USING OSTRACODES AND SEDIMENTS IN PALEOLAGOONS BEHIND THE UPPER CAMPBELL BEACH OF GLACIAL LAKE AGASSIZ TO RECONSTRUCT ITS HISTORY DURING THE EMERSON PHASE

A THESIS PRESENTED TO THE FACULTY OF GRADUATE STUDIES

UNIVERSITY OF MANITOBA IN PARTIAL FULFILMENT OF THE

REQUIREMENTS FOR THE DEGREE OF MASTER OF SCIENCE

© JASON D. MANN

GEOLOGICAL SCIENCES

FEBRUARY 1999

0-612-35074-6

THE UNIVERSITY OF MANITOBA

FACULTY OF GRADUATE STUDIES

*****

COPYRIGHT PERMISSION PAGE

USING OSTRACODES AND SEDIMENTS IN PALEOLAGOONS BEHIND THE UPPER CAMPBELL BEACH OF GLACIAL LAKE AGASSIZ TO RECONSTRUCT ITS HISTORY DURING THE EMERSON PHASE

BY

JASON D. MANN

A Thesis/Practicum submitted to the Faculty of Graduate Studies of The University

of Manitoba in partial fulfillment of the requirements of the degree

of

MASTER OF SCIENCE

TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

ABSTRACT

The level of Lake Agassiz was controlled by three interdependent, though individually dynamic, factors: the position of the late Wisconsinan Laurentide Ice Sheet, the hydrological budget, and the changing location and elevation of the overflowing outlet. During periods of stability at a particular lake level, distinct shorelines (consisting of erosional scarps, and constructional beaches) developed along the vast margin of the lake. The Upper Campbell beach is the best developed of all Lake Agassiz strandlines; it formed during the Emerson Phase (9.9 ka BP – 9.2 ka BP).

Four small lakes, modern remnants of Upper Campbell level backbeach lagoons (each separated by several tens of kilometers), were investigated to 1) obtain radiocarbon dates that constrain the age of the Upper Campbell beach; 2) provide a stratigraphic record of sedimentation from Lake Agassiz through the Holocene, and 3) provide paleohydrological insight -- by using ostracodes -- into conditions (e.g. water depth and temperature) in the lagoon during its periodic connection to Lake Agassiz, and through time thereafter. The four main sites studied include Brokenpipe Lake (near Dauphin, Manitoba), Ruby Lake (near Hudson Bay, Saskatchewan), Jay Jay Lake (southeast of La Ronge, Saskatchewan), and Gregory Lake (northwest of La Ronge Saskatchewan).

Because the Gregory Lake site is located northwest of an area where correlation of the Upper and Lower Campbell beaches is questionable, a regional analysis was undertaken of high resolution Global Positioning System beach elevations and isobases. The interpretation of

the Gregory lake site is presented from two perspectives; one that the site is located behind the Upper Campbell beach and the other that the site is located behind the Lower Campbell beach.

Unfortunately, only Brokenpipe Lake provided a relatively continuous record of sedimentation and ostracodes from its inception in Lake Agassiz to the present. At the Brokenpipe Lake site, deltaic and outwash sands and gravelly sands, deposited into Lake Agassiz by the Valley and Wilson Rivers, served as a headland for erosion and sediment supply for construction of the Upper Campbell beach. The Upper Campbell prograded northward as a subaqueous spit platform at about 346 m elevation, with multiple subaqueous spits built concurrently within the lagoon. The elevation of a relict shoreline around the lagoon of ~ 347 m, which is recorded in till on the western side, indicates that at the time of its inception, the Brokenpipe site lagoon was ~ 7-8 m deep. The final elevation of the Upper Campbell beach at the Brokenpipe site reached ~ 349 m, a result of wave action in Lake Agassiz. Evidence of differential isostatic rebound during the Holocene can be observed in a drowned tributary to the lake. The stratigraphy of the Brokenpipe Lake cores generally consist of dropstone-rich Lake Agassiz glaciolacustrine clay, overlain by gravel, massive to poorly and well-laminated, periodically organic-rich silty-clay and clayey-silt (dated 9350 +/- 70 yr BP [TO-5880] at its base and 9660 +/- 90 yr BP [TO-6206] at its top), and a calcareous marl-peat (dated 5980 +/- 70 yr BP [TO-6205]) -calcareous marl sequence.

Ostracodes such as *Cyclocypris ampla, Cypridopsis vidua, Candona ohioensis,* and *C. rawsoni* in the lower clayey-silt and silty-clay in the Brokenpipe core suggest that the lagoon was separated (though certainly periodic beach overwash was common) from the main body of

Lake Agassiz, where species like *Candona subtriangulata* occur. Also found in this interval of the core was *Cytherissa lacustris*, a species frequently found today in dilute, moderately deep, cold waters. The co-occurrence of *Ilyocypris gibba* and *I. bradyi* in this interval as well, being species that require flowing water during their life cycles, suggest that the lagoon was influenced by influxes of cold, deglacial, stagnant ice meltwater from the Riding Mountain uplands west of the site during this early stage. Several shallow channels, eroded into the till west of Brokenpipe Lake likely served as the conduits; today these channels are peat-filled.

Ostracodes in the upper marl-peat-marl sequence, mainly shallow water loving candonids such as *Candona decora, C. distincta,* and *C. acutula,* suggest that Lake Agassiz had retreated from the Brokenpipe area, forming the incipient stage of modern Brokenpipe Lake. The limited appearance of *Cypris pubera* -- an ostracode indicative of lake senescence -- near the marl-peat transition, and the 5980 yr BP radiocarbon date in the ostracode barren peat suggests that relatively drier mid-Holocene atmospheric conditions (Hypsithermal?) severely decreased water depths in Brokenpipe Lake and induced terrestrialization for a short period of time. Subsequently, a relatively cooler and wetter climate increased water depth in Brokenpipe Lake, restarting calcareous marl deposition and ostracode inhabitation.

The records from the other three sites, while yielding less information than the Brokenpipe Lake site, are summarized and correlated to the Brokenpipe site, and compatibly fit into a chronologic reconstruction of Lake Agassiz history.

Radiocarbon dates and paleoenvironmental reconstructions from other Upper Campbell related sites (described in the literature) suggest that the Upper Campbell formed between about 9.7 ka and 9.4 ka BP. Basal wood dates and the ostracode record from the Brokenpipe paleolagoon support this conclusion, and agree with estimates of the timing of renewed outflow through an eastern (Kaiashk) outlet from Lake Agassiz.

Acknowledgements

I would like to thank my advisor, Dr. James T. Teller (University of Manitoba) for accepting me as a student and making this project possible. Thank you for the financial support in and out of the field, and for always making discussion about Lake Agassiz more interesting and enjoyable. Thanks to my committee members, Drs. Bill Last (University of Manitoba) and Brandon Curry (Illinois Geological Survey), for taking the time to help me polish the thesis. Particularly, I thank you Brandon for taking me into your home as one of your own and teaching me the 'ropes' about ostracodes. Also thanks to Dr. Tim Fisher (Indiana University Northwest) for stimulating ideas and discussions, and for spending time in the field. Big thanks to Gaywood Matile (Manitoba Energy and Mines) and Dr. Harvey Thorleifson (Geological Survey of Canada) for listening to me ramble on, and for giving feedback whenever needed. Thanks to John Rayburn for venturing out in February for fieldwork, when most other students are hitting the beach (though I suppose we were as well....).

I sincerely thank my parents and sisters for their love and support throughout my several years of study for this degree. It would not have been possible to get this far without your help. Thanks all of my friends, near and far, for being constant when I may have been climbing the walls, and for tipping a few (!) beers now and then. And thanks to Brandy for making everything fun, and inspiring me to get it done!

CHAPTER 1 - INTRODUCTION

1.1 - THE LATE WISCONSINAN LAURENTIDE ICE SHEET

The Late Wisconsinan Laurentide Ice Sheet was as areally extensive as the present Antarctic Ice Sheet. It contained more than 50% of the glacial/interglacial difference in global water budget (Andrews, 1987), and had a direct influence on climate (Kutzbach and Wright, 1985). Proglacial lakes that formed at the ice margin not only influenced continental sedimentation and regional continental climate, but the influx of their overflow to the oceans was of possible climatic significance (e.g. Broecker *et al.*, 1989). Tracing this meltwater signal allows a direct linking of the ice sheet record and the oceans (Andrews, 1987).

Lake Agassiz was the largest proglacial lake that formed at the southern margin of the Laurentide Ice Sheet. Over the 4,000 years that Lake Agassiz existed during the last period of deglaciation, it submerged a total area of almost 1 million km^2 (an area spanning 12 degrees latitude and 16 degrees longitude; Fig. 1-1) and it impounded the drainage of more than 2 million km^2 (Teller, 1985). At any one period in time the size of Lake Agassiz was dependant on several factors. During advance and retreat of the Laurentide Ice Sheet in North America, previously established drainage routes were disrupted and dammed, and new basins were formed by erosion and deposition (Teller, 1987).

The volume and size of Lake Agassiz were the result of lake level. Major

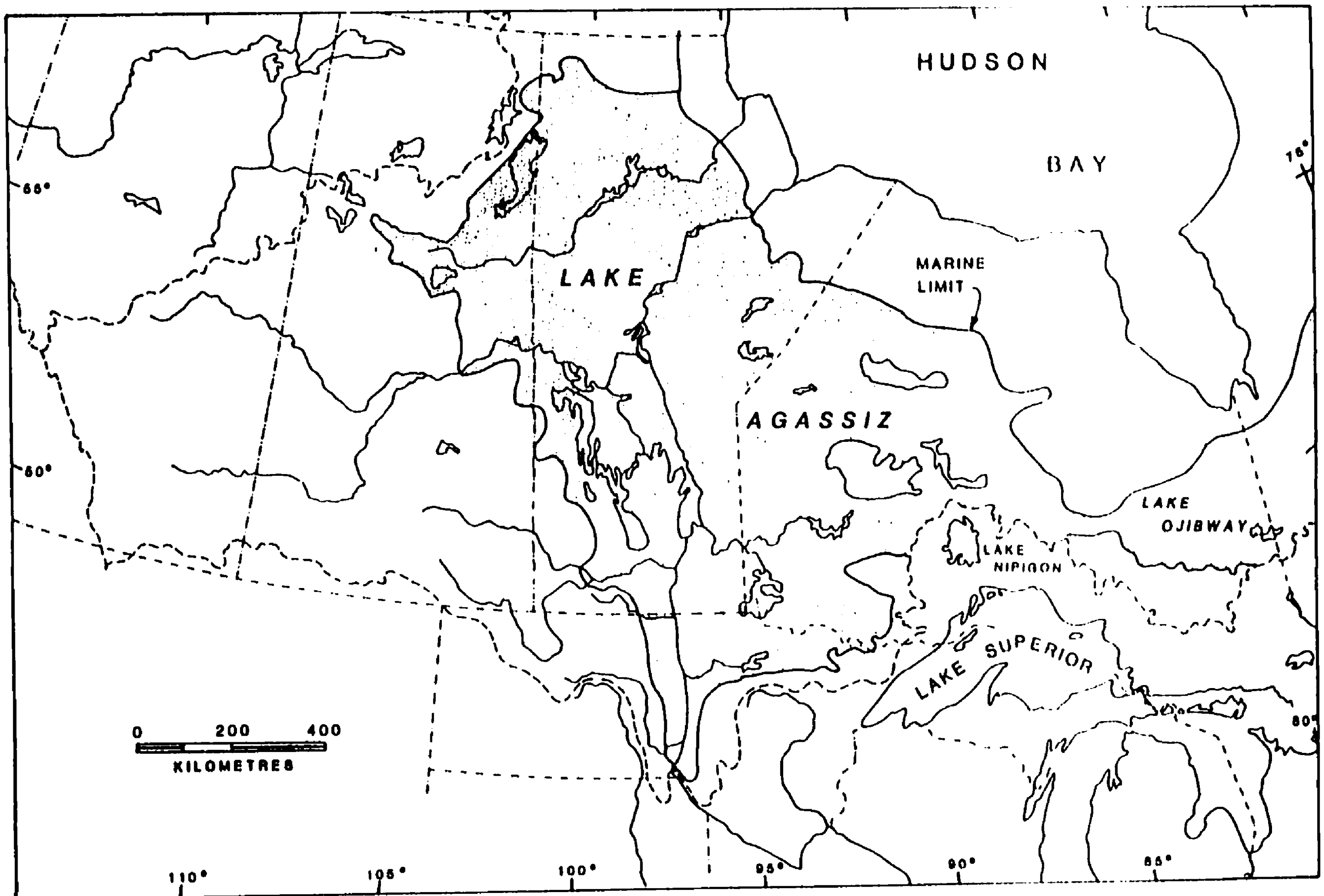


Figure 1-1: Total area above sea level covered by Lake Agassiz during all post-glacial phases. Outline of Lake Agassiz and Lake Ojibway drainage basins is shown by the dashed line. Also note the location of Lake Nipigon. (Teller, 1985, p.3, fig. 1)

elements that control lake level include three interdependent yet individually dynamic factors. These include the elevation of the overflowing outlet (e.g. Teller and Thorleifson, 1983; Clayton, 1983), differential isostatic depression and rebound within the basin (Johnston, 1946; Walcott, 1970, Vincent and Hardy, 1979; Teller and Thorleifson, 1983), and to a lesser extent the water budget (Teller 1990a, 1990b). These factors, in conjunction with the timing of ice retreat in the Lake Agassiz basin, are crucial in understanding the role Lake Agassiz played in routing meltwater from the southern margin of the Laurentide Ice Sheet to other watersheds, and to identifying the potential impact this meltwater had on the oceans into which it ultimately drained (Teller, 1988).

1.2 - GENERAL HISTORY OF LAKE AGASSIZ

1.2.1 - The Inception of Lake Agassiz and the Lockhart Phase (11.7 - 10.9 ka BP)

At about 14 ka BP, as the Red River-Des Moines lobe of the Laurentide Ice Sheet retreated from its maximum extent in Iowa, there were a number of readvances by surging -- followed by the accompanying retreats (Fenton et al., 1983; Clayton et al., 1985). Each successive advance fell short of the previous one. The intervening recessions seem to have sequentially retreated slightly farther north (Teller, 1987) into the Lake Agassiz basin, which up to this point had been filled with ice for more than 10,000 years (Teller, 1985). As the southern margin of the ice sheet was operating in such an oscillatory manner, there were several occasions where the glacial margin had wasted north of the divide between the Hudson Bay and Mississippi watersheds (Fenton et al., 1983). By doing so, there were periods of several hundred years where small lakes developed in the southern (upslope) portion of the Red River Lowland; each of these lakes were overridden by a subsequent ice surge (Teller, 1987). By about 11.7 ka BP, the southern margin of the Laurentide Ice Sheet had wasted north of the Red River Lowland - Mississippi drainage divide for the last time, and the first waters of Lake Agassiz began to collect there by about 11.5 ka, the beginning of the Lockhart Phase (Fig. 1-2).

Overflow during the Lockhart Phase was through the southern outlet to the Mississippi River basin. Deep-water sediments deposited during the Lockhart Phase are the silty clays of the Brenna Formation (Teller, 1976 and references therein) which sporadically contain ice-rafted dropstones (Teller, 1985). Several subaqueous (underflow) fans developed along the western margin of Lake Agassiz at this time as well. These were deposited by large influxes of western Canadian runoff and formed at the mouths of the Sheyenne, Pembina-Souris, and Assiniboine Rivers (Brophy and Bluemle, 1983; Kehew and Clayton, 1983; Klassen, 1983a). By the end of the Lockhart Phase (about 10.9-10.8 ka), Lake Agassiz had expanded at least 700 km north into central Manitoba, and the margin of the Laurentide Ice Sheet had nearly wasted far enough northward to allow overflow from Lake Agassiz to enter the Lake Superior basin through a lower elevation route (Teller, 1987). Shorelines formed during the Lockhart Phase include the Alice, Trail, and Herman strandlines (Fig. 1-3). Work done

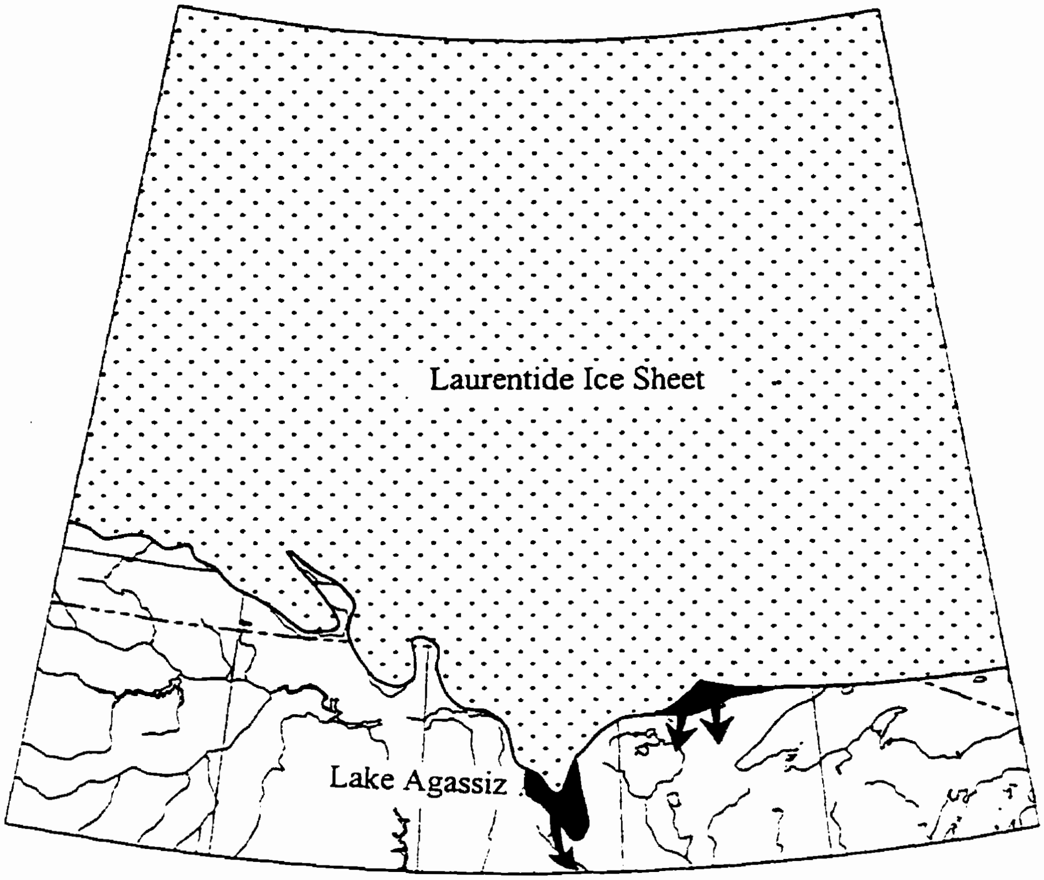


Figure 1-2: Lake Agassiz paleogeography at about 11.5 ka BP. (Thorleifson, 1996, p.68, fig. 31)

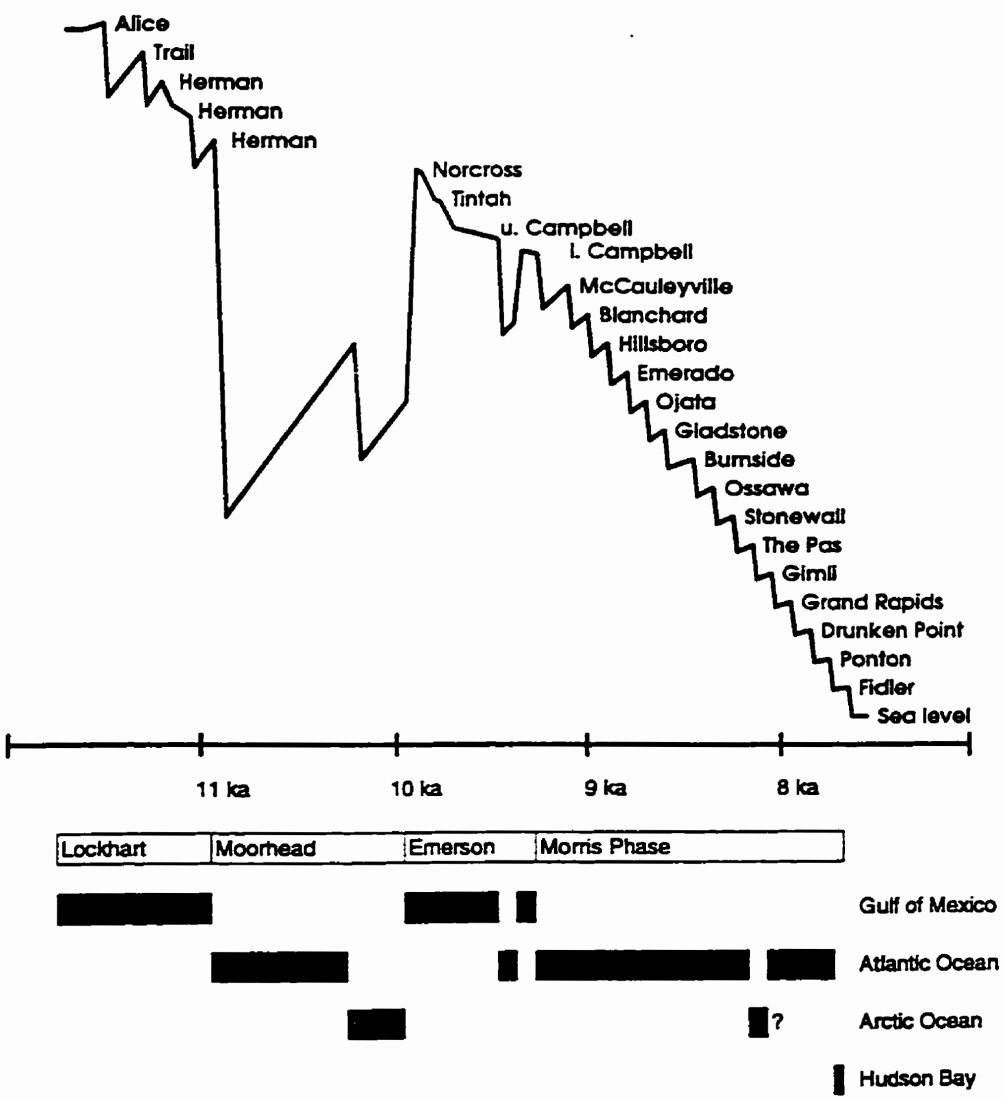


Figure 1-3: Schematic model for the trend and approximate relative magnitude of Lake Agassiz water level fluctuations, subdivision of Lake Agassiz history into four phases, and proposed routing of freshwater discharge from the lake. (Thorleifson, 1996, p.66, fig. 29)

by Upham (1895), Johnston (1946), and Elson (1967), summarized and updated by Thorleifson (1996), suggests that there were several Herman shorelines (Fig. 1-3), the latter being at "stepwise" levels below the first.

1.2.2 - The Moorhead Phase (10.9 - 9.9 ka BP)

The Lockhart Phase of Lake Agassiz ended when the margin of the Laurentide Ice Sheet retreated from the Steep Rock Moraine at the Minnesota-Ontario border (Thorleifson, 1996) to a position north of Thunder Bay, Ontario. This initiated deposition of the Eagle-Finlayson moraine, and a reorientation of the eastern end of the moraine to the Brule position (Zoltai, 1961, 1963, 1965; Prest, 1970; Fig. 1-4). The change in ice position allowed the drainage of Lake Agassiz overflow into the Lake Superior basin, and the initiation of declining Lake Agassiz levels (Fig. 1-5). The commencement of lake drainage to Thunder Bay through the Shebandowan outlet and Savanne outlet (numbers 18 and 19 in Fig. 1-6) dropped the level of Lake Agassiz below the southern outlet (Teller and Thorleifson, 1983; Teller, 1985; Thorleifson, 1996), thereby causing it to be abandoned.

The ability to pinpoint which of the eastern outlets (refer to Fig. 1-6) carried Agassiz overflow is dependant on knowing their elevations in relation to dated beaches in the Lake Agassiz basin during Moorhead time (Thorleifson, 1983; Teller and Thorleifson, 1983; Teller, 1985). As described by Teller (1985), the elevations of the eastern outlets during the Moorhead Phase would depend upon the relative amount of differential isostatic rebound

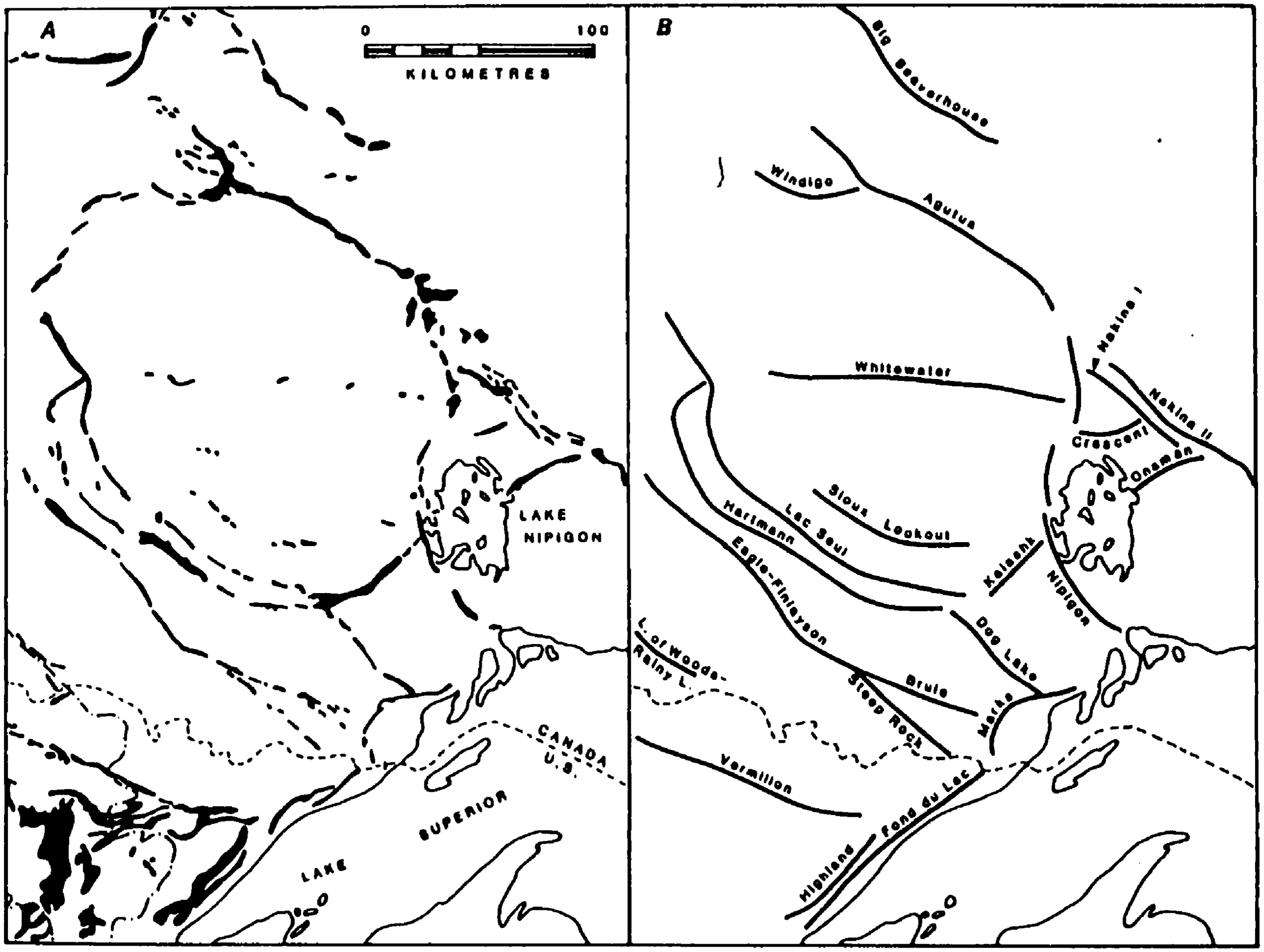


Figure 1-4: End moraines in the eastern outlet region of Lake Agassiz showing (A) actual outline of end moraines and (B) names applied to the ice-marginal zones. Dashed line in A represents easternmost boundary of calcareous till. (Teller and Thorleifson, 1983, p. 269, fig. 5)

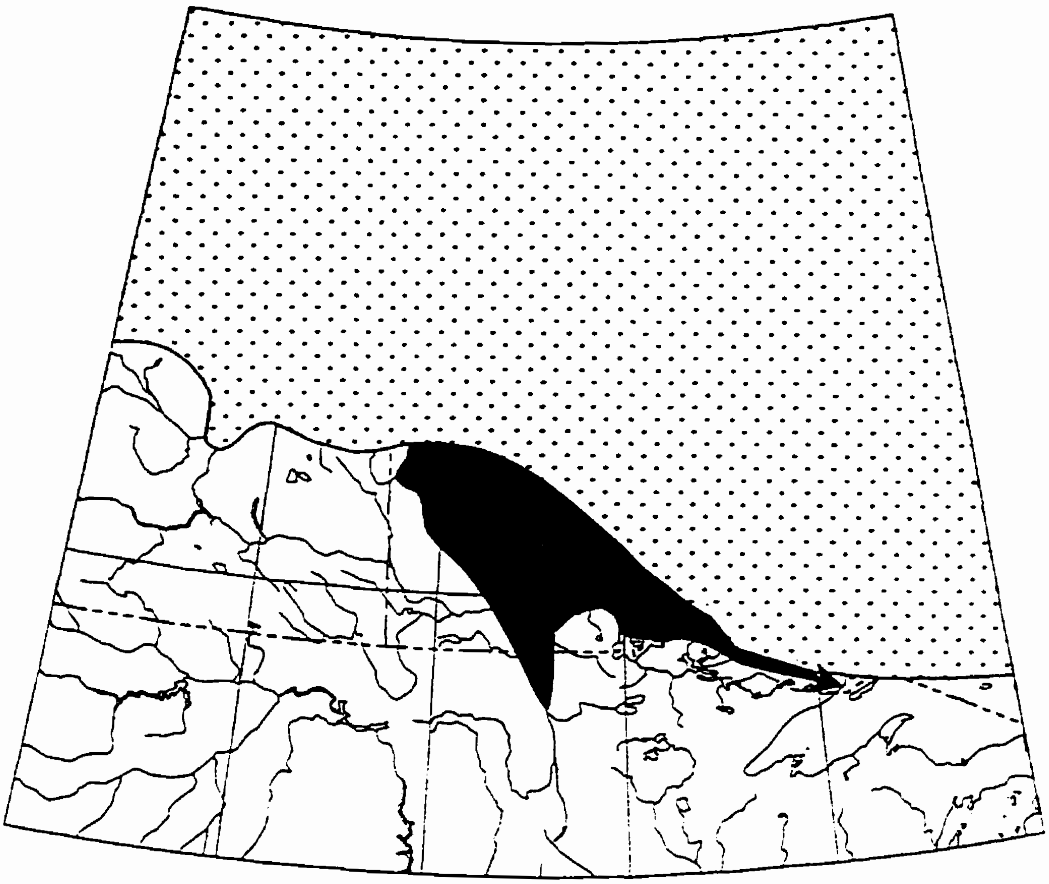


Figure 1-5: Lake Agassiz paleogeography at about 10.9 ka BP. (Thorleifson, 1996, p. 71, fig. 34)

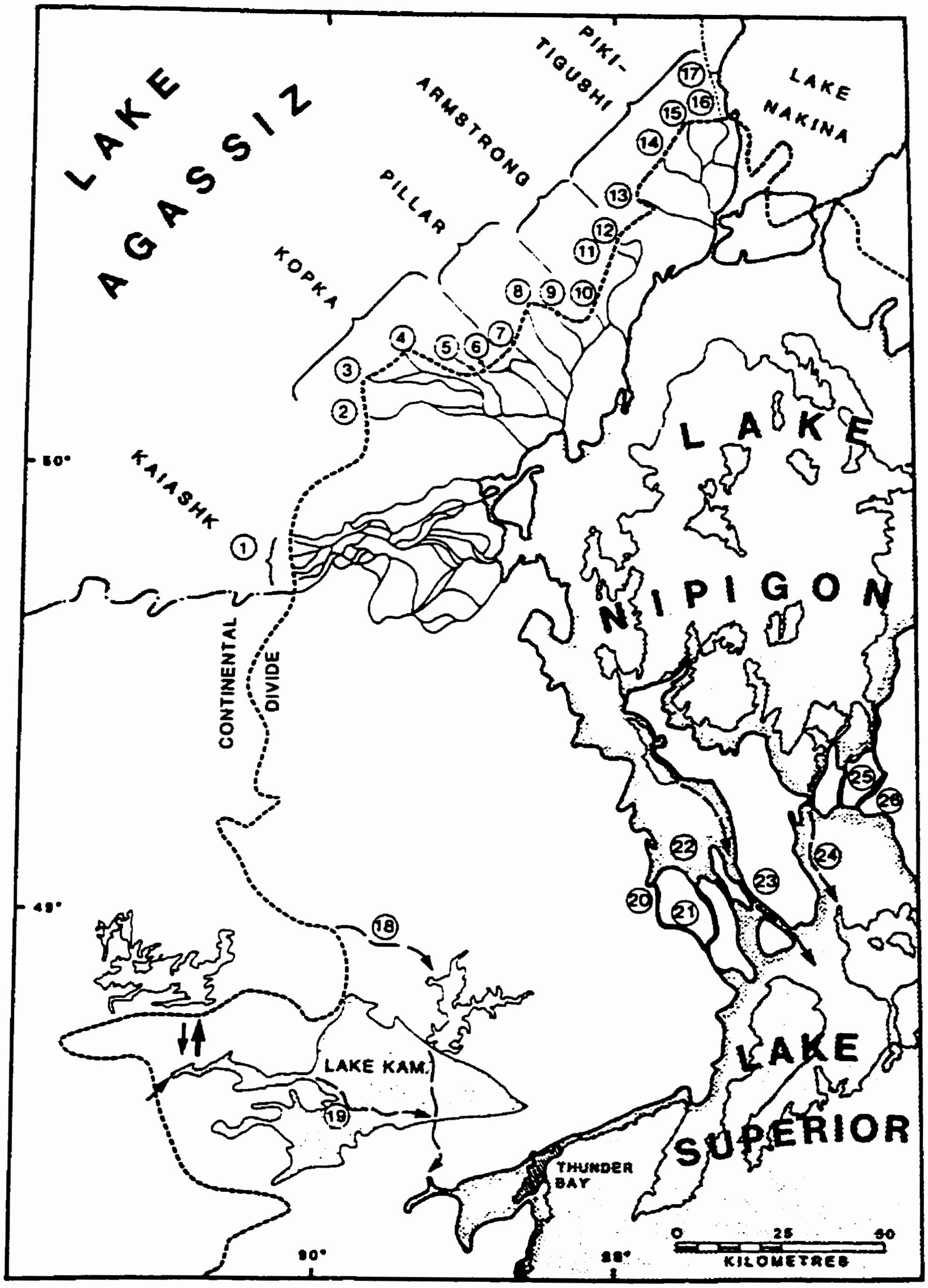


Figure 1-6: Maximum areas covered by Lakes Agassiz, Nipigon, Superior, and Nakina during the Nipigon Phase of Lake Agassiz. Note the locations of the Kaiashk outlets and Lake Kaministikwia (Lake Kam.). (Teller and Thorleifson, 1983, p. 272, fig. 7)

that had occurred by Moorhead time. If rebound had been significant, the more northerly (lower elevation) outlets would have overflowed to allow Agassiz to drop in level as it did during the Moorhead. If rebound had been relatively less, the margin of the Laurentide Ice Sheet would not have had to retreat as far north to expose eastern outlets that were at elevations low enough to carry Lake Agassiz overflow, causing the Moorhead lake level drop (Teller, 1985).

The radical lake level drop of more than 150 m, due to the reopening of this series of progressively lower eastern outlets, exposed much of the lake floor in the southern part of the Agassiz basin by about 10.5 ka BP (Teller, 1985). Subaerial organic material dated at ~ 10 ka BP has been found sandwiched between two offshore clay units in the southern Red River valley (Matile and Thorleifson, 1996), in southeastern Manitoba (Matile and Thorleifson, 1996), and in the Lake of the Woods regions (Bajc, 1987). In fact, subaerial vegetation probably became well established as far north as southeastern Manitoba (Risberg *et al.*, 1995; Teller *et al.*, 1996; Teller *et al.*, submitted), in the Lake Agassiz basin south of the 49th parallel (North Dakota and Minnesota), and in the Rainy River district of northwestern Ontario (Thorleifson, 1996). Buried by post Moorhead Phase sediments, fluvial sediments entrenched into older (Brenna Formation) Lake Agassiz sediments occur as far north as the 49th parallel (Arndt, 1977), and have also been observed in southeastern Manitoba (Matile, 1998, pers. comm.). Teller and Last (1981, 1982) observed a buried pedogenic horizon in the Lake Manitoba basin, which they interpreted to indicate that Lake Agassiz could have been no deeper than 20 m at

Winnipeg, during the lowest level of the Moorhead. A new idea suggests that differential isostatic uplift of the Savanne outlet and the Thunder Bay outlets by late Moorhead time caused a transgression of Lake Agassiz from the Grand Forks area to a position south of Fargo, thereby drowning sites in southeastern Manitoba (Thorleifson, 1996).

Surprisingly, there is no clear evidence for a low-water stage in northwestern Ontario. An uninterrupted series of glaciolacustrine clay varves were deposited during this time (Warman, 1991; Barnett, 1992). As well, there is no evidence for a major readvance of the Laurentide ice sheet at this time in northwestern Ontario; the continuous and uniform clay varves directly argue against this possibility (cf. Warman, 1991; Barnett, 1992). As such, the ice probably remained at the Eagle-Finlayson and Brule moraines (Fig. 1-4), and the Savanne outlet was probably the northernmost eastern outlet during the Moorhead (Thorleifson, 1996).

1.2.3 - The Emerson Phase (9.9 ~ 9.4 ka): The "Traditional" View

The well dated 9.9 ka BP Marquette readvance (Clayton and Moran, 1982; Fenton *et al.*, 1983; Drexler *et al.*, 1983; Clayton, 1983) of the Laurentide Ice Sheet into the Superior basin closed the eastern outlets of Lake Agassiz and caused the level of the lake to rise (Elson, 1967; Teller and Thorleifson, 1983) (Fig. 1-7). The Marquette readvance also isolated small lakes in the western part of the Superior Basin (Clayton, 1983; Teller, 1987). Most notably, the Superior Lobe ice barricade caused water to be ponded against the continental divide west of Thunder Bay, and against the Marks moraine (Fig. 1-4), in the Lake Kaministikwia basin (Fig. 1-6) (Teller and Thorleifson,

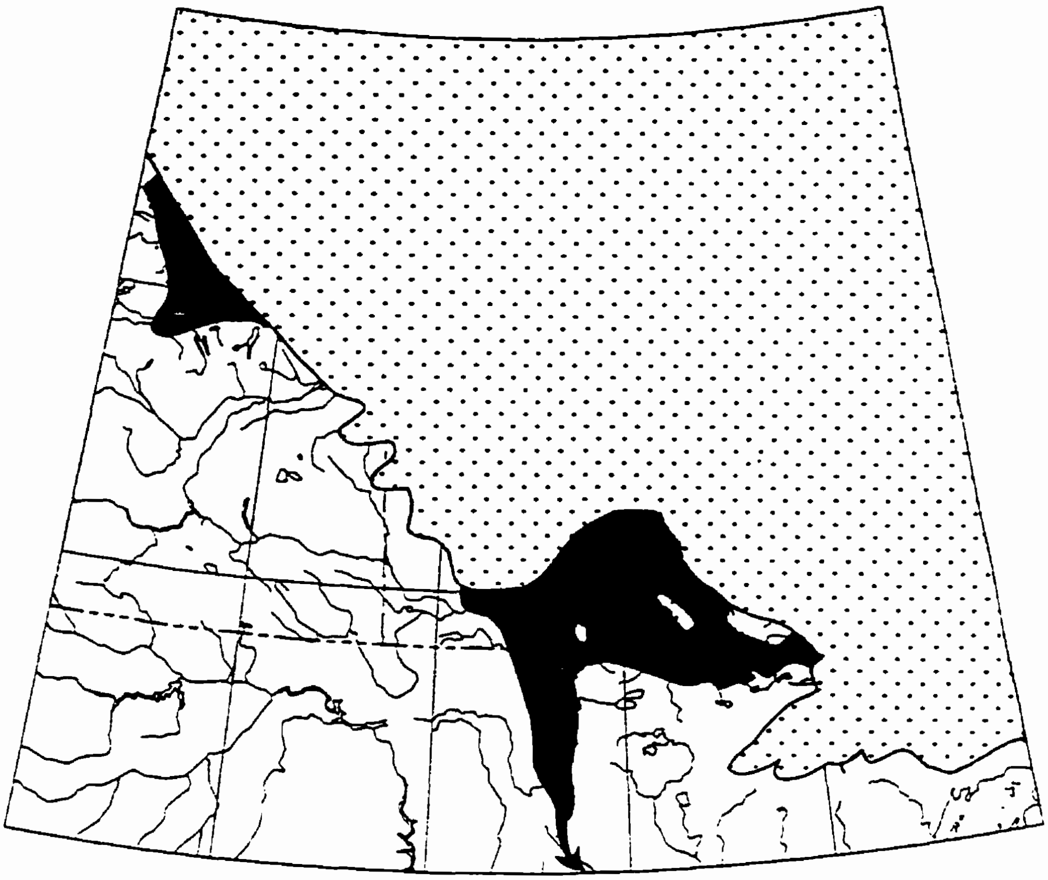


Figure 1-7: Lake Agassiz paleogeography at about 9.9 ka BP (Norcross level – Thorleifson, 1996, p. 73, fig. 36). Traditional interpretations (e.g. Teller *et al.*, 1983) state that the level of Lake Agassiz at this time was the Upper Campbell level (see fig. 1-9 for this 'traditional' paleogeography)

1983). The Marquette readvance marked the beginning of the Emerson Phase of Lake Agassiz.

Traditional interpretations of the history of Lake Agassiz (e.g. Elson, 1967; Fenton *et al.*, 1983; Teller, 1985) indicate that the southern outlet had been eroded to the Campbell level (Fig. 1-3) during the Lockhart Phase (11.7-10.9 ka BP), before the eastern outlets were opened at the end of the Lockhart Phase. This eastern outlet opening had lowered the lake level below that of the Campbell beach. Because of the Marquette readvance, the water level of Lake Agassiz at the start of the Emerson Phase rose to at least the Campbell level, again initiating Lake Agassiz overflow through the southern outlet, to the Mississippi River basin, and ultimately the Gulf of Mexico. The Marquette readvance also forced some of Lake Superior Basin overflow south into the Mississippi system; however overflow from Lake Kaministikwia travelled westward into Lake Agassiz, and a number of red varves were deposited in the easternmost part of the Lake Agassiz basin that record this event (Fig. 1-8) (Zoltai, 1961; Thorleifson, 1983; Teller and Thorleifson, 1983). At the Upper Campbell level, Lake Agassiz covered at least 350, 000 km^2 (Teller, 1985, 1987; Mann *et al.*, 1997, Mann *et al.*, 1998 in press), and was the most areally extensive level of Lake Agassiz, as well as the largest Pleistocene lake in North America (Fig. 1-9) (Teller, 1987). During the Upper Campbell level, the Sherack Formation silty clay (and correlatives) was deposited over the older lacustrine sediments and Moorhead Phase fluvio-organic unconformity (Harris *et* al., 1974; Teller, 1976; Fenton *et* al., 1983; Teller and Last, 1982). The final period of southward overflow of the

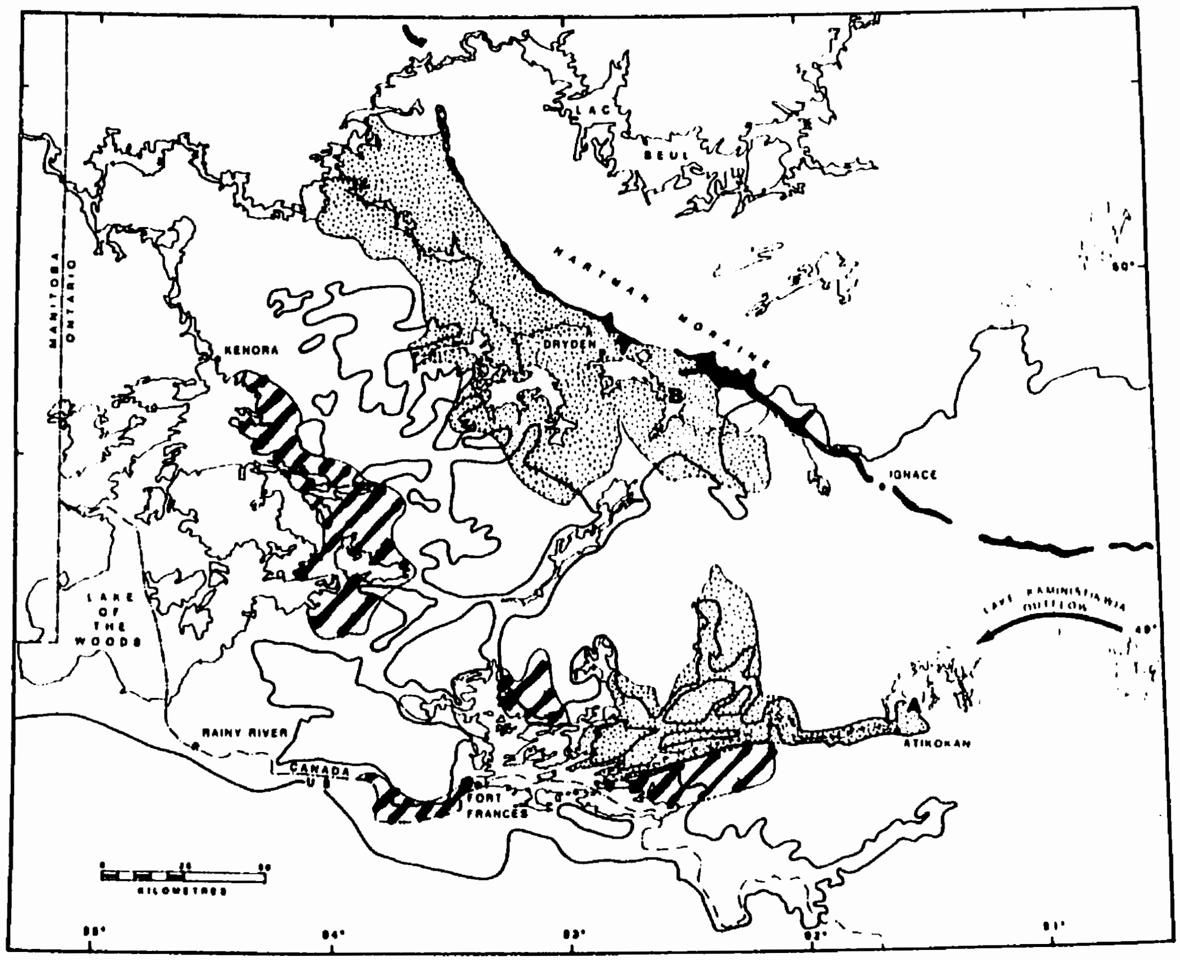


Figure 1-8: Areas where red lacustrine clay is found in northwestern Ontario, indicating where it is varved (coarse stipples) and where it is massive (diagonal stripes). The heavy line represents the Upper Campbell strandline. (modified from Teller and Thorleifson, 1983, p. 270, fig. 6)

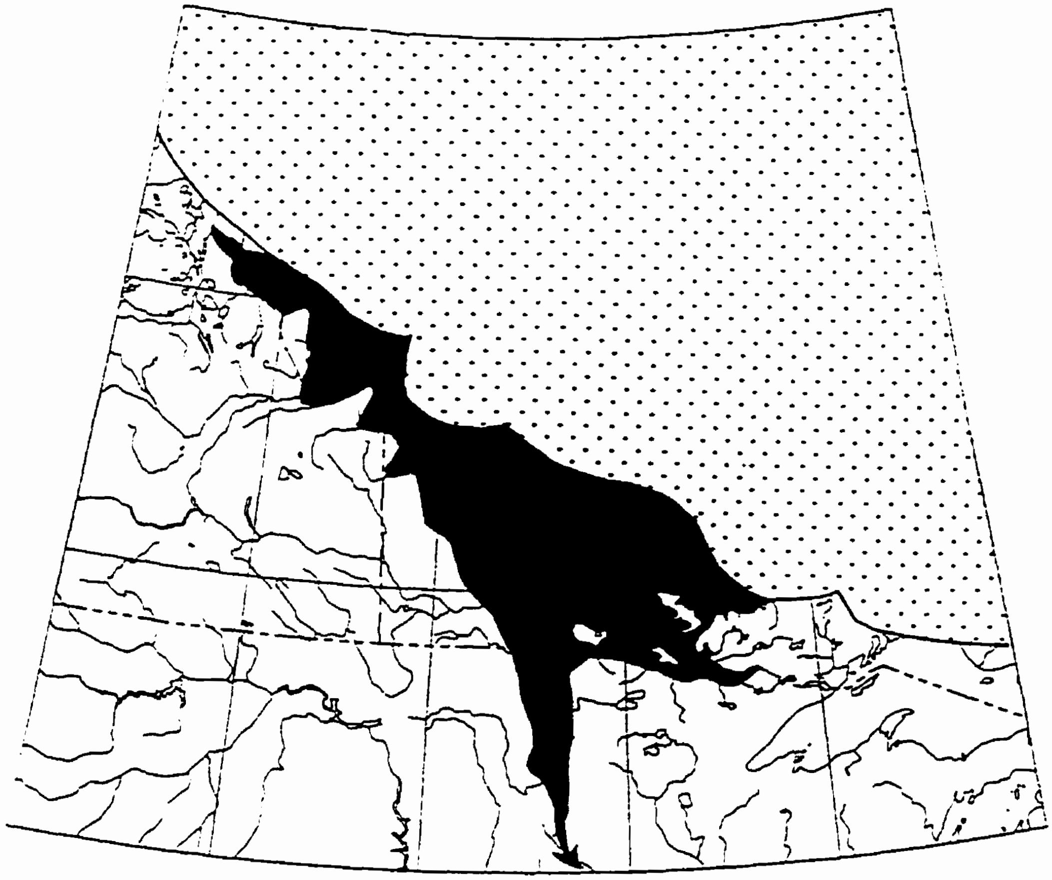


Figure 1-9: Lake Agassiz paleogeography at about 9.3 ka BP (Thorleifson, 1996, p. 74, fig. 37). Traditional interpretations (eg. Teller *et al.*, 1983) place a similar paleogeographic representation of the Lake Agassiz basin at 9.9 ka BP.

Emerson Phase Lake Agassiz had lake levels falling to, but not below, the Campbell level.

1.2.4 - The Morris Phase (9.5 - 7.5 ka BP)

Retreat of the ice margin from the Sioux Lookout moraine (Fig. 1-4) initiated the onset of the Morris Phase. The relatively higher elevation southern outlet was abandoned as the Kaiashk outlet was deglaciated (Teller and Thorleifson, 1983). Lake Agassiz lowered in a step-wise manner as progressively lower eastern outlets to the Nipigon basin were deglaciated (Fig. 1-3) (Teller and Thorleifson, 1983; Teller, 1985). From about 9.5 ka BP to about 8.5 ka BP, Lake Agassiz fell almost 200 m from the Lower Campbell (at the start of the Morris Phase) to the Gimli level (Fig. 1-3), which is the last and lowest beach formed when Lake Agassiz was draining into Lake Superior (Thorleifson, 1983). A number of overflow bursts from Lake Agassiz into the Nipigon Basin occurred during this period (e.g. Teller and Thorleifson, 1983; Teller and Mahnic, 1988).

As the Laurentide Ice Sheet retreated northward from the Lake Nipigon basin (Fig. 1-6) and the divide between the Superior and Agassiz watershed, the early part of the Morris Phase ended. By about 8.5 ka, Lake Agassiz overflow bypassed the Great Lakes completely (Fig. 1-10). This event marked the beginning of the latter part of the Morris Phase when Lake Agassiz overflow was directed into proglacial Lake Ojibway (Fig. 1-1), then into the Ottawa River valley (Teller, 1985). The combined Great Lakes and Lake Agassiz overflow entered the St. Lawrence River valley

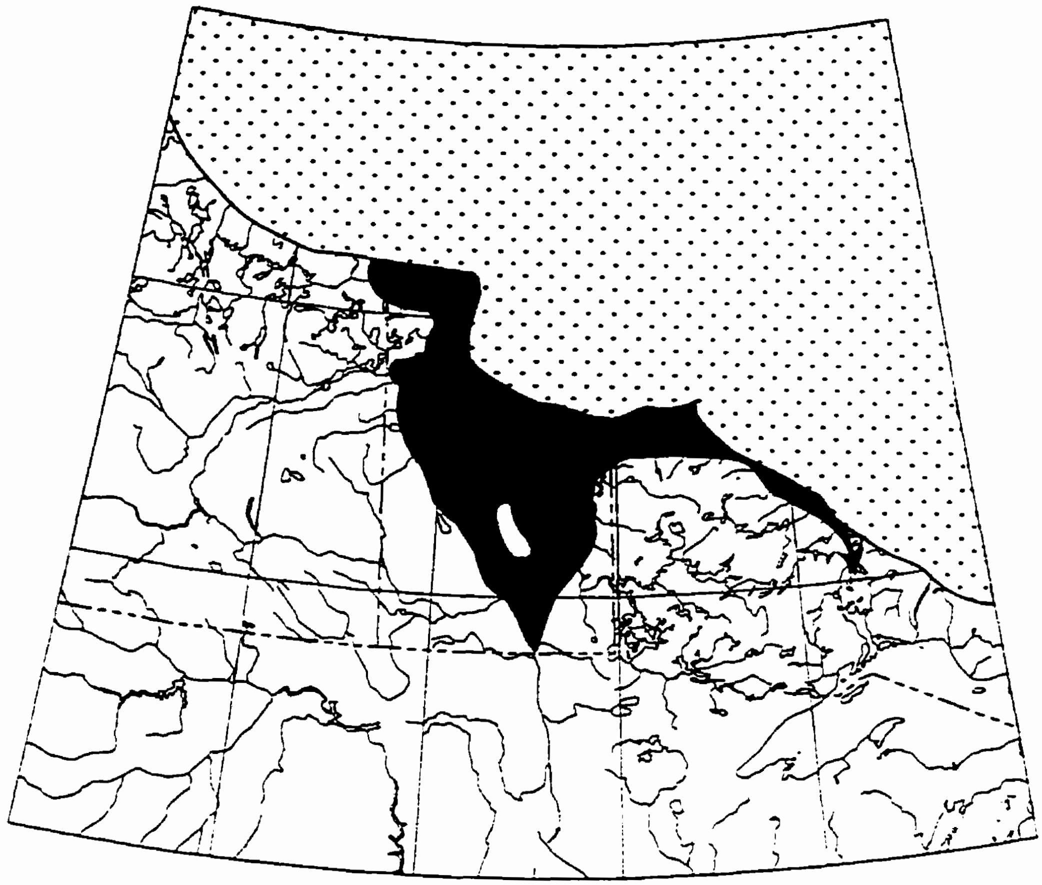


Figure 1-10: Lake Agassiz paleogeography at about 8.2 ka BP. (Thorleifson, 1996, p. 75, fig. 38)

near Montreal (Prest, 1970; Vincent and Hardy, 1979). By the time the southern margin of the Laurentide ice sheet had retreated into northeastern Manitoba and northern Ontario (Fig. 1-11), Lake Agassiz occupied the recently deglaciated Hudson Bay area and adjacent lowlands (Teller, 1985). Klassen (1983b) notes that by 7.7 ka, Lake Agassiz had begun to drain to the Tyrrell Sea by way of subglacial and ice marginal channels of the stagnant Hudson lobe. Lake Agassiz finally drained into Hudson bay by about 7.5 ka BP, through the Hudson Strait to the North Atlantic Ocean, rather than through the St. Lawrence River Valley (Klassen, 1983b).

1.2.5 - An Alternative Interpretation to the Early Emerson Phase

In a new interpretation of the Emerson Phase (Fig. 1-3), Thorleifson (1996) contends that:

1. The southern outlet was not eroded below the lowest Herman level during the Lockhart.

2. At the beginning of the Emerson Phase, Lake Agassiz first rose to the lowest Herman level (Fig. 1-3) which is at the final Lockhart Phase level. Water levels had to rise to the level where erosion stopped during the Lockhart Phase. Due to isostatic rebound during the Moorhead, differentially uplifted Herman strandlines were at elevations not attainable by early Emerson water levels.

3. Subsequent erosion downcut the southern outlet, establishing the Norcross and Tintah levels.

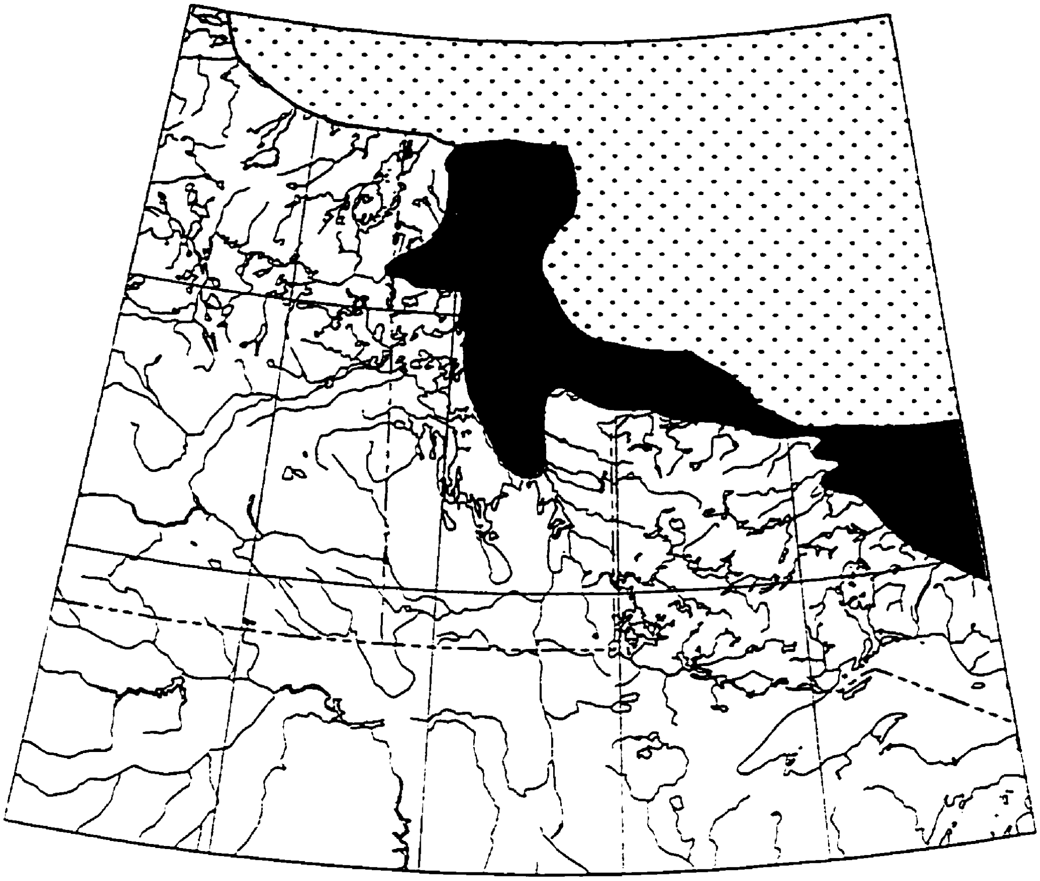


Figure 1-11: Lake Agassiz paleogeography at about 7.8 ka BP. (Thorleifson, 1996, p. 77, fig. 40)

4. The Campbell levels were next occupied, also resultant from erosion of the southern outlet during the Emerson.

5. Sub-Campbell levels formed during the early part of the Morris Phase (which lasted from 9.5-8.5 ka BP) as the successively lower eastern outlets carried overflow.

Thorleifson (1996) stated that the Marquette readvance initiated the Emerson Phase by closing the eastern outlets, causing Lake Agassiz water levels to rise to the Norcross level and overflow through the southern outlet (Figs. 1-7; 1-3). The southern outlet was downcut to the Upper Campbell level (Fig. 1-3). Later, ice retreat from the Lac Seul moraine (Fig. 1-4) in northwestern Ontario (Thorleifson, 1983) opened the relatively lower elevation eastern outlets. This outlet opening caused Agassiz water levels to fall from the Campbell level to the Lower Campbell. Eventually the southern outlet was abandoned altogether, which marked the inception of the early Morris Phase.

The problem with the traditional Emerson Phase reconstructions, with Lake Agassiz only rising to the Campbell level at the onset of the Emerson Phase, is that water depths were too shallow for clay transport and deposition over the Lake Agassiz - Lake Kaministikwia divide. An Emerson Phase maximum level at the Norcross, as described by Thorleifson (1996) offers several benefits. The position of the ice margin, during the Marquette readvance (Fig. 1-7), near the Interlake region in south-central Manitoba (Thorleifson, 1996), helps explain the absence of the Sherack Formation in the Interlake, whereas there are meters of the Sherack clays near Winnipeg. Also, there would have been sufficient water depths across the drainage divide west of Thunder Bay to facilitate red clay transport and sedimentation. In addition, the Emerson Norcross maximum offers an explanation for alluvial valley fills above the Campbell level in the western (Sheyenne, Pembina, and Assiniboine) river valleys (Thorleifson, 1996). Agassiz overflow exited via the southern outlet during the Norcross level, forming the Norcross and Tintah shorelines (Fig. 1-3) (Thorleifson, 1996). Lake Agassiz stabilized at the Campbell level (between about 9.7 - 9.4 ka BP, Fig. 1-9) as the southern outlet was eroded to a resistant boulder armour. Erosion of the southern outlet may have been limited by shallow Precambrian bedrock, exposed downstream of the southern outlet (H. Thorleifson, 1996, pers. comm.)

The fact that the elevations of the Upper and Lower Campbell beaches diverge northward from the southern outlet suggests that an outlet such as the eastern Kaiashk outlet (Fig. 1-6) near Lake Nipigon temporarily operated during the Emerson. Thorleifson (1996) suggests the Kaiashk outlet opened by ice retreat for about two centuries, then closed when the ice readvanced to the Sioux Lookout moraine (Fig. 1-4). Lake Agassiz rose to the Lower Campbell level due in part to the reduction of Agassiz overflow by differential uplift that had occurred between the Upper Campbell level and the Lower Campbell level. By 9.5 ka BP (or shortly thereafter), based on several dates between 9.0 and 9.4 ka BP along the north shore of the lake, the eastern outlets of Lake Agassiz reopened and Agassiz overflow was again routed to the Superior Basin (Zoltai, 1965; Elson, 1967; Mahnic and Teller, 1985).

1.2.6 - The Northwestern Outlet Hypothesis

1.2.6.1 - Introduction

Although not a new idea (see Upham, 1895; Elson, 1967, 1983; Teller and Thorleifson, 1983; Klassen, 1989) the northwest outlet has recently become a crucial topic in understanding the drainage history of Lake Agassiz. While Christiansen (1979) and Schreiner (1983) concluded that the lack of correlatable Agassiz offshore sediments and beach strandlines likely precluded a Lake Agassiz connection to the northwestern outlet, recent work (Fisher, 1993; Smith and Fisher, 1993; Fisher and Smith, 1994) indicates that Lake Agassiz outflow exited through the Clearwater River spillway at approximately 10 ka BP. In fact, Smith and Fisher (1993) advocated an entire Lake Agassiz watershed spillover out the Clearwater channel in northwestern Saskatchewan at 9.9 ka BP.

1.2.6.2 - The Northwestern Outlet Hypothesis: Discussion

At the onset of the Emerson Phase, after the eastern outlets were ice covered by the Marquette readvance, Lake Agassiz rose from the Moorhead low to the Emerson high water level. It has been traditionally accepted (see Section 1.2.3) that the stable Emerson highwater phase corresponds to the Campbell level (Teller and Thorleifson, 1983), which is projected to 438 m elevation in the northwestern arm of Lake Agassiz (Fisher and Smith, 1994). Fisher and Smith (1994) believe Lake Agassiz transgressed to 490 m in the northwestern arm of the Agassiz basin at the time of the Marquette readvance (Fig. 1-12), with the eastern outlets closed and the southern outlet not eroded lower than a

Herman level. A Lake Agassiz transgression to the projected Norcross level at 490 m overtopped the Beaver River moraine (Fig. 1-12), located southwest of the head of the northwest outlet (Clearwater Lower Athabasca Spillway - CLAS; Fig. 1-12). This allowed the drainage of Lake Agassiz out of the Churchill River basin into the Clearwater River and the Mackenzie basin. The coincident incision at the head of the Clearwater spillway lowered Lake Agassiz levels from 490m to 438m. Fisher and Smith (1994) believed that the final stable Emerson water plane was at 438 m elevation (the Upper Campbell level). They contend that after incision of the Clearwater spillway, Lake Agassiz continued to discharge through the northwest outlet until overflow could occur through the eastern outlets at the end of the Emerson Phase (see Section 1.2.4); this occurred after the northwest outlet was eroded to the Upper Campbell level.

While Fisher and Smith (1994) had distinct lines of evidence supporting their idea, if the ice sheet margin had not yet retreated far enough north to allow Agassiz to reach the Clearwater spillway, or if the southern outlet was eroded lower in elevation than the mouth of the Clearwater (perhaps during the Emerson highstand), no contribution to the Clearwater Lower Athabasca Spillway (CLAS) from the main body of Lake Agassiz could occur (Teller, 1990a). Fisher (1993) and Fisher and Smith

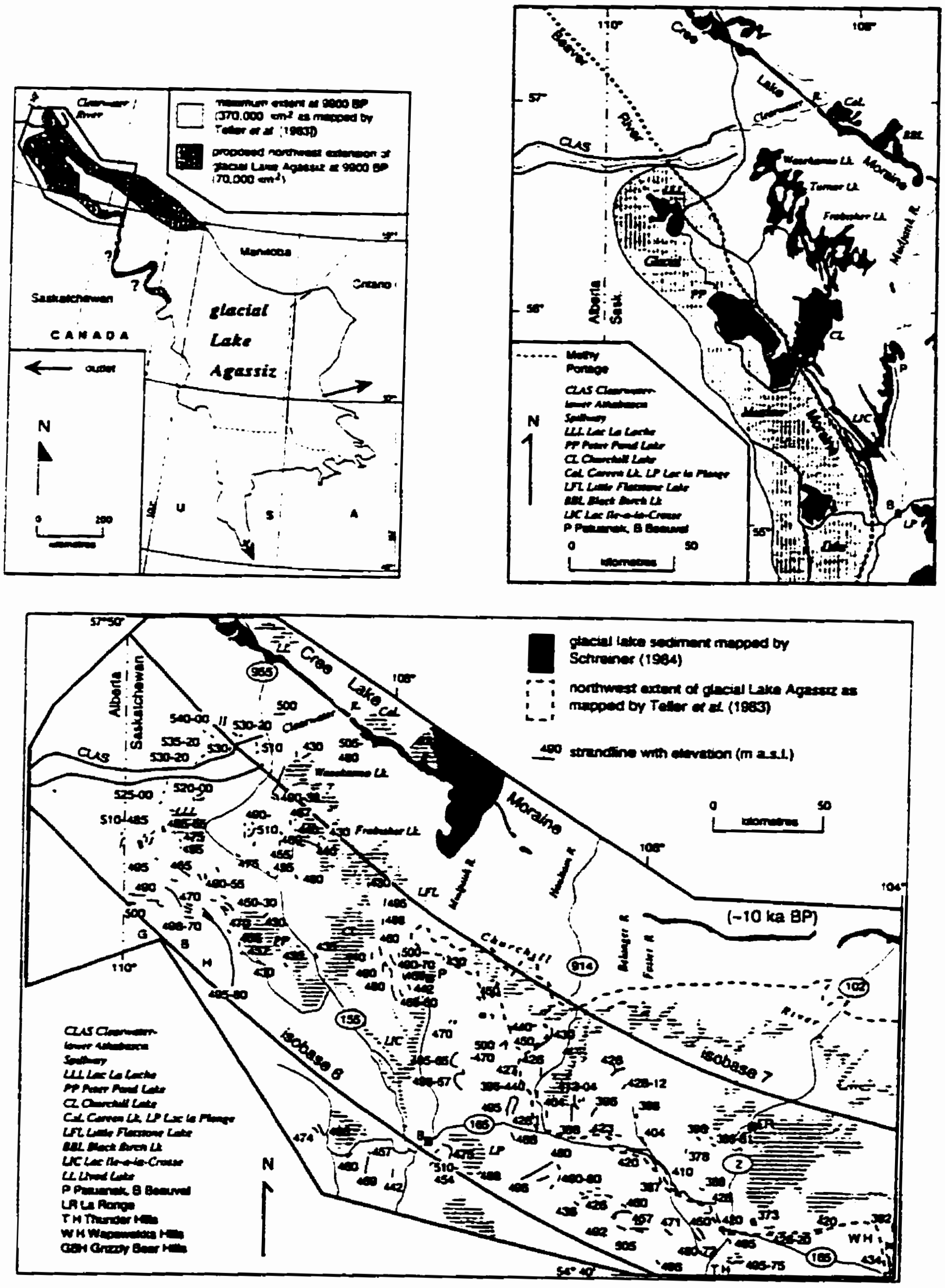


Figure 1-12: Locations in the northwestern arm of the Lake Agassiz basin and the Churchill River valley. The box located in the northwestern arm of the Lake Agassiz basin in the diagram at top left shows the expanded area shown in the large diagram at the bottom of the page. (Fisher and Smith, 1993, pgs. 845, 847, & 848, figs. 1, 3 & 4)

(1994) support the presence of Lake Agassiz in this northwestern extension by paleowater plane reconstruction and radiocarbon dates within the CLAS. Fisher (1993) and Fisher and Smith (1994) argue that because 1) the Emerson highstand of Lake Agassiz occurred at 9.9 ka BP, 2) radiocarbon dates in the CLAS date its incision at 9.9 ka BP, 3) geomorphic and sedimentological evidence in the upper Churchill River watershed suggests a transgression of a lake from the east, the waterbody that exited the CLAS must have been Lake Agassiz.

In the northwestern arm of the Lake Agassiz basin (Fig. 1-12), there are no continuous strandlines that are clearly associated with Lake Agassiz that extend to the head of the CLAS. Norcross strandlines, argued by Fisher and Smith (1994) to represent the water plane that connected Lake Agassiz to the northwestern outlet, have not been clearly correlated beyond Swan River, Manitoba, a locality several hundred kilometers southeast of the northwestern outlet. Closer to the northwest outlet than this, strandlines at elevations near the Norcross level are sporadic and discontinuous, and may be related to relatively small glacial lakes trapped between the Manitoba Escarpment and the Laurentide Ice Sheet, rather than to Lake Agassiz. Examples of this include Lake Swan (Nielsen, 1988) and Lake Saskatchewan (Christiansen, 1979; Christiansen *et al.*, 1995).

While glacial Meadow Lake probably did exist as a separate proglacial lake in the Upper Churchill River valley, and considering it did drain through the CLAS spillway (Christiansen, 1979; Schreiner, 1983; Fisher and Smith, 1994), drainage of that lake could not have been responsible for the formation of the mouth of the CLAS. The ice sheet was at the Beaver River moraine (50 km south of the spillway mouth) at the time Meadow Lake existed, thereby covering the mouth of the CLAS with ice (Fig. 1-12). Therefore, Meadow Lake could not have been responsible for the morphology of the northwest outlet spillway mouth, nor could it have incised the spillway from 490 to 438 m elevation. Thorleifson (1996) alternatively stated that occupation of the northwest outlet by Lake Agassiz occurred earlier than the Emerson Phase. His interpretation suggests that northwest outlet Lake Agassiz overflow maintained the Moorhead low water phase. Water would have exited the Clearwater starting sometime late in the Moorhead, ending at about 10 ka BP.

Because the landforms and sediments used in the interpretation of Fisher and Smith (1994) were not directly radiocarbon dated, nor were they correlated directly to the main Lake Agassiz basin, Fisher and Souch (1998) suggest that more information be collected to help prove whether a transgression of Lake Agassiz at 9.9 ka BP was responsible for the formation of the mouth of the CLAS spillway. Fisher and Souch (1998) speculate that the Emerson Phase maximum level of Lake Agassiz was the Upper Campbell level, and that this level was occupied twice during the Emerson Phase. Ice readvance closing the eastern outlets at 9.9 ka BP increased the level of Lake Agassiz to a first occupation of the Upper Campbell level. Emerson Phase lake level in the Churchill River Valley (Fig. 1-12) was controlled by a bedrock sill about 100 km southeast of the mouth of the CLAS (Fisher and Souch, in press), rather than the bedrock elevation between Wasekamio Lake and the CLAS (Fig. 1-12) as previously thought (Fisher and

Smith, 1994). After initial northwestern outlet erosion, rebound of the northwest outlet region occurring at a rate faster than outlet and sill erosion, would have forced Lake Agassiz waters south, eventually stabilizing at the Upper Campbell level for a second time. Fisher and Souch (1998) argue that this second southward transgression to the Upper Campbell level explains the young dates (~ 9.3 - 9.4 ka BP, e.g. Mann *et al.*, 1997; Risberg *et al.*, 1995) associated with the Upper Campbell strandline. The opening of the eastern outlets at the end of the Emerson Phase would have dropped Lake Agassiz levels below the Upper Campbell.

1.3 - OBJECTIVES:

The main goals of this thesis are, through limnological and geomorphological analyses of several paleolagoons located behind the Upper Campbell strandline, 1) to obtain radiocarbon dates that constrain the age of the Upper Campbell beach; 2) to provide a stratigraphic record of sedimentation from Lake Agassiz through the Holocene; 3) to develop a site specific paleoenvironmental history for each site (using ostracodes), from its inception in Lake Agassiz through the Holocene; 4) to correlate sediment facies and biofacies between Upper Campbell paleolagoon sites, spaced many tens of kilometers apart; 5) to evaluate the physical correlation of the Upper Campbell strandline; and 6) to assemble all the information amassed in steps 1-5 and apply it to the deglaciation history of the western Lake Agassiz basin during the Emerson Phase.

CHAPTER 2 - METHODOLOGY

2.1 - SITE SELECTION

In order to assess the age of the Upper Campbell beach and its relationship to the changing water levels of Lake Agassiz, a suitable basin containing datable material and ecologically useful material is needed. Ideally, these sites should be located as close to the Upper Campbell beach as possible. Thus, sites were chosen that were located immediately landward of the Upper Campbell strandline, in low areas that had at one time been lagoons of Lake Agassiz. If a record of the Holocene was to be found, these paleolagoons still had to contain water today, although one can never be entirely sure that lacustrine sedimentation was continuous, especially given that mid-Holocene (Hypsithermal) conditions are known to have been relatively warm and dry. Desiccation commonly results in the destruction of organic matter by oxidation and leaching.

