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A PALEOBOTANICAL STUDY OF HOLOCENE DROUGHT FREQUENCY IN SOUTHERN
ALBERTA

by

Robert E. Vance

B.Sc. Forestry, University of Alberta, 1975

M.A. Anthropology, University of Alberta, 1979

THESIS SUBMITTED IN PARTIAL FULFILLMENT OF
THE REQUIREMENTS FOR THE DEGREE OF
DOCTOR OF PHILOSOPHY
in the Department
of
Biological Sciences

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SIMON FRASER UNIVERSITY

August 1991

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ISBN 0-315-78338-9

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Degree: Doctor of Philosophy

Title of Thesis:

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ABSTRACT

Paleobotanical study of four sediment cores from Chappice Lake, a shallow, hypersaline lake in southeastern Alberta, produced a prehistoric drought record for the northern mixed-grass prairie. Analyses of pollen and plant macrofossils, sediments, and depositional patterns of modern pollen and plant macrofossils allowed the reconstruction of prehistoric lake level changes. Nine AMS radiocarbon determinations on seeds of upland plants provide a chronology for paleohydrological events. Lake level fluctuations are interpreted as a proxy record of regional climatic change, since lake level changes are a reflection of water table fluctuations that are ultimately driven by climate. Low water stands are considered to represent periods of heightened drought frequency.

Laminated lake sediments indicate past low water stands, since laminae are products of mineral and biological precipitation in high salinity brines. Preservation of these features results from the elimination of benthic fauna during hypersaline phases and protection of the sediment surface from wind disturbance by a dense overlying brine. Seeds and pollen of *Ruppia* are abundant during laminated intervals, indicating heightened lake salinity, as this aquatic macrophyte thrives in highly saline water. *Ruppia* populations decline during periods of recharge, when laminae also disappear.

Results suggest that droughts in the prairie region are not cyclic. Rather, the study area has experienced periods of: 1) extreme drought interrupted by short moist intervals; 2) extended drought without abatement, and; 3) long intervals during which droughts were rare. From as early as 8500 to 6000 BP Chappice Lake was much smaller than present and dried completely on occasion but also experienced short-lived high water stands, suggesting a generally warm and dry but variable climatic regime. Drying of the lake has not occurred in the last 6000 years, indicating that less severe macroclimatic conditions have prevailed since early to mid-Holocene time. However, predominantly low water, high salinity lake stands

(6000–2650 BP and 1060–600 BP) reveal that significant periods of heightened drought frequency have occurred over the last six millenia. These prolonged periods of frequent drought were separated by an uninterrupted high water stand during which droughts were uncommon.

The last 600 years of lake history are best characterized as a high water phase punctuated by a series of short-lived droughts, the most recent occurring in the last decade. During the 1980's, Chappice Lake has been reduced to its lowest historic lake levels, resulting in deposition of laminae analogous to those deposited during prehistoric low water stands.

ACKNOWLEDGEMENTS

This project was completed with the help and encouragement of many colleagues and friends. I would like to thank all members of my committee, Dr. R.W. Mathewes, Dr. R.C. Brooke, Dr. K. Fladmark, and Dr. A.C.B. Roberts, for their support, interest, and encouragement throughout my tenure at Simon Fraser University. Special thanks to Dr. R.W. Mathewes for providing an ideal setting to conduct research. Freedom to pursue lines of reasoning with a guiding hand when called upon are conditions under which most research students wish to work. Dr. Mathewes offered both, as well as lab, field equipment, and seasonal support staff of high quality. I would also like to thank Dr. R.G. Baker for serving as my external examiner and Dr. R.J. Hebda and Dr. M.C. Roberts for their efforts as public examiners. Their thoughtful questions and comments improved the final product considerably.

When I first considered conducting paleoenvironmental studies on the southern Canadian prairies I was employed by the Archaeological Survey of Alberta. This organization granted me both the time and resources to carry out much of the preliminary exploratory field work. Subsequent funding was obtained from Dr. Mathewes' NSERC research grant (A3835), graduate scholarships from Simon Fraser University, and research contracts with the Atmospheric Environment Service.

A critical step in this project was core collection. Southern Alberta may lie in the chinook belt, but winters can be windy and bitterly cold at times. Two coring expeditions to Chappice Lake happened to fall during cold spells, and I thank crew members J. Albanese, G. Chalut (of the Archaeological Survey of Alberta), A. Hanley and S. Vanderburgh (of Simon Fraser University) for their endurance and patience under trying circumstances. Vibracoring equipment was kindly loaned by Dr. M.C. Roberts of the SFU Geography Department. Dr. D. Smith (Department of Geography, University of Calgary) offered advice

and equipment used during vibracoring.

Acknowledgement of fieldwork assistance is incomplete without extending heartfelt thanks to J. Brumiey of Ethos Consultants, Medicine Hat who, from the early stages of this project, was an invaluable source of information and logistic support. His efforts in towing my field vehicle (haplessly equipped to withstand the rigors of a prairie blizzard) on and off Chappice Lake merit special thanks.

Few paleontological projects lack hours of tedious lab work, and this effort is no exception. I benefitted greatly from the patience and care that characterised those who assisted in some of the repetitious, time-consuming tasks required to finalize this work. Thanks to S. Chow for assistance processing pollen samples and M. Chiao and M. Pellat for help in isolating plant macrofossils.

Colleagues from a number of institutions provided invaluable assistance and expertise. Dr. E. Nelson of the SFU Archaeology Department kindly produced the first two AMS radiocarbon dates for this study. Dr. A. Beaudoin of the Archaeological Survey of Alberta arranged for a radiocarbon determination performed by Dr. D. Arnold of the Alberta Environment Centre. Dr. J. Clague of the Geological Survey of Canada brought years of experience to logging core sections and submitted samples to the Ottawa GSC lab for analyses. Dr. U.T. Hammer provided water chemistry data from his early collections at Chappice Lake. Dr. H. Hudson and R. Stein of the Research Council of Alberta offered maps, publications, and opinions on matters concerning the hydrogeological setting of the study site. Dr. W. Schofield and colleagues at UBC made my excursions to the herbarium both enjoyable and interesting. Thanks also to R. Lawford of the Atmospheric Environment Service for his interest and support, and to the Vancouver AES office for supplying meteorological records from stations across western Canada. R. Long, E. Carefoot, and B. Birch of the Instructional Media Centre (SFU) provided high quality photographic and drafting services.

I would be remiss in not mentioning the valuable support provided by many friends over the years of this project. How many times did they listen to complaints concerning problems encountered during one step or another? I thank them for their patience. Special thanks to Tony & Amanda Goerzen for childcare during the final flurry of activity to complete the dissertation. Finally, to Kate I extend heartfelt thanks for hours of editorial assistance and for her patience while enduring the rigors of life with a doctoral candidate.

DEDICATION

For my parents

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CHAPTER 1

INTRODUCTION

The Great Plains of North America stretches southward along the eastern flank of the Rocky Mountains from the Canadian prairie provinces and American midwest to the southern reaches of Texas (Moulton 1983). This expansive interior grassland once supported vast herds of bison that were the staple of nomadic aboriginals. Today the Great Plains is one of the most important agricultural areas on the continent, despite possessing an unreliable precipitation regime. During the late 1800's and decades of the 1920's, 1930's, and 1980's abnormally high temperatures and low precipitation produced prolonged droughts (Borchert 1950, 1970, Jones 1987, Kunkel 1989, Stark 1987, Stockton and Meko 1983) while short-term but nonetheless severe seasonal droughts occurred in 1961 and 1976 (Dey 1982).

Inhabitants of the northern Great Plains (also referred to as the northern interior plains or the northern mixed-grass prairie) are faced with both periodic shortfalls in precipitation and great seasonal temperature extremes. Historic decades of drought have exerted significant social and economic impacts on the region, most notably the hot, dry 'dirty thirties', when an estimated 250,000 farmers were driven off the southern Canadian prairies (Godwin 1988). More recently, it is estimated that the summer drought of 1980 cost the Canadian prairie provinces at least \$2 billion in lost crops and related revenues (Fraser 1980).

In light of such events, forecasts of increasing drought frequency for the region are particularly alarming. Computer models incorporating projected increases in 'greenhouse gases' predict increased temperatures and greatly increased evapotranspiration in the next few decades (Hare 1988). Although the course of future climate is by no means certain, computer models simulating climatic development suggest that the drought-prone northern interior plains may be subjected to even more frequent and severe drought events than those in recent history. Such predictions have gained credence recently with the six warmest years on record (global

average surface temperature) occurring in the last decade (Houghton and Woodwell 1989), and 1990 the warmest year yet (Pearce 1991). Drought on the northern Great Plains has become commonplace (Stark 1987) and the summer drought of 1988 ranks as one of the most severe this century (Kunkel 1989). An abnormally warm, dry climate has created severe water shortages and lake dessication (Warkentin 1988, Kunkel 1989), resulting again in widespread displacement of people (Farrow 1989).

One of the difficulties encountered when attempting to forecast climate is the limited time span of the meteorological record. On the northern Great Plains the record spans little more than a century. In order to avoid the dangers of forecasts based on a limited perspective, it is necessary to examine patterns of climatic change over as long a time span as possible to obtain a more reliable guide to the likely limits of climate (Gribbin and Lamb 1978). In other words, if we are to acquire some understanding of future climate on the northern Great Plains, it is necessary to understand past climate (Bryson 1980).

The purpose of this thesis is to expand the understanding of past climate by providing a proxy record of climatic change for the northern interior plains spanning the last 8500 years. Reconstruction of lake level fluctuations, using modern and fossil pollen and plant macrofossil analyses combined with sediment stratigraphies, is a recognized approach to studying drought periods (Watts and Winter 1966). Climatic histories based on changing lake levels have aided reconstruction of long-term climate in Europe (Gallaird 1985) and past atmospheric circulation patterns over the northern hemisphere (Harrison and Metcalfe 1985). To date however, detailed records of lake level fluctuations on the northern Great Plains are rare, and none have combined the above methods with efficient power-assisted core collecting and accelerator mass spectrometry (AMS) radiocarbon dating.

There are five main reasons why paleoecological studies in general, and lake level reconstructions in particular, do not form a significant part of the existing Great Plains data

base (Barnosky et al. 1987b). First, there are few suitable wetland study sites, as the extreme aridity of the area precludes permanent wetland development. Lakes tend to be shallow, flat-bottomed, and prone to drying.

Periodic dessication creates the second problem confronting paleoecologists, as it results in truncated sedimentary records. Many lakes of the area exhibit playa characteristics; those that do not dry seasonally likely dried completely during times of extended drought, as occurred in the late 1800's, 1920's, 1930's, and 1980's. Loss of portions of the sedimentary record through deflation creates significant gaps in paleoecological reconstructions.

A third complication involves the widespread occurrence of carbonaceous bedrock, till, and soils. Incorporation of ancient carbon from these deposits in lake sediment produces inaccurate conventional radiocarbon determinations when bulk sediment samples are dated. This has been a particularly thorny problem, as the cornerstone of a meaningful climate history is an accurate chronology.

A fourth problem concerns sample collection. Highly minerogenic sediments are characteristic of most lake basins in the region. This type of sediment is difficult to collect with the hand-operated coring equipment typically employed. The inability to penetrate minerogenic sediment also produces gaps in the sedimentary record.

Finally, paleoecologists relying upon pollen diagrams are faced with pollen spectra dominated by numerically-important taxa that include several species with wide ecological amplitudes. At present it is impossible to distinguish genera of grasses (Gramineae) and sedges (Cyperaceae), and species of sage (*Artemisia*). Taxonomic imprecision therefore constrains reconstructions based on pollen stratigraphies, particularly when aimed at documenting dynamics of upland grassland vegetation.

In this thesis, a number of new techniques and approaches have been employed to circumvent these long-standing problems. Surveying for a suitable study site focused on areas with groundwater inflow, as subsurface flow could buffer the basin from periodic desiccation. AMS dating techniques have been employed, allowing for careful selection of radiocarbon samples to avoid contamination from ubiquitous ancient carbon sources. Power-assisted coring equipment was used for sample collection to ensure complete sedimentary records were obtained. Finally, the research focus is reconstructing lake level fluctuations; as such, the utilization of local aquatic pollen, plant macrofossil, and sedimentary indicators of shoreline movements render taxonomic problems associated with reconstructing dynamics of upland vegetation a secondary concern. To develop an understanding of the critical pollen and plant macrofossil taxa used to document shoreline transgressions, a study of current subfossil deposition patterns is included. The result is a chronology of past lake level fluctuations on the northern interior plains that is as accurate as is currently possible.

Lake level fluctuations, once documented, are used as a guide to past climate. Closed saline lake basins of the region are well known for their responsiveness to changes in the hydrological balance (Last and Slezak 1986). Since precipitation and temperature are major components of the hydrological balance, closed saline lakes are also sensitive barometers of climate. Comparison of the lake level reconstruction produced in this thesis with previous regional records using other means, allows development of a record of significant Holocene climatic events on the northern interior plains, with an emphasis on the frequency and severity of drought.

There is no universally-agreed upon definition of drought (Chakravarti 1970). Drought is a relative term, not amenable to scientific description (Roots 1988); its definition dependent upon the inhabitants of a given location. For the purposes of this study, drought is defined as a prolonged period of reduced precipitation and/or increased temperature capable of lowering lake levels to the point that biota and sedimentation in the basin are altered. The

1980's serve as a guide for assessing the severity of past drought events, as lake draw-down during this decade was sufficient to affect sedimentation in a way that provides an analogue for past events.

The research described in the following pages represents the first attempt at combining recently developed methods and techniques to tap the detailed reservoir of proxy climatic data contained within saline lakes of the northern interior plains. The result is a record of past lake level dynamics, considered to be one of the most useful climatic indicators in continental regions (Kerr 1984), and presentation of a regional drought history with hypothesized atmospheric circulation patterns for significant climatic intervals. These patterns may help to hone computer models, and hopefully help predict future Great Plains' climate.

CHAPTER 2

SALINE LAKES OF THE NORTHERN INTERIOR PLAINS

2.1 Introduction

Inland saline lake basins are found on all continents and are virtually equivalent in volume to freshwater lakes. These lakes have been of economic importance since at least the time of Arab salt mining ventures in North Africa 2000 years ago (Williams 1981). In North America, saline lake basins on the northern interior plains have been mined for the last 65 years, with extraction of sodium sulfate salt alone now generating some \$30 million annually (Last and Slezak 1987).

Closed-drainage saline lake basins are unique ecosystems because of their rapidly fluctuating physio-chemical parameters. These dynamic and often extreme conditions place great physiological demands on their biota, resulting in greatly reduced biodiversity compared to freshwater lakes (Northcote and Larkin 1966). In this sense, they may be regarded as relatively "simple" ecosystems (Hammer 1978). Closed saline lakes also respond rapidly to climatic change (Street-Perrott and Harrison 1985) and are sensitive indicators of the hydrological budget (Last and Slezak 1987). Thus, their sediments contain a record of past climate (Eugster and Hardie 1978, Williams 1981) on both short and long-term time scales.

Despite their widespread occurrence, economic value, and scientific potential, the dynamics of salt lakes remain poorly understood (Williams 1981) and depositional regimes in saline lake basins are not well-documented (Last and Slezak 1987). Thus, a potential storehouse of detailed paleoenvironmental information remains relatively little explored. Recent studies of the inland saline lakes of the northern interior plains have helped fill this information gap. The purpose of this chapter is to review the work and outline aspects of the complex ecological and depositional regimes of saline lake environments with an emphasis

on those that have a bearing on the paleolimnological record.

2.2 Formation

The northern interior plains of North America contain some 3.5 million lakes and twice that number of semi-permanent sloughs (Last 1989), most of which are brackish, with total dissolved salt (TDS) content of 1–3 parts per thousand (ppt), or saline (TDS greater than 3 ppt). This staggering number is a clear indication of the region's combined geographic, geologic, climatic, and topographic features that offer an ideal setting for the development of saline lakes. Formation of saline lakes may be viewed as a tripartite process: a basin capable of ponding water is created, solutes are acquired, and evaporation rates must be high enough to concentrate brine (Last and Schweyen 1983).

Glaciation is responsible for the development of more lake basins than all other geologic processes combined (Reeves 1968), and recently glaciated terrain on the northern interior plains offers numerous opportunities for the development of internally drained basins. During deglaciation, the region's northeasterly drainage path was blocked by the Laurentide ice mass, creating extensive but short-lived proglacial lakes (Teller 1987). These lakes left their mark on the existing terrain in the form of extensive plains underlain by lacustrine deposits. Many proglacial lake basins have subsequently been dissected by deep valleys trending to the southeast, vestiges of meltwater spillways that channelled enormous amounts of water during episodic drainage. Gently rolling topography in other areas of the northern plains is a product of the stagnating Laurentide ice mass (Klassen 1989). Large blocks of ice, remnants of the main ice mass, melted in place leaving a multitude of potential catchments in depressions of hummocky disintegration moraine. Topographic lows of proglacial lake basins, isolated pools in hummocky terrain, or portions of meltwater channels closed by debris slides or alluvial fans, create topographic settings conducive to the development of internally drained

lakes. Solutes are then acquired from groundwater flow and runoff.

The 1500 m of sedimentary rocks underlying the northern interior plains contain many solute sources, but to date the exact source formations have not been identified, despite considerable interest and debate (see Last 1989 for a discussion of existing viewpoints). Basal Paleozoic rock, consisting mainly of carbonate–evaporite sequences (Last and Slezak 1986), is one potential source (Grossman 1968), although ionic proportions incompatible with those in existing saline lakes argue against this (Last and Slezak 1986). Overlying Mesozoic/Cenozoic sand and shale sequences are a more probable source. The solute enrichment mechanism advocated by most researchers is incorporation of Cretaceous bedrock fragments in the thick (in some instances up to 300 m) mantle of unconsolidated Quaternary deposits (Last and Slezak 1986) and overlying soils, followed by dissolution and movement of solutes by groundwater to lakes (Cole 1926, Freeze 1969a, Hendry et al. 1986, Last and Schweyen 1983, Rawson and Moore 1944, Whiting 1977).

Lake water in the region is dominated by Mg, Na, and SO_4 ions (Hammer 1978), a chemical signature compatible with this mechanism of mineral enrichment. Rawson and Moore (1944) noted that soluble salts in Saskatchewan soils are dominated by sulfates of Mg and Na, while Cole (1926) considered the most likely source of the common Na-rich brines to be cation exchange (Ca for Na) in glacial till. Wallick (1981) suggested that the chemistry of a shallow groundwater system in central Alberta is the product of chemical weathering and leaching of glacial drift, and Hendry et al. (1986) proposed that the sodium sulfate-rich brine of a southern Alberta groundwater system is derived from oxidation of organic sulfur in weathered till while Na enrichment results from cation exchange in till.

Once solutes begin accumulating in lake basins, the climatic regime is capable of rapidly concentrating and transforming brine. Lying in the rainshadow of the Rocky Mountains, the prairie region is subjected to high summer insolation and dry winds, producing

high evaporation rates. High evaporation/precipitation ratios, generally on the order of 3 to 10 (Last 1989), are typical of the region. This may be a contributing factor in the dominance of sodium sulfate-type water in the area, as precipitation of relatively insoluble carbonate minerals occurs during runoff under highly evaporative conditions, resulting in sodium sulfate-rich lake water (Gorham et al. 1983).

