	0	0	0
Cone scales	0	0	0	N/A	0	0	0	0
Picea glauca								
Cf. seed wing	0	0	0	N/A	0	0	0	0
Cone	0	0	0	N/A	0	0	0	0
Cone scales	0	0	0	N/A	0	0	0	0
Pinus sp.								
Needles	2	0	0	N/A	1	1	0	0
Twigs	0	0	0	N/A	0	0	0	0
Seeds	0	0	0	N/A	0	0	0	0
Seed wings	1	0	0	N/A	0	0	0	0
Cones	0	0	0	N/A	0	0	0	0
Cone scales	0	0	0	N/A	0	0	0	0
Rubus idaeus								
Stones	0	0	0	N/A	0	0	0	0

Sample Location	**10E**				**5E**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**	**2mm**	**1mm**	**250µm**	**125µm**
Taxa								
Terrestrial Herbs								
Lycopus americanus								
Nutlets	0	0	0	N/A	0	0	0	0
***cf. Poaceae sp.* Fragment**	0	0	0	N/A	3	0	0	0
cf. Poaceae sp. Seed	0	1	0	N/A	1	0	1	0
Aquatic Herbs								
Cyperaceae								
Achenes	0	2	0	N/A	2	6	2	0
Perigynia	0	1	0	N/A	0	0	0	0
Carex sp.								
Lenticular achenes	0	17	1	N/A	2	5	3	0
Trigonous achenes	0	0	0	N/A	0	0	0	0
cf.Equisetum fluviatile								
Part of node	0	0	0	N/A	0	0	0	0
cf. Nymphaeceae								
Seeds	0	0	0	N/A	0	0	0	0
Scirpus sp.								
Achenes	0	0	0	N/A	0	1	4	0
Scirpus americanus								
Achenes	0	0	0	N/A	0	0	0	0
Bryophytes								
Dicranum polysetum								
Leaves and stems	0	0	0	N/A	0	0	0	0
Pleurozium schreberi								
Leaves and stems	0	0	0	N/A	57	11	0	0
Bryozoans								
Cristatella mucedo								
Statoblasts	0	0	1	N/A	0	0	1	0
***Plumatella sp*.**								
Statoblasts	0	0	32	N/A	0	0	26	234
Cladocera								
Daphnia sp.								
Ephippium	0	0	417	N/A	0	0	221	60
Insect								
Various body parts	4	2	51	N/A	0	9	26	6
Fungi								
Cenococcum geophilum sclerotia	0	1	60	N/A	7	3	31	0
Fish								
Vertebrae	0	0	0	N/A	0	0	0	0

Sample Location	**10E**				**5E**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**	**2mm**	**1mm**	**250µm**	**125µm**
Taxa (continued)								
Ostracod								
Unidentified bivalve	0	0	0	N/A	0	0	1	0
Porifera								
Unidentified gemmules	0	0	26	N/A	0	0	0	20
Totals	**793**	**193**	**743**	**0**	**677**	**138**	**489**	**320**

Sample Location	**15W**				**20W**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**	**2mm**	**1mm**	**250µm**	**125µm**
Taxa								
Tree/Shrub								
Abies balsamea								
Needles	132	45	9	N/A	215	59	6	N/A
Twigs	0	0	0	N/A	0	0	0	N/A
Cf. seeds	0	0	0	N/A	1	0	0	N/A
Seed wings	0	0	0	N/A	0	0	0	N/A
Cones	0	0	0	N/A	0	0	0	N/A
Cone scales	0	0	0	N/A	0	0	0	N/A
Alnus viridis								
Samaras	2	3	0	N/A	1	2	0	N/A
Arctostaphylos uva-ursi								
Stones	4	4	0	N/A	0	0	0	N/A
Betulaceae								
Samaras	6	17	18	N/A	5	31	12	N/A
Betula papyrifera								
Twigs	0	0	0	N/A	0	0	0	N/A
Samaras	6	13	18	N/A	4	6	0	N/A
Fruit scales	18	2	0	N/A	18	1	0	N/A
Larix laricina								
Needles	21	18	14	N/A	15	24	0	N/A
Twigs	0	0	0	N/A	1	0	0	N/A
Seeds	0	0	0	N/A	0	0	0	N/A
Seed wings	1	0	0	N/A	1	0	0	N/A
Cones	0	0	0	N/A	0	0	0	N/A
Cone Scales	0	0	0	N/A	0	0	0	N/A
Picea sp.								
Needles	65	14	2	N/A	34	23	11	N/A
Twigs	6	0	0	N/A	5	0	0	N/A
Seeds	0	1	0	N/A	1	0	0	N/A
Seed wings	0	0	0	N/A	1	0	0	N/A
Cones	0	0	0	N/A	0	0	0	N/A
Cone scales	0	0	0	N/A	0	0	0	N/A
Picea glauca								
Cf. seed wing	0	0	0	N/A	1	0	0	N/A
Cone	0	0	0	N/A	0	0	0	N/A
Cone scales	0	0	0	N/A	0	0	0	N/A
Pinus sp.								
Needles	0	0	0	N/A	0	0	1	N/A
Twigs	0	0	0	N/A	0	0	0	N/A
Seeds	0	0	0	N/A	0	0	0	N/A
Seed wings	0	0	0	N/A	0	0	0	N/A
Cones	0	0	0	N/A	0	0	0	N/A
Cone scales	0	0	0	N/A	0	0	0	N/A
Rubus idaeus								
Stones	0	2	0	N/A	0	0	0	N/A