Along the western side of the Lake Agassiz basin, there are a number of paleolagoon sites behind the ~1000 kilometers of discontinuous Upper and Lower Campbell strandlines (Fig. 2-1) from the Dauphin area in Manitoba to north-central Saskatchewan. In this area, the Upper and Lower Campbell strandlines vary from sandy and gravelly beaches and erosional scarps to beaches that separated the main body of Lake Agassiz from the land with a lagoon. In general, discontinuities in the strandlines were usually caused by 1) the presence

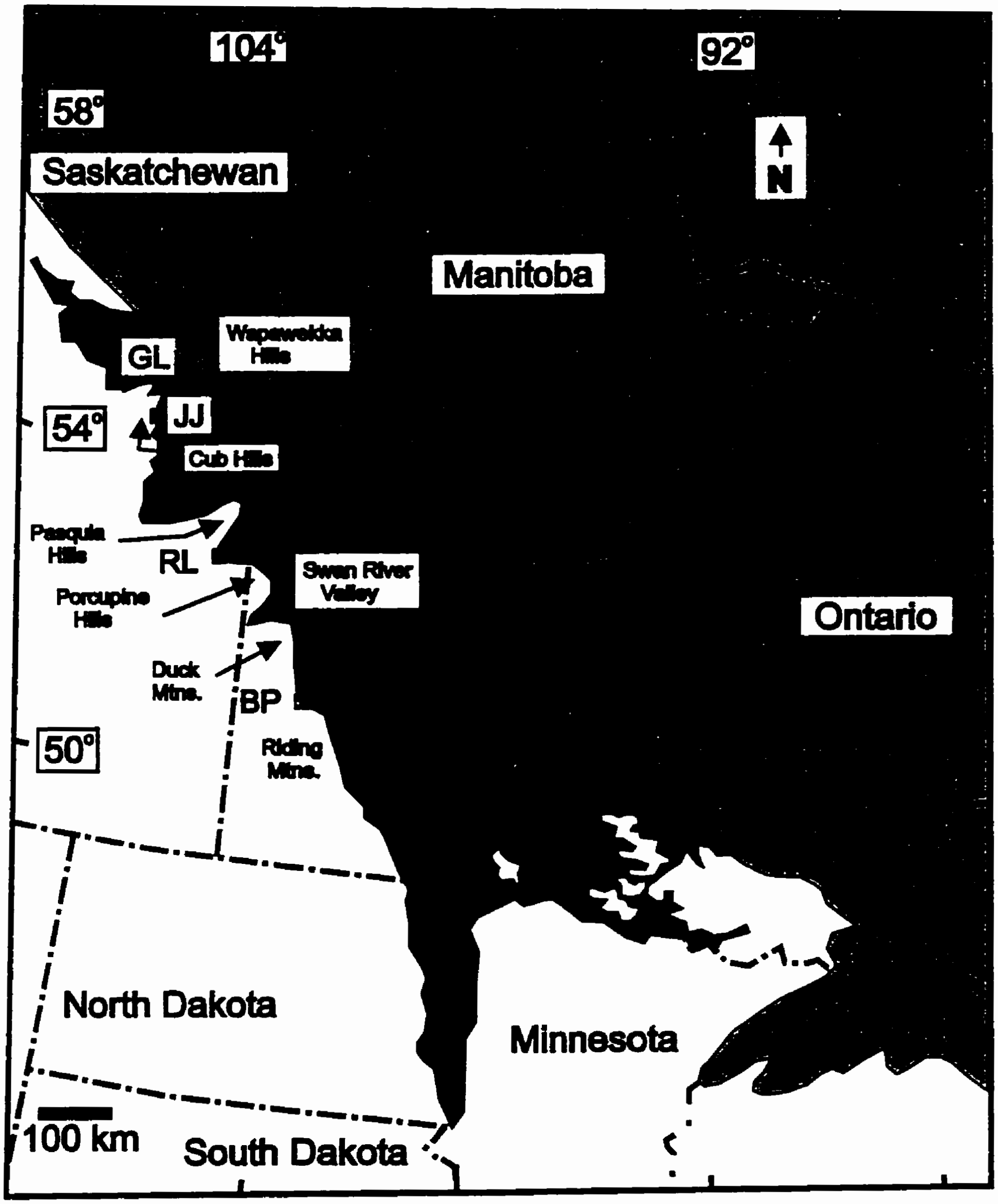


Figure 2-1: Location of new Upper Campbell lagoonal sites studied in this thesis. BP = Brokenpipe Lake, RL = Ruby Lake, JJ = Jay Jay Lake, GL = Gregory Lake. The Riding and Duck Mountains, with the Manitoban Porcupine Hills, make up the Manitoba Escarpment. Upper Campbell Lake Agassiz extent (dark shading) and Laurentide Ice Sheet (light shading) after Teller (1985, 1987).

of a glacier ice margin at the edge of the lake that did not allow the beach to be formed; 2) slumping of underlying unstable materials on steep slopes (e.g. Cretaceous shales along the Manitoba Escarpment -- Fig 2-1; Nielsen and Watson, 1985); 3) dissection by rivers and streams; 4) mining of the beaches in borrow pits and the construction of roads and villages; and 5) insufficient sediment supply for beach-building or erosion-resistant material.

The first stage in site selection was to map the Upper Campbell beach using air photos and previously published soils and geologic maps. This information was transferred to 1:250,000 topographic map sheets, which served as index sheets for the entire study area. Small ponds and lakes trapped behind the Upper Campbell strandline were identified from photos and maps. These sites were visited during the summer field season of 1995. Many (~10) sites were examined and hand cored to determine water depths, minimum lacustrine sediment thickness, and general sediment types. Localities deemed as good possibilities (those that appeared to have reasonable water depths and sediment records) were cored in February 1996 and 1997. Summer fieldwork in 1996 also involved three-dimensional (latitude, longitude, and elevation) surveying of the Upper and Lower Campbell beaches with a high resolution Global Positioning System (GPS) survey (Rayburn, 1997; Rayburn *et al.*, 1997). This was the final factor used to ensure that the sites studied were, in fact, located behind the Upper Campbell strandline, and not some other level (see Fig. 1-3). Four paleolagoonal lake sites (Fig. 2-1) were chosen for detailed analyses (the Brokenpipe Lake, Ruby Lake, Jay Jay Lake, and Gregory Lake sites), while the Hudson Bay Upper and Lower sites (see later sections on Ruby Lake, Chapter 4) were ranked as secondary sites.

2.2 - LAKE CORE COLLECTION

The coring system used for this project was a vibracorer constructed specifically for this study. Coring was conducted from a winter ice platform through a hole in the ice. The equipment design was similar (with some minor modifications) to that of Smith (1992) and Thompson *et al.* (1991). The heart of the vibracoring system (Fig. 2-2) is an eccentrically weighted elongate shaft that rotates at high RPM within a casing that is attached orthogonally to a 3-inch diameter, thin wall aluminum irrigation pipe core tube by a heavy steel clamp. The eccentric weight is driven by a flexible shaft connected to a 5.5 hp gasoline engine. The rotation of the weight in the vibrahead induces a vibration in the coretube that liquefies saturated unconsolidated sediments, thereby allowing the coretube to penetrate downward through the sediments. Core is held in the core tube by brass teeth in a steel bit which allow core to enter, but not extrude from the core tube (Fig. 2-3). Special clamps (Fig. 2-4) allow multiple sections of 3.05 m-long aluminum core tubes (thin-walled irrigation pipe) to be attached together. Penetration was inhibited or stopped by the presence of low-water content sediments (e.g. hard, well-compacted glaciolacustrine clays), and by hard, coarse, or fibrous impenetrable materials (e.g. roots, rooty peat, bedrock, soil hardpan, rocks or cobbles larger than the diameter of the core tube).

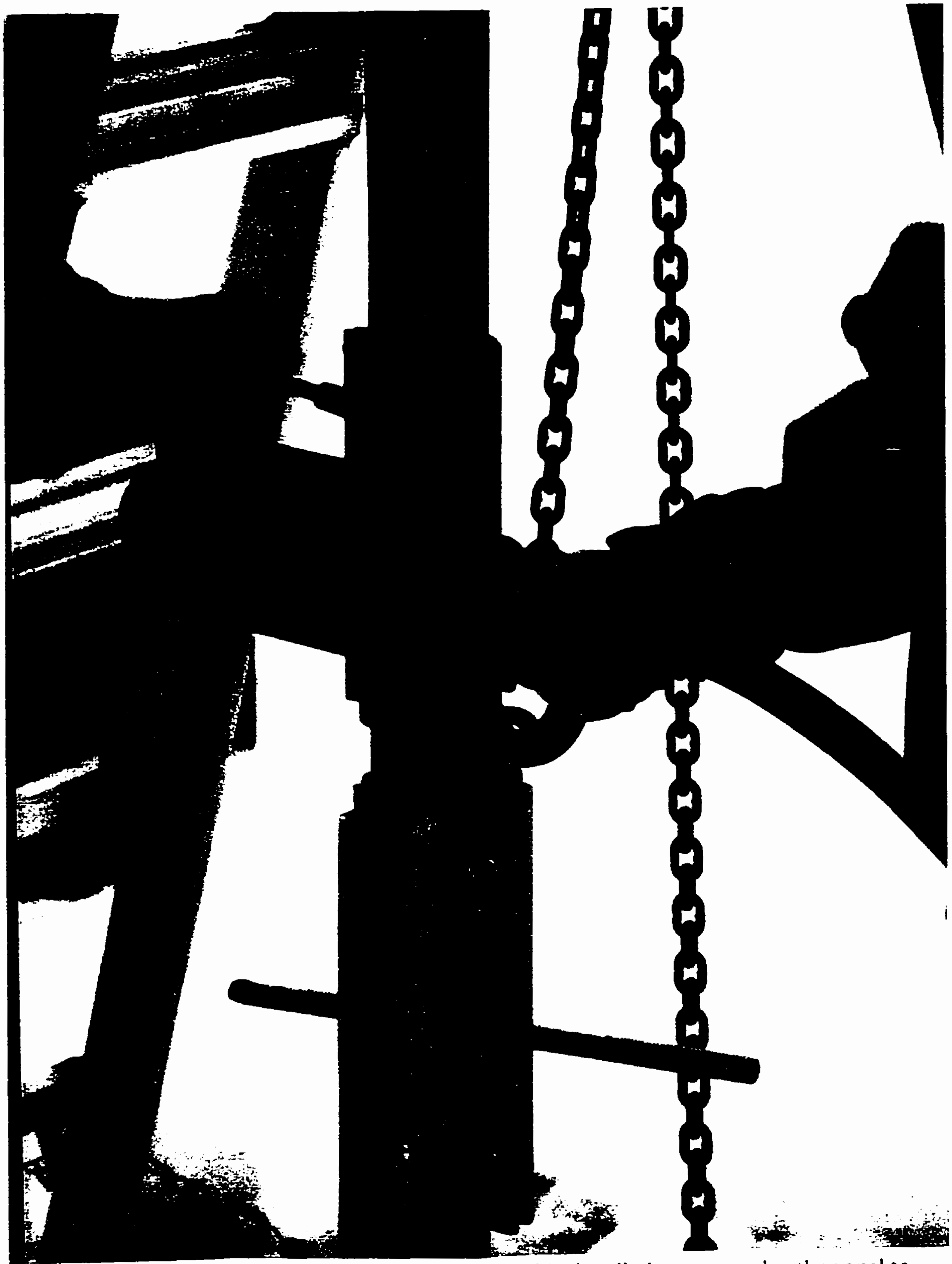


Figure 2-2: Photo showing the vibrahead (large black cylinder arranged orthogonal to the aluminum core tube) and flexshaft (right of the vibrahead, running out of the photo to the 5 hp gas motor – not shown) clamped to the aluminum core tube. The aluminum clamp below the vibrahead with the chain hoist attached to it is the hoist clamp.

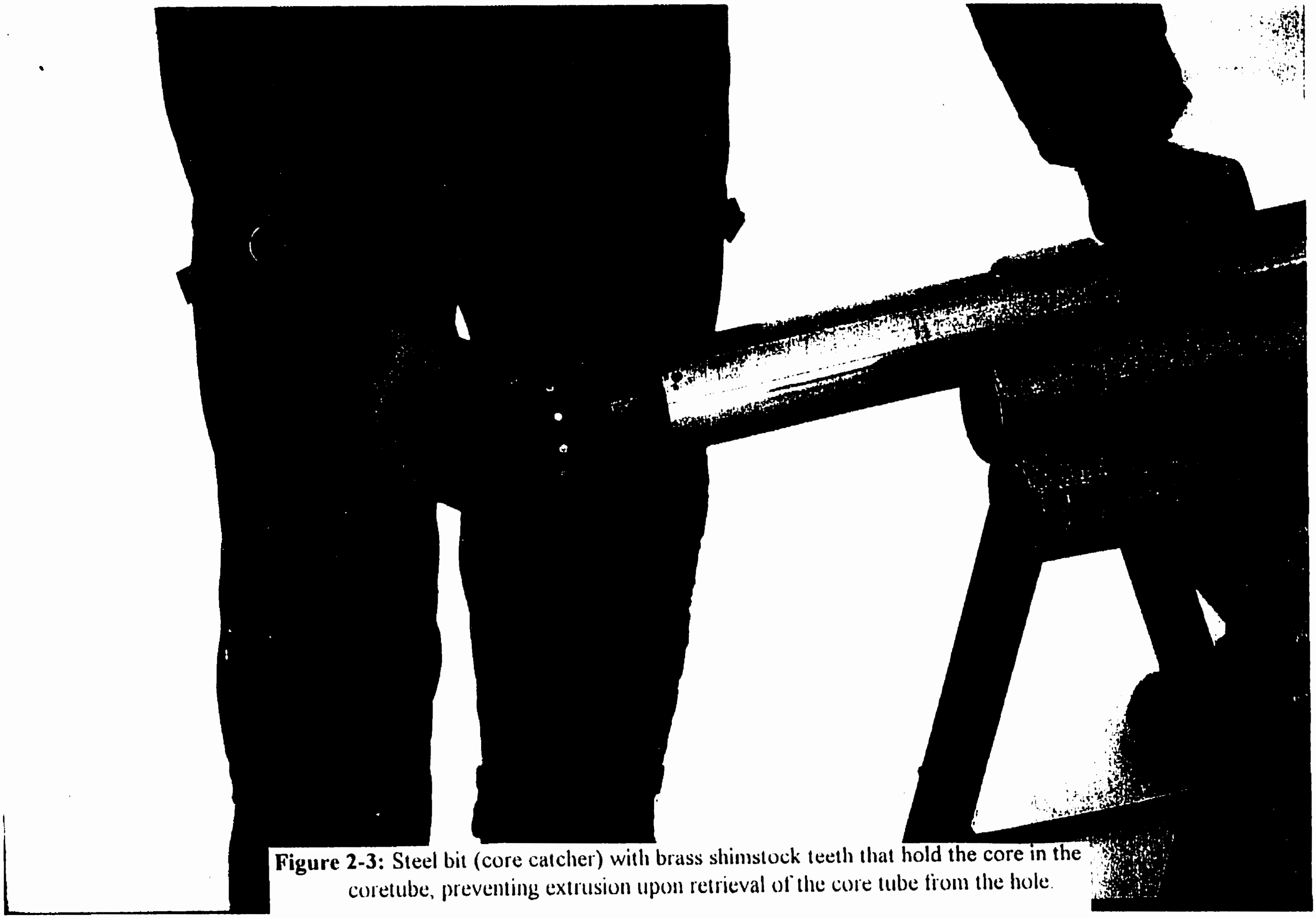


Figure 2-3: Steel bit (core catcher) with brass shimstock teeth that hold the core in the coretube, preventing extrusion upon retrieval of the core tube from the hole.

Figure 2-4: Aluminum core tube clamps that allow sections of 3.05 m long core tube to be joined together into a continuous coring stem.

The core tube was retrieved by a ratchet chain hoist and an ice-top platform made from a modified, adjustable welded aluminum ladder (Fig. 2-5).

Cores were taken from Brokenpipe Lake, Ruby Lake (and smaller ponds in the Ruby Lake area, see Chapter 4), Jay Jay Lake, and Gregory Lake; the coring locations within each lake are shown in Figure 2-6. Throughout the thesis, reference will be made to specific lakes, as well as the sampling and coring sites within individual lakes. Table 2-A summarizes the sampling activities at all the lakes in this study, including coring sites, water sampling sites, modern ostracode sampling sites, and field (Hydrolab) water testing sites.

2.3 - SAMPLE PREPARATION

2.3.1 - Core Subsampling and Description

The systematic description of the sediment cores, and the values for percent moisture, organics, and carbonate of the cores, are listed in Appendices A and B respectively. Moist samples weighed about 7 - 15 g. The sample masses varied based on the type of material and its water content, and was partly dictated by the amount of material that remained after sampling the same

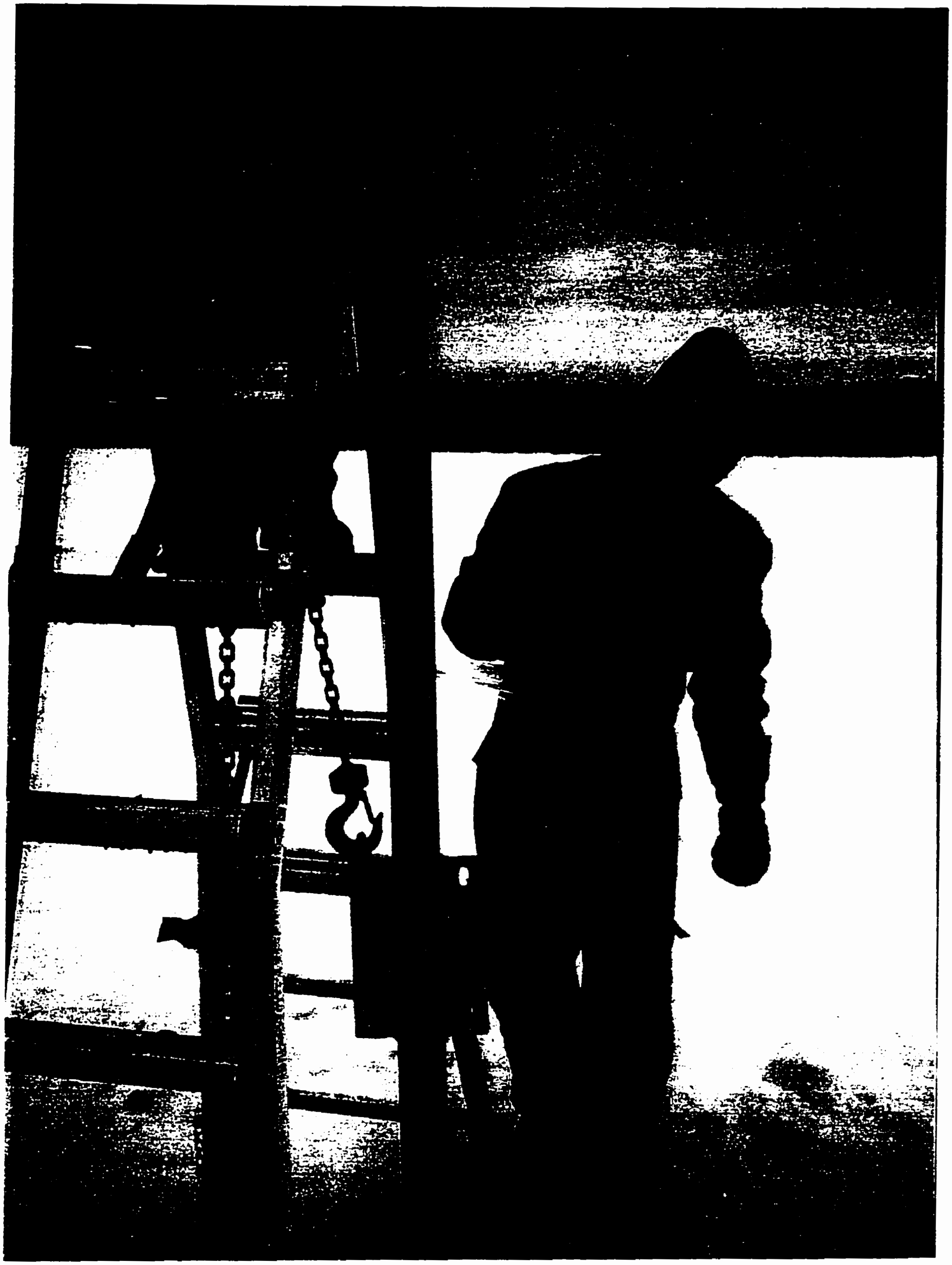


Figure 2-5: Aluminum ladder and chain hoist used to retrieve the coring stem from the hole.

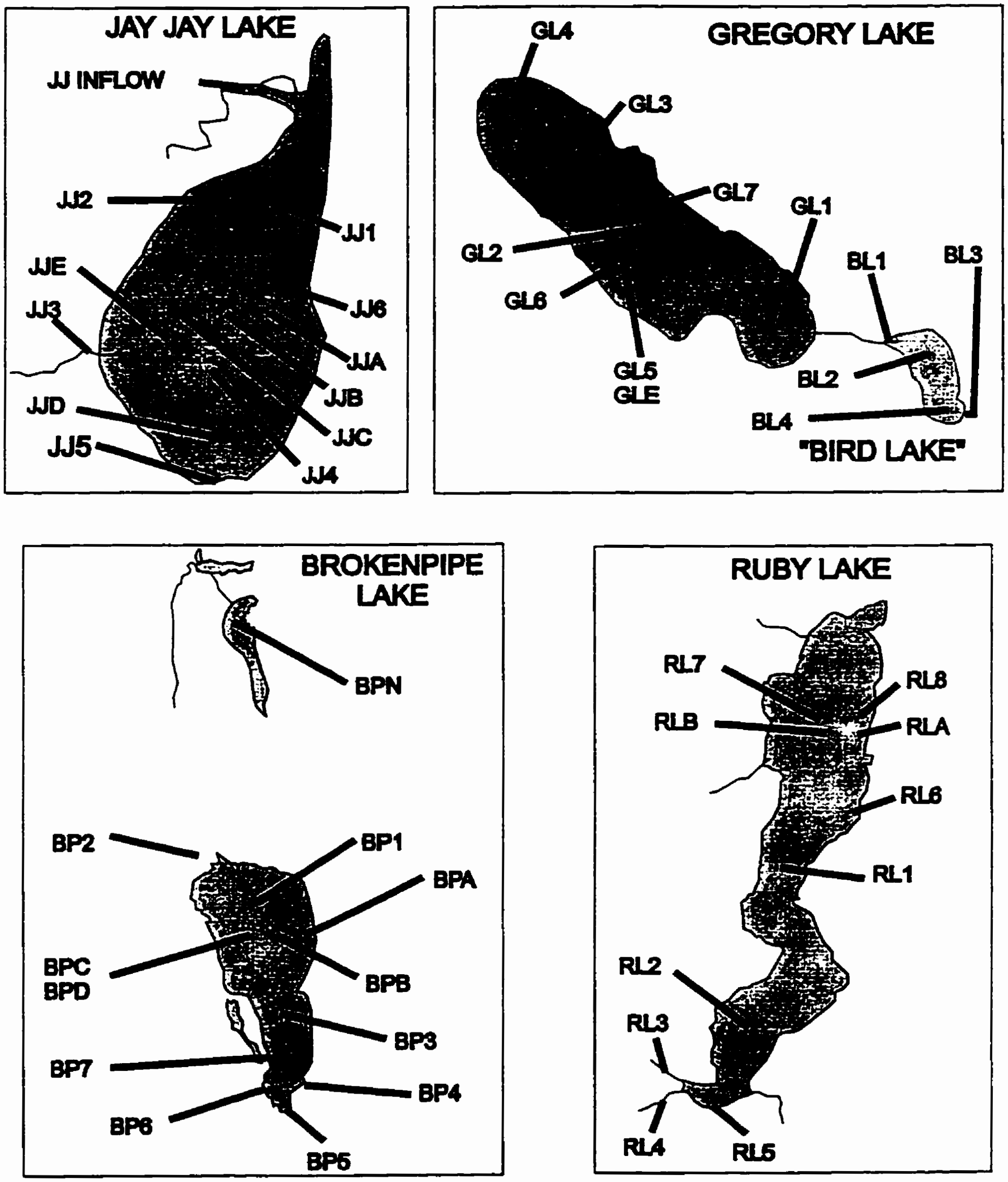


Figure 2-6: Sampling sites in the four major lakes of this study. Sampling activities include water and ostracode sampling, hand coring (probing), and vibracoring. See Table 2-A for details, and Figure 2-1 for the locations of these lakes along the western margin of the Lake Agassiz basin. Scale ~ 1:50 000

Table 2-A: Summary of Sampling Activities At All Lakesites

Lake	Site	Activity				
		Probe	Core	Water Sample	Ostracode Sample	Hydrolab Data
Brokenpipe	BPA		X			
	BPB		X			
	BPC		X			
	BPD		X			
	BPN		X			
	BP1			X	X	X
	BP2				X	X
	BP3			X	X	X
	BP4					X
	BP5				X	X
	BP6					X
	BP7				X	X
Ruby	RLA		X			
	RLB		X			
	RL1			X	X	X
	RL2					X
	RL3					X
	RL4					X
	RL5				X	
	RL6				X	X
	RL7			X	X	X
	RL8				X	X
Hudson Bay	HBU	X	X			
	HBL	X	X			

Table 2-A Continued	Core	Probe	Core	Water Sample	Ostracode Sample	Hydrolab Data
Jay Jay	JJA		X			
	JJB		X			
	JJC		X			
	JJD		X			
	JJE		X			
	JJ1			X	X	X
	JJ2	X			X	X
	JJ3	X				X
	JJ4			X	X	X
	JJ5				X	
	JJ6				X	
	JJ Inflow					X
Bird	BL1					X
	BL2			X	X	X
	BL3				X	X
	BL4			X	X	X
Gregory	GLE		X			
	GL1	X				
	GL2	X				
	GL3	X				
	GL4	X				
	GL5	X				
	GL6	X				
	GL7	X				

interval for fossil ostracode analysis. Percent moisture was calculated by determining the mass of water (in g) that evaporated in an oven at 90 °C (samples were dried for at least 12 hours) from a pre-weighed sample that was evaporated to dryness. Compaction due to the vibracoring process probably altered moisture content of the cores, especially the sandier sediments. Percent organic material was determined by the Loss On Ignition (LOI) technique(Dean, 1974); pre-weighed dry samples were placed in crucibles in a muffle furnace and heated for a minimum of 1.5 hours at 490 - 500 °C. The weight difference (in g) is assumed to be the weight of organic material present in that sample. For percent carbonate, the same samples in crucibles were raised from 500 °C to 1000 °C in the muffle furnace; temperatures were kept between 900 and 1000 °C for at least 1 hour (Last, 1980). The remains of these samples consisted entirely of non-carbonate ash. In some cases, samples combusted to determine LOI were treated with multiple applications of 10% hydrochloric acid until no reaction was visible; dried samples were then re-weighed to determine carbonate content.

2.3.2 - Modern Water Chemistry and Modern Ostracodes

The modern water chemistry and modern ostracode content of the study lakes were investigated in order to obtain a baseline of lake water chemistry and ostracode fauna to which the fossil ostracode fauna and interpreted lake chemistry could be compared. In this way, inferences about what controlled the hydrochemistry in Lake Agassiz lagoons can be made. Two water samples were taken from each of the four major lake sites, in order to determine the modern lake chemistry. Modern ostracodes were also collected with a brine shrimp net from various sites in each of the four major lake sites. To collect living ostracodes, the net was dragged along the surface of the sediment-water interface, and through the aquatic vegetation. These samples were put into labelled jars for subsequent lab treatment and identification to species (see later section).

In the case of Gregory Lake, where summertime accessibility was a problem, water and modern ostracode samples were collected from a nearby smaller lake (herein called "Bird Lake", Fig. 2-6) that lies south and directly downstream of Gregory Lake. Lake water samples were collected following standard guidelines outlined in Hem (1989) and Mackereth *et al.* (1978). A discrete spot-sampling strategy was employed and considered adequate to characterize water quality at the time when the sample was taken (Hem, 1989). Water was sampled from the epilimnion, away from the littoral zone and any obvious riverine input. Field water temperature, pH, and conductivity measurements were taken with a calibrated Hydrolab Digital 4041 unit. The unit was calibrated during and after fieldwork. Nalgene water collection jars (500 and 250 ml sizes) were acid-cleaned and rinsed with de-ionized water four times. The jars were taken into the field filled with de-ionized water. One unfiltered sample was used for determination of alkalinity in the laboratory. The other sample (used for major ion determination) was filtered in the field through Whatman GF/C glass filter paper. The samples were kept on ice (i.e. at 4°C) in a closed, dark cooler until they were put into storage by the Department of Fisheries and Oceans (DFO) at the University of Manitoba. The DFO lab used atomic absorption analysis to determine concentrations of Na, K, Ca, Mg, Cl, SO_4, HCO_3, CO_3, and dissolved CO_2.

2.3.3 - Fossil Ostracode Analyses

One centimeter (1 cm) thick subsamples were taken for ostracodes from a split of each core, at a maximum subsample spacing of 10 - 15 cm. Finer subsample spacing (2 cm) was used in most cases. The other half of the core was archived in the Physical Sedimentology lab at the University of Manitoba. Standard calcareous microfossil subsample processing procedures were used (Forester, 1988). Processed samples were washed through a set of at least 2 sieves (60 and 100 mesh). In some cases, a 20 mesh sieve was utilized to remove the coarser fraction found in many samples, and a 230 mesh sieve was often used to retain the finer fraction of the sample, and to check for any ostracode sieve loss. The sieved fraction was either air dried, or freeze dried if it had an appreciable amount of organic matter within it. Freeze dried samples were disaggregated and easy to work with. After sieving and drying, samples were picked for adult ostracodes with a fine sable brush using a reflected light microscope. Adult ostracodes were either placed into vials or mounted on micropaleontological slides for identification, using the publications of Delorme (1967, 1970a, 1970b, 1970c, 1970d 1971), experience gained personally with Dr. Brandon Curry of the Illinois Geological Survey, and through personal communication with Dr. Curry.

CHAPTER 3 - MODERN WATER CHEMISTRY OF THE LAKE SITES AND ITS RELATION TO WATERSHED GEOLOGY

3.1 - INTRODUCTION

In order to establish a baseline of modern hydrochemistry to which inferred hydrochemical interpretations based on fossil ostracode assemblages could be compared, water samples were taken from the four main lake sites. This chapter describes the major ion content and concentrations of ions in each of the lakes investigated. The composition of the water chemistry at each lake site, as the Upper Campbell beach began to form, must have been the same as that of Lake Agassiz. As the lagoon became separated from the main Lake Agassiz basin by the beach, and as Lake Agassiz withdrew from the area because of declining water levels -- the lake hydrology and hydrochemistry became more sensitive to the interaction and balance between several controls: 1) the periodic, beach overwash of stormy Lake Agassiz waters; and later, after retreat of Lake Agassiz 2) the watershed geology, as chemical weathering increased after deglaciation; 3) groundwater inflow and outflow, and its composition; and 4) moisture balance (precipitation vs. evaporation). For example, the variable influence of groundwater in lake water chemistry evolution of a Lake Agassiz remnant, during the post-Lake Agassiz era, has been shown in the isotopic evolution of Lake Manitoba waters (Last *et al.*, 1994) and by the Lake Manitoba ostracode record (Curry, 1997).

3.2 - MODERN WATER CHEMISTRY OF THE LAKE SITES

3.2.1 - Description of Trends in Water Chemistry Between Lake Sites

Lake water chemistry includes dissolved ion composition, and solute concentration (salinity). Ostracodes are particularly sensitive to both of these factors, which can influence both faunal diversity and quantity. The proportions of solutes and their total concentrations are influenced by both climatic and non-climatic processes (Forester, 1987; Gorham, 1961). Figure 3-1, after Forester (1987), outlines the major solute sources. These include water - rock (weathering) reactions, dissolution of precipitated minerals, atmospheric precipitation, equilibria between atmospheric and dissolved carbon dioxide, and dissolution of atmospheric dust. Processes of solute concentration include evaporative concentration or mixing of waters with different salinities or compositions, biologic processes, carbon dioxide outgassing, exchange reactions, and outflow (Forester, 1987). Factors that control the variance of solute composition and concentration of lake waters over the include the composition of soils and rock in the watershed, precipitation and/or dissolution of minerals, moisture balance (the ratio of evaporation to precipitation), and anthropogenic activities (Curry, 1995). Diagrams showing the general bedrock configuration in the western Lake Agassiz basin and a stratigraphic representation of the major pre-Quaternary Phanerozoic units in Manitoba are shown in Figures 3-2 and 3-3, respectively. The relationship of bedrock geology to till composition, ultimately the surficial material chemically weathered to donate ions to the lake watershed, is probably also important.

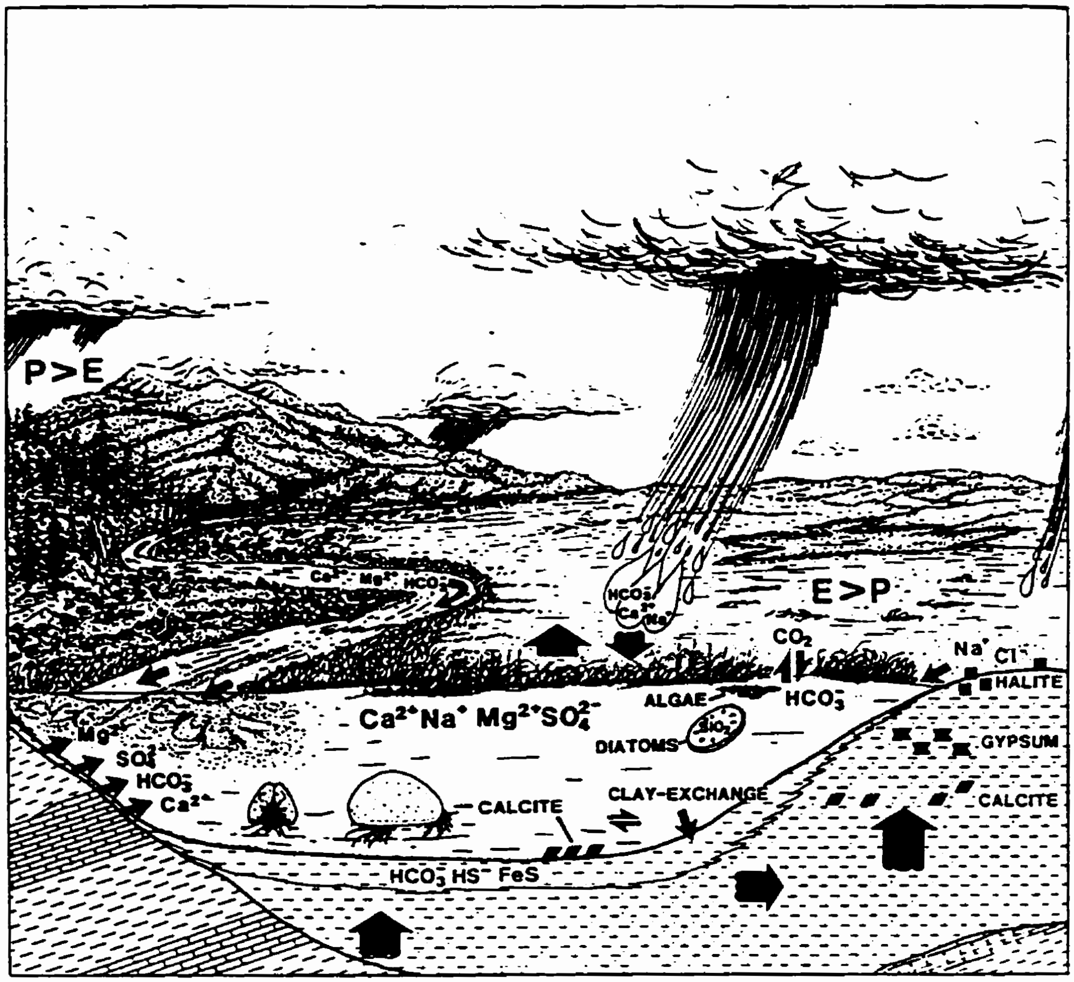


Figure 3-1: Generalized diagram showing common solute input and output sources for a saline lake having Ca-enriched saline water and no surface outlet. The water chemistry is a product of rock-water reactions (watershed geology), regional and local climate, and other factors. (Forester, 1987, p.264, fig. 4)

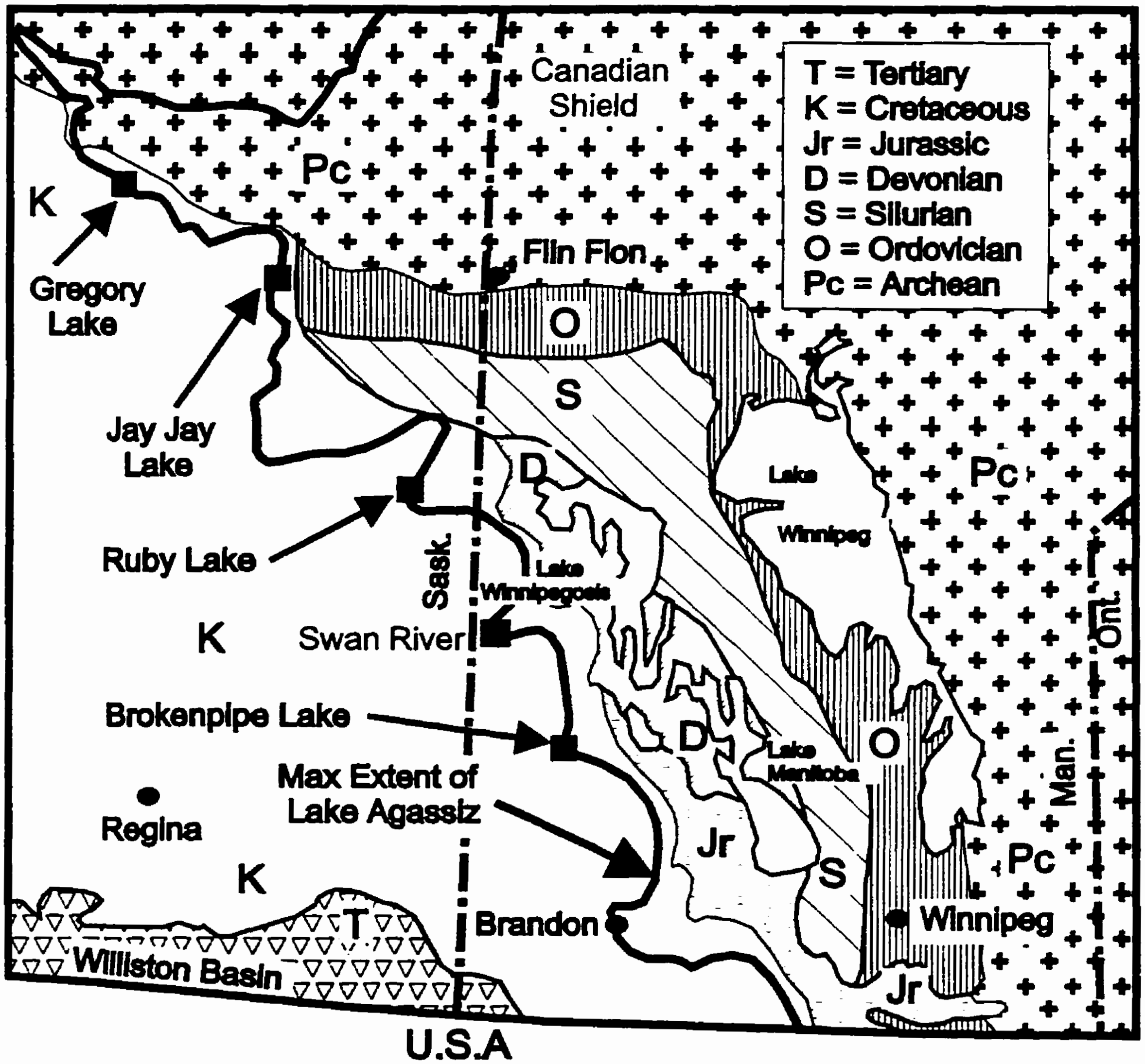


Figure 3-2: General bedrock configuration of the part of the Lake Agassiz basin discussed in detail in this thesis. See text for discussion (modified from Teller and Bluemle, 1983). Archean rocks are acid and basic crystallines and metamorphics, Ordovician rocks are predominantly dolomites, with shales and sandstones, Silurian rocks are dolomite, Devonian rocks are limestones, dolomites, potash, anhydrite, and shale, Jurassic rocks are shales, limestones, anhydrite, and gypsum, Cretaceous rocks are dominantly shales. Note that some lithologies (and their associated glacially derived sediments) are readily dissolved by weathering processes to supply ions to modern lake watersheds; these include gypsum, anhydrite, and calcareous rocks such as limestone.

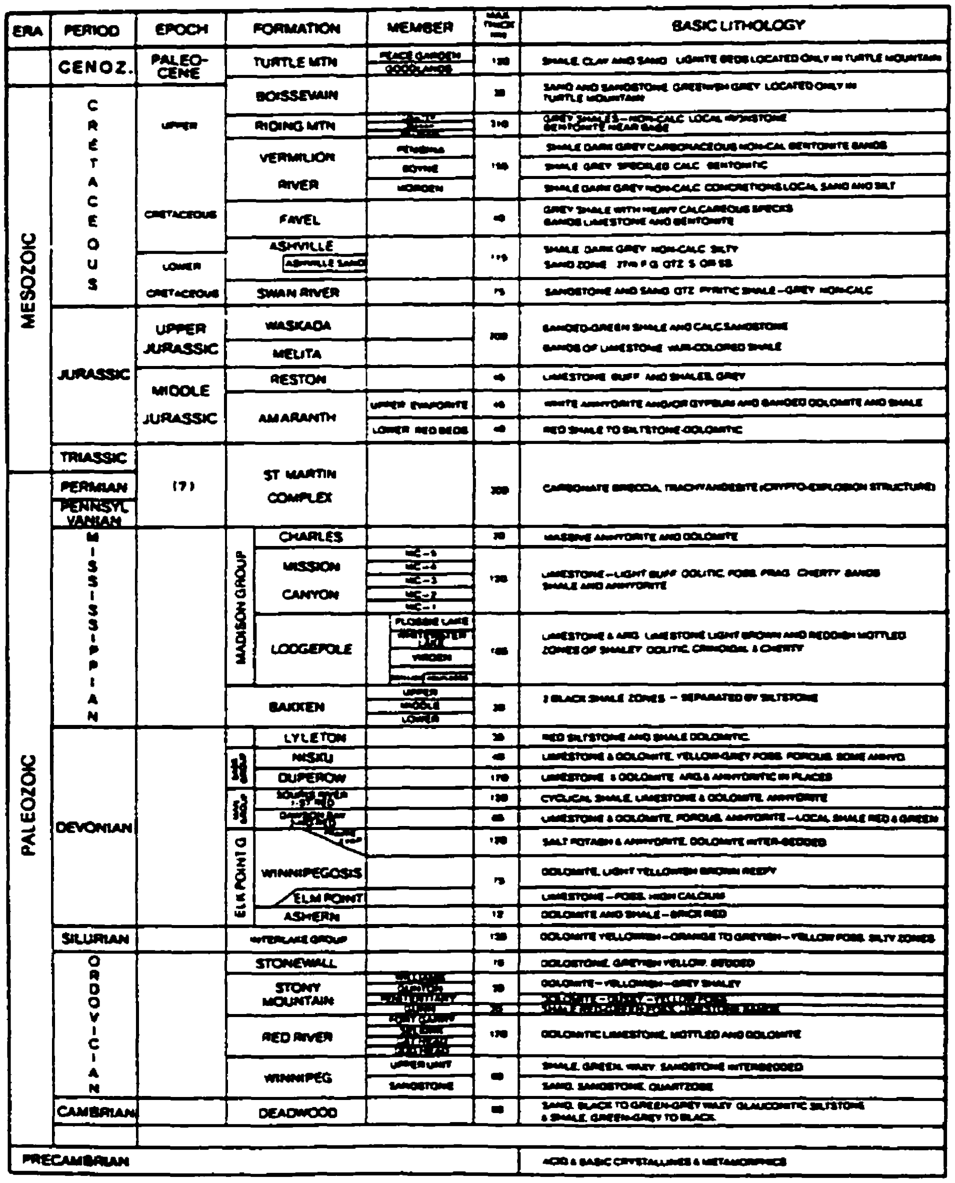


Figure 3-3: Major pre-Quaternary Phanerozoic units in Manitoba indicating the dominant lithology of each. (Teller and Bleumle, 1983, p. 13, fig. 4)

The major ions measured in this study include potassium, magnesium, chloride, sodium, calcium, and sulphate. The following description will also include alkalinity, conductivity, and pH. Alkalinity in this thesis is considered to be the sum of carbonate and bicarbonate, calculated (using standard techniques) from dissolved inorganic carbon (DIC) concentrations and pH. Tables of all the lab-determined and field-measured water chemistry values, and ion balances, are given in Appendix C.

a) Alkalinity, Conductivity, and pH:

Alkalinity is defined as the ability of water to react with H+ ions, and is a reaction controlled primarily by the concentration of carbonate and bicarbonate ions (Faure, 1991). The alkalinity of a lake is tied closely to the forms of carbon dioxide (CO_2) that it contains. Alkalinity, in this thesis, is defined as the sum of bicarbonate (HCO_3) and carbonate (CO_3), expressed in mg/l. While the relative composition of the different possible forms of carbon dioxide (HCO_3, CO_3, CO_2+H_2CO_3) is dependant on pH, total alkalinity is not necessarily dependant on the pH scale. Water on the acid side of the pH scale can have high alkalinity values (Cole, 1983). Waters with high total alkalinity are especially resistant to changes in pH. These are referred to as "buffer solutions". Addition of a strong base causes a reaction with carbonic acid to form a bicarbonate salt (and eventually carbonate), effectively utilizing all of the base. If an acid is added, it is utilized in the conversion of carbonate to bicarbonate and of bicarbonate to undissociated carbonic acid (Cole, 1983). These relationships explain large pH changes in lakes with low alkalinity; after the addition of respiratory carbon dioxide or the removal of carbon dioxide due to photosynthesis (Cole, 1983).

Alkalinity was measured in the laboratory for all four lakes involved in this study (Appendix C). These laboratory values were not used because the balance check on the alkalinity determinations was poor. The laboratory results showed that the lake with the lowest conductivity (Bird Lake) also had the highest alkalinity. This relationship does not make sense, as the measurement of conductivity is dependant upon the amount of dissolved salts and ions in solution, including those (HCO_3 and CO_3) that are responsible for alkalinity. This discrepancy in the data was likely an artifact of the lengthy storage of the water samples before analysis at the DFO lab. Therefore, bicarbonate and carbonate concentrations were calculated using the methodology outlined by Mackereth *et al.*, (1978). This calculation method accurately approximates bicarbonate and carbonate values based on the dissociation constants applicable to the carbon dioxide-bicarbonate-carbonate-hydroxide buffer system (Mackereth *et al.*, 1978). Factors in this system, which affect the dissociation constants, include pH, temperature, and ionic strength of the solution. Each of these parameters is accounted for by the method. All that is needed for the calculations are the pH, temperature, and total dissolved inorganic carbon (DIC) content of the water.

Figures 3-4 A and 3-4 B show the concentrations of Total Dissolved Solids (TDS), alkalinity, bicarbonate, pH, carbonate, and free carbon dioxide for the four lake sites in this study. Not surprisingly, based on the observations of generally decreasing major ion concentrations of the lake sites from south to north in the study area (likely

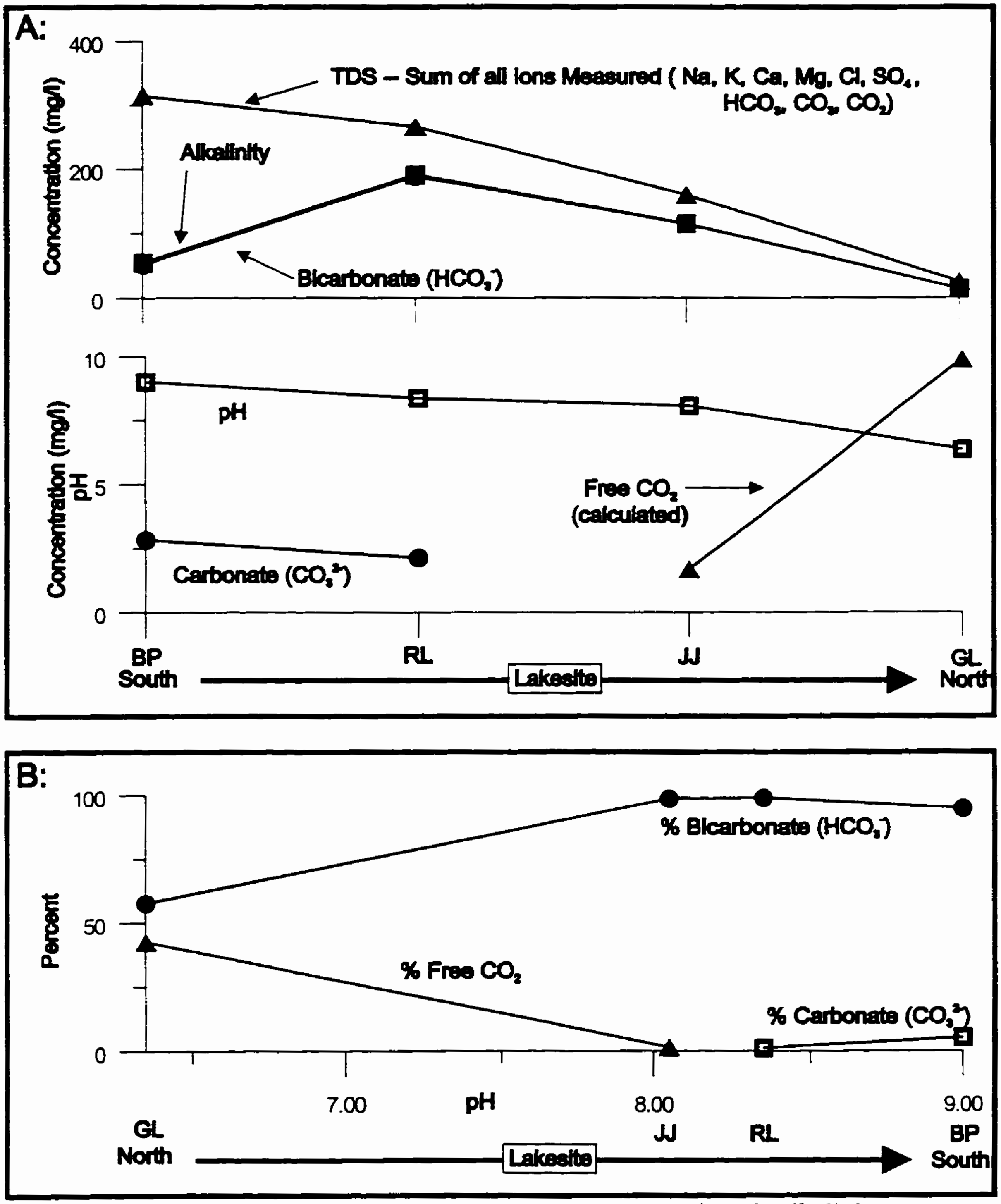


Figure 3-4 A and B: Figure A shows the concentrations of TDS, alkalinity, bicarbonate, carbonate, and free carbon dioxide, and average pH of the four main lakes studied in this thesis. Figure B shows the relative proportions for the three forms of carbon dioxide in relation to the pH of the lakes. BP = Brokenpipe, RL = Ruby Lake, JJ = Jay Jay, GL = Gregory Lake ("Bird" Lake). Note that alkalinity and bicarbonate curves overlap. See text for details.

reflecting a northward increase in the difficulty of chemical weathering ions from surficial materials into solution, ultimately a reflection of changes in surficial material composition with different bedrock provenance), TDS, alkalinity, and bicarbonate values generally show the same northward decreasing trend (Fig. 3-4 A). However, of note is the low alkalinity (and bicarbonate) values of the southernmost lake (Brokenpipe).

Alkalinity of lake water can usually give the researcher a good indication of the nature of the rocks (or geological materials) in the drainage basin and perhaps their state of weathering (Cole, 1983). In this sense, alkalinity often results from carbon dioxide and water dissolving sedimentary carbonate materials to form bicarbonate solutions. However, the relatively low alkalinity of Brokenpipe Lake is not a result of an absence of carbonate rock source material, nor is it a result of reduced weathering in the Brokenpipe Lake region. The low alkalinity is related to the intense carbon dioxide uptake and calcareous marl formation by a large population of *Chara sp.* algae that forms a thick mat over the entire bottom of the lake. Carbon dioxide uptake during photosynthesis eliminates bicarbonate from solution by precipitating carbonate, and results in increased pH of the waterbody (Cole, 1983). Figure 3-4 B shows the relative distribution of carbonate, bicarbonate, and free carbon dioxide of all the lake sites, in relation to lake pH. This data closely follows what is expected of the effect of pH on the dissociation of carbonic acid in lacustrine systems.

The conductivities of the waters of all the lake sites (Fig. 3-5) decrease from relatively higher values in the south (Brokenpipe Lake) to lower values in the north (Bird

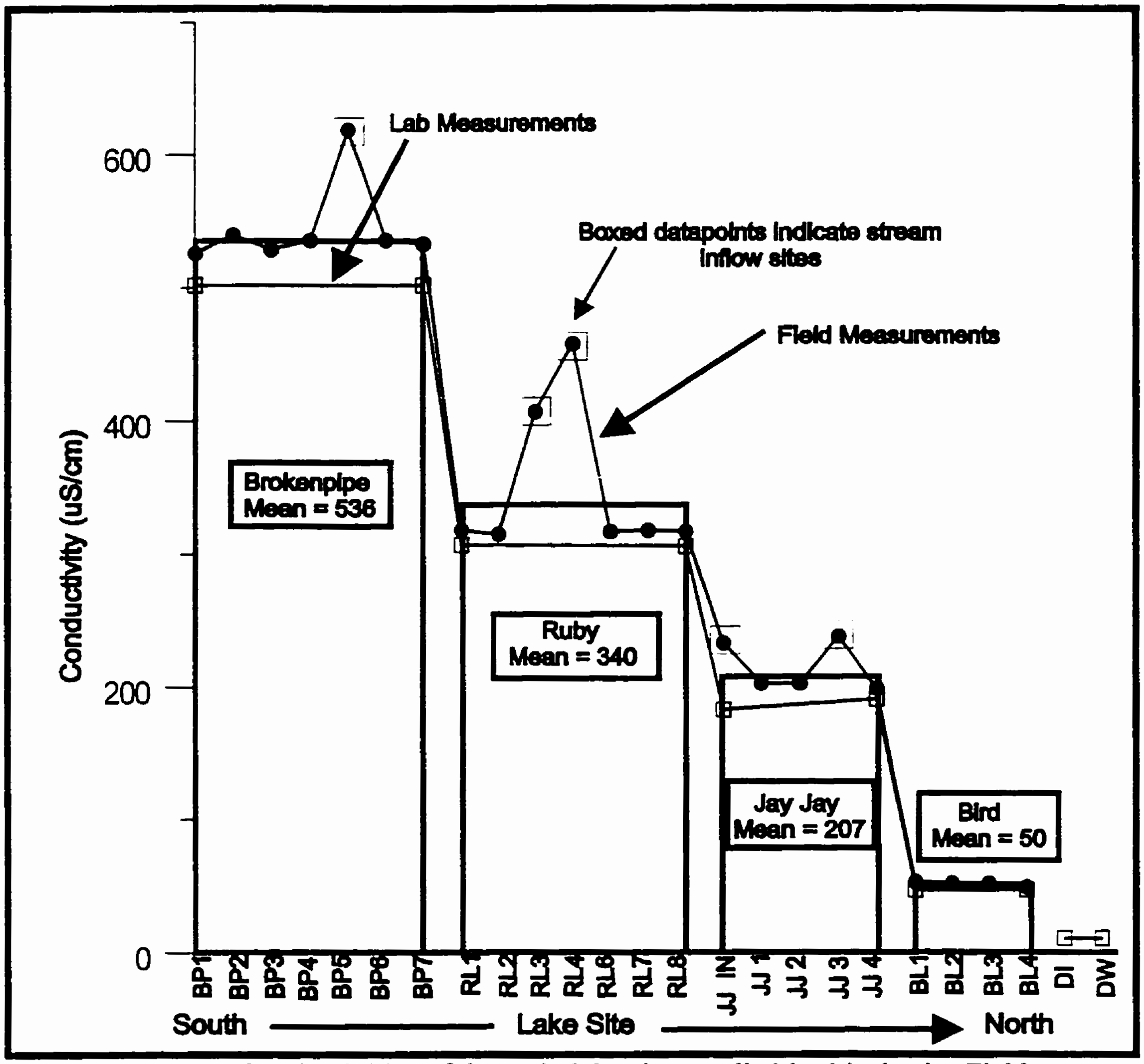


Figure 3-5: Conductivities of the main lake sites studied in this thesis. Field measurements taken with the Hydrolab unit, lab measurements were done by the DFO laboratory. Values in insets and bar heights represent averages. Lake sites correlate to those in Fig. 2-6, and Table 2-A. DI = lab deionized water sample, DW = lab distilled water sample. See text for details.

Lake). This is a reflection of the decreasing supply of easily weathered ions in surficial materials as the lake sites approach the Canadian Shield. Tills derived from the Shield have distinctly different lithologic and textural characteristics that do not promote effective chemical weathering. The high conductivity of Brokenpipe Lake is a reflection of its high concentration of sulphate (not because of relatively high alkalinity). Alternatively, Ruby and Jay Jay Lakes have relatively high conductivities due to their relatively high concentrations of bicarbonate, derived from weathering of the surrounding tills. Stream inflow to Brokenpipe, Ruby, and Jay Jay Lakes is relatively more conductive (saline) than the main lake waterbody (Fig. 3-5).

The decreasing pH of the lake sites (Fig. 3-6) from south (Brokenpipe Lake) to north (Bird Lake) in the study region is most easily explained by edaphic factors, since the lowest pH lake of the group is essentially neutral. Lakes located relatively closer to the Canadian Shield (e.g. Bird Lake) are subject to relatively high natural acidity in atmospheric and surface runoff waters, because the Canadian Shield has little soil cover. The soils that are present are thin, have low buffering capacity, and are subject to limited ion-exchange reactions, resulting in relatively less acid neutralizing capacity and higher natural acidity. In contrast, in areas with calcareous soils and tills (such as Jay Jay, Ruby, and Brokenpipe Lake areas), ion-exchange reactions are common; as such much natural acidity from atmospheric or surface runoff waters is neutralized before it enters the lake. Furthermore, in the more southern lakes, carbon dioxide uptake by aquatic plants eliminates bicarbonate, precipitates carbonates, and forms hydroxyl ions (Cole, 1983).

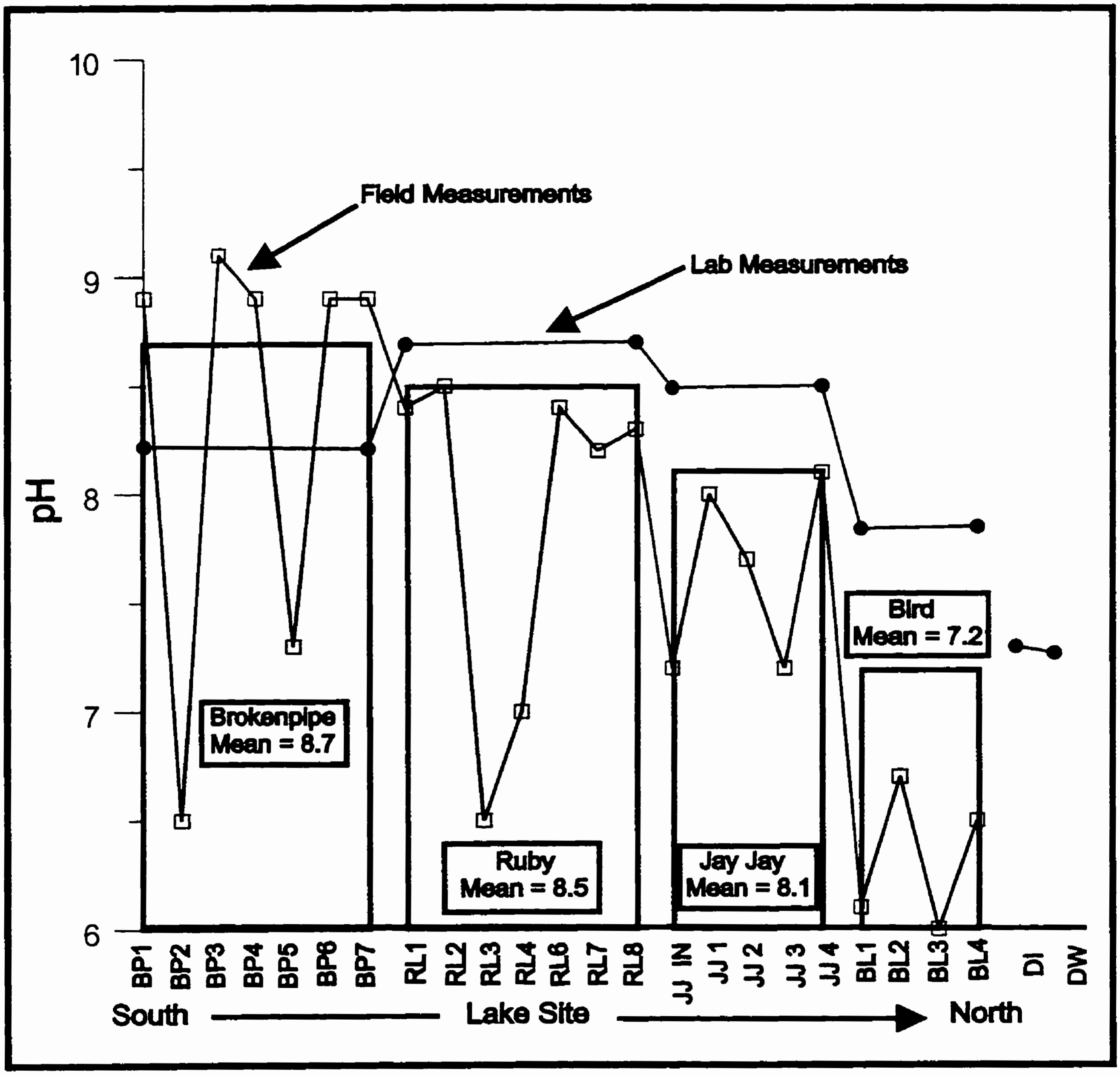


Figure 3-6: pH values for the main lake sites studied in this thesis. Field measurements taken with the Hydrolab unit, lab measurements were done by the DFO laboratory. Values in boxes and bar heights represent averages. Averages were calculated by omitting the lowest field measurements, as these were taken from marginal swampy areas of the lakes, rather than from within the main waterbody. Lake sites correlate to those in Fig. 2-6, and Table 2-A. DI = lab deionized water sample, DW = lab distilled water sample. See text for details.

All of these factors contribute to increases in pH. The overwhelming presence of *Chara sp.* at Brokenpipe Lake, and to a minimal extent at Ruby Lake, may partially account for the relatively high pH values in those lakes.

b) Major Ions:

A complete summary of the hydrochemistry of Brokenpipe, Ruby, Jay Jay, and Bird Lakes is given near the end of the chapter in Table 3-A. Potassium values (Fig. 3-7) are low in all lakes of this study. However, there is an overall progressive decrease in potassium content of the lakes from the south (Brokenpipe Lake) to north (Bird Lake) (Fig. 3-7). Brokenpipe Lake also has a notably higher potassium concentration than the other lake sites. The presence of potassium in the waters of the southern lakes likely a result of anthropogenic factors, rather than climate or weathering. Brokenpipe and Ruby lakes are nearer to agricultural lands; Jay Jay and Bird lakes are not in close proximity to agricultural lands. Agricultural fertilizer application to fields surrounding Brokenpipe and Ruby lakes is a likely source of potassium for these lakes.

Magnesium concentrations (Fig. 3-7) are quite low in the two northernmost lakes (Jay Jay and Bird), but are relatively higher in the two southernmost lakes (Brokenpipe and Ruby). Ruby Lake shows the highest values of magnesium for all sites.

Calcium concentrations (Fig. 3-7) are relatively high, but steadily decreasing from the south (Brokenpipe Lake) to the north (Jay Jay Lake). Bird Lake has relatively very low concentrations of calcium. Therefore, relatively high concentrations of calcium in the waters of Brokenpipe, Ruby, and Jay Jay Lakes may reflect the contribution of shallow

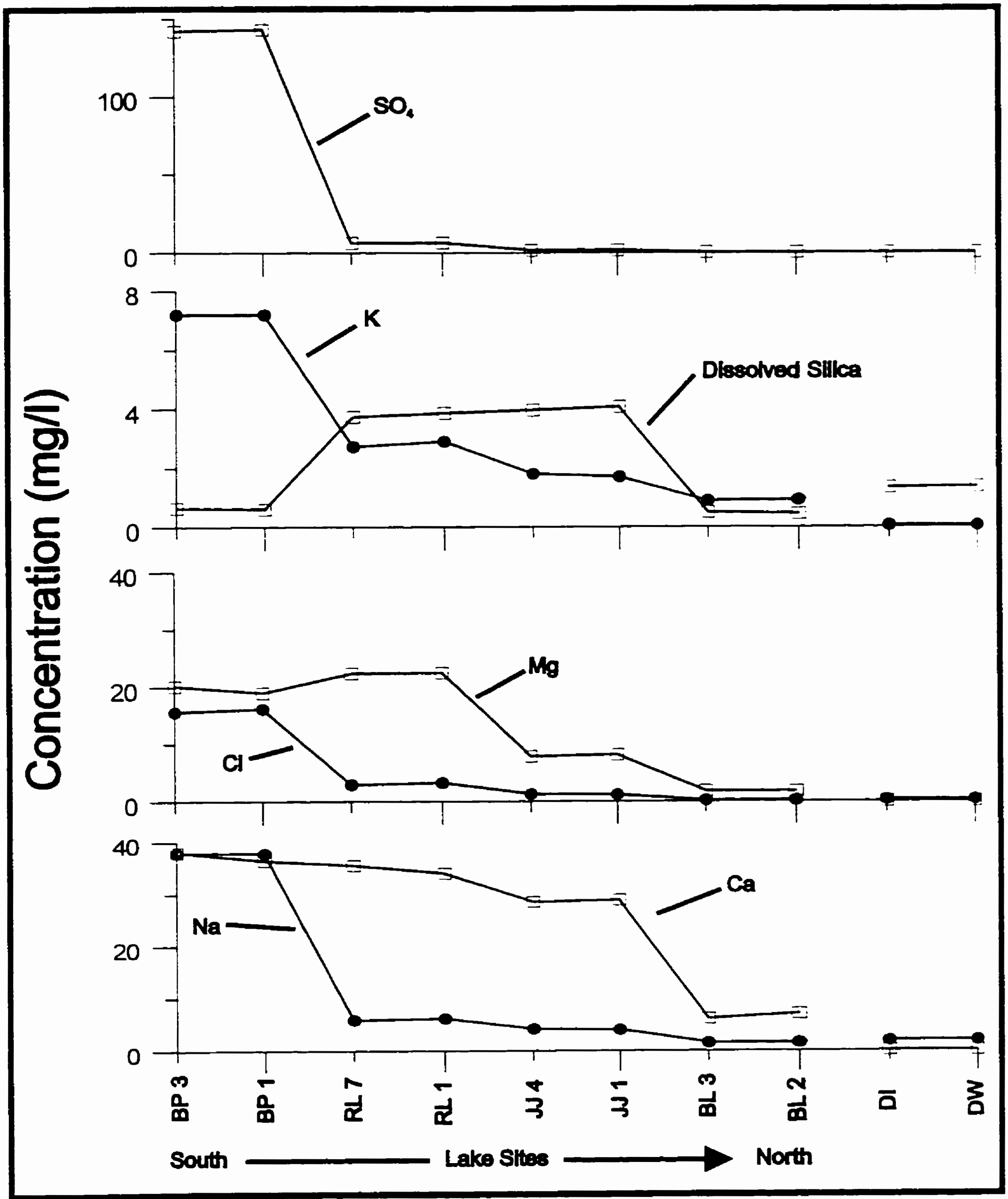


Figure 3-7: Concentrations of the major ions measured in the four main lakes of this thesis. Lake sites correlate to those in Fig. 2-6, and Table 2-A. Water chemistry is summarized in Table 3-A, data is tabulated in Appendix C. DI = lab deionized water sample, DW = lab distilled water sample. See text for details.

groundwater seepage through carbonate-rich till into the lakes. Weathering features are common in calcareous tills -- limestone hardpans (caliche) and white calcite or aragonite tubules that encase plant roots which have penetrated the till are indicative of chemical weathering and mobilization of calcium carbonate in solution.

Sodium concentrations (Fig. 3-7) are relatively high in Brokenpipe Lake, but are much lower and steadily decreasing for all lake sites to the north. An edaphic source for sodium is the leaching of igneous rocks (notably the feldspars). Once liberated, sodium tends to stay in solution (Cole, 1983). Since Brokenpipe Lake is far removed from the Canadian Shield (Fig. 3-2), and the tills in the region are high in carbonate (reflecting their origin from Paleozoic rocks), this is not a satisfactory explanation. Furthermore, lakes closer to the Canadian Shield (notably Bird Lake) have low sodium concentrations.

The concentration of the chloride ion (Fig. 3-7) is low at all sites, with the exception of Brokenpipe Lake, and somewhat elevated levels at Ruby Lake. Another possibility (for a sodium source as well) is an anthropogenic one; road salts (Cole, 1983; Taylor 1992) of $NaCl$ or $CaCl_2$ are commonly used to prevent ice-up on the highways. These salts are soluble, and are deposited into the lakes with spring runoff. Considering the fact that both Brokenpipe and Ruby Lakes (and only these two lakes out of the total of four sites) are situated near a paved highway, and that they have a relatively higher concentration of chloride ion in their waters, this seems to be a reasonable explanation.

Another possible source for the sodium and chloride is the influence of the east to north-easterly flowing groundwater of the evaporite - carbonate aquifer. Discharge of these waters occurs to the east of the Brokenpipe Lake site, along a northwest - southeast trending area to the east of the Mesozoic shale cover (Betcher *et al.*, 1995), which is the bedrock under the Brokenpipe Lake site. Many of these discharges are found on the western shore of Lake Winnipegosis. Diffuse seepage of these waters is known to have occurred into lakes Winnipegosis and Manitoba (Van Everdingen, 1971; Last, 1984; Last and Teller, 1983; Last *et al.*, 1994; Curry, 1997) and have had an influence in the solute composition and concentration in those lakes.

All the lake sites have low sulfate concentrations (Fig. 3-7), with the exception of the sulfate dominance observed in Brokenpipe Lake. Delorme (1969) comments on the fact that, in the prairies, the acidification of surface overflow by forest floor litter is absent. As such, sulfates and more soluble salts are leached first (i.e. prior to carbonates) and transported to lakes. This may be a factor contributing sulfate to the Brokenpipe Lake site, as the watershed is floored by Cretaceous shale bedrock. These shales are likely the ultimate source for sulphate in that is redeposited as gypsum in deeper soil horizons, within glaciolacustrine clays and till. Shallow groundwater seepage in locally focused recharge and discharge regimes may later dissolve the gypsum and deposit the sulphate in lakes that receive this shallow groundwater seepage. In fact, the presence of sulfate in shallow groundwater discharges and seeps, in till that was formed in close association with the Cretaceous shales, has been previously demonstrated (Little, 1973). Aside from water in the Swan River Formation (Fig. 3-3), the highest concentrations of sulphate in the Brokenpipe Lake area are in groundwater of unconfined sand and gravel deposits found in association with the till of the region (Little, 1973). A "gypsiferous mineralogic provenance" explained high sulfate concentrations in several lakes studied by Curry (1995).

While many of the major ions studied in the lakes show decreasing trends in concentration from the southernmost lake to the northernmost lake, the absolute concentrations are not particularly useful in deciphering clear relationships between ion concentration in a particular lake, and its original source. It is important to look at the concentrations of ions in terms of relative proportions (Fig. 3-8). Figure 3-8 illustrates that many of the northward-decreasing trends in ion concentrations are not maintained when the lake hydrochemistry is viewed as relative percentages. In particular, the decreasing trends in concentration of potassium and magnesium from the absolute concentrations essentially disappear, and the calcium trend becomes reversed when presented as relative proportions (Fig. 3-8). The relative proportion of sodium is highest in Brokenpipe Lake (Fig. 3-8), however the relative proportion of sodium in the other three lakes increases from the south (Ruby Lake) to the north (Bird Lake). These observations suggest that while most common ions are ultimately derived from weathering of bedrock and surficial sediments in the lake watershed, other considerations such as reaction thermodynamics of chemical weathering processes, biogenic factors, effective moisture, and hydrology all play a role in determining the salinity and proportions of ions in a lake. All of these factors complicate the description of the lake

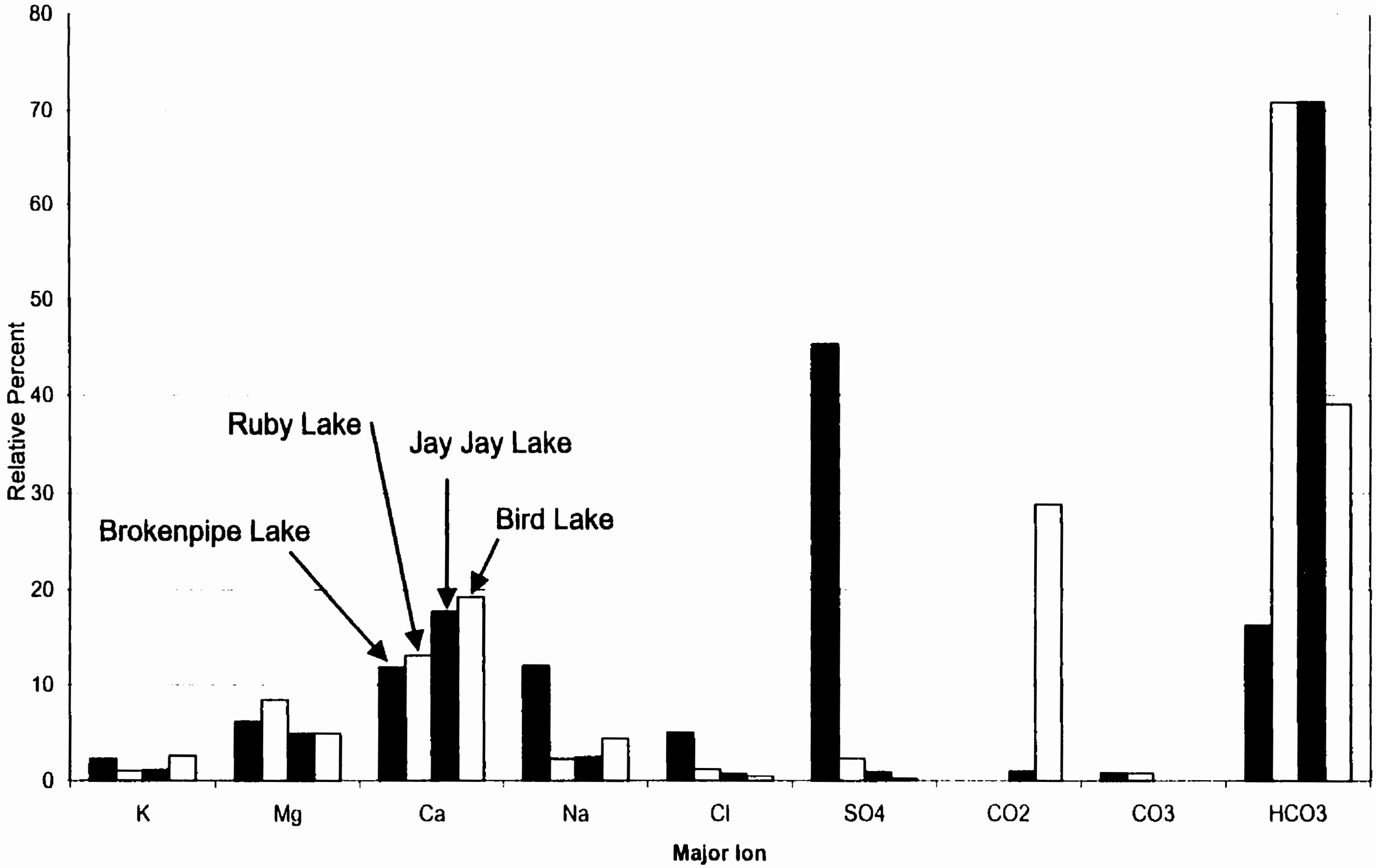


Figure 3-8: Relative percentages of major ions in each of the four main lakes investigated. Relative percents were calculated based on the TDS content of each lake. Data used is shown in Table 3-A.

hydrochemistry-lake watershed relationship, and the determination of the site-specific controls on the lake hydrochemistry.

3.3 - THE ROLE OF SHALLOW GROUNDWATER AND EFFECTIVE MOISTURE

Groundwater acquires its principal dissolved constituents during infiltration where net downward flow is balanced by infiltration and evaporation; during this time chemical processes of dissolution and precipitation primarily occur (Meyboom, 1967). The regional occurrence of gypsiferous shales tends to emphasize the sulphate content of surficial waters; this water percolates into the shallow bedrock, causing these aquifers to contain sulphate water (Meyboom, 1967). The relative proximity of the Brokenpipe Lake site to Cretaceous shales of the Manitoba Escarpment (which outcrop west of the lake and are the bedrock at depth below the lake), and the high sulphate concentrations noted in the waters of Brokenpipe Lake, may be an illustration of this generality.

The water compositions and concentrations of shallow, unconfined, surficial sand and gravel aquifers is absolutely dependant on their relative amount of aquifer outcrop area, the composition of the bedrock aquifer beneath them, and on the nature of the surficial material that surrounds or contains them. Unconfined aquifers in Manitoba include Pleistocene inter- and intra-till deposits, moraines, outwash areas, buried valley fills, alluvial sediments, and glaciolacustrine beach and deltaic deposits associated with Lake Agassiz (Betcher *et al.*, 1995). The TDS content of common surficial aquifers typically range from 200 to about 450 mg/l, if the aquifer has a large outcrop area.