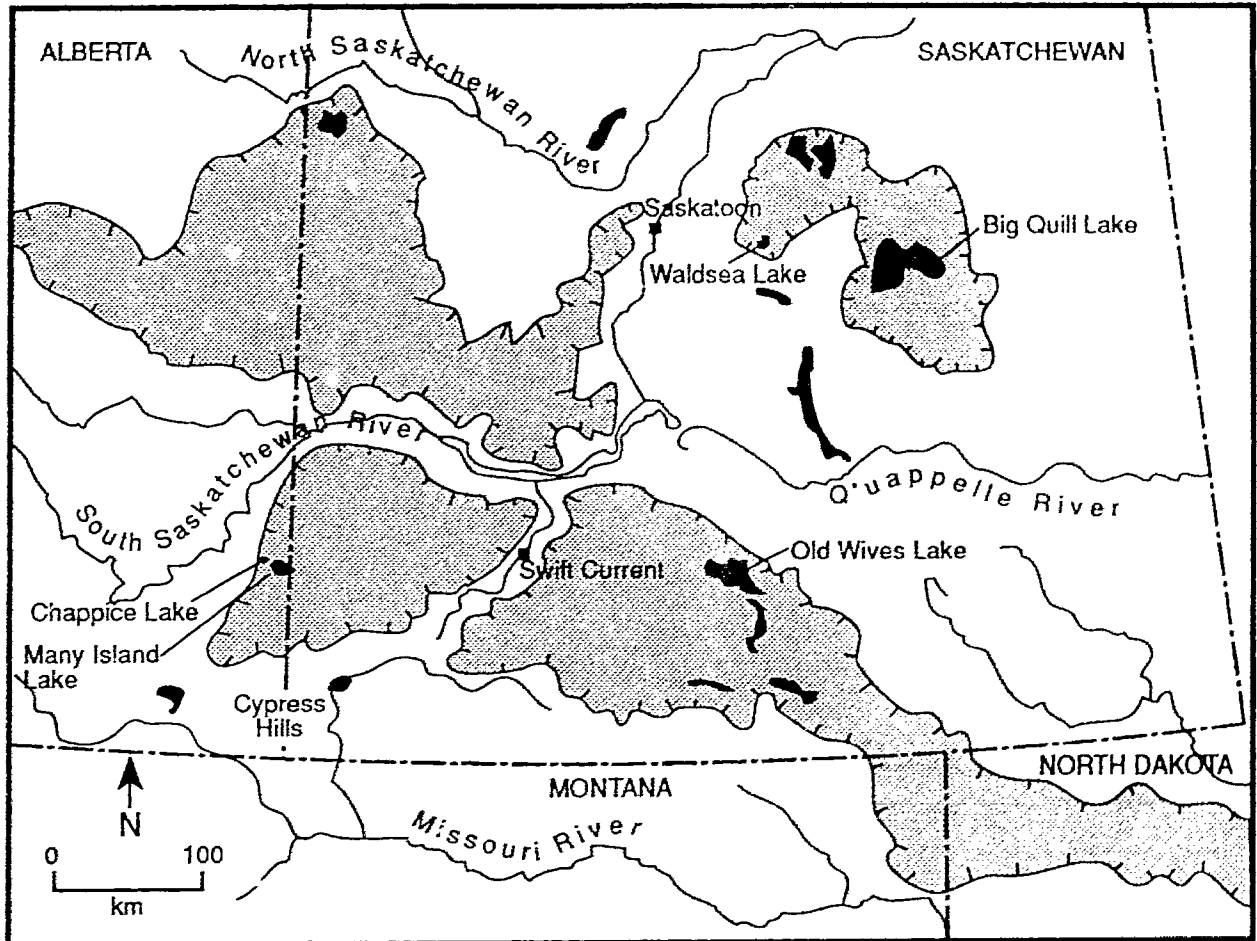
2.3 Regional Distribution

The vast majority of closed saline lake basins in the western interior of Canada are clustered within four regions of internal drainage situated between the major river courses of the area: The Quill Lake Plain lying between the South Saskatchewan and Qu'Appelle Rivers, the Saskatchewan Rivers Plain between the northern and southern arms of the Saskatchewan River system, and the Old Wives and Bigstick Lakes Plains between the South Saskatchewan and Missouri Rivers (Fig. 1). Evaporation/precipitation ratios tend to increase to the southwest (Hammer and Haynes 1978). Although lake size is highly variable, ranging from small prairie potholes (less than 1 ha) to expansive lakes some 300 km² (Last and Schweyen 1983), most tend to be flat-bottomed and shallow (mean depth of less than 10 m) with regular shorelines (Hammer 1978).

2.4 Hydrological Budget

In closed saline lakes, the hydrological budget is straightforward with precipitation, surface runoff, and groundwater inputs balanced by evaporative losses and groundwater outflow. Annual precipitation throughout the northern interior plains is low, averaging about 30 cm (Last and Schweyen 1983). Approximately 70% falls during the growing season (Rawson and Moore 1944), primarily in spring. It is difficult to determine what portion of the yearly precipitation actually reaches lake basins of the region via runoff, as a variety of

Figure 1. Areas of internal drainage on the northern Great Plains (shaded), showing location of study sites discussed in text (adapted from Hammer 1978, Last and Schweyen 1983).



environmental features in the area produce complex interactions between surface runoff and infiltration (Winter 1989). The generally low gradient of the land results in low runoff velocities and maximizes infiltration opportunity. However, the clay- and silt-dominated soils common to most areas retard infiltration, except through fractures, and underlying geologic formations generally possess low permeabilities. In addition, most precipitation and runoff occurs when ground-frost is as much as 1 to 1.3 m deep. Thus, despite contributing less than a third of the yearly precipitation budget, winter precipitation may exert a great

influence on lake levels, since little is lost to infiltration or evaporation (Whiting 1986). Winter runoff and spring precipitation also exert a dramatic impact on lake water composition and concentration, as inflow of relatively fresh water may cause temporary chemical and/or thermal brine stratification.

Calculating yearly water budgets for prairie lakes is extremely difficult, since groundwater movements are generally poorly known, despite contributing significantly to all saline lakes in the area (Last and Schweyen 1983). Input may occur via shallow (less than 2 m deep) local systems, deeper regional flow, or some combination of the two. Water originating from either system should have a distinctive chemical signature (Last and Slezak 1986), depending on the mineralogy of the surrounding geologic formation, the solubility of component minerals, and possible secondary reactions with subsurface rocks during transport (Freeze 1969a).

The contribution different flow regimes make to a particular basin is difficult to ascertain. The distribution of Quaternary sediments is vital, as they are thought to determine the location and development of saline lake basins by controlling the direction and flow of groundwater (Last and Schweyen 1983). Similarly, the close association of salt lakes with meltwater spillways and preglacial valleys suggests these features may act as groundwater conduits (Last 1989). Groundwater flow in two of the largest, shallow, sodium sulfate-rich lakes in the region, Old Wives and Big Quill Lakes (Fig. 1), has been studied in some detail.

Freeze (1969a, 1969b) outlined aspects of groundwater flow in the Old Wives Lake Plain through piezometric analyses, hydrogeochemistry, and surficial evidence. He concluded that the system is recharged on the permeable western uplands of the study area, moves northeastward through glacial drift, and discharges into Old Wives Lake. A buried preglacial valley system appears to be an integral part of the subsurface drainage in some parts of the

Old Wives Plain, and Freeze suggests his data "...offer piezometric proof that all such valleys are permanent groundwater discharge features" (1969b: 53). LaBaugh (1989) however, notes that interpretations concerning groundwater flow in and out of lakes based on samples from nearby wells, should be viewed with caution as the connection between the well site and actual flow pathways is not usually known.

Whiting (1977) developed a model of the Big Quill Lake water budget by monitoring hydrology and climate in association with lake level adjustments from 1963 to 1975. "Since the lake does not have a surface outlet the difference between the cumulative effect of surface inflow, precipitation, and evaporation and changes in the water content of the lake must give a measure of the groundwater discharge or recharge" (ibid: 7). By this reasoning, the groundwater contribution to the basin was calculated to be some 13% of the annual inflow, although difficulties in accurately measuring parameters of the water budget may have resulted in measurement errors to the degree that the measured difference should be viewed with caution (LaBaugh et al. 1987).

Big Quill Lake is fed by an aquifer associated with a meltwater channel and is underlain by a preglacial river-bed aquifer. Although the data are insufficient to clarify whether one or both aquifers contribute to the lake, the nature of lake level fluctuations suggests that more than one aquifer is involved. Periods of groundwater outflow are also implicated. In addition, the synchronous lake level changes of nearby Carlyle and Last Mountain Lakes point to similar, if not interconnected, groundwater systems.

While the groundwater component of the hydrological budget remains elusive, the distinctive chemical signature of groundwater from different formations offers some promise for detecting flow patterns, although in areas of hummocky terrain typically bearing alternating recharge and discharge areas, mixing may obscure the chemical signature (Freeze 1969a). Regional water chemistry surveys have led to suggestions that chloride-rich water is produced

by springflow (Hammer 1978), and water with a high carbonate content in an area typified by sulfate waters may be indicative of the presence of springs (Rawson and Moore 1944). Rozkowska and Rozkowski (1969) noted seasonal changes of ionic composition in brines of southeastern Saskatchewan, and suggested late summer enrichment of Mg and SO₄ ions was a product of increased groundwater input. LaBaugh et al. (1987) noted that as groundwater moves from a recharge to a discharge area, it is enriched in sulfate and major cations, particularly Mg.

Due to the varied chemistry and amount of groundwater input, prairie lakes display a great range of concentration, varying between 1 to 300 ppt, but averaging about 31 ppt (Last 1989). Evaporation is the one universal control of water levels in all closed saline lakes (Langbein 1961), and is particularly important in catchments with a high surface-to-volume ratio, where high evaporation rates are capable of rapidly producing extreme salinities (Blinn 1971).

Deep saline lakes exhibit yearly salinity cycles on the order of 2% to 10% (Northcote and Larkin 1966), while shallow perennial lakes experience surface water salinity increases on the order of 30% from May through September (Rawson and Moore 1944). Ephemeral basins display more dramatic evaporative concentration, with seasonal increases reported as high as 300% to 400% (Last 1984, Rozkowska and Rozkowski 1969). Brine evaporation is not a straightforward process however, since evaporation rates decline as brine concentration increases (Langbein 1961, Last 1984). High salinity brines absorb heat more rapidly at the surface (Hammer and Haynes 1978), and brine temperatures as high as 45°C on a hot summer day have been observed (Last 1987). Increasingly saline conditions have an effect on ice cover as well, since water of increased ionic concentration tends to freeze later and thaw earlier (Hammer and Haynes 1978), thereby lengthening the ice-free season (Northcote and Larkin 1966).

2.5 Saline Lake Biota

Inland saline lake basins are variable environments characteristically inhabited by a high proportion of opportunistic and fugitive species possessing effective dispersal mechanisms (Williams 1981). The high evaporation rates and great temperature extremes of the northern plains region bring about rapid changes in ionic concentration of lake water, placing great stress on the occupants of saline lake basins and reducing biodiversity.

Early surveys of saline lakes in Saskatchewan documented a pattern of reduced diversity with increasing salinity (Rawson and Moore 1944). Fish are not found in water where solutes are more concentrated than 20 ppt, and are generally restricted to concentrations of less than 7 ppt. Brine shrimp (*Artemia salina*) are one of the few inhabitants of hypersaline water (that is, where salinity exceeds 50 ppt), and have been noted in abundance as the sole multicellular inhabitants in water exceeding 115 ppt (ibid). Although algae are found in water ranging from fresh to excessively hypersaline (as high as 428 ppt), 14 species are restricted to hypersaline water in the area, with green algae dominating when concentrations exceed 100 ppt (Hammer et al. 1983). Productivity in Saskatchewan saline lakes is highest during algal blooms in both dilute and highly saline lakes (Haynes and Hammer 1978). One common green alga in hypersaline waters of the area (*Dunaliella salina*) is so abundant at times that it colors the water red (Hammer et al. 1983).

Benthic fauna are common in fresh-to-marginally saline lakes, but like other inhabitants are severely restricted by high salinities, and are absent from lake sediment when water concentration exceeds 100 to 125 ppt (Rawson and Moore 1944, Hammer et al. 1975). The presence of benthic organisms has a profound impact on the paleolimnological record, as these organisms commonly mix and disturb surficial deposits, thereby destroying details of the sedimentary record.

In terms of plant species, Ungar's (1974) survey of North American inland saline lake flora revealed considerable uniformity, suggesting that few have evolved adaptations required for survival in this setting. Although salinity is a critical limiting factor in species distribution, water depth and duration of flooding are also important (Kantrud et al. 1989). Under the low precipitation regime in the northern interior plains, moisture stress may also play a critical role in plant distribution (Ungar 1974). This, combined with temperature extremes and the dynamic nature of the basins (both in terms of water levels and salinity), likely accounts for the decreased diversity of halophytic vegetation in Canada compared to more southerly regions (Hammer and Heseltine 1988).

Because of severe and changeable growth conditions, vegetation zones are often conspicuous, with each zone usually dominated by a single species (Kantrud et al. 1989), leading most investigators to name zones after a single dominant or codominant species (Dodd and Coupland 1966, Keith 1958, Ungar 1974). Zonation within any individual basin fluctuates. Changing growth conditions, the rapid growth typical of many halophytes (Walker and Coupland 1968), and potential changes stored within ubiquitous seed banks (van der Valk and Davis 1976), means all assemblages are prone to rapid change. Seeds of halophytes are well known for long viability and dormancy under unfavourable conditions (Ungar 1978).

When standing water persists, aquatic species are found in most water bodies where salinity is less than about 75 ppt (Hammer and Heseltine 1988). *Ruppia maritima* (nomenclature of aquatic species follows Scoggan 1978), the most salt tolerant macrophyte, is found in shallow, highly saline water (Hammer and Heseltine 1988). *Potamogeton pectinatus* may occur in water nearly as saline (Kantrud et al. 1989), as high as 60 ppt (Hammer and Heseltine 1988). *Ruppia occidentalis* is found at depths as great as 8 m at similar salinities (ibid.). *Chara* spp. have been collected in moderately saline water, to about 25 ppt (ibid.). In less saline waters aquatic flora are more diverse (see Kantrud et al. 1989 and Pip 1988 for a listing of common species).

Scirpus paludosus (nomenclature of shoreline and upland taxa follows Moss 1983) is the most common emergent of hypersaline lake basins of the northern interior plains (Hammer and Heseltine 1988, Leiffers and Shay 1980, Shay and Shay 1986). Although able to tolerate high salinities, an extended period of inundation appears necessary for the development of dense, vigorously growing stands (Dodd and Coupland 1966). *Potamogeton pectinatus*, *Zanichellia palustris*, and *Chara* spp. are common associates of *Scirpus paludosus* but when dry, stands are quickly invaded by terrestrial chenopods and *Hordeum jubatum* (Leiffers and Shay 1980, Dodd and Coupland 1966). Other common emergents of hypersaline lakes in the area include *Scirpus pungens*, *Distichlis stricta*, *Puccinellia nuttalliana* and *Triglochin maritima*. *Typha latifolia* and *Scirpus acutus* are the dominant emergents of marginally saline-to-freshwater basins (Hammer and Heseltine 1988).

Dodd and Coupland (1966) recognized a series of vegetation zones surrounding saline basins in a survey of approximately 270 sites in the grassland region of Saskatchewan. Moving upslope from the lowermost reaches of the wetland, community dominants typically follow a sequence of *Salicornia rubra*, *Triglochin maritima*, *Puccinellia nuttalliana*, *Distichlis stricta*, *Hordeum jubatum*, and *Agropyron* spp. Although each of these species dominates a portion of the elevational gradient in nearly pure stands, transition zones of varying widths occur that may be dominated by two or more species. Dodd et al. (1964) provided conductivity and soils data for these sites that suggest vegetation zones are arranged along a salinity gradient, from high salinities on the basin floor to lower salinities characteristic of upland sites. Keith (1958) described a similar zonation pattern surrounding saline impoundments in southern Alberta. Kantrud et al. (1989) noted that both salinity and available moisture are important determinants of vegetational zonation in northern interior wetlands, and pointed out that groundwater seepage may at times interrupt this elevational gradient.

2.6 Sedimentary Processes

Saline lakes display a wide spectrum of sedimentary regimes (Last 1989). Sediments are a mixture of siliclastic material, organic detritus, and chemically-precipitated minerals that at times record hydrological change in a straightforward manner. Alternatively, sedimentary records may be complicated by diagenetic alterations, making interpretation difficult (Teller and Last 1990).

Essentially all brines in the northern interior plains are saturated or supersaturated with respect to calcium and magnesium carbonates, whereas highly saline brines are saturated with respect to more soluble alkaline salts. Chemical precipitates include a wide range of relatively insoluble carbonates as well as more soluble minerals such as gypsum ($\text{Ca}_2\text{SO}_4 \cdot 2\text{H}_2\text{O}$), mirabilite ($\text{Na}_2\text{SO}_4 \cdot 10\text{H}_2\text{O}$), thenardite (Na_2SO_4), bloedite ($\text{Na}_2\text{Mg}(\text{SO}_4)_{13} \cdot 15\text{H}_2\text{O}$) and halite (NaCl ; Last and Schweyen 1983). Organic matter content varies, likely influenced by rapid changes in brine concentration that bring about die-offs or blooms of organisms. Siliclastic sediment input is mainly through streamflow, although wind and sheetflow may be locally important. This material is predominantly silt- and clay-sized quartz, feldspars, carbonates, and clay minerals (ibid), closely reflecting the composition of surrounding glacial till (Last and Slezak 1987). Coarser grained sands and gravels generally remain in the near-shore region while finer particles are transported in suspension to deeper parts of the basin (Last and Schweyen 1983).

Many factors influence sedimentation, including climate, groundwater, geologic setting, and basin morphology (Teller and Last 1990). However, an important distinction between perennial and ephemeral lakes must be made since processes operating within each setting have a dramatic impact on the sedimentary record (Last and Schweyen 1983). Sedimentation in ephemeral basins is complex and dynamic. Interpretation of long-term sedimentary records is complicated by the fact that many perennial basins were likely ephemeral basins during the

mid-Holocene (Last and Slezak 1986).

Some deep, perennial lakes on the northern interior plains are meromictic. Waldsea Lake is relatively large (surface area of 4.64 km²) with a maximum depth of 14.3 m (Hammer et al. 1975). Thermal and solute gradients in the water column are apparent. From 1970 and 1977 the chemocline in Waldsea Lake moved between a depth of 4 to 7 m, in response to variations in the amount of spring freshet, intensity of summer rainfall, wind strength during the ice-free season, and groundwater input. Deposition in deep, stratified lakes like Waldsea is reasonably straightforward, with the input of siliclastic sediment determined by wind, streamflow, and slopewash whereas precipitation of evaporite minerals is dependent upon brine concentration and composition. Episodic sediment deposition may result in laminated sedimentary sequences, potentially bearing a record of short-term events. For example, Last and Schweyen (1985) suggested that aperiodic influxes of bicarbonate-rich runoff may stimulate massive inorganic carbonate sedimentation during deep-water phases, producing relatively pure aragonite laminae. Sediments in stratified deep-water basins are unlikely to contain benthic organisms in the oxygen- and light-poor, highly saline H₂S-rich environment of the hypolimnion. Sedimentary structures are thus likely preserved, resulting in generally fine-grained, organic-rich, laminated sequences (Last and Schweyen 1983).

Since only 10% of existing perennial lakes in the area are greater than 3 m in depth (ibid.) and only two lakes in Saskatchewan are known to be chemically stratified (Hammer and Haynes 1978), conditions necessary for the deposition of deep-water laminae are rare and perhaps transitory. Hammer et al. (1975) considered the meromictic condition of Waldsea Lake to be temporary because of the relatively large but shallow nature of the basin, its exposure to wind, and the relatively small salinity difference (less than 2) between the monimolimnion and mixolimnion.

In the far more numerous shallow basins, a dynamic sedimentary regime exists. Although chemical stratification may be less permanent than in deep lakes, evaporation and temperature change may promote chemical stratification of the water column. As previously noted, high summer temperatures combined with strong winds produce evaporation rates that are capable of concentrating brines rapidly. This can lead to mineral precipitation, beginning with relatively insoluble carbonates and eventually more soluble salts by a variety of potential pathways (Eugster and Hardie 1978). Temperature extremes may also stimulate mineral precipitation, as decreased temperature reduces solubility (Last 1984). For example, sodium sulfate-rich brines commonly precipitate mirabilite and thenardite crystals under a winter ice cover (Last 1989, Rawson and Moore 1944). This precipitate redissolves when temperature rises in spring, producing a layer of highly concentrated brine near the sediment-water interface. Rawson and Moore (1944) noted a thin (less than 1 mm) high salinity (118 ppt) bottom layer in Little Manitou Lake (maximum depth of 3.5 m). A similar high density bottom layer has been noted in shallow (1.5 m) Wallender Lake, B.C. (Blinn 1971), indicating that chemical gradients do exist in shallow saline lakes. Such gradients are capable of inhibiting thermal and oxygen circulation (Rawson and Moore 1944). Within this setting, discrete chemical, biological, and clastic sedimentary events may produce laminae that are sheltered from wind disturbance by the dense overlying brine, and protected from biotic disturbance by high salinities. In contrast, vernal circulation is complete in less saline shallow lakes. Undersaturated brines allow disturbance by wind and biological activity, resulting in the deposition of massive sediments that occasionally bear evidence of burrowing organisms (Teller and Last 1990).