Sample Location	**15W**				**20W**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**	**2mm**	**1mm**	**250µm**	**125µm**
Taxa								
Terrestrial Herbs								
Lycopus americanus								
Nutlets	0	0	0	N/A	0	0	0	N/A
cf. Poaceae sp. **Fragment**	0	0	0	N/A	0	0	0	N/A
cf. Poaceae sp. Seed	0	0	3	N/A	0	0	1	N/A
Aquatic Herbs								
Cyperaceae								
Achenes	2	1	5	N/A	0	2	4	N/A
Perigynia	0	0	0	N/A	0	0	0	N/A
Carex sp.								
Lenticular achenes	0	0	0	N/A	0	0	1	N/A
Trigonous achenes	0	0	0	N/A	0	0	0	N/A
cf.Equisetum fluviatile								
Part of node	0	0	0	N/A	0	0	0	N/A
cf. Nymphaeceae								
Seeds	0	0	0	N/A	0	0	0	N/A
Scirpus sp.								
Achenes	0	4	7	N/A	8	4	0	N/A
Scirpus americanus								
Achenes	1	2	0	N/A	0	0	0	N/A
Bryophytes								
Dicranum polysetum								
Leaves and stems	0	0	0	N/A	35	0	0	N/A
Pleurozium schreberi								
Leaves and stems	0	0	0	N/A	26	0	0	N/A
Bryozoans								
Cristatella mucedo								
Statoblasts	0	1	0	N/A	0	0	0	N/A
Plumatella sp.								
Statoblasts	0	0	63	N/A	0	0	15	N/A
Cladocera								
Daphnia sp.								
Ephippium	0	0	367	N/A	0	0	456	N/A
Insect								
Various body parts	11	5	23	N/A	5	5	81	N/A
Fungi								
Cenococcum geophilum sclerotia	10	7	184	N/A	3	1	240	N/A
Fish								
Vertebrae	0	0	0	N/A	0	0	0	N/A

Sample Location	**15W**				**20W**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**	**2mm**	**1mm**	**250µm**	**125µm**
Taxa (continued)								
Ostracod								
Unidentified bivalve	0	0	0	N/A	0	0	0	N/A
Porifera								
Unidentified gemmules	0	0	53	N/A	0	0	1	N/A
Totals	**285**	**139**	**766**	**0**	**380**	**158**	**829**	**0**

Sample Location	**25W**				**30W**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**	**2mm**	**1mm**	**250µm**	**125µm**
Taxa								
Tree/Shrub								
Abies balsamea								
Needles	14	52	10	N/A	353	207	18	N/A
Twigs	0	0	0	N/A	0	0	0	N/A
Cf. seeds	1	1	0	N/A	0	0	0	N/A
Seed wings	0	0	0	N/A	0	0	0	N/A
Cones	0	0	0	N/A	0	0	0	N/A
Cone scales	0	0	0	N/A	0	0	0	N/A
Alnus viridis								
Samaras	0	0	0	N/A	0	0	0	N/A
Arctostaphylos uva-ursi								
Stones	0	0	0	N/A	8	8	0	N/A
Betulaceae								
Samaras	0	2	0	N/A	2	9	0	N/A
Betula papyrifera								
Twigs	0	0	0	N/A	0	0	0	N/A
Samaras	0	0	0	N/A	0	0	0	N/A
Fruit scales	0	0	0	N/A	2	0	0	N/A
Larix laricina								
Needles	1	4	19	N/A	18	36	33	N/A
Twigs	0	0	0	N/A	0	0	0	N/A
Seeds	0	0	0	N/A	0	0	0	N/A
Seed wings	0	0	0	N/A	0	0	0	N/A
Cones	1	0	0	N/A	0	0	0	N/A
Cone Scales	1	0	0	N/A	0	0	0	N/A
Picea sp.								
Needles	13	28	2	N/A	210	84	11	N/A
Twigs	2	0	0	N/A	7	0	0	N/A
Seeds	1	1	0	N/A	1	4	0	N/A
Seed wings	0	0	0	N/A	0	0	0	N/A
Cones	0	0	0	N/A	0	0	0	N/A
Cone scales	0	0	0	N/A	0	0	0	N/A
Picea glauca								
Cf. seed wing	0	0	0	N/A	0	0	0	N/A
Cone	0	0	0	N/A	1	0	0	N/A
Cone scales	3	0	0	N/A	1	0	0	N/A
Pinus sp.								
Needles	0	0	2	N/A	0	1	1	N/A
Twigs	0	0	0	N/A	0	0	0	N/A
Seeds	0	0	0	N/A	0	0	0	N/A
Seed wings	0	0	0	N/A	0	0	0	N/A
Cones	0	0	0	N/A	0	0	0	N/A
Cone scales	0	0	0	N/A	0	0	0	N/A
Rubus idaeus								
Stones	0	0	0	N/A	0	7	1	N/A