Surficial aquifers without significant outcrop area tend to have a higher TDS, because they are subject to recharge from poorer quality waters from the surrounding glaciolacustrine clays, or tills (Betcher *et al.*, 1995).

The water compositions from interlake intra- and inter-till sand and gravel aquifers reflect an influx of water from the underlying carbonate bedrock and the surrounding carbonate-rich tills. These are calcium-magnesium-bicarbonate in nature, with a TDS less than 600 mg/l (Betcher *et al.*, 1995). Sand and gravel aquifers overlying bedrock areas that contain saline groundwater also reflect that fact by generally having increased salinities. This is well documented in areas underlain by the saline carbonate aquifer, such as west of Lakes Manitoba and Winnipegosis (Fig. 3-2), and west of the Red River in southern Manitoba (Betcher, *et al.*, 1995). The influx of groundwater from surrounding Cretaceous shale-rich tills probably is the control on the water compositions in these higher salinity cases (Betcher *et al.*, 1995).

Climate also affects the salinity and composition of ions in a lacustrine system, especially in terms of effective moisture. The moisture balance between precipitation and evaporation is often evaporatively coupled to water chemistry (De Deckker and Forester, 1988). Effective moisture is simply the value of annual precipitation (P) minus annual evaporation (E) in cm. Winter and Woo (1989) produced a map that illustrates P - E for all of Canada and the United States. All the lakes investigated in this thesis are occur where E>P, and the effective moisture values for all sites fall between -10 cm to -20 cm. All four lakes investigated are hydrologically open (i.e. they have inflow and outflow), and as such they will not record changes in climate as effectively as a hydrologically closed basin. Changes in climate directly sensed by the ostracodes will be subtle. The hydrochemical changes in salinity and ion composition, which are partially a function of climate (i.e. effective moisture), are recorded by the ostracodes because of their high sensitivity to these two parameters.

The evolution of a solute solution involves the process of evaporative concentration, which leads to mineral precipitation, and salt storage. Large changes in ionic composition occur when a mineral begins to precipitate, called the branchpoint. The result of these processes is that the water becomes enriched in a particular cation or anion, resulting in the depletion of the complimentary anion or cation (De Deckker and Forester, 1988). Thus, waters that increase in concentration (TDS) beyond the calcite branchpoint become either enriched or depleted in Ca^{2+} or HCO_3^-. For example, evaporative concentration causes enrichment in bicarbonate over calcium in the lake water because the deposition of calcite uses proportionally more dissolved calcium than dissolved bicarbonate (Curry, 1997). The predominance of groundwater inflow to a lake, for example, would be reflected in the lake hydrochemistry as dissolved bicarbonate > dissolved calcium. The calcite branchpoint occurs at low salinity (about 250 mg/l), and it marks a major change in water chemistry as evapotranspiration becomes an important mechanism to the hydrochemical evolution of a waterbody (Forester, 1987).

The relationship of climate to the ion concentration and composition of the modern lake sites cannot be discerned unless there is a large enough data set of lake water chemistries collected from across climatic gradients, in order to make comparisons and correlations. This approach was not taken in this thesis. However, in the upper mid-western part of the United States, Forester (1987) found that the calcite branchpoint corresponded with a region where annual precipitation was about equal to or less than annual precipitation. Though all four lake sites in this study fall essentially within this same climatic gradient (their moisture balances slightly E>P), the TDS concentrations of Brokenpipe and Ruby Lakes are greater than the calcite branchpoint, while the TDS concentrations of Jay Jay and Bird Lakes are not (Table 3-A). The evolution of Brokenpipe Lake and Ruby Lake hydrochemistry beyond the calcite branchpoint is complicated by the biogenic precipitation of calcareous marls.

Given that all the lakes were at one point lagoons of glacial Lake Agassiz during the Upper Campbell level, prior to the lagoonal stage they were occupied by cold, dilute (TDS <100 mg/l) Lake Agassiz waters (Curry, 1997). Through time, as the Upper Campbell beach formed and isolated the lake sites from the main body of Lake Agassiz, and after Lake Agassiz receded from their areas, the lake site hydrology eventually became predominantly governed by watershed geology, shallow groundwater flow systems, and to a lesser extent, climate (effective moisture). The evolution of each site from one occupied by Lake Agassiz waters to one with water chemistry dominated by shallow groundwater systems and watershed geologic sources of major ions (which is observed today) undoubtedly involved relatively cool and wet

Table 3-A: Summary of Water Chemistry for All Lakes Studied

Lake	Ion [Mg/l]									
	K^+	Mg^{2+}	Cl^-	Na^+	Ca^{2+}	SO_4^-	CO_2 ($+H_2CO_3$)	CO_3^{2-}	HCO_3^-	TDS
BP *	7.2	19.5	15.9	37.9	37.2	142.4	n/a	2.8	51.3	314.2
RL **	2.8	22.4	3.1	6.0	34.8	6.2	n/a	2.1	188.2	265.6
JJ ^	1.8	8.0	1.2	4.0	28.7	1.5	1.7	n/a	114.5	161.4
BL ^^	0.9	1.7	0.17	1.5	6.6	0.09	9.9	n/a	13.4	34.3

Note: All major ion concentrations are averages from two water samples taken at each lake in Mg/l. Carbon dioxide, carbonate and bicarbonate calculated from average of two dissolved inorganic carbon (DIC) readings and field pH following Mackereth *et al.* (1978).

TDS = Total Dissolved Solids (= sum of all ions listed above)

* Brokenpipe Lake

** Ruby Lake

^ Jay Jay Lake

^^ Bird Lake

n/a = not applicable

periods (glacial and post-glacial periods) and relatively warming and drying conditions (e.g. the Hypsithermal), which should be reflected in the paleolimnology of the fossil ostracodes found in the cores taken at each lake site.

3.4 - MODERN WATER CHEMISTRY, AS INDICATED BY OSTRACODES

The modern ostracode fauna was sampled from each lake, in order to compare the hydrochemistry of the modern lake to the modern ostracode fauna. This gives the researcher a tool with which to compare fossil ostracode faunas, and then infer paleohydrochemistry. The modern ostracode fauna of each study site was sampled during the mid-summer by using brine shrimp dip nets. Samples were stored in jars, and later processed and identified in the lab. Each ostracode fauna is listed in Table 3-B. The list includes dead ostracodes. These ostracodes had shells that contained soft parts and, therefore, are believed to have been alive during the same year as they were collected.

De Deckker and Forester (1988) hypothesized that ostracode diversity (i.e. the number of species present) would increase and then decrease with increasing salinity. Considering the fact that the TDS of the four study lakes are quite low, the relatively low species richness of the modern ostracode record suggests that this theory is valid. Furthermore, species richness decreases as salinity decreases; there are fewer and fewer ostracode species found as the lakes become more dilute (e.g. more ostracode species are found in Brokenpipe Lake than in Bird Lake).

All four lake sites (Brokenpipe Lake, Ruby Lake, Jay Jay Lake, Gregory Lake) today predominantly contain *Cyclocypris ampla* and *Cypridopsis vidua*. These

Table 3-B: Modern Ostracodes Collected at the Lake Sites

	Lakes			
Ostracode Species	Brokenpipe	Ruby	Jay Jay	Bird
Cyclocypris ampla	+	+	+	+
Cypridopsis vidua	+	+	+	+
Candona acutula	+	-	-	-
Candona decora	+	-	-	-
Candona distincta	+	+	-	-
Candona ohioensis	-	+	+	-
Candona paraohioensis	-	+	-	-
Candona candida	-	+	+	-
Limnocythere itasca	+	-	-	-

Note: + denotes presence, - denotes absence, Data includes "recently dead" ostracodes

ostracodes are characterized as tolerating wide variability of environmental conditions (Curry, in press). In particular, *Cypridopsis vidua* is commonly found in the littoral regions of lakes with appreciable groundwater discharge (Forester, 1991).

Cyclocypris ampla is noted as often being the only common ostracode in very dilute waters. As salinity increases, it tends to occur in association with one or several other ostracode species (Forester *et al,* 1989). The presence of *Candona candida* in Jay Jay and Ruby Lakes implies shallow permanent water, but also perhaps a substantial modern lotic input into these sites (Delorme, 1970a).

The candonids found in Brokenpipe Lake are tolerant of sulphate-rich waters, while the candonids found in Ruby Lake are not (Curry, 1998, personal communication). This agrees with the very high relative proportion of sulphate in Brokenpipe Lake, as opposed to Ruby Lake (Fig. 3-8).

Other ostracodes (Table 3-B) found during the collection of these modern samples were present in very low numbers, or their shells did not contain soft parts. It should be stressed that because the modern ostracode fauna was sampled at one time, from a small number of lakes, specific conclusions about how well the modern ostracodes represent the modern lake chemistry cannot be made. This sampling strategy is not regional in scale, and it is not likely representative of the complete ostracode fauna of each of the study lakes, and of lakes in the surrounding area. The modern ostracode fauna are typical of those observed in shallow permanent ponds and lakes with variable salinity and temperature.

CHAPTER 4 - GENERAL DESCRIPTION AND INFORMATION COLLECTED FROM EACH LAGOON SITE

4.1 - THE BROKENPIPE LAKE SITE: INTRODUCTION

Brokenpipe Lake (Fig. 4-1) lies in a former bay of glacial Lake Agassiz between the Riding and Duck mountains. Surficial deposits in the central portion of this ancient bay consist of lacustrine sands, silts, and clays; the periphery of the bay is dominated by ground moraine consisting of till and lake washed till (Fig. 4-2; Ehrlich *et al.*,1959; Klassen, 1979). Sediments deposited into Lake Agassiz are abundant in the area, and those originating from the Valley and Wilson Rivers (see discussion below) were of predominant importance in the development of the Upper Campbell beach at the Brokenpipe Lake site. Previous to this study (Mann, *et al.*, 1997), the Rossendale Gully (Teller, 1989), and the Swan River Valley (Nielsen, *et al.*, 1984) (Fig. 4-1), were the northernmost Upper Campbell Lake Agassiz paleolagoon sites ever examined in Manitoba.

4.1.1 - Geomorphological and Sedimentological Observations

The Upper Campbell beach in the immediate vicinity of Brokenpipe Lake underlies Highway 10 (Fig. 4-2). Several kilometers north of the lake site, the highway turns easterly, and is underlain by the Lower Campbell beach there. There is a flight of well developed beaches west of Brokenpipe Lake at elevations of about 380 m, about 32 m higher than the Campbell beaches at Brokenpipe Lake (number 1, Fig. 4-2). The elevations of these higher beaches correlate with the Norcross level beaches as

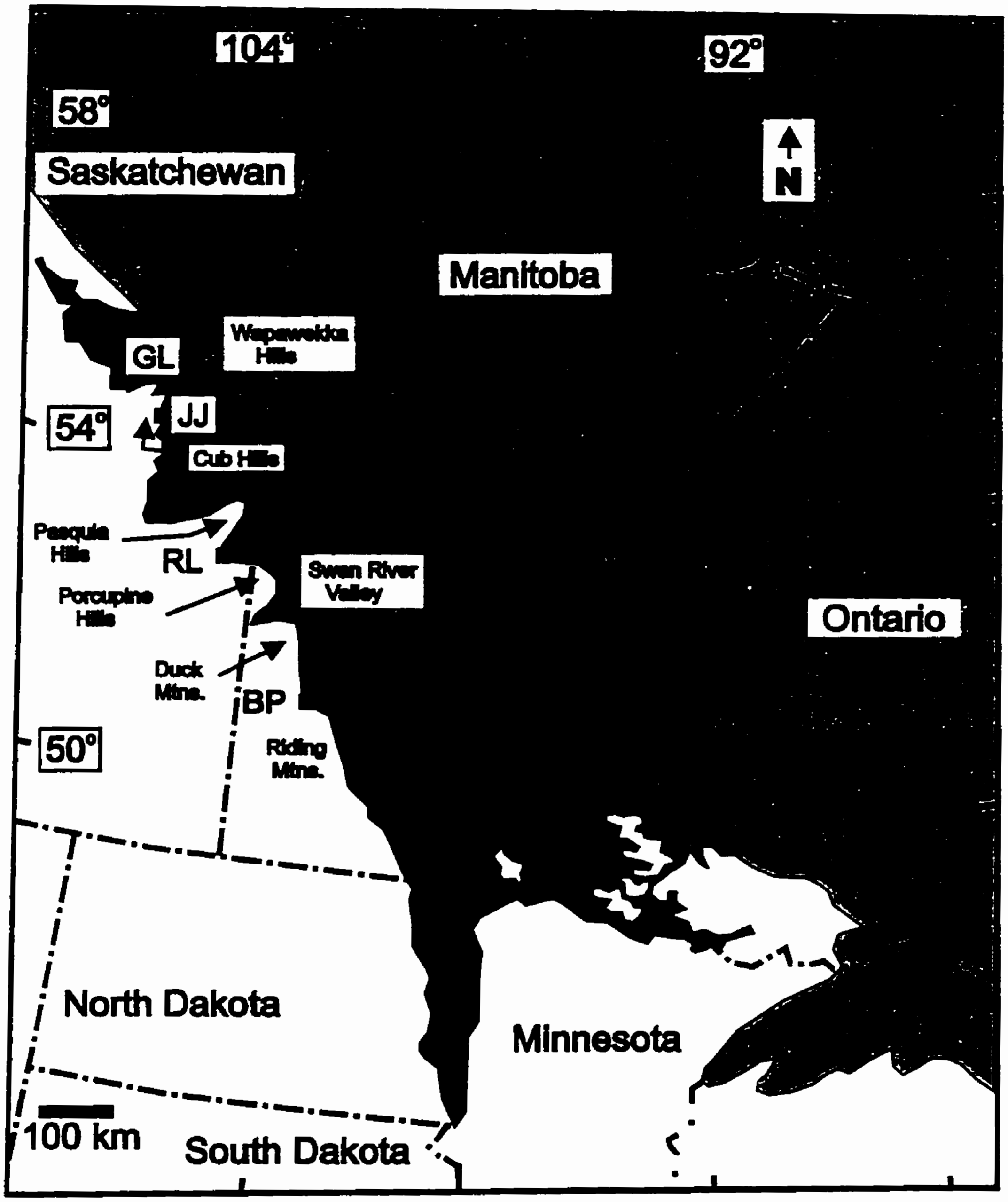


Figure 4-1: Location of new Upper Campbell lagoonal sites studied in this thesis. BP = Brokenpipe Lake, RL = Ruby Lake, JJ = Jay Jay Lake, GL = Gregory Lake. The Riding and Duck Mountains, with the Manitoban Porcupine Hills, make up the Manitoba Escarpment. Upper Campbell Lake Agassiz extent (dark shading) and Laurentide Ice Sheet (light shading) after Teller (1985, 1987).

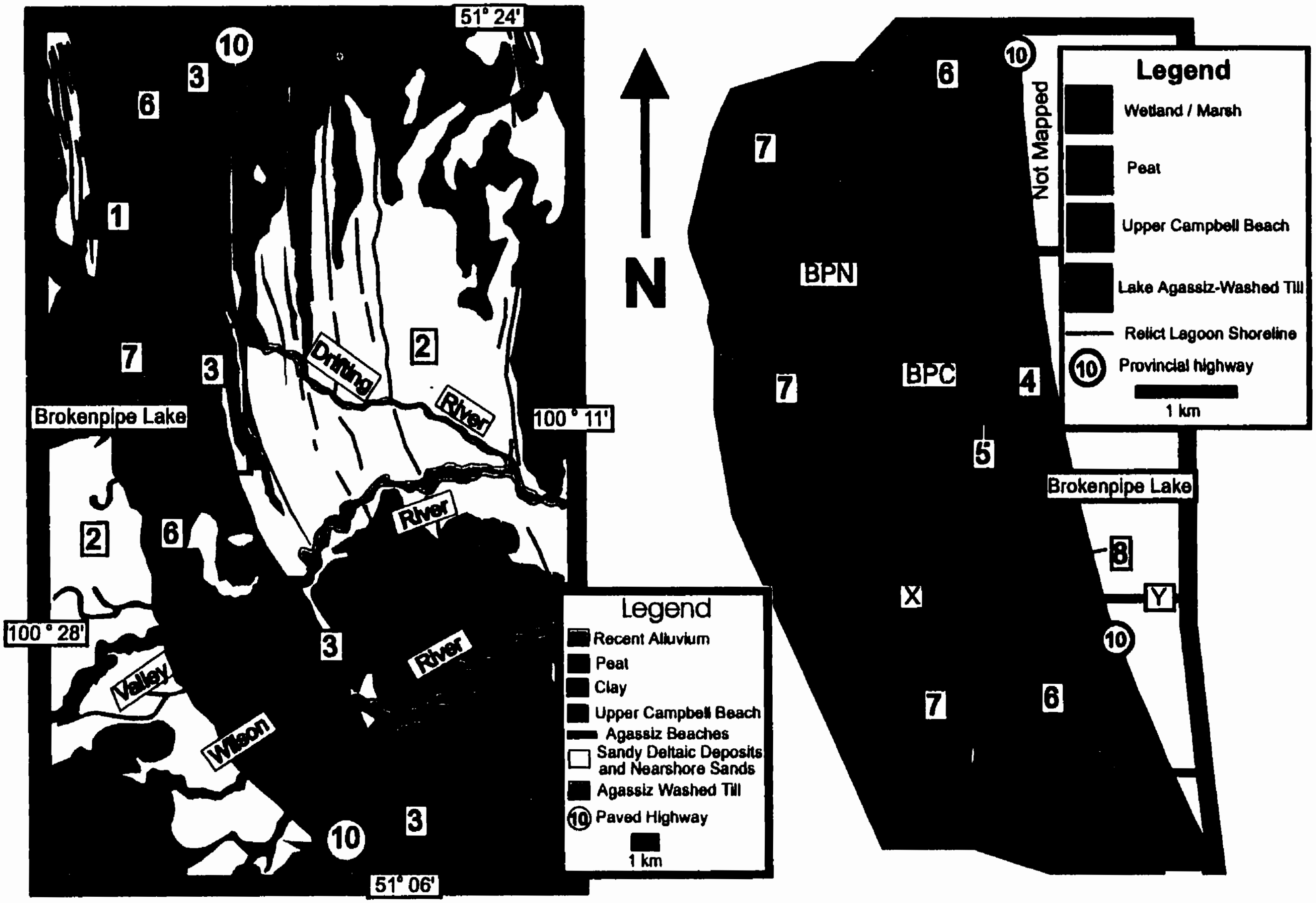


Figure 4-2: Geomorphological features of the modern Brokenpipe Lake area. Diagram at left after Ehrlich *et al*. (1959); Klassen (1979), diagram at right mapped from air photos. Numbers in boxes refer to features of the area discussed in the text, outline of area at right shown in left diagram. Note the locations of the BPC and BPN cores in the diagram at right.

described by the isobases and strandline diagram of Teller and Thorleifson (1983). To be absolutely certain of the identity of these strandlines, observation of the rebound gradient along the beach length is important. Herman beaches, if present in the Brokenpipe Lake area, may be at similar elevations to Norcross beaches. Fortunately, because they were formed earlier than Norcross beaches and were subject to additional isostatic adjustment, a Herman beach would show a steeper rebound gradient along its length than would a Norcross beach. Based on the information to date, it is likely that these high elevation beaches at the Brokenpipe Lake site correlate to the Norcross level of Lake Agassiz. Sandy deltaic deposits of the Valley and Wilson Rivers (number 2, Fig. 4-2), within which the Gilbert soil association is formed (Ehrlich *et al.*, 1959), reflect the different levels of Lake Agassiz that were present in the Brokenpipe Lake area. These sands are trimmed to the elevation of the Norcross strandlines, indicating that previously deposited sand was re-mobilized by the Norcross level of Lake Agassiz. The next large deposit of these sands is directly below the Upper Campbell. This implies that when Lake Agassiz stabilized at the Upper Campbell level, the deltaic and littoral sands continued to be deposited into Lake Agassiz from the Valley and Wilson Rivers. Deposition of this sand during the Upper Campbell level is also supported by the fact that there are younger Lake Agassiz beaches (i.e. basinward and lower in elevation than the Upper Campbell) formed in the sand deposit.

Therefore, the sand and gravel for the Upper Campbell beach (number 3, Fig. 4-2) probably originated as deltaic deposits from the Valley and Wilson Rivers. The Upper

Campbell beach prograded northward past the Brokenpipe Lake site. This is supported by the fact that the Upper Campbell beach is an erosional (scarp) feature south of the Wilson River. Between the Wilson and Valley Rivers it is a depositional (beach ridge) feature; north of the Valley River the Upper Campbell becomes a broad, flat-topped spit platform, with gravel topbeach and forebeach beds (Fig. 4-3).

Also attesting to a northward progradation of the Upper Campbell beach are several arcuate spit fingers (recognized on air photos) directly north of Brokenpipe Lake (number 4, Fig. 4-2). They extend from the backside of the main Upper Campbell beach. These spits were constructed during the initial stages of Upper Campbell beach building and northward progradation past the Brokenpipe Lake site. While it was under construction, the northernmost tip of the spit platform served as a headland, which was eroded by northward flowing longshore currents, constructing the spit fingers as a subaqueous extension of the main spit platform. The continued northward progradation and accretion of the main subaqueous spit platform beach (and the subsequent modification and subaerial buildup by Lake Agassiz waves) resulted in the elevation of the spit fingers being several meters below that of the main Upper Campbell beach.

Brokenpipe Lake lies within and is surrounded by a marshland-filled, back-beach basin that is several times larger than the lake itself (number 5, Fig.4-2). West of the low-lying marshland area, there are sporadic and discontinuous shoreline deposits that were formed during short pauses in the declining Lake Agassiz water levels.

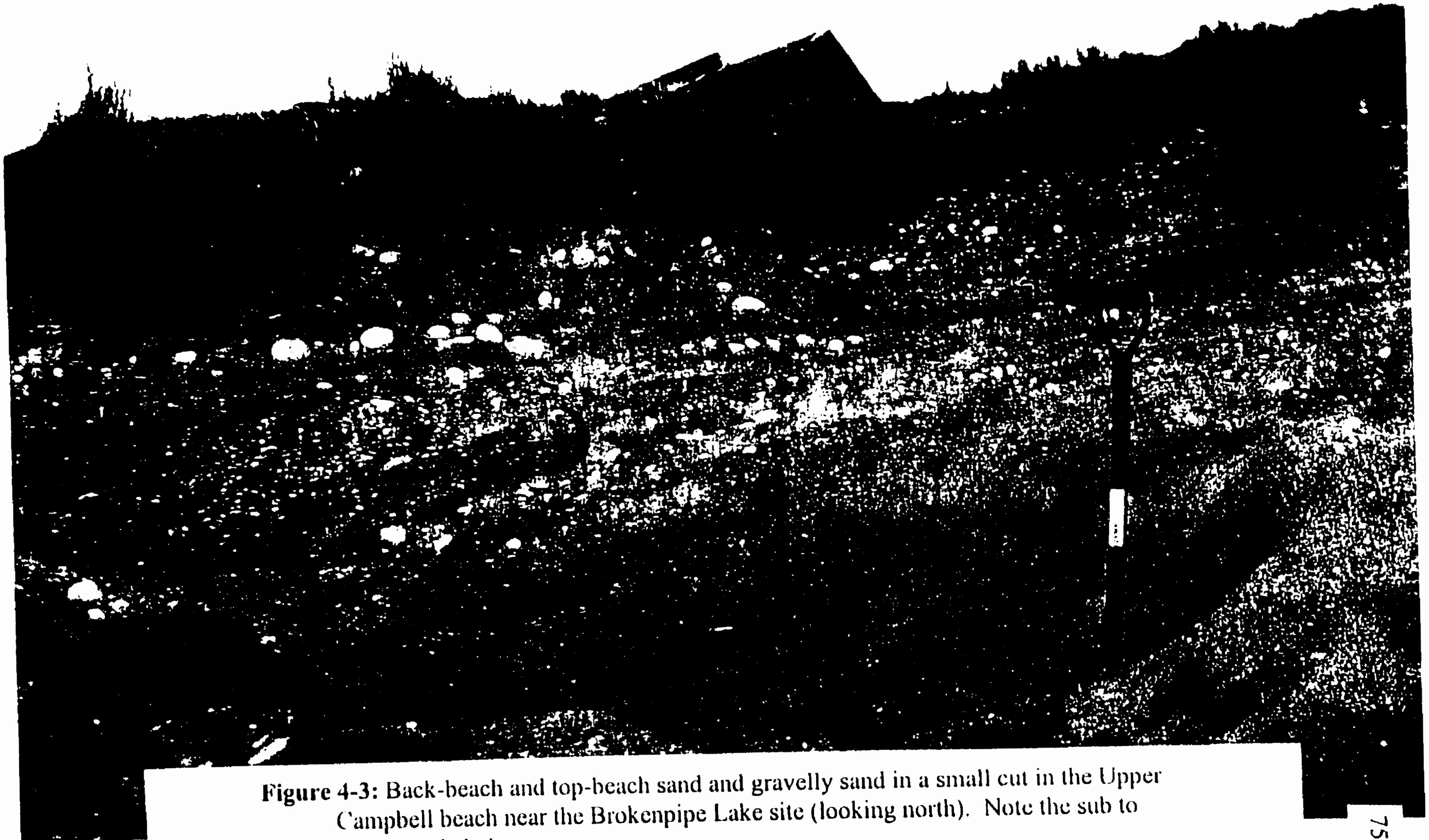


Figure 4-3: Back-beach and top-beach sand and gravelly sand in a small cut in the Upper Campbell beach near the Brokenpipe Lake site (looking north). Note the sub to well-rounded clasts.

These discontinuous shorelines include representations of Agassiz water levels between the Norcross and Upper Campbell levels (number 6, Fig. 4-2) (Ehrlich *et al.*, 1959). Below this moderately well-developed waterplane (left diagram in Fig. 4-2, labeled 6) only slightly westward of the Brokenpipe marsh, and at elevations very close to that of the top of the Upper Campbell beach, are relict shorelines formed in till. The relict shorelines are visible only in large scale (1:12,000) aerial photographs (right diagram, Fig. 4-2, labeled 6). From 1:50,000 topographic maps, the approximate elevations of these relict shorelines are 347 m, 352 m, and 355 m. These shorelines show distinctly in air photos as tonal contrasts, they divert the eastward flow of surface drainage by acting as barriers, and are visibly positive relief features in the field. In the agricultural fields where these shorelines occur, they have little, almost imperceptible, relief. Furthermore, travelling west from highway #10 on the W-E oriented section road south of the Brokenpipe site (labeled X-Y on Fig. 4-2), there are meter-scale steps up in the road; one at 800 m from the # 10 highway (the rise in elevation out of the low lying marsh area), one at 1000 m (the first relict shoreline), and another at 1200 m (the second relict shoreline).

There are 8 shallow peat-filled channels, each several meters in width, west of Brokenpipe Lake (number 7, Fig. 4-2). Each was formed in till, and carried drainage eastward from the Riding Mountain uplands. They have not carried a significant volume of water for a very long period of time, based on the assumption that it took at least several hundred years for the layer of approximately 15 - 25 centimeters of peat to completely fill them in. A drowned stream (number 8, Fig. 4-2) once flowed into the southern end of Brokenpipe Lake. This may show evidence of differential isostatic rebound that has been occurring during the Holocene, or it may represent a period of flow into the lake during a lowstand, and the subsequent drowning during a shift to a moister climatic regime.

4.1.2 - Cores From Brokenpipe Lake: Description

Brokenpipe Lake has a gently sloping littoral zone and a flat bottom. Maximum lake depth is ~ 2 m. The two cores selected for discussion are the BPC and BPN cores (Fig. 4-2). In terms of the BPC core, it is important to note that clay was collected from the base of both the BPC and BPD (Fig. 2-6) cores (taken less than 2 m apart), but the 30 cm collected in the BPD core was utilized for laboratory analysis. Because of the close proximity of cores BPC and BPD, and their identical stratigraphy, the longer clay sample recovered in the BPD core was stratigraphically attached at the base of the BPC core. An annotated description of the sedimentary units observed in the BPC and BPN cores is given in Tables 4-A and 4-B; the full description can be found in Appendix A. Figures 4-4 a - i show photographs of some of the interesting features of the BPC and BPN cores, which will be discussed in later chapters of the thesis. Figure 4-5 shows a generalized stratigraphic succession and the corresponding values for moisture, organic matter (LOI), and carbonate content for the BPC core.

Table 4-A: Stratigraphic Units in Core BPC

Depth Interval (cm)	Description
0 – 90	marl, with *Chara sp.* fragments, abundant molluscs, grading downward into peaty marl, gradually darkens downward, gradational basal contact
90 – 140	fine peat at 90 - 93 cm, wood frag. @ 90 cm (5980 +/- 70 yr BP; [TO-6205]) grading downward from a marly peat to a peaty marl at the base of unit, occasional molluscs, lower boundary gradational over 15 cm
140 – 196	marl, *Chara sp.* fragments; finer sized fragments than seen in upper marl unit, few visible molluscs, lower boundary gradational over 3 cm
196 – 230	clayey silt to silty clay, laminated except in upper 5 cm of unit, laminae are alternating light (silty) and dark (organic material +/- clay), thickest laminae are predominantly organic, wood frag. @ 203 cm, (9660 +/- 90 yr BP; [TO-6206]) lower boundary gradational
230 – 262	silty clay to clayey silt, laminated in 0.5-2.0 cm thick units, calcareous, no organic laminae, lower boundary distinct
262 – 265	sand, fg, well sorted, calcareous, mainly quartz, lower boundary distinct
265 – 281	silty clay to clayey silt, contains vegetal detritus, well laminated, increasing numbers of vegetal laminae (1 - 3 mm thick) downward, lower boundary distinct
281 – 285	clayey silt, non-laminated, calcareous, lower boundary distinct
285 – 289	sand, well sorted, calcareous, several laminae of clayey silt (up to 0.5 cm thick), basal contact distinct
289 – 291	vegetal detritus, slightly silty, some silty laminae, lower contact distinct
291 – 295	silty clay to clayey silt, contains vegetal detritus laminae (1-2 mm thick), wood fragments @ 293 cm (9350 +/- 70 yr BP; [TO-5880]), calcareous, lower boundary distinct
295 – 299	gravel, subrounded to rounded carbonate clasts, largest has long axis of 7 cm, calcareous matrix made up of mm-sized granule to fg sand sized carbonate and quartz sands, lower contact distinct
299 – 335	clay, occasional carbonate granule and pebble sized dropstones (up to 2 cm diameter), very stiff, except for floating grains clay is very smooth

Table 4-B: Stratigraphic Units in Core BPN

Depth Interval (cm)	Description
0 – 130	peat, medium to fine fibrous, molluscs throughout, 0 - 10 cm is lighter in color and is a mrly peat, bottom contact gradational over 1 cm
130 - 181	Gravelly silt, massive, carbonate subrounded clasts 2 - 5 mm in size, in a vcg - cg sand matrix, bottom contact distinct
181 - 210	gravel, massive, carbonate subrounded clasts 5mm - 1.5 cm in size in a vcg - cg sand matrix, grades downward into vcg sand matrix with extremely well-rounded clasts

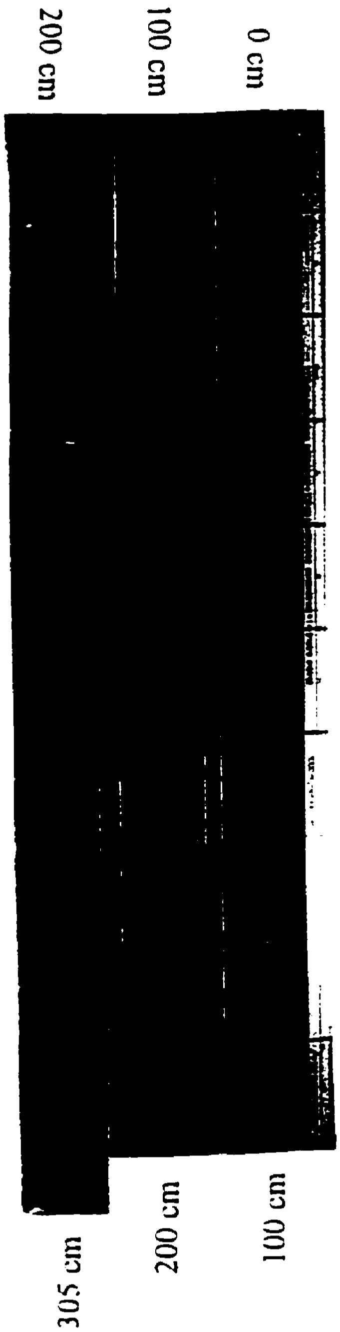


Figure 4-4a: Overall view of the BPC core, Brokenpipe Lake. For all photos of the BPC core (Fig. 4-4 a to g), and BPN core (Fig. 4-4 h and i), see Figures 2-6 and 4-2 for core locations. See Table 4-A for BPC unit descriptions, and Table 4-B for BPN unit descriptions. Distance between heavy dark lines on scale is 10 cm in all photos.

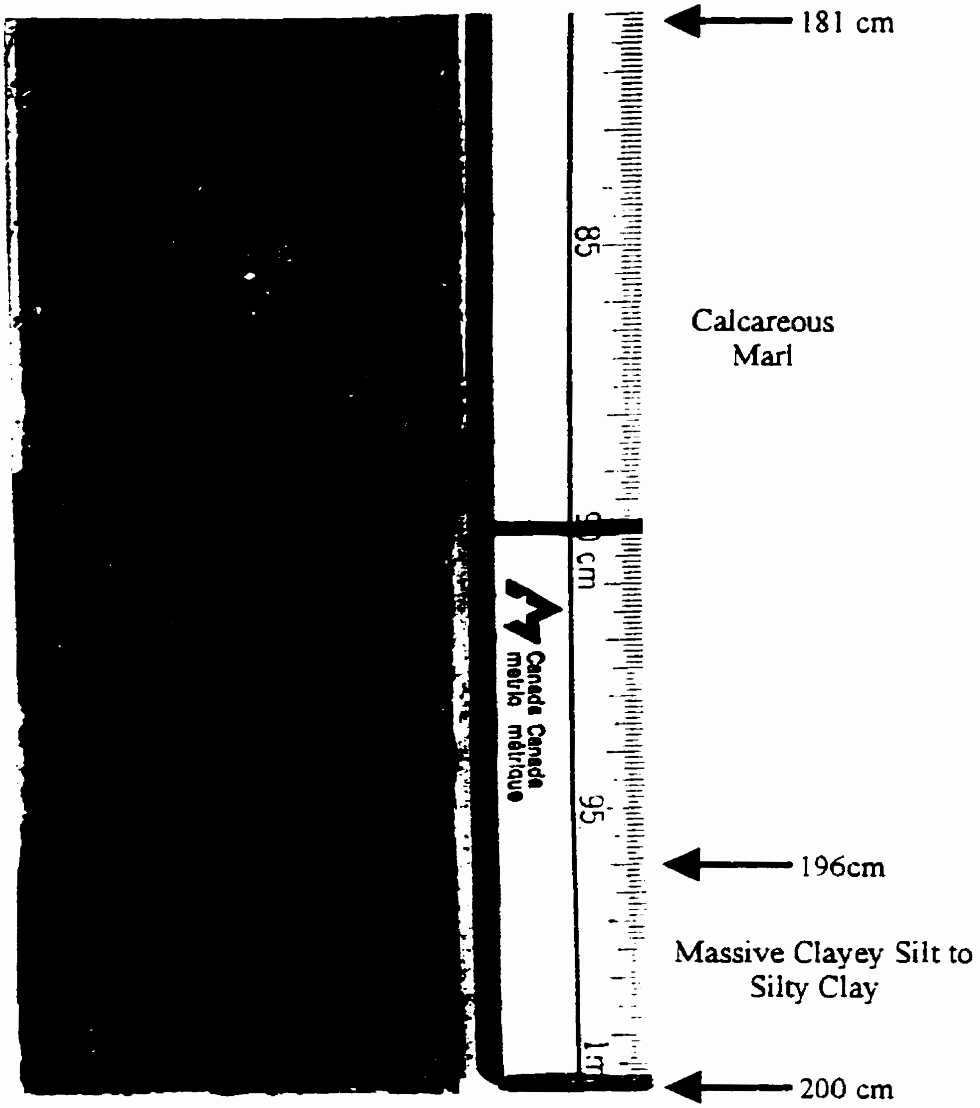


Figure 4-4b: Contact between silty clay and clayey silt of the lower BPC core (Brokenpipe Lake) and the overlying calcareous marl.

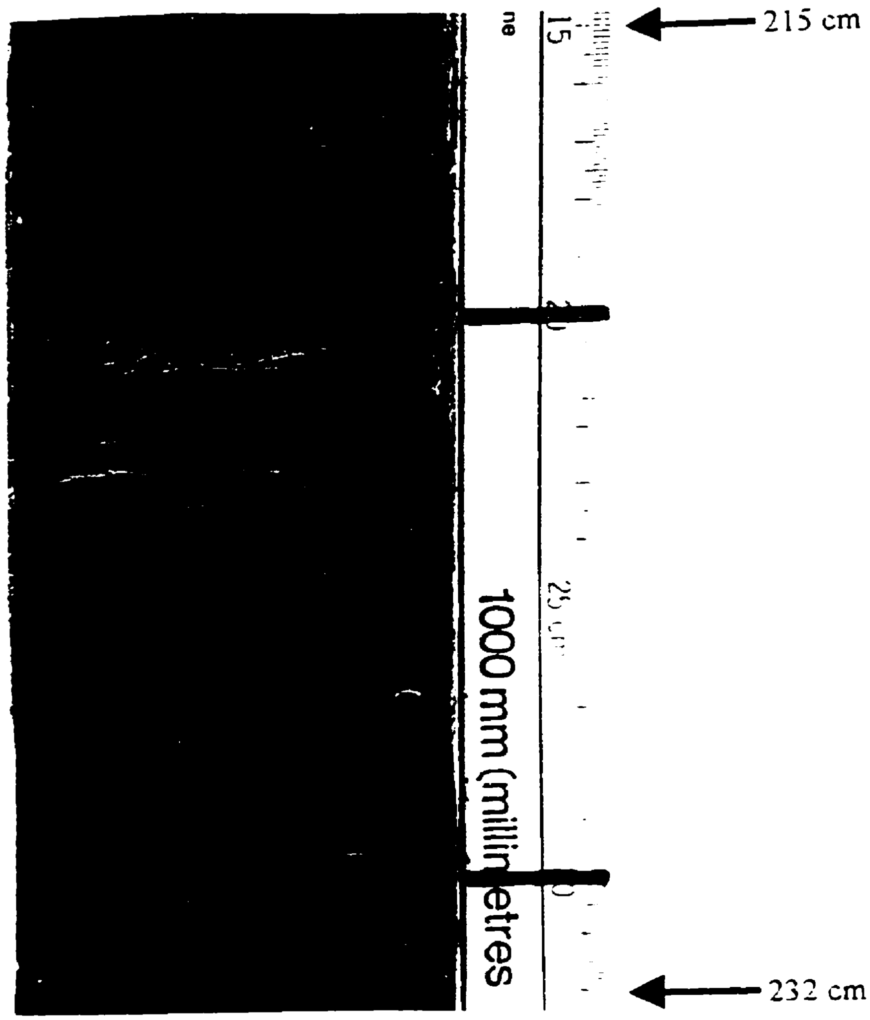


Figure 4-4c: Laminated clayey-silt to silty-clay; alternating laminae consist of silt (light) and organic-rich clay (dark). Photo is from the BPC core (Brokenpipe Lake)

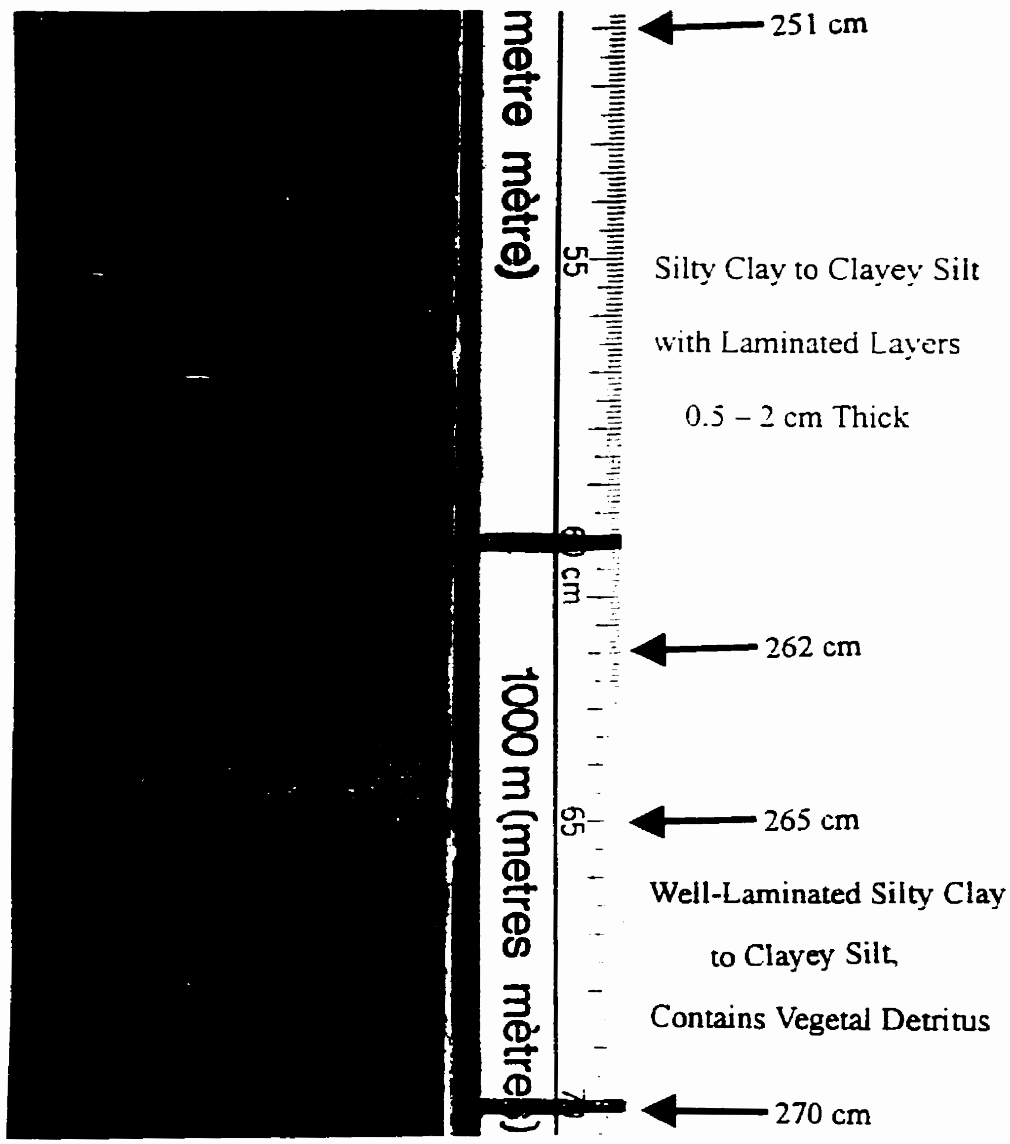


Figure 4-4d: Clean sand bed between finer-grained sediment. Photo is from the BPC core (Brokenpipe Lake)

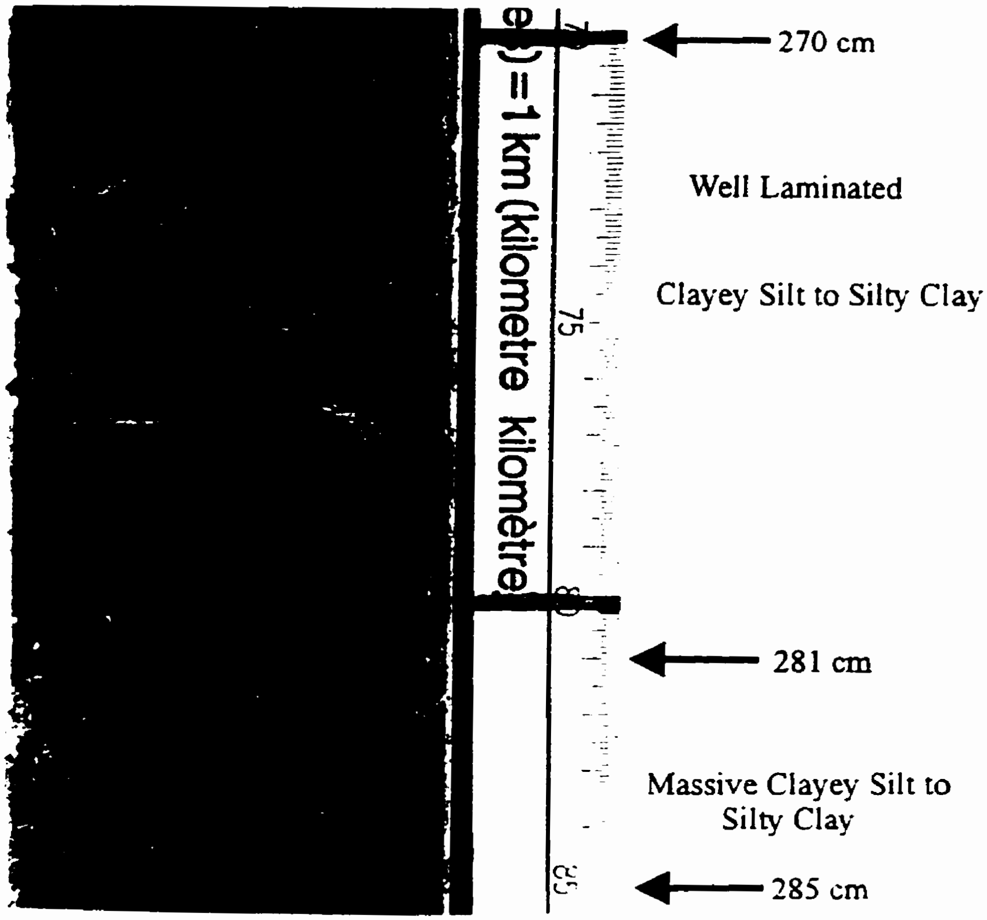


Figure 4-4e: Contact between laminated silty clay to clayey silt, and massive clayey silt. Photo is from the BPC core (Brokenpipe Lake)

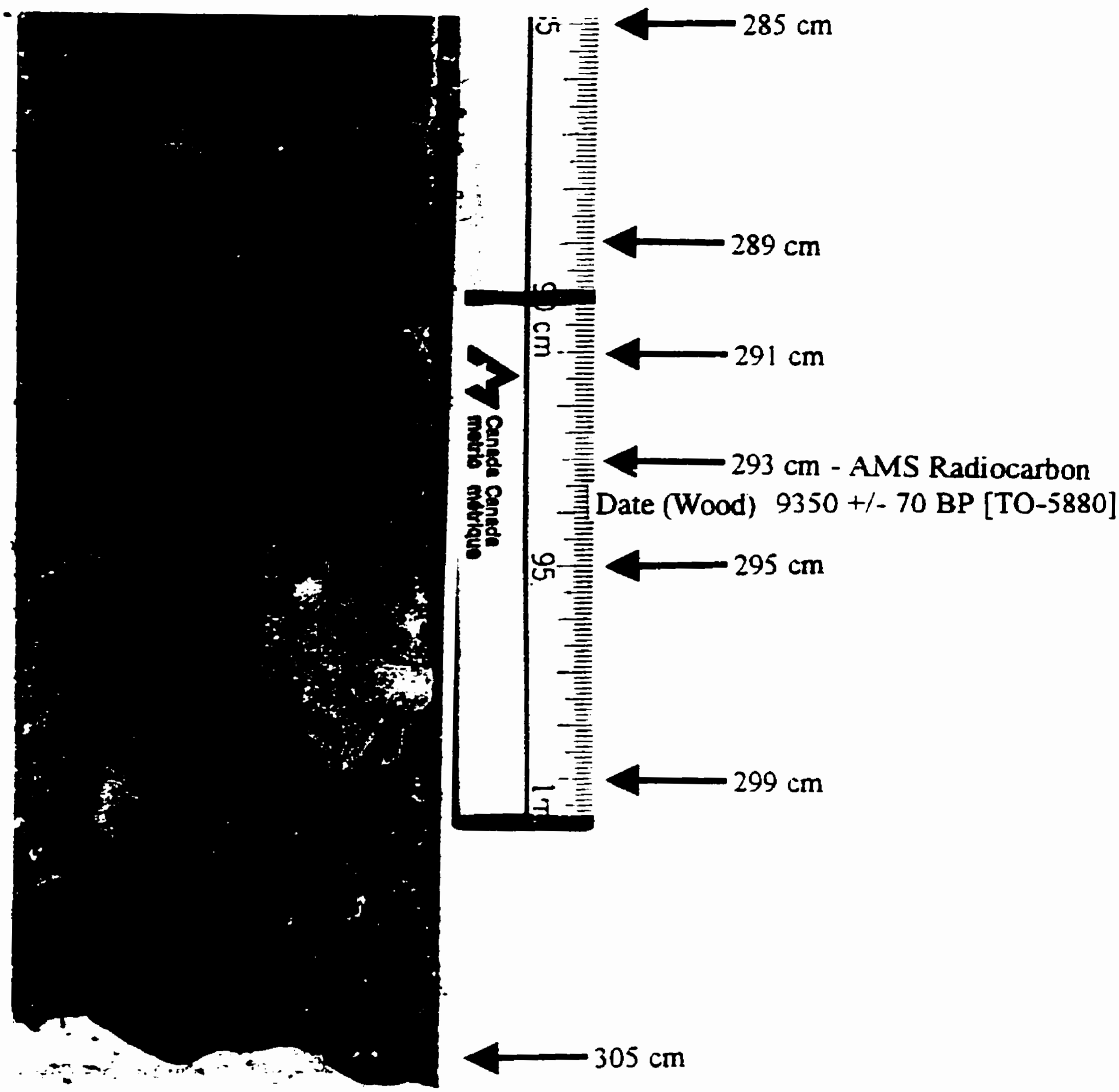


Figure 4-4f: Contact between laminated silty clay to clayey silt, underlying Upper Campbell beach gravel, and basal glaciolacustrine clay. Photo is from the BPC core (Brokenpipe Lake)

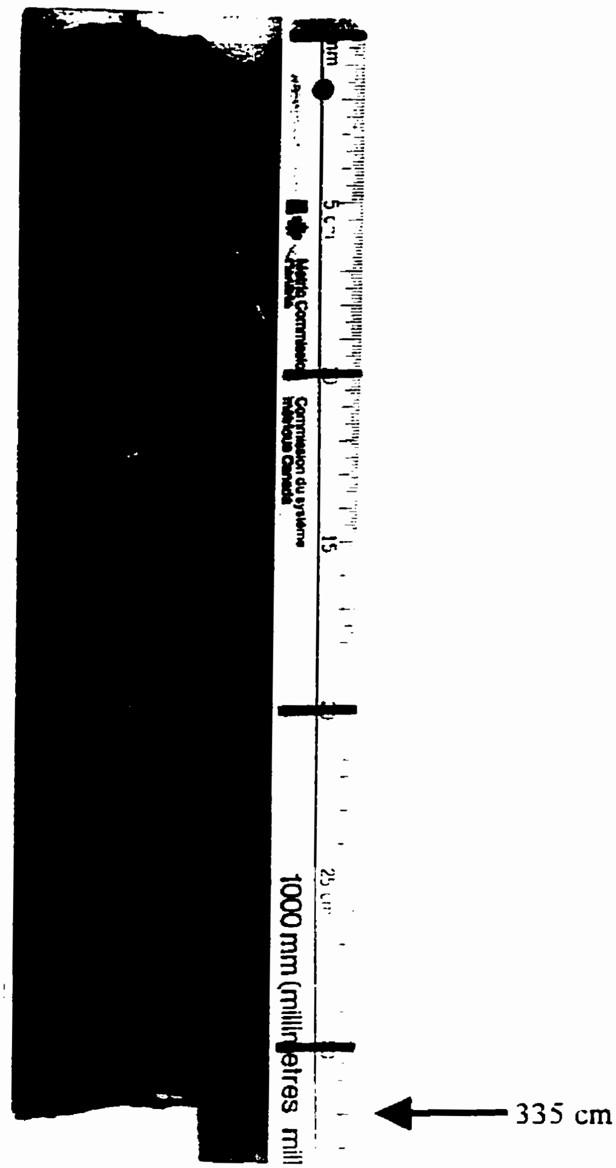


Figure 4-4g: Basal glaciolacustrine clayof the BPC core (Brokenpipe Lake). Unit contains mm-sized granule and pebble sized (up to 2 cm) dropstones. The clay is extremely compact and smooth. Deformation due to compression during extraction from the vibracore core catcher is noticeable at the base of the clay.

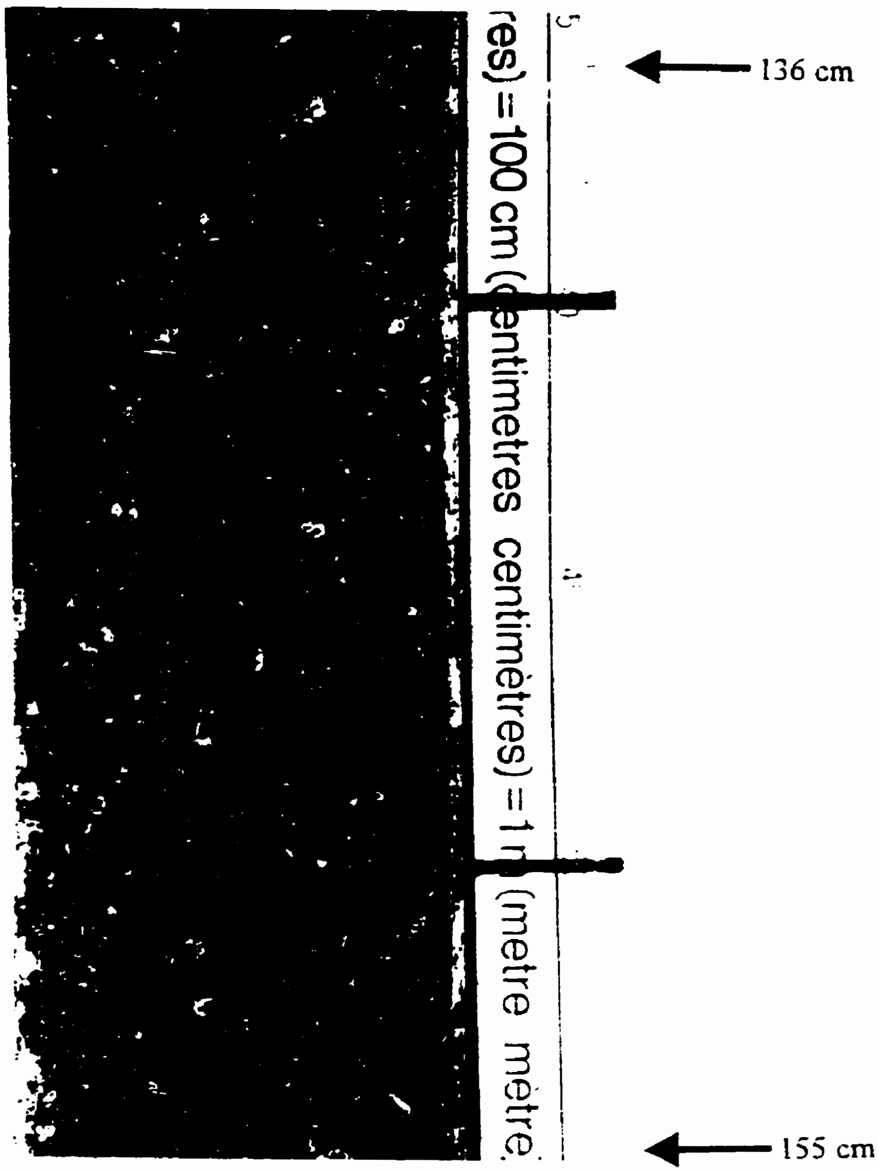


Figure 4-4h: Gravelly silt from the BPN (Brokenpipe Lake North) core. Clasts range from 2-5 mm in size, matrix is very coarse to coarse sand and silt.

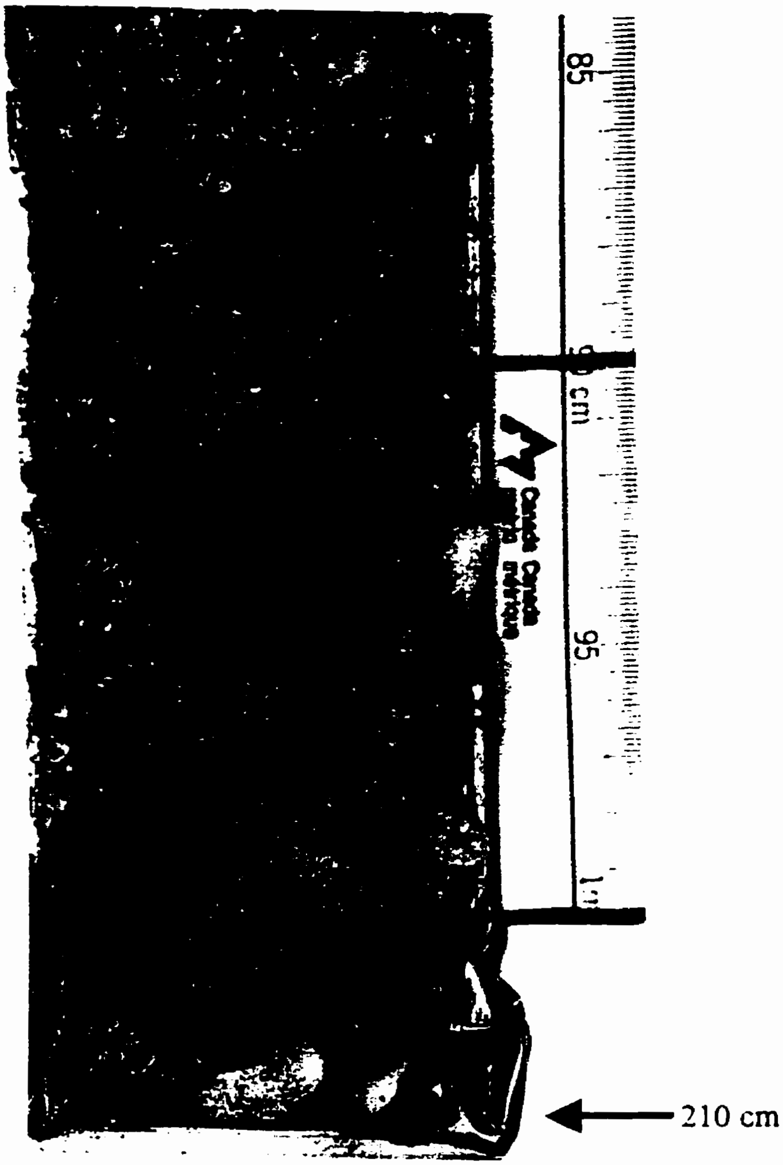


Figure 4-4i: Massive gravel from the base of the BPN (Brokenpipe Lake North) core. Clasts are well-rounded (5mm to 1.5 cm in size), in a very coarse grained to coarse grained sand matrix. Largest clasts are 7X5 and 5X3 cm in size.

4.1.3 - Description of the Moisture, Organic, and Carbonate Trends in Core BPC
a) Moisture:

In the upper marl portions of the core, where sediments are very permeable (e.g. 0 - 90 cm), moisture contents are generally high (about 65%) and unchanging (Fig. 4-5). Moisture content decreases quite visibly in the upper portion of the 90 - 196 cm zone, and then maintains this lower value to the base of the zone (196 cm). This is probably a reflection of: 1) decreased hydraulic conductivity of the peaty material versus a more marly material; and 2) dry-bulk density increases and the associated decreased hydraulic conductivity and porosity due to smaller marl fragments and compaction. Moisture content in the 196 - 295 cm interval (Fig. 4-5) is generally lower than that of the zone above. This reflects the change from marls and peaty-marls to lithologies such as sands, silts, and clays which would be expected to contain less moisture because of their hydrophobic nature, versus more hydrophyllic organic sediments. Several intervals in the 196 - 295 cm interval are of note. The 262 - 265, 285 - 289, and 295 - 299 cm zones show relatively reduced moisture contents. These zones are well sorted sands and poorly sorted gravel, which would be expected to have higher hydraulic conductivities and open pore spaces, enhancing evaporation of pore waters during core storage. Intervals of relatively high moisture contents (e.g. 289 - 295 cm) reflect moisture retention in zones with high proportions of vegetal detritus or

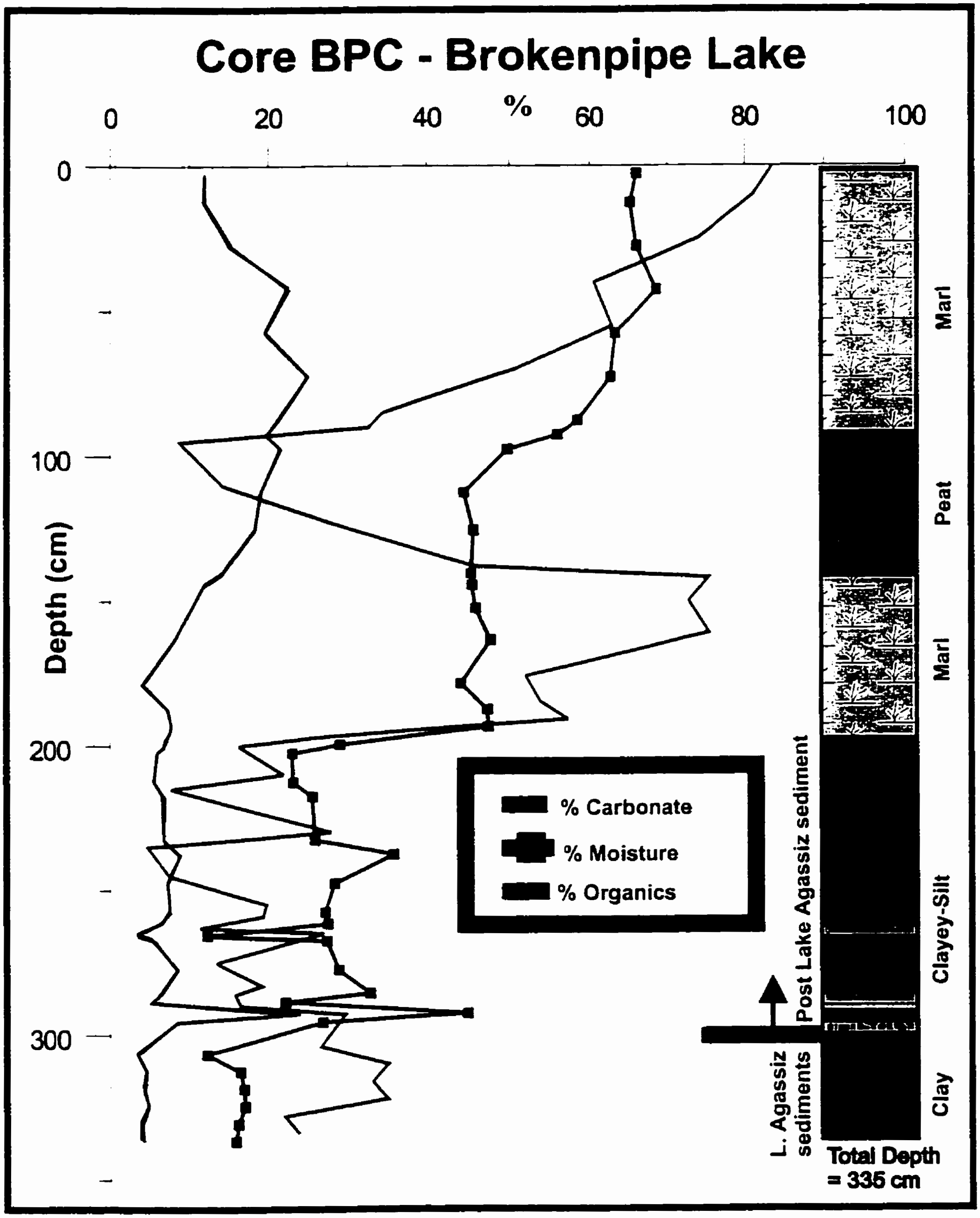


Figure 4-5: Moisture, organic, and carbonate percentage values for the BPC core (Brokenpipe Lake). Squares on the moisture curve indicate sample depths. Organic matter and carbonate contents were calculated by LOI from a dry sample. Generalized stratigraphic section at right.

organic material. Moisture content is low in the 299 - 335 cm clay unit, due to the reduced intergranular pore space available in the stiff, Lake Agassiz clay.

b) Organic matter (LOI):

Organic matter content of the 0 -90 cm zone is generally low (<30%), but generally increases with increasing depth (Fig. 4-5). This probably reflects the downward grading of the marl from a true marl at 0 cm to a more peaty marl at depth. The same can be said of the 90 - 140 cm interval. Organic matter content decreases from the upper portion of this zone to its base; this is probably a reflection of the downward gradation in the unit from a peat (90 - 93 cm) to more of a marly peat by 140 cm. Organic matter content of the 140 -196 cm marl unit continues a steady decrease from the unit above (90 -140 cm). The decreasing peat or plant matter content with depth is probably a reflection of the increasing proportion of siliciclastic sediments which comprise the bulk of the lower half of the BPC core. Thus, organic matter content of the 196 - 295 cm interval is generally low. Peaks of very low organic matter content are found in the clean sands at 262 - 265 and 285 - 289 cm, and a relatively high peak in organic content is found in the vegetal detritus layers in the 289 - 295 cm interval of the core.

c) Carbonate:

Carbonate content of the 0 - 90 cm zone is very high (Fig. 4-5) at shallow depths (about 70 - 85%), as would be expected in a calcareous marl. Carbonate content decreases with depth to a value of less than about 20% just above the marl - peat boundary. This may reflect a relative increase in the proportion of terrigenous sediments, as the organic matter content does not drastically change during this interval. Carbonate content is very low in the upper portion of the 90 - 140 cm unit, again probably due to an increased proportion of terrigenous sediments, and the peaty nature of the marl. Carbonate content increases with depth in the 90 - 140 cm unit, likely a reflection of increasing proportions of marl with increasing depth of the unit. Carbonate content is high (again about 55-80%) in the 140 - 196 cm unit, again reflecting a marl with a very small content of peaty organics. These values decrease but remain above about 50% with increasing depth in the zone. This may be a reflection of dilution by siliciclastic materials that originated during the transition from deposition of predominantly mineralogenic lithologies (found below the 140 - 196 cm marl zone) to calcareous marl deposition.

Carbonate content in the 196 - 295 cm zone (Fig. 4-5) is relatively low but highly variable. This possibly reflects the increased relative amount of carbonate silt found in lighter colored laminae of this unit versus the overall predominance of siliciclastic materials, though this hypothesis was never experimentally verified. Carbonate values are low and similar in the 299 - 335 cm zone (glaciolacustrine clay). The carbonate in the clay is probably a result of mm-sized carbonate dropstones and a small amount of carbonate silt.

4.1.4 - Fossil Ostracodes From Core BPC: Introduction

A waterbody where an ostracode can live is governed by its surrounding environment, and the environment is defined by the interaction of geological, hydrological, botanical, and climatic factors (Delorme, 1969; Forester, 1987; Smith, 1993). As discussed in Chapter 3 the geological materials in the watershed are usually the predominant sources for most common ions. While ground and surface water movement may be responsible for the transport of ions to ponds and lakes, the hydrology and successional development of the waterbody also plays a role in determining what ostracodes will live there. Changes in the total dissolved solids (TDS) concentration of a waterbody, through its successional stages, are first related to 1) ion input from increased leaching of watershed geological materials into a predominantly mineralogenic basin, leading to 2) increased nutrient and food production, which finally results in 3) organic sediment production and TDS reductions (Delorme, 1969), usually due to changes in hydrologic regimes (e.g. a shift from groundwater dominated input to surface runoff or a closed basin hydrology changing to an open basin hydrology).

Environmental factors (either chemical or physical) that limit populations are considered "limiting factors" (Delorme, 1969). Ostracodes are good indicators of changes (physical and chemical) in their habitats, as species occurrences often define discrete areas on graphs with thermal and chemical axes (Delorme, 1989; Forester, 1983, 1986, 1987; Smith, 1993; Curry, in press). The sensitivity of ostracodes to physical and chemical lacustrine systems is also linked to the needs, and stage, of their life cycles (Forester, 1991). For example, juvenile ostracodes typically have narrow growth ranges in temperature (and probably solute composition and TDS; Forester, 1991), but can survive as adults within wider ranges in environmental parameters (Colman *et al.*, 1990). Ostracodes are ubiquitous, and calcify shells commonly preserved in moderately buffered systems. They are small, and therefore are usually easily preserved in the geologic record of lakes with moderate sedimentation rates. They rapidly disperse, and would have quickly colonized large geographic areas after deglaciation. Ostracodes have low intraspecific competition, thereby allowing large population numbers to exist in the same habitat (Delorme, 1991).

4.1.5 - Fossil Ostracode Autecological Information For Core BPC

a) *Candona acutula*

Candona acutula is found in shallow water with abundant aquatic vegetation in the prairie-forest transition zone of the Canadian prairies (Delorme, 1970a). This ostracode is tolerant of broad fluctuations in water temperature (Delorme, 1991).

b) *Candona caudata*

In contrast to other long-lived ostracodes, such as *Cytherissa lacustris* (see section m below), *Candona caudata* has a short life span of several weeks to months, which assures several generations per year, as eggs will continue to hatch throughout the year (Delorme, 1982). *Candona caudata* is found in freshwater lakes and streams, and commonly in the hypolimnia of large lakes north of the frost line, with an upper TDS threshold of 3000 mg/l (Forester, 1991). Although it can live in waters with temperatures in excess of 20°C, *Candona caudata* does not live where such temperatures persist annually, nor does it live with cryophilic or thermophilic taxa (Forester, 1991). In general, *Candona caudata* lives in freshwater to slightly saline lakes and streams, or in stream supported, seasonally saline, carbonate enriched lakes (Brouwers and Forester, 1992), having major ion chemistries dominated by Ca-HCO_3 conditions, though it is tolerant of either Ca or CO_3 enrichment (Brouwers and Forester, 1992). While *Candona caudata* is commonly found in streams, and in spring-fed streams (Taylor, 1992), it is also found at considerable depth in lakes (deep enough to have bottom currents), at the southern edge of the boreal forest and in mixed woods zone on the Canadian prairie (Delorme, 1970a, 1991).

c) *Candona decora*

Occurrence of *Candona decora* is concentrated in lakes of the mixed woods zone and southern fringe of the boreal forest: it is rare in lakes of the true prairie zone (Delorme, 1970a).

d) *Candona distincta*

Candona distincta is found commonly in shallow, warm waters of small lakes and ponds of the mixed woods zone and southern margin of the boreal forest (Delorme, 1970a). Its presence implies relative closeness of the lake margin (Delorme, 1989; Forester, unpublished, after Colman *et al.*,1990).

e) *Candona inopinata*

Candona inopinata is common in groundwater discharge seeps, ponds, wetlands, and littoral zones of lakes where groundwater discharge is prevalent. It is also eurythermic, tolerates freshwater to slightly saline waters, and requires or tolerates environmental variability (Forester, 1991). *Candona inopinata* also occurs in streams in the southeastern part of the Canadian prairie (Delorme, 1970a).

f) *Candona ohioensis*

Candona ohioensis is found in lakes of the mixed woods zone and southern fringes of the Canadian prairie (Delorme, 1970a). It inhabits permanent to ephemeral prairie lakes and ponds of fresh to slightly saline waters, where chemical and temperature variability is high (Smith, 1993; Delorme, 1991). *Candona ohioensis* is found in waters where Ca ~ HCO_3, or dilute lakes (TDS<900 mg/l) with water compositions of HCO_3 > Ca with SO_4 enrichment or dominance, or Ca > HCO_3 and SO_4 dominance (Smith, 1993). *Candona ohioensis* has an overlapping distribution with *Limnocythere itasca* (Smith, 1993). *Candona ohioensis* is found in dilute waters with slightly higher bicarbonate concentrations than sulfate concentrations (where the reverse is true for *Limnocythere itasca*) (Smith, 1993).

g) *Candona paraohioensis*

Candona paraohioensis is found in lakes at the southern fringe of the boreal forest in permanent waterbodies (Delorme, 1970a).

h) *Candona rawsoni*

Candona rawsoni is generally considered to be a pond species that lives in permanent and ephemeral waterbodies throughout the Canadian prairies. However, it also thrives with ostracodes that live in deep, cold, and dilute waters in a "Great Lakes" type of association. *Candona rawsoni* lives in waters with salinities that range from fresh/slightly concentrated (TDS concentrations about 350 mg/l or slightly less) to high salinity (Karrow *et al.*, 1995; Delorme, 1969). *Candona rawsoni* does not live in freshwater lakes that have little or no seasonal salinity variation (Forester *et al.*, 1987).

Candona rawsoni is most productive in fresh to saline ephemeral prairie lakes, and in prairie wetlands (Colman *et al*. 1990). For example, this ostracode was found in Elk Lake (Forester *et al.*, 1987) during the mid-Holocene prairie period in Minnesota, but is not found there today (Colman *et al.*, 1990). *Candona rawsoni* is found in eurytopic HCO_3 enriched and HCO_3 depleted sulfate dominated waters (Smith, 1993), and tends to thrive in lakes with seasonally changing lake levels or seasonally ephemeral bays (Forester *et al.*, 1994).

Candona rawsoni is also found in moderate abundance in Lake Michigan today (Buckley, 1975) and in the Holocene fossil record of Lake Michigan (Forester *et al*, 1994), in association with cold water ostracodes that thrive in dilute settings, such as *Cytherissa lacustris* and *Candona subtriangulata*. As a euryhaline and eurythermal species that tolerates wide ranges in physical and chemical habitats (Delorme, 1989; Forester, 1986; Colman *et al*, 1990), it is common in environments of high variability and low predictability (Forester *et al.*, 1994). Thus, because it is suited to live in environmentally variable lakes, *Candona rawsoni* is particularly good at pioneering new, predominantly mineralogenic lake basins shortly after deglaciation. These lakes, in their incipient stages, undergo large changes in hydrology, chemistry, and temperature because of seasonally variable discharges of meltwater or melting of pack ice, and variable rates of weathering of the watershed.

i) ***Candona sigmoides***

Candona sigmoides occurs in the southeastern part of the Canadian prairie, commonly in intermittent streams (Delorme, 1970a).

j) ***Cyclocypris ampla***

Cyclocypris ampla is a nektic (swimming) ostracode tolerant of wide ranges in physical and chemical environmental factors. It is found commonly in permanent ponds of moderate salinity, in moderate to high salinity eutrophic lakes with abundant organic sediments (Delorme, 1969); and in cold, fresh tundra and boreal forest lakes to slightly saline lakes in the central Canadian prairies (Forester *et al.*, 1989; Delorme, 1991). It is often is the only common ostracode in very dilute water, as salinity increases it tends to occur with other ostracode species (Forester *et al.*, 1989). Typically, it inhabits the vegetated edges of lakes and ponds (Karrow *et al.*, 1995). The presence of *Cyclocypris ampla* may indicate proximity to a delta, or a combination of flowing and standing water (Delorme, 1991).

k) ***Cypridopsis vidua***

Cypridopsis vidua is another nektic ostracode with environmental preferences similar to those of *Cyclocypris ampla*. It is found in local groundwater discharge settings, including seeps, ponds, wetlands, and the littoral zone of lakes where groundwater discharge is occurring (Forester, 1991). It is eurythermic, freshwater to slightly saline water tolerant, and it requires/tolerates environmental variability (Forester, 1991; Delorme, 1989; Curry, in press). *Cypridopsis vidua* is common throughout the interior plains of Canada (Delorme, 1970b), and it commonly inhabits the vegetated edges of lakes and ponds (Karrow *et al*, 1995).

l) *Cypris pubera*

This large ostracode was a minor component to the ostracode record of the Brokenpipe Lake site lagoon. *Cypris pubera* is indicative of a lake in its senescence stage (Delorme, 1969), and is commonly found in lakes of the interior plains of Canada (Delorme, 1970c).

m) *Cytherissa lacustris*

Cytherissa lacustris is a long lived (>2 years) species which lives in sizeable permanent waterbodies (Loffler, 1986, 1997). *Cytherissa lacustris* has been found in the littoral zone of small lakes, but these are located in Alaska, with average summer water temperatures in the littoral zone of about 11 - 13 °C (Colman *et al*, 1990). *Cytherissa lacustris* lives at moderate water depths (> 3 m) in $CaCO_3$ water compositions of low salinity (Delorme, 1968, 1969, 1970d). In a Great Lakes study, Colman *et al.* (1990) found that the presence of *C. lacustris* suggested dilute (TDS ~ 196 - 365 ppm), cold water with maximum salinity of about 215 mg/l. *Cytherissa lacustris* is a limnetic species indicative of oligotrophic conditions; in fact, it will usually disappear from profundal zone due to increased sedimentation of organic ooze during eutrophication (Loffler, 1986). *Cytherissa lacustris* has an upper threshold of survival temperature of 23°C (Colman *et al*, 1990), and it only undergoes successful growth and moulting in lab temperature conditions below 18°C (Loffler, 1997).