In shallow basins there is always a possibility that during an extended drought, or as part of a yearly cycle, the brine will evaporate completely. Many shallow prairie basins experience such seasonal dessication, forming playas. Deposition in a playa setting is complex. During lake recession, large areas of mudflats are exposed and often become covered with

finely crystalline efflorescent salt crusts, formed by evaporation of brine films brought to the surface by capillary action and evaporative pumping (Slezak and Last 1985). Salt crusts may either remain in place and hinder further evaporation of pore water within underlying sediment (Last 1987), be removed by wind, or redissolved and transported into the basin by sheetwash (Slezak and Last 1985). Mudflats are also the scene of intrasedimentary crystallization of salts, a process that will disrupt laminations (ibid).

In the central playa area, the brine becomes increasingly concentrated during draw-down, at times exceeding 300 ppt (Last 1984, Last and Slezak 1987), leading to supersaturation and crystal growth at the brine surface (Last 1984). As evaporation proceeds to complete dessication, deposition of a salt bed 10 to 30 cm thick can occur (ibid). Under extremely arid conditions, the water table may drop well below the lake floor, promoting downward leaching of soluble salts, loss of the sedimentary record through deflation, and if extended over a sufficiently long period, soil formation (Teller and Last 1990). Intense evaporation on the remaining mudflat creates a dynamic diagenetic environment (ibid). During the next wet phase, up to 20 cm of the salt bed may be redissolved (Last 1984), producing a highly concentrated brine. Redissolution may be the main solute source during subsequent high water stands (Lerman and Jones 1973). Alternatively, if wind and sheetwash clastics have blanketed the basin following deposition a salt bed, dissolution will be minimized (Langbein 1961, Slezak and Last 1985), and salt deposits will be preserved within sediment.

There are other possible postdepositional processes operating in both playa and permanent salt lakes. Dilute groundwater flow may dissolve portions of evaporite deposits producing fissures that may subsequently be filled with fresh or highly concentrated brines, resulting in solution chimneys up to 9 m in depth (Last 1987). Sinter-like ridges and cones of soluble Na and Mg salts may be built up by groundwater discharge, or cool groundwater inflow may decrease brine temperature sufficiently to cause deposition of mirabilite or epsomite ($\text{MgSO}_4 \cdot 7\text{H}_2\text{O}$), producing an undulating lake bed (Last 1989). Oxidation of organics

and sulfate reduction by bacteria (or activity of other organisms) as well as poorly understood diagenetic reactions involving evaporite minerals and clay grains may generate other saline minerals, authigenic silicates, and carbonates (Last and Slezak 1987, Teller and Last 1990), thereby obscuring aspects of the paleolimnological record.

2.7 Responses to Climatic Change

Ungar (1974) cited a number of historical saline lake studies to elucidate the relationship between salinity, lake level, and climate. From ephemeral ponds in Kansas to perennial lakes in North Dakota and Utah the relationship is clear. Drought causes draw-down which leads to increased solute concentration in the remaining lake water. Periods of increased precipitation (particularly when accompanied by reduced evaporative stress) bring about higher water levels and reduced salinity. In fact, this relationship is the basis for the observation that a 0.3 m decline in the level of Great Salt Lake causes a 1.0% increase in solute concentration. Over this broad geographic area, fluctuations in salinity are primarily related to rising and falling water levels, which are ultimately controlled by climate.

Northern prairie wetlands also experience seasonal, annual, and long-term fluctuations in water depth (Kantrud et al. 1989) that affect dissolved salt concentration. Leiffers and Shay (1983) documented water depth and salinity fluctuations of two small ponds in central Saskatchewan over the course of a one-year cycle of infilling and drawdown. Periodic measurements show that water conductivity increases as water levels decline whereas rising water levels are accompanied by decreasing conductivities.

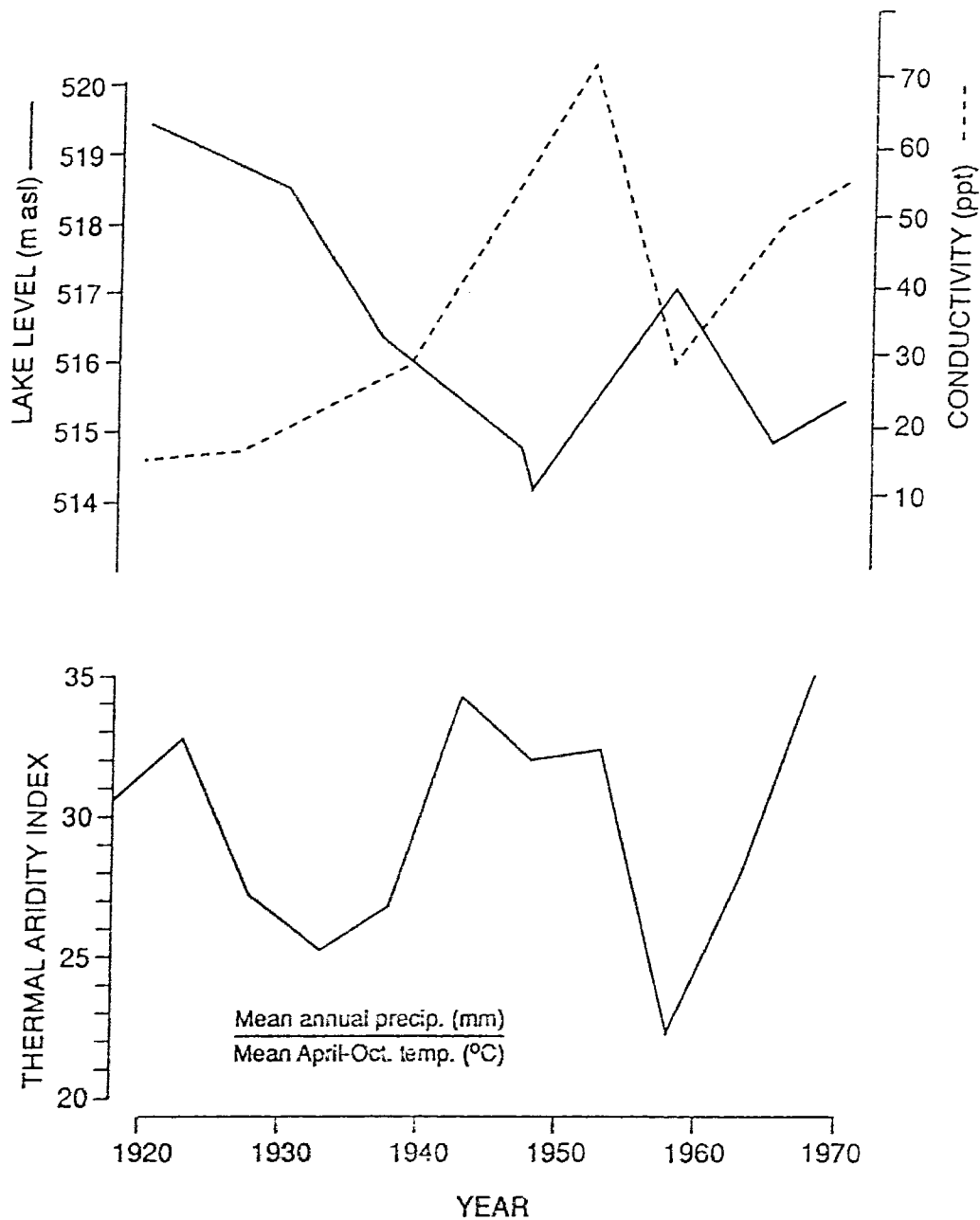
In addition to these seasonal adjustments in small ponds, annual and long-term trends in lake size and salinity have also been documented in larger bodies of water. Rawson and Moore (1944) compiled salinity data from eight lakes in southern Saskatchewan during their pioneering survey of saline lakes in the province. Despite varying in size, depth, and geologic

setting, all lakes showed an unwavering trend of increased salinity from 1920 to 1938, which the authors conclude was related to below normal precipitation. Increases in salinity were particularly rapid from 1937 to 1941, a response to the extreme aridity between 1932 and 1935. Although the magnitude of salinity increase varied between basins (24.4% to 97.5%) from 1920 to 1938, and a lag in lake response was evident, the overall increase in salinity (and decreased lake size) that accompanied increased temperature and reduced precipitation indicates that climate is the major determinant of lake size and salinity in the area. Hammer (1978) documented salinity changes in four Saskatchewan lakes spanning a fifty-year period, and concluded that long-term salinity changes are directly related to precipitation and runoff. These data once again span a wide range of lake size and setting yet consistently portray similar response patterns, although the responsiveness and variability exhibited by small lakes is more rapid and dramatic compared to larger waterbodies.

A detailed record of lake level fluctuations from one of these basins complements its salinity record and is instructive in elucidating the relationship between lake size, salinity, and climate. The Big Quill Lake record spans a fifty-year period (1920 to 1970) during which human activities had a minimal impact on the lake. This record is thus representative of long-term changes within a typical, perennial saline lake in the region. The trends are clear (Fig. 2). Salinity increases occurred during times of decreasing lake levels and salinity decreases accompanied rising lake levels, although salinity responses lagged somewhat behind water level fluctuations. This delayed response was particularly striking during the low water stand of 1940 to 1950. Increasing salinity was actually most pronounced during the following period of rising water levels. This is most likely related to redissolution of salts deposited along the lake margin that were subsequently transported to the lake by sheetflow.

Climatic data from the nearby Saskatoon weather station provide a guide to the regional climatic setting of these changes. A thermal aridity index (after Zoltai and Vitt 1990) was calculated in five-year intervals over the period of interest. The thermal aridity index is

Figure 2. Historic record of lake level and salinity changes at Big Quill Lake (Hammer et al. 1975, Whiting 1973) compared to a five-year thermal aridity index from the Saskatoon meteorological record.



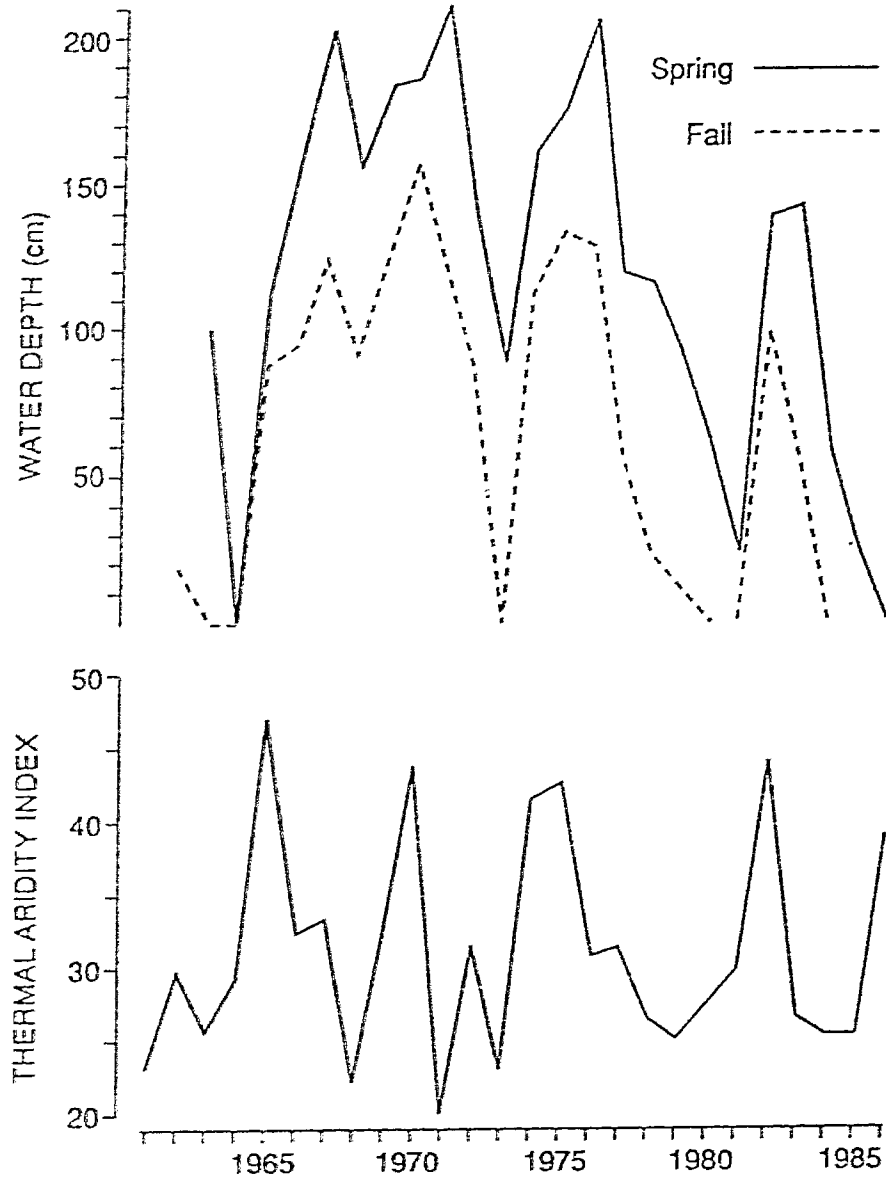
one method of estimating evaporation from temperature, and is employed in the absence of long-term evaporation measurements. Although no temperature index is a suitable measure of evaporation, mean summer temperature (April to October) provides a crude measure of evaporative stress. The thermal aridity index thereby modifies total precipitation according to growing season temperature, producing in a more reliable indicator of the precipitation–evaporation ratio.

The thermal aridity index curve plotted with the Big Quill Lake salinity and water level record (Fig. 2) outlines the close relationship between climate and changes within closed lake basins. As the thermal aridity index decreases (indicating increasingly arid conditions), lake levels decline. Conversely, as the index increases, lake levels rise. Lake response is not immediate, as there appears to be about a ten-year lag between an extreme in the thermal aridity index and its registration on the level of Big Quill Lake. This may be indicative of the buffering effect of groundwater. Although lake level response is not immediate, the direction of the response is closely and consistently related to the temperature–precipitation regime.

A shorter record of lake level changes from a much smaller saline basin in southwestern Saskatchewan (Fig. 3) complements the Big Quill Lake record. Climatic data from nearby Swift Current were used for calculation of a yearly thermal aridity index. Compared to Big Quill Lake, the lake level response is punctuated by abrupt changes, as would be expected from a small basin more responsive to climate, more dramatically influenced by transpiration of marginal phreatophytic vegetation, and more frequently measured. In spite of these differences, the lake's responses to yearly changes in weather are the same as those documented at Big Quill Lake.

These records of climate and lake level change underscore the potential that closed saline lake basins have as climatic indicators. Although groundwater regimes will buffer lake

Figure 3. Historic lake level fluctuations of a small catchment (Kantrud et al. 1989) compared to a yearly thermal aridity index calculated from the Swift Current meteorological record.



responses to climatic change. lake levels and the concentration of salts in lake water appear to be mainly determined by climate, particularly over decadal time scales. These data confirm that sedimentary records from saline lake basins contain detailed paleoclimatic records.

CHAPTER 3
POSTGLACIAL HYDROLOGY AND CLIMATE OF THE
NORTHERN MIXED-GRASS PRAIRIE

3.1 Introduction

Evidence of past lake level change has long been recognized as a source of information on paleohydrology and potentially, of paleoclimate (Street-Perrott and Harrison 1985). Geomorphological features such as strandlines of ancient beaches lying above current lake limits or submerged valleys within present-day lakes, conclusively indicate past water level fluctuations. Biogeographical distributions suggesting a close relationship between now-separate bodies of water imply that these basins were once connected, presumably during a high water stage (Richardson 1969). Such observations broadly outline past lake level fluctuations, but do not provide details on the nature of change. The most specific information on prehistoric spatial and temporal lake level variation is found within lacustrine sediment.

Richardson (1969) listed a series of stratigraphic markers that are of value in outlining past water level changes, based mainly on the observation that lake level fluctuations alter the areal extent and volume of water within a basin. Such changes affect water chemistry and biota, as well as physical and chemical characteristics of the sediment. While offering a potentially detailed record of past events, Richardson cautioned that "...almost no line of evidence is immune to misinterpretation..." (ibid: 90), and advocated a many-faceted approach to investigation, noting that variability in a number of indicators will be a more reliable barometer of past events than the abundance or rarity of any single sedimentary parameter.

A recent methodological summary for documenting past lake levels (Digerfeldt 1986) continues to advocate utilization of multiple lines of evidence to ensure accurate reconstructions. In addition, appreciation of the potential extent of within-basin variation led

Digerfeldt to urge investigators to analyze a transect of several cores within each basin. Multiple core analyses also make possible accurate estimates of past changes in lake area.

A preferred method therefore involves collection of a series of sediment cores extending from the current shoreline to deeper portions of the basin. Pollen and plant macrofossil analyses, in concert with physical and chemical sedimentary parameters, may then be used to outline paleohydrological change. A variety of indicators, correlated among separate core sections, provide compelling evidence of past lake level fluctuations. For example, as a lake reduces in size, emergent vegetation spreads inward. Subsequent shoreline recession will be recorded by increased representation of shoreline pollen and plant macrofossils in the central basin area. An increased load of coarse littoral sediment will be deposited in central portions of the basin as shorelines approach. In basins lacking surface outlets, reduced water volume will be accompanied by increased concentration of evaporitic salts, producing a distinctive signature in lake sediment as water chemistry and inhabitants of the basin change. It is unlikely that anything but lake lowering could account for this suite of depositional characteristics.

Unfortunately, interpretation is not always straightforward, as the record may be modified by a number of factors that have little to do with lake level change. Altered positions of inflowing streams or turbidity currents, redeposition of older sediment, as well as diagenetic processes can confound interpretation. In arid regions, additional complications arise from the characteristic erratic nature of climate. Short-term events (for example, a particularly heavy spring run-off transporting an abundance of clastic debris to the basin), may have a dramatic impact on sedimentation, whereas dessication and deflation may result in loss of portions of the sedimentary record (Richardson 1969).

Despite potential complications, studies of lake level fluctuations employing similar methods have contributed to paleoenvironmental studies in Europe (Gaillard 1985, Digerfeldt

1988), arid regions of Africa, Australia, and the western interior of Canada (Teller and Last 1990). Studies of regional lake level adjustments, particularly in closed drainage basins, have contributed significantly to understanding the nature and direction of past climatic change (Harrison 1988, Harrison and Metcalfe 1985, Street-Perrott and Harrison 1985, Street-Perrott and Roberts 1983).

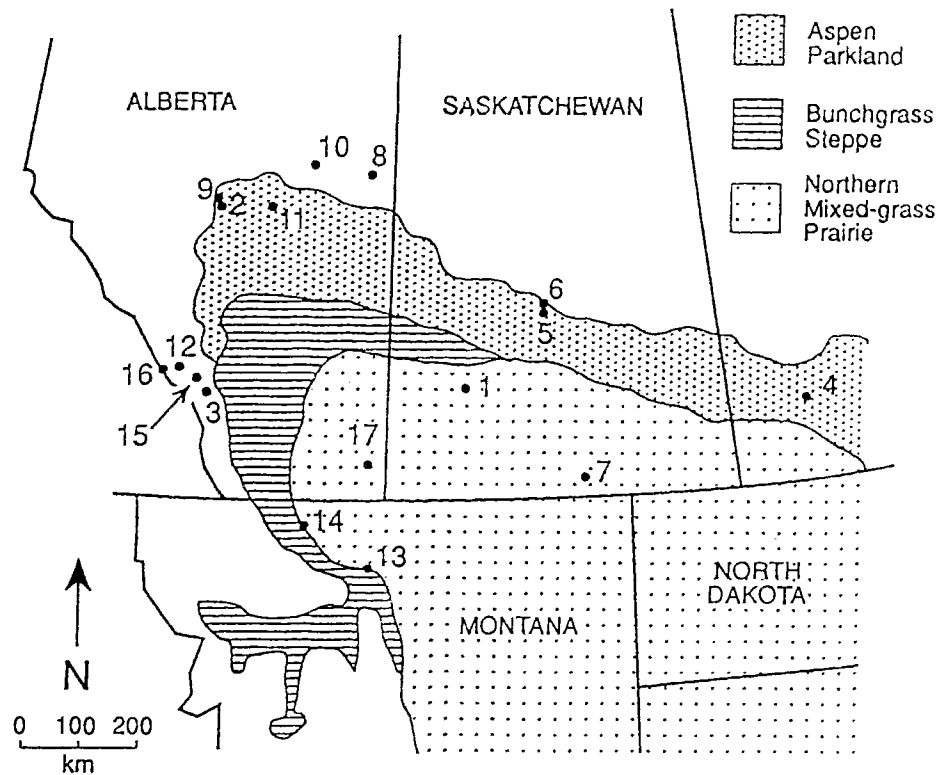
3.2 Limitations of Radiocarbon Chronologies

In the absence of wood remains or other suitably-sized organic fragments, paleoecologists working in the western interior of North America have been forced to rely mainly on conventional radiocarbon dating of bulk sediment. The resulting chronologies are often in error, due to the presence of ancient, radioactively dead carbon in widespread carbonaceous shale, coal and limestone deposits (Barnosky et al. 1987b). An appreciation of the magnitude of this problem was first established by dating near-surface sediment of known age.