Sample Location	**25W**				**30W**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**	**2mm**	**1mm**	**250µm**	**125µm**
Taxa								
Terrestrial Herbs								
Lycopus americanus								
Nutlets	0	0	0	N/A	0	0	0	N/A
***cf. Poaceae sp.* Fragment**	0	0	0	N/A	0	0	0	N/A
cf. Poaceae sp. Seed	0	0	0	N/A	0	0	0	N/A
Aquatic Herbs								
Cyperaceae								
Achenes	0	1	0	N/A	0	2	0	N/A
Perigynia	0	0	0	N/A	0	0	0	N/A
Carex sp.								
Lenticular achenes	0	1	0	N/A	0	0	0	N/A
Trigonous achenes	0	0	0	N/A	0	0	0	N/A
cf.Equisetum fluviatile								
Part of node	0	0	0	N/A	0	0	0	N/A
cf. Nymphaeceae								
Seeds	0	0	0	N/A	0	1	0	N/A
Scirpus sp.								
Achenes	0	2	0	N/A	0	4	0	N/A
Scirpus americanus								
Achenes	0	0	0	N/A	0	0	0	N/A
Bryophytes								
Dicranum polysetum								
Leaves and stems	0	0	0	N/A	3	0	0	N/A
Pleurozium schreberi								
Leaves and stems	0	0	0	N/A	0	0	0	N/A
Bryozoans								
Cristatella mucedo								
Statoblasts	0	0	1	N/A	0	0	0	N/A
***Plumatella sp*.**								
Statoblasts	0	0	22	N/A	0	0	49	N/A
Cladocera								
Daphnia sp.								
Ephippium	0	0	105	N/A	0	0	381	N/A
Insect								
Various body parts	0	1	32	N/A	0	23	81	N/A
Fungi								
Cenococcum geophilum sclerotia	0	0	16	N/A	9	11	120	N/A
Fish								
Vertebrae	0	0	0	N/A	0	1	0	N/A

Sample Location	**25W**				**30W**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**	**2mm**	**1mm**	**250µm**	**125µm**
Taxa (continued)								
Ostracod								
Unidentified bivalve	0	0	1	N/A	0	0	0	N/A
Porifera								
Unidentified gemmules	0	0	25	N/A	0	0	1	N/A
Totals	**37**	**93**	**235**	**0**	**615**	**398**	**696**	**0**