In general, *Cytherissa lacustris* is a good indicator of a region with a subarctic climate, with lake bottom waters undersaturated with respect to calcite (Delorme, 1970d; Delorme, 1989, after Forester *et al*, 1994). It is common in waters that show little seasonality in temperature changes, with average water temperatures below about 15 - 20 °C, and it is indicative of an area with a stenotypic boreal forest climate with precipitation > evaporation (Delorme, 1978, after Forester 1991).

n) *Heterocypris glaucus*

This nektic ostracode can be found in quite saline waters (TDS above 3000 mg/l), and is usually associated with sulfate dominated waters which are enriched in bicarbonate versus calcium (Smith, 1993).

o) *Ilyocypris bradyi*

Ilyocypris bradyi is commonly found in groundwater discharge settings including seeps, ponds, wetlands, and in the littoral zone of lakes with appreciable groundwater discharge (Forester, 1991). It is also common in the Canadian prairie, in permanent and intermittent streams, and in stream-fed lakes and ponds (Delorme, 1970d, 1991). *Ilyocypris bradyi* is commonly found in localities with a combination of flowing and standing water (Delorme, 1991), such as where streams flow into lakes. *Ilyocypris bradyi* is a eurythermic species, found in fresh to slightly saline waters, and it seems to tolerate or require environmental variability (Forester, 1991).

p) *Ilyocypris gibba*

Ilyocypris gibba is a species found in permanent streams with waters of moderate salinity (Delorme, 1969, 1991). It requires flowing water during all or part of its life cycle (Curry, in press). Like *Ilyocypris bradyi*, its presence may indicate proximity to a river mouth, or a combination of flowing and standing water (Delorme, 1991; Curry, 1997). *Ilyocypris gibba* is commonly found in the sediments of rivers and streams throughout the Canadian prairie (Delorme, 1970d).

q) *Limnocythere herricki*

Limnocythere herricki is found in waterbodies of the Canadian prairies or prairie-forest transition. These areas are typically dominated by long, cold winters, warm to cool summers, and high drought frequency; dry, subhumid conditions are common where *Limnocythere herricki* occurs (Forester *et al*, 1987). *Limnocythere herricki* commonly occurs in permanent or ephemeral freshwater lakes, ponds, or sluggish streams that undergo seasonal salinity variation, but remain fresh; this ostracode can tolerate a maximum salinity of about 1670 mg/l (Forester *et al*, 1987; Delorme, 1971). Many limnocytherids prefer fine (0.0625 mm) to coarse (1.0 mm) sand-sized particles as a substrate. This improves the flow of oxygenated waters between grains of the substrate where these ostracodes are found, thus they are less commonly found in lakes with clayey substrates (Delorme, 1991).

r) *Limnocythere itasca*

Limnocythere itasca is common in lakes of the boreal forest and parkland areas of the interior plains of Canada (Delorme, 1971a). It is commonly found in fresh to saline shallow, warm waters of lakes and ponds or in permanent to ephemeral prairie lakes and ponds. Seasonal variability in the chemical and temperature values for the lake is often large (Forester, 1991). The presence of *Limnocythere itasca* may imply relatively closer proximity to the lake margin (Delorme, 1989; Forester, unpublished, after Colman *et al*, 1990). *Limnocythere itasca* prefers waters with Ca ~ HCO_3, or dilute lakes (TDS<900 mg/l) with water compositions of HCO_3 > Ca with SO_4 enrichment or dominance, or Ca > HCO_3 and SO_4 dominance (Smith, 1993).

4.2 - THE RUBY LAKE SITE: INTRODUCTION

Three study sites in the Hudson Bay, Saskatchewan area -- Ruby Lake (labeled RL Fig. 4-1), Hudson Bay Upper, and Hudson Bay Lower -- all lie behind Campbell strandlines of Lake Agassiz (Fig. 4-6). Like the Brokenpipe Lake region, this locality is also a former bay of Lake Agassiz. It is situated in a low-lying area surrounded by the Porcupine Hills of Manitoba and Saskatchewan in the south, and the Pasquia Hills of Saskatchewan to the north (Fig. 4-1). The town of Hudson Bay, Saskatchewan (Fig. 4-6) is situated directly atop the Upper Campbell beach. In the Hudson Bay and surrounding areas, the Upper Campbell beach is a very large, well-defined sand and gravel berm with fore-beach stratification (Fig. 4-7).

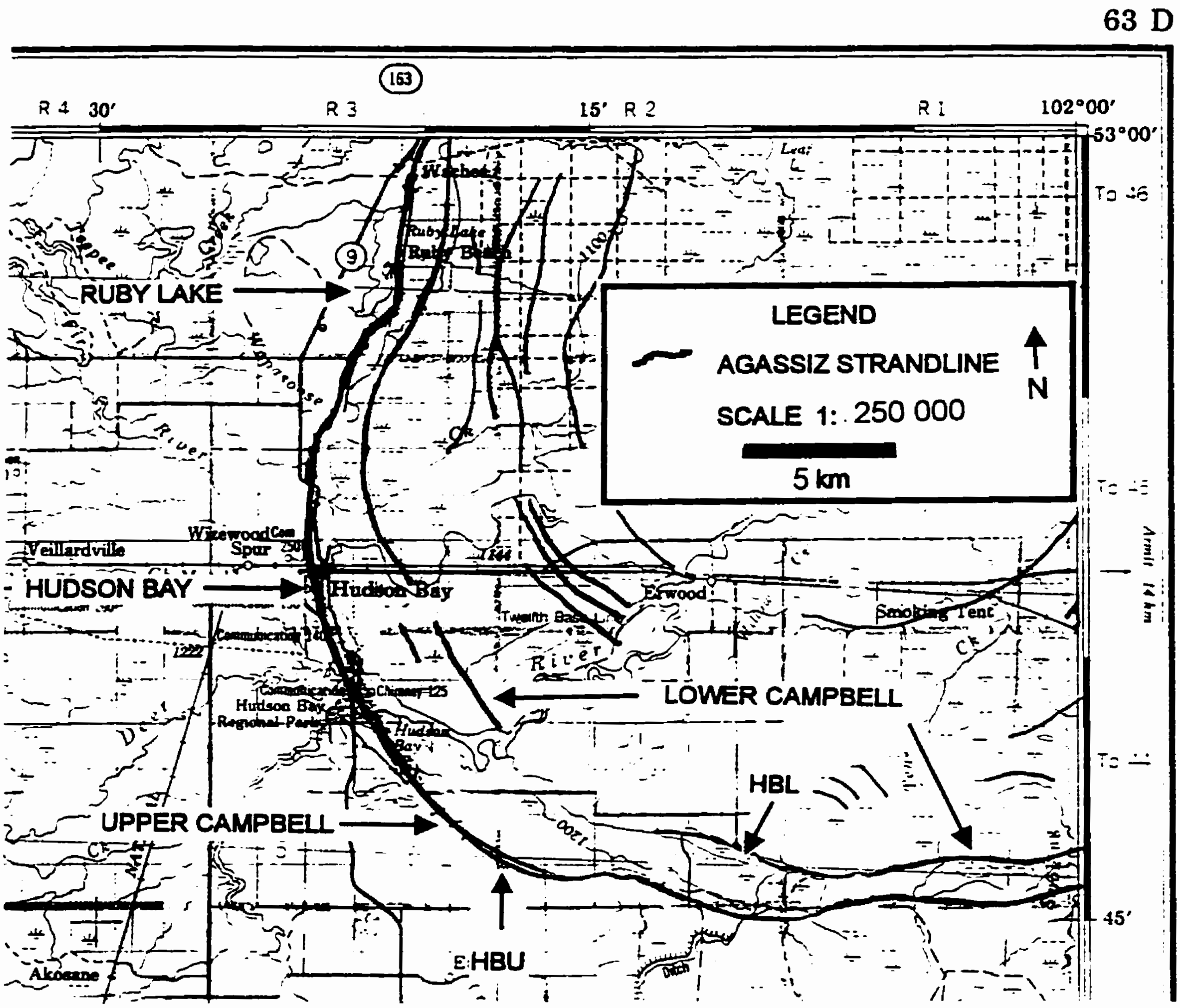


Figure 4-6: Lake Agassiz beaches in the Hudson Bay, Saskatchewan area. Note the locations of the HBU and HBL core sites, and the location of Ruby lake. Beaches drafted from Campbell (1987).

4.2.1 - Geomorphological and Sedimentological Observations

Aside from the excellently developed Upper Campbell beach at about 374 m elevation (Fig. 4-7), and Lake Agassiz beaches below the Upper Campbell level (Moran 1969), there is little geomorphological evidence for interpreting the history of Lake Agassiz in the Hudson Bay area. According to Moran (1969), the westernmost limit (highest elevation) beaches and glaciolacustrine deposits in the Hudson Bay region, which were deposited in Lake Agassiz, are marked by the well-developed Upper Campbell beach ridge (Fig 4-6). The lacustrine sediment and water-washed eroded till plain that occurs outside of the Lake Agassiz basin -- at elevations above the Upper Campbell strandline -- in the Swan (southeast of Hudson Bay) and Red Deer River (west - southwest of Hudson Bay) lowlands is attributed to a period of rapidly falling levels of other proglacial and supraglacial lakes immediately preceding the inundation of the Hudson Bay area by Lake Agassiz (Moran, 1969). During the occupation of the Hudson Bay area by these lakes, and by Lake Agassiz, meltwater associated with stagnant ice atop the Porcupine Hills and Pasquia Hills flowed off the uplands and deposited outwash on the lowlands. Some of this sediment also was deposited into the lakes, contributing to glaciolacustrine sedimentation (Moran, 1969).

4.2.2 - Cores From the Ruby Lake Area: Description

Four cores were sampled in the Ruby Lake area; two cores (RLA, RLB) from behind the Upper Campbell beach in Ruby Lake (Fig. 2-6), and two others from unnamed ponds behind the Upper Campbell (HBU core site) and Lower Campbell (HBL core site) beaches, respectively (Fig. 4-6). Only the RLA and HBL cores will be discussed in detail. An annotated description of the stratigraphy of the HBL and RLA cores is given in Tables 4-C and 4-D; a detailed description is in Appendix A. Photos of some features of these cores are shown in Figures 4-8 a-c.

Ruby Lake has an irregular bathymetry (Fig. 4-9), with the deepest portion of the lake directly west of the Upper Campbell strandline. The linear morphology of the bathymetry of Ruby Lake suggests that it is a drowned river channel that was dammed by the formation of the Upper Campbell strandline. This hypothesis is consistent with the observation that the easterly drainage of many rivers in the Hudson Bay area are controlled by the Lake Agassiz strandlines (Fig. 4-6) continue to flow down gradient, east of the Upper Campbell strandline.

Because the base of this "channel" (where the RLA core is located) is clay and weathered shale (98 - 123 cm interval of the RLA core; Fig 4-9, Table 4-D), and the same material comprises the entire sequence (aside from a thin layer of recent organics at the top of the section) in the RLB core (Fig. 4-9; Appendix A), which is located outside of this deep hole, it appears that the clay and weathered shale forms the basal unit in the stratigraphy of this site. The clay and weathered shale that occurs at the base of the succession may be weathered bedrock. However, the presence of dropstones and mm-sized granules in the matrix of the unit argues against this possibility. Thus, the shape of the bottom of the lake is more likely a reflection of the underlying uneven till surface, which was later draped by the glaciolacustrine deposits of clay and weathered shale found

Table 4-C: Stratigraphic Units in Core HBL

Depth Interval (cm)	Description
0 - 86	fine peat, mollusc shells throughout, distinct bottom contact
86 - 110	fg - mg sand, occasional 2 - 5 mm-sized pebbles, dark in color due to mixing of peat throughout unit, lower contact distinct
110 - 145	gravel, sub to well-rounded clasts commonly 3 mm to 3+ cm in size, vcg sand matrix, distinct lower contact
145 - 168	vfg sandy-silt, poorly laminated except for distinct alternating light (2-3 mm thick, silt) and dark (3 mm thick, more sandy) laminae in upper 10 cm of unit, distinct lower contact
168 - 190	silty clay matrix with common mm-sized granules, clasts of pieces of weathered shale bedrock, two 1 cm sized pebbles found in the unit

Table 4-D: Stratigraphic Units in Core RLA

Depth Interval (cm)	Description
0 - 1	mg-fg peaty sand, top of unit has a high % of peat, lower contact distinct
1 - 12	gravel with peat beds, upper 4 cm of unit is vfg sandy peat; 4-9 cm is gravel with one 4 mm thick peat laminae, subrounded clasts several mm to 1 cm in size, with vcg sand matrix; 9-12 cm is peat, distinct basal contact
12-56	40-56 cm is vcg sand with frequent mm-sized granules and occasional cm-sized clasts, grading upward to (25-40 cm) gravel, mostly mm-sized granules and vcg sand matrix with several cm-sized subrounded clasts, grading upward (12-25 cm) into gravel, subrounded clasts 0.5-1 cm in size (one is 4cm in dia.), vcg sand matrix, lower contact distinct
56 - 78	mg-cg sand with occasional mm-sized granules, very poorly bedded with gravel beds at 77-78, 69-71, and 57-59 cm; gravel has well rounded 0.5-1 cm clasts in a vcg sand and mm-sized granule matrix, two 3 mm thick vfg sandy silt laminae at 56 and 65 cm, lower contact distinct
78 - 95	silty vfg sand with a gravel layer at 80 - 84 cm, gravel clasts are 0.5 cm in size, subrounded, matrix is vcg sand, two vfg sandy silt layers 3-4 mm in thickness are found at 89 and 92 cm, unit has yellow-orange staining which may reflect surface and groundwater flow through the unit
95 - 98	vfg sandy silt, vegetal detritus laminae at 97 cm
98 - 123	silty clay, blocky structure, blocks are relatively stiffer and smoother than the matrix, one 1 cm-sized stone found near core tube edge in the clay at 102 cm, matrix has frequent mm-sized granules, unit represents weathered shale fragments in a clay matrix

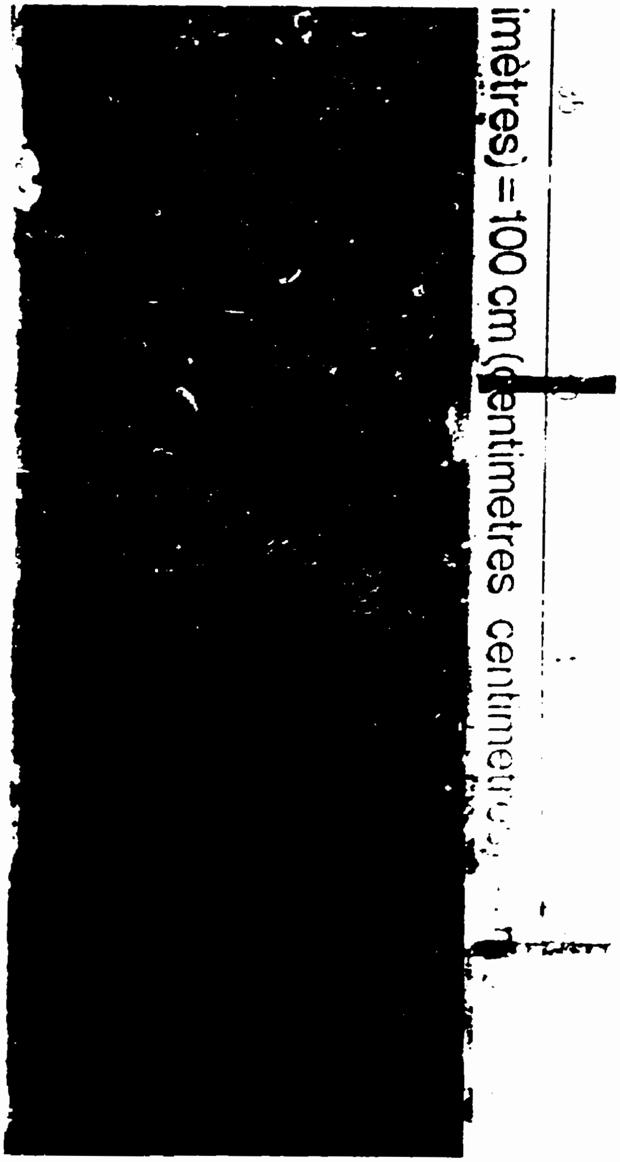


Figure 4-8a: Notice the deformation of sandy silt / gravel contact in the HBL core due to the vibracoring process. For the HBL core (Fig. 4-8 a and b) and the RLA core (Fig. 4-8 c), see Figures 2-6, 4-6,and 4-9 for core locations. See Table 4-C for HBL unit descriptions,and Table 4-D for RLA unit descriptions. Distance between heavy dark lines on scale is 10 cm in all photos.

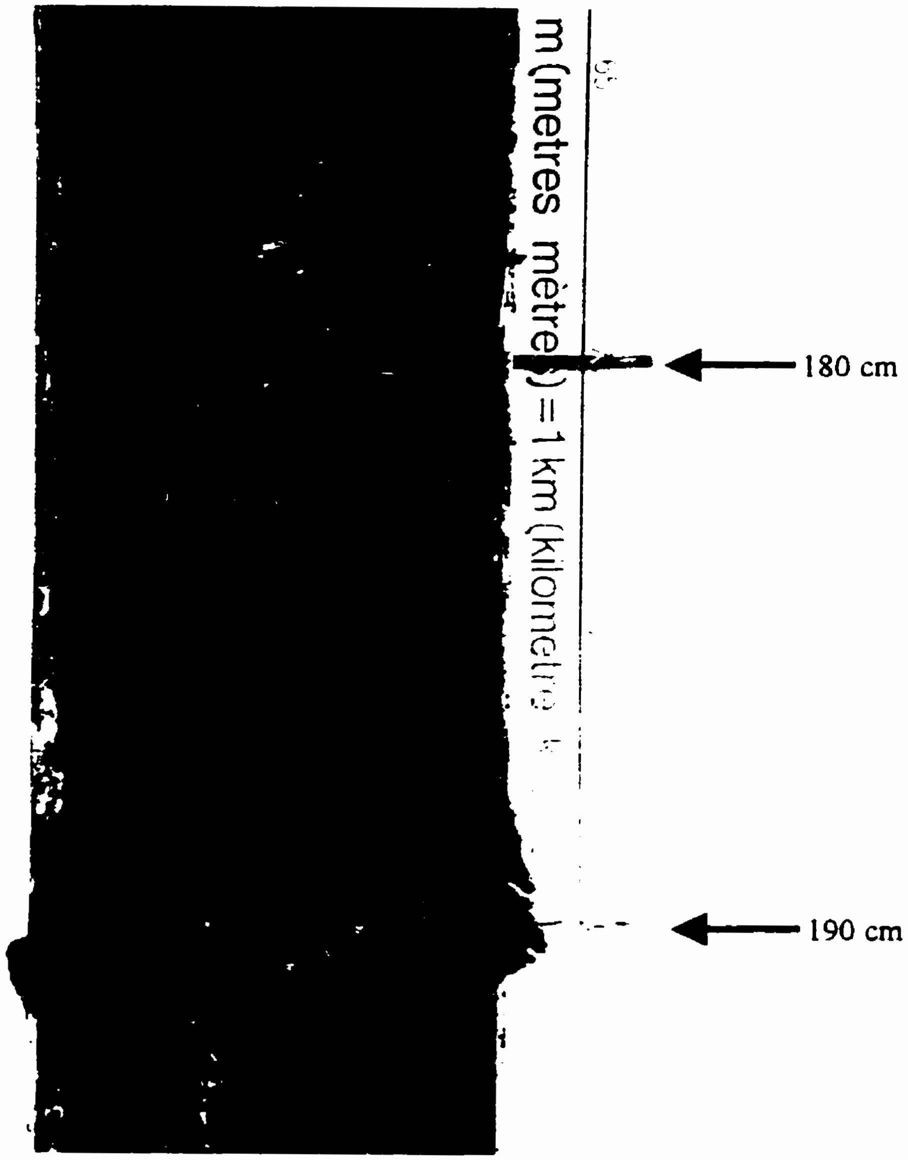


Figure 4-8b: Weathered shale clasts in a silty clay matrix at the base of the HBL core; matrix also contains mm-sized granules.

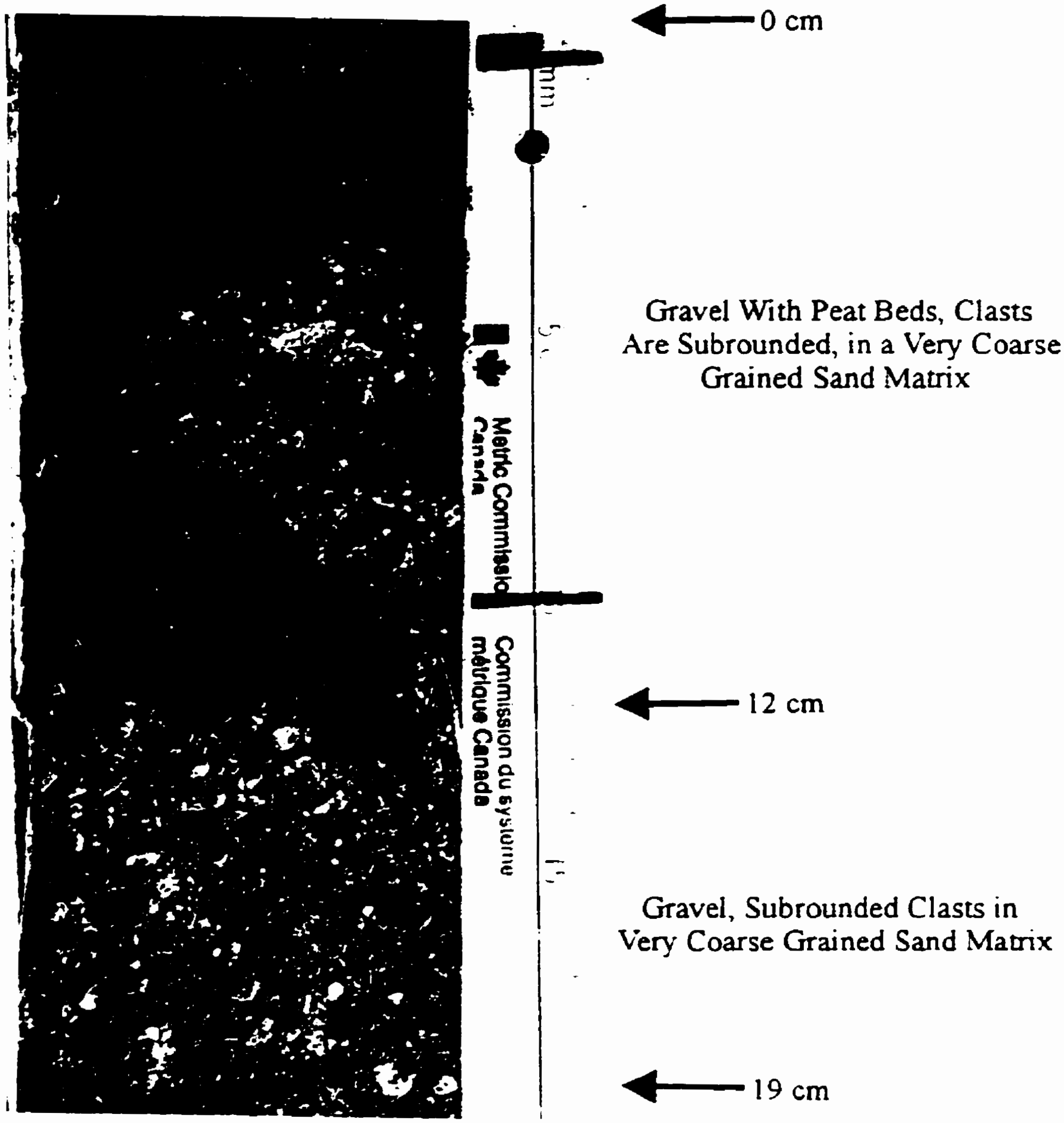


Figure 4-8c: Gravel in the upper portion of the RLA core.

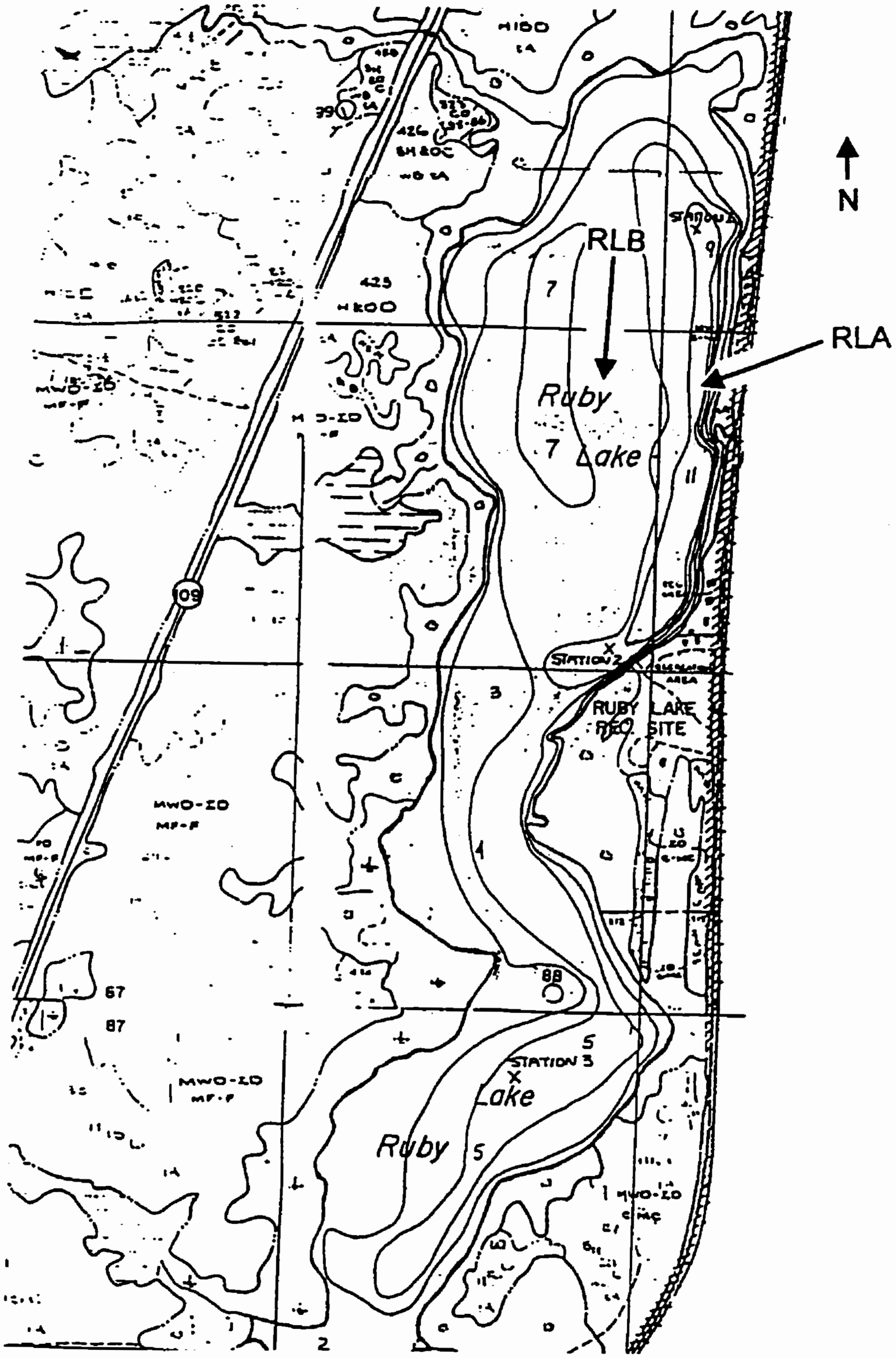


Figure 4-9: Bathymetric map of Ruby Lake. Contours are in feet (2 foot intervals). The lake is located 8 km north of the town of Hudson Bay (see Fig. 4-6). Note the locations of the RLA and RLB cores. (map drafted by P. Rosen and J. Crabb, 1987)

at the base of the RLA core and in the RLB core. Till occurs in the areas surrounding Ruby Lake itself (Stonehouse and Ellis, 1983; Moran, 1969).

4.3 - THE JAY JAY LAKE SITE: INTRODUCTION

Like both Brokenpipe and Ruby Lakes, Jay Jay Lake lies in a low area situated between relatively high elevation bedrock escarpments. The Cub Hills are located to the south, and the Wapawekka Hills to the north (Fig. 4-1) of Jay Jay Lake. Directly east of Jay Jay Lake is the Upper Campbell strandline, and the Lake Agassiz basin. Langford (1973) produced a detailed surficial geology map of this area (the 73I NTS mapsheet) at a scale of 1:250 000. Langford (1973) noted that there are no features demarcating the position of the retreating Laurentide Ice Sheet. Instead, he suggested that progressive thinning of active glacial ice left behind the many sub-glacial and ice-marginal meltwater features found in lowland areas (Langford, 1973). As ice thinning progressed, the Wapawekka/Cub Hills region would eventually be covered by isolated patches and thin lobes of ice in the lowland areas (Langford, 1973). The area east of the Cub Hills would have a slightly different history, as ice retreat in that area was directly followed by the northward expansion of Lake Agassiz.

4.3.1 - Geomorphological and Sedimentological Observations

The following discussion of geomorphological features only includes those directly related to forming a history of deglaciation and site evolution in the Jay Jay Lake area. The numbers of each observation listed below is duplicated in small boxes in Figure

4-10 to show the locations of these features.

According to Langford (1973), ice marginal channels are identified by their orientation parallel to topographic contour, while subglacial channels will tend to cross elevation contours. In some cases, subglacial channels in the Jay Jay Lake region have orientations and geometries such that water flowing through them would have been forced up-gradient by ice sheet surface hydraulic head (Langford, 1973). For example, the meltwater channel labeled 1a (Fig. 4-10) was formed subglacially (Langford, 1973). As the ice sheet wasted, exposing more of the top of the Cub Hills, an ice-marginal channel formed (labeled 1b, Fig. 4-10) that drained into the preexisting subglacial channel (1a).

The channels located at 1c (Fig.4-10) are ice marginal meltwater channels. Langford (1973) considered that these channels, which follow the elevation contour closely, formed against a lobe of ice that persisted between the Wapawekka and Cub Hills during ice wasting and retreat in that area. The meltwater from these channels flowed eastward, underneath the main ice sheet (Langford, 1973). The channels located near the Dowd Lakes, labeled 1d (Fig. 4-10), probably began as sub-glacial channels (Langford, 1973). However, the large lobed outwash plains at the terminus of each channel suggests that they later carried deglacial meltwater as subaerial channels, that drained into a body of standing water (Langford, 1973).

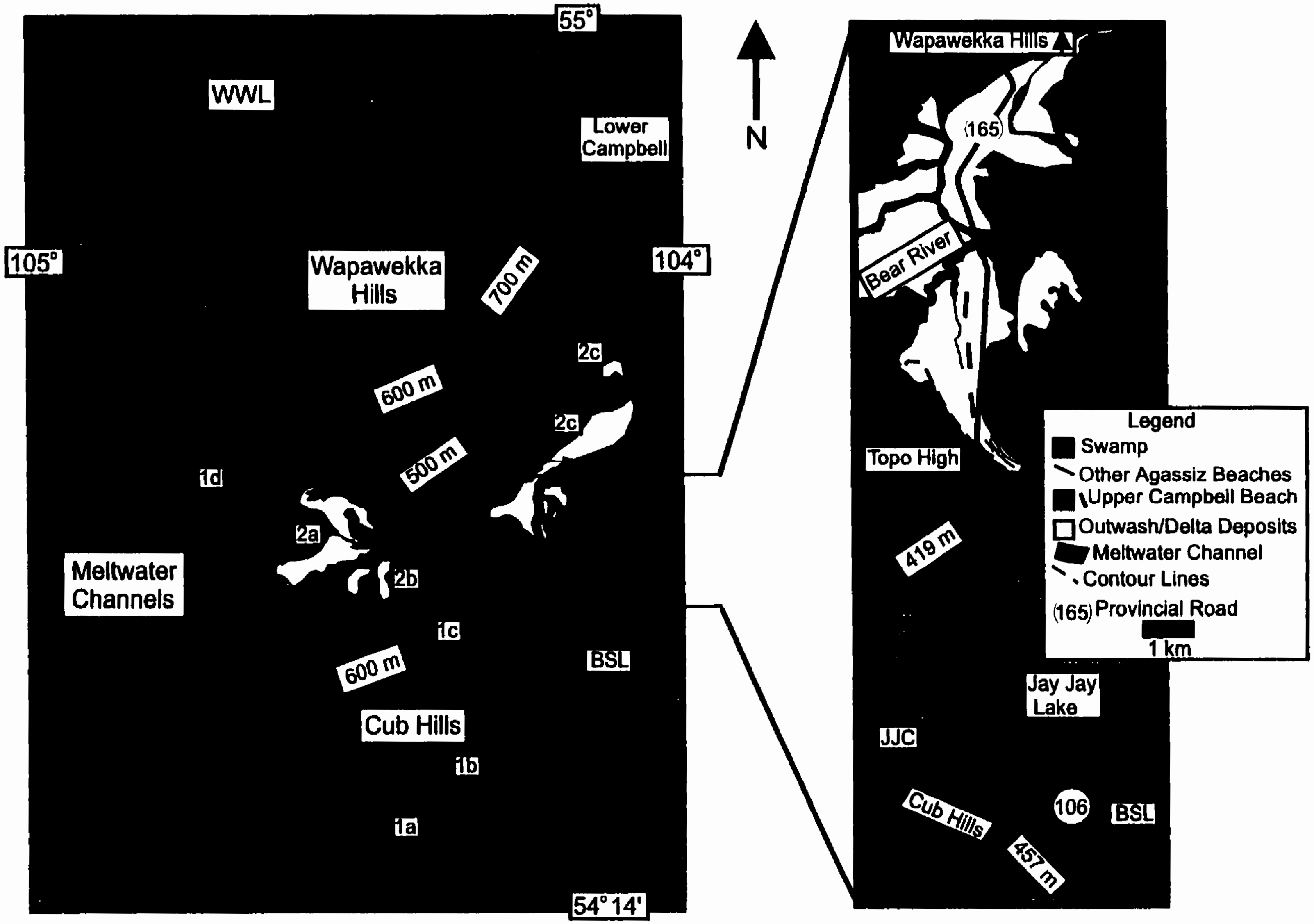


Figure 4-10: Geomorphological features of the modern Jay Jay Lake area. Diagrams after Langford (1973). Numbers in boxes refer to features discussed in the text. Note the location of the JJC core in the diagram at right. BSL = Big Sandy Lake

In the vicinity of Jay Jay Lake there are several outwash plains that are important to the deglacial history of the region, and to the incipient stages of Lake Agassiz. The lobate outwash plains at the terminus of each of the Dowd Lakes channels, labeled 2a (Fig. 4-10) were likely deposited into a body of water, because the distal edges of these plains are lobed and wave cut (Langford, 1973). Langford (1973) hypothesized that the body of water was a short-lived, higher (than the Upper Campbell) stage of Lake Agassiz. The lack of any strandlines or other more geographically extensive high-elevation outwash plains or deltas in the Jay Jay Lake area suggests that the body of water into which the Dowd Lakes outwash was deposited was less expansive than Lake Agassiz, trapped by ice against higher elevation land to the west. Two smaller outwash deposits, labeled 2b (Fig. 4-10), located near the Dowd Lakes outwash plains, are at a slightly lower elevation than those outwash plains connected to the Dowd meltwater channels (2a, Fig. 4-10). The location of these smaller deposits at a slightly lower elevation is a reflection of a lowering lake level from that present when the Dowd Lakes (Fig. 4-10) outwash plains were deposited. Another set of extensive outwash deposits -- (labeled 2c Fig.4-10) or deltaic deposits (cut by the Bear River, Fig. 4-10) -- mark the edge of the Lake Agassiz basin, where the steep topographic gradients of the Wapawekka and Cub Hills decline and dip gently basinward into Lake Agassiz. The sinuosity of the Bear River, north of Jay Jay Lake (Fig. 4-10) shows this gradient change quite clearly. The river is contained in a slightly meandering channel as it makes its way to lower elevations from the topographically high Wapawekka Hills. However, as the Bear River crosses into the Lake Agassiz basin (east of provincial road 165, Fig. 4-10), the sinuosity increases, distinguished by many meander cut-offs and oxbow lakes -- a characteristic feature of rivers with extremely low gradients.

The glaciofluvial outwash deposits commonly consist of less than a meter of sand over poorly stratified sand and gravel. It is important to note that since the deposition of these glaciofluvial outwash plains appears to have occurred at a break in slope, and in absence of detailed sedimentological information, it is unclear whether these outwash plains were emplaced into Lake Agassiz as deltas, or were deposited subaerially as glaciofluvial outwash plains which were later modified by Lake Agassiz. However, the outwash plains in question at the Bear River (north of Big Sandy Lake -- Fig. 4-10) do have lobed delataic margins that suggest deposition in water (Langford, 1973).

The Upper and Lower Campbell beaches form distinct shorelines in the Wapawekka Hills area. In the vicinity of Jay Jay Lake, the Upper Campbell forms a large baymouth bar which at one time separated the main body of Lake Agassiz from a former bay of Lake Agassiz (located between the Wapawekka and Cub Hills), the area within which Jay Jay Lake is found today (Fig. 4-10). This large sand and gravel bar is flat topped and several tens of meters wide, and is composed of well-sorted flat-bedded top-beach sand (Fig. 4-11). Progradation of this large baymouth bar was likely from north to south. A reasonably large source of material for the construction of the beach is located

Figure 4-11: Well-sorted, planar bedded medium to coarse grained sand in a small cut in the upper part of the Upper Campbell beach near the Jay Jay Lake site.

north of Jay Jay Lake; the outwash plains at 2c (Fig. 4-10) and glaciofluvial outwash (delta?) at the mouth of the Bear River (Fig. 4-10) probably served as the headland for erosion and were reworked and transported southward to build the beach. The morphology of the Lower Campbell beach east of the Wapawekka Hills, being arcuately curved and flared towards the southwest, also suggests the dominant direction of longshore current in the northern part of the Jay Jay Lake area was from the north east to the southwest. This interpretation also fits with the evidence and reasoning presented by Nielsen *et al.* (1984) that the circulation of another fairly constricted, western lake margin Lake Agassiz bay was anti-clockwise.

4.3.2 - Cores From Jay Jay Lake: Description

Five cores were taken from the Jay Jay Lake site (see Fig. 2-6 for core locations). None of the cores encountered Lake Agassiz clays, till or diamicton, which was expected beneath the Lake Agassiz sequence. Thus, it appears that none of the cores penetrated completely through the post-Lake Agassiz sediment sequence. After comparing the relatively variable stratigraphy each of the cores with one another, core JJC (Figs. 2-6, 4-10, Table 4-E) appears to have the most conformable post-Lake Agassiz sequence at this site (Fig 4-12).

The cores from Jay Jay Lake generally include a surficial organic sequence about 1 m thick, with the rest of the core varying from uncorrelatable beds of silty fine sand to clayey silty fine sand in generally irregular successions (see full descriptions of all Jay Jay cores in Appendix A). An overall photo of the JJC core is given in Figure 4-13.

Table 4-E: Stratigraphic Units in Core JJC

Depth Interval (cm)	Description
0 - 10	fine organic sediment (humified peat), fine rootlets common, slightly calcareous in lower part with increasing lighter coloring (like unit below), occasional molluscs, lower boundary gradational over 3 cm
10 - 56	gyttja, increasingly calcareous downward, gastropods and seeds common, no siliciclastic grit, lower boundary distinct but gradational over 1 cm
56 - 75	very clayey silt with coarse sand, no laminae visible, non-calcareous, grading downward into clayey silt with fine sand
75 - 81	slightly clayey and sandy silt, no visible laminae, very few coarse sand grains, non-calcareous, lower boundary gradational
81 - 121	vfg-mg sand, silty, laminated, non-calcareous
121 - 312	silt to clayey silt and vfg-mg sand, bedded and laminated, zones of siltier and sandier laminae vary from 1 to 30 cm in thickness (no systematic variation), lower boundary distinct
312 - 332	silt to very fine sand, poorly laminated, lower contact distinct
332 - 336	clayey silt, 2 mm organic-rich laminae at top and occasional thinner organic laminae below, lower boundary distinct
336 - 350	clayey silt, more clayey than overlying unit, laminae (1 -2 mm) of vfg-mg sand throughout, lower boundary distinct
350 - 391	fg-mg sand and silty vfg sand, laminated, zones of silty versus sandy laminae < 0.5 - 4 cm, coarsest laminae in lower 6 cm of interval, lower contact distinct
391 - 400	silty fine sand, laminated

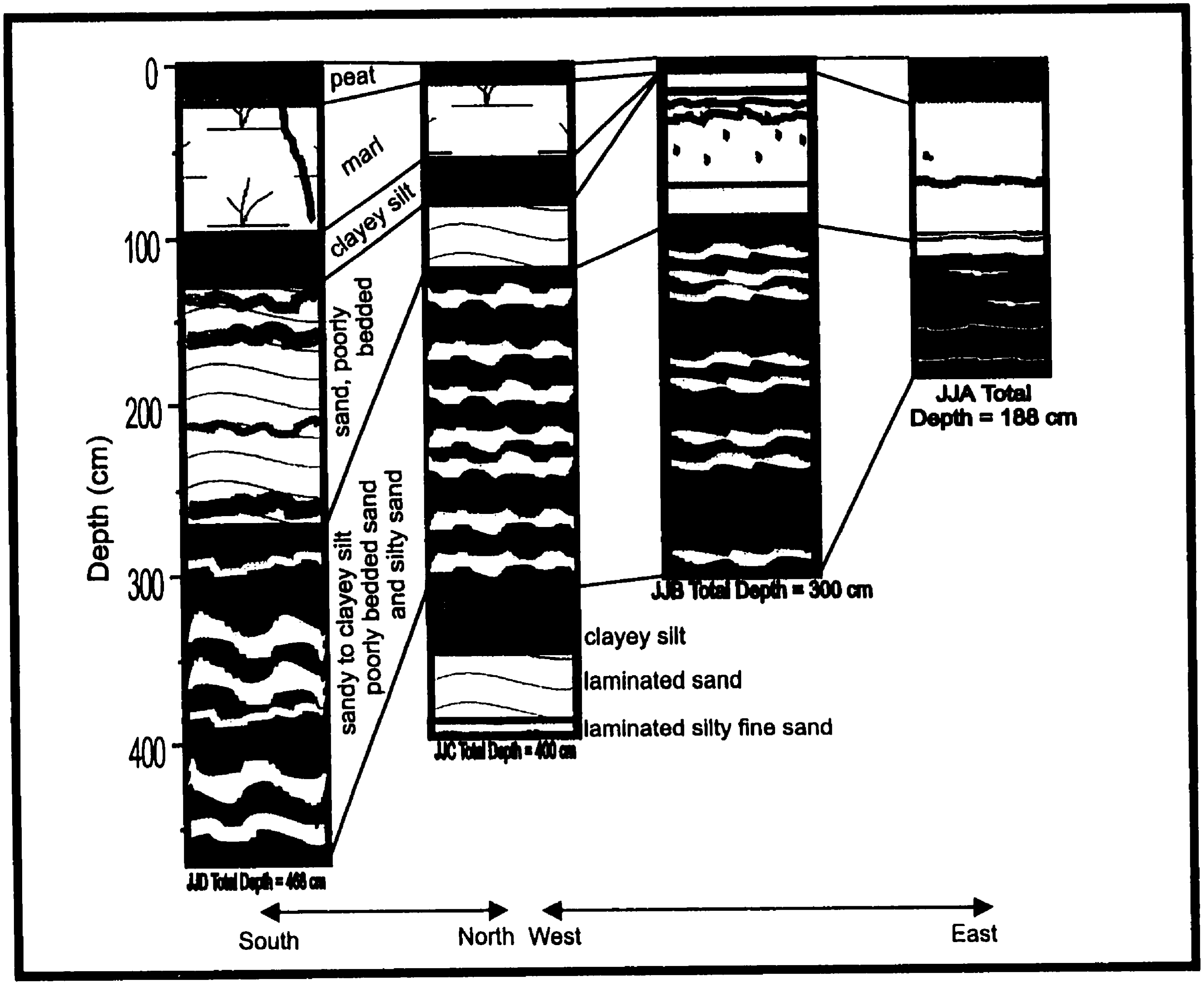


Figure 4-12: Correlation of the JJA, JJB, JJC and JJD cores from Jay Jay Lake. Relative core locations shown in Figure 2-6, see Appendix A for complete core descriptions.

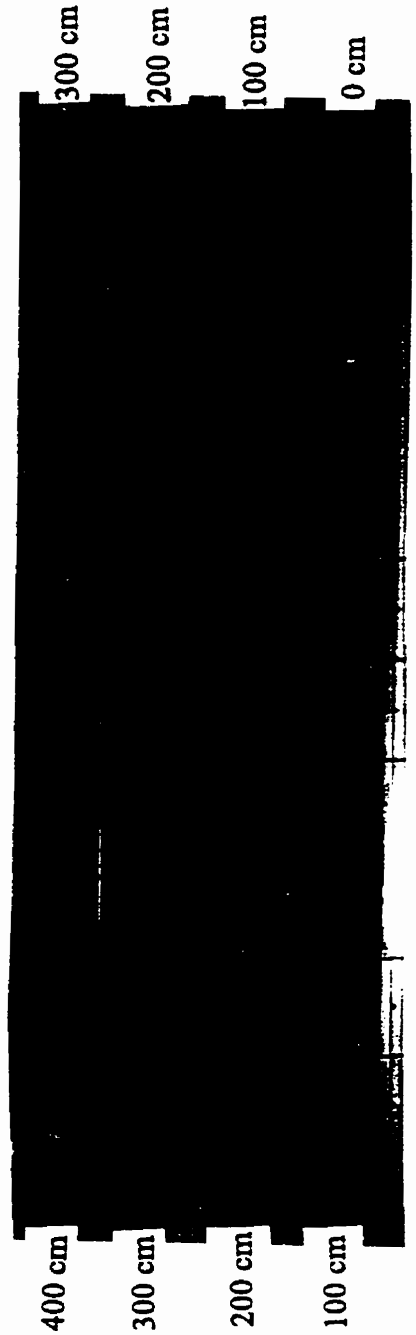


Figure 4-13: Overview of the JJC (Jay Jay Lake) core, illustrating the irregular succession of alternating massive to poorly bedded and laminated sand (light) and massive to poorly bedded and laminated clayey silty sand (dark gray) common in all of the Jay Jay lake cores. Sand size varies from coarse to fine, and it is poorly to well sorted. Note the deformation of contacts due to the vibracoring process, and the sample extraction sites (dark bands). The location of the JJC core is shown in Figures 2-6 and 4-10. Table 4-E provides a description of the core. Distance between heavy dark lines on the scale is 10 cm.

4.3.3 - Description of the Moisture, Organic, and Carbonate Trends in Core JJC

a) Moisture

Moisture content (Fig. 4-14) is high in the upper peat and marl portion of the JJC core. Decreasing moisture content with increasing depth in the marl unit probably represents decreasing available pore space due to smaller grain size and increased bulk density (compaction) of the gyttja-marl. Moisture content drops to below 20% and stays as such throughout the mineralogenic (clay, silt, and sand; 56 - 400 cm) portion of the core.

b) Organics

Not surprisingly, organic content is highest in the peat (0 - 10 cm; Table 4-E) of the JJC core (~ 20 - 48%; Fig. 4-14). Organic content steadily decreases through the gyttja-marl zone (10 - 56 cm; Table 4-E). Organic content is very minor in the mineralogenic portion of the JJC core (56 - 400 cm).

c) Carbonate

Carbonate values (Fig. 4-14) are low in the peat (0 - 10 cm), but increase dramatically to more than 30% in the gyttja-marl unit (10 -56 cm; Table 4-E). The high carbonate content of the 10 - 56 cm zone is a result of the biogenic production of marl carbonates in Jay Jay Lake. Carbonate content drastically decreases in the 56 - 400 cm interval. Periodic increases in carbonate content in the lower 56 - 400 cm interval of the JJC core are likely the result of a relatively higher proportion of carbonate silt in those

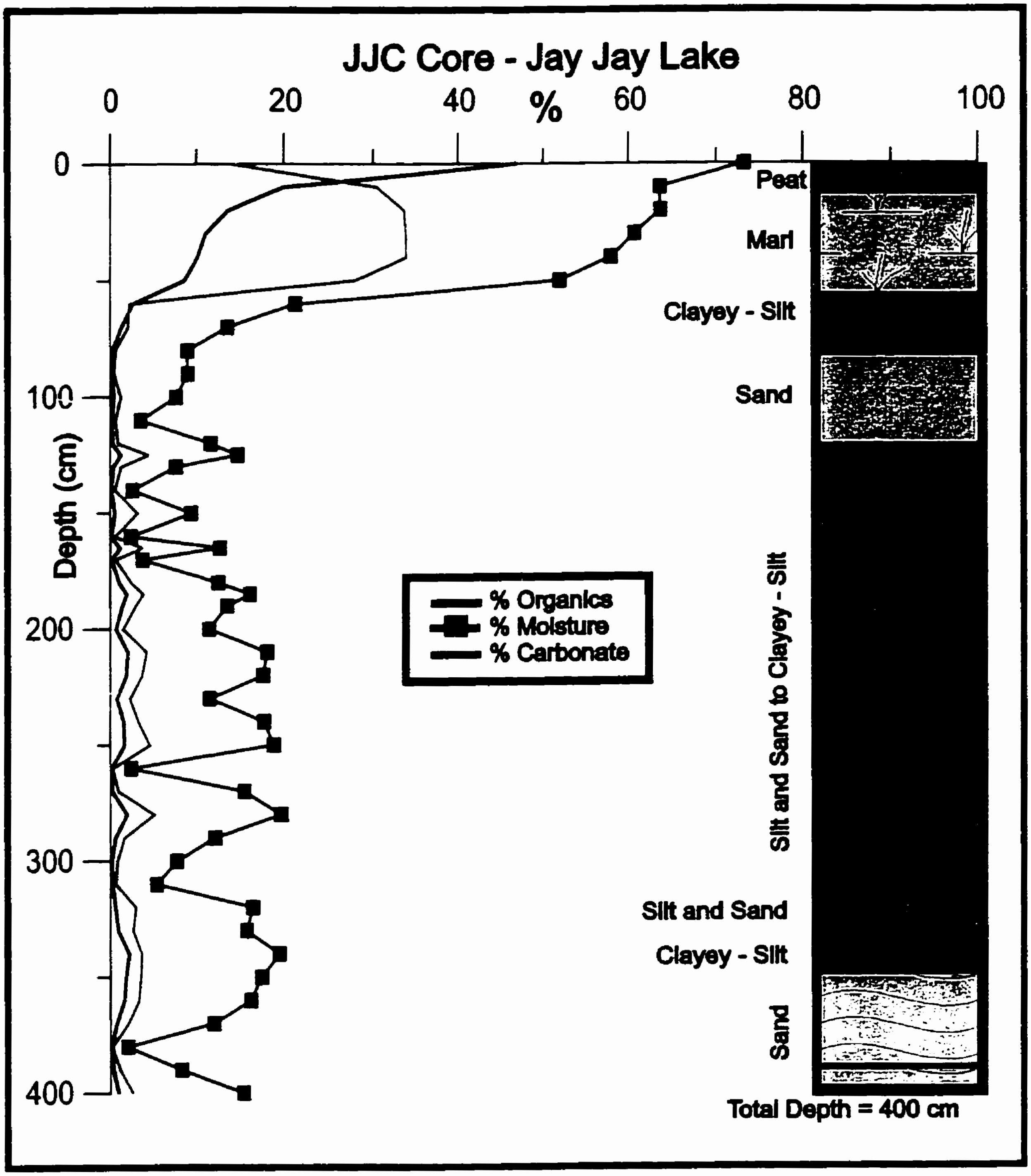


Figure 4-14: Moisture, organic, and carbonate percentage values for the JJC core (Jay Jay Lake). Squares indicate sample depths. Organic matter and carbonate contents were calculated by LOI from a dry sample. Generalized stratigraphic section at right.

samples. Interestingly, in the lower 56 - 400 cm interval of the JJC core, the moisture, carbonate and organic curves are covariant. This implies that the relatively siltier zones of the core not only contain more moisture, but they contain relatively more organic material as well. This generality was usually confirmed when looking through samples (under a reflected light microscope) for ostracodes.

4.3.4 - Fossil Ostracodes From Core JJC

Some preliminary ostracode identification from the JJC core was initially done by Brandon Curry (1996, personal communication). Most likely due to mechanical abrasion and destruction by the fine to medium grained sand and silts of the JJC core, or perhaps due to dissolution resulting from deposition in waters undersaturated with respect to calcite (perhaps in conjunction with slow sedimentation rates), ostracode preservation was poor in the JJC core. If a period of relatively low salinity occurred under calcite saturation (e.g. a dilution triggered by a change in hydrology), conditions required for the dissolution of calcite are the result.

Ostracode subsamples (Table 4-F) taken from lower sections of the JJC core (230 - 370 cm; see Table 4-E and Appendix A for core description), usually in clayey silty sand, revealed an assemblage of *Candona rectangulata* , *Heterocypris glaucus*, and *Limnocythere herricki*. *Candona rectangulata* is characteristic of a tundra setting (Brandon Curry, 1996, personal communication), and is today found in lakes north of 60^{o} N (Maher *et al.*, 1998). *Heterocypris glaucus* and *Limnocythere herricki* are both found today in prairie lakes and ponds, and indicate conditions when E>P (Forester *et al.*, 1987;

Table 4-F: Fossil Ostracodes Found in Core JJC

Depth (cm)	Ostracode Species
0	*Cyclocypris ampla*, ostracode juveniles (candonids?), unidentified mollusc fragments
10	*Cyclocypris ampla*, *Candona ohioensis*, *Candona acutula*, ostracode juveniles (candonids?), unidentified mollusc fragments
20	*Cyclocypris ampla*, *Candona ohioensis*, *Candona acutula*, ostracode juveniles (candonids?), unidentified mollusc fragments
30	*Cyclocypris ampla*, *Candona ohioensis*, *Candona acutula*, ostracode juveniles (candonids?), unidentified mollusc fragments
40	*Cyclocypris ampla*, *Candona ohioensis*, *Candona acutula*,*Cypridopsis vidua*, ostracode juveniles (candonids?), unidentified mollusc fragments
50	*Cyclocypris ampla*, *Candona ohioensis*, *Candona acutula*, *Cypridopsis vidua*, ostracode juveniles (candonids?), unidentified mollusc fragments
60	*Ilyocypris gibba*, *Cyclocypris ampla*, *Cypridopsis vidua*, *Candona acutula*, unidentified mollusc fragments
70	unidentified mollusc fragments
100	unidentified mollusc fragments
230	ostracode juveniles (candonids?)
240	ostracode juveniles (candonids?), some partially dissolved
250	*Limnocythere herricki*
280	*Cyclocypris ampla*, *Cypridopsis vidua*
340	*Candona rectangulata*
370	*Heterocypris glaucus*, *Candona rectangulata*, *Limnocythere herricki*

Note: Ostracodes were sampled every 10 cm from 0 to 400 cm depth in the core, gaps indicate barren zones.

Smith, 1993). In particular, *Heterocypris glaucus* is characteristic of a more saline, sulfate dominated, bicarbonate>calcium environment (Smith, 1993). More detailed information the autecology of some of these ostracodes is given in Section 4.1.5.

The ostracodes found in the upper portion (0 - 60 cm; see Table 4-E and Appendix A for core descriptions) of the JJC core include *Cyclocypris ampla*, *Candona ohioensis*, *Candona acutula*, *Cypridopsis vidua*, and *Ilyocypris gibba* (Table 4-F). The autecological data for all of these ostracodes is listed in Section 4.1.5. The 60 - 230 cm interval of the JJC core was barren of ostracodes.

4.4 - THE GREGORY LAKE SITE: INTRODUCTION

Gregory Lake is located in an area that consists of ground moraine (Simpson, 1988), characterized by relative topographic highs and intervening lake and bog-filled depressions (Fig. 4-15). Unlike Brokenpipe, Ruby, and Jay Jay Lakes, Gregory Lake (Fig. 4-1) is not located in a former bay of Lake Agassiz, nor is it bordered by any significant uplands. The Precambrian Canadian Shield lies about 20 km due north of Gregory Lake. Gregory Lake is positioned behind a well developed strandline that continues west of the Wapawekka Hills (Fig. 4-1), toward the northwestern arm of Lake Agassiz. It is unclear whether this strandline belongs to the Upper or Lower Campbell level of Lake Agassiz (see next chapter).

The Campbell strandline is an erosional (versus constructional) feature in the Gregory Lake area, likely due to relatively increased exposure along this section of the Lake Agassiz basin. It is a small scarp eroded into a ridge of ground moraine that is

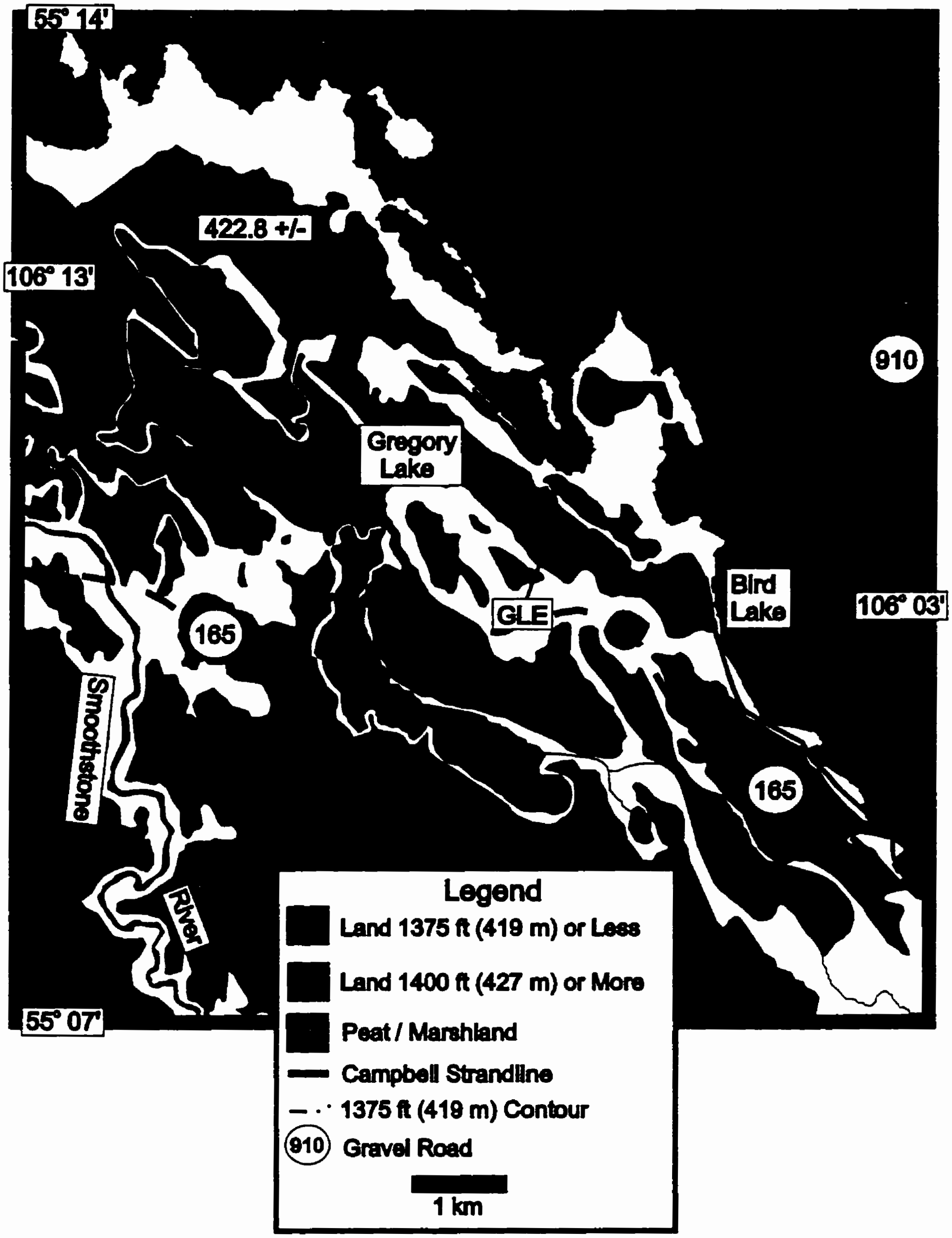


Figure 4-15: Geomorphological features of the modern Gregory Lake area. The region is characterized by ridged and hummocky ground moraine. Note the location of the GLE core. See text for details.

located southeast of Gregory Lake and east of "Bird" Lake, to the eastern side of Gregory Lake (Fig. 4-15). The strandline near Gregory Lake is at an elevation of ~ 422 m (Rayburn *et al.*, 1997; Rayburn, 1997), and is nearly continuous from the west side of the Wapawekka Hills (Fig 4-1), about 35 km away. Northwest of Gregory Lake, into the northwestern arm of Lake Agassiz, glaciolacustrine strandlines become a complex array of hundreds of small scraps that have been mapped by Fisher (1993). The elevation of these strandlines is difficult to determine due to their inaccessibility, and therefore it is difficult to correlate them to any known level of Lake Agassiz from the Wapawekka Hills/Gregory Lake areas.

4.4.1 - Cores From Gregory Lake: Description

After considerable probing and reconnaissance of the Gregory Lake (see Fig. 2-6 and Table 2-A), one site was cored. This vibracore was taken at the GLE site (Figs.2-6, 4-15), the deepest area found in the lake. From bottom to top, the general succession includes about 1 m of silty clay with dropstones, overlain by about 4 m of sand, silt, and clayey silt, which is in turn capped by about 2 m of peat (Table 4-G). The GLE core is described in detail in Appendix A. Figures 4-16 a-d illustrate some of the features of stratigraphic units in the GLE core. The GLE core was completely barren of ostracodes, probably due to water undersaturated with respect to calcite (note extremely dilute modern water chemistry). Ostracode preservation was also likely hindered by 1) mechanical destruction during deposition of the sands at the GLE site; and 2) shell dissolution under acid conditions of organic matter decomposition and methanogenesis.

Table 4-G Stratigraphic Units in Core GLE

Depth Interval (cm)	Description
0 - 183	peat, mg-fg fibrous, uppermost 93 cm not recovered due to high water content, increasingly 'gyttja like' with increasing depth in the unit, basal 20 cm of unit slightly sandy, sand content decreases upwards
183 - 198	fg-mg sandy peat, upper boundary gradational into overlying unit, bulk radiocarbon date (196-198 cm; 8850+/-80 [TO- 6893])
198 - 325	fg-mg clean sand, 320-325 cm of unit is fg-mg sand with occasional laminae of fg sandy silt, middle part of unit is poorly bedded fg-mg sand, 198-203 cm of unit is stained and has root traces, basal contact distinct
325 - 348	clayey fg sandy-silt, interval 12-18 cm from bottom of the unit is more sandy (fg-mg sandy-silt)
348 - 398	fg-mg sand, lower 25 cm of unit has mm-sized granules, lower 15 cm of unit has cm-scale bedding, upper boundary gradational over 1 cm
398 - 423	clayey fg sandy silt, poorly to non-laminated at base of unit, grading upwards to silty fg-mg sand in upper 10 cm of unit, occasional vcg sand granules
423 - 469	clayey fg sandy silt at base of unit (12 cm thick) grading upwards into less clayey, more sandy fg-mg sandy silt (20 cm thick and poorly laminated), grading upward into 14 cm thick zone of silty fg-mg sand with broken laminae of silty clay, upper contact of unit is distinct
469 - 472	fg-mg clean sand, no visible laminae, distinct contact
472 - 474	clayey fg-mg sandy silt, lower boundary distinct
474 - 480	vfg sandy silt, very poorly laminated, lower boundary distinct
480 - 573	silty clay, very poorly laminated to non-laminated, cm-sized dropstones at 560, 551 and 507 cm, very slightly sandy (vfg sand) in all but upper 25 cm of unit

Figure 4-16a: Poorly bedded fine to medium grained sand in core GLE (Gregory Lake). The location of the GLE core is given in Figures 2-6 and 4-15. Table 4-G provides a description of the core. Distance between heavy dark lines on the scale is 10 cm.

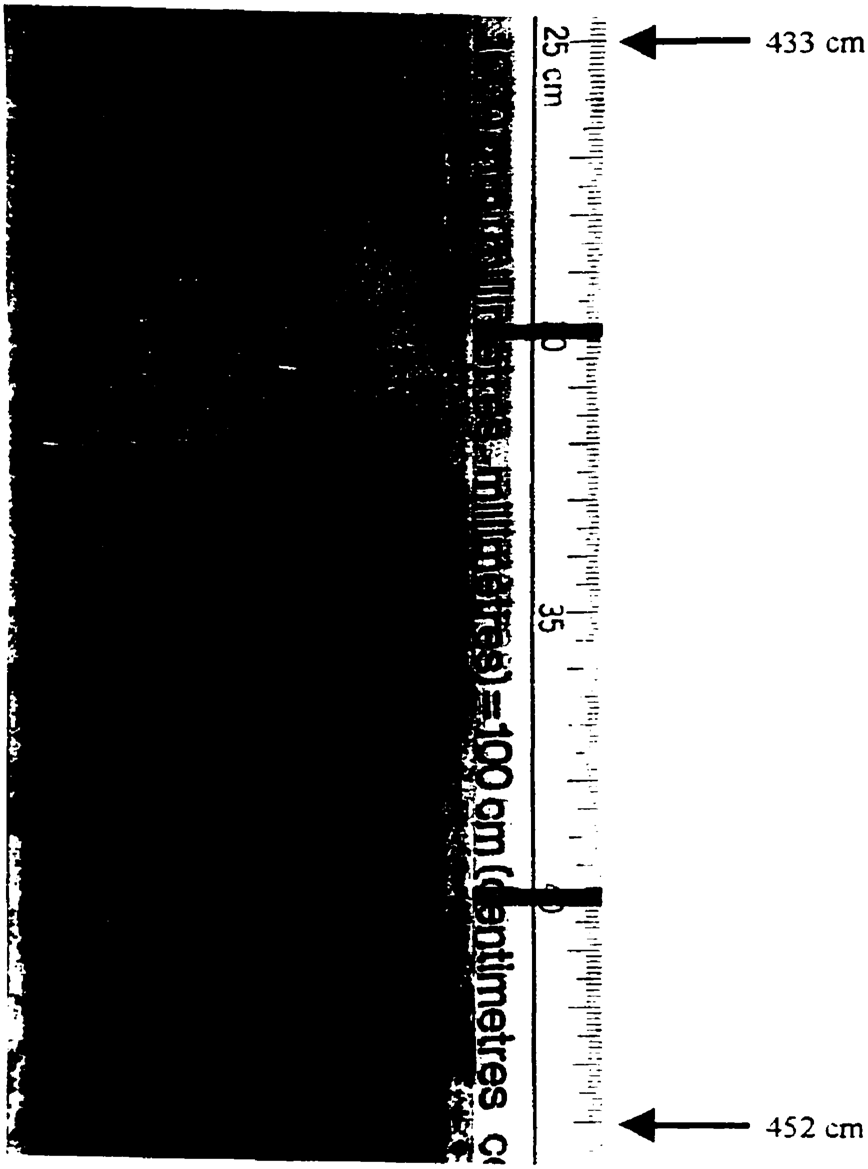


Figure 4-16b: Poorly laminated fine to medium grained sandy silt grading into overlying laminated silty fine to medium grained sand in the GLE core.

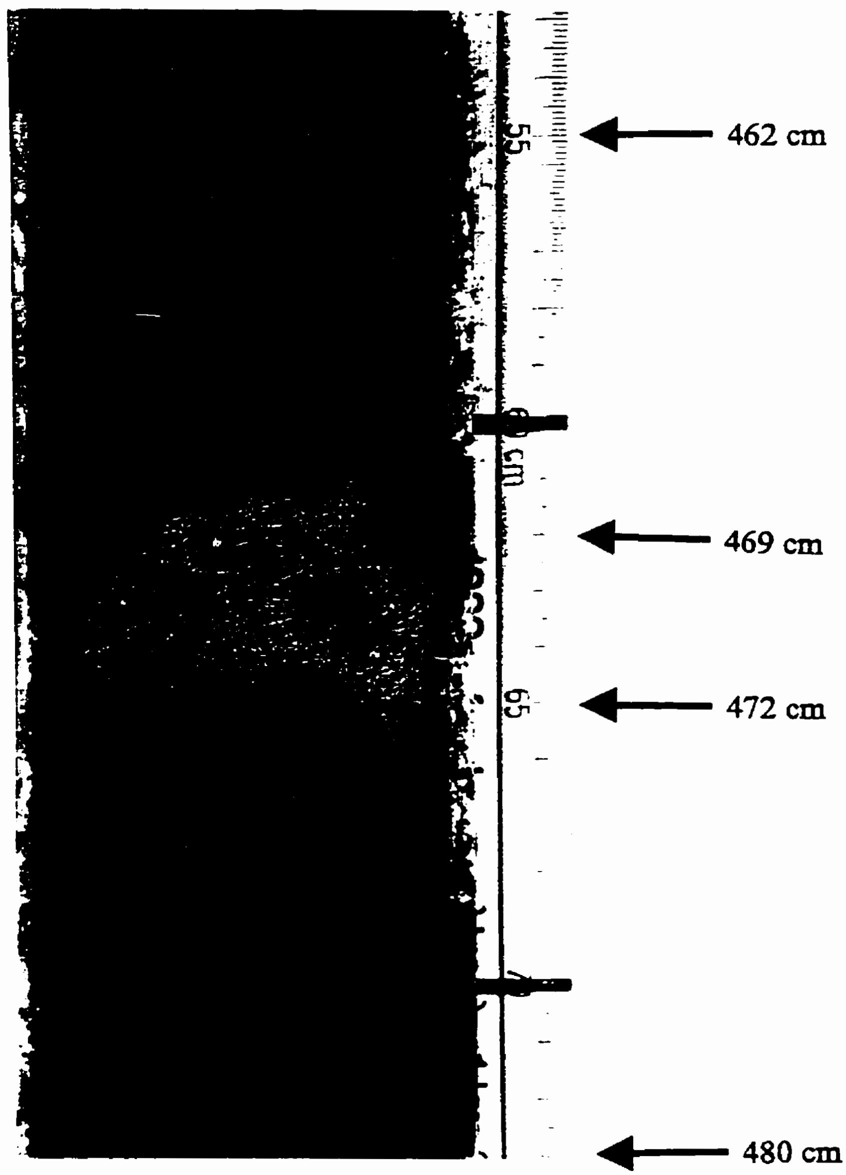


Figure 4-16c: Fine to medium grained clean sand between massive clayey fine to medium grained sandy silt (below) and massive clayey fine grained sandy silt (above) in the GLE core (Gregory Lake).

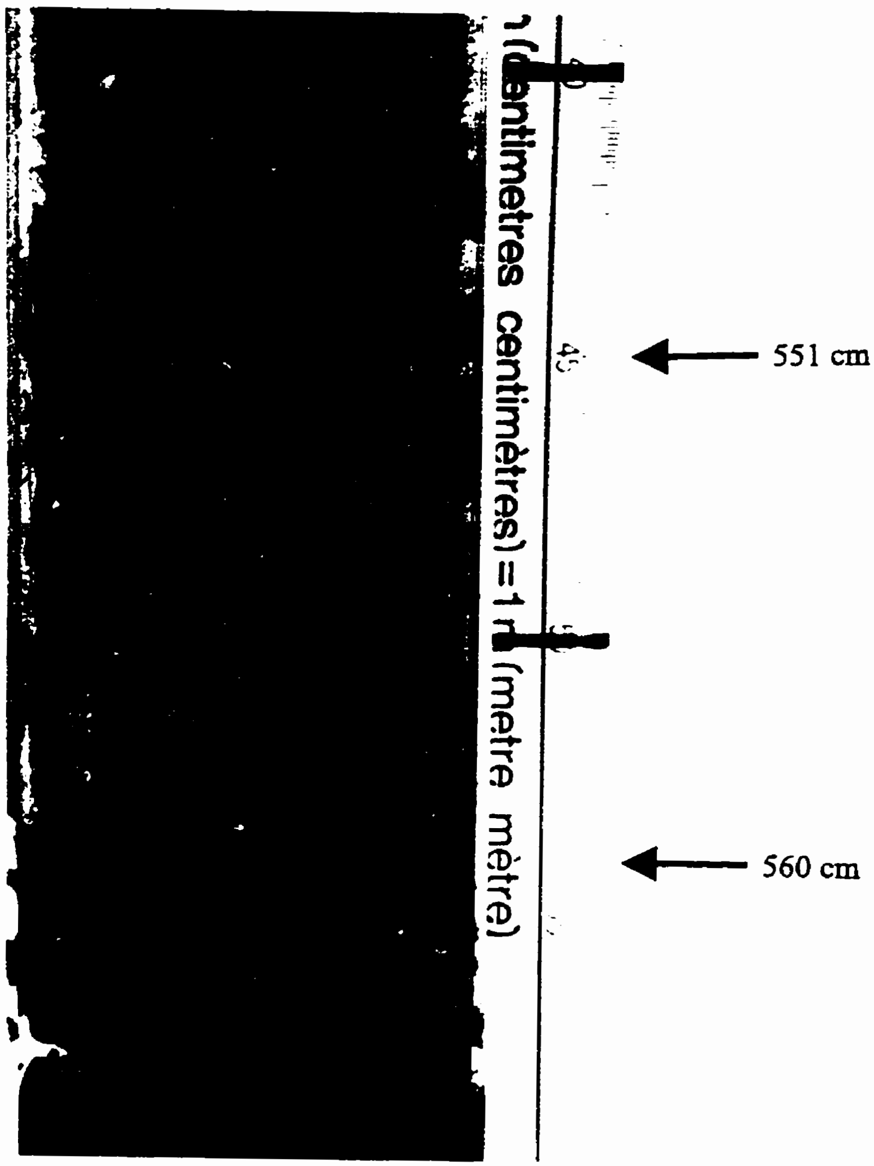


Figure 4-16d: Poorly to non-laminated silty clay, unit is generally very slightly sandy. Dropstones (cm-sized) occur at 560 and 551 cm depth. Photo is from the GLE core (Gregory Lake)

4.4.2 - Description of the Moisture, Organic, and Carbonate Trends in Core GLE

Figure 4-17 shows the plots of percent moisture, organic matter, and carbonate for the GLE core (Appendix B), along with a representative stratigraphic section of the core. The upper 93 cm of the core (peat) was not recovered in the field.

a) Moisture:

Of all the cores discussed and otherwise documented in this thesis, the GLE core is the only one that froze while in the field. Because of this, the moisture content in the GLE core may be relatively lower than natural. After freezing, some of the moisture which was originally in the pore spaces of the sediment does not all necessarily return back into those pore spaces. The moisture content in the GLE core (Fig. 4-17) is highest (~85%) in the upper organic peat (94 - 183 cm; Table 4-G), gradually decreases with depth in the peat (likely due to compaction of the peat and increased bulk density) and begins to decline in the sandy peat (183 - 198 cm; Table 4-G). The decrease in moisture in the sandy peat is likely an artifact of the increasing proportion of sand.

In the lower sand, silt, and clay portions of the core (183 - 573 cm; Table 4-G), moisture content is very low, likely due to water loss from freezing, and slow evaporation (during storage) from the open pore spaces of the sands which predominate this portion of the core. Relatively higher moisture contents are noted in zones which have a relatively higher proportion of silts and clays (e.g. 325 - 348 cm, 480 - 573 cm; Table 4-G).