Mott (1973) assessed the degree of contamination in sediment from Clearwater Lake, southwestern Saskatchewan (Fig. 4), by carbon-dating the sedimentary horizon marking the onset of European cultivation practices (as indicated by abrupt changes in the pollen and sedimentary record). Although European settlement occurred less than 100 years ago, sediment samples from this horizon produced radiocarbon ages of 1170 ± 190 BP (radiocarbon years before present) on the organic fraction and 1260 ± 190 BP on the inorganic portion. This ca. 1000-year error was caused by radioactively dead carbon in the lake derived from ancient carbonate sources in the catchment basin.

Unreliable radiocarbon chronologies have also plagued paleoenvironmental investigations at Lake Wabamun, Alberta (Hickman et al. 1984). In this case, coal-bearing deposits within the catchment basin appear to be the source of contamination. Questionable radiocarbon chronologies have also hampered paleoecological studies on the eastern slopes of Alberta

Figure 4. Map of the northern Great Plains, showing distribution of vegetation (adapted from Rissler et al. 1981) and study sites discussed in text.



Map No.	Study Site	Reference
1	Clearwater Lake	Mott 1973
2	Lake Wabamun	Hickman et al. 1984
3	Chalmers Bog	Mott and Jackson 1982
4	Lake Manitoba	Teller and Last 1981
5	Waldsea Lake	Last and Slezak 1988
6	Deadmoose Lake	Last and Slezak 1986
7	Ceylon Lake	Last 1990
8	Moore Lake	Schweger et al. 1981
9	Lake Isle	Hickman and Klarer 1981
10	Lofty Lake	Schweger and Hickman 1989
11	Hastings Lake	Forbes and Hickman 1981
12	Wedge Lake	MacDonald 1982
13	Lost Lake	Barnosky 1989
14	Guardipee Lake	Barnosky 1989
15	Toboggan Lake	MacDonald 1989
16	Opabin Lake	Reasoner and Hickman 1989
17	Chappice Lake	this study

(White et al. 1985), due to ^{14}C -deficient carbon in widespread coal deposits and calcareous bedrock. No site exemplifies the potential magnitude of error better than Chalmers Bog. While developing a pollen record for this basin, Mott and Jackson (1982) attempted to circumvent chronological problems by dating an aquatic moss recovered in near-basal sediment. The dates ($18,400\pm 380$ BP and $18,500\pm 1090$ BP) were used as support for the existence of tundra-like vegetation in a late glacial ice-free corridor. Subsequent AMS dates (Gove et al. 1980) on the same species of moss (*Drepanocladus crassicosatus*) and terrestrial plant macrofossils in a nearby basin indicated that the early Chalmers Bog dates were in error, perhaps by as much as 6000 years (MacDonald et al. 1987). These results accentuate the need for judicious sample selection when using conventional radiocarbon dating techniques, in addition to emphasizing the advantages conferred by utilizing AMS technology to date small samples.

3.3 Existing Studies on Past Lake Level Fluctuations

Ongoing investigations into the modern sedimentation and history of saline lake basins in the western interior of Canada have advanced understanding of these complex sedimentary environments. This research, mainly by W.L. Last and his colleagues, has made use of a variety of physical and chemical sediment parameters from multiple sampling locations within each basin. Large lakes with significant surface outflow and inflow, as well as small, closed playa lakes have been investigated. The results of more than a decade of study offer insight into the paleohydrology of several catchments.

Teller and Last (1981) analyzed 51 sedimentary cores to reconstruct past water level fluctuations of Lake Manitoba, a large (4700 km^2), shallow (maximum depth of 6.3 m) basin situated on the eastern grassland margin. Although complicated by an error-ridden chronology (due to contamination from pre-Quaternary carbon) and differential isostatic rebound, physical,

mineralogical, and chemical aspects of the sedimentary record indicate low water levels and periodic dessication occurred between 9200 and 4500 BP, with a major period of dessication evident between 5500 and 4500 BP (Last and Teller 1983).

Sedimentary records from lake basins in central and southern Saskatchewan are truncated in mid-Holocene time (Last and Slezak 1988). This suggests that the low water interval registered at Lake Manitoba was a regional event, precluding the existence of lakes in this more arid environment during the early Holocene. Waldsea Lake, a deep (14.5 m), saline, meromictic, internally-drained lake in central Saskatchewan, contains a sedimentary record spanning the last 4000 years (Last and Schweyen 1985). Physical, mineralogical, and biological aspects of this record were documented in 31 core sections, with temporal control developed from conventional radiocarbon determinations on organic sediment. The lake's early history was punctuated by periodic dessication, as shown by deposition of abundant mirabilite crystals, *Artemia* (brine shrimp) eggs, and vegetation mats in near-basal sediment (the presence of lithified crusts, carbonate sands and grits, and gypsum grains is also indicative of periodic draw-down). By 3000 BP, organic-rich clays with an abundance of irregularly spaced aragonite laminae were deposited, suggesting that anoxic conditions prevailed at the sediment-water interface in a relatively deep, stratified lake. The lake has remained at reasonably high levels since 3000 BP, with the exception of a low-water stand between 2500 and 2000 BP (as indicated by an abundance of gypsum silt/sand laminae, calcareous hardgrounds and interclasts, and organic fibre mats in coarse-textured sediment) and a brief low water stand at about 700 BP (as suggested by basin-wide deposition of a coarse clastic layer). A pollen diagram developed from a core taken in the central basin area displays little variation over the last 4000 years, with the exception of a brief interval of marginally-increased Gramineae and reduced *Pinus* percentages between 2500 and 2000 BP, suggesting that increased local grassland openings may have accompanied this low-water stand.

Nearby Deadmoose Lake has been investigated by similar techniques, although the irregular morphology of this closed basin produced a more complex stratigraphic sequence than Waldsea Lake (Last and Slezak 1986). Although sediments are inadequately dated, preliminary results point to an early phase of lower-than-present water levels, likely during the mid-Holocene. In addition, the presence of coarse clastics, carbonate-cemented hardgrounds and crusts, vegetation mats, and gypsum laminae in cores from shallow lake margins suggest lake level dropped about 10 m, creating a series of satellite basins at 1080 ± 110 BP (a conventional radiocarbon date on wood fragments). As at Waldsea Lake, the Deadmoose Lake pollen record displays little systematic variation.

The sedimentary record of a playa lake basin in southern Saskatchewan has been explored through analysis of five core sections and multiple auger holes (Last 1990, 1987, Last and Slezak 1988). Ceylon Lake is a shallow (maximum depth is 2.1 m), closed basin that currently experiences pronounced seasonal water level and salinity changes (TDS may undergo as much as a four-fold increase from May to August), stimulating massive salt precipitation. The sedimentary record outlines three developmental phases.

The presence of calcite and absence of Mg-bearing carbonates in basal calcareous clay suggests Ceylon Lake was a relatively fresh, deepwater lake in the early postglacial period. Massive sediment with high organic matter content and abundant shells indicate productivity was high and bottom-dwelling organisms existed. Overlying black anoxic silty-clay, containing occasional gypsum laminae and intrasedimentary salts, represents a shallow, brackish-water phase of the lake. Carbonate hardgrounds in this unit indicate periodic drying occurred. The transition from silty clay to salt (mainly mirabilite and thenardite) in the upper 3 m of the record documents the change from a mud-dominated to a salt-dominated playa. Recent standard radiocarbon determinations date the onset of salt deposition at approximately 5000 BP (Last 1990).

In central Alberta, Lake Wabamun has been the scene of intensive paleoecological study for the past two decades (Fritz and Krouse 1973, Hickman et al. 1984, Holloway et al. 1981). Despite lacking a precise chronology, analyses of a number of indicators in several cores from this large, relatively shallow lake outline broad-scale Holocene water level changes. Fossil pollen and diatom stratigraphies, as well as changes in sediment lithology, indicate that a low water stand occurred from between 8800 to 4200 BP, with lowest lake levels achieved between 7400 and 4200 BP.

While questions concerning the Wabamun Lake chronology limit paleohydrological interpretation, several other basins with more secure chronologies have been studied in central Alberta. Fossil pollen and diatom stratigraphies, combined with analyses of sedimentary pigments and lithology have been employed in many investigations. Although the research has not been aimed primarily at documenting past lake level fluctuations, it is a valuable addition to the paleohydrological record.

The Moore Lake record (Anderson et al. 1989, Schweger et al. 1981, Schweger and Hickman 1989) outlines major early to mid-Holocene paleohydrological developments in central Alberta. The 9.85 m long sedimentary core consists mainly of laminated clay-gyttja. Near-basal sediment, dated by conventional radiocarbon means, indicates the lake began filling around $11,300 \pm 170$ BP. From 9200 to 5800 BP, the presence of *Ruppia* pollen (its only appearance in the record) suggests low water levels and increased salinity, as *Ruppia* is restricted to saline basins of the province today (Husband and Hickman 1985). A shift from planktonic to epipelagic and epiphytic diatoms during this period also indicates low water levels, as does the reduction of pyrite deposition, due to oxygenation of the previously anoxic sediment-water interface. These changes suggest a water level reduction of some 15 m occurred throughout the early to mid-Holocene.

Other lake basins in central Alberta experienced similar postglacial water level reductions. Lake Isle was reduced in depth between about 8000 and 4000 BP (Hickman and Klarer 1981), an interpretation based on rising calcium and total carbonate levels in addition to the appearance of *Ruppia* pollen. At Lofty Lake, increased Cheno-Am pollen deposition between about 8500 and 7000 BP (Lichti-Federovich 1970) is suggestive of lake lowering, and recent stratigraphic evidence indicates the lake was considerably reduced in size between 8700 and 6200 BP (Schweger and Hickman 1989).

Conventional radiocarbon determinations on basal sediment from both the central and shallow northeastern basin of Hastings Lake indicate that the main lake began filling by 4800 BP, but the northeastern basin was not flooded until about 3900 BP. Fluctuating, low water intervals during this period are suggested by irregular inorganic sedimentation in near-basal sediment (Forbes and Hickman 1981).

Many other central Alberta lakes experienced a similar history of early Holocene dessication followed by mid-Holocene filling. A recent survey of paleohydrological research in the region (Schweger and Hickman 1989) summarizes basal radiocarbon dates, sediment lithology, pigment content and chemistry, as well as pollen and diatom stratigraphies from the above-mentioned basins and others to outline regional patterns of lake development. Data from 28 sites across central Alberta indicate that a significant lowering of the regional water table occurred during the early Holocene. Basins deeper than 8 m filled in the early postglacial period, but water levels dropped, dissolved salt content increased, and highest productivity levels were attained in the early Holocene. Shallow basins began filling by 7500 BP, but current levels were not attained until approximately 3000 BP. The use of hand-operated piston corers is no guarantee that the entire sedimentary record was recovered and reliance on conventional radiocarbon determinations of bulk sediment introduces imprecision to the chronology of events; however, Schweger and Hickman's (1989) summary of a variety of evidence from a number of basins outlines major, climatically-driven

paleohydrological changes in central Alberta.

Investigations along the eastern slopes margin of the northern mixed-grass prairie document a similar pattern of postglacial change. At Chalmers Bog, tephra deposition (presumed to be Mazama ash) coincides with pollen and stratigraphic evidence of lake lowering (Mott and Jackson 1982). Early to mid-Holocene lake lowering is also suggested by increased carbonate content in sediment from nearby Wedge Lake (Harrison 1988, MacDonald 1982). Chronological control in this case is provided by a conventional radiocarbon date on near-basal *Populus* wood fragments, and a tephra layer presumed to be Mazama ash (MacDonald 1982).

With the advent of AMS radiocarbon dating techniques extremely small samples of terrestrial plant remains can be dated, presenting an opportunity for substantial improvement over the traditional practice of dating bulk sediment (for example, see Brown et al. 1989). Three recent studies along the margin of the northern Great Plains employed AMS technology and offer refined chronologies, although they are not directed specifically at documenting past lake level change and analyses are confined to a single core from each basin.

Barnosky (1989) utilized pollen and plant macrofossils combined with sediment lithology to reconstruct postglacial vegetation change in central Montana. At Lost Lake, interbedded clay, silt, and sand (containing pockets of plant debris and angular clasts thought to represent mudcracks), with an abundance of *Ruppia* fruits and significant amounts of *Ruppia* and Chenop-Am pollen, were interpreted as representing a marshy and intermittently dry basin between 9400 and 8300 BP. Between 8300 and 6000 BP a shallow, saline, semi-permanent lake existed, as indicated by laminated sand, silt, and clay deposits with increased Chenop-Am pollen and an abundance of *Ruppia* and *Chenopodium rubrum* seeds. Lake infilling from 6000 to 3600 BP is suggested by reduced *Ruppia* seed input and the regular appearance of

Potamogeton fruits in finely laminated silt and clay. Decreased Cheno-Am pollen percentages, beginning about 3600 BP, imply a further increase in lake depth. Deposition of silty clay sediment containing nodules of gypsum throughout the last 3600 years indicate hypersaline conditions have persisted to the present.

Guardipee Lake was shallow and brackish during its early history, as thinly bedded marl and clay bearing an abundance of *Sarcobatus* and *Ruppia* pollen was deposited after 11,500 BP. The presence of *Potamogeton* pollen throughout this period indicates complete dessication did not occur. From about 9300 to 100 BP however, the lake was probably dry much of the time, since compact inorganic sediment, low pollen deposition, and the near absence of pollen from aquatic taxa (with the exception of *Ruppia*), is characteristic of these deposits (Barnosky 1989).

MacDonald (1989) utilized pollen and plant macrofossil stratigraphies to outline the paleoenvironmental history of a site near low elevation treeline in southwestern Alberta. Increased percentages and influx of Cheno-Am and *Typha* pollen at Toboggan Lake between 8400 and 7600 BP are suggestive of water level fluctuations as overall lake levels receded. Between 7600 and 5500 BP Cheno-Am influx declined slightly and *Typha* pollen disappeared from the record. MacDonald interpreted these changes as indicating "...complete dessication of some previously seasonal lakes and the stabilization of lakes at lower levels" (ibid: 164). From 5500 BP to the present, reduced Cheno-Am pollen input suggests filling of the basin has occurred.

Reasoner and Hickman (1989) used AMS dating of plant macrofossils for chronological control of sedimentary records from two high elevation sites in the southern Canadian Rocky Mountains. Although their pollen and plant macrofossil analyses were aimed primarily at reconstructing Holocene treeline movements, the diatom stratigraphy of Opabin Lake (now situated above timberline) outlines past lake level fluctuations. From 5500 to 3000 BP, diatom

numbers declined dramatically, indicating water levels had dropped to the point that the lake was ephemeral. A return to high diatom numbers, between 3000 and 2000 BP, is thought to represent refilling of the basin. A subsequent decline in diatom abundance at about 1000 BP (coinciding with the onset of predominantly clastic sedimentation), is indicative of increased turbidity, possibly a response to "...more extensive Late Neoglacial ice..." (ibid: 312).

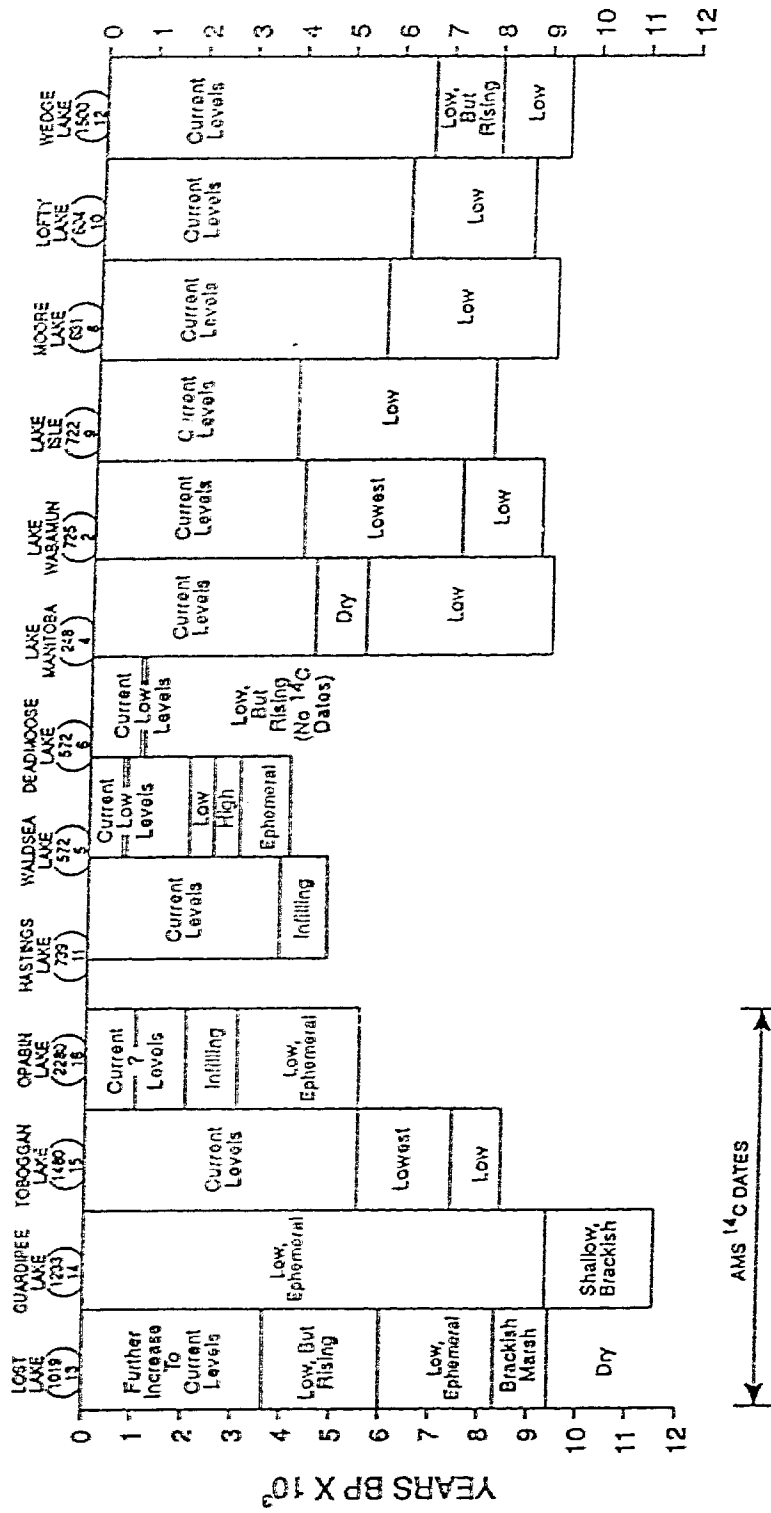
3.4 Summary of Holocene Lake Level Fluctuations

A summary of Holocene lake level changes described in the preceding section (Fig. 5) shows that all sites on the fringe of the northern mixed-grass prairie experienced a prolonged low water stand throughout the early Holocene. The onset and duration of this low water stand does not appear to be uniform. This is not surprising considering the variety of samples and radiocarbon techniques used, the variable response times of indicators utilized in the reconstructions, and the large geographic area under consideration.