Sample Location	**35W**				**40W**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**	**2mm**	**1mm**	**250µm**	**125µm**
Taxa								
Tree/Shrub								
Abies balsamea								
Needles	46	29	4	N/A	0	9	1	0
Twigs	0	0	0	N/A	0	0	0	0
Cf. seeds	1	0	0	N/A	0	0	0	0
Seed wings	1	0	0	N/A	0	0	0	0
Cones	0	0	0	N/A	0	0	0	0
Cone scales	0	0	0	N/A	0	0	0	0
Alnus viridis								
Samaras	0	0	0	N/A	0	0	0	0
Arctostaphylos uva-ursi								
Stones	0	0	0	N/A	0	0	0	0
Betulaceae								
Samaras	0	2	1	N/A	0	0	0	0
Betula papyrifera								
Twigs	0	0	0	N/A	0	0	0	0
Samaras	0	3	0	N/A	0	0	0	0
Fruit scales	0	0	0	N/A	0	0	0	0
Larix laricina								
Needles	12	5	29	N/A	0	0	9	0
Twigs	0	0	0	N/A	0	0	0	0
Seeds	0	0	0	N/A	0	0	0	0
Seed wings	0	0	0	N/A	0	0	0	0
Cones	0	0	0	N/A	0	0	0	0
Cone Scales	1	0	0	N/A	0	0	0	0
Picea sp.								
Needles	35	27	2	N/A	2	17	2	0
Twigs	4	0	0	N/A	2	0	0	0
Seeds	1	2	0	N/A	0	1	0	0
Seed wings	1	0	0	N/A	0	0	0	0
Cones	0	0	0	N/A	0	0	0	0
Cone scales	0	0	0	N/A	0	0	0	0
Picea glauca								
Cf. seed wing	0	0	0	N/A	0	0	0	0
Cone	0	0	0	N/A	0	0	0	0
Cone scales	2	0	0	N/A	0	0	0	0
Pinus sp.								
Needles	0	0	1	N/A	0	0	0	0
Twigs	0	0	0	N/A	0	0	0	0
Seeds	0	0	0	N/A	0	0	0	0
Seed wings	0	0	0	N/A	0	0	0	0
Cones	0	0	0	N/A	0	0	0	0
Cone scales	0	0	0	N/A	0	0	0	0
Rubus idaeus								
Stones	0	0	0	N/A	0	0	0	0

Sample Location	**35W**				**40W**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**	**2mm**	**1mm**	**250µm**	**125µm**
Taxa								
Terrestrial Herbs								
Lycopus americanus								
Nutlets	0	0	0	N/A	0	0	0	0
***cf. Poaceae sp.* Fragment**	0	0	0	N/A	0	0	0	0
cf. Poaceae sp. Seed	0	0	0	N/A	0	0	0	0
Aquatic Herbs								
Cyperaceae								
Achenes	0	0	0	N/A	0	0	0	0
Perigynia	0	0	0	N/A	0	0	0	0
Carex sp.								
Lenticular achenes	0	0	0	N/A	0	0	0	0
Trigonous achenes	0	0	0	N/A	0	0	0	0
cf.Equisetum fluviatile								
Part of node	1	0	0	N/A	0	0	0	0
cf. Nymphaeceae								
Seeds	0	0	0	N/A	0	0	0	0
Scirpus sp.								
Achenes	1	3	0	N/A	0	0	0	0
Scirpus americanus								
Achenes	0	0	0	N/A	0	0	0	0
Bryophytes								
Dicranum polysetum								
Leaves and stems	7	0	0	N/A	0	0	0	0
Pleurozium schreberi								
Leaves and stems	51	0	0	N/A	0	0	0	0
Bryozoans								
Cristatella mucedo								
Statoblasts	0	0	0	N/A	1	0	2	0
***Plumatella sp*.**								
Statoblasts	0	0	2	N/A	0	0	0	267
Cladocera								
Daphnia sp.								
Ephippium	0	0	158	N/A	0	0	73	92
Insect								
Various body parts	0	2	25	N/A	0	0	14	23
Fungi								
Cenococcum geophilum sclerotia	0	1	3	N/A	0	0	3	0
Fish								
Vertebrae	0	0	0	N/A	0	0	0	0

Sample Location	**35W**				**40W**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**	**2mm**	**1mm**	**250µm**	**125µm**
Taxa (continued)								
Ostracod								
Unidentified bivalve	0	0	0	N/A	0	0	1	0
Porifera								
Unidentified gemmules	0	0	6	N/A	0	0	9	0
Totals	**164**	**74**	**231**	**0**	**5**	**27**	**114**	**382**

Sample Location	**45W**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**
Taxa				
Tree/Shrub				
Abies balsamea				
Needles	6	6	0	N/A
Twigs	0	0	0	N/A
Cf. seeds	0	0	0	N/A
Seed wings	0	0	0	N/A
Cones	0	0	0	N/A
Cone scales	0	0	0	N/A
Alnus viridis				
Samaras	0	0	0	N/A
Arctostaphylos uva-ursi				
Stones	0	0	0	N/A
Betulaceae				
Samaras	0	0	0	N/A
Betula papyrifera				
Twigs	0	0	0	N/A
Samaras	0	1	0	N/A
Fruit scales	0	0	0	N/A
Larix laricina				
Needles	1	4	0	N/A
Twigs	0	0	0	N/A
Seeds	0	0	0	N/A
Seed wings	0	0	0	N/A
Cones	0	0	0	N/A
Cone Scales	0	0	0	N/A
Picea sp.				
Needles	8	19	0	N/A
Twigs	0	0	0	N/A
Seeds	1	0	0	N/A
Seed wings	1	0	0	N/A
Cones	0	0	0	N/A
Cone scales	1	0	0	N/A
Picea glauca				
Cf. seed wing	0	0	0	N/A
Cone	0	0	0	N/A
Cone scales	0	0	0	N/A
Pinus sp.				
Needles	0	1	1	N/A
Twigs	0	0	0	N/A
Seeds	0	0	0	N/A
Seed wings	0	0	0	N/A
Cones	0	0	0	N/A
Cone scales	0	0	0	N/A
Rubus idaeus				
Stones	0	0	0	N/A