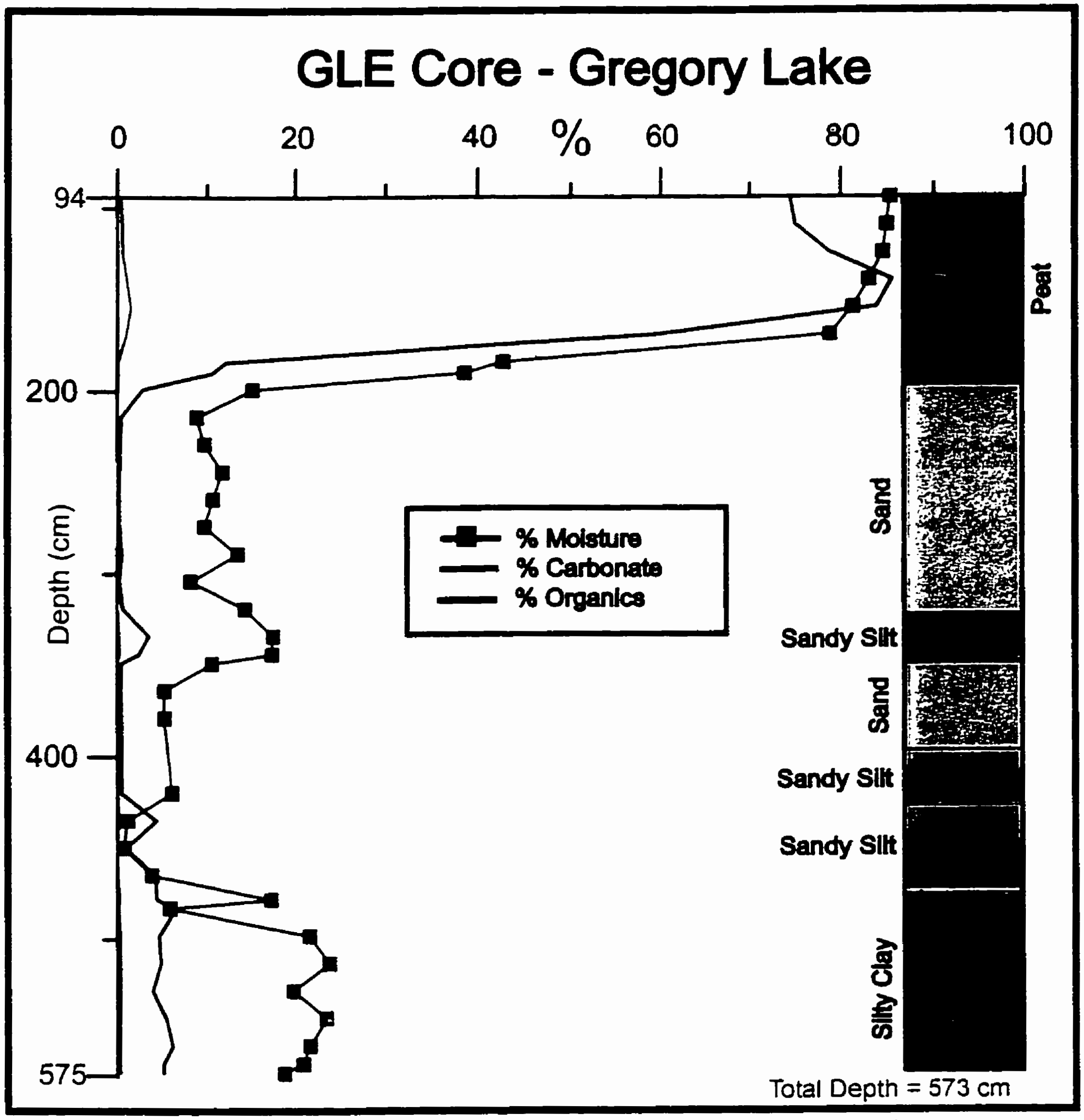


Figure 4-17: Moisture, organic, and carbonate percentage values for the GLE core (Gregory Lake). Squares indicate sampling depths. Organic matter and carbonate contents were calculated by LOI using a dry sample. Generalized stratigraphic section at right.

b) Organic matter:

Organic material content in the GLE core (Fig. 4-17) is high (~70 - 85%) in the upper portion of the peat (94 - 183 cm; Table 4-G). Throughout the 183 - 198 cm interval of the core, organic content drastically decreases as sand content increases. In the lower sand, silt, and clay portion of the core (198 - 573 cm; Table 4-G), organic content is low, except for the relatively silty (e.g. 325 - 348 cm; Table 4-G) and clayey (e.g. 480 - 573 cm; Table 4-G) intervals. Samples, which were searched for ostracodes under a reflected light microscope, verified this observation.

c) Carbonate:

As expected from the very dilute water presently in Bird Lake, carbonate content is low for the GLE core. Carbonate content (Fig 4-17) ranges from 0.01 to ~ 1.5 %. Not surprisingly, this is likely due to Gregory Lake's close proximity to the poorly buffered Canadian Shield, and its surroundings that consist of surficial materials that are poor ion donators (i.e. largely derived from the Shield). The modern chemistry of Bird Lake (Chapter 3), having very low total alkalinity, low TDS, and relatively low pH reflects this fact.

CHAPTER 5 - CORRELATION OF THE CAMPBELL BEACHES AROUND THE WAPAWEKKA HILLS, SASKATCHEWAN

5.1 - Correlation of the Campbell Beaches in the Wapawekka Hills Area: Introduction

The Upper and Lower Campbell beaches of Lake Agassiz have only recently been mapped and correlated using a +/- 0.5 m resolution Global Positioning System (GPS) survey from the international border, to the northwestern arm of Lake Agassiz, west of the Wapawekka Hills in north-central Saskatchewan (Fig. 5-1) (Thorleifson, unpublished; Matile, unpublished; Rayburn *et al.*, 1997; Rayburn, 1997). Strandline correlation is crucial to Lake Agassiz study, as it has implications for defining the extent, level, amount of differential isostatic rebound within the basin, and chronology of the lake and of its back beach lagoons. The Upper and Lower Campbell beaches are found as a pair throughout most of the Lake Agassiz basin, usually either as constructional beaches or erosional scarps. In places, one, or both of the strandlines may be absent. The absence of a beach may be due in part to initially poorly formed beaches, destruction by slumping (especially around areas underlain by Cretaceous shales, such as the Porcupine Hills and Pasquia Hills; Nielsen and Watson, 1985) and erosion, or dissection by rivers.

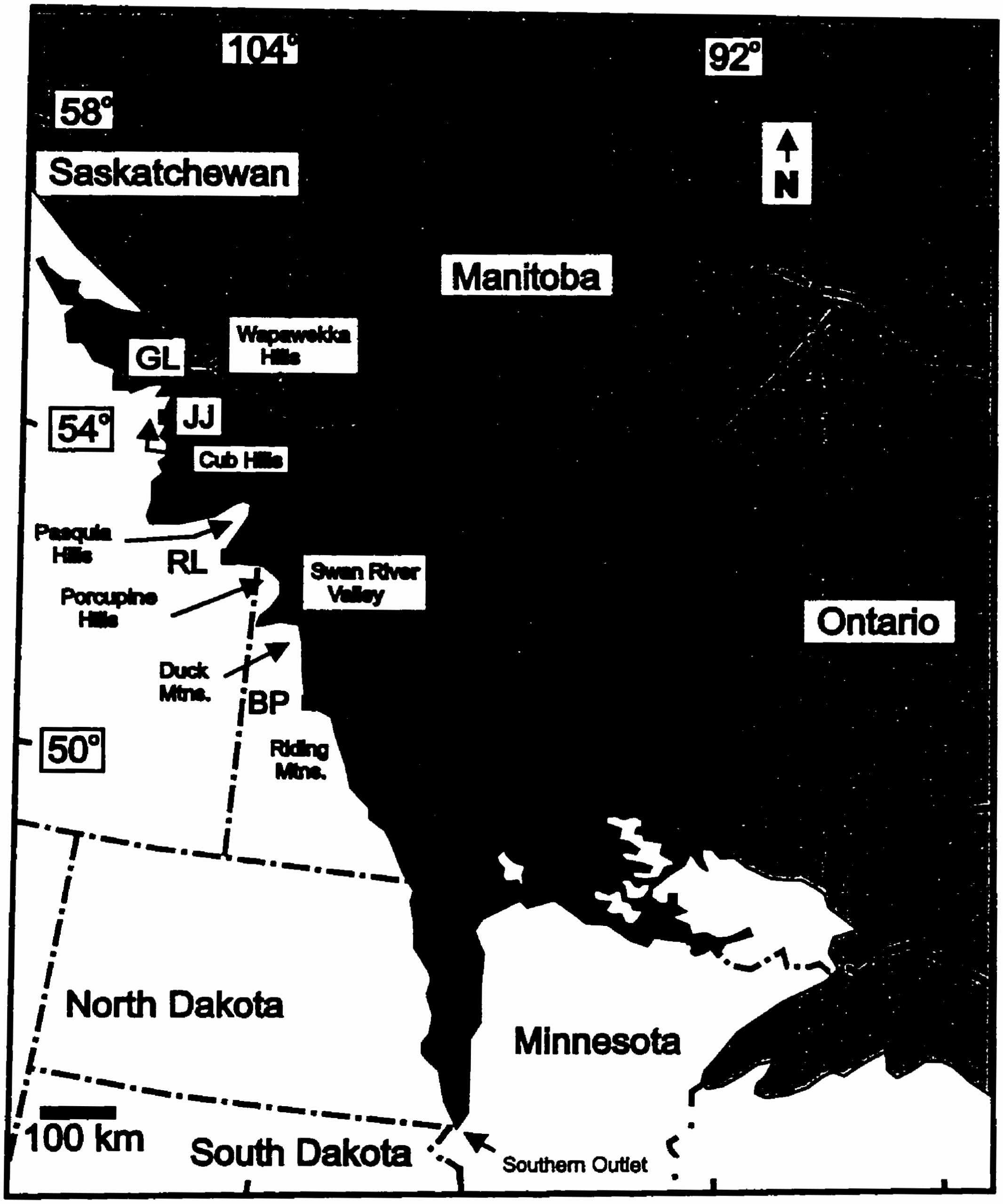


Figure 5-1: Location of new Upper Campbell lagoonal sites studied in this thesis. BP = Brokenpipe Lake, RL = Ruby Lake, JJ = Jay Jay Lake, GL = Gregory Lake. The Riding and Duck Mountains, with the Manitoban Porcupine Hills, make up the Manitoba Escarpment. Upper Campbell Lake Agassiz extent (dark shading) and Laurentide Ice Sheet (light shading) after Teller (1985, 1987).

5.1.1 - Discussion

The Upper and Lower Campbell beaches are correlatable from the southern outlet to the Wapawekka Hills (Fig. 5-1). Correlation of the beaches becomes difficult at the north and west slope of the Wapawekka Hills. Figures 5-2 A and B show the key area of this correlation problem and were assembled with information gathered from surficial geological maps (Langford, 1973; Campbell, 1988), soils maps (Head *et al.*, 1981) , and 1: 50 000 air photo mapping on 1: 50 000 topographic mapsheets. From Figure 5-2 A, it is apparent that there are two clear strandlines leading from the main Lake Agassiz basin east of the Wapawekka Hills, around the Wapawekka Hills, to the area south of Wapawekka Lake. In Figure 5-2 A, the Lower Campbell beach closely follows the 1375 foot (~ 419 m) contour. The Upper Campbell lies above this, at about 435 - 440 m elevation in this area (Langford, 1973; Rayburn, 1997). Also of note is the small island that existed during the Upper and Lower Campbell levels, south of Wapawekka Lake, and just west of 104° 30' longitude. Strandlines are conspicuous on this island in air photos.

Both the Upper and Lower Campbell strandlines can be followed to the area just east of the Wuchewun River (Fig 5-2 A). From this point westward, however, correlation becomes quite uncertain. Dissection of the strandlines by the many rivers flowing northward through the area, including the Wuchewun, Nipekamew, and Meeyomoot Rivers, is extensive. Any scraps, or discontinuous pieces of strandline (Fig. 5-2 B) in and around the Nipekamew River area -- an area characterized by a drumlin field which

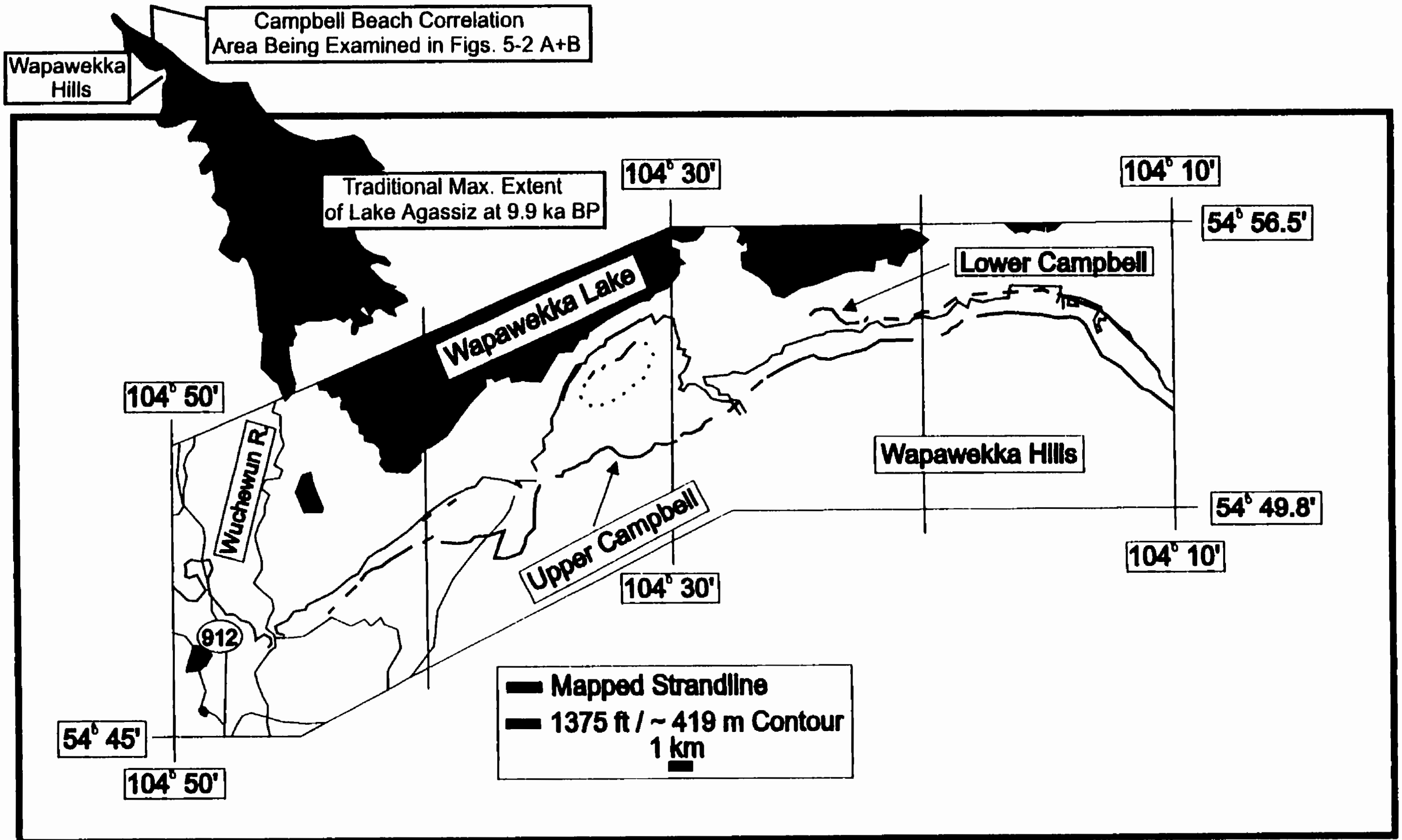


Figure 5-2 A: The eastern half of the key area where the correlation of the Upper and Lower Campbell beaches from the main basin of Lake Agassiz, toward the northwestern arm of Lake Agassiz becomes difficult. Diagram compiled from 1:50 000 air photo mapping on 1:50 000 topographic mapsheets. See text for details. (maximum extent of Lake Agassiz in locator map after Teller, 1985, 1987)

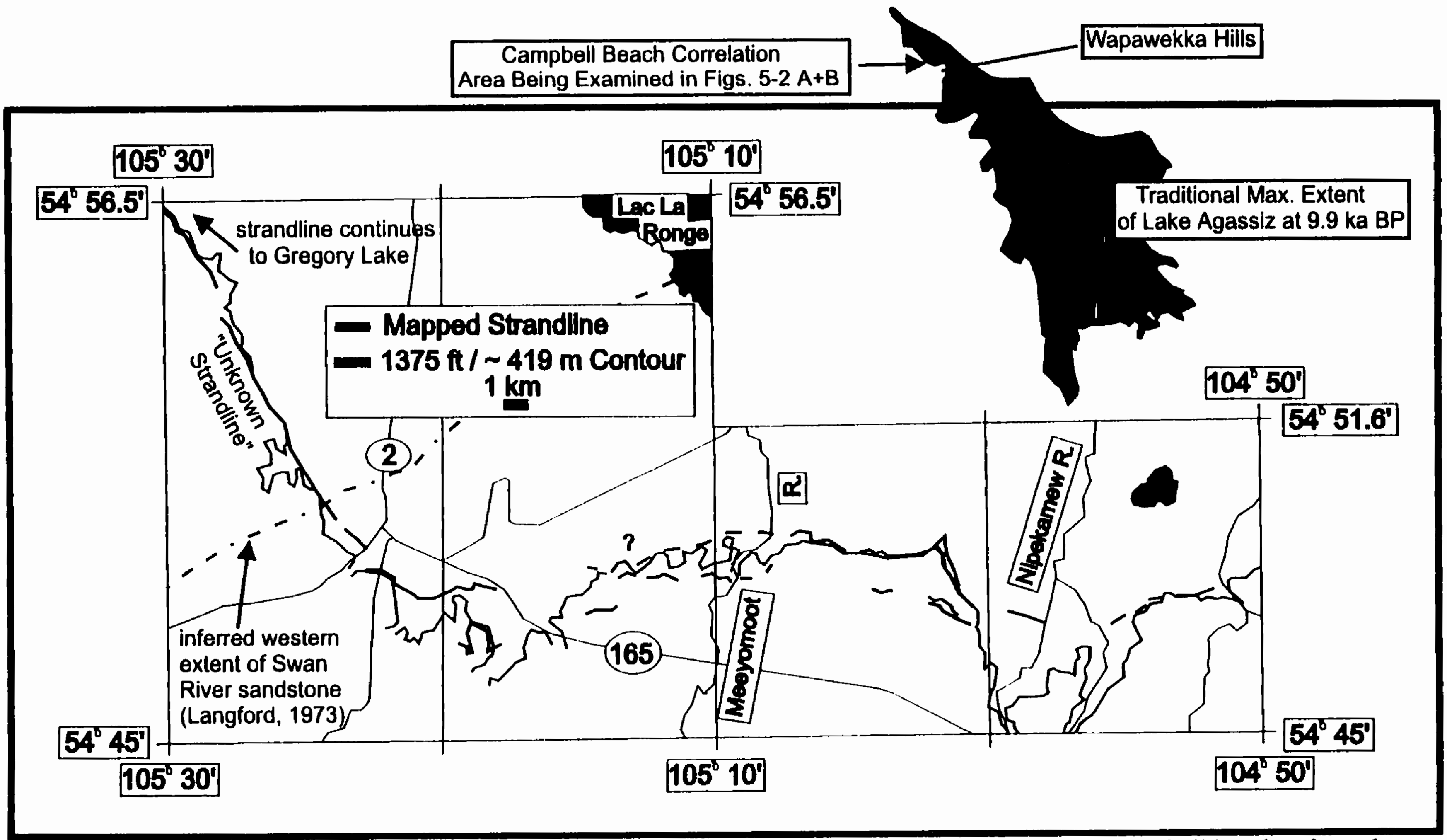


Figure 5-2 B: The western half of the key area where the correlation of the Upper and Lower Campbell beaches from the main basin of Lake Agassiz, toward the northwestern arm of Lake Agassiz becomes difficult. Diagram compiled from 1:50 000 air photo mapping on 1:50 000 topographic mapsheets. See text for details. (maximum extent of Lake Agassiz in locator map after Teller, 1985, 1987)

would have formed a shallow bay of many islands (Langford, 1973) at the Upper Campbell level -- are all approximately 1375 feet (~ 419 m) in elevation. The one and only clear strandline that continues toward the western arm of Lake Agassiz (west of junctions of Highways 2 and 165; Fig. 5-2 B) is a well developed constructional beach where sand from re-worked Swan River sandstone (Cretaceous bedrock in this area) and sand reworked from sandy tills of Swan River Sandstone provenance was formed into a distinct sand berm; it is associated with nearshore lacustrine sand deposits as well (Langford, 1973). In the area west of the junctions of highways 2 and 165, this beach is surrounded by low relief, drumlinoid moraine or hummocky glaciofluvial terrain (Langford, 1973; Campbell, 1988). In the Gregory Lake area (farther northwest along the same strandline), the beach becomes an erosional wave-cut bench eroded into ridged, hummocky ground moraine (Simpson, 1988). This change to an erosional strandline probably reflects increased exposure in this part of the Lake Agassiz basin (Matile, 1998, personal communication) and perhaps a closer proximity to the relatively sediment-starved Canadian Shield and distance from the a source of sand from sandy tills of Swan River sandstone provenance. In the Wapawekka Hills and Lac la Ronge area, the Swan River sandstone is covered by a thin till veneer, and exceptionally sandy tills in the Wapawekka Hills area are useful to infer the extent of Swan River sandstone bedrock (shown in Fig. 5-2 B; after Langford, 1973). The sandstone is inferred to extend about 15 km west of the junction of Highways 2 and 165, and the western boundary of the sandstone occurrence (Langford, 1973) runs diagonally from the southern tip of Lac la

Ronge through the junction of Highways 2 and 165 (Fig. 5-2 B).

Undoubtedly, the continuity and excellent morphology of this beach in the Lac la Ronge area (Fig. 5-2 B) suggests it formed over a relatively long period of time, had a reasonable sediment supply (weathered and outcropping Swan River sandstone near the Wapawekka Hills and extremely sandy tills of sandstone provenance) and that Lake Agassiz had a reasonable fetch. The elevation of this singular strandline, west and south of Lac la Ronge (Fig. 5-2 B), like the Lower Campbell beach east and north of the Wapawekka Hills (Fig. 5-2A) closely follows the 1375 ft (~ 419 m) elevation contour. Thus, at first glance, this implies that this single strandline that continues toward the northwest arm of Lake Agassiz (hereafter referred to as the "unknown" strandline, see Fig. 5-2 B) is part of the Lower Campbell strandline.

Unfortunately, the story is not so simple. The elevations of the unknown strandline, when compared to the isobases of Teller and Thorleifson (1983) suggest that this strandline is part of the Upper Campbell (see Section 5.1.3). However, if the unknown strandline is considered to be part of the Upper Campbell, it is important to reconcile the absence of the Lower Campbell strandline in the region west of Lac la Ronge.

The opening of a lower eastern outlet (the Kaiashk) late in the Emerson Phase, and the subsequent closure of that outlet by a small ice readvance in the eastern outlet region (Thorleifson, 1996) caused the reoccupation of the southern outlet, and the rise of Lake Agassiz to the Lower Campbell level. This scenario explains the fact that elevations of the Upper and Lower Campbell beaches diverge from the southern outlet (Teller and Thorleifson, 1983). If the Lake Agassiz basin was open (i.e. not covered by readvancing ice) for the duration of the Upper and Lower Campbell levels (as suggested by accepted Lake Agassiz history), intuition suggests that wherever the Upper Campbell strandline (erosional or constructional) is present in a particular locality an expression of the Lower Campbell (erosional or constructional) should be found as well.

Limiting factors on the formation of a beach include time, fetch, and sediment supply. Since 1) the Upper and Lower Campbell levels were of similar duration (Teller and Thorleifson, 1983), 2) they would have formed in a lake with a similar geographic extent (i.e. fetch and therefore beach building energy was similar for both lake levels), and 3) within the same region of the Lake Agassiz basin, both beaches were likely governed by similar (though not necessarily equal) sediment budgets, both Campbell beaches commonly co-occur throughout most of the Lake Agassiz basin. If the Upper and Lower Campbell levels of Lake Agassiz were both present in the western arm of Lake Agassiz, and considering that well-developed Upper and Lower Campbell strandlines are very commonly found paired together throughout much of the Agassiz basin (even to localities within about 30 km of the unknown strandline, south of Wapawekka Lake; Fig. 5-2 A), it does not make sense that the Upper or Lower Campbell strandline would be absent.

If the unknown strandline is the extension of the Upper Campbell strandline, the Lower Campbell in this region may perhaps be a diffuse, discontinuous assemblage of small, poorly-formed erosional strandline scraps on the flanks of hummocky or drumlinoid moraine plain which to date have yet to be mapped. The discontinuous nature of strandlines in drumlinoid moraine of the Wapawekka Hills area (Langford, 1973), and in the northwestern arm of the Lake Agassiz basin (Fisher, 1993) suggests that the shoreline of Lake Agassiz -- also probably because of thin drift cover and the irregular physiography of the nearby Canadian Shield -- is not continuous near the Shield, but was formed dispersed among a complex assemblage of shallow water and small islands in an "archipelago". The same can be said for lower elevation strandlines south of Lac la Ronge and in the Gregory Lake area as hummocky ground moraine is pervasive in that region.

The Laurentide Ice Sheet can be used to explain, in absence of new GPS elevation data, that the unknown strandline is an expression of the Lower Campbell level. The isolation of the northwestern region from the main Lake Agassiz basin, by ice abutted against the Wapawekka Hills, may have occurred during general retreat. This intervention of the Laurentide Ice Sheet would preclude Lake Agassiz' occupation of the northwestern region, thereby making the formation of beaches impossible. If this occurred during the Upper Campbell level but not during the Lower Campbell level, one would expect the absence of the Upper Campbell strandline and the presence of the Lower Campbell strandline in the northwestern arm of the Lake Agassiz basin. The reverse is not true, however, as the ice sheet would destroy any nearby shoreline (i.e. the Lower Campbell) as it advanced to close the basin (to form the Upper Campbell). If the Laurentide Ice

Sheet was in contact with the Wapawekka Hills for the duration of the of the Upper Campbell level, the Upper Campbell strandline would end there.

Although they were dealing more specifically with the Nipawin Delta, deposited during the Upper Campbell level of Lake Agassiz, this appears to be what was suggested by the paleogeographic reconstruction of Christiansen *et al.* (1995). As the ice pulled back from the Wapawekka Hills at the end of the Upper Campbell level, Lake Agassiz would have expanded northwestward, eventually stabilizing at the Lower Campbell level. In this scenario, the Upper Campbell is continuous throughout the Lake Agassiz basin, as far as the north side of the Wapawekka Hills, where the Laurentide Ice Sheet would have served as a barrier. The Lower Campbell strandline is continuous throughout the Agassiz basin, but it persists west of the Wapawekka Hills, into the northwestern arm of the Lake Agassiz basin. While Schreiner (1983) made no interpretation, to explain the absence of a Campbell beach south and west of Lac la Ronge, he concluded that the one continuous portion of strandline that continues toward the northwest arm of Lake Agassiz -- the unknown strandline -- is probably at elevations below the Upper Campbell beach.

5.1.2 - Analysis Using Independantly Collected Global Positioning System (GPS) Elevation Data: Introduction

Based on the facts that: 1) the elevation of the unknown strandline plots at Upper Campbell elevations (according to the isobases of Johnston, 1946, and Teller and Thorleifson, 1983) -- but that it intuitively seems more reasonable for the Lower

Campbell to continue northwestward from the Wapawekka Hills area (i.e. the Laurentide Ice Sheet abutted against the Wapawekka Hills precluded occupation of the northwestern arm of the Lake Agassiz basin by Lake Agassiz at the Upper Campbell level) -- and 2) GPS elevation data on the Upper and Lower Campbell beaches along the western margin of the Lake Agassiz basin weakly suggest that isobases curve to a more west-east orientation, Rayburn (1997) argued that the isobases in the Lake Agassiz basin (Johnston, 1946; Teller and Thorleifson, 1983) need to be re-oriented; as a result this created relatively more isostatic depression in the area northwest of the Wapawekka Hills. The isobase re-orientation proposed by Rayburn (1997) causes elevations of the unknown strandline to become closer to those elevations associated with the *known* Lower Campbell strandline, thereby facilitating its correlation to the main Lake Agassiz basin Lower Campbell beach.

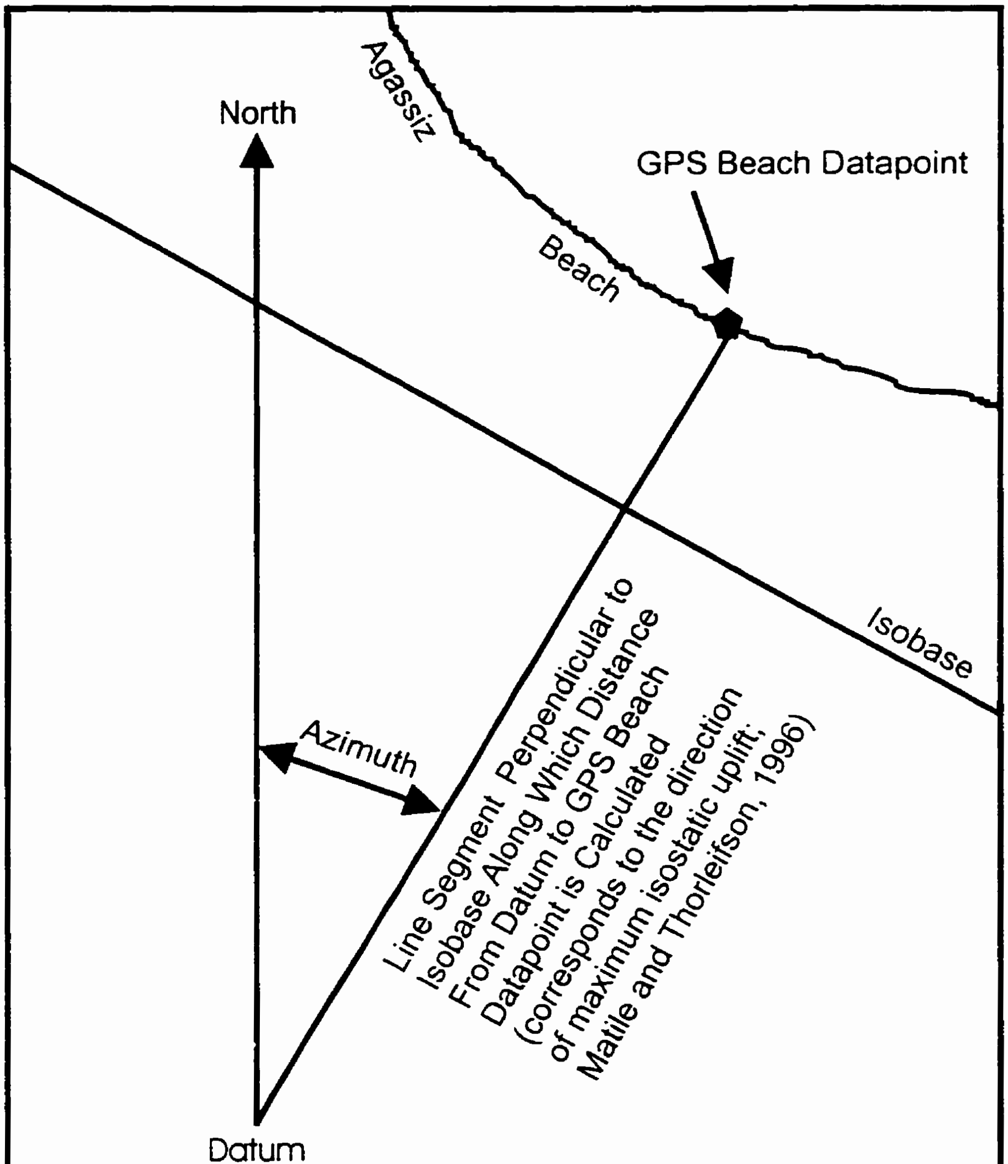
5.1.3 - A Simple Procedure to Test the Validity of Isobase Orientation Changes on GPS Beach Elevation Data

Following a simple procedure first used by Matile and Thorleifson (1996), the GPS beach elevation dataset collected on the Upper and Lower Campbell beaches in the western Lake Agassiz basin (Rayburn, 1997; Rayburn *et al.*, 1997) was analysed in a rigorous way, in order to determine 1) whether the unknown strandline is better related to the Upper or Lower Campbell levels of Lake Agassiz, and 2) whether the GPS data collected to date is conclusive enough to support an isobase orientation change. Figure

(5-3) shows a schematic diagram of how the data was analysed. In all cases, locations were dealt with in terms of Universal Transverse Mercator (UTM) coordinates, and elevations in meters.

A datum was chosen that was south and west of the points in UTM Zone 13 of the dataset, in order to keep all easting and northing distances as positive values. The same was done for the datapoints in UTM Zone 14. The distance (in terms of eastings and northings, in kilometers) from that datum to each of the beach GPS datapoints in both smaller datasets (UTM Zone 13 and UTM Zone 14) was calculated. Next, the easting and northing distances were plugged into the formula shown in Figure (5-3), in order to determine the distance of the datum to the GPS beach datapoint along a line segment perpendicular to isobase, for different Azimuth orientations of the line segment. By changing the Azimuth of this line segment, the orientation of the isobase is also being changed. The data was then plotted as distance (in kilometers) from the datum along a line perpendicular to isobase at some Azimuth orientation (on the x-axis) versus elevation (in meters) on the y-axis.

Figures (5-4A-D) show the plots of distance from the datum along a line perpendicular to isobase, with changing Azimuth orientations. Second order polynomials were fit to the data, and correlation values (r^2) for the Upper Campbell and Lower Campbell beaches are given in Table 5-A. Based on other data in the Lake Agassiz basin, the orientation that best matches the direction perpendicular to the isobases (the direction of maximum isostatic uplift; Matile and Thorleifson, 1996) of Johnston (1946)



Distance from datum to GPS beach datapoint along the line perpendicular to isobase = COS(Az*PI/180)*north + SIN(Az*PI/180)*east

Where:
north = northing distance (km) from datum to GPS beach point
east = easting distance (km) from datum to GPS beach point
Az = Azimuth

Figure 5-3: Theory behind determining distance from the datum to the beach GPS data point as the azimuth of the line segment perpendicular to isobase changes (thereby altering the trend of isobase). Method after Matile (pers. comm.).

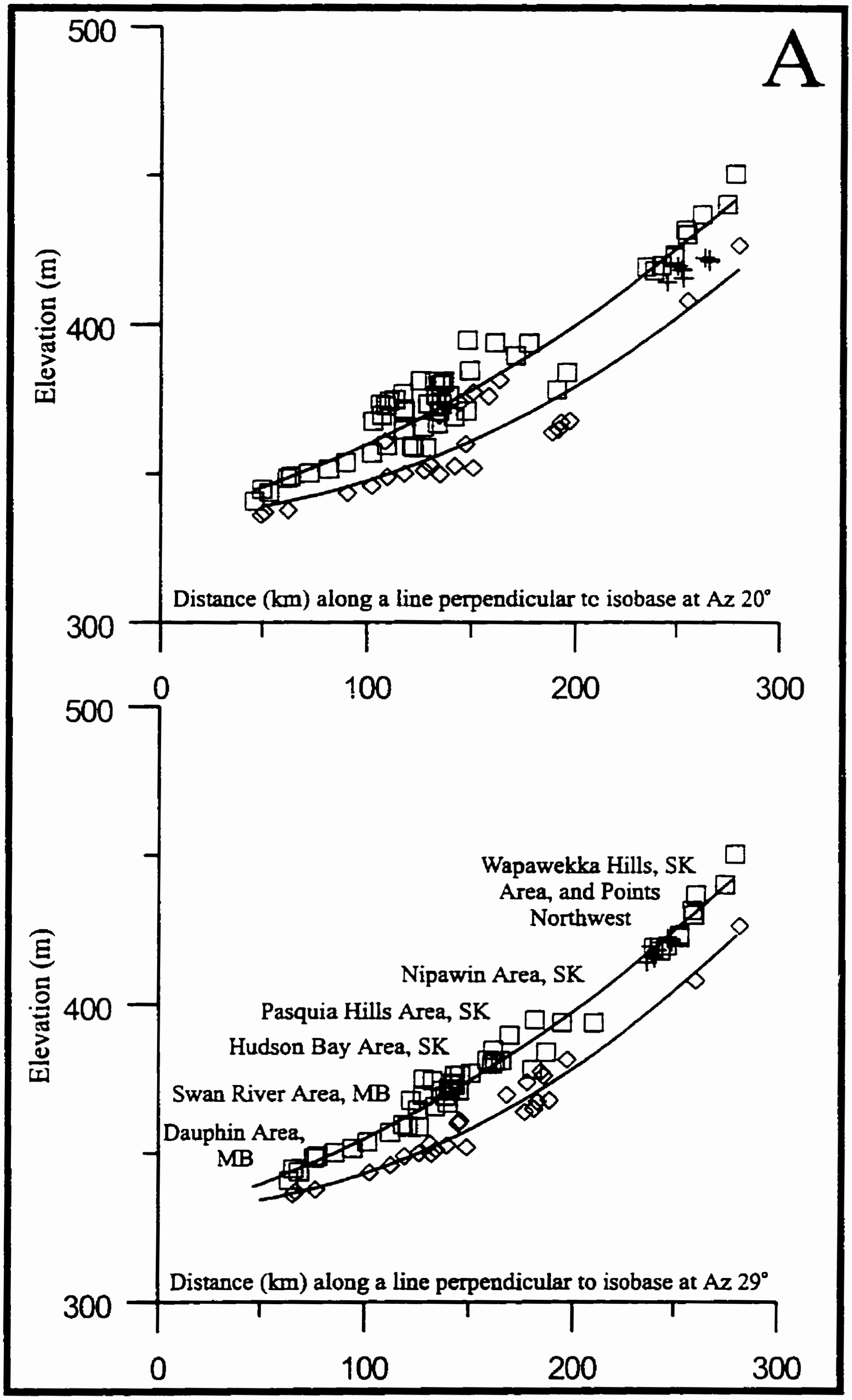


Figure 5-4 A-D: Plots of all points in the GPS dataset of Rayburn (1997) and Rayburn *et al.*, (1997). Squares = Upper Campbell, Diamonds = Lower Campbell, Crosses = Strandline west of the Wapawekka Hills. See text for details.

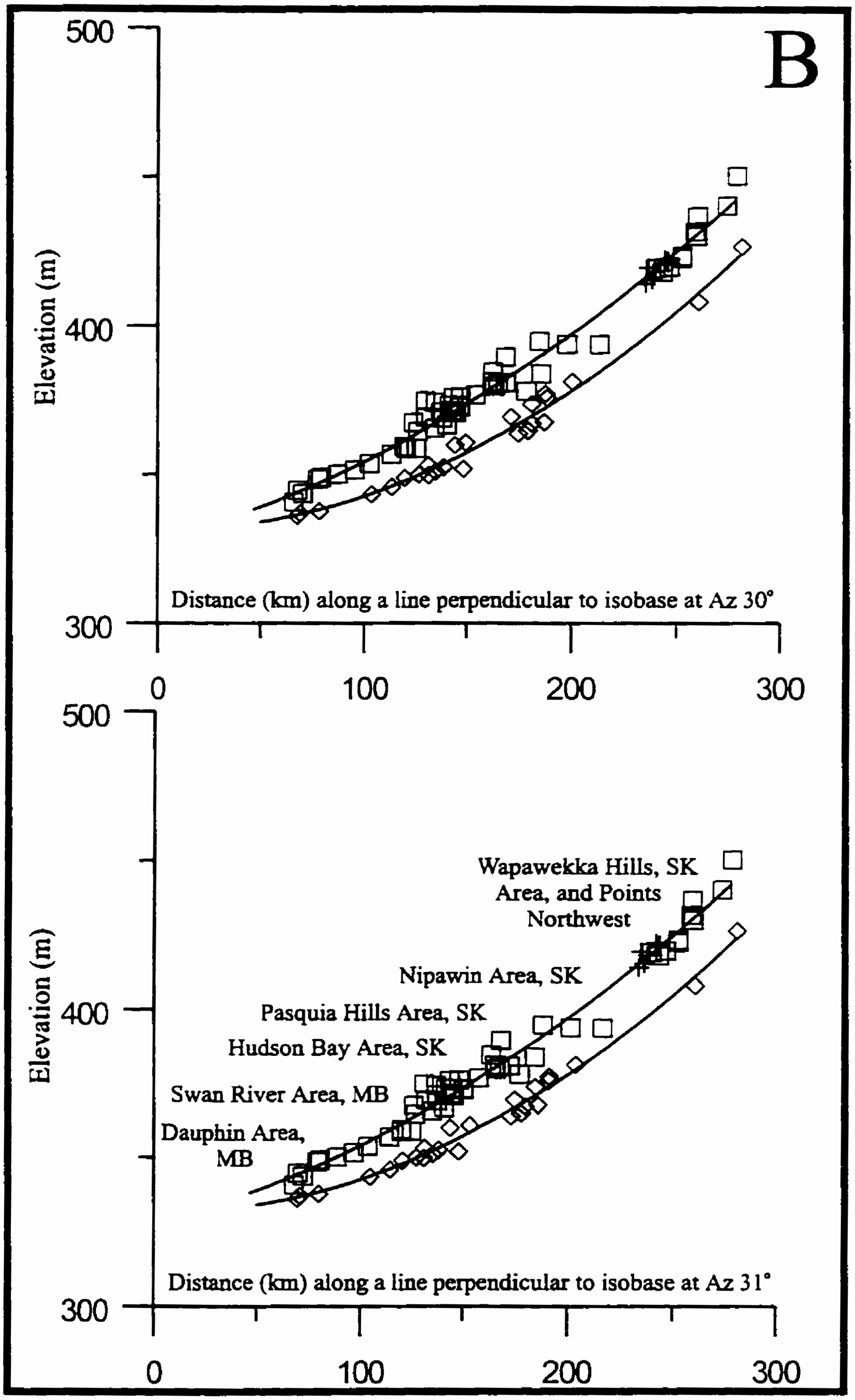


Figure 5-4 A-D continued: Plots of all points in the GPS dataset of Rayburn (1997) and Rayburn *et al.*, (1997). Squares = Upper Campbell, Diamonds = Lower Campbell, Crosses = Strandline west of the Wapawekka Hills. See text for details.

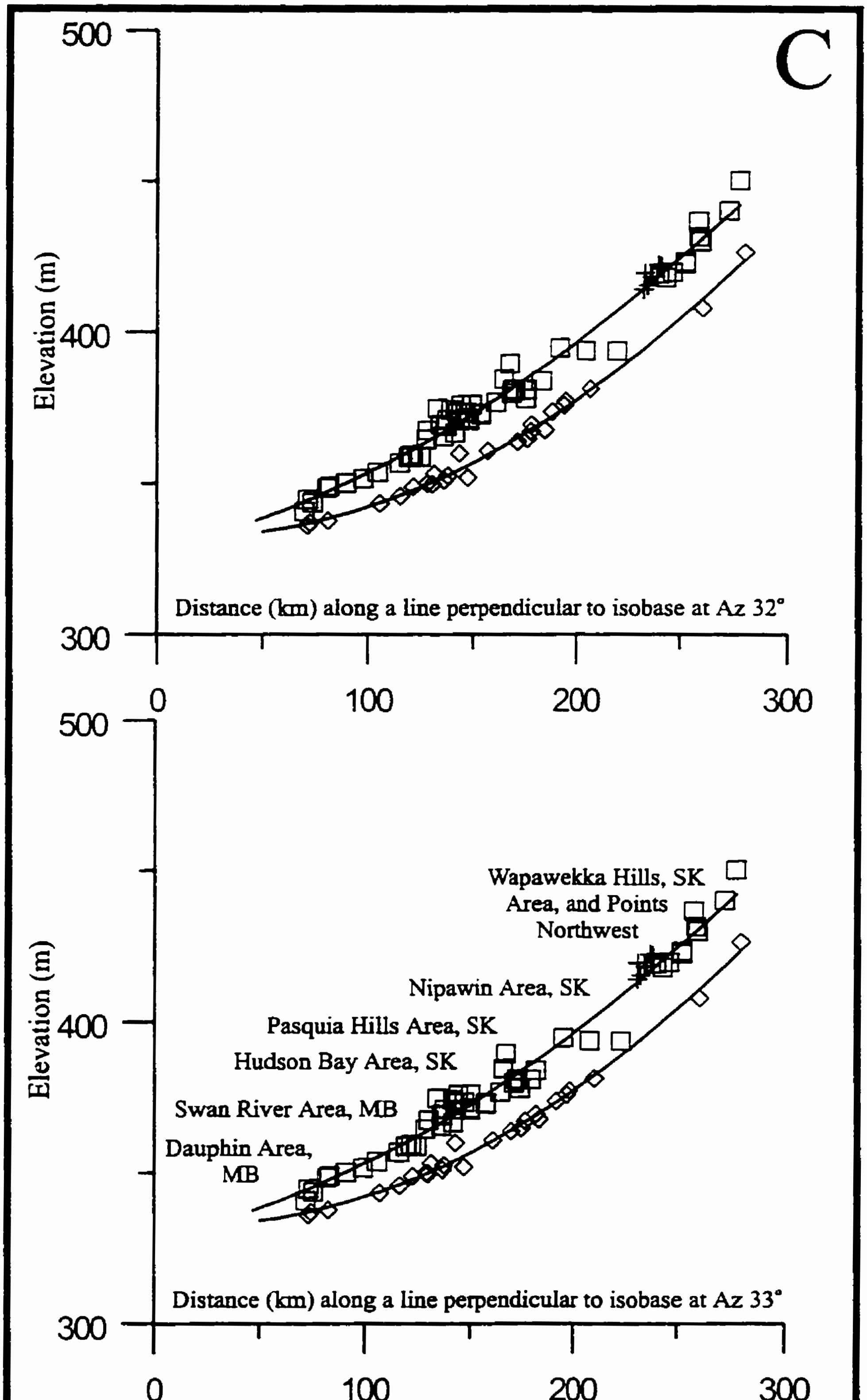


Figure 5-4 A-D continued: Plots of all points in the GPS dataset of Rayburn (1997) and Rayburn *et al.*, (1997). Squares = Upper Campbell, Diamonds = Lower Campbell, Crosses = Strandline west of the Wapawekka Hills. See text for details.

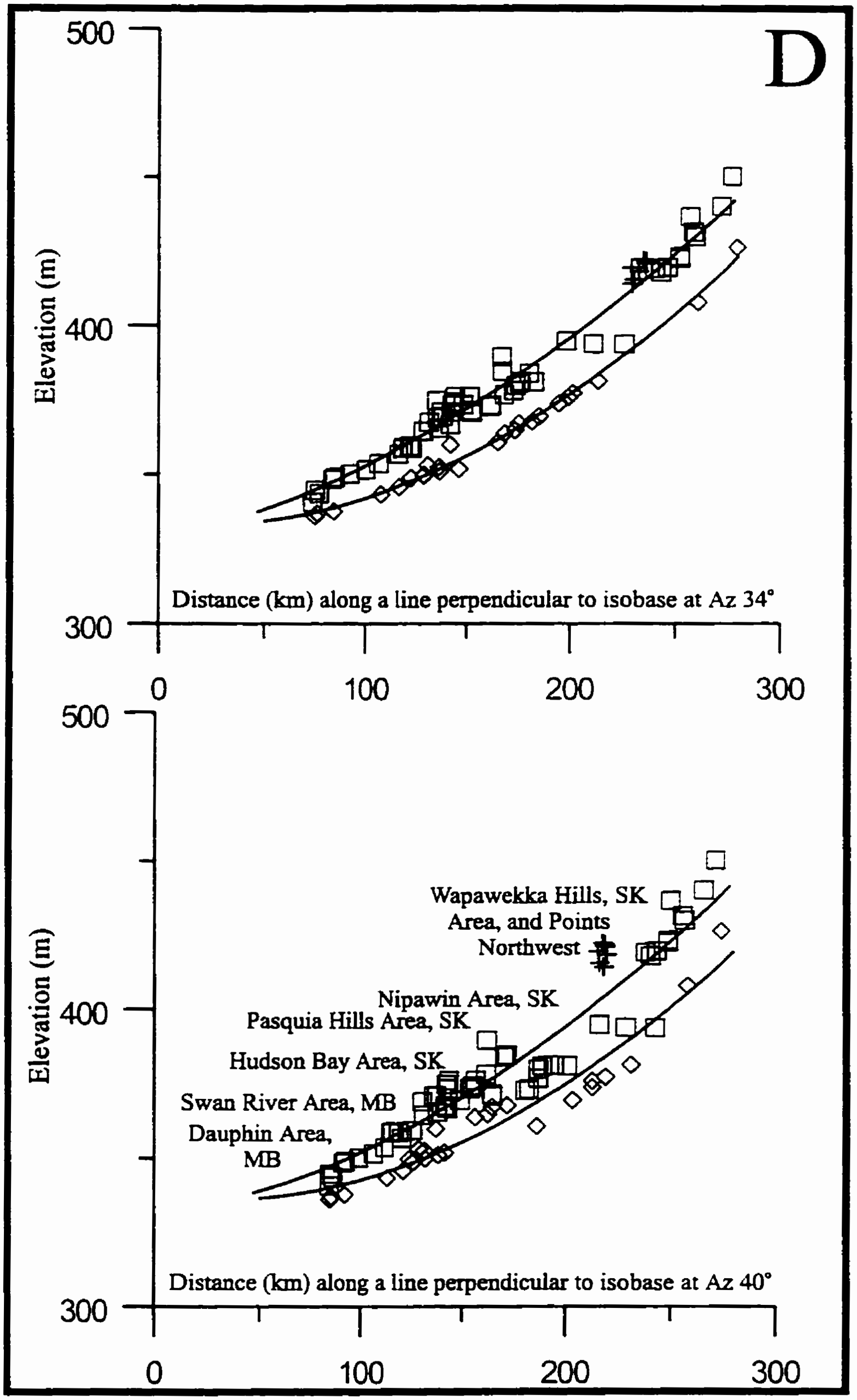


Figure 5-4 A-D continued: Plots of all points in the GPS dataset of Rayburn (1997) and Rayburn *et al.*, (1997). Squares = Upper Campbell, Diamonds = Lower Campbell, Crosses = Strandline west of the Wapawekka Hills. See text for details.

Table 5-A: Correlation values for GPS analysis of the Campbell Beaches

Azimuth	Upper Campbell (r^2)	Lower Campbell (r^2)
15°	0.870379	0.703947
18°	0.901493	0.769960
20°	0.920523	0.813783
26°	0.964123	0.929750
27°	0.968771	0.944790
28°	0.972507	0.958066
29°	0.975266	0.969385
30°	0.976991	0.978565
31°	**0.977622**	0.985445
32°	0.977104	0.989883
33°	0.975382	**0.991765**
34°	0.972405	0.991008
40°	0.925831	0.931102

Note: Azimuth refers to azimuth direction of the line perpendicular to Isobase (the direction of maximum isostatic uplift) and Teller and Thorleifson (1983) is about Az 30° (Matile, pers. comm.). Interestingly, for this dataset, as the Azimuth of the line perpendicular to isobase increased within a few degrees of 30°, the highest r^2 values for the Upper and Lower Campbells were reached. The maximum r^2 value for the Upper Campbell is achieved at a direction perpendicular to isobase of Az 31°, and for the Lower Campbell at Az 33° (Table 5-A). Therefore, by this analysis, it is proposed that the orientation of isobase in this portion of the Agassiz basin is probably somewhere between Az 121° and Az 123°. Note that this is very close to the Az 120° orientation (line perpendicular to isobase is at Az 30°) suggested by Johnston (1946) and Teller and Thorleifson (1983), and possibly reflects the subtle bending of the isobases to the northwest, as suggested by the Teller and Thorleifson (1983) model.

The upper plot in Figure 5-4 B illustrates that the GPS data, and traditional isobase orientations (Johnston, 1946; Teller and Thorleifson, 1983) place the datapoints taken from the unknown strandline (crosses) at elevations expected for the Upper Campbell. The upper plot in Figure (5-4 A) shows the data plotted against a line perpendicular to isobase at an orientation of Az 20°; isobases would have a more west - northwestward orientation (Az 110°) versus the traditional (Johnston, 1946; Teller and Thorleifson, 1983) isobase orientation (Az 120°). This is in part the change in direction of isobase orientation that Rayburn (1997) suggested in the northwestern Lake Agassiz basin. With a line perpendicular to isobase at an orientation of Az 20° (isobases at Az 110°), correlation values for the Upper and Lower Campbell curves (r^2 values of 0.920523 and 0.813783 respectively, Table 5-A) are poor, and there is a visible increase in the spread of the data over that seen in the plots which have an Azimuth direction perpendicular to isobase that is within a few degrees of Az 30° (see Figs. 5-4 A-D). However, notice in the upper plot of Figure 5-4 A that the datapoints from the unknown strandline (crosses) plot at elevations closer to those of the known Lower Campbell. This more west-east isobase trend (versus Teller and Thorleifson, 1983) was the isobase re-orientation suggested by Rayburn (1997), which helped explain his conclusion that the unknown strandline was the extension of the Lower Campbell level of Lake Agassiz, and that the Upper Campbell level was cut off by the Laurentide Ice Sheet at the Wapawekka Hills.

The preceding analysis has unfortunately not revealed a solid conclusion about which Campbell level the unknown strandline belongs to, however it has suggested that changing isobase, in order to explain the GPS data, is not necessarily the prudent approach (it results in plunging correlation values (r^2) for the Campbell beaches). It appears that the isobase configuration of the Lake Agassiz basin could perhaps benefit from a re-evaluation based on GPS beach data. However, it also seems that any such changes to isobase must be subject to more data collection and research in the Agassiz basin, particularly in north-central Saskatchewan.

In any case, Rayburn (1997) effectively raised some key considerations and undoubtedly showed that isobase configuration, at least based on the GPS data collected in north-central Saskatchewan and the rest of the Agassiz basin, needs to be closely scrutinized. Only in this way will the correlation of the Campbell beaches from the main

Agassiz basin, around the Wapawekka Hills, into the northwestern arm of Lake Agassiz be solved conclusively.

5.1.4 - Summary

The formation of the Campbell beaches in the Wapawekka Hills area, and in the northwestern arm of Lake Agassiz may have occurred in one of two ways:

1) The Upper Campbell formed while the Laurentide Ice Sheet was abutted against the Wapawekka Hills, and the Lower Campbell formed after the ice sheet had retreated off the Wapawekka Hills. Thus, the Upper Campbell does not continue west of the Wapawekka Hills, and if any major Emerson Phase beach exists in the northwestern arm of Lake Agassiz, it is the Lower Campbell (Rayburn, Mann, and Teller, 1996; Rayburn, 1997). This scenario is not supported by the regional analysis of GPS elevation data on the Upper and Lower Campbell beaches, when applied to traditional isobases (Teller and Thorleifson, 1983) for the Lake Agassiz basin.

2) Both the Upper and Lower Campbell strandlines formed after the ice sheet had retreated north from the Wapawekka Hills (i.e. this ice sheet configuration against or near the Wapawekka Hills had no direct effect on the formation and distribution of Campbell strandlines). This scenario is supported by the GPS beach elevation data and traditional isobases (Teller and Thorleifson, 1983), although it is difficult to reconcile the absence of the Lower Campbell strandline west of the Wapawekka Hills since Lake Agassiz occupied that area as an open embayment for both of the Campbell levels.

Without a change in isobase configuration (which is not supported by the regional

GPS elevation data analysis presented above), the unknown strandline that continues to the northwest of the Wapawekka Hills appears to be a continuation of the Upper Campbell. Questions relating to the absence of the Lower Campbell strandline west of the Wapawekka Hills, strandline correlation, and isobase re-orientation in the Wapawekka Hills area, remain to be answered -- these are subjects for future data collection and consideration.

CHAPTER 6 - INTERPRETATION

6.1 - OSTRACODE BIOZONES OF THE BROKENPIPE LAKE BPC CORE

6.1.1 - INTRODUCTION

Figure 6-1 A shows the relative abundance diagram for adult ostracode valves identified in the BPC core. The ostracode counts are tabulated in Appendix D. The total number of adult ostracodes (1 adult = 1 carapace; 1 carapace = 1 left and 1 right adult valve) per interval is indicated by the curve at the extreme left of the diagram. The percentages of relative abundances are based on this total number.

The BPC core of Brokenpipe Lake has three different ostracode biozones that correspond to several distinctly different stratigraphic units. Sediment the base of the BPC core (299 - 335 cm; Table 4-A; Chapter 4) is undoubtedly glaciolacustrine, and presumed to have been deposited in Lake Agassiz. Ostracodes are almost non-existant in this portion of the core, although three shells of *Heterocypris glaucus* were discovered from 6 large subsamples within this unit (Fig. 6-1 A). Above this clay unit are two distinct biozones with distinctive faunal assemblages, including Biozone A (196 - 299 cm), and Biozone B (0 - 196 cm) (Fig. 6-1 A). These two biozones are coincidentally marked by the change in lithologies in the BPC core from silts and clays (196 - 295 cm) to marls and peaty-marls (0 - 196 cm) -- see section 4.1.2 and Table 4-A; Chapter 4. It is not surprising that the change in lithology, which implies a fundamental change in depositional environment and therefore environmental conditions, is accompanied by changes in faunal assemblages.

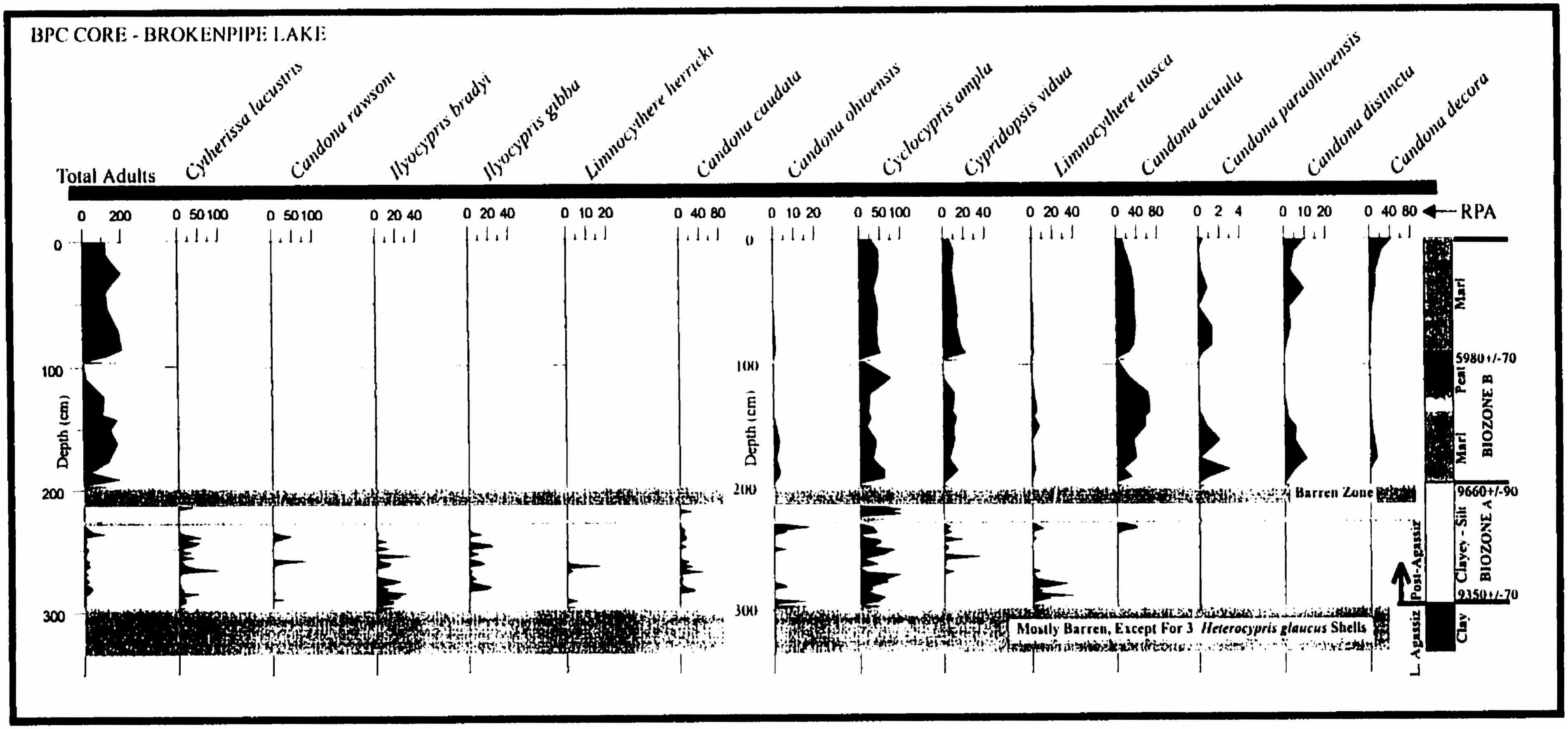


Figure 6-1 A: Ostracode relative percent abundance (RPA) diagram for the BPC core. General stratigraphy, radiocarbon ages, and biozones shown at far right.

Based on the autecology of the ostracodes in the BPC core (Section 4.5.1; Chapter 4), deposition during both Biozones A and B occurred in a body of water separate from Lake Agassiz. Ostracode autecology implies that environmental conditions during deposition of Biozones A and B were quite different. Although the sediments of the 299 - 335 cm interval of the BPC core (Section 4.5.1; Table 4-A; Chapter 4) were likely deposited in Lake Agassiz, the ostracode found in the Lake Agassiz sediment probably did not live in Lake Agassiz. In the literature to date, only the ostracode *Candona subtriangulata* has been found in Lake Agassiz sediments (Last *et al.*, 1994; Curry, 1997), where water temperature was cold, and TDS was <100 mg/l (Curry, 1997).

Some ostracodes appear in both biozones, suggesting that these individual species are able to adapt to wide-ranging environmental conditions. The relative distribution of the all the ostracode species found in the BPC core is given in Table 6-A.

6.1.2- The Lake Agassiz Sequence (Core BPC 335 - 299 cm)

The only ostracode found in the stiff, silty, dropstone-laden glaciolacustrine clays (Table 4-A; Figure 4-4 g; Chapter 4) was *Heterocypris glaucus*. To date, there has been no report of this ostracode being found in Lake Agassiz sediments. Because *Heterocypris glaucus*, with its modern occurrence being associated with elevated salinities (Smith, 1993), has never been found in Lake Agassiz sediments (nor is it common in modern lakes with salinity as low as that of Lake Agassiz), it is uncertain whether it actually lived in Lake Agassiz at the Brokenpipe Lake site.

Table 6-A: Ostracode Assemblages in Core BPC

Biozone A: (196 - 299 cm)	Biozone B: (0 - 196 cm)
Ostracodes limited to Biozone A:	**Ostracodes limited to Biozone B:**
Cytherissa lacustris	*Candona acutula*
Candona caudata	*Candona decora*
Ilyocypris bradyi	*Candona distincta*
Ilyocypris gibba	*Candona paraohioensis*
Candona rawsoni	*Cypris pubera* ***
Limnocythere herricki	
Candona inopinata **	
Candona sigmoides **	
Ostracodes in both Biozones A and B:	**Ostracode in Lake Agassiz Sequence**
Cyclocypris ampla	*Heterocypris glaucus*
Cypridopsis vidua	
Limnocythere itasca	
Candona ohioensis	
Candona acutula **	

** denotes minor occurrence in Biozone A

*** denotes minor occurrence in Biozone B

Considering that Curry (1997) has concluded that the presence of *Candona subtriangulata* in Lake Agassiz sediments of Lake Manitoba cores suggests Lake Agassiz waters were extremely dilute (TDS<100 mg/l) and cold, *Heterocypris glaucus* is not an ostracode one would expect to find in Lake Agassiz.

The co-occurrence of *Candona subtriangulata* and *Candona rawsoni* -- a euryhaline "prairie" taxa which lives in salinities roughly two orders of magnitude greater than that of *Candona subtriangulata* (Curry, 1997) -- in the Lake Agassiz portion of the Lake Manitoba sequence suggests, as similarly concluded in a study of Lake Michigan (Forester *et al.*,1994), that glacial meltwater at the time of deposition of these ostracodes was not flowing directly into the lake. The mean TDS occurrence of *Heterocypris glaucus* is slightly higher than that of *Candona rawsoni*, although the upper and lower tolerance limits for TDS of these two species are nearly identical (Delorme, 1989). Following this reasoning, the occurrence of *Heterocypris glaucus* (like *Candona rawsoni*) in Lake Agassiz sediments may also indicate that glacial meltwater was not directly entering Lake Agassiz at the Brokenpipe site. The occurrence of *Candona rawsoni* in Lake Agassiz sediments of the Lake Manitoba cores (Curry, 1997), and the occurrence of *Heterocypris glaucus* in Lake Agassiz sediments of the Brokenpipe Lake site, also suggests that these ostracodes were living at the lower (more dilute) end of their salinity tolerances.

A possible explanation for the presence of *Heterocypris glaucus* in the Lake Agassiz sediments of the BPC core is that the ostracode was living in peripheral pools of meltwater and surface runoff trapped in depressions around the shore of Lake Agassiz. These peripheral pools probably had relatively elevated temperature and salinity (due to significant evaporation and/or significant saline groundwater discharge) over that of the main body of Lake Agassiz. The *Heterocypris glaucus* shells were subsequently washed lakeward into Lake Agassiz to be deposited with the glaciolacustrine clays.

Isotope and trace element data from the *Heterocypris glaucus* shells in the Lake Agassiz clays provide information that will indicate if this ostracode actually lived in Lake Agassiz itself, or if it was reworked from a shallow, lake-marginal pool. Additional shells are necessary for these analyses.

6.1.3 - Summary of Biozone A (Core BPC 196 - 299 cm)

The presence of *Cytherissa lacustris*, a key indicator species in Biozone A, implies the lagoon during deposition of Biozone A was generally cold, with dilute waters of moderate depth. *Cytherissa lacustris* is a long-lived (>2 years) ostracode, and it is found throughout most of Biozone A (Fig. 6-1 A). The geomorphology of the basin, as reconstructed from relict lagoon shorelines and sediment cores (see Section 6.2.7) suggests that the water in the lagoon during deposition of Biozone A was between about 7.5 an 8 m deep. This result is in agreement with the modern depth requirements of *Cytherissa lacustris*. Because the lagoon during Biozone A was not in direct connection with Lake Agassiz (only *Candona subtriangulata* has previously been found in Lake Agassiz sediments; e.g. Last *et* al., 1994; Curry, 1997), the cool water temperatures in the lagoon were maintained by 1) general deglacial climate, and/or 2) periodic storm beach overwash of cold Lake Agassiz waters, whose average temperature was probably closer to 5°C (Mann *et al*, 1997; Curry, 1997).

On its own, *Cytherissa lacustris* is a good indicator of a lake with bottom waters undersaturated with respect to calcite, and TDS < ~300 mg/l. *Cytherissa lacustris* is also common in waters that show little seasonality in temperature change (with average water temperatures below about 15 - 20 °C), and it is indicative of a climate with precipitation > evaporation. This information implies that the cool, dilute salinity conditions in the lagoon during Biozone A must have been at least partially maintained by the cold, and wet climatic conditions common of a general deglacial climate, undoubtedly aided by the "lake effect" (Hu *et* al., 1997) of the nearby, cold mass of water of Lake Agassiz.

Other ostracodes from Biozone A are important to discuss as well. Ostracodes such as *Ilyocypris bradyi, Ilyocypris gibba,* and *Cyclocypris ampla* (Fig. 6-1 A) suggest flowing water was entering the lagoon during the deposition of Biozone A. The source of the water was likely stagnant ice on the Riding Mountain uplands, west of Brokenpipe Lake. Like *Cytherissa lacustris*, the presence of *Candona caudata* may suggest the lagoon was relatively cold, as it has been found in proglacial lake sediments in other localities (Brandon Curry, 1998, personal communication). However, *Candona caudata* can be quite cosmopolitan, as it has also been found living in spring-fed streams (Taylor, 1992).

Ostracodes such as *Candona rawsoni, Limnocythere herricki, Candona inopinata,* *Candona sigmoides, Cypridopsis vidua, Limnocythere itasca,* and *Candona ohioensis* (Fig. 6-1 A) suggest that the lagoon during deposition of Biozone A was subject to environmental variability (variations in temperature and salinity), though the spatial distribution of where different ostracode species lived has implications for this interpretation (Fig. 6-1 B). The presence of *Candona rawsoni* with *Cytherissa lacustris* suggests it is in the Great Lakes type of association (Colman *et al.*, 1990; Buckley, 1975); that is it is living within its lower range of survival for both temperature, and salinity. The environmental variability during Biozone A was predominantly a response of a relatively newly deglaciated watershed -- overshadowed by a deglacial climate, Lake Agassiz lake effect, and Lake Agassiz beach overwash -- to variable discharges of relatively warmer and more saline surface runoff, shallow saline groundwater discharge, and to a lesser extent evaporation. Any direct signature or connection of Lake Agassiz with the lagoon (such as periodic storm beach overwash events), in terms of the ostracodes found in Biozone A, was masked by this interaction of this surface and shallow groundwater input, climatic conditions, and the occupation of subenvironments of the lagoon by different groups of ostracodes.

In Biozone A, those ostracodes which suggest the lagoon was dominated by dilute waters (i.e. *Cytherissa lacustris*) were probably living within their upper tolerance range of salinity for survival, in the relatively more stable profundal areas of the lagoon. At the other end of the spectrum, ostracodes that indicate shallow, relatively warmer and more saline conditions (e.g. *Cyclocypris ampla,* and *Cypridopsis vidua*),

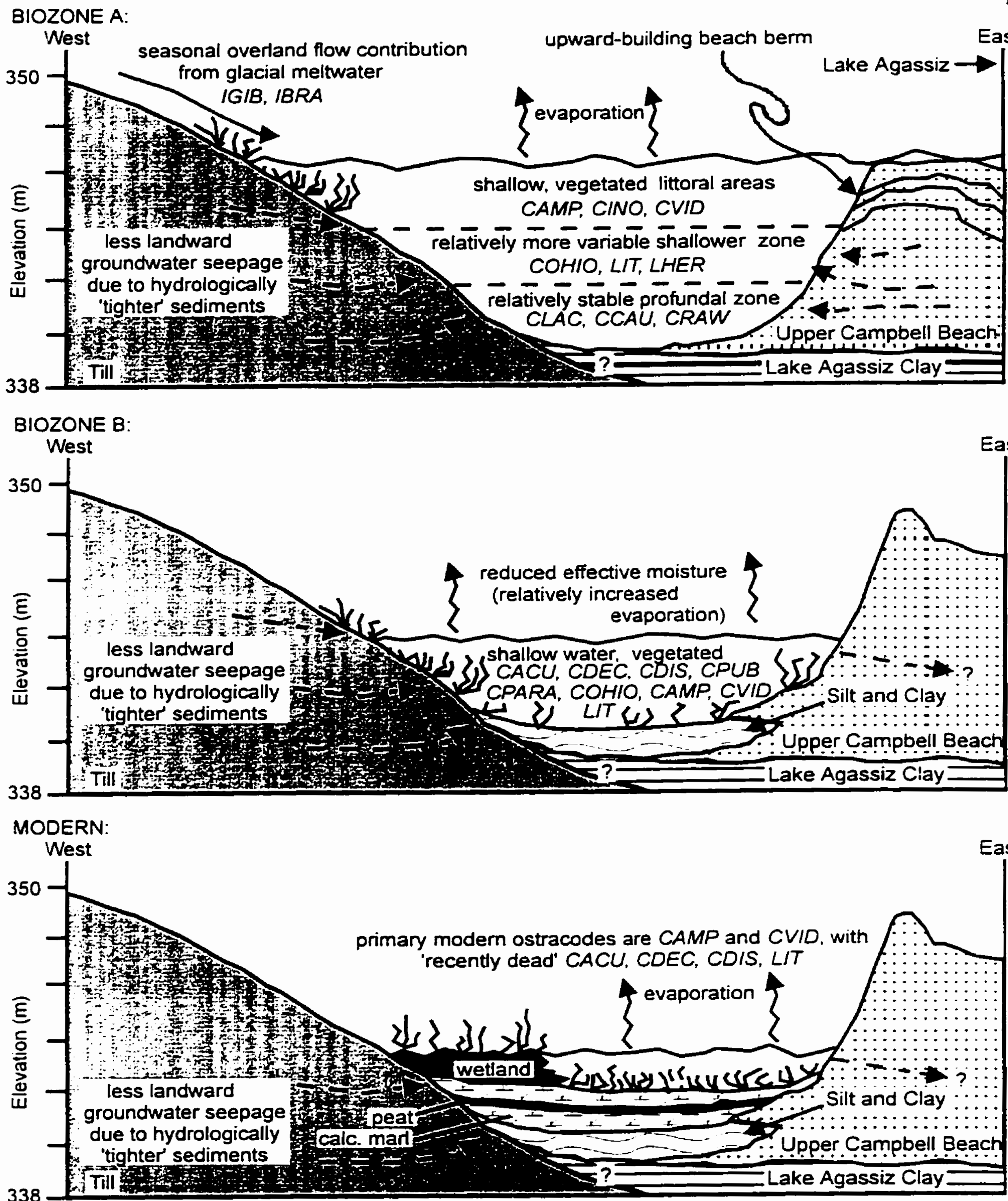


Figure 6-1 B: Schematic representation of a cross section through the Brokenpipe Lake site during deposition of Biozones A and B (upper two sections), and today (lowest section). *CAMP = Cyclocypris ampla, CVID = Cypridopsis vidua, CINO = Candona inopinata COHIO = Candona ohioensis, CCAU = Candona caudata, CRAW = Candona rawsoni CACU = Candona acutula, CDEC = Candona decora, CDIS = Candona distincta, CPARA = Candona paraohioensis, CPUB = Cypris pubera, CLAC = Cytherissa lacustris LIT = Limnocythere itasca, LHER = Limnocythere herricki, IGIB = Ilyocypris gibba, IBRA = Ilyocypris bradyi.* W - E distance across section is 1.75 km.

were living in the shallow, vegetated littoral areas of the lagoon (Fig. 6-1 B). In summary, the ostracodes of Biozone A suggest that the lagoon was; 1) not in constant connection with Lake Agassiz, where only *Candona subtriangulata* is found, 2) of moderate depth (>3m), 3) relatively cool (< 15-20 °C), and dilute (TDS concentrations were probably between about 200 and 300 mg/l) at depth, 4) subject to appreciable lotic input and evaporation, which in turn caused 5) moderate variation in temperature and salinity in shallower, vegetated areas of the lagoon.

6.1.4 - Summary of Biozone B (Core BPC 0 - 196 cm)

All of the ostracodes found only in Biozone B (*Candona acutula, Candona decora, Candona distincta, Candona paraohioensis,* and *Cypris pubera,* Fig. 6-1 A) are indicative of shallow water conditions and moderate to elevated (at least in relation to Biozone A) temperature and salinity. They are all tolerant of broad water temperature and salinity fluctuations, which generally are a result of shallow water depths. The fundamental shift (disappearance) of ostracode species in Biozone A, which indicate moderate water depths, low salinity, and cool temperatures (notably *Cytherissa lacustris*), to those in Biozone B which all indicate elevated temperature, salinity and shallower water depths, implies that effective moisture was reduced (i.e. conditions became drier - less precipitation and more evaporation) from that prevalent during deposition of Biozone A, and that the level of Lake Agassiz had fallen below levels that would have influenced the hydrology of the Brokenpipe site. Essentially, the lagoon had become a stand-alone lake during the deposition of Biozone B.

The large and increasing proportions of *Candona acutula* and *Candona decora* in Biozone B, coupled with the disappearance of *Ilyocypris gibba* and *Ilyocypris bradyi* suggests that water currents and water levels decreased -- assumedly because of a water budget shift to groundwater and atmospheric sources -- as Lake Agassiz no longer influenced basin hydrology, and also due to the cessation of overland deglacial meltwater flow into the Brokenpipe site (Fig. 6-1 B). The elevation of vegetation the fringing wetland was probably increasing, further buffering the incipient Brokenpipe Lake from water currents induced by inflowing streams. It is probable that the drowning of the stream channel at the southern end of the lake occurred at this time. The brief appearance of *Cypris pubera*, an ostracode indicative of lake senescence (*Cypris pubera* was only found at 150 and 123 cm in the BPC core) further supports the idea that not only was the lake shallowing, but terrestrialization of the lake basin (marked by a shift from marl production to peat production (Fig. 6-1 A) occurred at least once.