Evidence from AMS radiocarbon-dated sites on the western grassland fringe (Guardipee, Lost, and Toboggan Lakes) suggests that this prolonged low water stand may have begun in late glacial time and persisted until about 6000 BP. Lake levels subsequently rose, culminating about 3500 BP, when current levels were attained. Opabin Lake, the lone high elevation site with an AMS chronology, contains evidence of a low water interval only in mid- to late Holocene time. The lack of an early Holocene low water interval at this site may be related to delayed local deglaciation.

In central Alberta shallow lake basins, like two closed lake basins in Saskatchewan, were dry throughout the early Holocene. Although lacking an early Holocene record due to high aridity, these catchments are well-suited for documenting late Holocene fluctuations as they are situated in a climatically-sensitive region. Subtle recent changes, such as rising lake levels between 4000 and 3000 BP and lake lowering around 1000 BP (recorded at Waldsea

Figure 5. Summary of postglacial lake level changes on the northern Great Plains. Elevation (m asl) and location on Figure 4 indicated in brackets below each site name.



and Deadmoose Lakes), suggest that they alone responded to what appears to have been a relatively minor change, compared to earlier events. The only other basin bearing evidence of lake level change at this time is Opabin Lake, located in a climatically-sensitive, high altitude environment.

3.5 Implications for Paleoclimatology

Whereas water level changes within a single basin cannot be considered a reliable indicator of climate change, regionally-synchronous lake level changes are driven by climate. In addition, if information is derived from closed basins, the record will be a sensitive barometer of climate (Street-Perrott and Harrison 1985). Regional lake level changes are clearly depicted in Figure 5. One test of their sensitivity to climate involves a comparison with an independently-derived record of Holocene climate. A paleoclimatic reconstruction based on vegetation change depicted in regional pollen studies provides such a test.

Summaries of postglacial vegetation and climatic change in the northwestern interior of North America (Barnosky et al. 1987b, Ritchie 1987, Vance 1986) outline a developmental sequence consistent with the lake level changes described above. Although relying mainly on radiocarbon chronologies based on dating bulk sediment samples by conventional means, the large number of study sites considered provide a consistent picture of the timing of major changes. As is the case with paleohydrological data, regional vegetation change must ultimately be driven by climate.

Following an early period of plant migration after deglaciation, grassland rapidly succeeded forest throughout much of the northwestern interior between 11,000 and 9500 BP (Mott and Christiansen 1981, Ritchie 1976, Schweger 1972). By 9000 BP, grassland extended north of its current limit in Manitoba (Ritchie and Lichti-Federovich 1968), Saskatchewan (Mott 1973), and Alberta (Schweger and Hickman 1989). This period of grassland expansion

(from 9000 to 6000 BP) was concurrent with rising upper treeline in the Rocky Mountains (Beaudoin 1986, Hills et al. 1985, Kearney and Luckman 1983, Luckman and Kearney 1986, Reasoner and Hickman 1989), increased fire frequency (MacDonald 1982, 1989), expanded grassland openings in low elevation subalpine forest (Hills et al. 1985, MacDonald 1989), reduced coniferous forest cover on the Cypress Hills (Sauchyn 1990), and development of xerophytic grassland on the northern Great Plains (Barnosky 1989). By 6000 BP, grassland throughout central Alberta, Saskatchewan, and southern Manitoba had reached its northernmost extension, and subsequently began to recede southward (Ritchie 1987). During the following millennia the upper limit of tree growth was lowered (Beaudoin 1986, Hills et al. 1985, Kearney and Luckman 1983, Luckman and Kearney 1986, Reasoner and Hickman 1989), grassland openings in subalpine forest (Hills et al. 1985, MacDonald 1989) and on the Cypress Hills (Sauchyn 1990) were reduced in size, fire frequency declined (MacDonald 1982, 1989), and the lower limit of tree growth was depressed (Barnosky 1989).

The northern limit of grassland in Alberta, Saskatchewan, and Manitoba reached approximately its current position by 3000 BP (Ritchie 1987, Vance et al. 1983). Since then, varied responses are evident at some sites, presumably depending upon the sensitivity of local vegetation to climatic change. High altitude sites, near the temperature-sensitive upper limit of tree growth, convey a somewhat confusing signal over the last 5000 years. Some sites display little change once treelines were established near their current limits about 4500 BP (Kearney and Luckman 1983, Luckman and Kearney 1986). In contrast, *Picea/Pinus* pollen ratios at other sites are suggestive of higher than present treelines until 1500 BP, followed by a minimum in elevation at 1000 BP (Beaudoin 1986, Reasoner and Hickman 1989).

The sequence of vegetation change described above reflects significant paleoclimatic events. By 9000 BP, rising treelines imply temperatures were warmer than present (Luckman and Kearney 1986). This is also suggested by the northward movement of the climatically controlled grassland-aspen parkland ecotone, although decreased precipitation was also likely a

factor (Ritchie 1987).

An early to mid-Holocene period of increased warmth and reduced precipitation would also account for increased fire frequency, expanded grassland openings in subalpine forest on the eastern flanks of the Rocky Mountains, and reduced coniferous tree cover on the Cypress Hills. However, a period of decreased lower treeline limit at Toboggan Lake suggests that increased precipitation may have interrupted this prolonged dry spell, at least in southwestern Alberta, between 9400 and 8400 BP (MacDonald 1989).

By 6000 BP peak aridity had passed. Declining upper treelines signify reduced temperature. Southward movement of the grassland-aspen parkland ecotone is consistent with a reversal of early Holocene temperature and precipitation trends. Modern distribution of vegetation zones was achieved by 3000 BP, signifying the onset of a temperature and precipitation regime similar to present. This conclusion may be somewhat misleading however, since minor climatic events in the late Holocene may not have brought about the scale of vegetation change required for registration in pollen stratigraphies, except at sites located in climatically-sensitive regions. Upper treelines are such sites, and evidence from these localities reveal that changes have occurred during the last 3000 years.

Pollen data have also been used to estimate the magnitude of temperature and precipitation changes that occurred during postglacial time. Multivariate analyses of modern pollen and meteorological data allow development of multiple regression equations, which then may be applied to fossil pollen assemblages to quantitatively reconstruct aspects of past climate (Webb and Bryson 1972). This technique, when applied to the Lofty Lake pollen data (Vance 1986), suggests that growing season temperature (May through August) rose rapidly in the early Holocene, attaining values some 10% above present between 10,000 and 6000 BP, before declining slightly to 3000 BP, and gradually decreasing to present levels by about 1000 BP. Growing season precipitation rose rapidly in the late glacial, but declined abruptly in the

early Holocene, reaching minimum values (some 18% below present) between 8000 and 6000 BP. A gradual rise from this minimum brought growing season precipitation to current levels by 4000 BP, with little change evident to the present.

Ritchie (1983) employed similar techniques to outline past growing season temperature and precipitation changes in the Riding Mountain area of Manitoba. Results are similar to those from Lofty Lake, with the period between 10,000 and 6500 BP characterized by summer temperatures approximately 20% warmer than present and effective precipitation reduced by 10% to 20%, compared to today. Precipitation rose to current levels by 6000 BP and temperature declined from an early Holocene maximum, but remained above present to 3000 BP, with a rapid decline evident at 2000 BP. Even though these studies are preliminary efforts, they are important as they offer insight into the magnitude of change in the early to mid-Holocene, as well as outlining possible differences between past climatic regimes within the western Canadian interior.

Zoltai and Vitt (1990) also reconstructed quantitative aspects of postglacial climate based upon the distribution of basal radiocarbon dates from peatlands in the western interior of Canada. Distinct differences in the distribution of peatlands older and younger than 6000 BP indicate that mean July temperature was slightly higher than present and growing degree days had increased by some 6% to 20% compared to today prior to 6000 BP, likely a consequence of a longer growing season. In addition, precipitation was considerably lower than present (19%) on the Saskatchewan Plain, but 5% higher than present in Manitoba during early to mid-Holocene time.

3.6 Critical Gaps in the Paleoclimatic Record

Although this review conveys a reasonably consistent pattern of major postglacial climatic trends, many details of the record require further investigation. Outstanding issues that are best addressed with a high resolution proxy record from the climatically-sensitive northern mixed-grass prairie include:

- (1) The nature and pace of past climatic change in the drought-prone grassland biome. To date, there are few proxy records for the region; fewer still extend beyond 4000 BP. Although major changes are apparent in records from sites on the grassland perimeter, little direct evidence of Holocene variability within the grassland region exists. A proxy record from the region would help fill this major gap in the data base.
- (2) Although the warm and dry climate of the early to mid-Holocene is the most clear-cut signal in the proxy data, many details of this widespread, extended drought remain unknown. Was this period punctuated by brief intervals of less severe aridity? If so, how long were they and when did they occur?
- (3) Middle to late Holocene time was evidently marked by declining aridity, although it is not clear when present climatic conditions were established. Evidence from sites bordering grassland suggests current conditions were attained by 6000 BP (MacDonald 1982, 1989, Ritchie 1987), whereas data from one site within grassland suggests warmer conditions prevailed until 3600 BP (Barnosky 1989), and evidence from high elevation sites suggest temperatures remained above current levels until as late as 1500 BP (Beaudoin 1986, Reasoner and Hickman 1989). In addition, the precipitation record for this period conveys a mixed signal. The record from one high elevation site suggests decreased precipitation from 5000 to 3000 BP (Reasoner and Hickman 1989), whereas lower elevation sites bear evidence of precipitation increases at this time (Last and Schweyen 1985, Last and Slezak 1986).

(4) Although more subtle late Holocene climatic changes are registered only at sites near upper treeline growth and on the arid Saskatchewan Plain, the signal is varied. A depression in upper treeline about 1000 years ago, combined with evidence of glacial activity (Reasoner and Hickman 1989), suggests a cooler, moister climate. However, coincident lake level lowering in Saskatchewan (Last and Slezak 1986) is indicative of increased temperature and/or reduced precipitation. Additional information on this recent period must be derived from a climatically-sensitive area, as few sites register the change.

(5) A critical question concerning prospects for future climate is how prehistoric climatic events compare to historic climatic extremes. For example, were the droughts of the 1880's and 1930's of similar magnitude and extent, compared to earlier events outlined in the proxy record? A high resolution paleoecological record is required to address this question.

CHAPTER 4

STUDY SITE

4.1 Site Selection

As noted in the introduction, continuous long-term sedimentary records from the northern mixed-grass prairie are rare, as drought periodically dries most wetlands. As a result, the survey for suitable sites in this study focused on spring-fed basins, since it is possible that groundwater inflow would protect the catchment from periodic dessication.

During the summer of 1983, previously inventoried springs exhibiting continuous, vigorous discharge (Borneuf 1971, 1983, Borneuf and Stevenson 1970, Tokarsky 1973) in association with wetland deposits in southern Alberta were located and assessed in terms of their suitability as paleoecological study locales. Coring was carried out in 1983 and 1984 at the eight most promising sites. All sites, with the exception of Chappice Lake, contained short sedimentary records (usually less than 2 m, Table 1), confirming that most are short-lived, likely drying completely during drought in the late 1800's, 1920's or 1930's. The lengthy core section obtained from Chappice Lake was the sole long-term sedimentary record recovered from the area of interest.

Table 1. Spring-fed catchments sampled during study site survey.

Site	Lat.	Long.	Core Length (m)
Hammond's Pond	49° 42'	113° 48'	1.5
Rhinehart Lake	49° 06'	113° 08'	1.7
Tyrell Lake	49° 23'	112° 15'	1.8
Forty Mile Coulee	49° 42'	111° 30'	0.7
Bow Island Pond	49° 46'	111° 24'	0.6
Ellis Spring	50° 10'	110° 42'	1.5
Old Channel Spring	50° 12'	110° 38'	1.4
Chappice Lake	50° 10'	110° 22'	9.75

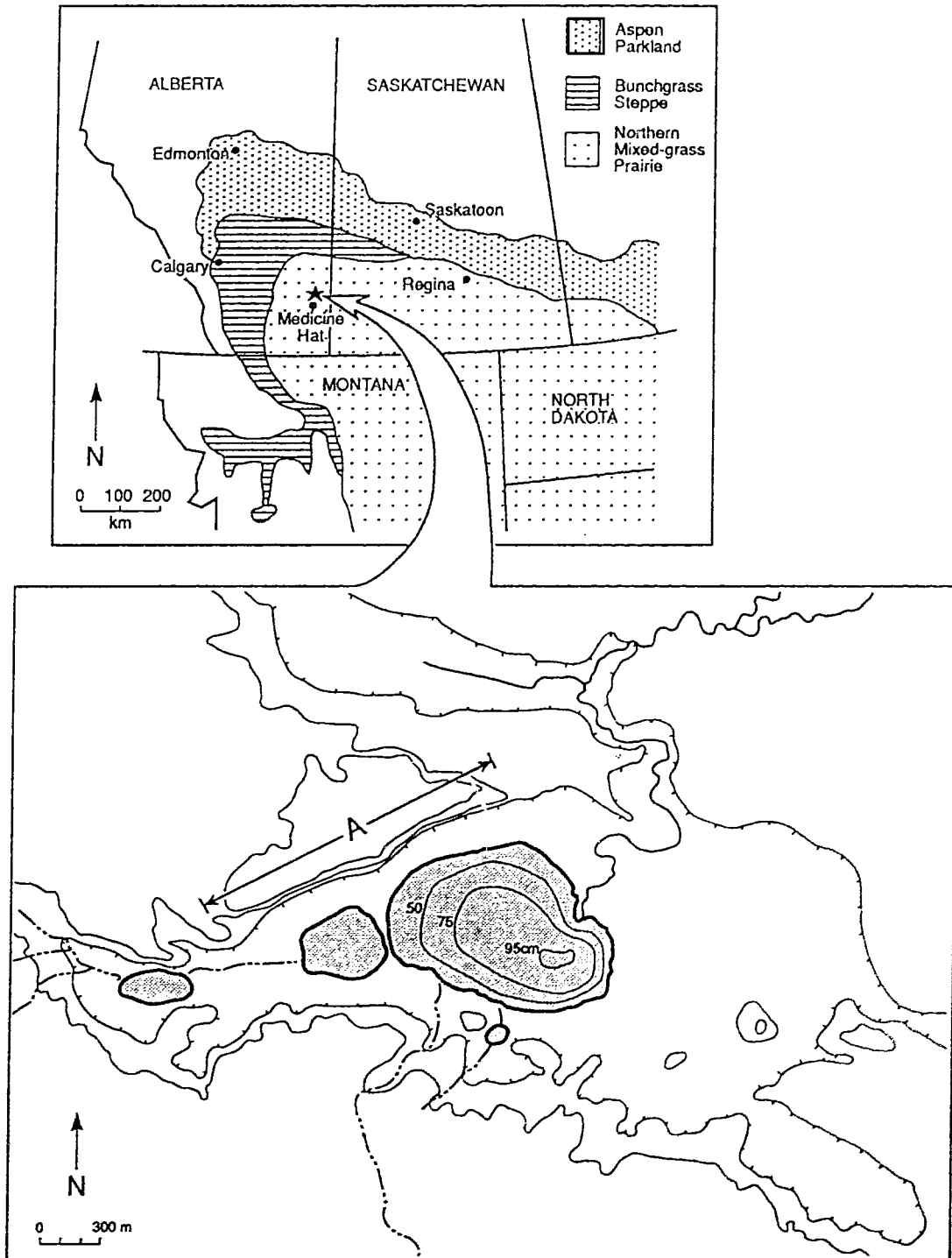
4.2 Bathymetry and Chemistry

Chappice Lake is relatively large (1.5 km²), flat-bottomed, and currently separated into two water-bodies by a low relief sand ridge (Fig. 6). Although the lake is situated within a glacial meltwater channel (731 m asl), the margins of the basin have rather gentle slopes, with the exception of steep-sided bluffs (some 30 m in height) flanking the northwest margin of the lake (bar A in Figure 6) and low elevation knolls lining the lake's southeastern margin.

Besides sheetflow, the only path of surface flow to the lake today is via a small stream draining roughly 100 km² of gently rolling topography north of the basin. Topographic maps based on aerial photography taken in 1970, when the lake was at a high water stand, show a permanent and temporary stream draining slopes south of the basin, as well as a network of temporary streams entering the lake from the west. Over the past 40 years streamflow has been impeded by construction of a number of small dams on many of the inflowing stream systems (the exact construction dates are unknown, although some appear on aerial photography taken as early as 1955). Surface flow to the lake was severely restricted in October of 1978 when a dam was built on the main inflowing stream within 1.5 km of the lake. In addition to surface flow, subsurface inflow is evident, as seepage was noted at several points around the lake perimeter in 1988 and 1989. These regions of low but continuous flow are readily identified as they support patches of vigorous plant growth. Chappice Lake has no surface outlet.

Samples of near-surface water were collected in 1 L polyethylene bottles during January and August of 1989. These samples were submitted to Quanta Trace Laboratories of Burnaby, B.C. for determination of ionic chemistry, pH, and conductivity (Table 2). Results show that Chappice Lake currently contains alkaline, hypersaline, sodium sulfate-type water. Sodium sulfate concentration had evidently reached saturation by 1989, as precipitation of a 10–15 cm

Figure 6. Chappice Lake location map, bathymetry, and surrounding topography (as outlined by the 7.6 m contour interval). Path of the main inflowing streamcourse is indicated by a solid line, dashed lines mark positions of ephemeral streams. Bar A demarcates the steep-sided bluff referred to in text.



layer of thenardite crystals (see Chapter 6) now occurs beneath winter ice. This accounts for the increased salt concentration in the winter water sample, compared to the sample collected in summer.

Table 2. Chappice Lake water chemistry.

	August, 1989	January, 1989
Physical Tests		
pH	8.8	8.5
conductivity ($\mu\text{S}/\text{cm}$)	80000	54000
dissolved solids (ppm)	159000	73100
Anions ($\mu\text{g}/\text{ml}$)		
Chloride Cl	7680	8750
Fluoride F	<500	<500
Nitrite $\text{NO}_2\text{-N}$	<500	<500
Nitrate $\text{NO}_3\text{-N}$	<50	<50
Phosphate $\text{PO}_4\text{-P}$	<2000	<2000
Sulfate SO_4	74800	23200
Alaklinity as CaCO_3 ($\mu\text{g}/\text{ml}$)		
Carbonate CO_3	1400	600
Bicarbonate HCO_3	4100	5900
Major Elements ($\mu\text{g}/\text{ml}$)		
Aluminum Al	<0.5	<0.5
Barium Ba	<0.05	<0.05
Calcium Ca	29	40
Iron Fe	1	1
Lithium Li	4	5
Magnesium Mg	2330	2590
Manganese Mn	<0.05	<0.05
Phosphorus P	<2	<2
Potassium K	1800	2090
Silicon Si	2.1	3.4
Sodium Na	37200	13700
Strontium Sr	0.22	0.28
Titanium Ti	0.05	0.05
Zirconium Zr	0.1	0.1

4.3 Climate

In general, climate throughout the northern mixed-grass prairie is uniform and truly continental with cold winters, short but warm summers, and a light precipitation regime. The Medicine Hat mean annual temperature (1941–1970) is 5.1°C (a minimum of -12.1°C in January and a maximum of 20.2°C in July) and yearly precipitation averages 348 mm (Hare and Thomas 1979).

Southeastern Alberta receives approximately 70% of its total yearly precipitation as spring rainfall (the highest monthly precipitation between 1941 and 1970, 63.5 mm, was recorded in in June), followed by a dry summer and fall. Droughts may occur at any time of the year, but historically are most common in July and August (Borchert 1950). Strong westerly and southwesterly winds are characteristic of the region (Longley 1972).

The character of southeastern Alberta's climate is largely dictated by its position on the eastern flank of the Rocky Mountains (Longley 1968). Westerly hemispheric upper air flow pushes Pacific air masses toward the Rocky Mountains causing orographic uplift and substantial moisture loss. These dry, modified Pacific air masses pass over the crest of the Rockies and warm as they descend into Alberta.