Sample Location	**45W**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**
Taxa				
Terrestrial Herbs				
Lycopus americanus				
Nutlets	0	0	1	N/A
***cf. Poaceae sp.* Fragment**	0	0	0	N/A
cf. Poaceae sp. Seed	0	0	0	N/A
Aquatic Herbs				
Cyperaceae				
Achenes	0	0	0	N/A
Perigynia	0	0	0	N/A
Carex sp.				
Lenticular achenes	0	0	0	N/A
Trigonous achenes	0	0	0	N/A
cf.Equisetum fluviatile				
Part of node	0	0	0	N/A
cf. Nymphaeceae				
Seeds	0	0	0	N/A
Scirpus sp.				
Achenes	0	0	0	N/A
Scirpus americanus				
Achenes	0	0	0	N/A
Bryophytes				
Dicranum polysetum				
Leaves and stems	0	0	0	N/A
Pleurozium schreberi				
Leaves and stems	2	0	0	N/A
Bryozoans				
Cristatella mucedo				
Statoblasts	0	1	13	N/A
***Plumatella sp*.**				
Statoblasts	0	0	0	N/A
Cladocera				
Daphnia sp.				
Ephippium	0	0	44	N/A
Insect				
Various body parts	0	1	1	N/A
Fungi				
Cenococcum geophilum sclerotia	0	0	0	N/A
Fish				
Vertebrae	0	0	0	N/A

Sample Location	**45W**			
Sieve Size	**2mm**	**1mm**	**250µm**	**125µm**
Taxa (continued)				
Ostracod				
Unidentified bivalve	0	0	1	N/A
Porifera				
Unidentified gemmules	0	0	55	N/A
Totals	**20**	**33**	**116**	**0**

Total Identified Macrofossil Count

Taxa	Total
Tree/Shrub	
Abies balsamea	
Needles	**4130**
Twigs	**9**
Cf. seeds	**9**
Seed wings	**6**
Cones	**0**
Cone scales	**4**
Alnus viridis	
Samaras	**10**
Arctostaphylos uva-ursi	
Stones	**31**
Betulaceae	
Samaras	**120**
Betula papyrifera	
Twigs	**0**
Samaras	**54**
Fruit scales	**42**
Larix laricina	
Needles	**1859**
Twigs	**30**
Seeds	**2**
Seed wings	**5**
Cones	**10**
Cone Scales	**16**
Picea sp.	
Needles	**1325**
Twigs	**138**
Seeds	**30**
Seed wings	**12**
Cones	**0**
Cone scales	**1**
Picea glauca	
Cf. seed wing	**1**
Cone	**1**
Cone scales	**7**
Pinus sp.	
Needles	**188**
Twigs	**16**
Seeds	**0**
Seed wings	**2**
Cones	**0**
Cone scales	**0**
Rubus idaeus	
Stones	**10**

Taxa (continued)	**Total**
Terrestrial Herbs	
Lycopus americanus	
Nutlets	**1**
***cf. Poaceae sp.* Fragment**	**3**
cf. Poaceae sp. Seed	**19**
Aquatic Herbs	
Cyperaceae	
Achenes	**50**
Perigynia	**2**
Carex sp.	
Lenticular achenes	**56**
Trigonous achenes	**1**
cf.Equisetum fluviatile	
Part of node	**1**
cf. Nymphaeceae	
Seeds	**4**
Scirpus sp.	
Achenes	**41**
Scirpus americanus	
Achenes	**6**
Bryophytes	
Dicranum polysetum	
Leaves and stems	**47**
Pleurozium schreberi	
Leaves and stems	**274**
Bryozoans	
Cristatella mucedo	
Statoblasts	**28**
Plumatella sp.	
Statoblasts	**1035**
Cladocera	
Daphnia sp.	
Ephippium	**3937**
Insect	
Various body parts	**635**
Fungi	
Cenococcum geophilum sclerotia	**1152**
Fish	
Vertebrae	**1**
Ostracod	
Unidentified bivalve	**4**

Taxa (continued)	**Total**
Porifera	
Unidentified gemmules	**484**
Identified Macrofossil Total	**15849**