6.1.5 - Ostracodes Common to Biozones A and B

a) *Cyclocypris ampla*

Cyclocypris ampla is suggestive of wide variability in environmental parameters, and in the Biozone A association with ostracodes which inhabit deeper, colder and more dilute waters (e.g. *Cytherissa lacustris*), a reasonable interpretation for the presence of *Cyclocypris ampla* in Biozone A could be argued several ways: 1) it survives in dilute, cold water, is a "pioneering" type of species, and lived as such; 2) it probably thrived in the shallower, warmer, perhaps vegetated margins of the lagoon, and was washed lakeward after death to be deposited at the BPC core site; or 3) it thrived because of the aquatic niche set up by inflowing overland deglacial runoff to the lagoon, creating a combination of standing and flowing water (Fig. 6-1 B). It is likely that the occurrence of *Cyclocypris ampla* in Biozone A is a reflection of a combination of all of these three possibilities.

A Biozone B association of this ostracode with the shallow water candonids, suggests *Cyclocypris ampla* was living within the realm of higher temperatures and salinities, yet still within its normal range. A Biozone B occurrence of *Cyclocypris ampla* suggests a more eutrophic system with a highly vegetated littoral zone, shallower water depths than in Biozone A, resulting in higher water temperatures and moderate to high salinities (Fig. 6-1 B). It is unlikely that its presence in Biozone B is supported by the influx of water from the streams west of the site because *Cyclocypris ampla* is no longer found in association with *Ilyocypris bradyi* and *Ilyocypris gibba* in Biozone B (Figs. 6-1 A and 6-1 B). Lowered lagoon water levels, and the cessation of any appreciable overland flow entering the lagoon during the deposition of Biozone B created a different chemical environment in the lagoon as well, marked by the onset of marl accumulation (Fig. 6-1 A). Increased chemical weathering of the recently deglaciated watershed, and the onset of environmental conditions suitable for plant inhabitation combined to increase the proportion of solutes (namely bicarbonate and sulfate) to the early Brokenpipe Lake, which helped to initiate calcareous marl deposition.

b) *Cypridopsis vidua*

Cypridopsis vidua and *Cyclocypris ampla* are common littoral zones. Their presence with profundal species like *Cytherissa lacustris* in Biozone A (Fig. 6-1 B) suggests the shells were reworked to deep water prior to burial at the BPC core site.

In Biozone B, like *Cyclocypris ampla*, *Cypridopsis vidua* was exploiting its ability to tolerate more saline waters, and higher temperatures associated with shallower water depths prominent during deposition of Biozone B (Fig. 6-1 B).

c) *Limnocythere itasca*

The presence of *Limnocythere itasca* in Biozone A implies that chemical and temperature variation in the lagoon was moderate (Fig. 6-1 B). Overall, during the times when *Limnocythere itasca* was present in Biozone A , the lagoon was probably dilute, and perhaps had a slight predominance or enrichment of SO_4 over HCO_3 and Ca (Smith, 1993).

The presence of *Limnocythere itasca* in the Biozone B association suggests something slightly different. It survived well in the shallow, warm waters with relatively elevated salinities, prominent during Biozone B deposition (Fig. 6-1 B). In this association, its presence may also suggest a relatively closer proximity of the BPC core site to the shoreline of the lagoon (i.e. shallower water depths than during Biozone A).

The fact that *Limnocythere itasca* is found higher in the section (123 - 191 cm, Fig. 6-1 A) in Biozone B than *Candona ohioensis* (150 - 191 cm, Fig. 6-1 A) further implies that water depths in the lake were decreasing.

d) *Candona ohioensis*

The presence of both *Candona ohioensis* and *Limnocythere itasca* in Biozone A and B (Figs. 6-1 A and 6-1 B) perhaps suggests water chemistry varied between predominance or enrichment of SO_4 over HCO_3 and Ca (as *Limnocythere itasca* suggests), and situations of enrichment of HCO_3 (and Ca) over SO_4 (as *Candona ohioensis* suggests). These kinds of changes might occur due to sulfate reduction by bacteria in wetlands. This is a precursor of methanogenesis, which may account for the barren zone in Biozone B. Stable isotopes would perhaps indicate whether this did occur.

6.1.6 - A Paleoenvironmental Interpretation of the BPC Core, Based on Species Ratios and Weighted Means of Ostracode Autecological Data

A semi-quantitative analysis based on ostracode relative abundance was used to indicate changes in the chemical and physical environment of Brokenpipe Lake. Each centimeter interval sampled in each Biozone is represented by relative abundance changes among the ostracode species that are present. Changes in relative abundance may reflect the hydrochemistry of the host water, or water depth. Forester *et al.* (1994), for example, used ostracode species ratios to show relative changes in the distance from paleoshoreline or hydrochemistry of Lake Michigan during the Holocene.

For the Brokenpipe lagoon, a flowing water index (FWI) ratio was composed to indicate changes in influence of flowing water relative to standing water conditions at the

BPC core site. Species that tolerate both flowing and standing water (conditions met where streams flow into lakes, for example) and moderate salinity conditions, including *Cyclocypris ampla*, *Ilyocypris bradyi*, and *Ilyocypris gibba*, were ratioed against *Cytherissa lacustris*, a species that is indicative of profundal environments, and relatively more dilute waters.

FWI = *Cyclocypris ampla* + *Ilyocypris bradyi* + *Ilyocypris gibba*/ *Cytherissa lacustris* + 1

Adding 1 to *Cytherissa lacustris* avoids dividing by zero, if there are no *Cytherissa lacustris* shells found in a particular interval.

Large values imply that there were currents in the lagoon, either from a large volume of stream input, or because the BPC core site was near to the input of the water during the deposition of that interval. Low values imply there were no currents, either because the predominance of lotic input was relatively reduced, or because the lagoon was dominated by more purely lacustrine conditions (deeper water). This ratio was calculated only for the 237 - 293 cm interval (Biozone A) of the BPC core, the interval which contains relatively constant amounts of *Cytherissa lacustris*, and the ilyocyprids. Results are shown in Figure 6-2.

Along with the ratio measure discussed above, TDS and bottom water temperature were also investigated. I used mean TDS and mean bottom water temperature values for each ostracode given by Delorme (1989, 1991). The weighted

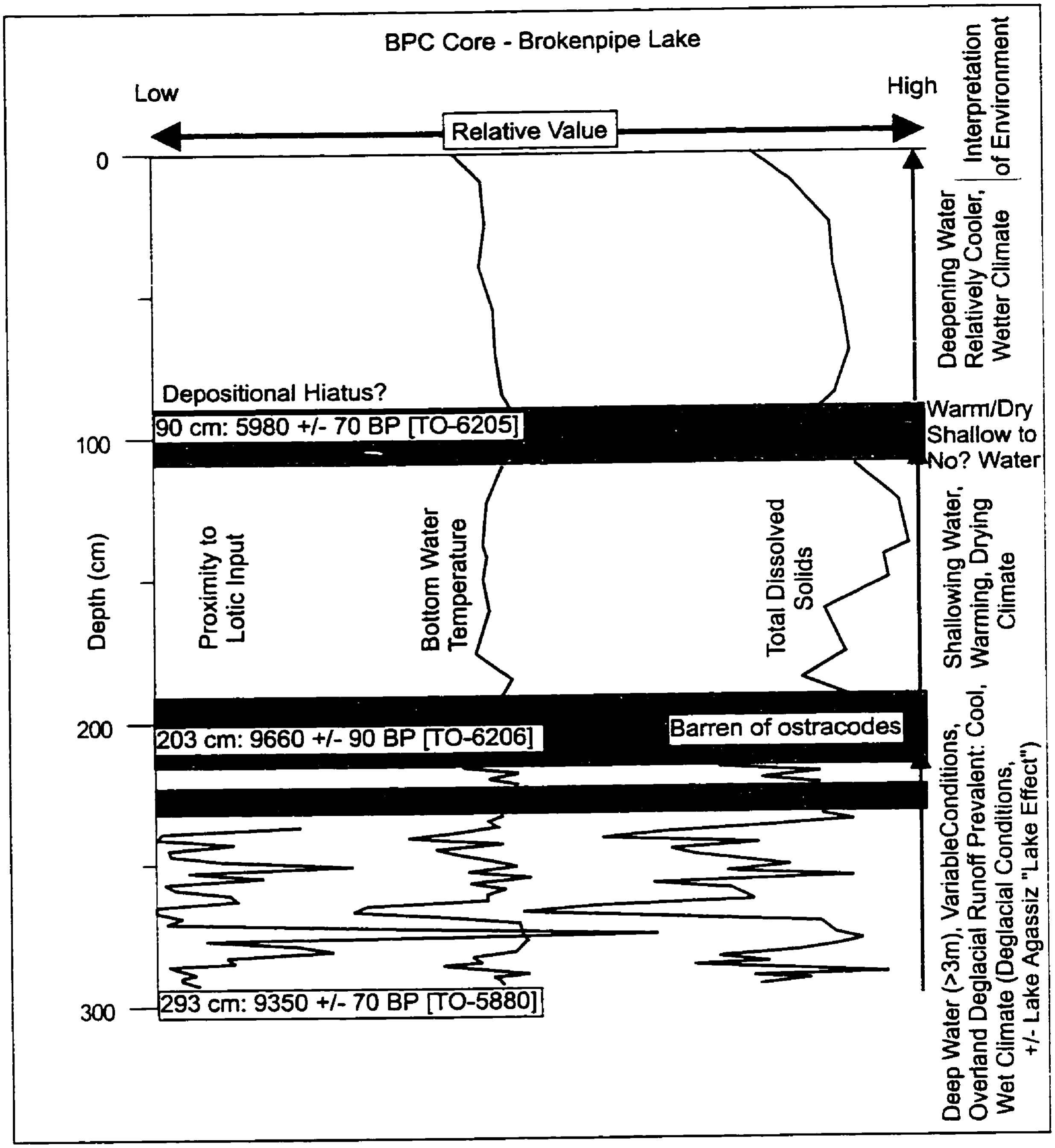


Figure 6-2: Representation of ostracode-determined changes in proximity or relative volume of lotic input, bottom water temperature, and total dissolved solids concentration for the BPC core. Curves represent relative values, see text for explanation. AMS wood radiocarbon dates shown in boxes at left side of the diagram.

mean value for each parameter was calculated per interval, by multiplying Delorme's values by the number of valves for each species. These products were summed for each sample interval, and divided by the total number of valves in that interval (see Delorme, 1971b). The results are shown in Figure 6-2.

Prior to these calculations, the ranges in TDS and bottom water temperature for each species was compared with one another in each interval. The ranges overlapped, suggesting that the species did co-habitate the lagoon, and were not reworked.

The values of the three indices are variable and co-variant in Biozone A (Fig. 6-2). The covarience implies that an increase in overland flow into the lagoon increased both bottom water temperature, and TDS of the lagoon. Any intermittent connection of the lagoon with Lake Agassiz, resulting in periodic influxes of dilute, cold water, may be masked by the increased lotic input of relatively saline waters (or proximity of this lotic input to the BPC core site), and relatively saline shallow groundwater seepage to the lagoon.

The extremely large relative increase in lotic input at 275 cm relates to an interval (271 - 277 cm) where there are no examples *Cytherissa lacustris* present, and there are large numbers of *Ilyocyprids*, and *Cyclocypris ampla*. Also, the 262 - 265 cm and 285 - 289 cm intervals are particularly interesting. These intervals contain clean sand beds that punctuate the general clayey-silt and silty-clay lithologies of the lower portion of the BPC core (see Table 4-A, Figs. 4-4 d and 4-4 f, Chapter 4). The question arises whether the sand was derived from stream alluvium deposited during floods, or from reworked beach sediments derived from over-beach inwashing of a stormy Lake Agassiz. The ostracodes do not help make the differentiation. Each of the discrete sand beds contain ostracodes that indicate there was a lotic input into the lagoon at the time of deposition of the sand intervals (e.g. the *Ilyocyprids* and the only occurrences of *Candona inopinata* and *Candona sigmoides* in the entire core in the 262 - 265 cm sand pulse; Fig. 4-4 d), and small numbers of other ostracodes are present in the sand pulses (e.g. *Candona caudata, Candona rawsoni, Cyclocypris ampla, Cypridopsis vidua, Limnocythere herricki*) that also suggest that the lagoon was not in direct connection to Lake Agassiz. If these sand units are representative of two distinct short-lived Lake Agassiz overwash events into the lagoon (rather than stream flood), relatively saline lotic input (channelized overland flow), perhaps in connection with relatively saline shallow groundwater inflow to the lagoon, has predictably served to disguise these Agassiz overwash events in the ostracode record.

The upper section of the BPC core (0 - 196 cm, Biozone B) supports the previous interpretations, based on the ostracode autecology, that Brokenpipe Lake was shallowing, and becoming characterized by ostracodes that flourish under variable temperature and salinity conditions. Of note in Figure 6-2 is the fact that both the bottom water temperature and Total Dissolved Solids (TDS) curves increase to the barren interval (corresponding to peat deposition) near 95 cm depth. These trends, along with the appearance of *Cypris pubera* during this interval, suggest that the deposition of the peat was a consequence of a warming, relatively drier climate, shallowing water depths, and terrestrialization of the Brokenpipe Lake site. The peat represents a hiatus in lacustrine deposition during Biozone B. After the peat deposition, both TDS and bottom water temperature follow a decreasing trend, likely a reflection of a shift to a relatively cooler and wetter climate, perhaps in conjunction with increasing groundwater levels, increased the water depth at the Brokenpipe site (causing decreases in bottom water temperatures and TDS) and the re-initiation of calcareous marl deposition. These events are discussed in more detail in the summary section to follow.

6.1.7 - Diatoms From the BPC Core

Dr. Jan Risberg, while on a Post-Doctoral research grant at the University of Manitoba, investigated the diatom assemblages in 12 intervals of the BPC core. Samples taken from 304 cm, 293 cm, and 290 cm did not yield any specific interpretation, though the lower sample (304 cm, from the Lake Agassiz clay with dropstones) may indeed reflect deeper Lake Agassiz conditions (Risberg, 1996, personal communication). Samples taken from 283 cm, 275 cm, 255 cm, 245 cm, 235 cm, and 210 cm (Biozone A) yielded diatoms that suggest variable conditions; in some samples brackish water forms are present. *Paralia sulcata*, a diatom found at the Wampum site (see Chapter 7) and in the Lake Agassiz sequence in Lake Manitoba, is periodically found in this sampling group (Risberg, 1996, personal communication). This perhaps indicates that there was an infrequent connection of the lagoon (during deposition of Biozone A) to Lake Agassiz. In particular, the sample taken from 210 cm contained very few diatoms; this indicates variable water depths and salinity (Risberg, 1996, personal communication). The upper part of the BPC core was sampled at 185 cm, 150 cm, and 110 cm (Biozone B). These samples indicate indifferently saline to freshwater conditions in a shallow water lake with no connection to Lake Agassiz (Risberg, 1996, personal communication). In general, Risberg's diatom study agrees with the interpretations based on ostracodes.

6.2 - SPECULATIVE SUMMARY OF DEGLACIATION FOR THE NORTHWESTERN LAKE AGASSIZ BASIN

6.2.1 - INTRODUCTION

Based on data collected from the lakes in this thesis, Upper Campbell and Lower Campbell strandline correlation and analysis, and the arguments of the preceding sections and chapters, a paleogeographic reconstruction of the deglaciation of a large portion of the northwestern Lake Agassiz basin was developed. Throughout the following discussion, several site specific paleolagoon diagram sequences will be interwoven with the larger, overall progressive sequence of deglaciation in the northwestern portion of the Lake Agassiz basin. The reconstruction presented in this study is not intended to preclude alternative interpretations, nor does it make any concrete inferences about the drainage of Lake Agassiz through the northwest outlet. Figures 6-12 and 6-13, presented at the end of the following summary, correlate the depositional events described in the text at each lagoon site to the changes in Lake Agassiz water levels from the Norcross level, through to the Lower Campbell level. Figures 6-12 and 6- 13 include the possibility of the Gregory Lake site being located behind either the Upper or Lower Campbell beach.

6.2.2 - Period 1: Figure 6-3 A (~11.5 ka BP)

During this time interval, Lake Agassiz overflow was shifting from exiting the southern outlet (during the Herman levels) to the eastern outlets, as northward ice retreat uncovered them. The level of Lake Agassiz responded to this outlet reorganization by

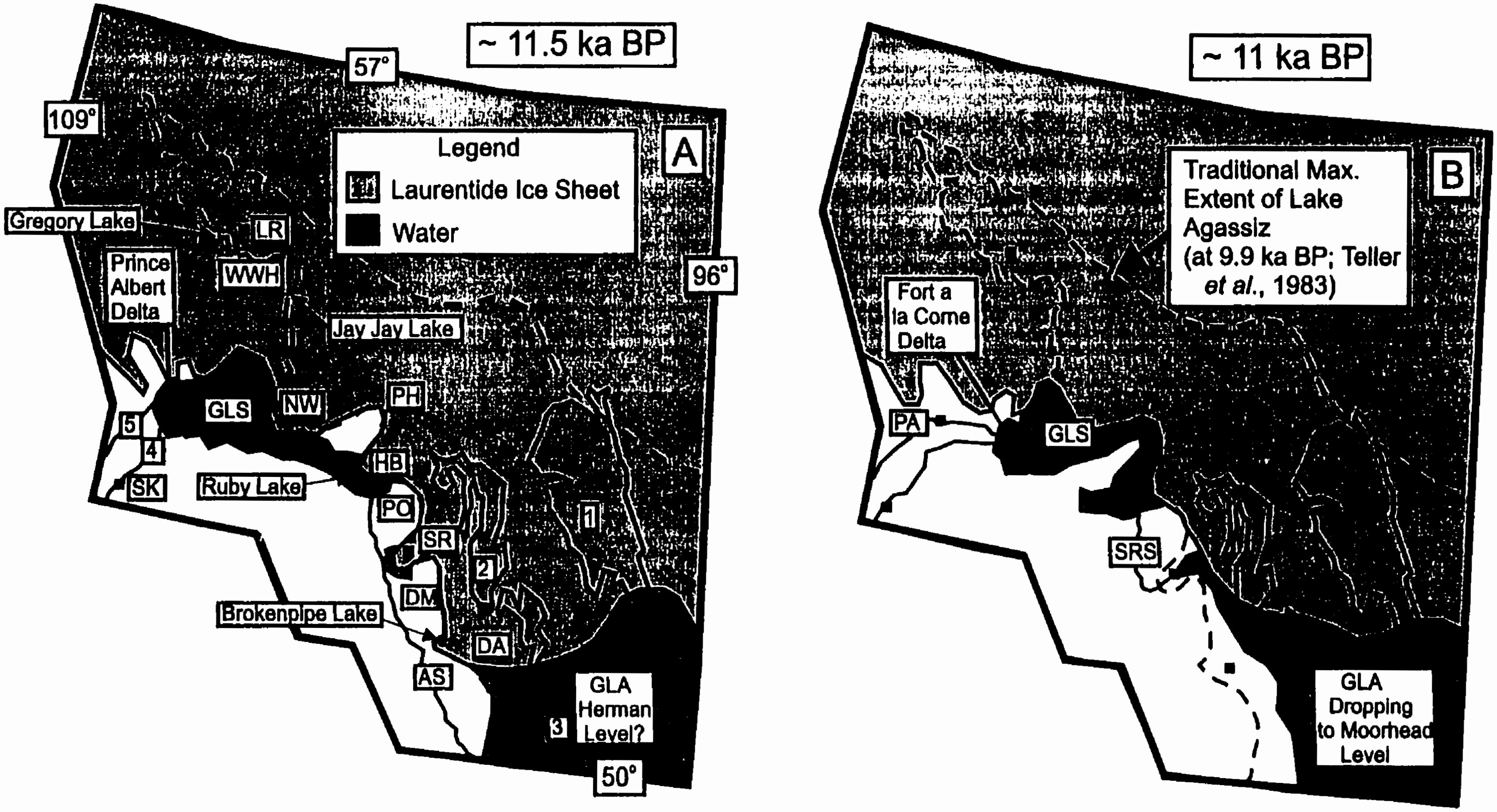


Figure 6-3 A-B: Speculative Laurentide Ice Sheet position and extent of glacial Lakes Agassiz (GLA) and Saskatchewan (GLS) during part of the last deglaciation (about 11.5 - 9.3 ka BP), see text for details. 1= Lake Winnipeg, 2 = Lake Winnipegosis, 3 = Lake Manitoba, 4 = South Saskatchewan River, 5 = North Saskatchewan River. Locations in Saskatchewan include: SK = Saskatoon, PA = Prince Albert, LR = La Ronge, WWH = Wapawekka Hills, NW = Nipawin, PH = Pasquia Hills, HB = Hudson Bay. Locations in Manitoba include: PO = Porcupine Hills, SR = Swan River, DM = Duck Mountains, DA = Dauphin. AS = Assiniboine Spillway, SRS = Swan River Spillway. (compiled from this thesis; Christiansen,1979; Christiansen *et al.*, 1995; Nielsen, 1987; Thorleifson, 1996; Langford, 1973; Teller *et al.*, 1983; Teller, 1985, 1987)

beginning its decline to the Moorhead level. The Laurentide Ice Sheet was abutted against the Manitoba Escarpment south of Swan River, Manitoba, at the Duck Mountain. In Saskatchewan, glacial Lake Saskatchewan was at about 1600 feet in elevation (488 m), and meltwater flowed into it through several channels, in particular the North and South Saskatchewan Rivers, depositing the Prince Albert delta (Christiansen, *et al.*, 1995). Lake Saskatchewan drainage at this time (Fig. 6-3 A) was into Lake Agassiz via the Assiniboine River spillway (Christiansen *et al.*, 1995).

6.2.3 - Period 2: Figure 6-3 B (~ 11 ka BP)

Continued northwestward ice retreat allowed Lake Agassiz to expand northward into the Swan River Area (Fig. 6-3 B). Lake Agassiz continued to drain through the eastern outlets, and at the Brokenpipe Lake site the level of the lake was declining. Brokenpipe Lake was subaerially exposed by the time Lake Agassiz reached the Moorhead level. Both the Jay Jay Lake and Gregory Lake sites were covered by ice at this time.

Lake Agassiz was separated from Lake Saskatchewan by ice abutted against the Porcupine Hills of Manitoba and Saskatchewan (Fig. 6-3 B). Lake Saskatchewan was at about 1500 feet in elevation (457 m), and water flow through the North and South Saskatchewan Rivers into Lake Saskatchewan constructed the Fort a la Corne delta at this time (Christiansen *et al.*, 1995). Lake Saskatchewan drained around the Pasquia Hills into a glacial lake which covered the Hudson Bay, Saskatchewan area, including Ruby

Lake. From Lake Saskatchewan, meltwater entered into the Swan River Spillway (Fig. 6-3 B), constructing a delta west of the town of Swan River, Manitoba, where the spillway entered Lake Agassiz (Christiansen *et al.*, 1995).

In the Hudson Bay area, the clay with weathered shale clasts, found at the base of the RLA, RLB, HBU and HBL cores (Fig. 4-8 b) may have been deposited in Lake Saskatchewan at this time. Shale bedrock, exposed in slumps and scarps along the sides of the Pasquia Hills, was carried by meltwater (Moran, 1969) into Lake Saskatchewan. There is at least one strandline in the Hudson Bay vicinity, on the Pasquia Hills (see Fig. 6-3 A for location), that is many meters in elevation above the Upper Campbell beach (Ayres *et al.*, 1978). It is likely that this strandline, and the associated high-elevation glaciolacustrine sediments, are related to a smaller proglacial lake that was trapped at high elevations against the bedrock upland of the Pasquia Hills, and the Laurentide Ice Sheet, namely glacial lake Saskatchewan (e.g. Christiansen *et al.*, 1995).

Variably-textured colluvial sands, clays and shales intermixed with sandy and gravelly beach deposits (Ayres *et al.*, 1978) are found at elevations well above the Upper Campbell strandline on the southeastern and northwestern sides of the Pasquia Hills (up to 427 m in this area, about 30 m higher than the Upper Campbell beach). Also, high elevation (~ 610 m) bedrock scarps, and a veneer of glaciolacustrine sediments (Campbell, 1987) at elevations up to about 550 m suggests that there was another, high elevation proglacial lake present in this region, pinned by ice at elevations higher than those associated with the strandlines and glaciolacustrine sediments of Lake Agassiz. In the Pasquia Hills region, reconstructions of proglacial Lake Saskatchewan (Christiansen, 1979; Christiansen *et al.*, 1995) place Lake Saskatchewan at an elevation of 457 m against the Pasquia Hills, while covering the Hudson Bay region with water (Fig. 6-3 B).

6.2.4 - Period 3: Figure 6-3 C (~10.9 - 10 ka BP)

As ice continually retreated from the Lake Agassiz basin, the northwestern margin of the Laurentide Ice Sheet paused for a period of time and constructed the Beaver River Moraine. The ice margin position shown in Figure 6-3 C is similar to that shown in Figure 17 of Christiansen (1979). During this time, southwest of the Beaver River Moraine, glacial Lake Meadow formed, and ultimately reached an elevation of about 488 m (Schreiner, 1983). Lake Meadow (Fig. 1-12) drained by way of the Clearwater-Lower Athabasca Spillway (CLAS, see Section 1.2.6; Chapter 1) no later than 9935 BP (Schreiner, 1983), and most probably at around 10.5 ka BP (Fisher and Smith, 1994; Teller, 1985). By this time, the Wapawekka Hills had not yet been deglaciated, as the ice maintained its position at the Beaver River Moraine and Wapawekka Hills to keep waters from glacial Lakes Meadow and Agassiz separated.

It is possible (Thorleifson, 1996), however, that at some point between about 10.5 ka BP (the draining of glacial Lake Meadow) and about 10 ka BP (the Marquette readvance of the Laurentide Ice Sheet), that ice retreat from the Wapawekka Hills and the Beaver River Moraine opened a corridor for the drainage of the Moorhead level Lake Agassiz through the northwest outlet (CLAS).

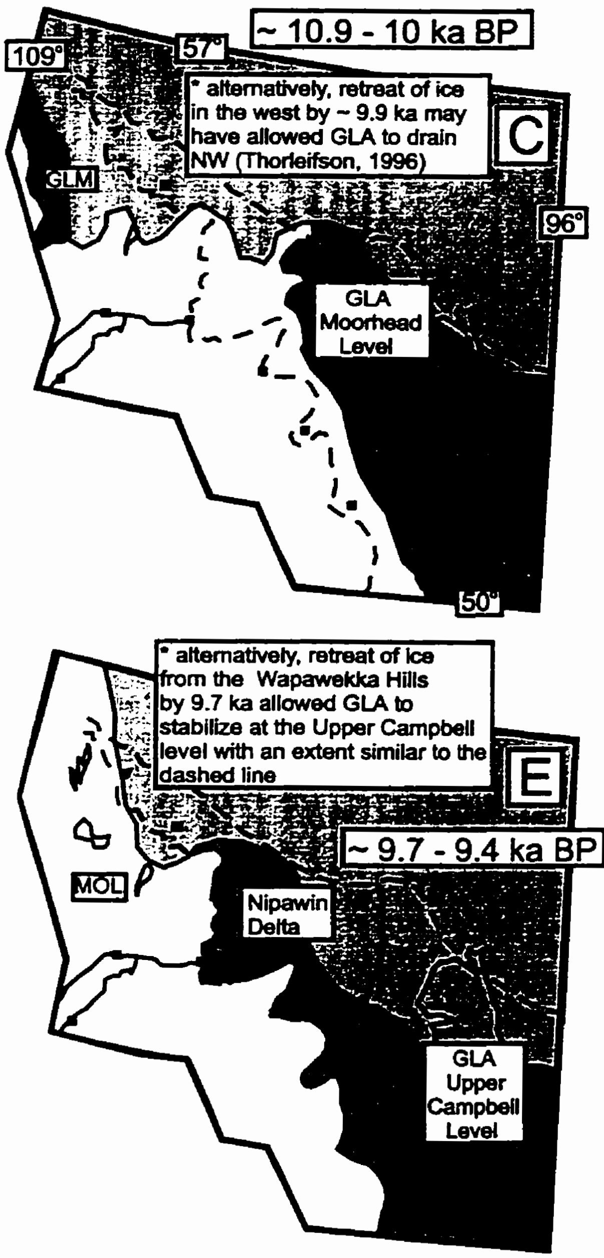


Figure 6-3 C-F: Speculative Laurentide Ice Sheet position and extent of glacial Lake Agassiz (GLA) during part of the last deglaciation (about 11.5 - 9.3 ka BP), see text for details. MOL = Montreal Lake. GLM = glacial Lake Meadow.

6.2.5 - Period 4: Figure 6-3 D, 6-12 & 6-13 (~9.9 ka BP)

A major ice readvance at about 10 ka BP (coincident with the Marquette readvance in the Lake Superior basin; Section 1.2.3) saw drastic ice margin advances that were relatively more extensive and widespread in the main Agassiz basin to the east, and less so in the Wapawekka Hills region of the western Agassiz basin. The difference in ice margin advance distances between the northwestern arm of Lake Agassiz and the main Agassiz basin was likely due to topographic control (by the Wapawekka, Cub, and Thunder Hills, for example) on the advancing ice sheet. The ice sheet in the main Lake Agassiz basin was buoyed by deep water of the lake, making ice advance in this portion of the basin relatively rapid, and the ice thin, as described by Clayton *et al.* (1985) for all late-glacial readvances in the Lake Agassiz basin. As a consequence, the advancing ice sheet in the main Lake Agassiz basin probably calved many icebergs. If the ice had previously retreated from the Beaver River Moraine and Wapawekka Hills in north-central Saskatchewan (which would be the case if Lake Agassiz drained through the northwestern outlet during the Moorhead level; Thorleifson, 1996), the readvance of ice west of the Wapawekka Hills area could not have overridden the Beaver River Moraine, or the mouth of the northwest outlet because these features do not show any evidence of being overridden by ice (Fisher and Smith, 1994). A 9935 +/- 170 BP [S-964] date from a lacustrine sandy silt containing wood and shells (Schreiner, 1983) suggests that at the latest, the final draining of Meadow Lake could have been occurring at this time, though other reconstructions of the chronology of lake Meadow place its final draining at an earlier (10.5 ka BP) date (Christiansen, 1979; Elson, 1967; Teller, 1985; Fisher and Smith, 1994).

6.2.5.1 - Ice Margin History in the Wapawekka Hills Area

Schreiner and Alley (1975) suggested a local readvance occurred during deglaciation in the La Ronge area, a region dominated by sandy till. A glacier readvance over areas with lacustrine clay was proposed to explain the presence of clay-rich tills in the La Ronge area (Schreiner and Alley, 1975). Schreiner and Alley (1975) described this readvance as being relatively limited and topographically controlled (Langford, 1977). If the hypothesis of Schreiner and Alley (1975) for a local readvance in the Wapawekka Hills area is true, Langford (1977) expected to see widespread occurrences of sandy till overlain by lacustrine sediments, in turn overlain by clayey till (formed by ice overriding lacustrine sediments) -- which is not the case in the field. Elson (1967) had observed a till over lacustrine deposits in the Sturgeon-Weir River basin (Fig. 6-4). Unfortunately, at a regional scale, the relationships between Pleistocene units in the Churchill and Reindeer River Valleys (Fig. 6-4) are much too complicated to be accounted for by a single glacial readvance (Langford, 1977). Further examination of the tills of the area, and some of the associated structureless sand and gravel units led Langford (1977) to conclude that many of these deposits were a result of downslope flow in a lacustrine environment. Furthermore, there is no evidence for terminal moraines to indicate a readvance east of the Wapawekka Hills (Langford, 1977).

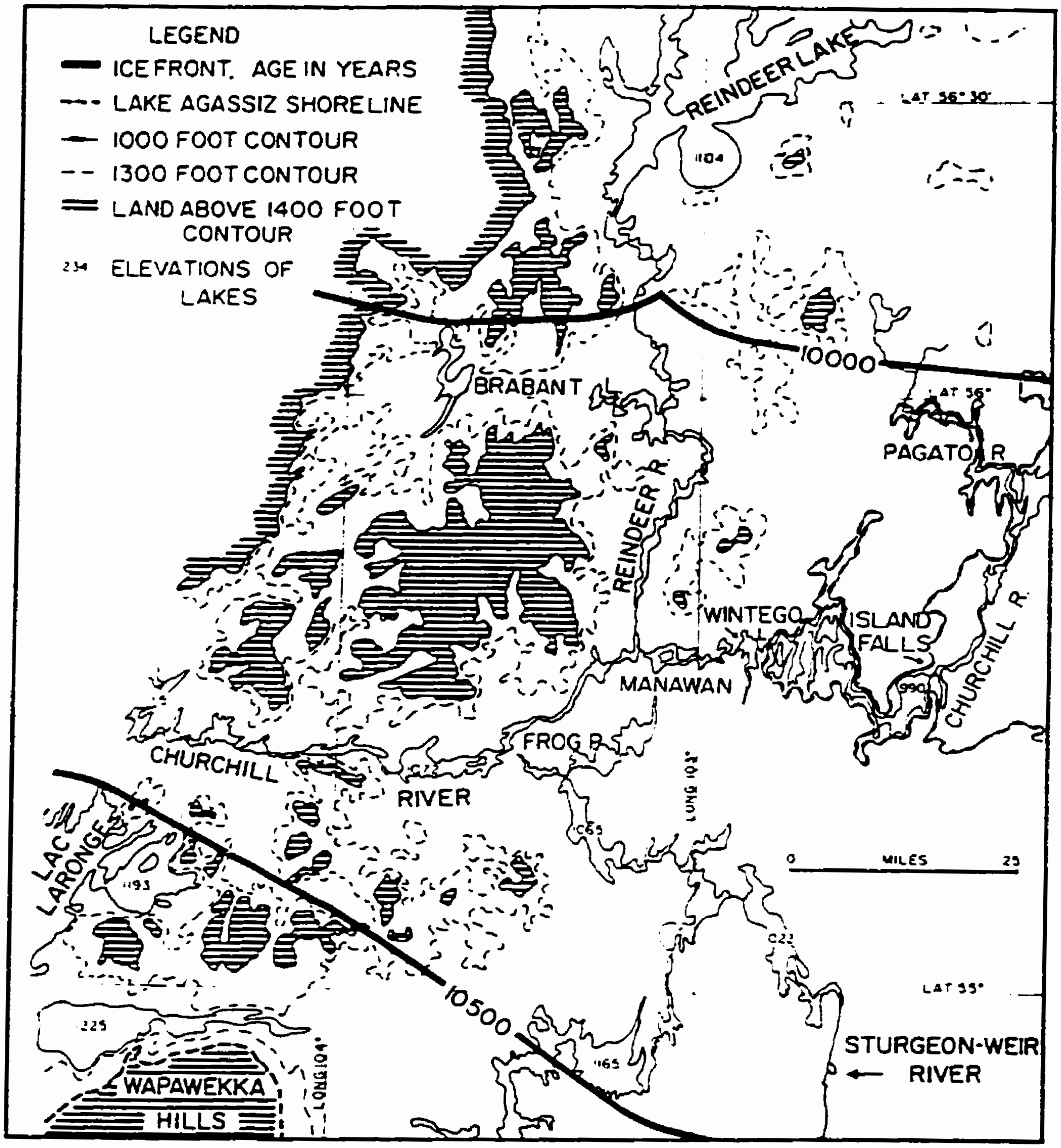


Figure 6-4: The area of the Churchill and Reindeer Rivers in Saskatchewan showing contours, geographic locations, and retreating Laurentide Ice Sheet positions (after Prest, 1970). (Langford, 1977, p.1287, fig. 1)

There is, however, a ridge on the north side of the Wapawekka Hills, at 1550 - 1600 ft (~ 473 - ~ 488 m) elevation which Rayburn (1997) interpreted as a moraine. The GPS elevation of the apex of this ridge is about 488.6 m (~1602 feet). From the level of the small lake trapped between the Wapawekka Hills and the ridge, the land rises steeply by approximately 15 m to the apex of the ridge, and the feature is continuous for more than ~ 3 km. It is generally composed of sand and gravelly sand (angular to sub-angular clasts) and contains boulders. There has been no detailed analysis of this feature to confirm that it is a moraine, but it lies well above the Upper Campbell strandline, and has no correlatives. It has been mapped as a beach strandline on soils and surficial geology maps (Head *et al.*, 1981; Simpson, 1988), however its size, morphology and sediments suggest that it is a deposit emplaced in association with ice against the Wapawekka Hills. In air photos, and in oblique air photos (Fig. 6-5) this deposit is a distinct ridge.

Ice margin positions, and chronologies, have been mapped in northwestern Saskatchewan by several workers (e.g. Dyke and Prest, 1987; Christiansen, 1979; Schreiner, 1984; Christiansen *et al.*, 1995), and traditionally the 10 ka BP ice margin position has been placed (west of the Wapawekka Hills) at the Cree Lake Moraine and at the southern end of Reindeer Lake, some 150 km north of the Wapawekka Hills (Fig. 6-6).

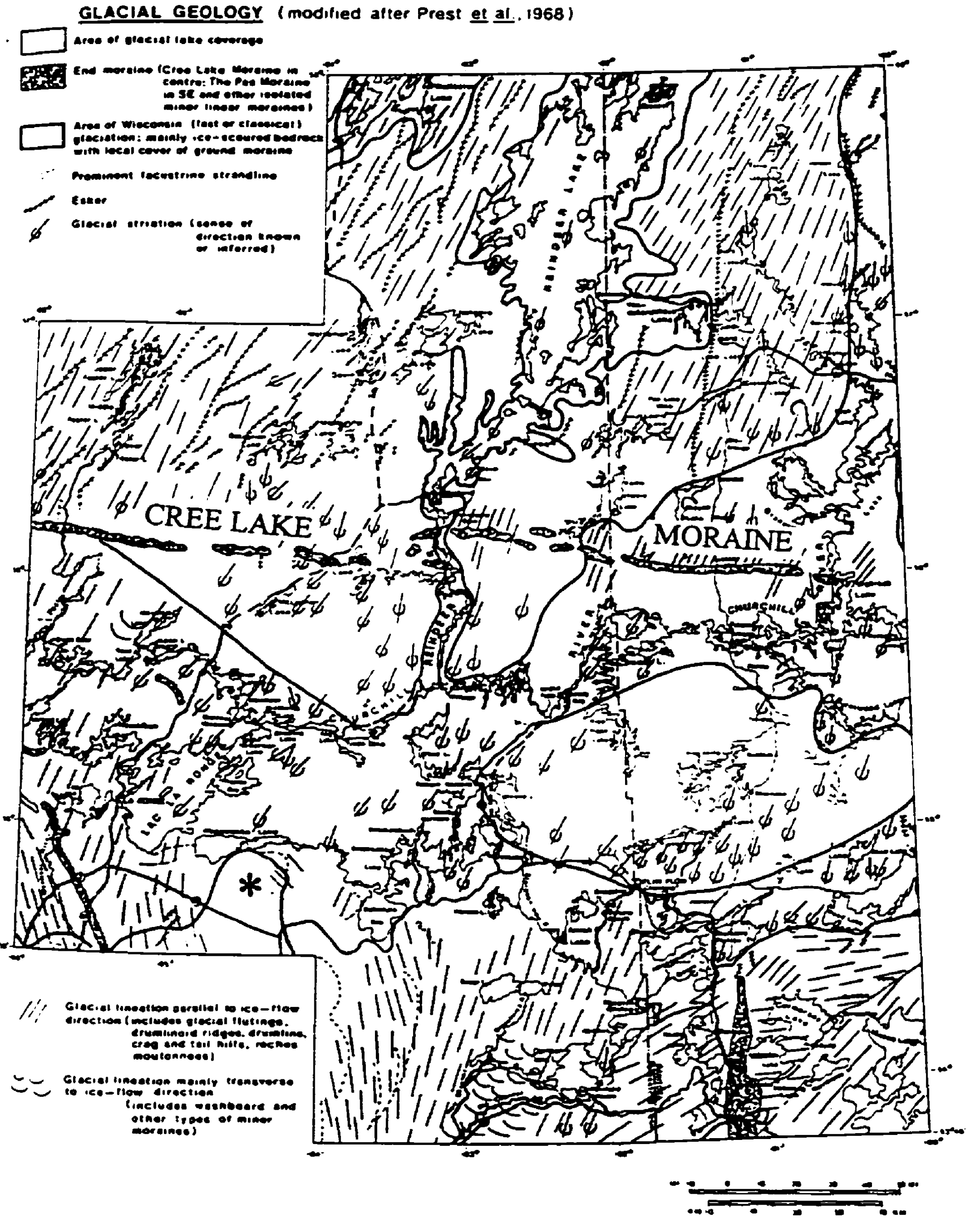


Figure 6-6: Glacial features of the northeastern Saskatchewan and northwestern Manitoba areas. The Wapawekka Hills are located by the '*'. (Kupsch, 1974, p. 17, Map 3)

Dates on either side of the Cree Lake Moraine itself are much younger (Mott, 1971) than 10 ka BP. These dates are minimum ages for deglaciation (see Fig. 6-7; 8640 +/- 240 yr BP [GSC-1446] south of the moraine, and 8230 +/- 250 yr BP [GSC-1466] north of the moraine; Mott, 1971) and suggest that the ice sheet could have been forming the Cree Lake Moraine after the end of the Emerson. As such, the ice sheet margin may have been in a position nearer the Wapawekka Hills during the Emerson Phase. In fact, Thorleifson (1996) places the ice margin at the Cree Lake Moraine about 8.2 ka BP, also suggesting the Laurentide Ice Sheet was covering or was in close proximity to the Wapawekka Hills during the Emerson Phase. Furthermore, since the 10 ka BP age assigned to the Cree Lake Moraine is based on Lake Agassiz correlations by Prest (1970) and Klassen (1989) (Lemmen *et al.*, 1994), the young dates and circular nature of the argument (assessing an age for the Cree Lake Moraine based on what Lake Agassiz is *believed* to be doing) should raise doubt as to the accepted Cree Lake Moraine ice front date of 10 ka BP.

Ice margin positions published by Prest (1970) suggest that when Agassiz was at 427 m elevation at the Wapawekka Hills, the ice front was at the Cree Lake Moraine and at the southern end of Reindeer Lake, 150 km to the northeast of the Wapawekka Hills (Fig. 6-6). Models of isostatic rebound developed by Broecker (1966), Walcott (1970), Teller and Thorleifson (1983), and data given by Elson (1967) suggests that with the ice front at Reindeer Lake (Figs. 6-6; 6-7), that area would be 45 to 90 m lower with respect to the Wapawekka Hills than it is now (Langford, 1977). If this were the case, most of

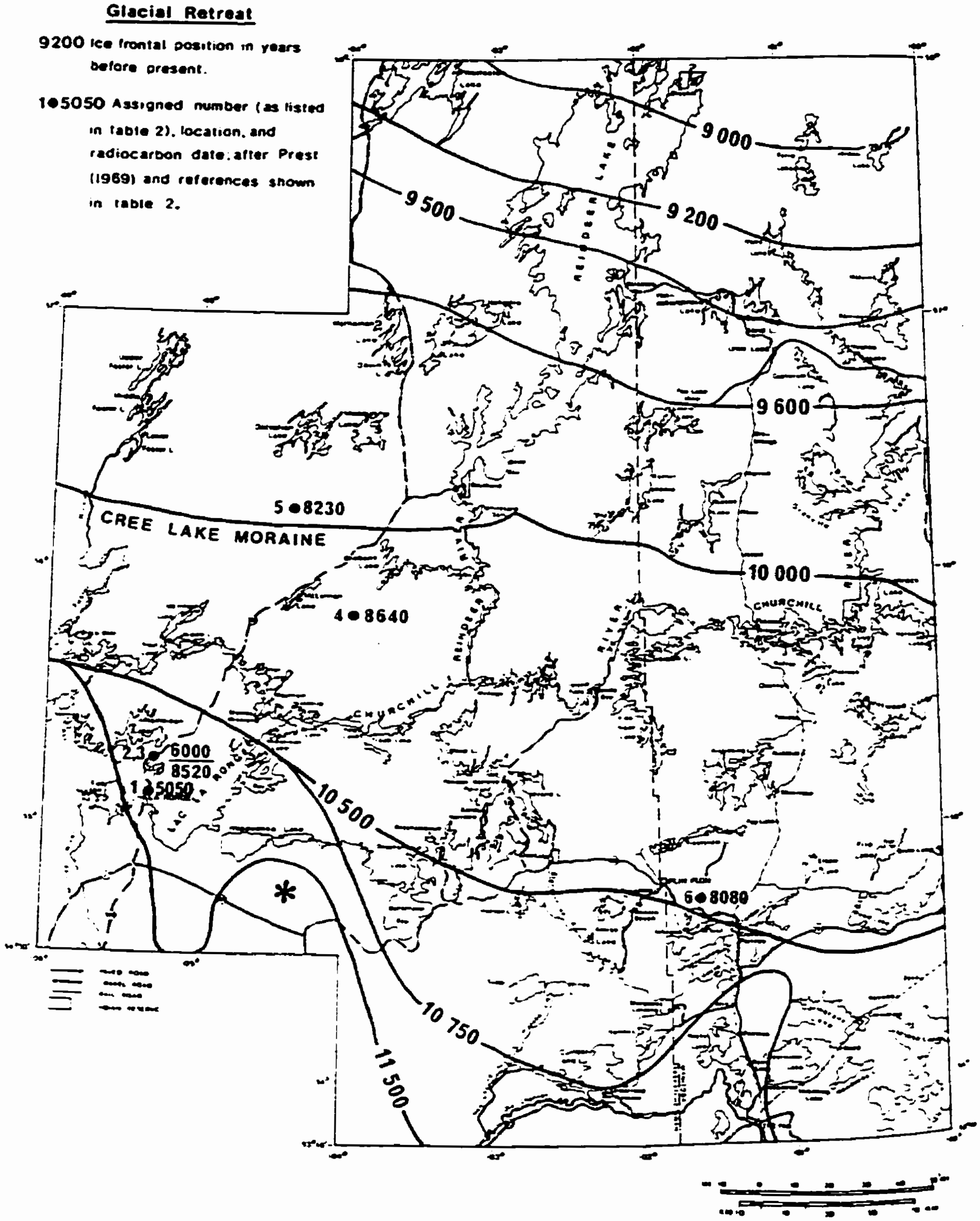


Figure 6-7: Laurentide Ice Sheet positions during retreat in northeastern Saskatchewan and northwestern Manitoba. The Wapawekka Hills are located by the '*'. (Kupsch, 1974, p. 22, Map 4).

the land now above 427 m, perhaps as high as 519 m would have been flooded by Lake Agassiz (Langford, 1977). The absence of extensive lacustrine deposits above an elevation of 427 m in the Reindeer Lake area (Fig. 6-6) suggests that this did not occur (Langford, 1977). The ice sheet could therefore be interpreted to have been in a position closer to the Wapawekka Hills, rather than being at the south end of Reindeer Lake (and at the Cree Lake Moraine) at about 10 ka BP (i.e. the ice was at the 11 500 BP or 10 750 BP position in Fig. 6-7 at 10 ka BP). Thus, *if* the Laurentide Ice Sheet had retreated north of the Wapawekka Hills by about 10 ka BP, it had not withdrawn very far from them. The ensuing ice readvance at ~ 10 ka BP would have pushed up against the Wapawekka Hills, or may have overridden the Hills completely. If the ice had previously retreated from this area to allow Lake Agassiz drainage through the northwest outlet during the Moorhead Phase (Thorleifson, 1996), the subsequent readvancing ice margin could not have overridden the Beaver River Moraine (Fig. 1-12), or the mouth of the northwest outlet (Fisher and Smith, 1994).

After the ice had readvanced in the Wapawekka Hills region about 10 ka BP, a relatively small proglacial lake began to form in an interlobate area, in the vicinity of modern Montreal Lake, south of the Gregory Lake site (Fig. 6-3 D). This small proglacial lake formed in a depression trapped between the Thunder Hills, and the ice sheet on the northern sides. This glacial lake was entrapped at elevations between about 515 m and 580^{+} m against the Thunder Hills (where strandlines are found today; Langford, 1973), a lobe of ice west of the Wapawekka Hills, and an eastern lobe that infilled the low area between the Wapawekka and Thunder Hills. Langford (1973) first saw evidence for this lake, and he called it glacial Montreal Lake. Lacustrine deposition in glacial Montreal Lake was a draping of fine sands about 30 cm thick; in places the sand is deposited directly over till (Langford, 1973). Because of the distribution of these sediments, Langford (1973) suggested that glacial Montreal Lake was ice-dammed to the north, and that since its strandlines along the Thunder Hills are higher in elevation than all known Lake Agassiz strandlines in this region, it likely predated Lake Agassiz' occupation of the area. Glacial Montreal Lake must have also been a product of the last ice readvance from the Thunder Hills area, since its sediments and shorelines are well-preserved.

THE GREGORY, JAY JAY, AND RUBY LAKES AREAS

If ice did readvance to the Wapawekka Hills and south into Manitoba at about 10 ka BP, the Gregory, Jay Jay and Ruby Lake areas probably were all buried by ice at this time (Fig. 6-3 D). The ice margin in this part of the main Lake Agassiz basin abutted against the Manitoba Escarpment at least as far south as Swan River in southern Manitoba. This basis of this conclusion is the absence of Norcross strandlines (which formed at this time) to the north of Swan River. The ice readvance marked the inception of the Emerson Phase, and the initial Emerson highstand of Lake Agassiz reached the Norcross level (Thorleifson, 1996; see Section 1.2.5, Chapter 1).

THE BROKENPIPE LAKE SITE

With Lake Agassiz at the Norcross waterplane, strandlines formed (Labeled 1, Fig. 4-2, Chapter 4) at elevations in excess of 381m near the Brokenpipe Lake site. This is about 32 m higher than the Upper Campbell strandline, placing the Brokenpipe Lake site under at least 37.5 m of water at this time. When Lake Agassiz was at the Norcross level, outwash sand plains (Labeled 2, Fig 4-2, Chapter 4) previously deposited subaerially by the Valley and Wilson Rivers during the Moorhead Phase were trimmed to the Norcross shoreline elevation. Because of the great nearshore depth of Lake Agassiz at the Norcross level, new outwash sands deposited into Lake Agassiz by the Valley and Wilson Rivers during the Norcross level were carried lakeward, and were dispersed by Lake Agassiz water currents.

The stiff, smooth, clay with ice-rafted dropstones found in the 299 - 335 cm interval of the BPC core (Section 4.1.2, Figure 4-4 g, Table 4-A) was probably deposited in Lake Agassiz during the Norcross Phase, and perhaps even during declining lake levels from the Norcross highstand to the Upper Campbell level (Fig 6-12 and 6-13). The proximity of the BPC core to Upper Campbell beach berms precludes deposition of the clays in shallow water at the Upper Campbell level. Alternatively, this clay may have been deposited during the Lockhart Phase, when Lake Agassiz was at a Herman level, though it is likely that Brokenpipe Lake was covered by ice at that time (Teller, 1985).

The presence of *Heterocypris glaucus* in the Lake Agassiz sediments of the BPC core (299 - 335 cm) suggests (see Section 6.1.2) that meltwater from the Riding

Mountain Uplands was not directly entering Lake Agassiz. The ostracodes were living in Lake Agassiz-marginal pools that collected saline groundwater and overland meltwater; the valves were redeposited post-mortem into the nearshore environment of Lake Agassiz and were deposited at the BPC core site.

6.2.6 - Period 5: Figures 6-3 D 6-8 A to C, 6-10 A, 6-12 & 6-13 (~ 9.8 ka BP) WAPAWEKKA HILLS AREA

Glacial Montreal Lake probably began to decrease in size, due to shrinkage, thinning, and retreat of the ice that formed its northern margins. Ice thinning and retreat undoubtedly opened new low elevation drainage routes, and the water of glacial Montreal Lake probably began to drain eastward toward the Lake Agassiz basin, via a network of large subglacial tunnel valleys (mapped by Langford, 1973) which are found throughout the area. Lake Agassiz had yet to expand as far northward as the Wapawekka Hills area.

A speculative chronologic sequence of events that document the development of the Jay Jay Lake site, and the first inundation of Lake Agassiz in the Jay Jay Lake/Wapawekka Hills area, is given in Figure 6-8 A - E. As ice thinning progressed, (Fig. 6-8 A), and an ice lobe remained in the low-lying area between the Cub and Wapawekka Hills. A small, short-lived, probably partially supraglacial lake developed, pinned by the ice along the east, and was fed by drainage from the Dowd Lakes channels.

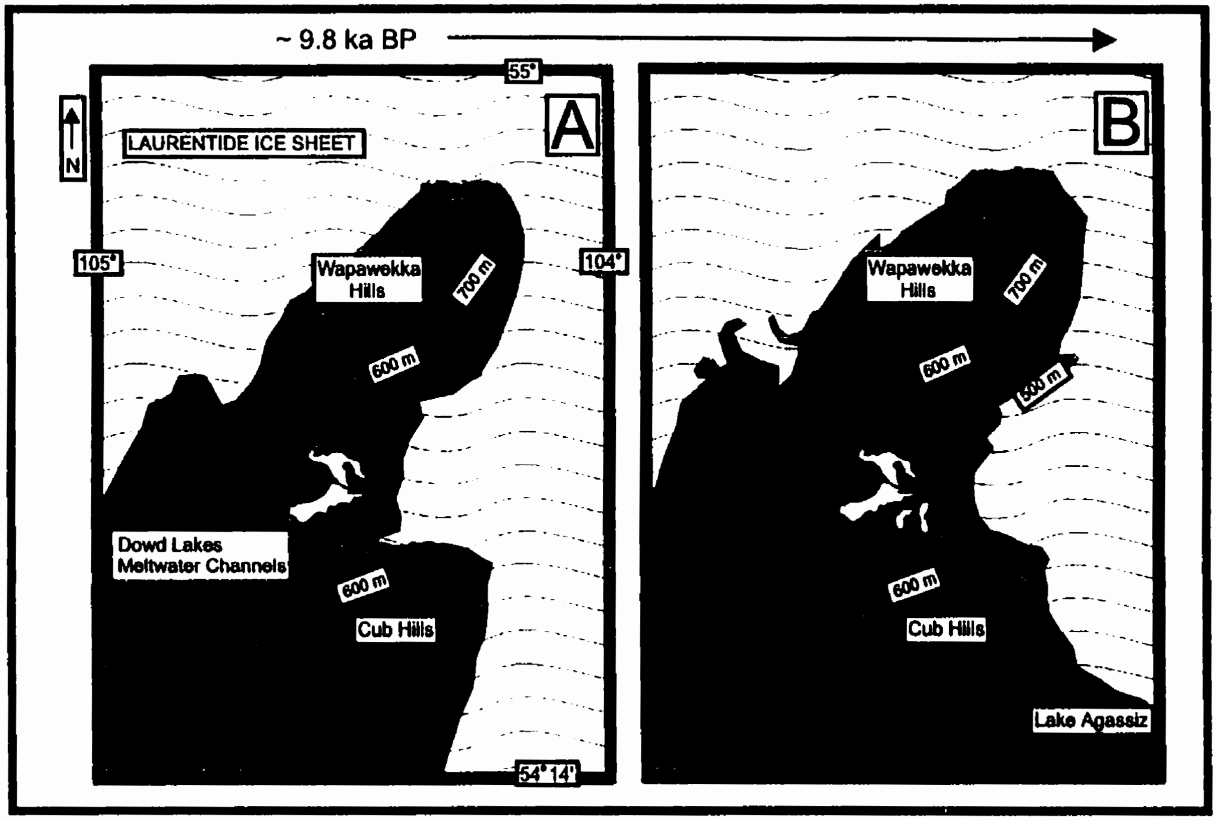


Figure 6-8 A-E: Speculative sequence of deglaciation in the Jay Jay Lake/Wapawekka Hills area, after the ~ 10 ka BP ice readvance. Events are summarized in the text. (after Langford, 1973)

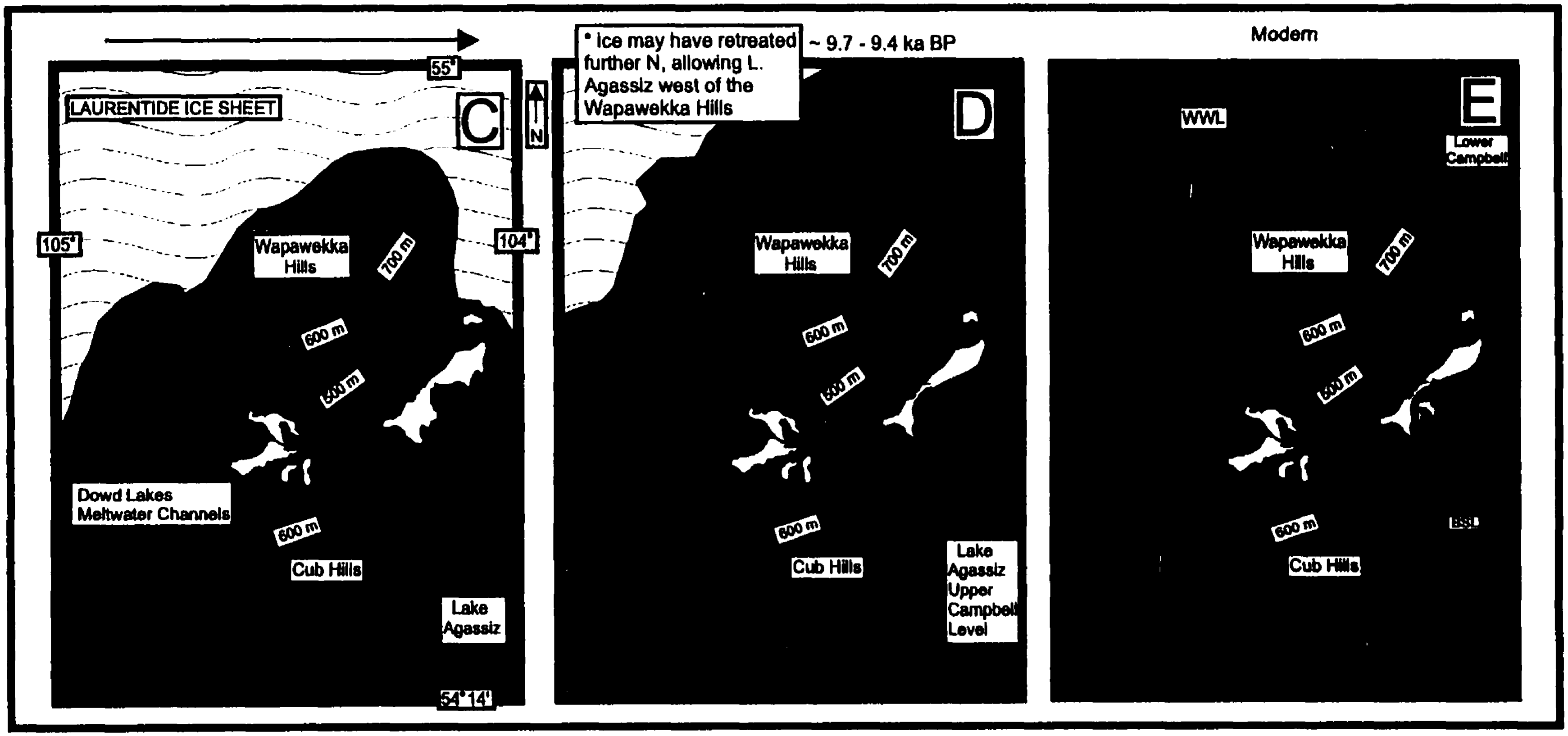


Figure 6-8 A -E continued: Speculative sequence of deglaciation in the Jay Jay Lake/Wapawekka Hills area, after the ~ 10 ka BP ice readvance. If the singular strandline west of the Wapawekka Hills is the continuation of the Upper Campbell level from the main Lake Agassiz basin to the east, then the ice abutted against the Wapawekka Hills shown in Fig. 6-8 E would have retreated from the Hills by 9.7 ka BP. Events are summarized in the text. (after Langford, 1973)

At the terminus of each of the Dowd Lakes channels, lobate outwash plains formed in this perched lake (Fig. 6-8 A). At the southern edge of the ice lobe, against the Cub Hills, an ice marginal channel carried drainage from the small perched lake into the subglacial hydrologic system. The meltwater channels on the Cub Hills by this time were probably inoperative.

Continued ice thinning reduced the size of the lobe between the Cub and Wapawekka Hills (Fig. 6-8 B), thereby decreasing the level of the small lake. Small outwash lobes were deposited on the southern side of the lake, and the Dowd Lakes channels likely continued to contribute sediment and water as well. The lake was then drained to the subglacial hydrologic system by a second, lower elevation ice marginal channel against the Cub Hills (Fig. 6-8 B). Ice wasting east of the Cub Hills allowed Lake Agassiz to begin its northward expansion into the region, probably at a level higher than the Upper Campbell since the lake level was declining from the Norcross highstand.

The ice lobe that persisted between the Wapawekka and Cub Hills disintegrated -- probably aided by calving along the ice margin (Fig. 6-8 C)-- and Lake Agassiz expanded into the region, although it never reached the level of the previous ice-marginal lakes associated with the Dowd channels. During this stage, the low-lying area that today contains Jay Jay Lake was a bay of Lake Agassiz. Lacustrine sediments mapped by Langford (1973) that surround and may underlie Jay Jay Lake are probably a reflection of Lake Agassiz deposition, although they could also be at least partially related to deposition in the small perched lake which had also previously occupied at least part of this site. Outwash plains at the margin of the Lake Agassiz basin were deposited and, based on the observation that the Upper Campbell strandline was cut into these outwash plains (Fig. 6-9) and that the outwash deposits have lobed deltaic margins (Langford, 1973), these outwash plains are tentatively interpreted to have been deposited into Lake Agassiz. If this was the case, the initial level of water inundation of the Wapawekka/Cub Hills region by Lake Agassiz may have been at a level slightly higher (perhaps partially due to isostatic depression of the area by nearby ice) than the Upper Campbell level. However, as noted in Section 4.3.1, because these outwash plains occur at a break in slope, and in absence of detailed sedimentology, they may have been emplaced subaerially.

If the ice had retreated from the Wapawekka Hills by ~ 9.7 ka BP (Fig. 6-8 D), the well-developed "unknown strandline" (see Fig. 5-2 B, Chapter 5) to the west is the Upper Campbell in this area. This Upper Campbell scenario is supported by the GPS elevation data (Chapter 5). Therefore, the absence of a Campbell beach west of the Wapawekka Hills cannot be explained by the ice margin against the Wapawekka Hills during the Upper Campbell level. As discussed in Chapter 5, the presence of only one beach west of the Wapawekka Hills alternatively suggests the ice margin was at the Wapawekka Hills during the Upper Campbell level (as shown in Fig. 6-8 D), making this singular beach the Lower Campbell. Although this situation is unresolved, I am siding with the GPS data analysis (Chapter 5) and traditional isobases (Teller and Thorleifson, 1983) that suggest the "unknown strandline " correlates with the Upper Campbell. It is

Figure 6-9: Glaciofluvial and glaciolacustrine features east of the Wapawekka Hills. The dashed line shows the location of Highway 106. 'O' indicates outwash plain deposits, the Upper Campbell strandline is located by the small arrows. The Lower Campbell beach is marked by the 'H'. The north arrow is about 1.6 km long. (Langford, 1973, Plate 6)

proposed that possibilities for the absence of the Lower Campbell west of Lac la Ronge include: 1) the beach was not preserved; 2) the Lower Campbell beach formed as a diffuse, discontinuous assemblage of strandlines in a shallow water archipelago (see Chapter 5), or 3) there is a need for more detailed fieldwork. Thus, if Gregory Lake is considered to be located behind the Upper Campbell strandline, deposition of the silty clay at the base of the GLE core (325 - 573 cm; see Section 4.4.1, Fig. 4-16 d, Table 4-G, Fig. 6-12) occurred during the initial expansion of Lake Agassiz into the northwestern arm of the basin (at ~9.7 ka BP), while the lake was at a level higher than the Upper Campbell (as suggested by the relationship of the Upper Campbell strandline and outwash plains on the southern margin of the Wapawekka Hills; Fig. 6-9).

HUDSON BAY AREA

Farther south in the Lake Agassiz basin, ice which had retreated from the western margin of the Lake Agassiz basin at the Manitoba Escarpment would have allowed the expansion of lake Agassiz into locations that had been occupied with ice since the inception of the Emerson Phase (~ 10 ka BP). The very fine grained sandy silt and organic material, found in the 95 - 98 cm interval of the RLA core (Table 4-D) may reflect declining lake levels, approaching the Upper Campbell level.

By the Emerson Phase reconstruction of Thorleifson (1996), earlier reconstructions and strandline mapping by Teller and Thorleifson (1983) and Johnston (1946), and the conclusions of Moran (1969), the highest *stable* level of Lake Agassiz in the Hudson Bay area was the Upper Campbell. Based on the observation in the

Wapawekka Hills area (Fig. 6-3 A for location) of several sizeable outwash plains north of Jay Jay Lake -- deposited on the margin of the Lake Agassiz basin and into which the Upper Campbell beach is formed (Langford, 1973; Fig. 6-9) -- it appears that during ice retreat after the initial Emerson highstand (the Norcross level), Lake Agassiz expanded into these areas along the western side of the basin at a level above the Upper Campbell.

THE BROKENPIPE LAKE AREA

As with the Norcross level, in the Dauphin area (Fig. 6-10 A - D), as Lake Agassiz levels declined from the Norcross to the Upper Campbell level after ~ 9.9 ka BP (Fig. 6-10 A), the stiff clay with dropstones in the 299 - 335 cm interval of the BPC core (Section 4.1.2, Photo 4-4 g, Table 4-A, Chapter 4) was deposited (if not during the earlier Lockhart Phase; see also Figs 6-12 & 6-13).

6.2.7 - Period 6: Figures 6-3 E, 6-8 D, 6-10 B, 6-12 & 6-13 (~9.7 - 9.4 ka BP)
Rapid ice retreat from the Emerson Phase maximum (Fig. 6-3 D) to the position shown in Figure 6-3 E probably occurred as early as about 9.7 ka BP. During this time, water flowing into Lake Agassiz from the North and South Saskatchewan rivers deposited the Nipawin Delta at Nipawin, Saskatchewan (Christiansen, *et al.*, 1995). In north-central Saskatchewan, in the La Ronge area, a lobe of ice may have filled the low area between the Wapawekka and Thunder Hills, precluding the inundation of the Churchill River valley by Lake Agassiz waters at this time. Lake Agassiz had likely stabilized at the

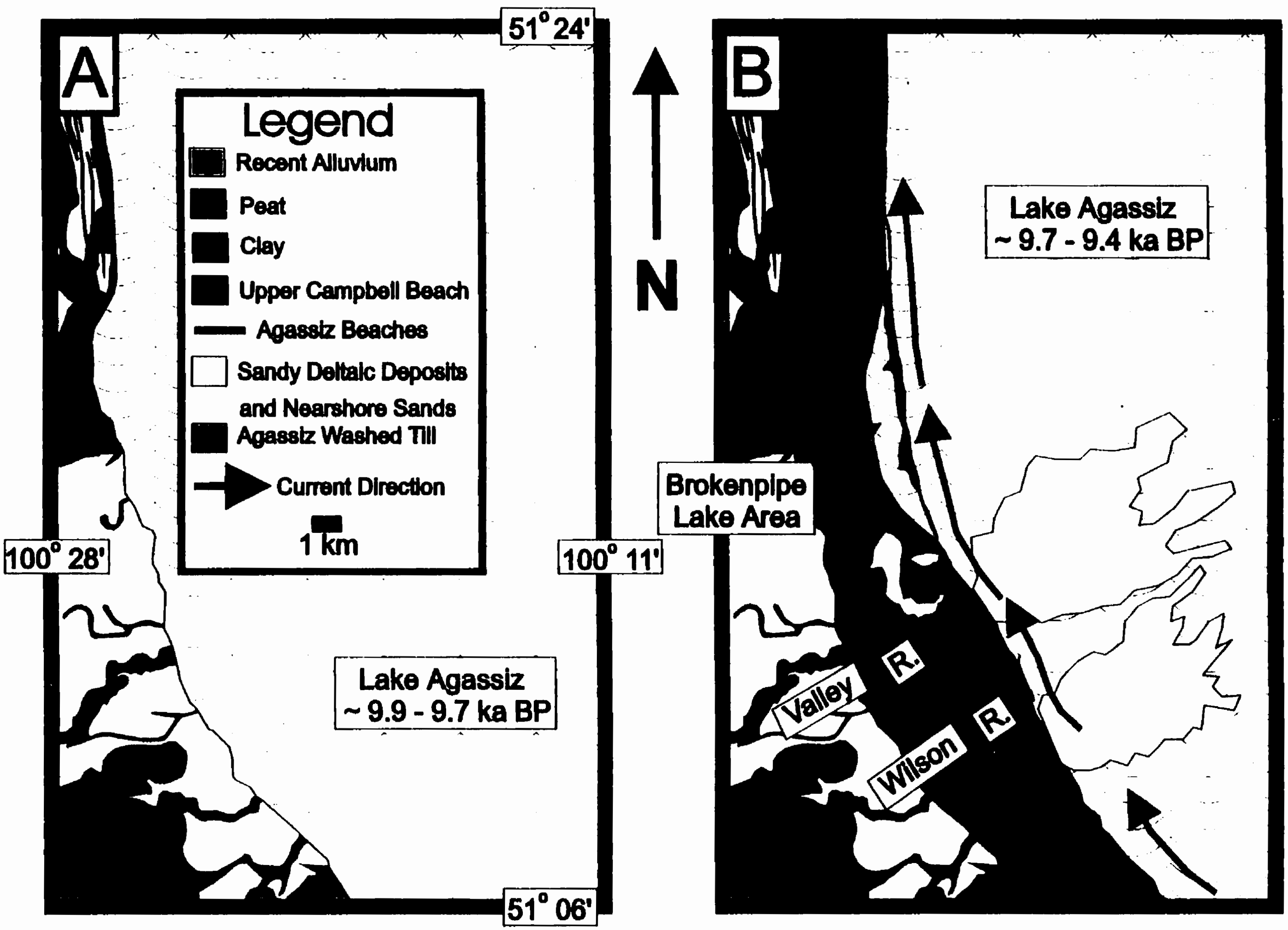


Figure 6-10 (A-D): Evolution of the Brokenpipe Lake Site During the Emerson Phase of Lake Agassiz. See text for explanation. (modified from Ehrlich *et al.*, 1959; Klassen, 1979)

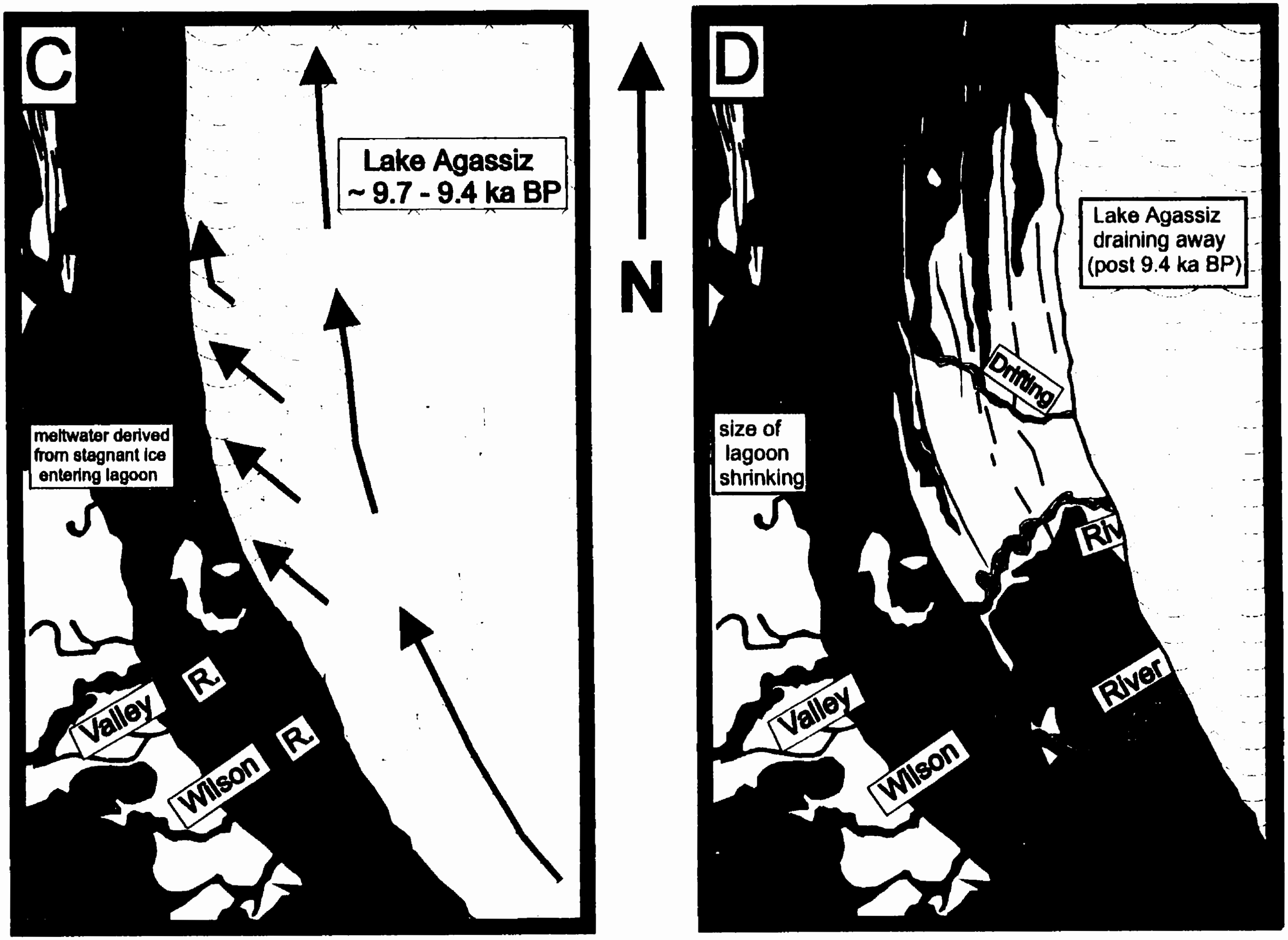


Figure 6-10 (A-D) continued: Evolution of the Brokenpipe Lake Site During the Emerson Phase of Lake Agassiz. See text for explanation. (modified from Ehrlich *et al.*, 1959; Klassen, 1979)

Upper Campbell level by about 9.5 ka BP.

THE GREGORY LAKE SITE

If Gregory Lake is located behind the Upper Campbell strandline (see Chapter 5), ice would have had to retreat from the Wapawekka Hills by ~ 9.7 ka BP (to the alternative --dashed -- ice margin position suggested in Fig. 6-3 E), and deposition of the variable sequence of poorly bedded and poorly laminated sand and silty sand near the base of the Gregory Lake sequence (198 - 480 cm; see Figs. 4-16 a & b, Table 4-G, Section 4.4.1) occurred during the stable period of Lake Agassiz at the Upper Campbell level (Fig 6-12). Therefore, the northwestern terminus of correlatable, Lake Agassiz Upper Campbell strandline occurs at Gregory Lake. The strandline in the Gregory Lake area is erosional, and from this location to the northwest is was diffused among a shallow water archipelago of ground moraine. In fact, given the nearly equal elevation of Gregory Lake versus the strandline to the west and the absence of a topographic barrier to form a closed basin (Fig. 4-15, Chapter 4) at the Gregory Lake site, it is highly probable that Lake Agassiz inundated Gregory Lake for the duration of the Upper Campbell level. Sand was reworked from the ground moraine of the region and was deposited in the relatively low-lying depression of Gregory Lake.

THE JAY JAY LAKE AREA

Regardless of whether the ice margin wasted to the north side of the Wapawekka Hills, trapping the Upper Campbell level to the east (Fig. 6-8 D) or, if the ice had retreated from the north side of the Wapawekka Hills by ~ 9.7 ka BP (i.e. Gregory Lake is behind the Upper Campbell strandline), in the Jay Jay Lake area, Lake Agassiz longshore currents constructed a large baymouth bar at the Upper Campbell level, which separated the small bay between the Wapawekka and Cub Hills (today within which Jay Jay Lake is located) from the main body of Lake Agassiz (Fig. 6-8 D). In the Swan River area, Nielsen *et al.* (1984) concluded that anti-clockwise circulation was probably prevalent in this fairly constricted bay of Lake Agassiz. Similarly, progradation of the beach at the Jay Jay Lake site was likely from the north. Outwash previously deposited by overland flow of deglacial meltwater from the highlands of the Wapawekka Hills and by the Bear River (located just north of Jay Jay Lake) served as the headland for erosion. Though formed after the Upper Campbell, the Lower Campbell beach, which extends around the east side of the Wapawekka Hills toward the Jay Jay Lake area as an arcuate, flared spit (Fig. 6-9), also suggests anti-clockwise circulation and southward progradation of this beach.