High altitude westerly winds influence air mass movements in temperate regions throughout the world, and are therefore an important determinant of local temperature and rainfall patterns (Lamb 1972). The position and strength of westerly airflow is, of course, variable. One source of variation is produced by seasonal migration, a movement that alters the mean position of air masses over the continent. In summer, high altitude westerlies move northward, restricting Arctic air to the north, thereby ensuring a near-continuous flow of mild, dry Pacific air into southern Alberta. Summer rainfall, mainly the product of convective heating, is highly localized but can be intense during short-term storms (Longley 1972).

In winter, westerly winds shift southward. As a result, Arctic air more frequently reaches southern Alberta, its arrival often heralded by snowstorms and followed by spells of frigid temperatures under clear skies. Incursions of Arctic air are generally short-lived however, as most systems are moving away from the Rockies, steered by upper atmospheric air flow. As a result, Pacific air generally returns to Alberta and temperatures rise. Particularly rapid temperature increases are produced when cold Arctic air is replaced by warm Pacific air, producing a "chinook" wind (Nkemdirim 1986). Southwestern Alberta experiences the greatest number of chinooks in western Canada, and these outbreaks of Pacific air may result in temperature increases of 7 to 17°C within a few hours (Hare and Thomas 1979).

As most precipitation in Alberta results from storm centers that develop along the boundary between Arctic and Pacific air (Longley 1972), seasonal movements of the steering upper westerlies produce seasonal precipitation maxima in different regions of the province. Peak June precipitation in southern Alberta marks the time when the Pacific/Arctic air mass boundary is situated over the northern Great Plains. Severe spring snowstorms also occur during westerly movement, generated by periodic incursions of Arctic air as the westerlies move northward.

In addition to seasonal shifts in upper westerly flow, long-term fluctuations in amplitude and position have been documented (Lamb 1972). Blocking, caused by a reduction in upper westerly windspeed and a consequent north-south meandering pattern of airflow (referred to as a meridional circulation pattern), may result in the persistence of a high pressure ridge over the northern mixed-grass prairie. Meridional circulation produced the summer droughts of 1961, 1967, and 1988 (Dey 1982, Trenberth 1989, Trenberth et al. 1988) and the winter drought of 1976 (Dey 1982). The opposite extreme in circulation, that is, high speed upper westerly flow restricting north-south air mass movement (a zonal circulation pattern), drives a near-continuous stream of Pacific air into the northern Great Plains. Zonal circulation

prevailed during the late 1800's and 1930's, periods characterized by year-round high winds and aridity (Borchert 1950, 1970, Lamb 1982, Namias 1983). In other words, severe drought is not linked to "...either a zonal or meridional circulation but, rather, with particularly great persistence of either type" (Borchert 1970:8).

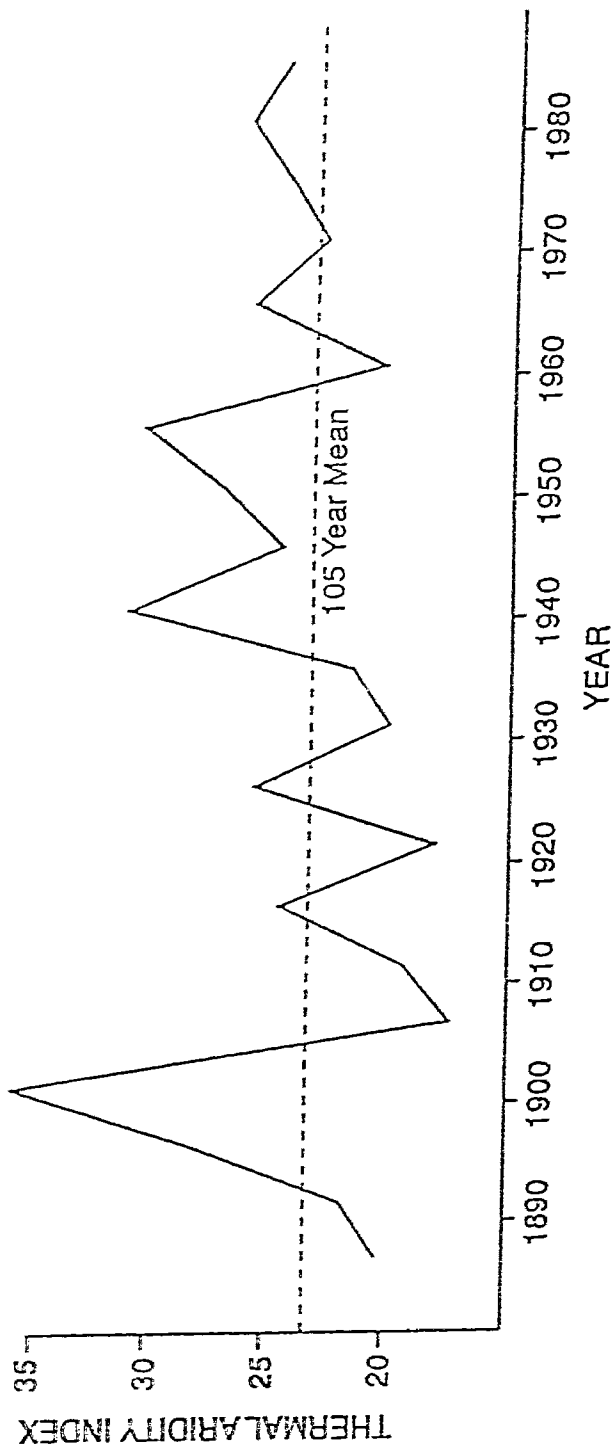
The mean yearly temperature and precipitation of the 105-year Medicine Hat meteorological record (one of the longest in western Canada) is 5.4°C and 336 mm, respectively. Calculation of a thermal aridity index in five-year intervals from this data (Fig. 7), offers insight into historic climatic variations in the study area. By this measure, the droughts of the 1980's do not appear as severe as those that occurred in the first three decades of this century or during the early 1960's. In contrast to severe drought in 1908, the early 1920's, and throughout the 1930's, the years at the turn of the century, as well as the 1940's and 1950's, were relatively cool and moist.

4.4 Vegetation

Chappice Lake is located near the northern boundary of the mixed-grass prairie (Fig. 6), an essentially treeless plain possessing a near-continuous cover of grass and sedge, with forbs and shrubs occurring as associated subdominants (Clayton et al. 1977). In Canada, the mixed-grass prairie occupies an area of about 260,000 km² (ibid), forming a semi-circular arc along the international border in Alberta and Saskatchewan. Although the vegetation has been altered by historic practices of grazing, cultivation, and fire suppression, a description of little-disturbed, pre-agricultural vegetation of the 1940's is provided by Coupland (1950, 1961). Nomenclature throughout this section follows Moss (1983).

Coupland (1950, 1961) noted *Stipa comata* (spear grass), *S. curtisetia* (barkworth), *Bouteloua gracilis* (grama grass), *Agropyron smithii* (western wheat grass), *A. dasystachyum* (northern wheat grass), and *Koeleria macrantha* (june grass) as the dominant grasses. The

Figure 7. Five-year mean thermal aridity index calculated from the Medicine Hat meteorological record.



most abundant sedge was *Carex stenophylla*. These species comprised 84% to 95% of the total cover.

The dominant forbs included *Artemisia frigida* (pasture sagewort), *A. ludoviciana* (prairie sagewort), *Phlox hoodii* (moss phlox), *Sphaeralcea coccinea* (scarlet mallow), *Anemone occidentalis* (chalice flower), *Gutierrezia sarothra* (broomweed), *Haplopappus spinulosa* (spiny iron plant), and *Heterothera villosa* (golden aster). The most common shrubs, restricted to areas of sufficient moisture, were *Rosa arkansana* (prairie rose) and *Artemisia cana* (sagebrush). *Selaginella densa* (club moss) formed dense mats in the dry prairie understory. On very dry sites, *Coryphantha vivipara* (ball cactus) and prickly pear cactus (*Opuntia fragilis* and *O. polyacantha*) were common.

4.5 Soils

Brown chernozems are the characteristic soils of the mixed-grass prairie (Clayton et al. 1977). About 120 km north of Chappice Lake, Brown Chernozemic soil grades into dark brown chernozems underlying bunchgrass steppe, and ultimately black chernozems of the aspen parkland (approximately 300 km north of Chappice Lake).

Local occurrences of saline soil within the Brown Chernozemic soil zone are readily distinguished by the presence of salt-tolerant species, including *Distichlis stricta*, *Hordeum jubatum*, *Sarcobatus vermiculatus*, *Salicornia europaea* var. *rubra*, and *Suaeda* spp. (ibid). All of these halophytes are found in the Chappice Lake basin (see Chapter 5). Wyatt and Newton (1926) noted that rolling topography in the vicinity of Chappice Lake was characterized by accumulations of fine-grained sediment in low lying areas, whereas gravels were mainly restricted to knolls. They suggested that "...the rather sandy mixed area around Sam and Chappice lakes was undoubtedly formed as a result of the sorting action of the lake water..." (ibid: 33), although meltwater discharge likely transported much of the sand to

the area. As a result, soil is poorly developed in the Chappice Lake basin and dry prairie vegetation is established in saline sand. On knolls surrounding the lake above the meltwater channel floor, Brown Chernozemic soils are better developed.

4.6 Geology

Bedrock units underlying glacial drift in the study area are mainly sandstone and shale, remnants of a series of marine transgressions. Three Late Cretaceous formations are found in the Chappice Lake catchment (Stevenson and Borneuf 1977). The oldest, the Foremost Formation, is composed of brown and gray sandstone, siltstone, and shale containing fossils of marine and brackish water molluscs (Barendregt 1988). It is overlain by the Oldman Formation, a pale gray to light greenish gray sandstone and shale dominated by freshwater invertebrate fossils. These two formations record regression of the Pakowki Sea (Russell and Landes 1940). An abrupt contact with the uppermost Bearpaw Formation, a dark gray to brownish gray shale, marks the final transgression of the Bearpaw Sea.

Following recession of the Bearpaw Sea, uplift of the Rocky Mountains during early Cenozoic time ended almost 100 million years of cyclic inundation and recession of inland seas in southern Alberta (Beaty 1972) and initiated 25 to 30 million years of erosion that moulded the major topographic features of today's landscape. Topographic highs, such as the Milk River Ridge and the Cypress Hills, survived this erosive period due to protection afforded by resistant deposits on easily eroded Cretaceous bedrock (Barendregt 1988).

Although prominent features of the regional landscape are products of these early events, recent continental glaciations were a major force in shaping details of the current topography, particularly in the Chappice Lake catchment. Although glacial extent is delimited by landforms, timing of the most recent retreat of the southwestern margin of the Laurentide ice sheet remains unclear, due to a deficiency of suitable material for radiocarbon dating

(Stalker 1977) and the unreliability of radiocarbon determinations on available materials (Clayton and Moran 1982). Recent compilations of deglaciation dates suggest that ice retreated rapidly from its maximum extent (south of Chappice Lake) between about 13,000 and 8,000 BP (Dyke and Prest 1987). During retreat, a series of proglacial lakes formed along the glacial margin as the ice mass impeded the northeast trending drainage pattern. A large proglacial lake existed in the Chappice Lake area between 13,000 and 11,300 BP (Teller 1987) and is likely associated with the Fox Valley moraine (Christiansen 1979), which runs from Chappice Lake eastward into Saskatchewan. Continued ice recession and lake drainage produced a complex system of tributaries to the main drainage channel which runs west of Many Island Lake (Catto 1980). Chappice Lake (15 km west of Many Island Lake) is situated within rolling hills of a moraine, approximately 30 m above the surrounding prairie, within a remnant of this meltwater system.

4.7 History

With the exception of periodic visits by nomadic North American natives, occupation of the land surrounding Chappice Lake occurred relatively recently. The first Europeans arrived in the area during the late 1880's (Jones 1987), attracted by the vast expanse of grassland considered ideal rangeland for cattle. Although ranchers vigorously resisted early attempts to convert land to grain farming, by 1910 the area was swamped with a rush of settlers responding to calls from real estate boosters in southern Alberta, "...a land blessed of the Gods – a land over which bending nature ever smiles and into whose cradle she emptied her golden horn..." (Haste 1908).

Early grain farms were concentrated along the CPR railroad line 40 km south of Chappice Lake. The promise of a blessed land never materialized for most of these early farmers, as the good crop years of 1915 and 1916 were followed by severe drought years from 1917 to 1922 (Jones 1987). This disastrous turn in the weather stimulated a rate of

depopulation that rivalled the initial rush of settlement. Severe drought during the early 1920's was followed by the infamous dry years of the 1930's, causing further homestead abandonment. Although the dry years of the 1930's affected the entire central North American grassland, the drought of the early 1920's was most devastating in southeastern Alberta. As an example, the postmaster in the hamlet of Vale, 5 km north of Chappice Lake, once served some 50 local farmers – by 1924 only 3 remained (ibid.).

Chappice Lake was a centre of activity in 1921 when Charles Hatfield, a rainmaker who championed himself as 'master of the clouds', signed a contract with the Medicine Hat Agricultural Association to generate the greatest increase in precipitation possible, for the tidy sum of \$2000 an inch. Hatfield chose Chappice Lake as the venue for his endeavors as it was the largest body of water at the highest elevation in the vicinity of Medicine Hat during the dry early 1920's. Rain was so plentiful in the first few weeks following the rainmaker's efforts that local citizenry petitioned him to schedule downpours at night only. The rain did not last however. Heavy May rains were followed by hot, dry weather throughout the summer. Hatfield returned most of the money and was not invited to return the following year (ibid).

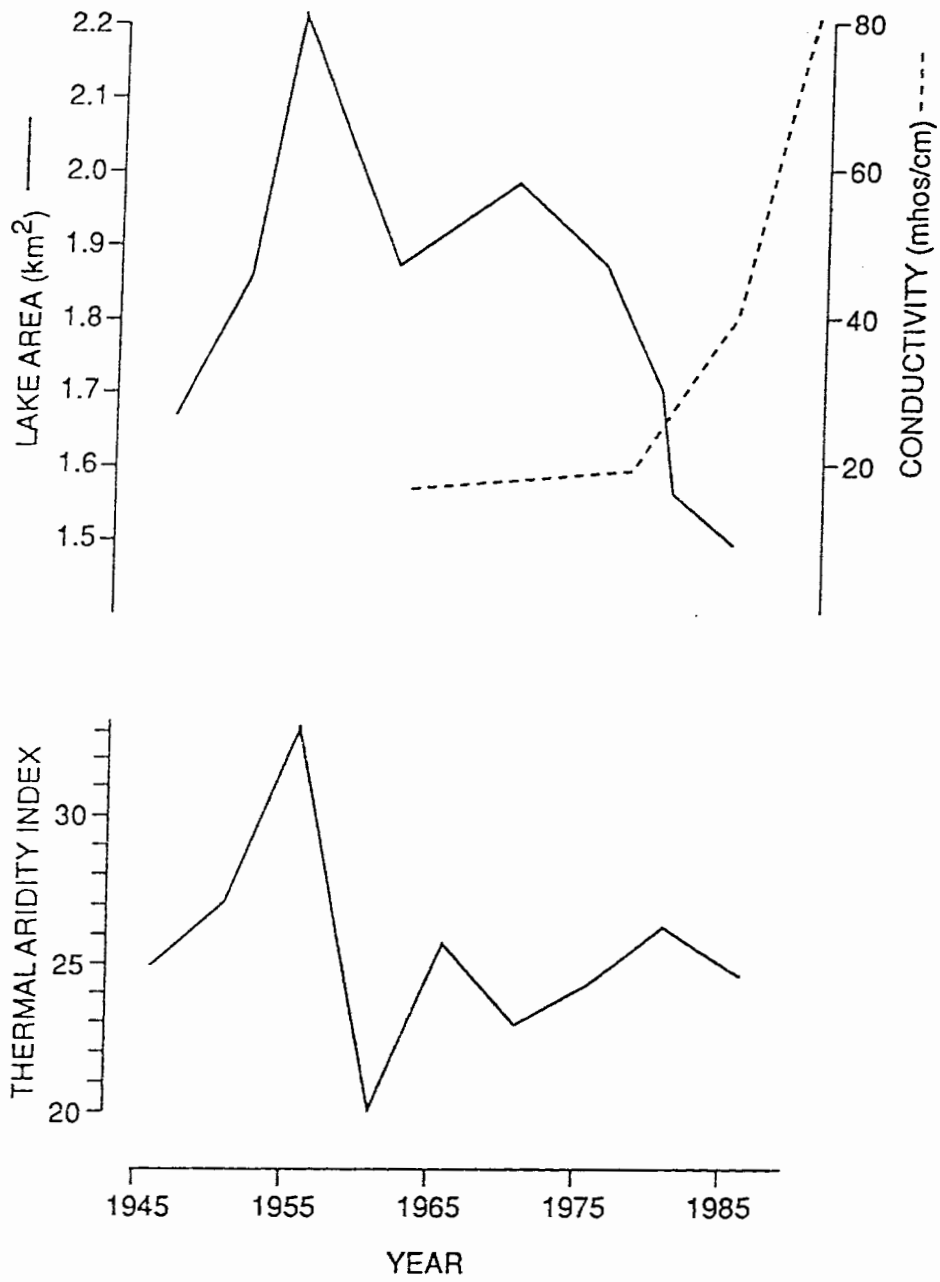
It is not clear exactly when farming began within the Chappice Lake watershed, although the area directly north of the lake was under cultivation in 1947 (the year of the earliest aerial photography). Sporadic air photo coverage to the present reveals that fields along the northern lake margin have been cultivated periodically since the late 1940's.

4.8 Historic Salinity and Water Level Changes

Historic water level changes in Chappice Lake have been documented using aerial photographs taken at various times of the year in 1947, 1952, 1955, 1962, 1970, 1976, 1980, 1981, and 1985. Lake area was determined on each photo set using a planimeter. This

information, when plotted with the Medicine Hat thermal aridity index and conductivity measurements (Hammer and Heseltine 1988, Hammer personal communication 1990), summarizes the lake's response to recent weather changes (Fig. 8). Recent damming of the main inflowing stream has affected the lake, causing a rapid drop in lake level and a rise in concentration of dissolved salts in the late 1970's. In spite of these modifications, lake level and salinity responses to climate are identical to those outlined in Chapter 2. Times of high rainfall and/or low temperature produce high lake levels. During the mid-1950's, for example, the now separate basins of Chappice Lake were connected, doubling current lake area. With reduced rainfall and higher temperatures of the early 1960's (producing a lower thermal aridity index), lake area declined. Warm and dry weather of the 1980's also resulted in lake recession and high salinities as salts were concentrated within a progressively smaller water-body. In addition, many small ponds and sloughs around Chappice Lake have dried up totally in recent years.

Figure 8. Historic water level and salinity changes at Chappice Lake compared to the five-year mean thermal aridity index of the Medicine Hat meteorological record.



CHAPTER 5

MODERN POLLEN AND PLANT MACROFOSSIL DEPOSITION

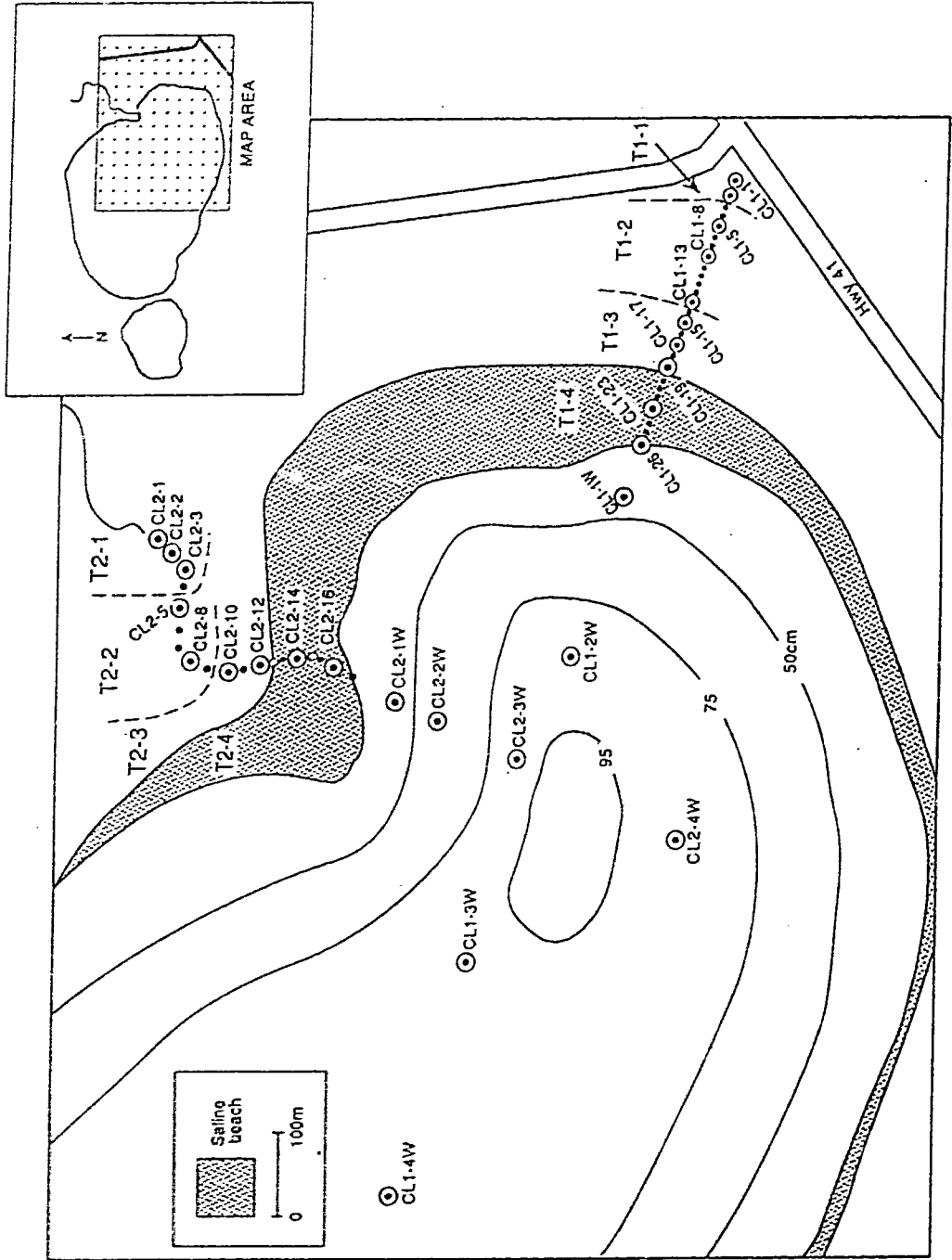
5.1 Introduction

In order to develop the most accurate interpretation possible of the fossil record, an appreciation of the processes involved in modern subfossil deposition at the study site is required. To evaluate the distribution and abundance of halophytic vegetation around Chappice Lake, two transects were surveyed in June of 1988. Both began in upland prairie vegetation and extended downslope toward the central basin area (Fig. 9). Surface samples were collected from within each sample plot along both transects to allow characterization of modern pollen and plant macrofossil deposition within each vegetation zone.

The location of transect CL1 was chosen to survey dry upland vegetation, several former beaches marking past high water stands, and the current shoreline. Extension of transect CL1 to the central lake area provides information on current lacustrine pollen and seed assemblages at increasing distances from shoreline. Variation in vegetation, pollen, and macrofossil assemblages from the central lake area to the upland prairie environment should thus provide an analogue of changes that would accompany lake dessication.

Transect CL2 followed the main inlet streamcourse to characterize source vegetation as well as modern pollen and plant macrofossil deposition along this transport route. As in CL1, transect CL2 was extended lakeward to assess the influence of streamflow on lacustrine pollen and macrofossil assemblages.

Figure 9. Location of transects and sample plots used during the vegetation survey. Pollen spectra were developed from sediment collected in all CL2 sample plots. Open circles in transect CL2 indicate samples analyzed for plant macrofossil remains. Open circles along transect CL1 indicate sample plots where pollen and plant macrofossil analyses were conducted.



5.2 Methods

5.2.1 *Vegetation*

A series of 1 m² sample plots were placed at regular intervals (15 m along CL1 and 30 m along CL2) throughout the terrestrial portion of each transect. Within each sample plot the occurrence of all plant species was noted along with visual estimates of percent cover for each species. A 100 ml surface sample (scraped from the upper 3 cm of sediment) was collected from within each sample plot. Lacustrine samples (100 ml) were collected from surficial sediment raised with an Eckman dredge.

Upland plants were identified following Moss (1983), while aquatic taxa were identified with the key developed by Scoggan (1978). Problematic identifications were verified with voucher specimens located in the University of British Columbia herbarium. Distinctions between many species of grass (Gramineae) and sedge (Cyperaceae) were not attempted since diagnostic floral parts had not developed at the time of sampling.

5.2.2 *Pollen and Plant Macrofossils*

Sediment samples collected during the vegetation survey were subsampled for pollen analysis. Four cm³ of the sandy sediment was required in most cases. The remainder of the sample was gently washed with tap water through nested screens (1 mm, 250 μm, and 180 μm openings). Plant macrofossils were picked from the organic debris on each screen and identified under a stereoscopic microscope. Identifications were made using keys, drawings, and photographs provided by Beijerinck (1976), Martin and Barkley (1961), and Montgomery (1977), as well as the macrofossil reference collection located in the palynology laboratory, Department of Biological Sciences, Simon Fraser University.

Pollen samples from alkaline prairie sediments require careful processing to ensure complete palynomorph recovery. Each acid wash must be carried out with caution as reactions

are often vigorous and may inhibit centrifugation. The following protocol consistently concentrated a sufficient number of microfossils from sediment containing greater than 2000 pollen grains/cm³.

To enable calculation of pollen concentration, a *Lycopodium* marker tablet (batch number 201890, 11,300±400 spores) was added to each sample and dissolved in warm 10% HCl. Repeated washes in warm 0.1M Na₄P₂O₇, were followed by a 10 minute hot KOH treatment. The remaining residue was washed with distilled water (dH₂O) through a 250 μm screen, and any material retained on the screen was examined for macrofossils. Most sand was removed by suspending the sample in 10 ml of dH₂O, swirling for 60 seconds and allowing the mixture to settle for 30 seconds. The supernatant was then immediately poured into a centrifuge tube, leaving behind the coarse pollen-free residue. Following centrifugation, the fine-grained, pollen-rich portion was placed in a hot HF bath for 1 hour and subjected to a 3-minute acetolysis treatment. Since lacustrine sediments contained a significant amount of fine clastic debris a warm 0.1M Na₄P₂O₇ wash was required, followed by gravity filtration through a 6 μm nitex screen (Cwynar et al. 1979). Dehydration with successive washes in 50%, 75%, and 95% ethanol preceded a final wash in tertiary butyl alcohol. Samples were mixed with silicone and evaporated at 40°C overnight. After stirring, the residue was mounted on a glass microscope slide.

All pollen counts were made at 400X with a Zeiss microscope. Pollen identifications were carried out with the aid of keys by Faegri and Iverson (1975), McAndrews et al. (1973), or Moore and Webb (1978), and verified using the pollen reference collection in the palynology laboratory, Department of Biological Sciences, Simon Fraser University. All pollen and spores were tabulated during a series of traverses (equally distributed between the margin and middle of the slide if pollen concentration was sufficient) until a sum of >300 upland pollen grains was attained. The Cyperaceae family (whose members occupy both upland and shoreline habitat in the area), aquatic plant taxa, and spores were tallied and their

frequencies calculated outside the pollen sum. Pollen and plant macrofossil diagrams were constructed using the Michigrana 86 graphics program (Futyma and Meacham 1986).

All pollen samples collected from transect CL2 were processed and counted. Since pollen spectra from within each vegetation assemblage in this transect displayed little variation, only 14 of 30 pollen samples from CL1 were analyzed (Fig. 9). Thirteen of 21 macrofossil samples in CL2 and 14 of 30 samples in CL1 were analyzed, as a distinctive macrofossil spectrum was evident for each vegetation assemblage at this point.

5.3 Results and Discussion

5.3.1 *Vegetation*

Vegetation along transect CL1 (Table 3) has been grouped into five assemblages based on the abundance of the dominant taxa *Opuntia*, *Scirpus*, *Distichlis*, *Puccinellia*, *Suaeda*, and *Ruppia*. These assemblages are numbered T1-1 to T1-5, from the upland fringe of the basin to open water, respectively (Fig. 9).

Grass dominates the T1-1 assemblage (*Agropyron pectiniforme* was a prominent species), exhibiting the highest cover in the transect. T1-1 is representative of a dry prairie upland, as indicated by an abundance of *Opuntia fragilis*, many of which were in flower at the time of sampling. Other species found in the two plots falling within this assemblage (*Artemisia frigida*, *Astragalus tennellus*, and *Lactuca pulchella*) contributed little to overall cover. *Coryphantha vivipara* and *Rosa acicularis* were abundant nearby (within 5 m).

In T1-2, dry prairie upland indicators decline, replaced by species reflecting increased soil moisture such as *Scirpus pungens* var. *pungens* and *Scirpus nevadensis* (Hammer and Heseltine 1988, Moss 1983). Increasing salinity, compared to T1-1, is indicated by *Distichlis stricta* and *Puccinellia nuttalliana* presence (Hammer and Heseltine 1988, Ungar 1974).

Table 3. Summary of CLL vegetation, with percent frequency and percent cover (in brackets) listed for each species.

	T1-1	T1-2	T1-3	T1-4	T1-5
<i>Opuntia fragilis</i>	50 (30)				
Gramineae spp.	100 (60)				
<i>Artemisia frigida</i>	50 (2.5)				
<i>Astragalus tenellus</i>	50 (2.5)	8.3 (<1)			
<i>Lactuca pulchella</i>	50 (2.5)	16.6 (<1)			
<i>Scirpus pungens</i> var. <i>pungens</i> }		83.3 (58)			
<i>Scirpus nevadensis</i> }					
<i>Sonchus arvensis</i>		25 (3.8)			
<i>Tragopogon dubius</i>		8.3 (<1)			
Compositae seedlings }		8.3 (<1)			
Chenopodiaceae seedlings }					
<i>Descurainia sophia</i>		16.6 (<1)			
<i>Bassia hyssopifolia</i>		8.3 (<1)			
<i>Puccinellia nuttalliana</i>		8.3 (1.6)	100 (35)		
<i>Distichlis stricta</i>		83.3 (17.6)	20 (4)		
<i>Suaeda calceoliformis</i>			60 (<1)		
<i>Salicornia europaea</i> var. <i>rubra</i>			20 (<1)		
<i>Ruppia maritima</i>				42.9 (10)	75
Number of sample plots	2	12	5	7	4
Average percent cover of type	90	82.5	22.8	0	0

Cover in T1-2 is virtually continuous, except near the T1-3 border, where barren sand patches are common. Species noted nearby T1-2 sample plots include *Opuntia fragilis*, *Ranunculus cymbalaria*, *Achillea millefolium*, *Artemisia cana*, *A. frigida*, *A. campestris*, *Salsola kali*, *Medicago alba*, *Taraxacum*, and Cruciferae spp.

A further increase in soil salinity is evident in T1-3, as overall cover declines from T1-2, and *Salicornia europaea* var. *rubra* becomes common, one of the most salt-tolerant species of western North America (Ungar 1974). The presence of *Suaeda calceoliformis*, combined with increased *Puccinellia nuttalliana* cover, is also indicative of extreme soil salinity (Dodd and Coupland 1966).

Evidently soil underlying T1-4 is too saline for even the most salt tolerant species, since vegetation cover is further reduced. Sparse patches of *Puccinellia nuttalliana*, outside the T1-4 sample plots and associated with groundwater seepage, were the only living living plants encountered. Clumps of *Ruppia* flotsam and colonies of Archaeobacteria occasionally interrupt the otherwise barren salt-encrusted sand.

Bottom samples raised with an Eckman dredge in the aquatic segment of transect CL1 contained only one aquatic macrophyte, *Ruppia maritima*. Several specimens collected were in flower.

Vegetation surveyed in transect CL2 displays a similar distributional pattern, although a gradient of increasing soil salinity from upland to shoreline is less well-defined due to intermittent stream discharge. The diverse vegetation of T2-1 (Table 4) is dominated by three high salinity indicator species: *Scirpus pungens* var. *pungens*, *S. paludosus*, and *Atriplex subspicata* (Liefers and Shay 1980, Moss 1983). Included in T2-1 are other halophytes, such as *Puccinellia nuttalliana* and *Sonchus arvensis* (Hammer and Heseltine 1988), as well as taxa noted nearby that are associated with marginally-saline to freshwater environments, like *Scirpus acutus* (Hammer and Heseltine 1988, Sloan 1970) and *Triglochin maritima* (Ungar 1974).

Table 4. Summary of CI.2 vegetation, with percent frequency and percent cover (in brackets) listed for each species.

	T2-1	T2-2	T2-3	T2-4	T2-5
<i>Scirpus paludosus</i>	25 (22.5)				
<i>Sonchus arvensis</i>	50 (10)				
Gramineae spp.	25 (7.5)				
<i>Juncus balticus</i>	25 (<1)				
<i>Atriplex subapicata</i>	50 (17.5)				
<i>Ranunculus</i> sp.	25 (<1)				
<i>Glaux maritima</i>	25 (<1)				
<i>Polygonum</i> spp.	25 (<1)				
<i>Scirpus pungens</i> var. <i>pungens</i>	100 (40)	40 (22.6)	100 (40)	40 (8)	
<i>Puccinellia nuttalliana</i>	11 (3.3)	20 (2)	100 (40)	40 (6.2)	
<i>Chenopodium salinum</i>		20 (<1)			
<i>Suaeda</i> seedlings	25 (1)		100 (3.3)	20 (2)	
<i>Atriplex prostrata</i>		20 (<1)	66.7 (10)		
Compositae sp.			33.3 (<1)		
<i>Salicornia</i> seedlings			33.3 (3.3)		
<i>Chenopodium</i> seedlings				20 (2)	
<i>Ruppia maritima</i>					75
Number of sample plots	4	5	3	5	4
Average percent cover of type	100	37.5	100	16	-

In T2-2, vegetation is reduced in diversity and cover compared to T2-1, although the overall reduction in cover is somewhat misleading as two plots fall within small ponds. Where vegetation is established, growth is vigorous. *Scirpus pungens* var. *pungens* dominates T2-2, with *Puccinellia nuttalliana*, *Chenopodium salinum* and *Atriplex prostrata* occurring as subdominants. Noted nearby the T2-2 sample plots were *Triglochin maritima*, *Scirpus acutus*, *Aster* sp., *Solidago* sp., *Eurotia lanata*, and *Medicago alba*. A mixture of species associated with varying salinities, as outlined in T2-1, continues in this assemblage.

The dense cover *Scirpus pungens* var. *pungens* and *Puccinellia nuttalliana*, combined with abundant *Suaeda* and *Salicornia* seedlings and reduced cover of marginal halophytes in T2-3, indicates that high salinity soil underlies this assemblage. The presence, but further reduced cover, of these species in T2-4 suggests a further increase in salinity. As in transect C1.1, the only macrophyte recovered from the lacustrine portion of transect C1.2 was *Ruppia maritima*.

5.3.2 Pollen and Plant Macrofossils

The pollen and plant macrofossil record from transect C1.1 (Fig. 10) summarizes current micro- and macrobotanical variability. The summary pollen diagram (Fig. 10a) includes only taxa encountered in the vegetation survey or recognized in plant macrofossil assemblages, although the frequency of each was calculated using a pollen sum that included all upland taxa. Thus, Figure 10a is a local pollen record; extralocal taxa, such as *Pinus*, *Picea*, *Populus*, *Betula*, and *Salix*, as well as other taxa that appear only sporadically in low amounts have not been included on the modern pollen diagrams.

The most distinctive aspect of T1-1 pollen spectra is the sole occurrence of Cactaceae pollen (10% in one sample). Liguliflorae type Compositae and Leguminosae pollen peak in abundance in T1-1 (5% for both), as do *Selaginella densa* spores (18%). These taxa may

therefore be considered indicators of a dry prairie environment.

Gramineae and Cyperaceae pollen dominate the T1-2 pollen spectra (22-58% and 8-36%, respectively), a reflection of their dominance in the current vegetation. Pollen concentration peaks in T1-2 (10,000 to 25,000 grains/cm³).

Pollen concentration drops to a minimum in T1-3 (<5,000 grains/cm³). Chenop-Am pollen attains its highest value of the transect here (35%), in accordance with its abundance in the sparse vegetation cover. *Ruppia* pollen consistently occurs in significant amounts throughout T1-3 (20-50%), likely recording a recent high water stand.

Pollen concentration remains low throughout T1-4 in pollen spectra dominated by *Ruppia*. Peak *Ruppia* percentages (60-75%) occur in T1-4, but drop to between 30-60% in T1-5 lacustrine sediment samples. *Artemisia* pollen becomes an important element of the modern pollen spectra for the first time in T1-4 and T1-5, reflecting its abundance in upland vegetation surrounding Chappice Lake. Total pollen concentration throughout T1-5 is generally less than 10,000 grains/cm³.

The peak in *Ruppia* pollen representation on the saline beach and in the sample collected closest to the existing shoreline (CL1-1W) is the most important aspect of modern pollen distribution, in terms of reconstructing past lake level changes. *Ruppia maritima* thrives in shallow, saline water (Hammer and Heseltine 1988) and this preference, combined with *Ruppia* flotsam washed ashore, must contribute to near-shore abundance. Thus, high *Ruppia* pollen percentages may be equated with near-shore, shallow, high-salinity conditions.

Patterns of pollen distribution between near-shore lacustrine samples and collecting sites more distant from shoreline are also apparent in three other native plant taxa. Chenop-Am and *Sarcobatus* are marginally more abundant in near-shore samples, while Labiatae, although only a minor constituent of the modern pollen rain, appears only in the sample collected adjacent to the current shoreline. As a result, Chenop-Am, *Sarcobatus*, and Labiatae may be

considered indicators of proximity to shoreline. Since none display dramatic increases in near-shore samples however, they must be regarded as weak indicators. Other taxa commonly found in shoreline vegetation, such as grass and sedge, do not exhibit high pollen values in near-shore sediment. This is likely due to the prevalence of these taxa in the mixed-grass prairie, as the contribution from upland representatives could overwhelm the near-shore signal. The abundance of grass, sedge, and chenopods in shoreline vegetation surrounding the lake further obscures the near-shore pollen signature, as strong prevailing westerly winds aid pollen dispersal from the opposite shoreline, resulting in increased representation of these taxa in the central lake area.

The plant macrofossil diagram (Fig. 10b) documents distinct changes in subfossil representation between each vegetation assemblage, reflecting both the composition of current vegetation and recent shoreline recession. As is the case with modern upland pollen spectra, Cactaceae is a distinctive indicator, as its seeds are restricted to T1-1. *Euphorbia* and Cruciferae seeds are also most abundant in T1-1, although both taxa are found at low concentrations in other vegetation zones. The presence of *Ruppia* and Chenopodiaceae seeds in dry prairie sediment likely records previous high water stands.

Macrofossil deposition in T1-2 is dominated by Gramineae (both undifferentiated florets and *Puccinellia* type) and Cyperaceae, a reflection of their dominance in current vegetation. Undifferentiated Chenopodiaceae seed deposition is also high in T1-2 (a further indication of past high water conditions), but reaches peak concentrations in T1-3, as does *Salicornia* seed concentration, a pattern consistent with the distribution of these taxa in the existing vegetation.

Ruppia seeds are abundant in both T1-3 and T1-4, accompanied by the only occurrence (albeit minor) of *Hordeum jubatum* seeds. Total seed deposition is reduced in the latter two assemblages, compared to T1-1 and T1-2.

Macrofossil deposition is low throughout the lacustrine portion of CL1. Of the few macrofossils recovered, Chenopodiaceae, Compositae (*Erigeron* type), Cruciferae, Cyperaceae, and *Chara* oogonia are all more abundant in near-shore samples (CL1-1W and -2W) than in samples collected in the central lake area. *Ruppia* seeds are abundant on the current saline beach and in the lacustrine sample closest to the current shoreline, but absent in mid-lake samples. Thus, proximity to shoreline is clearly reflected in macrofossil distribution in contrast to pollen spectra that, with the exception of *Ruppia*, bear only weak indicators of a near-shore location.