The stratigraphy of the JJC core (Table 4-E) and ostracodes found in the JJC core (Section 4.3.4, Chapter 4) do not suggest that the JJC core penetrated into the Lake Agassiz record at the Jay Jay Lake site, although the core does record sedimentation at the Jay Jay Lake site as a lagoon of Lake Agassiz. Sediments deposited in the lagoon consist of clayey silt and sand, frequently punctuated by relatively coarser materials (56-400 cm of the JJC core, see Table 4-E), that presumably record Lake Agassiz beach storm overwash at the Upper Campbell level (Figs. 6-12 & 6-13). The lower ostracode assemblage of *Heterocypris glaucus*, *Limnocythere herricki*, and *Candona rectangulata* (with *Cyclocypris ampla* and *Cypridopsis vidua*; Table 4-F) in the lower portion of the JJC core (56 - 400 cm; Section 4.3.4), suggest that the Jay Jay Lake lagoon was subject to lotic input, and was situated in a tundra-like setting which had negative effective moisture (E>P; the lagoon perhaps had TDS values between about 1 000 to 2 000 mg/l; today the TDS is an order of magnitude lower; Brandon Curry, 1998, personal communication).

THE HUDSON BAY AREA

In the Hudson Bay area, outwash from the surrounding uplands was likely a component of lacustrine deposition at this time (Moran, 1969). At Ruby Lake (Fig. 4-6, Section 4.2) very fine grained sandy silt with organic material (95 - 98 cm, core RLA; Table 4-D), likely records declining lake levels to the Upper Campbell, as does the sandy silt in the HBL core (145 - 168 cm; Fig. 4-8 a; Table 4-C; see Fig. 4-6 for location) The sandy silt at the Ruby Lake site grades upward into sand with gravel lenses (78 - 95 cm, core RLA; Table 4-D), perhaps an indication of the inception of the Upper Campbell beach -- a large constructional feature composed of sand and gravel (Fig. 4-7, Chapter 4) -- in the Hudson Bay area. Above this, a coarsening upward sequence of sand and gravel (12-56 cm in the RLA core; Table 4-D) probably represents subaqueous slumping or beach overwash into the lagoon during the Upper Campbell level (Fig. 4-8 c). Figures 6-12 and 6-13 summarize these events.

THE BROKENPIPE LAKE AREA

As Lake Agassiz water levels dropped from the Norcross to the Upper Campbell level in the Brokenpipe Lake area, continued outwash deposition from the Valley and

Wilson Rivers into Lake Agassiz formed distinct deposits at the mouths of these rivers. At the Upper Campbell level in the Brokenpipe region (Fig. 6-10 B), these sands and gravels were reworked by northward Lake Agassiz longshore currents, and as it formed, the Upper Campbell beach prograded northward past the Brokenpipe Lake site. The geomorphological observations of the Upper Campbell strandline (Section 4.1.1) supports this conclusion. The outwash deposited into a standing body of water (Lake Agassiz) became the headland for longshore transport and spit construction. The progradation at the Brokenpipe site occurred as a subaqueous spit-platform at about 346 m elevation, concurrently building multiple subaqueous spits in the lower-lying, back-beach area at the Brokenpipe Lake site (Section 4.1.1).

The poorly sorted gravel of the 295-299 cm interval of the BPC core (Fig. 4-4 f; Table 4-A), with sub-rounded to rounded carbonate clasts (the largest with a long axis of 7 cm), in a mm-sized granule to fine sand-sized carbonate and quartz matrix, and the massive gravel with large well-rounded clasts at the base of the BPN core (Table 4-B; Fig. 4-4 i) relate to the incipient progradation of the Upper Campbell beach past the Brokenpipe Lake site (Fig. 6-10 B). The well-rounded nature and large size of many of the gravel clasts, which are marked with percussion cones, supports the idea that the gravel had a fluvial origin (Klein, 1963, after Nielsen *et al.*, 1984) from the Valley and Wilson Rivers, then was transported and reworked by longshore Lake Agassiz currents several kilometers northward to be incorporated into the Upper Campbell beach at the Brokenpipe site.

Brokenpipe Lake today is in a low-lying, wetland area, and is in fact a remnant of a much larger lagoon that formed relict shorelines in till on the western side of the wetland area at about 347, 352, and 355 m elevation (Fig. 4-2). These relict shorelines mark pauses in the Lake Agassiz water level as it declined from the Norcross to the stable Upper Campbell level. Once the growth of the Upper Campbell beach spit platform had completed the main cut off of the large lagoon from the main body of Lake Agassiz, and the main basin of the initial lagoon (Fig. 6-10 C) at the Brokenpipe site was between about 7.5 and 8 m deep. At the Upper Campbell level, water depth in the lagoon was computed by calculating the elevation difference from the 347 m relict shoreline elevation -- which corresponds with the elevation of the Upper Campbell spit at the Brokenpipe site -- to the top of the Upper Campbell gravel in the 295-299 cm interval of the BPC core; Table 4-A).

The massive to well-laminated muds of the 196-295 cm interval of the BPC core (Table 4-A; Figs. 4-4 c to e), were probably deposited in this relatively deep lagoonal environment, separated from direct connection with Lake Agassiz by the Upper Campbell beach (see also Figs. 6-12 & 6-13). The fine laminae found in some portions of this interval of the BPC core (e.g. Fig. 4-4 c) represent suspension sedimentation in quiet water. Lagoon circulation was dominated by wind-driven nearshore waves and surface currents. The more massive zones of clayey-silt or silty-clay and the clean sand beds (at 262-265 cm and 285-289 cm; Figs. 4-4 d & f, Table 4-A) imply increased sedimentation rates, either by a periodic connection with Lake Agassiz or by flooding of overland runoff.

The silty gravels in the 130-181 cm interval of the BPN core (Table 4-B, Fig. 4-4 h) also were deposited during the initial highwater lagoonal period. They are correlatives with the clayey silts and silty clays deposited in the 196 - 295 cm zone of the BPC core (Fig. 6-11). This is suggested by their relatively smaller clast size (than the gravels below in the 181 - 210 cm zone of the BPN core; Table 4-B), and the silt content of the matrix. During silt deposition of the highwater lagoonal period, which deposited the massive and laminated clayey-silts and silty-clays found at the BPC core site, gravel slumped and/or was winnowed from the backside of the nearby Upper Campbell strandline, which then flowed into the lagoon as debris flows, and was deposited at the BPN coring site (see also Figs. 6-12 and 6-13).

During the Upper Campbell lagoon period, overland drainage was a major component to the hydrological budget of the lagoon. The peat-filled channels eroded into till on the west side of the Brokenpipe Lake site carried meltwater into the lagoon. Diverging shorelines of the lagoon at the northernmost channel (see right diagram in Fig. 4-2) indicate that the channels were in operation only during the earlier stages of the lagoon.

As previously noted (see Section 6.1.3, Figs. 6-1 A and 6-1 B), the ostracodes found in Biozone A (196 - 295 cm) of the BPC core are indicative of several key things:
1) The presence of *Cytherissa lacustris* implies the lagoon was generally cold, with dilute waters of moderate depth (>3m).

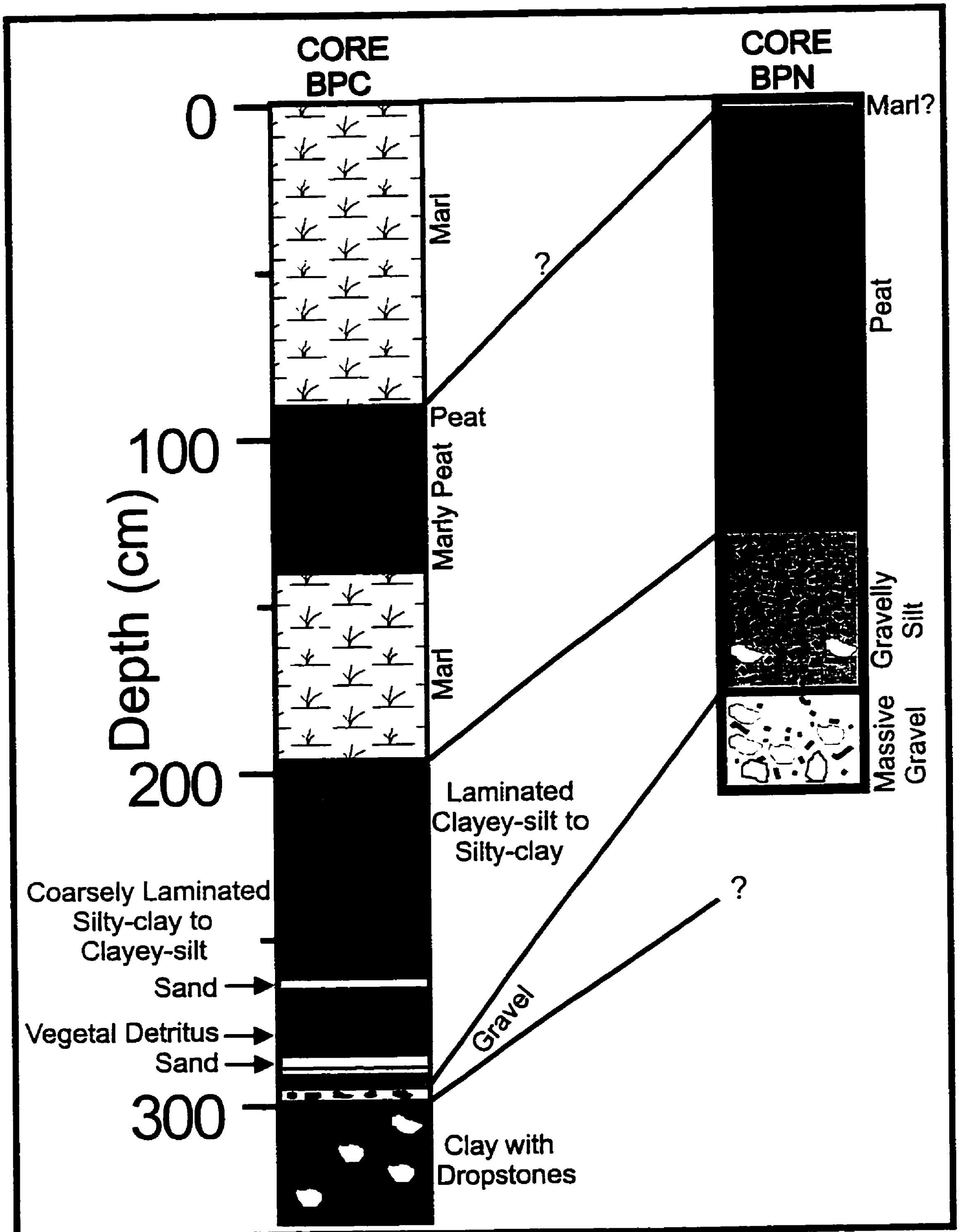


Figure 6-11: Stratigraphic correlation of the BPC and BPN cores. See Figs. 2-6 and 4-2 for core locations, and Tables 4-A and 4-B for core descriptions.

2) The presence of *Ilyocypris bradyi* and *Ilyocypris gibba* suggest flowing water. I propose that the currents entered the lagoon from the shallow channels eroded into till on the western side of Brokenpipe Lake. The source of the water was likely stagnant ice on the Riding Mountain uplands. *Cyclocypris ampla* and *Cypridopsis vidua* also tolerate flowing water, but they also may have inhabited the warmer, more vegetated margins of the lagoon, and were washed lakeward to be deposited at the BPC core site.

3) Other ostracodes such as *Candona rawsoni, Limnocythere herricki, Candona inopinata, Candona sigmoides, Cypridopsis vidua, Limnocythere itasca,* and *Candona ohioensis* suggest that the lagoon during this period was subject to variation in temperature and salinity. This variability was likely predominantly a response of the lagoon to variable inputs of runoff and groundwater, rather than changes in climate such as evaporation or temperature (though these probably were factors as well). Any direct signature or connection of Lake Agassiz with the lagoon, in terms of the ostracodes found in Biozone A, was masked by this interaction of surface and groundwater input -- probably because the influx of Lake Agassiz water was of short duration and of a relatively small volume.

The reduction in the number of ostracodes occurring in the upper portion of Biozone A to almost barren conditions (e.g. samples taken from 196 cm to 230 cm; Appendix D) was probably the result of dissolution. The few ostracodes found in this interval were commonly perforated by dissolution holes, and generally had dissolution-thinned carapaces. There are two possibilities for the destruction (dissolution) of these ostracode carapaces, 1) methanogenesis during the diagenesis of organic sediment, (in conjunction with a reduced sedimentation rate) was occurring -- the oxidation of methane would increase the amount of dissolved CO_2 in the water, thereby reducing the pH of the water at the bottom of the lake, enhancing dissolution of recently deposited ostracode shells, or 2) the lagoon was undersaturated with respect to calcite, and a reduction in the sediment accumulation rate would have accelerated ostracode valve dissolution.

6.2.8 - Period 7: Figure 6-3 F, 6-8 E, 6-10 D, 6-12 & 6-13 (~9.4 ka BP to Present)

Ice retreat in the main Lake Agassiz basin opened the eastern Kaiask outlet late in the Emerson Phase, lowering lake levels from the Upper Campbell level to below the Lower Campbell level (Thorleifson, 1996). Closure of that outlet by ice advance to the Sioux Lookout moraine (Thorleifson, 1996) near the end of the Emerson allowed Lake Agassiz to reoccupy the southern outlet, and rise in level to the Lower Campbell (see Section 1.2.5).

THE GREGORY LAKE AREA

In the Gregory Lake area, if ice had not retreated from the north side of the Wapawekka Hills by the end of the Upper Campbell level -- the configuration shown in Fig. 6-3 E (~ 9.4 ka BP) --, the well-developed, singular strandline that continues south and west of Lac la Ronge (the "unknown strandline", see Fig. 5-2B, Chapter 5) must be considered to be a continuation of the Lower Campbell strandline. In this scenario, the final retreat of ice from the Wapawekka/Thunder Hills area at about 9.4 ka BP allowed the northwestward expansion of Lake Agassiz into the Gregory Lake area. Lake Agassiz stabilized at the Lower Campbell level shortly after 9.4 ka BP, and probably only maintained this level until about 9.3 ka BP (Matile, 1998, personal communication).

The silty clay at the base of the GLE core (480 - 573 cm, Table 4-G; Figs. 4-16 d & 6-13) was deposited during the initial expansion of Lake Agassiz into the Gregory Lake area. Poorly bedded and laminated sand and silty sand (198-480 cm Table 4-G, see also Fig. 6-13) of the GLE core was deposited by Lake Agassiz currents eroding the predominantly sandy till ground moraine, in a shallow water, archipelago setting.

Regardless of the strandline correlation west of the Wapawekka Hills, the sandy peat of the GLE core (183-198 cm , Table 4-G) represents a transitional period between predominantly sand deposition during the higher energy initial beach-forming stage and subsequent isolation from Lake Agassiz due to declining water levels. The very gradational boundaries of this unit suggest a gradual transition, and continuous deposition at the site during the change from sand to organic deposition. There is a bulk organic AMS radiocarbon date of 8850 +/- 80 yr BP [T0-6893] from the base of this sandy organic unit (181 - 183 cm, Table 4-G).

The modern levels of Gregory Lake and Bird Lake are maintained by a series of 3 beaver dams at the southeastward draining outlet of Bird Lake. This "natural" control may have played a role in maintaining lake levels high enough for sedimentation and infilling by organic material (peat). Otherwise, nearly complete drainage of inter-till depressions like this after Lake Agassiz retreated from the Campbell level may have precluded or seriously hampered the formation of the series of swamps and small lakes (of which Bird and Gregory Lakes are a part) by producing periodically dry conditions.

THE JAY JAY LAKE AREA

Decline of Lake Agassiz water levels from the Upper Campbell would have totally isolated the Jay Jay lagoon from Lake Agassiz, forming an independant lake. Sediment deposition at the Jay Jay lake site changed from predominantly minerogenic sand and silt to organics (see Figs. 6-12 and 6-13). Specifically, the depth interval from 50 to 60 cm in the JJC core crosses the boundary (56 cm, Table 4-F) between predominantly organic sediments (peat and gyttja) and predominantly minerogenic sediments (clays, silts, and sands). In addition to finding all of the ostracodes as in the 0-50 cm interval, the 60 cm subsample also contained *Ilyocypris gibba*, an ostracode that is particularly indicative of lotic conditions (see Section 4.1.5). Deposition of the silts and clays in the lagoon was likely from suspension, the materials ultimately derived from overland runoff and channelized stream flow over the relatively recently deglaciated Wapawekka and Cub Hills, and the highlands to the west of the Jay Jay Lake site. *Ilyocypris gibba* would have lived in these flowing water environments, and would have been washed into the lagoon. The change in deposition from the clayey silt and sand of the 56-75 cm interval of the JJC core (Table 4-E) to organics, (gyttja and peat, Table 4-E) and the disappearance of *Ilyocypris gibba* (Table 4-F) suggests that the lotic contribution to the lagoon hydrology ebbed, leading to stagnation, increased TDS, lowering water levels, and gyttja/peat deposition.

Subsamples from the upper part of the core (0 - 50 cm, Table 4-F) included *Cyclocypris ampla, Cypridopsis vidua, Candona acutula,* and *Candona ohioensis.* The assemblage implies the lake was shallow, variable in temperature and salinity, and received shallow groundwater. The deposition of the gyttja-marl (10-56 cm, Table 4-E) was followed by lowering water levels, which culminated in the inception of vegetation encroachment and peat deposition (0-10 cm, Table 4-E) at the Jay Jay Lake site.

THE HUDSON BAY AREA

In the Hudson Bay area, the Ruby Lake RLA core site recorded the decline of Lake Agassiz level from the Upper Campbell to the Lower Campbell as a unit of gravel with peat beds (1-12 cm, Table 4-D, see also Figs. 6-12 and 6-13). The peat began to accumulate on the bottom of the very shallow lake, and periodic slumping of the beach into Ruby Lake may have occurred. As time progressed, sandy peat was deposited above this unit, representing sedimentation in an established Ruby Lake.

The drop in Lake Agassiz from the Upper to the Lower Campbell level was recorded quite differently at the HBL site, which unlike Ruby Lake, is located behind the Lower Campbell beach (Fig. 4-6). Poorly laminated sandy silt in the HBL core (145 - 168 cm) was deposited during the Upper Campbell level at this site, and during the decline in lake level from the Upper Campbell. The overlying gravel (110 - 145 cm in the HBL core) noted the inception of the Lower Campbell beach at this site (Table 4-C; Fig. 4-8 a, see also Figs. 6-12 and 6-13). Deposition of peaty sand in the newly formed

Lower Campbell lagoon followed (86 - 110 cm in the HBL core; Table 4-C). After Lake Agassiz declined from the Lower Campbell at about 9.3 ka BP (Matile, 1998, personal communication), vegetation encroachment and peat deposition ensued (0 - 86 cm of the HBL core, Table 4-C).

THE BROKENPIPE LAKE SITE

Water levels in the Brokenpipe lagoon dropped in concert with the decline of Lake Agassiz water levels from the Upper Campbell to the Lower Campbell (Fig. 6-10 D). This event ended the deep lagoonal period at the Brokenpipe Lake site, and Brokenpipe Lake became an independant waterbody. More importantly, as Lake Agassiz water levels dropped well below the Upper Campbell, the strandline was perhaps breached at the location where the Drifting River today crosses the strandline (Fig. 6-10 D), thereby draining most of the water from the Brokenpipe area. Northward drainage of overflow from the lagoon was initiated through the breach in the Upper Campbell strandline at the Drifting River, which is observed today. Through time, differential isostatic rebound has differentially uplifted areas to the north of the Brokenpipe Lake site, slowing this northward drainage by reducing the gradient and causing drowning of the tributary at the southern end of Brokenpipe Lake (see Section 4.1.1). The water level of modern Brokenpipe Lake and its surrounding marsh is likely maintained by groundwater hydrology. The history of this relatively shallower stage of the Brokenpipe Lagoon (initiated by the decline of Lake Agassiz from the Upper to the Lower Campbell), to the modern day, is reflected within three zones of the upper portion of the BPC core (see

Figs. 6-12 and 6-13).

The shift from deposition of primarily mineralogenic clays and silts in the lagoonal environment to a biogenically induced calcareous marl (140 - 196 cm in core BPC; Table 4-A) implies a dramatic change in environmental parameters that governed this lake. Lakes that accumulate marls are characteristically oligotrophic; their organic productivity is low (Cole, 1983). Low organic productivity is a result of inhibited availability of nutrients under the highly buffered alkaline carbonate-bicarbonate system, and not because of a low input of nutrients (Vreeken, 1981). The buffering capacity of the system is usually supplied by the $Ca(HCO_3)_2$ that was ultimately derived from a carbonate reservoir within the lake's watershed (Vreeken, 1981), though groundwater may also play a role.

Marl production in Brokenpipe Lake was the result of CO_2 loss related to the needs of specialized hard-water plants; CO_2 was taken out of the water by the *Chara sp.* to meet photosynthetic demands. As the CO_2 is withdrawn from the water, the dissolved $Ca(HCO_3)_2$ decomposes into CO_2, H_2O and precipitates $CaCO_3$. The study of Vreeken (1981) noted that the presence of marl producing lakes is not necessarily related directly to the presence of a calcareous-rich bedrock substrate; however marl sites tend to be linked to carbonate-rich till surroundings, such as at the Brokenpipe Lake site.

Hydrologically, the most important factor related to the presence and continued deposition of marl is that the site must be perenially wet, in order to ensure the survival of the marl-producing aquatic vegetation (Vreeken, 1981). Commonly, marl sites are therefore lined by an impervious clay substrate, although it is important that the site is not completely impervious to lateral subsurface groundwater flow (Vreeken, 1981).

Average marl accumulation rates in the Vreeken (1981) study range from 1.5 - 11.0 cm/100 years. Vreeken (1981) concluded that a narrow range in accumulation rates may reflect a constancy in marl accumulation parameters (i.e. perennially wet conditions, continuous calcareous supply). The onset of marl production suggests two major factors; plant colonization on the landscape was rapid, and pedogenesis and carbonate weathering were strong (Vreeken, 1981). Thus, in the case of the Brokenpipe Lake site, the formation of the marl suggests two major changes: 1) colonization of plants in the Brokenpipe lagoon site which probably was facilitated by ameliorating climate (as Lake Agassiz drained from the area -- reducing lake effect -- and as the ice sheet retreated northward), and/or a decrease in water depth at the site -- due to northward drainage of the site after the Upper Campbell beach was breached at the Drifting River, and 2) a hydrological shift in the lagoon from dominance by early post-glacial runoff and a possible periodic connection with beach overwash of Lake Agassiz waters (during the deeper lagoonal period), to dominance by groundwater and run-off from an increasingly weathered carbonate till in the region surrounding Brokenpipe Lake during this shallower water second level of the lake.

The 90 to 140 cm interval of the BPC core (Fig. 4-4 a) consists mainly of peaty marl that grades upward into fine peat in the 90 - 93 cm zone (Table 4-A). Regional cessation of marl accumulation in southern Ontario was determined to be a reflection of

Holocene climatic changes, and corresponding hydrological changes (Vreeken, 1981). For example, during a post-glacial warm up period, hydrological regimes may have changed from perenially wet to seasonally wet types and back to perennially wet following the Hypsithermal warm period (Vreeeken, 1981). Young and Mansue (1979) also suggested the predominance of marl formation is related to wetter climatic conditions, though this observation may not be applicable to all marl-forming sites.

A progressively decreasing supply of carbonate rich waters may have also played a role in the cessation of marl production at the Brokenpipe Lake site, though decalcification of the landscape is a complex process, and includes time as a factor (Vreeken, 1981). Decalcification of shallow hydrological systems feeding marl producing sites would eventually deplete the carbonate reservoir needed for marl production, and initiate the change from marl production to peat deposition (Vreeken, 1981).

The gradual change from a marl to a peat from bottom to top in this unit (90 - 140 cm in the BPC core) could be related to a shift towards less precipitation and more evaporation (i.e. less effective moisture), resulting in decreased water table levels and groundwater inflow to the site, and a corresponding water level drop. The hydrological regime of the Brokenpipe lagoon site could have shifted from one of perennial wetness to periodic dryness, thereby decreasing overland runoff and lowering piezometric groundwater levels, which could result in a reduced supply of carbonate-rich waters to the lake. As the depth of open water continually decreased with sedimentation infilling the basin and water level decreases, the encroachment of emergent macrophytes (such as cattails) probably occurred. Production of vegetation increased, and peaty sediments began to accumulate.

In consideration of the 5980 +/- 70 yr BP [TO-6205] wood age from 90 cm in the upper peat of this BPC core peat unit, the gradual shift from marl production to peat production could be explained by a mid-Holocene warming, such as the Hypsithermal. Specifically, a reduction in effective moisture (i.e. more evaporation and less precipitation), independent of temperature, perhaps best explains the gradual shift from marl to peat production in Brokenpipe Lake. Pollen records from the Brokenpipe Lake area (Ritchie, 1964, 1969, 1983), and sedimentological records of nearby Lake Manitoba (Teller and Last, 1979), indicate a distinct warming, which reaches a peak around 6000 yr BP. Along with this are reduced precipitation values, from 10 000 yr BP, up to about 6500 yr BP (Ritchie, 1983). The pollen records of lakes (e.g. Ritchie, 1964, 1969) well above the highest level of Lake Agassiz on the Riding Mountain uplands, including one situated just outside of the Wilson River watershed (see Figure 6-10 D for location of the Wilson River to the Brokenpipe site), were an integral part of the dataset which led Ritchie (1983) to produce these inferences about the paleoclimate of the central Lake Agassiz basin. Furthermore, regional data supports the interpretation that the deposition of this peat was a result of a relatively drier climate. Proxy evidence from lakes outside of the Lake Agassiz basin, within the northern Great Plains region (e.g. Xia *et al.*, 1997; Last and Sauchyn, 1993; Last, 1990; Vance *et al.*, 1993) report mid-Holocene periods of relatively drier climates, causing high salinity and/or low lake stands that are temporally similar, or overlap the dated peat at the Brokenpipe Lake site. By this time, Lake Agassiz had completely drained and therefore would not have had an influence on regional climate as it may have had during periods of time when it was geographically extensive (Hu *et al.*, 1997). Ostracode shells within the peat are not well preserved. In fact, the barren zone in the peat at 90 cm in the BPC core probably represents a hiatus, or a period when the lake became a fen. Brokenpipe Lake likely reformed no later than about 3 500 yr BP, based on the reconstruction from Lake Manitoba that suggests increasing regional water levels at that time (Curry, 1997).

The 90 - 0 cm interval of the BPC core (Table 4-A) is characterized by an upward gradation from peaty marl to marl (Fig. 4-4 a). The gradual shift from peat production back to marl production would occur in the opposite way as marl production changed to peat production in the 90-140 cm zone of the BPC core. A relative cooling of the climate after about 6000 yr BP probably was characterized by reduced evaporation and relatively increased precipitation; this increase in effective moisture would have certainly had an effect on Brokenpipe Lake. Water levels in Brokenpipe Lake would have risen, as increased groundwater discharge and overland flow increased carbonate leaching of watershed soils to produce carbonate-rich waters suitable for the production of the marl. The site would have increased in depth enough to support a *Chara sp.* dominated flora, and shifted back from perhaps being periodically dry, to being perenially wet.

Ostracodes of Biozone B (0 - 196 cm, Section 6.1.4) are all indicative of variable conditions of temperature and salinity, characteristic of hydrologically open shallow lakes and wetlands. There are neither ostracodes that indicate the lake was of any appreciable depth, nor any ostracodes that suggest there were substantial currents in the lagoon. Modern analog analysis by the use of weighted means (Fig. 6-2) suggests that the bottom water temperature and salinity of Brokenpipe Lake increased from the bottom of Biozone B to the deposition of the peat layer at 90 - 93 cm. The appearance of *Cypris pubera* implies that the lake was undergoing a senescence stage (Delorme, 1969), and increasing proportions of *Candona distincta* and *Candona decora* also suggest shallowing conditions, consistent with the ostracode-derived trends of increasing bottom water temperatures and salinity. Following the deposition of the peat, both the trends of the salinity and bottom water temperature curves of the lagoon reverse themselves. This reversal in the lower and upper parts of Biozone B support the conclusion that a warming climate peaked in dryness at about 6000 yr BP (resulting in deposition of the peat), and then deteriorated to conditions slightly wetter conditions, which are observed today.

The history of the Brokenpipe lagoon from the drop in water level at the end of the Upper Campbell to the modern day is represented by a larger interval (0- 130 cm) in the BPN core (Table 4-B). When water levels dropped with the retreat of Lake Agassiz from the Upper Campbell level -- initiating northward drainage of the Brokenpipe lagoonal area -- water depth was deep enough in the Brokenpipe Lake area (i.e. the BPC core site) to deposit the first marl (see above). However, the BPN core location is stratigraphically several meters higher, so water depth here was much less at this time.

Thus, rather than establishing of a marl-forming *Chara sp.* population (as at the BPC site), terrestrial vegetation encroached on the BPN site, resulting in peat growth. The peat of the 0 - 130 cm zone in the BPN core is correlative to the entire upper 196 cm of the BPC core (Fig. 6-11). The upper 10 cm of this peat, being lighter in color and marly, is a reflection of recently higher water levels, and perhaps the inception of recent marl production in this part of the marsh.

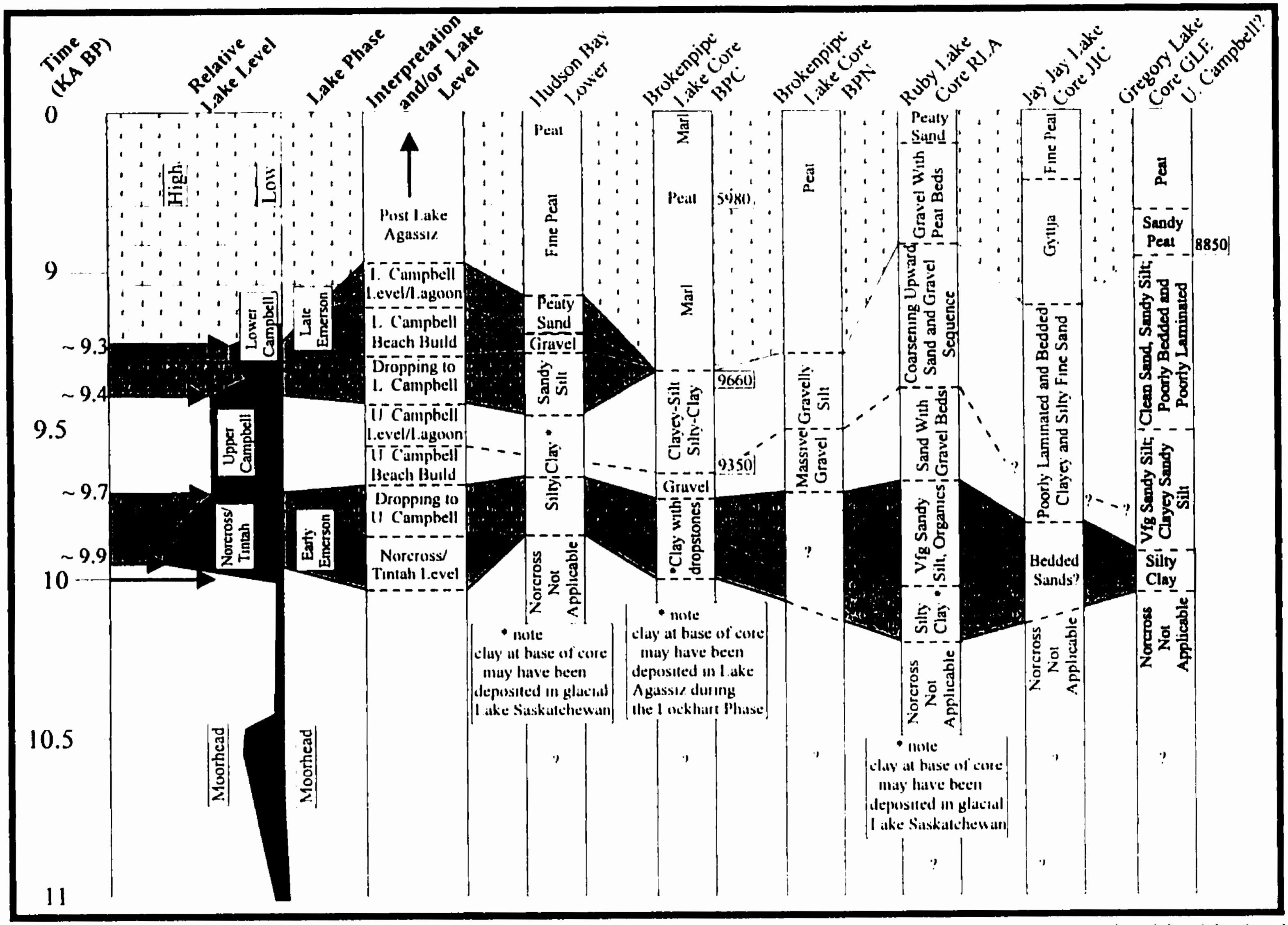


Figure 6-12: Speculative correlation of Lake Agassiz history and depositional events of the lagoons examined in this thesis. In this scenario, the Gregory Lake site is considered to be located behind the Upper Campbell strandline. See text for details.

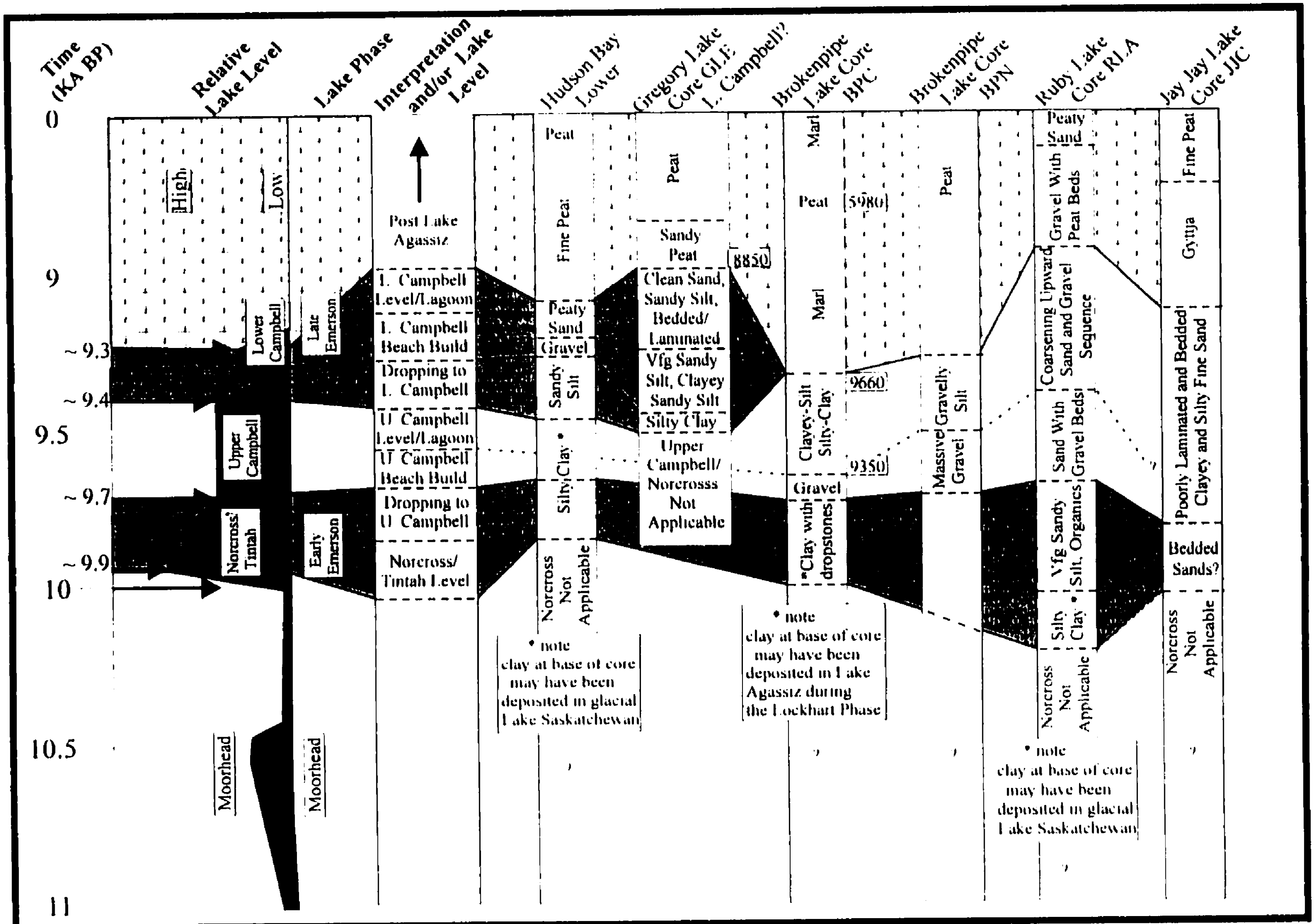


Figure 6-13: Speculative correlation of Lake Agassiz history and depositional events of the lagoons examined in this thesis. In this scenario, the Gregory Lake site is considered to be located behind the Lower Campbell strandline. See text for details.

CHAPTER 7 - RELATIONSHIP OF THE BROKENPIPE LAKE SITE TO OTHER UPPER CAMPBELL LAGOONAL SITES

INTRODUCTION

The sequence at the Brokenpipe Lake paleolagoon provides the most detailed stratigraphic sequence in the northwestern part of the Lake Agassiz basin studied in this thesis. It also includes three radiocarbon dates, and a detailed ostracode record. Because there are few such sites known in the Lake Agassiz basin, it is important to compare the Brokenpipe record with other such dated, Upper Campbell-related locations. Figure 7-1 shows the locations of all the sites that will be discussed below. A summary diagram (Fig. 7-2) shows how the Brokenpipe Lake stratigraphy correlates with the other Upper Campbell sites discussed below, and the relative level changes of Lake Agassiz. Taking site specific interpretations into account, a reasonable correlation can be made that illustrates that essentially the same lake level changes of Lake Agassiz are recorded at all the sites, although in each case, the lake level changes are demonstrated in different ways.

7.1 - Wampum West (Teller *et al.*, 1996; Teller *et al.*, submitted)

The Wampum West site lies behind the Upper Campbell beach at the edge of an island in Lake Agassiz that existed in the southeastern corner of Manitoba during the Emerson Phase (WW in Fig. 7-1). The stratigraphy of several cores and the paleoenvironmental reconstruction of this site based on diatoms document the evolution of the site, and its relationship to the changing levels of Lake Agassiz.

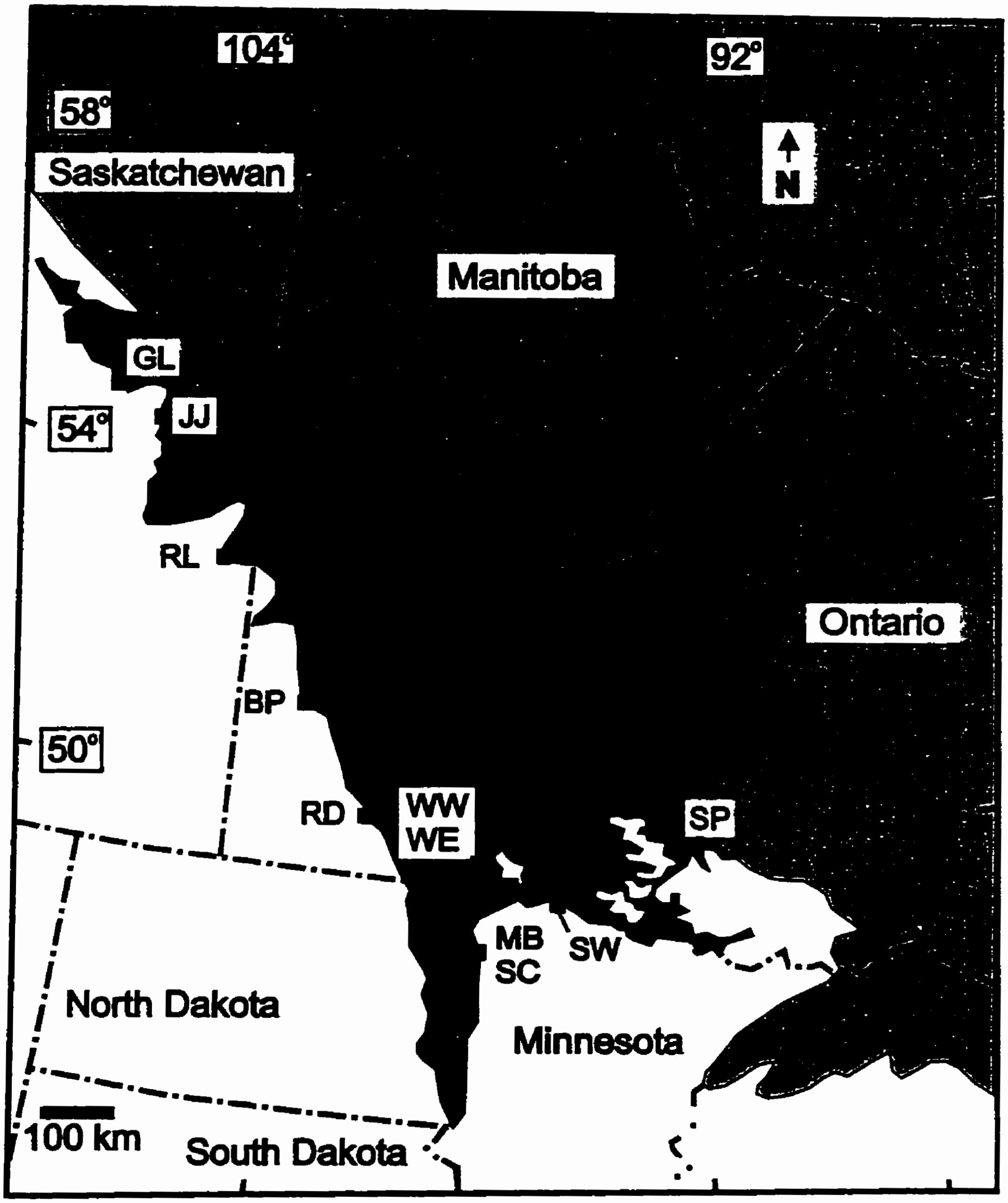


Figure 7-1: Locations of all Upper Campbell related sites discussed in the text. (GL = Gregory Lake, JJ = Jay Jay Lake, RL = Ruby Lake, BP = Brokenpipe Lake, RD = Rossendale, WW = Wampum West, WE = Wampum East, SP = Sioux Pond, SW = Swift, MB = Mosbeck, SC = Snake Curve) Upper Campbell LakeAgassiz extent (dark shading) and Laurentide Ice Sheet position (light shading) after Teller (1985, 1987).

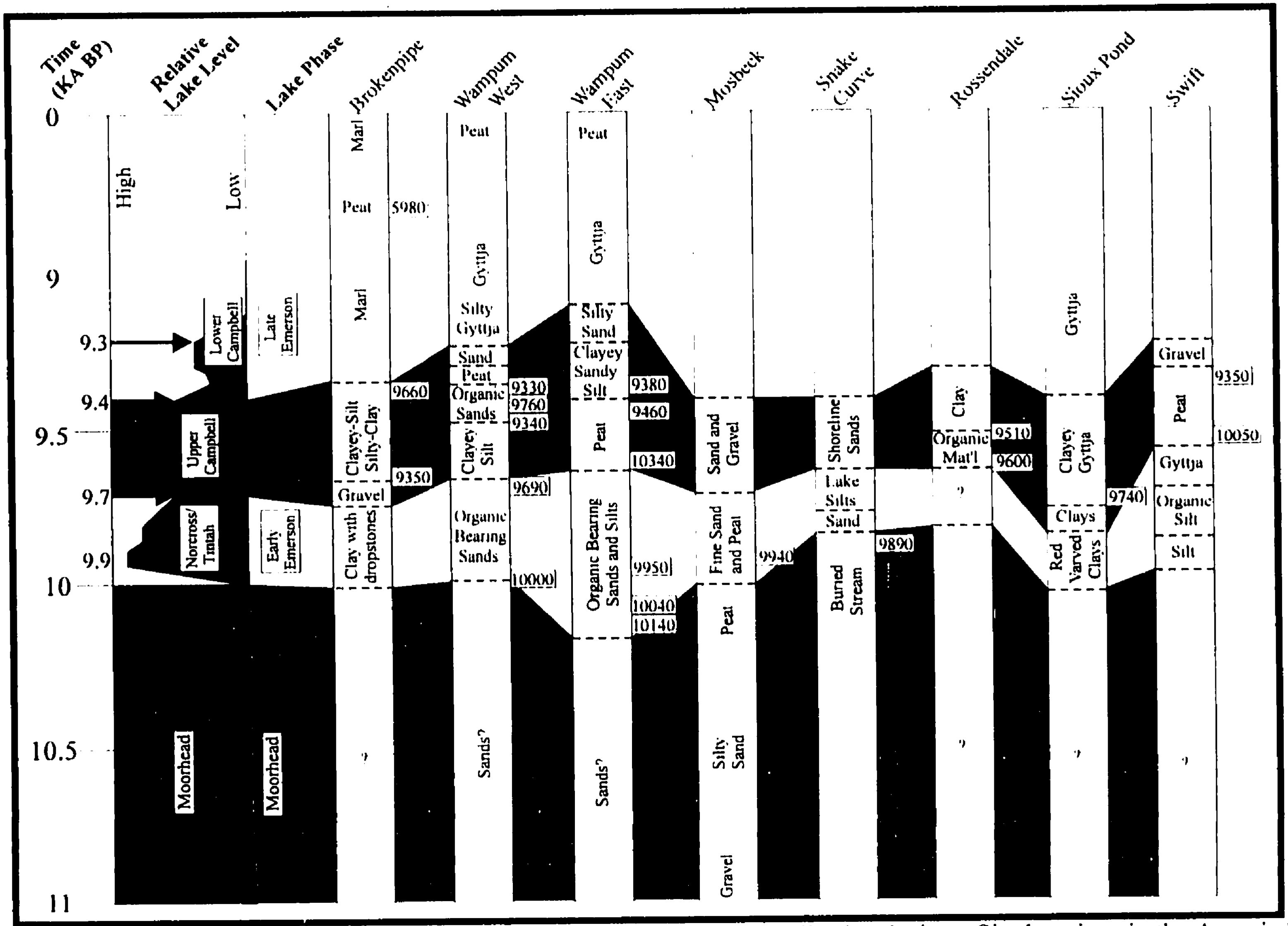


Figure 7-2: Summary diagram and correlation of events at several Upper Campbell-related sites. Site locations in the Agassiz basin are shown in Figure 7-1. Wampum sites after Teller *et al.* (1996); Risberg *et al.* (1995); Teller *et al.* (submitted), Swift and Sioux Pond sites after Bjorck and Keister (1983), Mosbeck Site after Ashworth *et al.* (1972), Snake Curve Section after Moran *et al.* (1971), and Rossendale Gully after Teller (1989). See text for details.

The cores studied in detail from this site generally have peat sandwiched between sands and shoreline sands which often contain appreciable organic material. This buried peat, which correlates to peat buried directly under sands of the Upper Campbell strandline, is dated at 9330 +/- 80 yr BP [TO-4856]. Two wood dates from organic bearing sands below this zone are 9690 +/- 70 yr BP [TO-4869] and 10 000 +/- 110 yr BP [TO 4286] (Fig. 7-2). The fact that the peat unit is sandwiched between shoreline sediments, and in consideration of the radiocarbon dates, Teller *et al.* (1996) concluded that the Upper Campbell beach was formed during a transgressive phase of Lake Agassiz, and was formed between 9700 and 9400 BP at this site.

Recent interpretations (Teller *et al.*, submitted) about this site suggest that clayey silts, which underlie the beach sands and buried peat, but are above the 9690 +/- 70 BP date, are representative of the Norcross level of Lake Agassiz in this area. The organic bearing sand (dated between 9690 +/- 70 and 10 000 +/- 110 yr BP) found in the lower section of cores (below the clayey silt) at this site is therefore representative of wave reworking during the post-Moorhead transgression of Lake Agassiz over the Wampum area at about 10 000 yr BP, culminating in the deposition of these clayey silts at about 9690 yr BP (Teller *et al.*, submitted). Therefore, the 9690 +/- 70 yr BP date from the organic bearing sands suggests that Lake Agassiz did not stabilize at the Upper Campbell level until about 9500 yr BP. Two more wood dates, 9340 +/- 90 yr BP [TO-4855] and 9760 +/- 80 yr BP [TO-4854] are also from the Wampum West site. Both are from directly below the sands of the Upper Campbell beach. The 9760 yr BP date is probably of reworked peat deposit, and is rejected. Thus, the 9330 yr BP and 9340 yr BP dates, from two different cores located directly below the sands of the Upper Campbell beach, in conjunction with the 9690 yr BP date in the stratigraphically lower organic bearing sands, suggest that this beach was constructed here during stable Upper Campbell water levels between ca. 9500 yr BP and 9300 yr BP. By about 9300 yr BP, Lake Agassiz water levels had probably fallen below the Upper Campbell at the Wampum West site, thereby abandoning it (Teller *et al.*, submitted).

The peak in salinity conditions of the lagoon, as suggested by diatoms, occurred between 7200 and 7000 yr BP; salinity also increased between 6000 and 4000 yr BP. This information suggests that climate warmed and dried during these periods, increasing evaporation of lagoonal waters and increasing salinity. This change in water chemistry may be related to mid-Holocene (Hypsithermal) climate warming. Terrestrialization of the Wampum West site occurred after 4000 yr BP.

7.1.1 - Ostracodes from Wampum West

Samples from the basal portion of the Wampum West C core, the core utilized in the detailed studies of Teller *et al.* (1996) and Teller *et al.*, (submitted), were analysed by me for fossil ostracodes. A total of 84 samples, at 2 cm intervals, were processed in the 721 - 895 cm interval of the core. Unfortunately, the samples available were quite small, thus the ostracode counts were very low, and this data could not be treated in a quantitative way. Table 7-A (after Matile and Risberg, 1996, personal communication) describes the stratigraphy of the lower portion of the WWC core, and Table 7-B lists the ostracodes recovered from the WWC core. In most cases, whole and fragmented mollusc shells were present in the samples, even in those that did not contain any ostracodes. The autecology of ostracodes found in the Wampum West C are similar to that of co-stratigraphic diatoms, as outlined by Teller *et al.* (1996), or Teller *et al.*, (submitted). The ostracodes found in the studied portion of the core do not show evidence for any direct connection with Lake Agassiz. They lived in a lagoonal environment that underwent fluctuations in temperature and salinity, although in general the waters remained fresh. The lone occurrence of *Cytherissa lacustris* at 747 cm -- the only ostracode in the assemblage that suggests relatively deep, dilute cold water -- with the other ostracodes, suggests that *Cytherissa lacustris* was living under conditions at the upper end of its salinity and environmental variability tolerance range.

The variability in temperature and salinity in the lagoon may have been due to periodic overwash of Lake Agassiz (as suggested by diatom paleoenvironment reconstruction by Teller *et al.* (1996) on the same strandline, east of this site), although this connection appears to be masked by conditions predominantly lagoonal in nature in the Wampum West ostracode record. The widely separated, lone occurrences of one *Cyclocypris ampla* at 805 cm, and one shell each of *Candona paraohioensis* at 859 cm and *Candona acutula* at 867 cm suggests that these particular ostracodes were deposited into the Wampum West lagoon from smaller depressions around the periphery of the Wampum West site, during the transgression of Lake Agassiz to the Norcross level.

Table 7-A: Generalized Stratigraphy of the Basal Section of the WWC Core

Depth (cm)	Description
703 - 724	Alga gyttja, contains shells, grey in color, basal contact gradational
724 - 737	Algal gyttja, contains shells, greyish-brown, lower boundary sharp
737 - 766	Silty gyttja, grey, contains shells
766 - 770	Medium grained sand, low organic content, lower boundary sharp
770 - 773	Peat, black, lower boundary relatively sharp
773 - 803	Fine to medium grained sand, organic-rich
803 - 810	Fine to medium grained sand with sparse coarse sand grains, lower boundary sharp
810 - 816	Clayey silt with sparse angular granules
816 - 834	Interbedded clayey silt and fine to coarse sand with rare pebbles
834 - 871	Fine to medium grained sand with rare coarse sand grains, rare shell fragments
871 - 895	Laminated silty clay (1-10 mm thick) and silt (<1-4 mm thick)

Description after Matile and Risberg (pers. comm.)

Table 7-B: Fossil Ostracodes Found in Core WWC

Depth (cm)	Ostracode Species (no. carapaces found)	Depth (cm)	Ostracode Species (no. carapaces found)
721	*Cyclocypris ampla* (2)	757	*Cyclocypris ampla* (2)
723	*Candona ohioensis* (1.5)	759	*Cyclocypris ampla* (5) *Cypridopsis vidua* (1) *Candona paraohioensis* (1.5)
725	*Candona ohioensis* ? (broken)	761	*Cyclocypris ampla* (7.5) *Cypridopsis vidua* (2.5) *Candona paraohioensis* (4)
727	*Candona ohioensis* (1)	763	*Cyclocypris ampla* (5) *Cypridopsis vidua* (1.5) *Candona paraohioensis* (3.5) *Candona decora* (1)
735	*Candona ohioensis* ? (broken)	765	*Candona distincta* (1.5) *Candona paraohioensis* (0.5)
737	*Candona ohioensis* (1.5)	767	*Candona paraohioensis* (1)
739	*Candona ohioensis* (2.5) *Cyclocypris ampla* (5) *Candona paraohioensis* (1.5)	769	*Cyclocypris ampla* (0.5)
741	*Candona ohioensis* (2) *Cyclocypris ampla* (5) *Limnocythere herricki* (3.5)	771	*Cypridopsis vidua* (1) *Candona punctata* (1.5)
743	*Cyclocypris ampla* (3.5) *Candona paraohioensis* (1.5)	773	*Cyclocypris ampla* (0.5) *Cypridopsis vidua* (1)
745	*Cyclocypris ampla* (1.5) *Candona paraohioensis* (1.5)	775	*Candona acutula* (1.5)
747	*Cytherissa lacustris* (0.5) *Cyclocypris ampla* (6) *Cypridopsis vidua* (1) *Candona paraohioensis* (0.5)	789	*Cyclocypris ampla* (4) *Cypridopsis vidua* (2) *Candona paraohioensis* (1)
749	*Cyclocypris ampla* (1.5) *Cypridopsis vidua* (2) *Candona paraohioensis* (1)	805	*Cyclocypris ampla* (1)
751	*Cyclocypris ampla* (7.5) *Cypridopsis vidua* (1)	859	*Candona paraohioensis* (0.5)
753	*Cyclocypris ampla* (3) *Cypridopsis vidua* (5) *Candona paraohioensis* (2.5)	867	*Candona acutula* (0.5)
755	*Cyclocypris ampla* (1)		

Note: ostracodes were sampled every 2 cm from 721-895 cm in core; gaps in table indicate barren zones

7.2 - Wampum East (Risberg *et al.*, 1995; Teller *et al.*, submitted)

The Wampum East site is about 2 km east of the Wampum West site, behind the same strandline (Fig. 7-1). A 9-cm-thick sequence of humified peat (dated 10 340 +/- 100 yr BP [TO-4285] at its base and 9460 +/- 90 yr BP [TO-4284] at its top), deposited between silty sand below and silty clay to clayey silt above, is key to interpreting the development of this site in relation to Lake Agassiz.

The older radiocarbon date poses a problem for making the traditional interpretation (e.g. Fenton *et al.*, 1983) that the transgression of Lake Agassiz to the Upper Campbell level occurred at 9900 yr BP. This traditional interpretation suggests that the peat was deposited behind a newly constructed Upper Campbell beach, during the Emerson Phase, and as such it should have a date of ~ 9.9 - 10.0 ka BP. Risberg *et al.*, (1995) reasoned that unless the basal peat date is contaminated by old detrital organics (e.g. Nambudiri *et al.*, 1980) or by old dissolved bicarbonate (e.g. Teller, 1989) it appears unlikely that the peat formed behind the Upper Campbell beach. If the basal peat date was closer to 9900 yr BP, the traditional scenario would be a feasible interpretation for the site. A subsequent rise in Agassiz level after 9460 yr BP would have drowned the Wampum East site, ending peat deposition and initiating deposition of the overlying silty clay and clayey silt.

To account for the older basal peat date, Risberg *et al.* (1995) suggested that this peat was formed in a depression on the emergent floor of Lake Agassiz, during the

Moorhead Phase. Outlets in use at this time could have been the eastern (traditional interpretation) or northwestern (non-traditional interpretation - Risberg *et al.*, 1995; Thorleifson, 1996). In either case, this scenario suggests that 1) Agassiz waters were below the level of the Wampum East site during the Moorhead, and that 2) the Upper Campbell beach would not have been constructed until after the peat layer, and overlying silty clay was deposited (Risberg *et al.*, 1995). This would have presumably occurred after 9460 yr BP (Risberg *et al.*, 1995).

New interpretations at this site (Teller *et al.*, submitted) suggest that: 1) the buried peat, overlying clayey-sandy-silt, and silty sand, represent later Emerson Phase (Upper Campbell level) deposition; 2) the 10 340 yr BP date, associated with the 9460 yr BP date in the buried peat under the Upper Campbell beach sand is likely reworked from older peaty deposits because it stratigraphically overlies a 9950 yr BP date; and 3) organic-bearing sands and silts in the lower section of Wampum East cores represent wave reworking during the post-Moorhead transgression of Lake Agassiz over the Wampum area to (perhaps) the Norcross level (at about 10 000 yr BP) -- dates of 10 140 +/- 80 [TO 4872], 10 040 +/- 70 [TO 4871], and 9950 +/- 90 [TO 4874] yr BP in the lower organic-bearing silts and sands at Wampum East support this hypothesis.

To further support this line of reasoning, Teller *et al.* (submitted) also suggest that at both the Wampum West and Wampum East sites, the Emerson Phase organic-bearing silts and sands may consist of two depositional episodes (Fig. 7-2). One would be related to the transgression of Lake Agassiz over the Moorhead low water surface, and the other related to deposition in the protected bay behind a growing Upper Campbell beach (Teller *et al.*, submitted). The dates in the organic sand and silt sequence that are older than about 10 000 yr BP may represent reworked organics from the Moorhead surface that were incorporated in nearshore sands of the early Emerson (Norcross and Tintah levels) Phase (i.e. 10.0 - 9.7 ka BP). Teller *et al.* (submitted) also observes that many of the organic fragments in the lower organic-bearing sands and silts appear to have been reworked, while those stratigraphically above do not.

Paleoenvironmental reconstruction of the Wampum East Upper Campbell lagoon, by using diatoms, suggests that the early lagoon was variably fresh to brackish. This may have been a consequence of dilute Agassiz waters -- entering the lagoon during storms as beach overwash -- mixing with more saline lagoonal waters and overland runoff (Risberg *et al.*, 1995).

7.3 - The Swift and Sioux Pond Sites (Bjork & Keister, 1983)

Bjork and Keister (1983) studied two Emerson Phase Lake Agassiz sites in order to evaluate the changes in lake level during the Emerson Phase, and to look at the paleoenvironment of the sites by using diatoms. The Swift site (SW in Fig. 7-1) is an Upper Campbell lagoonal site in Minnesota. The Sioux Pond site (SP in Fig. 7-1) is a small, overgrown pond situated atop the Sioux Lookout moraine, in northwestern Ontario. It was located well within the boundaries of Lake Agassiz during the Emerson Phase, and thus records the Emerson Phase in a different way than does the lagoon at the Swift Site.

The stratigraphy at the Swift site consists of 20+ cm of Lake Agassiz silt at the base, overlain by 5 cm of organic silt, 2 cm of gyttja, a 38 cm thick sequence of woody peat, and finally 2 m of well sorted, rounded pebble gravel. Two radiocarbon dates on wood and peat are 10 050 +/- 100 yr BP [WIS-1325] at the base of the peat, and 9350 +/- 100 yr BP [WIS-1324] at the top of the peat.

Bjorck and Keister (1983) suggest two possible scenarios to account for deposition of the peat, and the subsequent transgression of Lake Agassiz at the Swift site: 1) the peat formed in a small depression within the beach ridge between two "Campbell" level high-water phases of Lake Agassiz -- as such the peat represents a regressive phase of the lake, or 2) the peat represents a transgression of Lake Agassiz between 10 000 yr BP and 9350 yr BP, ending in deposition of the gravel -- in this scenario the organic silt represents a "soil" formed during a low stand of Agassiz, between higher levels of Lake Agassiz, the first recorded by the basal silt, and the second by the gyttja-peat-gravel section (see Fig. 7-2).

Based on findings at both the Wampum West and East sites, and arguments presented by Risberg *et al.* (1995), Thorleifson (1996), and Teller *et al.* (submitted), a reinterpretation is perhaps in order. The basal silts of this sequence, while interpreted by Bjorck and Keister (1983) to have been deposited in Lake Agassiz, are interpreted here to have specifically been deposited during the early Emerson Norcross highstand of Lake Agassiz. Declining lake levels led to the deposition of the organic silt and gyttja at the Swift site, and finally the peat. I believe the basal 10 050 yr BP date is probably reworked, though it is chronologically representative of the transgression of Lake Agassiz from the Moorhead low to the early Emerson (Norcross) Lake Agassiz highstand (Teller *et al.*, submitted). Subsequent peat deposition at this site was, presumably near or before about 9350 yr BP, covered by gravel shoreline sediments of the Upper Campbell beach. Wright (1972) reported that this Campbell strandline was abandoned at 9200 +/- 600 yr BP [W-1057], based on a radiocarbon date on the same Campbell beach ridge west of the Swift site; Bjorck and Keister (1983) also placed this event at about 9300 yr BP.

The Sioux Pond site (SP in Fig. 7-1) probably was exposed during the Moorhead low level of Lake Agassiz (Bjorck and Keister, 1983). The stratigraphy of this site consists of 48 cm of basal Lake Agassiz red varved clays, overlain by 23 cm of reddish Agassiz varved clay, 5 cm of a massive grey clay, 5 cm of silty organic clay, 6 cm of clay-gyttja with mollusc shells, and finally 32 cm of a mollusc rich gyttja. The varved red clays are thought to represent a period of confluence between Lake Kaministikwia in the Superior basin and the Lake Agassiz basin, during a high level (Norcross) of Lake Agassiz (Thorleifson, 1996).

A radiocarbon date of 9740 +/-100 yr BP [WIS-1329] in the clay-gyttja that overlies the organic-rich clay, the grey clay, and the varved clay -- in combination with the diatom paleoenvironmental reconstruction of the clay-gyttja unit -- suggests that by 9740 yr BP the Sioux Pond site was isolated from Lake Agassiz and that the transgression of Lake Agassiz into the Sioux Pond site (which was also responsible for the deposition of the organic clay) occurred at about 9900 - 10 000 yr BP (Bjorck and

Keister, 1983) (see Fig. 7-2). This scenario agrees with the interpretation by Thorleifson (1996) that the early Emerson Lake Agassiz maximum level was the Norcross. Presumably the red varved clays at the base of the Sioux Pond sequence were deposited during this 9900 - 10 0000 yr BP transgression as well, when the Marquette readvance of ice into the Superior basin forced water to overflow westward into Lake Agassiz (Teller and Thorleifson, 1983). This allowed the confluence of Lakes Agassiz and Kaministikwia, and the deposition of the red clays across the drainage divide that separated the Lake Agassiz basin from the Great Lakes basin (Thorleifson, 1996).

Interpretation of the diatoms at the Sioux Pond site is not so simple. Diatoms are sparse in the Lake Agassiz sediments, as a consequence of high sediment accumulation rates and unfavorable biotic conditions (Bjorck and Keister, 1983). The interpretation that the 9900 - 10 000 yr BP transgression of Lake Agassiz was to the Norcross level (Thorleifson, 1996) necessitates looking at the autecology of diatoms found in the grey clays, and in the other units overlying the red varves at the Sioux Pond site. Diatoms in the grey clay, whose autecology suggests that the Sioux Pond site was separated from Lake Agassiz at that time, are similar to those found in the overlying organic clays (clayey-gyttja in Fig. 7-2), which were interpreted to have been deposited during the transgression of Lake Agassiz over the Sioux Pond site (Bjorck and Keister, 1983). It is the small occurrence of planktonic diatoms found in the organic clay (overlying the grey clay) that suggested to Bjorck and Keister (1983) that Lake Agassiz again inundated the Sioux Pond site after it had regressed from the area. However, diatoms in the organic clay that are 1) similar to those in the underlying grey clay, 2) terrestrial, and 3) planktonic such as those related to Lake Agassiz (which suggests that Lake Agassiz may have transgressed over the Sioux Pond site for a second time, after it had abandoned the site post-deposition of the underlying grey clay (Bjorck and Keister, 1983), opens the opportunity to present an alternative interpretation.

Bjorck and Keister (1983) clearly suggested that the Sioux Pond site was probably never under significant water depths during the Emerson Phase, and it also was in close proximity to the shoreline (perhaps enhancing the possibility of producing the mixed diatom assemblage of the organic clays). Also, as the change in diatom species composition is small from the grey clay to the overlying organic clay (which also reflects minor changes in sedimentation), Bjorck and Keister (1983) suggested that the diatom variation between the grey and organic clays could be regarded as insignificant. The Sioux Pond site would have been finally separated from Lake Agassiz by a combination of differential uplift and declining Agassiz lake levels, after which the overlying gyttja was deposited (Fig. 7-2).

7.4 - The Mosbeck Site (Ashworth *et al.*, 1972)

The Mosbeck site is an exposure in a drainage ditch, on the eastern margin of the southern Agassiz basin, 3 km west of the Upper Campbell beach, and about 2 km north and east of a Campbell spit (MB in Fig. 7-1). During the Moorhead low, a silty sand with snail shells was deposited at the Mosbeck site, perhaps in a small, isolated body of water on the emergent floor of Lake Agassiz (Fig. 7-2). Continued low levels of Lake Agassiz during the Moorhead Phase caused the water table to decline, and the encroachment of terrestrial vegetation into this small pond. These two factors combined to cause deposition of a woody peat over the silty sand.

As Lake Agassiz transgressed over the Mosbeck site at the end of the Moorhead Phase, a layer of organic-rich silty fine sand shoreline sediment was deposited. The organic materials were reworked from the underlying peaty and woody material associated with the Moorhead surface. Ashworth *et al.* (1972) suggest this is a transgressional shoreline sediment during deepening to the Upper Campbell level. Thorleifson (1996) argued that this transgression reached the Norcross level. The sedimentology, the 9940 yr BP radiocarbon date, and the arguments of Teller *et al.* (submitted) at the Wampum sites, support Thorleifson's idea. Furthermore, Ashworth *et al.* (1972) note that evidence from the Snake Curve section (16 km southeast of this site, see description below and Fig. 7-2), indicates that the early Emerson transgression of Lake Agassiz rose to a level above the Upper Campbell.

Above the organic-rich silty fine sand, an interbedded sandy gravel and silty sand was deposited. This is the shoreline sediment of Lake Agassiz at the Mosbeck site, likely deposited when Lake Agassiz was at the Upper Campbell level. A radiocarbon date on wood from interbedded sandy gravel and silty sand of the Emerson shoreline sediments is 9940 +/- 160 yr BP [I-3880]. The wood was likely eroded from the woody peat deposited during the Moorhead low water phase.

7.5 - The Snake Curve Section (Moran *et al.*, 1971)

The Snake Curve section is located on a riverbank cut of the Red Lake River near Red Lake Falls, Minnesota. It is also found on the eastern side of Lake Agassiz, on the Upper Campbell shoreline (SC in Fig. 7-1). When the level of Lake Agassiz dropped to the Moorhead level, the Snake Curve site was subaerially exposed. During this time, the ancestral Red Lake River formed, flowing west over the Snake Curve site, downcutting 25 feet into lake plain sediments. A sequence of flat-bedded gravel and sandy gravel with cross bedded gravelly sands were deposited in this ancient river bed. Gastropods found in these gravels, sandy gravels, and gravelly sands suggest neither purely lacustrine nor purely fluvial conditions. The paleoenvironment was likely one of a stream or small river flanked by woodlands and small ponds (Moran *et al.*, 1971). A date at the base of these sediments places the age of this buried stream at 9890 +/- 150 yr BP [I-4853]. This date is in agreement with reconstructions on the age of the Moorhead Phase (10.9 - 9.9 ka BP). The riverbed deposits of the ancient Red Lake River are overlain first by flat bedded and ripple cross-stratified shoreline sands, followed by lacustrine silt, and finally thinly-bedded, well sorted shoreline sands. This sequence implies that the transgression of Lake Agassiz from the Moorhead low to start the Emerson Phase: 1) reached a level about the same as the Upper Campbell, depositing the first shoreline sands at the Snake Curve site, 2) transgressed to an even higher lake level (Norcross?), drowning the Snake Curve section, and depositing the lacustrine silts, and 3) lowered in level from the Emerson transgressive maximum (Norcross) back to the Upper Campbell level, stabilized there, and deposited the second package of shoreline sands.

7.6 - The Rossendale Gully (Teller, 1989)

The Rossendale site (RD in Fig. 7-1) is situated on the western side of the Agassiz basin in a small underfit valley that leads to a lagoonal depression about 1 km west of the Upper Campbell beach (Teller, 1989). The stratigraphy of the site consists of organic detritus and wood overlain by grey clay, silty marl, grey clay, silty to clayey sand, and clayey silt. Wood dates from the basal organic-rich layer are 9600 +/- 70 yr BP [TO-534] and 9510 +/- 90 BP yr BP [GSC-4490]. Paleoenvironmental reconstruction of the mollusc and ostracode communities found in the basal organics and overlying marl suggest that the deposition of these units (and presumably the overlying clays and clayey sands) was unrelated to any direct connection with Lake Agassiz. The Rossendale site is at about the same elevation as the Upper Campbell strandline, and by the time this organic sequence at the Rossendale gully was deposited, it is likely that the Upper Campbell beach had been built, serving as a barrier to isolate the Rossendale site from the main body of Lake Agassiz (Teller, 1989). Thus, by utilizing these radiocarbon dates, and thereby accepting a younger deglacial chronology over the previously accepted old chronology (by eliminating old radiocarbon dates from moss at this site which is known to accumulate old carbon), the age of the Upper Campbell beach was interpreted by Teller (1989) to be closer to 9500 - 9600 yr BP rather than 9900 yr BP, as previously thought.

7.7 - SUMMARY:

In summary (Fig. 7-2), by looking at several Lake Agassiz Campbell sites together, it appears that several conclusions are fairly certain:

1) A transgression of Lake Agassiz, dated 10 000 - 9900 yr BP, occurred after the Moorhead low-water phase, drowning previously subaerial sites.

2) The early Emerson (9900 - 9700 yr BP) maximum Lake Agassiz level was the Norcross. The Tintah level was occupied shortly thereafter, as Lake Agassiz began its drop in lake level to the Upper Campbell.

3) While Lake Agassiz was at the Norcross level (and presumably Tintah level as well), the Brokenpipe Lake site was the most ice-proximal of all sites considered in Figure 7-2. It is the only site which clearly contains ice-rafted detritus (sediments associated with ice-proximal deposition) during deposition of the Norcross (and Tintah?) level. This piece of information supports the conclusion of Thorleifson (1996) that the ice margin at this time was well south in Manitoba (perhaps into the Interlake area), and was very close to (if not in contact with) the Manitoba Escarpment near the Brokenpipe Lake site (see Figs. 1-7 & 6-3 D).

4) The Upper Campbell level was occupied by Lake Agassiz between about 9700 yr BP and 9400 yr BP, and stable Upper Campbell water levels were achieved by 9500 yr BP. Shortly thereafter, Lake Agassiz dropped below the Upper Campbell, stabilizing at the Lower Campbell by about 9400 yr BP.

CHAPTER 8 - SUMMARY

All the lake sites (Brokenpipe, Ruby, Jay Jay, Gregory/Bird Lakes) included in this study were initially occupied by cold, dilute Lake Agassiz waters. Through time, the Upper Campbell beach developed and formed lagoons at the Brokenpipe, Ruby and Jay Jay lake sites, as they were separated by the beach berm from the main body of Lake Agassiz. After Lake Agassiz fell from the Upper Campbell, the hydrology and hydrochemistry of these sites eventually became predominantly governed by watershed geology, shallow groundwater flow systems, and to a lesser extent, climate. This is reflected in the modern water chemistry of the lakes. The fossil ostracode record and sedimentation history of each site can be used to infer the relationship between the lagoon and Lake Agassiz.

Based on a detailed analysis of the geomorphology and ostracode record of the Brokenpipe Lake site, several conclusions were reached. The Upper Campbell beach grew as a broad, flat-topped, subaqueous spit platform in a northward direction past the Brokenpipe lake site from about 9.7 - 9.4 ka BP. The initial lagoon at the Brokenpipe site was moderately deep (~7-8 m), relatively cool and dilute, although its temperature and salinity fluctuated due to variable influxes of meltwater through small channels in the till west of the site, and possibly seepage through till of shallow groundwater systems into the site. Sedimentation during this deeper lagoonal period was predominantly silty to clayey fine sands and clayey silts, forming massive to finely laminated units. Several discrete layers composed of sand were possibly deposited by beach overwash of a stormy Lake Agassiz into the lagoon, though based on the ostracodes found in these sands, an alluvial source is equally as plausible. As Lake Agassiz receded from the Brokenpipe Lake area, isolating the lagoon to form Brokenpipe Lake, the lake became much shallower, more constricted, and relatively more saline, although it still maintained its variability in temperature salinity, now due to shallower water depths and groundwater seepage through till, rather than as a result of meltwater (lotic) inflow from stagnant ice on the Riding Mountain uplands, or Lake Agassiz overwash. Biogenic carbonate marl production by *Chara sp.* algae was predominant in the lake at this time.

The ostracode fauna in the silty and clayey fine sands, and clayey silts (Biozone A) shows abrupt changes, suggesting large changes in salinity and water temperature were common during deposition of these sediments. Ostracode faunal abundances are much more constant in the upper calcareous marls (Biozone B). This suggests variations in water temperature and salinity were less pronounced, and were similar to the variability observed in modern Brokenpipe Lake.

During the mid-Holocene (centered around 6 ka BP), a period of reduced effective moisture caused a cessation of marl production and a shift from perennial wetness to periodic dryness at the Brokenpipe site. Additional radiocarbon ages may indicate a significant hiatus at this stratigraphic interval. The return to relatively wetter conditions re-started calcareous marl production and ostracode inhabitation. Based on other work, sedimentation may have resumed at about 4 ka BP, which appears to have continued to the present.

Lake Agassiz lagoons in the Hudson Bay, Saskatchewan area (Ruby Lake and Hudson Bay Lower sites), while contributing neither radiocarbon dates nor ostracodes yield other important pieces of information. Deposition of weathered shale in a proglacial lake environment indicates that either 1) stagnant ice meltwater and associated outwash from the Pasquia and Porcupine Hills uplands washed the shales into Lake Agassiz (or another, smaller proglacial lake that preceded the inundation of Lake Agassiz in the Hudson Bay area; e.g. Lake Somme; Moran, 1969), or 2) a proglacial lake pinned by the Laurentide Ice Sheet at higher elevations than Lake Agassiz (e.g. Lake Saskatchewan; Christiansen, 1979; Christiansen *et al.*, 1995), while still receiving stagnant ice meltwater and outwash from the uplands, was covering the Hudson Bay area prior to occupation by Lake Agassiz. Subsequent deposition of sand and gravel at the Ruby Lake and Hudson Bay Lower core sites occurred when the margin of Lake Agassiz stood at the Campbell levels. Deposition occurred as a result of 1) the incipient construction of the Campbell strandlines, 2) sedimentation from suspension, or 3) slumping and density underflows, initiated by Lake Agassiz beach overwash during storms, and the instability of the sand and gravel strandline in the lagoonal environment.

As the Laurentide Ice Sheet continued to retreat after the Marquette readvance, thinning ice in the Wapawekka/Cub Hills region probably would have disintegrated to isolated patches, forming several sub-glacial and ice-marginal meltwater features. The progradation of the Upper Campbell strandline at the Jay Jay Lake site occurred as a broad, flat-topped baymouth bar which cut off the small bay of Lake Agassiz between the Wapawekka and Cub Hills (of which today Jay Jay Lake is a part), to form a lagoon. Sedimentation in the lagoon, likely deposited by overland meltwater flow, was dominated by fine sand and silt reworked from the sandy till of the region. Beach overwash coarse sands from Lake Agassiz are interbedded with the finer sediments in Jay Jay Lake. None of the cores from the Jay Jay Lake site fully penetrated the post- Lake Agassiz sequence. The ostracodes in the lower part of the JJC core suggest that there was some appreciable amount of lotic input to the Jay Jay lagoon, although some of these ostracodes may have been living in a lacustrine (versus lotic) association. In general, these Jay Jay Lake fossil ostracodes indicate a cold, tundra type of condition with E>P during the earlier lagoonal stages. Ostracodes from the upper organic portions of the JJC core indicate that after retreat of Lake Agassiz from the area, the lake was shallow, variable in temperature and salinity, and perhaps subject to shallow groundwater discharges from the surficial sediments of the area. Lowering water levels culminated in the inception of vegetation encroachment and peat deposition at the Jay Jay Lake site.

Reconstruction of the deglacial history at the Gregory lake site shows that the Gregory lake site marks the end of the well-developed, continuous Campbell strandline in the northwestern arm of Lake Agassiz. From this point westward, the strandline becomes a discontinuous assemblage of many small strandlines formed on an archipelago of islands consisting of ridges and mounds of ground moraine. The succession of massive to periodically bedded and laminated clay, silt, and silty fine sand deposited at the

Gregory Lake site is the result of its early inundation by Lake Agassiz. An overlying bed of clean sand, likely deposited during beach building at the Upper Campbell level, was probably reworked from sandy till. After Lake Agassiz retreated from the Gregory Lake area, vegetation encroachment and peat deposition began.

The manipulation of Global Positioning System (GPS) data points taken from the Upper and Lower Campbell beaches along the western margin of Lake Agassiz has not solved strandline correlation discrepancies in the Wapawekka Hills area. By using traditional isobases, the portion of previously uncorrelated, well-developed strandline that continues west of the Wapawekka Hills, into the northwestern arm of Lake Agassiz is part of the Upper Campbell beach. Thus, Gregory Lake is located behind the Upper Campbell strandline. However, the absence of a well-developed Lower Campbell beach suggests that the Lower Campbell beach continues west of the Wapawekka Hills into the northwestern arm of Lake Agassiz, and that the Upper Campbell strandline ends at the northern side of the Wapawekka Hills (where the Upper Campbell level of Lake Agassiz was blocked by ice).

Manipulation of isobase orientation as applied to the GPS beach data suggests that the traditionally accepted isobases are still the most appropriate model for the Lake Agassiz basin, and changes in isobase orientation are not necessarily warranted by the data at this time. The analysis of GPS beach data points, and the interpretations made about the Campbell strandline correlation in the Wapawekka Hills area are not conclusive, and solution of the Upper/Lower Campbell correlation problem at the

Wapawekka Hills is subject to future endeavors.

Analysis of Upper Campbell lagoon sites in this study, and from Upper Campbell-related sites studied by others suggests that a major transgression of Lake Agassiz occurred from 10 - 9.9 ka BP, after the Moorhead low-water phase. The transgressing waters drowned and reworked organic-rich sediments and peat of previously subaerial sites. Following a readvance of the Laurentide Ice Sheet at about 10 ka BP, the early (9.9 - 9.7 ka BP) Emerson Phase maximum level of Lake Agassiz probably reached above the Upper Campbell to the Norcross level. The Tintah level was occupied shortly thereafter, as Lake Agassiz began its drop in lake level to the Upper Campbell. The Upper Campbell level was occupied between about 9.7 ka BP and 9.4 ka BP. However, Lake Agassiz probably dropped below the Upper Campbell by about 9.4 ka BP.

A body of water the size of Lake Agassiz would surely have influenced its surrounding regional climate, the lakes, rivers, and oceans into which it flowed, and probably would have affected the dynamics of the nearby Laurentide Ice Sheet as well. Several recent studies investigated the chronology and outlet geometry of Lake Agassiz outflow to other watersheds (and the potential climatological impacts of these events), the extent, level, and depositional history of the many different phases of the lake, and the paleoecology and paleolimnology of Lake Agassiz itself, and of its backbeach lagoons.

There is still a large requirement for continued work in the Lake Agassiz basin; additional radiocarbon age control, GPS-based shoreline measurements and correlation, stratigraphic and sedimentological investigations, GIS-based basin modelling, as well as further paleoecological study in northern regions would be useful. While many different pieces of an immensely large puzzle have been put together to produce what seems to be a reasonable history of events, there are still fundamental questions that need be answered, and interpretations that need to be refined or tested, in order to produce the best vision of what the history of this immense glacial lake was, and how it influenced late-glacial conditions in central North America.

REFERENCES

Andrews, John T., 1987, The late Wisconsinan glaciation and deglaciation of the Laurentide Ice Sheet; *In* Ruddiman, W.F., and Wright, H.E., Jr. (eds.), *North America and adjacent oceans during the last deglaciation*, Geological Society of America, The Geology of North America **K-3**, p. 13-39.

Antevs, E., 1951, Glacial clay in Steep Rock Lake, Ontario, Canada; *Geological Society of America Bulletin 62*, p. 1223-1262.

Arndt, B. M., 1977, Stratigraphy of offshore sediment, Lake Agassiz, North Dakota; *North Dakota Geological Survey Report of Investigation 60*, 58 pp.

Ashworth, A. C., Clayton, L., and Bickley, W. B., 1972; The Mosbeck site: a paleoenvironmental interpretation of the late Quaternary history of Lake Agassiz based on fossil insect remains; *Quaternary Research*, **2**, p. 176-188.

Ayres, K. W., Anderson, D. W., and Ellis, J. G., 1983, The soils of the northern provincial forest in the Pasquia Hills and Saskatchewan portion of the Pas map areas (63E and 63F Saskatchewan); *Saskatchewan Institute of Pedology, Publication SF4, Publication Extension 260*, 110 pp.

Bajc, A. F., 1987, Molluscan paleoecology and Superior basin water levels, Marathon, Ontario; Msc.Thesis, *University of Waterloo*, 271 pp.

Barnett, P. J., 1992, Quaternary geology of Ontario; *In:* Thurston, P. C., Williams, H. R., Sutcliffe, R. H., and Stott, G.M., (eds.), *Geology of Ontario*, Ontario Geological Survey Special Volume 4, p. 1011-1088.

Betcher, R., Grove, G., and Pupp, C., 1995, Groundwater in Manitoba: hydrogeology, quality concerns, management; *National Hydrology Research Institute*, contribution number CS-93017, 45 pp.

Bjorck, S., and Keister, C. M., 1983, The Emerson Phase of Lake Agassiz , independantly registered in northwestern Minnesota and northwestern Ontario; *Canadian Journal of Earth Sciences*, **20**, p. 1536-1542.

Broecker, W. S., 1966, Glacial rebound and the deformation of the shorelines of proglacial lakes; *Journal of Geophysical Research*, **71**, p. 4777-4783.

Brophy, J.A., and Bluemle, J.P., 1983, The Sheyenne River: its geological history and effects on Lake Agassiz; *In:* Teller, J.T., and Clayton, Lee (eds.), *Glacial Lake Agassiz*, Geological Association of Canada Special Paper 26, p. 173-186.

Brouwers, E. M., and Forester, R. M., 1992, Ostracode assemblages from modern bottom sediments of Vitus Lake, Bering Piedmont Glacier, southeast Alaska; *Geologic Studies in Alaska by the U.S. Geological Survey*, p. 228-235.

Buckley, S. B., 1975, Study of post-Pleistocene ostracode distribution in the soft sediments of southern Lake Michigan; PhD. Thesis, *University of Illinois at Urbana-Champaign.*

Campbell, J. E.,1987, Surficial geology of the Pasquia Hills area, (63 E, F), Saskatchewan; *Saskatchewan Research Council, Sedimentary Resources,* 1: 250,000 Map.

Campbell, J. E., 1988, Surficial geology of the Wapawekka area (73I) Saskatchewan; *Saskatchewan Research Council, Sedimentary Resources*, 1:250,00 Map.

Christiansen, E. A., 1979, The Wisconsin deglaciation of southern Saskatchewan and adjacent areas; *Canadian Journal of Earth Sciences*, **16**, 913-938.

Christiansen, E. A., Sauer, E. K., and Schreiner, B. T., 1995, Glacial Lake Saskatchewan and Lake Agassiz deltas in east-central Saskatchewan with special emphasis on the Nipawin Delta; *Canadian Journal of Earth Sciences,* **32** (3), p. 334-348.

Clayton, L., 1983, Chronology of Lake Agassiz drainage to Lake Superior; *In:* Teller, J.T., and Clayton, Lee (eds.), *Glacial Lake Agassiz*, Geological Association of Canada Special Paper 26, p. 291-307.

Clayton, L., and Moran, S. R., 1982, Chronology of Late-Wisconsinan glaciation in middle North America; *Quaternary Science Reviews*, **1**, p. 55-82.

Clayton, L., Teller, J.T., and Attig, J.W., 1985, Surging of the southwestern part of the Laurentide Ice Sheet; *Boreas*, **14**, p. 235-241.

Cole, G. A., 1983, Textbook of Limnology; Waveland Press Inc., Illinois, 401 pp.

Colman, S. M., Jones, G. A., Forester, R. M., and Foster, D. S., 1990, Holocene paleoclimatic evidence and sedimentation rates from a core in southwestern Michigan; *Journal of Paleolimnology*, **4**, p. 269-284.

Curry, B. B., 1995, Late Pleistocene lithofacies, paleolimnology, and ostracode fauna of kettles on the Illinoian till plain; PhD Thesis, *University of Illinois at Urbana-Champaign.*

Curry, B. B., 1997, Paleochemistry of Lakes Agassiz and Manitoba based on ostracodes; *Canadian Journal of Earth Sciences*, **34**, p. 699-708.

Curry, B. B., 1998 (in press), An environmental tolerance index for ostracodes as indicators of physical and chemical factors in aquatic habitats; *Palaeogeography, Palaeoclimatology, Palaeoecology.*

Dean, W. E., 1974, Determination of carbonate and organic matter in calcareous sediments and sedimentary rocks by loss on ignition: comparison with other methods; *Journal of Sedimentary Geology*, **44**, p. 242-248.

De Deckker, P., and Forester, R. M., 1988, The use of ostracodes to reconstruct continental paleoenvironmental records; *In:* De Deckker, P., Colin, J. P., and Peypouquet, J. P., (eds.), *Ostracoda in the Earth Sciences*, p. 175-199.

Delorme, L. D., 1967, Field key and methods of collecting freshwater ostracodes in Canada; *Canadian Journal of Zoology*, **45**, p. 1275-1281.

Delorme, L. D., 1968, Pleistocene freshwater Ostracoda from Yukon, Canada; *Canadian Journal of Zoology*, **46**, p. 859-876.

Delorme, L. D., 1969, Ostracodes as Quaternary paleoecological indicators; *Canadian Journal of Earth Sciences*, **6**, p. 1471-1476.

Delorme, L. D., 1970a, Freshwater ostracodes of Canada, Part III: family Candonidae; *Canadian Journal of Zoology*, **48**, p. 1099-1127.

Delorme, L. D., 1970b, Freshwater ostracodes of Canada, Part II: subfamily Cypridopsinae, and Herpetocypridinae, and family Cyclocyprididae; *Canadian Journal of Zoology*, **48**, p. 253-266.

Delorme, L. D., 1970c, Freshwater ostracodes of Canada, Part I: subfamily Cypridinae; *Canadian Journal of Zoology*, **48**, p. 153-168.

Delorme, L. D., 1970d, Freshwater ostracodes of Canada, Part IV: families Ilyocyprididae, Notodromadidae, Darwinulidae, Cytherideidae, and Entocytheridae; *Canadian Journal of Zoology*, **48**, p. 1251-1259.

Delorme, L. D., 1971a, Freshwater ostracodes of Canada, Part V: families Limnocytheridae, Loxoconchidae; *Canadian Journal of Zoology*, **49**, p. 43-64.

Delorme, L. D., 1971b, Paleoecological determinations using Pleistocene freshwater ostracodes; *Centre de Recherche Pau-SNPA Bulletin*, **5** (supplement), p. 341-347.

Delorme, L. D., 1982, Lake Erie oxygen, the prehistoric record; *Canadian Journal of Fisheries and Aquatic Sciences*, **39**, p. 1021-1029.

Delorme, L. D., 1989, Methods in Quaternary ecology 7: freshwater ostracodes; *Geoscience Canada*, **16** (2), p. 85-90.

Delorme, L. D., 1991, Ostracoda; *In:* Thorp, J. H., and Covich, A. P., (eds.), *Ecology and Classification of North American Freshwater Invertebrates*, Academic Press, p. 691-722.

Drexler, C. W., Farrand, W. R., and Hughes, J. D., 1983, Correlation of glacial lakes in the Superior basin with eastward discharge events from Lake Agassiz; *In:* Teller, J.T., and Clayton, Lee (eds.), *Glacial Lake Agassiz*, Geological Association of Canada Special Paper 26, p. 309-329.

Dyke, A. S., and Prest, V. K., 1987, Late Wisconsinan and Holocene history of the Laurentide Ice Sheet; *Geographie Physique et Quaternaire*, **41**, 237-263.

Ehrlich, W. A., Pratt, L. E., and Leclaire, F. P., 1959, Report of reconnaissance soil survey of Grandview map sheet area; *Manitoba Soil Survey*, Soils Report No. 9, 96 pp.

Elson, J. A., 1967, Geology of glacial Lake Agassiz; *In:* Mayer-Oakes, W.J. (ed.) *Life, land, and water*, University of Manitoba Press, Winnipeg, p. 36-95.

Elson, J. A., 1983, Lake Agassiz - Discovery and a century of research; *In:* Teller, J.T., and Clayton, Lee (eds.), *Glacial Lake Agassiz*, Geological Association of Canada Special Paper 26, p. 21-41.

Faure, G., 1991, Principles and applications of inorganic geochemistry; Mac Millan Publishing Co., NY.

Fenton, M. M., Moran, S. R., Teller, J. T., and Clayton, L., 1983, Quaternary stratigraphy and history in the southern part of the Lake Agassiz basin; *In:* Teller, J.T., and Clayton, Lee (eds.), *Glacial Lake Agassiz*, Geological Association of Canada Special Paper 26, p. 49-74.

Fisher, T. G., 1993, Glacial Lake Agassiz: the N.W. outlet and paleoflood spillway, N.W. Saskatchewan and N.E. Alberta; PhD Thesis, *University of Calgary*, 206 pp.

Fisher, T. G., and Smith, D. G., 1994, Glacial Lake Agassiz: its northwest maximum extent and outlet in Saskatchewan (Emerson Phase); *In:* Teller, J. T., and Kehew, A. K., (eds.) Late glacial history of large proglacial lakes and meltwater runoff along the Laurentide Ice Sheet, *Quaternary Science Reviews*, **13** (9-10), p. 845-858.

Fisher, T. G., and Souch, C., 1998 (in press), N.W. Lake Agassiz outlet channels, isostatic tilting, and a migrating drainage divide, Saskatchewan, Canada; *Geomorphology*.

Forester, R. M., 1987, Late Quaternary paleoclimate records from lacustrine ostracodes; *In:* Ruddiman, W.F., and Wright, H.E., Jr. (eds.), *North America and adjacent oceans during the last deglaciation*, Geological Society of America, The Geology of North America **K-3**, p. 261-276.

Forester, R. M., 1983, Relationship of two lacustrine ostracode species to solute composition and salinity; implication for paleohydrochemistry; *Geology*, **11**, p. 435-438.

Forester, R. M., 1986, Determination of the dissolved anion composition of ancient lakes from fossil ostracodes; *Geology*, **14**, p. 796-798.

Forester, R. M.,1988, Nonmarine calcareous microfossil sample preparation and data acquisition procedures; *United States Geological Survey*, Technical Procedure HP-78 RI, p. 1-9.

Forester, R. M., 1991, Pliocene climate history of the western United States derived from lacustrine ostracodes; *Quaternary Science Reviews*, **10**, p. 133-146.

Forester, R. M., Delorme, L. D., and Bradbury, J. P., 1987, Mid-Holocene climate in northern Minnesota; *Quaternary Research*, **28**, 263-273.

Forester, R. M., Delorme, L. D., and Ager, T. A., 1989, A lacustrine record of late Holocene climate change from south-central Alaska; *Geophys. Monogr.*, **55**, p. 33-40.

Forester, R. M., Colman, S. M., Reynolds, R. I., and Keigwin, L. D., 1994, Lake Michigan's late Quaternary limnological and climate history from ostracode, oxygen isotope, and magnetic susceptibility; *Journal of Great Lakes Research*, **20**, p. 93-107.

Gorham, E., 1961, Factors influencing the supply of major ions to inland waters, with special reference to the atmosphere; *Geological Society of America Bulletin*, **72**, p. 795-840.

Harris, K. L., Moran, S. R., and Clayton, L., 1974, Late Quaternary stratigraphic nomenclature, Red River Valley, North Dakota and Minnesota; *North Dakota Geological Survey Miscellaneous Series 52*, 47 pp.

Head, W. K., Anderson, D. W., and Ellis, J. G., 1981, The soils of the Wapawekka map area (73I) Saskatchewan; *Saskatchewan Institute of Pedology Publication SF6, Extension Publication 303*, 108 pp.

Hem, J. D., 1989, Study and interpretation of the chemical characteristics of natural water; *United States Geological Survey Water Supply Paper 2254*, 3rd edition, Department of the Interior, Washington, D.C., 263p.

Hu, F. S., Wright, H. E. Jr., Ito, E., and Lease, K., 1997; Climatic effects of glacial Lake Agassiz in the midwestern United States during the last deglaciation; *Geology*, **25** (3), p. 207-210.

Johnston, W.A., 1946, Glacial Lake Agassiz with special reference to the mode of deformation of the beaches; *Geological Survey of Canada Bulletin*, **7**, 20 pp.

Karrow, P. F., Ceska, A., Hebda, R. J., Miller, B. B., Seymour, K. L., and Smith, A. J., 1995, Diverse nonmarine biota from the Whidbey Formation (Sangamonian) at Point Wilson, Washington, *Quaternary Research*, **44**, p. 434-437.

Kehew, A.E., and Clayton, L., 1983, Late Wisconsinan floods and development of the Souris-Pembina spillway system in Saskatchewan, North Dakota, and Manitoba; *In:* Teller, J.T., and Clayton, Lee (eds.), *Glacial Lake Agassiz*, Geological Association of Canada Special Paper 26, p. 187-209.

Klassen, R. W., 1979, Pleistocene geology and geomorphology of the Riding Mountain and Duck Mountain area, Manitoba-Saskatchewan; *Geological Survey of Canada Memoir 396*, 52 pp.

Klassen, R. W., 1983a, The Assiniboine delta and the Assiniboine-Qu'Appelle valley system - implications concerning the history of Lake Agassiz in southwestern

Manitoba; *In:* Teller, J.T., and Clayton, Lee (eds.), *Glacial Lake Agassiz*, Geological Association of Canada Special Paper 26, p. 211-229.

Klassen, R. W., 1983b, Lake Agassiz and the late glacial history of northern Manitoba; *In:* Teller, J.T., and Clayton, Lee (eds.), *Glacial Lake Agassiz*, Geological Association of Canada Special Paper 26, p. 97-115.

Klassen, R. W., 1989, Quaternary geology of the southern Canadian interior plains; *In:* Fulton, R.J. (ed.) *Quaternary geology of Canada and Greenland*, Geological Society of America, The Geology of North America **K-1**, p. 138-173.

Kupsch, W. O., 1974, The Churchill-Reindeer Rivers area, - evolution of its landscape; *Musk-ox*, **15**, p. 10-29.

Kutzbach, J. E., and Wright, H. E., Jr., 1985, Simulation of the climate of 18,000 years BP; results from the North American/North Atlantic/European sector and comparison with the geologic record of North America; *Quaternary Science Reviews*, **4**, p.147-188.

Langford, F. F., 1973, The geology of the Wapawekka area, Saskatchewan; *Saskatchewan Department of Mineral Resources*, Report 147, 36 pp.

Langford, F. F., 1977, Northern extent of Lake Agassiz in eastern Saskatchewan; *Canadian Journal of Earth Sciences*, **14**, p. 1286-1291.

Last, W. M., 1980, Sedimentology and post-glacial history of Lake Manitoba; PhD. Thesis, *University of Manitoba*, 687 pp.

Last, W. M., 1984, Modern sedimentology and hydrology of Lake Manitoba; *Canadian Environmental Geology*, **5**, p. 177-190.

Last, W. M., 1990, Paleochemistry and paleohydrology of Ceylon Lake, a salt-dominated playa basin in the northern Great Plains, Canada; *Canadian Journal of Paleolimnology*, **4**, 219-238.

Last, W. M., and Teller, J. T., 1983, Holocene climate and hydrology of Lake Manitoba; *In:* Teller, J.T., and Clayton, Lee (eds.), *Glacial Lake Agassiz*, Geological Association of Canada Special Paper 26, p. 211-229.

Last, W.M., and Sauchyn, D. J., 1993, Mineralogy and lithostratigraphy of Harris Lake, southwestern Saskatchewan; *Canadian Journal of Paleolimnology*, **9**, p. 23-39.

Last, W. M., Teller, J. T., and Forester, R. M., 1994, Paleohydrology and paleochemistry of Lake Manitoba, Canada: the isotope and ostracode records; *Journal of Paleolimnology*, **12**, p. 269-282.

Lemmen, D., Duk-Rodkin, A., and Bednarski, J., 1994, Late glacial drainage systems along the northeastern margin of the Laurentide Ice Sheet; *In:* Teller, J. T., and Kehew, A. K., (eds.) Late glacial history of large proglacial lakes and meltwater runoff along the Laurentide Ice Sheet,*Quaternary Science Reviews*, **13** (9-10), p. 805-828.

Little, J., 1973, Groundwater availability in the Dauphin area; *Manitoba Department of Natural Resources, Water Resources Branch*, Groundwater Availability Studies Report No.10, Winnipeg.

Loffler, H., 1986, Ostracod analysis; *In:* Berglund, B. E., (ed.), *Handbook of Holocene Paleoecology and Paleohydrology*, John Wiley and Sons, p. 693-702.

Loffler, H., 1997, The role of ostracods for reconstructing climatic change in Holocene and Late Pleistocene lake environment in central Europe; *Journal of Paleolimnology*, **18**, p. 29-32.

Mackereth, F. J. H., Heron, J., and Talling, J. F., 1978, Water analysis: some revised methods for limnologists; *Freshwater Biological Association*, pub. 36, p. 35-42.

Maher, L. J. Jr., Miller, N. G., Baker, R. G., Curry, B. B., and Mickelson, D. M., 1998, Paleobiology of the sand beneath the Valders diamicton at Valders, Wisconsin. *Quaternary Research*, **49**, p. 208 - 221.

Mahnic, P., and Teller, J. T., 1985, History of sedimentation in the northwestern Lake Superior basin as related to Lake Agassiz overflow; *CANQUA Symposium on the paleoenvironmental reconstruction of the Late Wisconsin deglaciation and the Holocene*, Program with Abastracts, p. 44.

Mann, J. D., Rayburn, J., and Teller, J. T., 1997, Brokenpipe Lake Manitoba: a remnant of an Emerson Phase Lake Agassiz lagoon; *Geological Society of America North-Central Section Annual Meeting*, Abstracts with Programs, Madison, Wisconsin, p. 57-58.

Mann, J. D., Leverington, D., Rayburn, J., Grant, N., and Teller, J. T., 1997; Calculating the volume and heat budget of glacial Lake Agassiz; *Geological Society of America Annual Meeting*, Abstracts with Programs, Salt Lake City, Utah, p. A-111.

Mann, J. D., Leverington, D. W., Rayburn, J., and Teller, J. T., (1998 in press), The volume and paleobathymetry of glacial Lake Agassiz; *Journal of Paleolimnology.*

Matile, G.L.D., and Thorleifson, L.H., 1996, Prairie NATMAP: Evidence for Lake Agassiz lake level fluctuations from southeastern Manitoba; *In:* Geological Association of Canada/Mineralogical Association of Canada Joint Annual Meeting - 1996, Program with Abstracts.

Meyboom, P.,1967, Interior plains hydrogeological region; *In:* Groundwater in Canada, Economic Geology Report, Geological Survey of Canada, **24**, p. 131-158.

Moran, S. R., 1969, Geology of the Hudson Bay area, Saskatchewan; PhD. Thesis, *University of Illinois at Urbana Champaign*, 190 pp.

Moran, S. R., Clayton, L., and Cvancara, A. M., 1971, New sedimentological and paleontological evidence for the history of Lake Agassiz: Snake Curve section, Red Lake County, Minnesota; *Proceedings of the North Dakota Academy of Science*, **24** (part 2), p. 61-73.

Mott, R. J., 1971, Radiocarbon dates from Saskatchewan; *Geological Survey of Canada Paper 71-1 (B)*, p. 126-128.

Nambudiri, E. M. V., Teller, J. T., and Last, W. M., 1980, Pre-Quaternary microfossils-a guide to errors in radiocarbon dating; *Geology*, **8**, p.123-126.

Nielsen, E., 1988, Surficial geology of the Swan River area; *Manitoba Energy and Mines Geological Services Report GR 80-7*, 51 pp.

Nielsen, E., and Watson, D. M., 1985, Stratigraphy and age of landslides along the Porcupine Hills; *Manitoba Energy and Mines Report of Field Activities*, p. 232-234.

Nielsen, E., Gryba, E. M., and Wilson, M. C., 1984, Bison remains from a Lake Agassiz spit complex in the Swan River valley, Manitoba: depositional environment and paleoecological implications; *Canadian Journal of Earth Sciences*, **21**, p. 829-842.

Prest, V.K., 1970, Quaternary geology of Canada; *In:* Douglas, R.J.W. (ed.), *Geology and economic minerals of Canada*, Geological Survey of Canada Economic Geology Series, No. 1, 5th ed., p. 676-758.

Rayburn, J., 1997, Correlation of the Campbell strandlines along the northwestern margin of glacial Lake Agassiz; Msc. Thesis, *University of Manitoba*, 189 p.

Rayburn, J., Mann, J. D., and Teller, J. T., 1997, Using Lake Agassiz beaches to re-evaluate isobases along the western Laurentide Ice Sheet; *In: Geological Society of America North-Central Section Annual Meeting*, Program with Abstracts, p. 67-68.

Risberg, J, Matile, G, and Teller, J.T., 1995, Lake Agassiz water level changes as recorded by sediments and their diatoms in a core from southeastern Manitoba, Canada; *Pact 50*, II.1, p. 85-96.

Ritchie, J. C., 1964, The Riding Mountain area; *Canadian Journal of Botany*, **42**, p. 181-197.

Ritchie, J. C., 1969, Absolute pollen frequencies and carbon-14 age of a section of Holocene Lake sediment from the Riding Mountain area of Manitoba; *Canadian Journal of Botany*, **47**, p. 1345-1349.

Ritchie, J. C., 1983, Paleoecology of the central and northern parts of the glacial Lake Agassiz basin; *In:* Teller, J.T., and Clayton, Lee (eds.), *Glacial Lake Agassiz*, Geological Association of Canada Special Paper 26, p. 157-170.

Rittenhouse, G., 1934, A laboratory study of an unusual series of varved clays from northern Ontario; *American Journal of Science*, **28**, p. 110-120.

Schreiner, B. T., 1983, Lake Agassiz in Saskatchewan; *In:* Teller, J.T., and Clayton, Lee (eds.), *Glacial Lake Agassiz*, Geological Association of Canada Special Paper 26, p. 75-96.

Schreiner, B. T., 1984, Quaternary geology of the Precambrian shield, Saskatchewan; *Saskatchewan Geological Survey,* Report 221, 106 pp.

Schreiner, B. T., and Alley, D. W., 1975, Quaternary geology, Lac La Ronge area (73P) Saskatchewan; *Saskatchewan Research Council, Preliminary Map No. 1,* marginal notes.

Simpson, M. A., 1988, Surficial geology of the Ile A La Crosse area (73O) Saskatchewan; *Saskatchewan Research Council Sedimentary Resources*, 1:250,000 Map.

Smith, A. J., 1993, Lacustrine ostracodes as hydrochemical indicators in lakes of the north-central United States; *Journal of Paleolimnology*, **8**, p. 121-134.

Smith, D. G., 1992, Vibracoring: recent innovations; *Journal of Paleolimnology*, **7**, 137-143.

Smith, D. G., and Fisher, T. G., 1993, Glacial Lake Agassiz: the northwestern outlet and paleoflood; *Geology*, **21**, p. 9-12.

Stonehouse, H. B., and Ellis, J. G., 1983, The soils of the Hudson Bay-Swan Lake map areas: 63D and 63C Saskatchewan; *Saskatchewan Institute of Pedology Publication S5*, 190 pp.

Taylor, L. C., 1992, The response of spring-dwelling ostracodes to intra-regional differences in groundwater chemistry associated with road salt practices in southern Ontario: a test using an urban-rural transect; Msc. Thesis, *University of Toronto.*

Teller, J.T., 1976, Lake Agassiz deposits in the main offshore basin of southern Manitoba; *Canadian Journal of Earth Sciences*, **13**, p. 27-43.

Teller, J.T., 1985, Glacial Lake Agassiz and its influence on the Great Lakes; *In:* Karrow, P.F., and Calkin, P.E. (eds.), *Quaternary evolution of the Great Lakes*, Geological Association of Canada Special Paper 30, p. 1-16.

Teller, J.T., 1987, Proglacial lakes and the southern margin of the Laurentide Ice Sheet; *In* Ruddiman, W.F., and Wright, H.E., Jr. (eds.), *North America and adjacent oceans during the last deglaciation*, Geological Society of America, The Geology of North America **K-3**, p. 39-69.

Teller, J.T., 1988, Lake Agassiz and its contribution to flow trough the Ottawa-St. Lawrence system; *In:* Gadd, N.R. (ed.), *The late Quaternary development of the Champlain Sea Basin*, Geological Association of Canada Special Paper 35, p. 281-289.

Teller, J. T., 1989, Importance of the Rossendale site in establishing a deglacial chronology along the southwestern margin of the Laurentide Ice Sheet; *Quaternary Research*, **32**, p. 12-23.

Teller, J.T., 1990a, Volume and routing of late glacial runoff from the southern Laurentide Ice Sheet; *Quaternary Research*, **34**, p. 12-23.

Teller, J.T., 1990b, Meltwater and precipitation runoff to the north Atlantic, Arctic, and Gulf of Mexico from the Laurentide Ice Sheet and adjacent regions during the Younger Dryas; *Paleoceanography*, **5**, p. 897-905.

Teller, J. T., and Last, W. M., 1979, Postglacial sedimentation and history in Lake Manitoba, Canada; *Manitoba Department of Mines, Natural Resources and Environment*, Report 79-41, 185 pp.

Teller, J. T., and Last, W. M., 1981, Late Quaternary history of Lake Manitoba, Canada; *Quaternary Research*, **16**, p. 97-116.

Teller, J. T., and Last, W. M., 1982, Pedogenic horizons in lake sediments; *Earth Surface Processes and Landforms*, **7**, p. 367-379.

Teller, J.T., and Bluemle, J.P., 1983, Geological setting of the Lake Agassiz region; *In:* Teller, J.T., and Clayton, Lee (eds.), *Glacial Lake Agassiz*, Geological Association of Canada Special Paper 26, p. 7-20.

Teller, J.T., and Thorleifson, L.H., 1983, The Lake Agassiz-Lake Superior connection; *In:* Teller, J.T., and Clayton, Lee (eds.), *Glacial Lake Agassiz*, Geological Association of Canada Special Paper 26, p. 261-290.

Teller, J. T., and Mahnic, P., 1988, History of sedimentation in the northwestern Lake Superior basin and its relation to Lake Agassiz overflow; *Canadian Journal of Earth Sciences*, **25**, p. 1660-1673.

Teller, J. T., Matile, G., Risberg, J., and Zoltai, S., 1996, Evolution of a small lake basin associated with the Upper Campbell beach of Lake Agassiz in southeastern Manitoba (prairie NATMAP); *Geological Association of Canada/Mineralogical Association of Canada Joint Annual Meeting*, 1996 Program With Abstracts, p. A-95.

Teller, J.T., Risberg, J., Matile, G., Sandgren, P., and Zoltai, S., 1998 (submitted), Postglacial history and paleoecology of Wampum, Manitoba, a former lagoon in the Lake Agassiz basin; *Geological Society of America Bulletin*.

Thompson, T. A., Miller C. S., Doss, P. K., Thompson, L., and Baedke, S., 1991, Land-based vibracore analysis: tips, tricks, traps; *State of Indiana Geological Survey Occasional Paper 58*, 13 pp.

Thorleifson, L. H., 1983, The eastern outlets of Lake Agassiz; Msc. Thesis, *University of Manitoba*, 87 pp.

Thorleifson, L.H., 1996, Review of Lake Agassiz history; *In:*Teller, J.T., Thorleifson, L.H., Matile, G., and Brisbin, W.C., (eds.), *Sedimentology, geomorphology, and history of the central Lake Agassiz basin*, Geological Association of Canada Field Trip Guidebook for GAC/MAC Joint Annual Meeting - 1996, p. 55-84.

Upham, W., 1895, The glacial Lake Agassiz; Monograph - United States Geological Survey, Monograph 25, 658 pp.

Vance, R., Clague, J., and Mathewes, R., 1993, Holocene paleohydrology of a hypersaline lake in Alberta; *Journal of Paleolimnology*, **8**, p. 103-120.

Van Everdingen, R. O., 1971, Surface water composition in southern Manitoba reflecting discharge of saline subsurface waters and subsurface dissolution of evaporites; *In:* Turnock, A. C. (ed.), *Geoscience Studies in Manitoba*, Geological Association of Canada Special Paper 9.

Vincent, J-S., and Hardy, L., 1979, The evolution of glacial Lakes Barlow and Ojibway, Quebec and Ontario; *Geological Survey of Canada Bulletin*, **316**, 18 pp.

Vreeken, W. J., 1981, Distribution and chronology of freshwater marls between Kingston and Belleville, Ontario; *Canadian Journal of Earth Sciences*, **18**, p. 1228-1239.

Walcott, R.I., 1970, Isostatic response to loading of the crust in Canada; *Canadian Journal of Earth Sciences*, **7**, p. 716-726.

Warman, T. A., 1991, Sedimentology and history of deglaciation in the Dryden, Ontario area, and their bearing on the history of Lake Agassiz; Msc.Thesis, *University of Manitoba*, 256 pp.

Winter, T. C., and Woo, M. K., 1989, Hydrology of lakes and wetlands; *In:* Wolman, M. G., and Riggs, H. C., (eds.), *Surface Water Hydrology;* Geological Society of America, The Geology of North America **O-1**, p. 159-188.

Xia, J., Haskell, D. R., Engstrom, D. R., and Ito, E., 1997, Holocene climate reconstructions from tandem trace element and stable isotope composition of ostracodes from Coldwater Lake, North Dakota, U.S.A., *Journal of Paleolimnology*, **17**, p. 85-100.

Young, R. A., and Mansue, L. J., 1979, Relations between Holocene climatic variations and sedimentation rates in western New York state; 22nd Conference on Great

Lakes Research, *International Association for Great Lakes Research*, Rochester, NY.

Zoltai, S.C.,1961, Glacial history of part of northwestern Ontario; *Geological Association of Canada Proceedings*, **13**, p. 61-83.

Zoltai, S.C., 1963, Glacial features of the Canadian Lakehead area; *Canadian Geographer*, **7**, p. 101-115.

Zoltai, S.C., 1965, Glacial features of the Quetico-Nipigon area, Ontario; *Canadian Journal of Earth Sciences*, **2**, p. 247-269.

APPENDICIES

Appendix A: Core Descriptions

Appendix A contains detailed systematic descriptions of all stratigraphic units for all cores collected during this study. Locations of the lakes along the western margin of the Lake Agassiz basin are shown in Figure 4-1. Locations of all cores within individual lakes are shown in Figure 2-6, as well as in Figures 4-2, 4-6 and 4-9, 4-10, and 4-15 for Brokenpipe Lake, Ruby Lake and surrounding area, Jay Jay Lake, and Gregory Lake, respectively. Core lengths are uncorrected for compaction due to vibracoring.

Brokenpipe Lake, Manitoba

1) Core BPA

Water depth ~ 1.6 m, total core length = 90 cm. Core BPA was extruded in the field on the ice, and described there.

0 - 12.5 cm
Organics, high water content, undifferentiated

12.5 - 27.5 cm
Marl, contains abundant *Chara sp.* coarse fragments, molluscs abundant, highly calcareous

27.5 - 45 cm
Marl, grading downward into a peaty marl, boundary is clearest at 45 cm

45 - 57.5 cm
Peat, fine humified material, few fibers greater than 1 cm in length, molluscs abundant

57.5 - 82.5 cm
Peat, slight fg - mg sand content, molluscs abundant

82.5 - 90 cm

Peaty gravel, 0.5 - 1 cm sized subrounded clasts with a mg - cg sand matrix, molluscs less apparent

2) Core BPB

Water depth ~ 1.6 m, total core length = 310 cm.

0 - 93 cm

Marl, contains abundant *Chara sp.* fragments, molluscs throughout, grades from a marl at the top of the unit to a darker peaty marl over the 55 - 93 cm interval, highly calcareous, lighter colored more marly band in the peaty material at 90 - 93 cm, varies in color from 2.5 Y 6/2 (light brownish grey) in the upper portion of the unit to 10 YR 2/1 (black) in the lower peaty marl, lower boundary distinct

93 - 155 cm

Peat (100 - 125 cm), humified, pasty texture rather than fibrous, no molluscs apparent in the upper portion of the unit, grades downward over the 125 - 155 cm interval into a marly peat that contains molluscs, varies in color from 7.5 YR 2/0 (black) to 2.5 Y 3/2 (very dark greyish brown), lower boundary gradational over 2 cm

155 - 212 cm

Marl, contains abundant *Chara sp.* fragments, finer fragment sizes than in the 0 - 93 cm unit, molluscs present, highly calcareous, 5 Y 5/1 grey, lower boundary gradational over 4 cm

212 - 254 cm

Clayey silt to silty clay, laminated, light laminae (0.5 - 1 mm) are silty, dark laminae are organic-rich, highly organic laminae at 226 cm (2 mm thick), 235 cm, 240 cm, 244 cm, 5 Y 4/1 (dark grey), lower boundary distinct

254 - 285 cm

Silty clay to clayey silt, rhythmically laminated in 0.5 - 2 cm units, calcareous, vfg sand lenses at 275 cm and 278 cm, 2.5 Y 4/2 (dark greyish brown) to 2.5 Y 3/2 (very dark greyish brown), lower boundary distinct

285 - 287 cm

Fg sand, well sorted, all quartz, lower boundary distinct

287 - 299 cm

Silty clay to clayey silt, well laminated, 3 mm thick vegetal detritus layer at 293 cm, silty laminae 5 mm in size at 290 cm and 292 cm, silt is highly calcareous, clay is less so, 2.5 Y 3/2 (very dark greyish brown), lower boundary distinct

299 - 305 cm

Clayey silt, non-laminated, 5 Y 2/1 (black), lower boundary distinct

305 - 307 cm

Fg sand, well sorted, basal contact distinct

307 - 310 cm

Vegetal detritus, slightly silty, several 1 - 2mm thick silty laminae

3) Core BPC

Water depth ~ 1.6 m, total core length = 335 cm

0 - 90 cm

Marl, contains abundant *Chara sp.* fragments, grading downward to calcareous fine peat (marly peat), abundant molluscs, most calcareous zones are gritty with calcareous fragments and are lighter colored 10 YR 5/4 (yellowish brown), darker zones less calcareous and contain many fibers (more 'peaty'), 10 YR 3/2 (dark yellow brown), lightest colored zone is 0 - 20 cm, gradually becoming dark downward, 50 - 60 cm and 78 - 83 cm are somewhat lighter (perhaps more marl content), non-laminated, basal contact is very gradational

90 - 140 cm

Fine peat at 90 – 93 cm, grading downward into a marly peat to a peaty marl, occasional molluscs, non-laminated, 10YR 2/1 (black), lower boundary gradational over 15 cm, AMS radiocarbon date from wood fragment at 90 cm of 5980 +/- 70 BP [TO-6205]

140 - 196 cm

Marl, finer *Chara sp.* fragment sizes than seen in 0 - 20 cm of the core, few molluscs visible except in upper 5 cm, 10 YR 4/1 (dark grey) except upper part is slightly darker (more peaty material there), non-laminated, lower boundary gradational over 3 cm

196 - 230 cm
Clayey silt to silty clay, laminated except in upper 5 cm, laminae are 0.5 - 1 mm in size, alternating light (silty) and dark (organic material +/- clay), thicker laminae are 1 - 3 mm and are dominantly organic material (especially at 207, 214, and zone at 219 - 230 cm), wood fragment at 203 cm, 10 YR 3/1 (very dark grey) to 5 Y 2/2 (black), calcareous, lower boundary very gradational, AMS radiocarbon wood fragment date from 203 cm of 9660 +/- 90 BP [TO-6206]

230 - 262 cm
Silty clay to clayey silt, laminated in 0.5 - 2 cm thick units, calcareous, no organic laminae, 5 Y 4/1 (dark grey) to 5 Y 2/1 (black), lower boundary distinct

262 - 265 cm
Sand, fg, well sorted, mainly quartz, calcareous, lower boundary distinct

265 - 281 cm
Silty clay to clayey silt, contains vegetal detritus, well laminated, increasing numbers of vegetal laminae (1 - 3 mm thick) downward with a high percentage of vegetal laminae in the 275 - 280 cm interval, one 2 mm laminae of silt at 276 cm, clay is weakly calcareous, silt is very calcareous, 10 YR 3/1 (very dark grey) to 5 Y 2/2 (black), lower boundary distinct

281 - 285 cm
Clayey silt, non-laminated, calcareous, 5 Y 2/2 (black), lower boundary distinct

285 - 289 cm
Sand, fg, well sorted, calcareous, several laminae of clayey silt (up to 0.5 cm thick), distinct basal contact

289 - 291 cm
Vegetal detritus, slightly silty, some silty laminae, 5 YR 2/1 (black), lower contact distinct

291 - 295 cm
Silty clay to clayey silt, vegetal detritus, finely laminated (1 - 2 mm thick), occasional wood fragments at 293 cm 5 YR 2/1 (black), distinct lower contact, AMS radiocarbon wood fragment date from 293 cm of 9350 +/- 70 BP [TO-5880]

295 - 299 cm

Sandy pebbly gravel, slightly silty, subrounded to rounded clasts, largest clast has a long axis of 7 cm, calcareous, distinct lower contact

299 - 335 cm

Clay, occasional granules and pebbles (up to 2 cm diameter), very compact, except for floating grains clay is very smooth, 5 Y 3/2 (dark olive grey)

4) **Core BPN** (from small lake north of Brokenpipe Lake)

Water depth ~1.3 m, total core length = 210 cm.

0 - 130 cm

Peat, fine to medium fibrous, several cm-sized fibres, molluscs throughout, 5 YR 2/1 (black), bottom contact gradational over 1 cm; 0 - 10 cm interval is slightly lighter in color, and is a marly peat -- this is probably a gradation from the underlying peat to an overlying marl, elements of a marl containing *Chara sp.* fragments were dragged down the sides of the core tube in the upper 10 cm of the core, but no true marl was recovered

130 - 181 cm

Gravelly silt, massive, upper 15 cm of unit grades downward from containing molluscs and peat fibres, to a gravelly silt below, clasts are 2 - 5 mm in size, subrounded, and in a vcg - cg sand matrix, 5 Y 3/2 (dark olive grey), bottom contact distinct

181 - 210 cm

Gravel, massive, bottom 15 cm of unit is all vcg sand sized pebbles and granule matrix with extremely well rounded clasts generally 1 - 1.5 cm, largest clasts in this interval measure 7 X 5 cm and 5 X 3 cm, 181 - 195 cm interval has clasts that are generally 1 - 1.5 cm in size, silt is also present

Hudson Bay Area, Saskatchewan

1) **Core HBU**

Water depth ~ 1.3 m, total core length = 73 cm.

0 - 20 cm

Vf fibrous peat, pasty texture, no molluscs apparent, 10 YR 2/1 (black), lower contact gradational over 5 cm

20 - 25 cm
Vfg sandy silty clay, massive, with occasional mm-sized granules, dark in color due to organic (peat) content, some irregular lighter silty areas 0.5 cm in size (silt balls?), 5 Y 2/1 (black), bottom contact irregular and smeared over 2 cm

25 - 73 cm
Very silty clay with blocky structure near bottom of unit, smoother near top of unit, with many mm-sized granules, cm-sized stones noted at 26 cm, 39 cm, and 55 cm, gradational upper contact smeared with organic matter over 5 cm, unit has several pure silt balls, and a pure silt vertical slit starting at 35 cm and ending at 64 cm, which may reflect a root trace, or dried out cracked clays carrying groundwater, clayey part is 5 Y 4/1 (dark grey), silty part is 5 Y 5/3 (olive)

2) Core HBL
Water depth ~ 1.1 m, total core length = 190 cm.

0 - 86 cm
Vf peat, with occasional cm-long fibres, mollusc shells present throughout, 7.5 YR 2/0 (black), bottom contact distinct

86 - 110 cm
Fg - mg sand, occasional 2 - 5 mm sized pebbles, dark in color due to mixing of peat throughout unit, 5 Y 3/2 (dark olive grey) to 5 Y 2/2 (black) in color, lower boundary distinct

110 - 145 cm
Gravel, sub to well rounded clasts commonly 3 mm to 3+ cm in size, vcg sand matrix, bottom contact distinct

145 - 168 cm
Vfg sandy silt, poorly laminated, except for distinct alternating light (2 - 3mm thick, silt) and dark (3 mm thick, more sandy), laminae in upper 10 cm of unit, 10 YR 2/1 (black), basal contact distinct

168 - 190 cm
Silty clay, with 2 mm to 1 cm sized clasts of weathered shale, unit becomes more compact towards the base, mm-sized granules common, 1 cm sized pebbles at 172 cm (near core tube edge) and 175 cm, 2.5 Y 2/0 (black)

Ruby Lake, Saskatchewan

1) Core RLA

Water depth ~ 3.66 m, total core length = 123 cm.

0 - 1 cm

Mg - fg peaty sand, top of unit has a high % of peat, 5 Y 3/1 (very dark grey), lower contact distinct

1 - 12 cm

Gravel with peat beds, upper 4 cm of unit is vfg sandy peat, peat is vf fibrous (almost pasty), 4 - 9 cm is gravel with one 4 mm peat laminae in the middle (peat as above), clasts are subrounded, several mm to 1 cm in size, with vcg sand matrix, 9 - 12 cm is peat (as above), unit is dark in color due to peat content, 5 Y 3/2 (dark olive grey), basal contact distinct

12 - 56 cm

Base of unit (40 - 56 cm) is vcg sand with frequent mm-sized granules and occasional cm-sized clasts, grading upward to (25 - 40 cm) gravel, mostly mm-sized granules and vcg sand matrix with several cm-sized, subrounded clasts, grading upward (12 - 25 cm) into gravel, clasts are subrounded, generally 0.5 - 1 cm in size (one is 4 cm in diameter), vcg sand matrix, color varies from 5 Y 4/2 (olive grey) at the base, to 5 Y 6/1 (grey) toward the top of the unit, color change denotes the change in gravel clast size and relative framework content versus matrix content in the gravels (lighter color = relatively more framework vs matrix, and larger clast size), lower contact distinct

56 - 78 cm

Mg - cg sand with occasional mm - sized granules, very poorly bedded with gravel layers at 77 - 78 cm, 69 - 71 cm , and 57 - 59 cm, gravel has well rounded 0.5 - 1 cm sized clasts in a vcg sand and mm-sized granule matrix, there are two 3 mm thick vfg sandy silt laminae at 56 and 65 cm, 5 Y 3/2 (dark olive green), lower contact distinct

78 - 95 cm

Silty vfg sand with a gravel layer at 80 - 84 cm, gravel clasts are 0.5 cm in size, subrounded, matrix is vcg sand, two vfg sandy silt layers 3 - 4 mm in thickness are found at 89 and 92 cm, unit has a yellow - orange staining which may reflect surface and ground water flow through the unit during a period oxidation and sub-aerial exposure, color is 2.5 Y 4/4 (olive brown) at the base to 5 Y 5/4 (olive) toward the top

95 - 98 cm
Vfg sandy silt, with one 3 mm thick vegetal detritus layer at 97 cm, 5 Y 3/2 (dark olive grey)

98 - 123 cm
Silty clay, contains weathered shale clasts, one 1 cm diameter stone found near core tube edge in the clay at 102 cm, matrix contains frequent mm - sized granules, 5 Y 3/2 (dark olive grey)

2) Core RLB
Water depth ~ 2.8 m, total core length = 31 cm.

Ruby Lake Core B had a length of approximately 31 cm, it consisted entirely of material as described in the 98 - 123 cm interval of the Ruby Lake A core, high water content sediments from the water - lake bottom interface were not recovered

Jay Jay Lake, Saskatchewan

1) Core JJA
Water depth ~ 1.7 m, total core length = 188 cm.

0 - 19 cm
Peat, pasty texture rather than fibrous, no molluscs apparent, 7.5 YR 2/0 (black), lower boundary distinct

19 - 100 cm
Fg - mg sand, upper 8 cm stained from peat and contains mollusc shells, 1 cm-sized stone at 58 cm, silty laminae 2 mm thick at 73 cm, massive, occasional mm-sized granules, moderately sorted, lower boundary distinct

100 - 188 cm
Clayey silt, cm-sized stone at 122 cm, upper 10 cm of unit has several cm-scale fg-mg moderately sorted sand laminae, vfg sandy silt lenses at 125 - 135 cm and 145 - 148 cm, mg clean well sorted sand layer at 165 - 167 cm (badly disturbed due to vibracoring operations), small mm-scale fg-mg silty sand layers at 155 cm, 160 cm, 170 cm, 175 cm, 179, and 182 cm, overall color 5 Y 4/1 (dark grey)

2) Core JJB

Water depth ~ 1.7 m, total core length = 300 cm.

0 - 5 cm

Peat, non-fibrous, no molluscs apparent, 10 YR 2/1 (black), lower boundary distinct

5 - 18 cm

Fg-mg sand, moderately sorted, massive, has considerable organic (gyttja) content at the top of the unit that decreases to <5% at the bottom, 2.5 Y 3/2 (very dark greyish brown), lower boundary distinct

18 - 71 cm

Fg-mg sand, massive, 0.5-1 cm sized stones common, oxidized zone at 25 - 40 cm 10 YR 3/3, (several clayey silt laminae 3 mm to 1 cm thick in the 25 - 40 cm zone, lower boundary distinct

71 - 95 cm

Fg-mg sand, poorly sorted, has some silt content, 2 mm thick clayey silt laminae are present at 74 cm and 76 cm, the lower 10 cm of the unit shows poor bedding (2 cm thick), the rest of the unit is massive, lower boundary distinct

95 - 300 cm

Clayey silt, 5 Y 4/1 (dark grey), punctuated by zones and lenses of well sorted mg-cg sands with occasional mm-sized granules at 95 - 97, 255 - 257, 258 - 259, 261 - 262, 264 - 265 and 283 - 284 cm, and zones of vfg-mg moderately sorted sand at 99 - 100, 105 - 118, 122 - 125, 139 - 140, 143 - 144, 149 - 152, 154 - 155, 169 - 170, 174 - 175, 177 - 178, 182 - 192, 227 - 229, 242 - 244 and 290 - 300 cm

3) Core JJC

Water depth ~ 1.8 m, total core length = 400 cm.

0 - 10 cm

Fine organic sediment (humified peat), fine rootlets common, no laminae visible, slightly calcareous in lower part with increasing lighter coloring (like marl below), occasional molluscs (bivalves and snails) shells, 10 YR 2/1 (black), lower boundary gradational over 3 cm

10 - 56 cm

Marl/gyttja, organic sediment that becomes increasingly calcareous downward, no laminae visible, gastropods common, seeds, no siliciclasitic grit, 10 YR 7/1 (light grey), lower boundary distinct but gradational over 1 cm

56 - 75 cm
Very clayey silt with coarse sand, no laminae visible, non-calcareous, 10YR 3/1 (very dark grey), grading downward into clayey silt with finer sand

75 - 81 cm
Slightly clayey and sandy silt, no visible laminae, very few coarse sand grains, non-calcareous, 10 YR 3/2 (very dark greyish brown), lower boundary gradational

81 - 121 cm
Sand, vfg - mg, silty, laminated, non-calcareous, 10YR 4/1 (dark grey)

121 - 312 cm
Silt to clayey silt and sand (vfg - mg), bedded and laminated, zones of siltier and sandier laminae vary from 1 to 30 cm (no systematic variation), 10 YR 5/1 (grey) to 10 YR 4/1 (dark grey), lower boundary distinct

312 - 332 cm
Silt to very fine sand, poorly laminated, 10 YR 5/1 (grey), lower boundary distinct

332 - 336 cm
Clayey silt, 2 mm organic-rich laminae at top and occasional thinner organic laminae below, 10 YR 5/1 (grey), lower boundary distinct

336 - 350 cm
Clayey silt, more clayey than overlying unit, laminae (1 - 2 mm) of vfg - mg sand throughout, 10 YR 5/1 (grey), lower boundary distinct

350 - 391 cm
Sand (fg - mg) and silty vfg sand, laminated, zones of silty versus sandy laminae <0.5 - 4 cm, coarsest laminae in lower 6 cm, rare organic fibers, 2.5 Y 5/2 (greyish brown), lower boundary distinct

391 - 400 cm
Silty fine sand, laminated, 2.5 Y 4/2 (dark greyish brown)

4) Core JJD

Water depth ~ 1.8 m, total core length = 468 cm.

0 - 22 cm

Peat, non-fibrous, humified pasty texture, molluscs present in lower 8 cm of unit, 2.5 Y 2/0 (black), lower boundary gradational to more peaty gyttja over 1 cm

22 - 97 cm

Gyttja, coarser in the upper 40 cm of the unit, molluscs present, upper 10 cm of unit is darker colored (some peat content), interval over 55 - 95 cm has a peat-filled crack (probably a root trace), 5 Y 6/1 (grey), lower boundary distinct

97 - 129 cm

Fg sandy clayey silt, massive, fg sandy silt zone 100 - 110 cm, 5 Y 3/1 (very dark grey)

129 - 272 cm

Fg-mg moderately sorted sand, fg sandy silt zones at 140 - 142, 145 - 147, and 149 - 155 cm, cm-scale fg sandy silt layers at 210, 235, 243, and 261 cm, fg-cg poorly sorted sand layers and lenses with occasional mm-sized granules at 170 - 175, 181 - 188, 198 - 208, 215 - 230 (which is poorly bedded), 245 - 250, and 265 - 272cm, overall 5 Y 4/1 (dark grey), lower boundary distinct

272 - 468 cm

Vfg sandy silt to clayey silt, cm-sized pebble at 357 cm, sporadic pure silt beds 0.5 cm in size (good example at 338 cm), 359 - 361 cm interval is fg-mg poorly sorted sand, cm-sized lenses and laminae of fg-mg poorly sorted sands at 290, 325, 355, 400, and 435 cm, fg-cg poorly sorted sands with occasional mm-sized granules located at 410 - 412, 420 - 422, 430 - 432, 443 - 445, 448 - 450, 453 - 456, 458 - 460, and 463 - 465 cm, pure silt laminae at 380 - 390 cm, overall 2.5 Y 4/2 (dark greyish brown) to 2.5 Y 4/4 (olive brown)

Gregory Lake, Saskatchewan

1) Core GLE

Water depth ~ 1.8 m, total core length = 573 cm.

0 - 183 cm

Peat, mg - fg fibrous, uppermost 93 cm not recovered due to high water content, was increasingly gyttja like towards water - lake bottom interface, basal 20 cm of unit slightly sandy, sand content decreased upwards, 2.5 Y 2/0 (black)

183 - 198 cm
Fg - mg sandy peat, upper boundary gradational into overlying unit, 5 Y 3/1 (very dark grey), AMS bulk organic radiocarbon date from 196-198 cm of 8850 +/- 80 [TO-6893]

198 - 325 cm
Fg - mg clean sand, basal 5 cm of unit is fg - mg sand with occasional laminae of fg sandy silt, middle part of unit is poorly bedded fg - mg sands, upper 5 cm of unit is stained (5 Y 4/4 olive) with root traces over upper 14 cm of the unit, overall color is 5 Y 5/2 (olive grey), basal contact distinct

325 - 348 cm
Fg sandy silt, interval 12 - 18 cm from bottom of the unit is more sandy (fg - mg sandy silt), 5 Y 4/2 (olive grey)

348 - 398 cm
Fg - mg sand, lower 25 cm of unit has mm - sized granules, lower 15 cm of unit has cm - scale bedding, upper boundary gradational over 1 cm

398 - 423 cm
Clayey fg sandy silt, poorly to non-laminated at base of unit, grading upward to silty fg - mg sand in upper 10 cm of unit, occasional vcg sand granules, overall unit is 5 Y 4/2 (olive grey), upper 10 cm of unit is 5 Y 4/1 (dark olive grey)

423 - 469 cm
Clayey fg sandy silt at base of unit (12 cm thick) grading upward to less clayey, more sandy fg - mg silt (20 cm thick and poorly laminated), grading upward to 14 cm thick zone of silty fg-mg sand with broken laminae of silty clay, lower part of unit is 5 Y 3/2 (dark olive grey), upper part of unit is 5 Y 4/2 (olive grey), upper contact of unit is distinct

469 - 472 cm
Fg - mg clean sand, no visible laminae, distinct contact

472 - 474 cm
Clayey fg - mg sandy silt, 5 Y 4/2 (olive grey), lower boundary distinct

474 - 480 cm
Vfg sandy silt, very poorly laminated, 5 Y 5/2 (olive grey), lower boundary distinct

480 - 573 cm

Silty clay, very poorly laminated to non-laminated, cm sized dropstones near base of unit, very slightly sandy (vfg sand) in all but upper 25 cm of unit, 5 Y 3/2 (olive grey)

Appendix B: % Moisture, % Organics and % Carbonate Data

Appendix B contains percentage values calculated for moisture, organic matter, and carbonate for three cores from this study. Section 2.3.2 (Chapter 2) details the methodology for obtaining the data. Percent organic matter and carbonate were calculated from dry samples by using loss on ignition. Data is shown for the BPC core from Brokenpipe Lake, the JJC core from Jay Jay Lake, and the GLE core from Gregory Lake. The locations of these lakes along the western margin of the Lake Agassiz basin are shown in Figure 4-1. Figure 2-6 shows the aforementioned core sites within each individual lake, as do Figures 4-2, 4-10, and 4-15 for Brokenpipe Lake, Jay Jay Lake, and Gregory Lake respectively. Figures 4-5, 4-14, and 4-17 show this percentage data plotted as curves (alongside a representative stratigraphic section) for the BPC (Brokenpipe Lake), JJC (Jay Jay Lake), and GLE (Gregory Lake) cores respectively.

BPC DATA

Depth (cm)	% Moisture	% Organics	% Carbonate
0	65.39	11.35	83.48
10	64.61	11.42	81
25	65.41	14.87	73.92
40	67.99	21.99	60.56
55	62.78	18.97	62.86
70	62.16	24.42	50.57
85	57.95	20.66	34.29
90	55.43	19.23	32.6
95	49.24	20.99	8.71
110	43.96	18.39	14.16
123	45.15	17.72	27.91
138	44.85	13.59	45.67
142	44.96	11.3	75.29
150	45.41	9.63	72.83
161	47.21	7.45	75.36
176	43.56	3.6	51.99
185	46.84	6.72	53.89
191	46.92	7.1	57.27
197	28.41	6.25	26.59
200	22.43	5.22	16.11
210	22.5	4.91	21.85
215	24.92	6.2	7.5
230	25.25	6.16	27.87
235	35.26	8.2	4.61
245	27.75	6.64	7.38
255	26.54	6.93	19.68
259	26.9	5.44	19.25
263	11.66	2.64	11.15
265	26.8	5.07	26.94
275	28.27	7.84	13.28
283	32.25	5.84	19.05
286	21.54	4.53	15.73
290	44.4	23.46	16.48
293	26.23	7.68	29.65
304	11.58	2.84	26.51
310	15.84	3.95	35.07
316	16.31	3.73	33.1
322	16.43	4.17	35.01
328	15.57	3.4	21.98
334	15.26	3.55	23.84

JJC DATA

Depth (cm)	% Moisture	% Organics	% Carbonate
0	73.19	46.88	13.92
10	63.59	19.9	30.44
20	63.65	13.54	33.65
30	60.66	11.01	33.84
40	57.92	10.07	33.89
50	51.93	8.57	27.86
60	21.21	2.64	2.25
70	13.48	1.33	2.1
80	8.89	0.27	0.63
90	8.95	0.23	0.48
100	7.63	0.46	1.28
110	3.54	0.26	0.67
120	11.6	0.29	0.96
125	14.66	1.26	4.45
130	7.59	0.39	1.28
140	2.59	0.15	0.6
150	9.34	0.58	3.15
160	2.43	0.12	0.59
165	12.63	1.18	3.72
170	3.74	0.21	0.68
180	12.42	1.06	2.44
185	16.11	1.93	3.84
190	13.45	1.43	2.94
200	11.37	0.7	1.51
210	18.07	2.06	4.22
220	17.61	1.9	3.7
230	11.44	0.79	2.29
240	17.69	1.56	3.25
250	18.8	1.66	4.57
260	2.39	0.18	0.3
270	15.39	0.27	1
280	19.63	1.94	5.07
290	12.05	0.61	1.61
300	7.66	0.21	0.85
310	5.38	0.28	0.68
320	16.37	0.63	2.94
330	15.67	0.92	2.61
340	19.39	2.2	3.64
350	17.41	1.83	3.52
360	16.16	1.63	3.31
370	11.91	0.74	2.05
380	2.03	0.19	0.25
390	8.21	0.38	1.15
400	15.3	0.96	2.58

GLE DATA

Depth (cm)	%Moisture	% Organics	% Carbonate
94	85.21	74.49	0.45
109	84.85	75.04	0.61
124	84.46	78.81	0.66
139	82.97	85.40	1.11
154	81.27	83.76	1.54
169	78.80	59.61	1.04
184	42.66	12.13	0.13
190	38.58	10.50	0.12
199	15.13	2.76	0.08
214	8.79	0.44	0.03
229	9.68	0.29	0.05
244	11.70	0.26	0.01
259	10.61	0.23	0.01
274	9.65	0.26	0.02
289	13.44	0.51	0.02
304	8.13	0.24	0.02
319	14.25	0.65	0.05
334	17.41	3.47	0.19
344	17.29	2.32	0.24
349	10.45	0.46	0.04
364	5.19	0.38	0.04
379	5.20	0.46	0.08
420	6.08	0.51	0.08
435	1.13	4.30	0.18
450	0.74	0.90	0.03
465	3.83	4.23	0.10
478	17.18	4.39	0.25
483	5.84	6.48	0.18
498	21.49	4.63	0.34
513	23.69	4.84	0.33
528	19.67	3.92	0.39
543	23.36	5.43	0.35
558	21.54	6.14	0.35
568	20.74	5.11	0.31
573	18.65	5.13	0.38

Appendix C: Lake Water Chemistry Data

Appendix C contains raw water chemistry data for all lakes in this study. Figure 4-1 gives the location of each lake along the western margin of the Lake Agassiz basin. Figure 2-6 shows where each collection site is located within each lake. The table labeled 'DFO DATA' shows the water chemistry data from the Department of Fisheries and Oceans laboratory. Site notation in the DFO table includes a 'W' in the name, meaning 'water sample'. For example, a water sample collected from site BL2 in Bird Lake (see Fig. 2-6) is named 'BLW-2'. The 'DI BLK' and 'DWBLK' in the DFO table refer to de-ionized water and distilled water samples ('blanks') which were run with the lake samples as a comparison. The table labeled 'HYDROLAB FIELD DATA' shows the water data collected with the portable, calibrated Hydrolab unit, while in the field. Figures 3-4, 3-5, 3-6, and 3-7 display the water chemistry data graphically.

DFO DATA

SITE	DIC um/l	DOC um/l	SiO2 mg/l	Na mg/l	K mg/l	Ca mg/l	Mg mg/l	Cl mg/l	SO4 mg/l	Cond uS/c	pH	ALK ueq/l	OrgAcid ueq/l
BLW-2	440	1450	0.448	1.51	0.92	7.04	1.71	0.18	0.09	47	7.85	498	81
BLW-3	450	1340	0.494	1.48	0.89	6.09	1.75	0.15	0.09	47	7.84	454	81
JJW-1	1920	1440	4.07	3.95	1.7	28.9	8.1	1.12	1.47	191	8.5	197	95
JJW-4	1930	1400	3.96	4.11	1.8	28.5	7.8	1.22	1.43	183	8.49	192	93
RLW-1	3150	1770	3.86	6.1	2.9	34	22.4	3.24	6.32	306	8.7	327	73
RLW-7	3150	1700	3.73	5.8	2.72	35.5	22.3	2.89	6.03	307	8.69	330	79
BPW-1	860	1560	0.598	37.8	7.2	36.4	18.9	16.1	142.9	502	8.21	104	1430
BPW-3	920	1570	0.639	37.9	7.2	38	20.1	15.6	141.9	502	8.22	107	1470
DI BLK	130	30	1.33	1.86	0.05	0.03	0.01	0.26	0.09	11	7.3	96	12
DW BLK	140	60	1.36	1.95	0.04	0.02	0.01	0.27	0.09	11	7.27	100	12

Ion Balances

Site	Anions ueq/l	Cations ueq/l	An-Cat	> 10% imbalance	CalCond uS/cm	Cond uS/cm	CalCond-Cond	>10% imbalance	Cal Alk ueq/l	Alk ueq/l	Cal Alk-Alk	>10% imbalance
BLW-2	517	581	-64	Yes	52	47	5	Yes	429	498	-69	Yes
BLW-3	526	535	-9		50	47	3		439	454	-15	
JJW-1	2096	2324	-228		220	191	29	Yes	1939	197	1742	Yes
JJW-4	2105	2288	-183		218	183	35	Yes	1948	192	1756	Yes
RLW-1	3512	3878	-366		379	306	73	Yes	3216	327	2889	Yes
RLW-7	3500	3927	-427	Yes	380	307	73	Yes	3214	330	2884	Yes
BPW-1	5716	5199	517		595	502	93	Yes	857	104	753	Yes
BPW-3	5781	5382	399		605	502	103	Yes	917	107	810	Yes
DI BLK	139	85	54	Yes	10	11	-1		117	96	21	Yes
DW BLK	147	88	59	yes	11	11	0		126	100	26	Yes

HYDROLAB FIELD DATA

Lake	Site	Conductivity (uS/cm)	pH	Temperature (^{o}C)
Brokenpipe				
	BP1	526	8.9	22
	BP2	540	6.5	21.7
	BP3	529	9.1	22.2
	BP4	536	8.9	22.1
	BP5	619	7.3	21.3
	BP6	536	8.9	22.2
	BP7	533	8.9	22.1
Ruby				
	RL1	318	8.4	20.1
	RL2	315	8.5	20.4
	RL3	407	6.5	17.8
	RL4	458	7	16.9
	RL6	317	8.4	20.2
	RL7	318	8.2	20
	RL8	317	8.3	19.9
Jay Jay				
	JJ1	203	8	19.5
	JJ2	203	7.7	20
	JJ3	238	7.2	11.9
	JJ4	199	8.1	20.3
	JJ INFLOW	233	7.2	15.3
Bird				
	BL1	53	6.1	19.7
	BL2	52	6.7	19.9
	BL3	52	6	20.9
	BL4	49	6.5	20.2

Appendix D: Fossil Ostracodes From the BPC Core (Brokenpipe Lake)

Appendix D summarizes the fossil ostracode counts for the BPC core (Brokenpipe Lake). The location of Brokenpipe Lake along the western margin of the Lake Agassiz basin is given in Figure 4-1. The location of the BPC core within Brokenpipe Lake is given in Figures 2-6, and 4-2. The ostracode data is tabulated as numbers of adult carapaces – one carapace equals one left and one right adult shell -- (the table labeled 'BPC #'), and as relative percentages (the table labeled 'BPC %'). Ostracode species are listed as follows in the tables:

IBRA = *Ilyocypris bradyi*

CRAW = *Candona rawsoni*

LHER = *Limnocythere herricki*

CCAU = *Candona caudata*

COHI = *Candona ohioensis*

IGIB = *Ilyocypris gibba*

CAM = *Cyclocypris ampla*

CVID = *Cypridopsis vidua*

CLAC = *Cytherissa lacustris*

LIT = *Limnocythere itasca*

CINO = *Candona inopinata*

CSIG = *Candona sigmoides*

CACU = *Candona acutula*

CPARA = *Candona paraohioensis*

CDIS = *Candona distincta*

CDEC = *Candona decora*

Other than *Candona sigmoides* and *Candona inopinata*, each represented by one adult carapace in the entire BPC core, the tables do not include the occurrences of scarce ostracodes. Other scarce ostracodes in the BPC core included those species that were represented by 3 or less shells in the *entire* BPC core. These are the two *Cypris pubera* shells (one each found at 150 cm and 123 cm in the BPC core), and the three *Heterocypris glaucus* shells (one found at 316 cm, the other two at 328 cm).

Section 4.1.5 (Chapter 4) details the autecology of all ostracodes found in the BPC core. Figure 6-1 is the relative abundance diagram for the ostracodes, excluding the scarce ostracodes -- *Candona inopinata*, *Candona sigmoides*, *Cypris pubera*, and *Heterocypris glaucus*.

BPC %

DEPTH (cm)	IBRA	CRAW	LHER	CCAU	COHI	IGIB	CAM	CVID	CLAC	LIT	CINO	CSIG	CACU	CPARA	CDIS	CDEC	TOTAL
0							28.6	6.4					11.5	0.42	9.8	43.1	99.8
10							46.1	10.3					17.3		4.5	21.8	100
25							46.5	8.3					30.3	0.25	3	11.6	100
40							33.9	10.9					34.7	0.84	9.6	10	99.9
55							44.7	13.6		0.76			34.4		2.3	4.2	100
70					0.53		41.4	13.2		1.3			38.2	1.3	2.9	1.1	99.9
85					0.99		44.9	17.6		0.99			33.3	1.2	0.5	0.5	100
90					0.79	0.4	50.4	21.8		1.9			23.8	0.39		0.39	99.9
95																	0
110							76.3						23.7				100
123							25.2	10.7		1.9			62.1				99.9
138							20.5	8.5		4.5			64.5		1	1	100
142							26.8	12.5		1.7			55.6	0.28	1.7	1.4	100
150					1.4		16.5	9.7		7.2			55.9	0.7	5.4	3.2	100
161					2.8		39.5	10.2					32.7	2	4.8	8	100
176					1.1		31	5.3		1.1			37.8		10.7	13	100
185					3		60.6	13.6		3			12.1	3	4.5		99.8
191					2.1		58.9	5.8		0.52			29.6	1	2.1		100
196																	0
197																	0
200																	0
202																	0
204																	0
206							100										100
208																	0
210																	0
211																	0
213																	0
215																	0
216							66.7		33.3								100
218							100										100
220				25			75										100
222							100										100
224																	0
226																	0
228																	0
230																	0
233				1.9	17.6		31.4	5.9		5.9			37.3				100

BPC %

DEPTH (cm)	IBRA	CRAW	LHER	CCAU	COHI	IGIB	CAM	CVID	CLAC	LIT	CINO	CSIG	CACU	CPARA	CDIS	CDEC	TOTAL
235				9.5	6		34.5	1.2		10.7			38.1				100
237	2.3	14.9		8.1	4.9	3.2	41.2	6.8	5.4	1.4			11.8				100
239		46.7		6.7		13.3			33.3								100
241		25		12.5			6.2		56.3								100
243	10.7	3.6		10.7			46.4	17.9	7.1	3.6							100
245							50		50								100
247						20	40		40								100
249	13			13		23.9	30.4	4.3	15.2								99.8
251					5.9	2.9	85.3	2.9	2.9								99.9
253							68.8		31.2								100
255	36.4					9.1	54.5										100
257							25	37.5	37.5								100
259	3.4	82.8		3.4			3.4	3.4	3.4								99.8
261	14.3	17.1		20		11.4	28.6		8.6								100
263	12.5	3.1	1.6	6.3		15.5	42.2	1.6	14.1			3.1					100
265			18.2	27.3					36.3		18.2						100
267									100								100
269			2.6	48.7		2.6	17.9	10.3	10.3	7.7							100
271							100										100
275	25			5.4		7.1	58.9			3.6							100
277	16.7						66.6			16.7							100
279	1.5			3	4.5	10.4	41.8		4.5	34.3							100
281	12			10.9	6.5	22.8	23.9		4.3	19.6							100
283	7.7	1.5		29.2		18.5	29.2		12.3	1.5							99.9
285	30.9	7.1		28.6			16.7		7.1	9.5							99.9
286	25						25		50								100
289	16.7						50			33.3							100
290		28.6					57.1		14.2								99.9
291	16.6						50		16.6	16.6							99.8
293	15.8		5.3		15.8		42		15.8	5.3							100
295																	0
297																	0
304																	0
310																	0
316																	0
322																	0
328																	0
334																	0

BPC #

DEPTH (cm)	IBRA	CRAW	LHER	CCAU	COHI	IGIB	CAM	CVID	CLAC	LIT	CINO	CSIG	CACU	CPARA	CDIS	CDEC	TOTAL
0							33.5	7.5					13.5	0.5	11.5	50.5	117
10							56	12.5					21		5.5	26.5	121.5
25							92.5	16.5					60.5	0.5	6	23	199
40							40.5	13					41.5	1	11.5	12	119.5
55							59	18		1			45.5		3	5.5	132
70					1		77	24.5		2.5			71	2.5	5.5	2	186
85					2		90.5	35.5		2			67	2.5	1	1	201.5
90					1	0.5	63.5	27.5		2.5			30	0.5		0.5	126
95																	0
110							14.5						4.5				19
123							27	11.5		2			66.5				107
138							20.5	8.5		4.5			64.5		1	1	100
142							47	22		3			97.5	0.5	3	2.5	175.5
150					2		23	13.5		10			78	1	7.5	4.5	139.5
161					5		69.5	18		0			57.5	3.5	8.5	14	176
176					1.5		40.5	7		1.5			49.5		14	17	131
185					1		20	4.5		1			4	1	1.5		33
191					4		112.5	11		1			56.5	2	4		191
196																	0
197																	0
200																	0
202																	0
204																	0
206							0.5										0.5
208																	0
210																	0
211																	0
213																	0
215																	0
216							1		0.5								1.5
218							3										3
220				1.5			4.5										6
222							3										3
224																	0
226																	0
228																	0
230																	0
233				0.5	4.5		8	1.5		1.5			9.5				25.5

BPC # DEPTH (cm)	IBRA	CRAW	LHER	CCAU	COHI	IGIB	CAM	CVID	CLAC	LIT	CINO	CSIG	CACU	CPARA	CDIS	CDEC	TOTAL
235				4	2.5		14.5	0.5		4.5			16				42
237	2.5	16.5		9	5.5	3.5	45.5	7.5	6	1.5			13				110.5
239		3.5		0.5		1			2.5								7.5
241		2		1			0.5		4.5								8
243	1.5	0.5		1.5			6.5	2.5	1	0.5							14
245							2		2								4
247						0.5	1		1								2.5
249	3			3		5.5	7	1	3.5								23
251					1	0.5	14.5	0.5	0.5								17
253							5.5		2.5								8
255	2					0.5	3										5.5
257							1	1.5	1.5								4
259	1	24		1			1	1	1								29
261	2.5	3		3.5		2	5		1.5								17.5
263	4	1	0.5	2		5	13.5	0.5	4.5			1					32
265			1	1.5					2		1						5.5
267									0.5								0.5
269			0.5	9.5		0.5	3.5	2	2	1.5							19.5
271							0.5										0.5
275	7			1.5		2	16.5			1							28
277	0.5						2			0.5							3
279	0.5			1	1.5	3.5	14		1.5	11.5							33.5
281	5.5			5	3	10.5	11		2	9							46
283	2.5	0.5		9.5		6	9.5		4	0.5							32.5
285	6.5	1.5		6			3.5		1.5	2							21
286	1						1		2								4
289	0.5						1.5			1							3
290		1					2		0.5								3.5
291	0.5						1.5		0.5	0.5							3
293	1.5		0.5		1.5		4		1.5	0.5							9.5
295																	0
297																	0
304																	0
310																	0
316																	0
322																	0
328																	0
334																	0

IMAGE EVALUATION TEST TARGET (QA-3)

150mm

6"

APPLIED IMAGE . Inc
1653 East Main Street
Rochester, NY 14609 USA
Phone: 716/482-0300
Fax: 716/288-5989