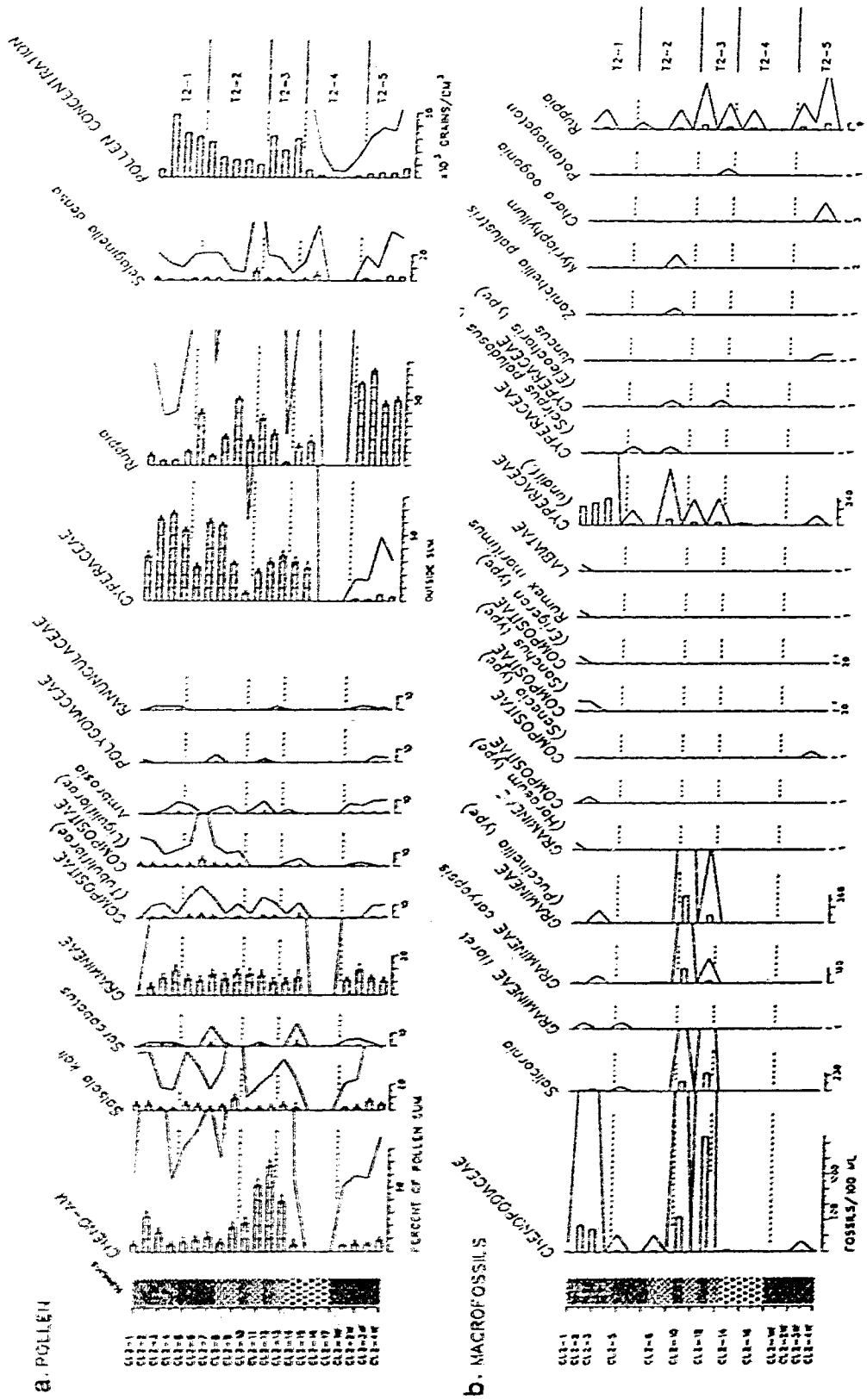
All pollen spectra in CL2 (Fig. 11a) are dominated by Cheno-Am, Gramineae, Cyperaceae, and *Ruppia*, although the proportional representation of each varies between vegetation assemblages. In T2-1 Cyperaceae pollen dominates and with the exception of CL2-1, pollen concentration is high, attaining the highest level recorded in this study (48,000 grains/cm³).

Cyperaceae remains the dominant pollen taxon in T2-2, although *Ruppia* pollen percentages increase (particularly in CL2-5 and -8) to the highest levels of the terrestrial portion in CL2 (50%), reflecting the availability of suitable habitat in small ponds situated in this portion of the streamcourse. Pollen concentration in T2-2 is generally less than half that recorded in T2-1.

Cheno-Am pollen dominates the T2-3 pollen spectra, reaching the highest recorded levels (65%) in the transect and reflecting its abundance in the current vegetation. Pollen concentration rises to 30,000 grains/cm³ in this assemblage.

The T2-4 pollen rain is incompletely characterized, as three samples from the saline beach (CL2-15, -16, and -17) contain pollen densities too low to tabulate (<2000 grains/cm³).

Figure 11. Modern pollen and plant macrofossil distribution along transect CL2 (note scale changes in 11b).



R. Vance

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STRATIGRAPHIC UNITS WITH 85% CONFIDENCE INTERVALS AUREOLATE CURVES EMPLOY UNITS AT DE EDUCATION

This reflects the sparse vegetation cover in addition to indicating a poor environment for preservation. Of the two samples tallied, Cheno-Am, Gramineae, Cyperaceae, and *Ruppia* pollen dominate.

Ruppia dominates the modern pollen spectra of lacustrine sediment samples in CL2 and, as in CL1, attains highest levels (63-74%) in samples collected closest to the current shoreline. Gramineae comprises 10-20% of pollen spectra throughout T2-5 and is marginally more abundant in near-shore samples. Cyperaceae and Cheno-Am decline from levels attained in the terrestrial portion of CL2.

Cheno-Am, *Salsola kali*, Cyperaceae, and *Selaginella densa* attain slightly higher percentages in samples retrieved some distance lakeward from the saline beach. Pollen concentration in lacustrine sediment is lower than the the terrestrial portion of transect CL2 (with the exception of the saline beach), attaining a maximum of 6000 grains/cm³ in CL2-4W.

Macrofossil assemblages along transect CL2 (Fig 11b) are dominated by Chenopodiaceae, Gramineae, and Cyperaceae, reflecting the abundance of these three taxa in the current vegetation. As in transect CL1, distributional patterns of current local vegetation are clearly reflected in plant macrofossil assemblages. Cyperaceae and Chenopodiaceae dominate the T2-1 assemblage. Although Cyperaceae is the dominant member of the current vegetation, members of the Chenopodiaceae family (with the exception of *Atriplex subspicata*) are not as prevalent. The abundance of Chenopodiaceae seeds and the appearance of *Ruppia* seeds in T2-1 may therefore be related to recent high water stands or periodic stream discharge.

Cyperaceae dominates the macrofossil assemblage of T2-2, although total macrofossil concentration declines, compared to T2-1. Occurrences of *Myriophyllum* and *Eleocharis* type seeds in T2-2 are indicative of recent stream activity, as these taxa are commonly found in fresh-to-marginally saline habitats (Hammer and Heseltine 1988).

Chenopodiaceae seeds occur in extremely high concentrations in T2-3 (1200 seeds/100 ml) and with Gramineae dominate the macrofossil assemblage, as these taxa do in the current vegetation. *Salicornia* seeds appear in high concentrations only in T2-3 (a maximum of 250 seeds/100 ml) where its seedlings are abundant.

Very few macrofossils were recovered on the saline beach. This setting is evidently not conducive to preservation of botanical remains, as the paucity of macrofossils mirror the sparse modern pollen collection. Repeated wet/dry cycles are known to be detrimental to pollen preservation (Holloway 1989). Thus, periodic stream discharge and seasonal lake level fluctuations combined with high evaporation rates likely contribute to the paucity of subfossil remains in T2-4.

Few macrofossils were found in lacustrine sediment samples collected in transect CL2. Thus, it appears that streamflow has had little impact on recent macrofossil distribution.

5.4 Conclusion

The plant cover surrounding Chappice Lake is similar in composition and distribution to the halophytic vegetation of other saline basins of western Canada (Dodd and Coupland 1966, Hammer and Heseltine 1988, Keith 1958, Rawson and Moore 1944) and the United States (Sloan 1970, Ungar 1970). Low diversity is characteristic of these extreme environments.

Dry, upland prairie vegetation produces a distinctive pollen signature, due to the presence of Cactaceae and heightened Liguliflorae type Compositae, Leguminosae, and *Selaginella densa* representation compared to shoreline and lacustrine assemblages. The most clear-cut indicator in the modern pollen rain of a near-shore, saline, shallow water environment is an abundance of *Ruppia* pollen. Cheno-Am, *Sarcobatus*, and Labiatae pollen display marginally increased representation in near-shore samples, and are therefore best considered weak indicators of proximity to shoreline.

The distribution of plant macrofossils along both transects bears a strong imprint of the local vegetation. Dry upland prairie vegetation is clearly distinguished by the presence of Cactaceae, *Euphorbia*, and Cruciferae seeds, as shoreline vegetation is by high Chenopodiaceae and Cyperaceae seed concentrations. Near-shore lacustrine samples contain increased numbers of Chenopodiaceae, Compositae (Erigeron type), and *Ruppia* seeds, compared to samples collected a greater distance from shoreline, indicating that seeds of these taxa are clear-cut indicators of proximity to shoreline. In addition, *Chara* oogonia are restricted to near-shore samples, suggesting they too are useful indicators of the shallow water, near-shore environments.

Results obtained in this study indicate that intermittent streamflow appears to have little impact on present pollen and seed distribution in the basin. This conclusion should be viewed with caution however, as recent construction of retaining weirs on this stream and its tributaries has undoubtedly reduced discharge.

CHAPTER 6

HOLOCENE PALEOHYDROLOGY OF CHAPPICE LAKE

6.1 Introduction

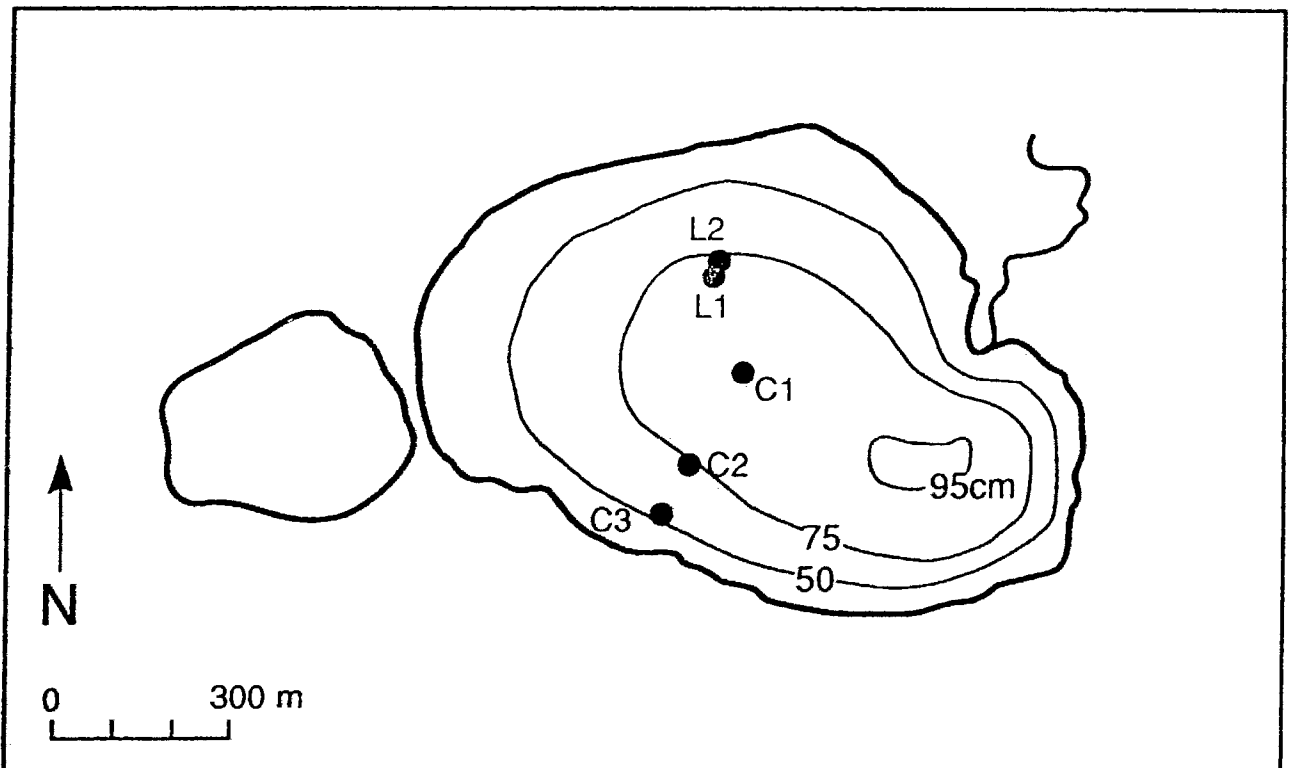
The process of reconstructing the Holocene paleohydrology of Chappice Lake was conducted in two stages. Small diameter sediment cores were used for preliminary pollen and plant macrofossil analyses, as well as establishing a preliminary chronology. Encouraging results stimulated collection of three large diameter cores for detailed macrofossil analyses and recovery of sufficient seed collections for additional AMS dates. The use of a vibracorer during the second winter coring trip also allowed collection of basal sediment that extended the overall sedimentary record by 4000 years.

6.2 Methods

6.2.1 *Coring*

In February of 1984, two 5 cm diameter cores (L1 and L2) were raised in the central basin area of Chappice Lake with a 5 cm diameter square-rod piston sampler (Wright 1967). To avoid collection of resuspended and mixed sediment from past periods of vigorous stream discharge, coring was conducted some distance away from the current stream delta (Fig. 12). Core L2 was extracted within 5 m of core L1. Depth from the ice surface to the sediment/water interface was 2.36 m. Both cores were extruded in the field, and each 1 m section was wrapped in cellophane and brown paper.

Figure 12. Chappice Lake coring locations. Bathymetry based on August 1989 lake levels.



In January of 1989 three large diameter (7.6 cm) sediment cores were extracted in a transect extending from the central basin area to the southern shoreline using a powered vibracorer (Lanesky et al. 1979). Continuous core sections were taken 350, 180, and 80 m from the current shoreline (C1, C2, and C3, respectively, Fig. 12). Depth through ice and water to the sediment surface was 86 cm at C1, 73 cm at C2, and 64 cm at C3.

Vibracores were limited only by the available length of coring pipe, in contrast to Livingstone cores that, without the benefit of power assistance, were unable to penetrate basal sediments. Immediately after each core was raised to the surface the aluminum collecting tubes were cut into approximately 1.5 m lengths and capped to prevent dessication.

To ensure complete recovery of unconsolidated material near the sediment/water interface, two additional one-meter cores were taken adjacent to C1 and C2 with a Brown sampler (B1 and B2). Cores B1 and B2 were capped and secured in an upright position for transport to the laboratory. All cores were stored at 3°C prior to analysis.

6.2.2 Core Description

Cores were split longitudinally in the laboratory for notation of Munsell colour and textural variation. Initially all cores were black but colour development associated with prolonged oxidation revealed previously-obscured sections of laminated sediment. Detailed logs of cores L1, C1, C2, C3, and portions of L2 were therefore finalized when oxidation was complete (after several days' exposure). Comparison of the upper sections of cores C1 and C2 with sediment collected using a Brown sampler revealed that the uppermost sediments were obtained without disturbance. As a result, sediment collected with the Brown sampler was not used in subsequent analyses.

Cores C1, C2, and C3 were sampled every 10 cm for determination of bulk density, dry weight, and loss-on-ignition following the methods outlined by Bengtsson and Enell (1986). To characterize the main stratigraphic units, additional samples were removed from L1, C1, C2, and C3 at varying intervals for grain size determination.

6.2.3 Pollen and Plant Macrofossil Analysis

Samples 1 cm³ in size were removed from the interior of cores L1 and C1 for pollen analysis. Sampling distance was a maximum of 25 cm, although efforts to sample discrete sedimentary units in detail reduced the sampling interval to 3 cm in some sections. Pollen processing and counting followed the protocol outlined in Chapter 5, with the exception that a minimum of 500 pollen grains (excluding Cyperaceae and aquatics) were tallied in all L1 samples, whereas a pollen sum of 300 was employed when tallying C1 samples.

Longitudinal halves of cores L1, C1, C2, and C3 were screened for macrofossil analysis with the exception of the 3-4 m section of L1, where the entire core was screened to ensure recovery of a sufficient number of seeds for AMS dating. The standard sampling interval was 10 cm, although periodic adjustments were made to permit sampling of discrete sedimentary units. Following determination of the volume of each sample prior to screening,

processing and identification of plant macrofossils followed the methods outlined in Chapter 5.

6.2.4 Radiocarbon Dating

A total of twelve samples were submitted for radiocarbon dating. Two samples of silty clay sediment from core L1 (0.00–0.20 m and 9.65–9.75 m) were submitted for conventional radiocarbon determinations. Prior to submission the outer surface was removed in an effort to minimize contamination. Samples were air dried, packaged in foil, and shipped to the laboratory.

The remaining ten radiocarbon samples consisted of non-aquatic seeds picked during macrofossil screening. These samples consisted mainly of Chenopodiaceae and Cyperaceae seeds, although one sample was supplemented with a *Potentilla* seed and a second with a *Potentilla* twig fragment to raise their weight to the minimum required. All specimens were handled exclusively with metal tweezers and thoroughly rinsed in dH₂O. Each sample was air dried and stored in a clean glass vial capped with a plastic stopper in preparation for shipment to the laboratory.

6.3 Results

6.3.1 Radiocarbon Dates

Results of ¹⁴C age determinations on the twelve samples submitted for analysis are listed in Table 5. Two bulk sediment samples were pretreated by removing rootlets and immersion in a warm HCl bath before being dated by conventional radiocarbon techniques. Sample AECV-355C (62 g of silty clay representing the upper 0.20 m of core L1), produced an uncorrected radiocarbon age of 1250±110 BP, indicating that conventional radiocarbon determinations are in error by as much as 1200 years due to the presence of ancient carbon.

Table 5. Results of radiocarbon determinations.

CORE: DEPTH (m)	LAB NUMBER	MATERIAL	WEIGHT	AGE (uncorrected years BP)*
L1: 0.00-0.20	AECV-355C ¹	silty clay lake sediment	62 g	1250±110
L1: 3.72-3.82	RDDL-829 ¹	Chenopodiaceae seeds	NA	1060±75
L1: 9.15-9.20	RDDL-830	Chenopodiaceae seeds	NA	3600±75
L1: 9.65-9.75	BETA-9067 ¹	silty clay lake sediment	84 g	9570±80
C1: 1.79-2.01	TO-1658 ¹	Chenopodiaceae seeds	2.2 mg	too small to date
C1: 3.95-4.30	AA-4922 ¹	Chenopodiaceae seeds	1.2 mg	2290±120
		plus 1 <u>Potentilla</u> seed		
C1: 7.60-7.70	TO-1659	Chenopodiaceae and Cyperaceae seeds, plus a <u>Potentilla</u> twig fragment	5.6 mg	6260±70
C1: 8.60-8.66	AA-4923	Cyperaceae seeds	1.5 mg	7325±70
C2: 2.76-2.94	TO-1660	Chenopodiaceae and Cyperaceae seeds	3.5 mg	5110±130
C2: 5.92-6.00	TO-1661	Cyperaceae seeds	25.0 mg	7050±70
C2: 6.79-7.00	TO-1662	Cyperaceae seeds	12.2 mg	7760±70
C3: 1.24-1.44	TO-1663	Chenopodiaceae and Cyperaceae seeds	5.8 mg	3340±70

*With the exception of L1: 0.00-20 and L1: 9.65-9.75, all dates are AMS determinations.

- ¹Alberta Environment Centre
- ¹Radio-Isotope Direct Detection Laboratory (McMaster University)
- ¹Beta Analytic Inc.
- ¹Isotope Laboratory (University of Toronto)
- ¹NSF Facility for Radioisotope Analysis (University of Arizona)

To circumvent these 'hard-water' effects, only AMS dates produced by dating non-aquatic organic remains were used for chronological control.

Ten samples were submitted for ^{14}C AMS radiocarbon determinations at three different laboratories. Only one sample (C1: 179–210 cm) produced a residue "...much too small for analysis" (Beukens, personal communication 1990). All nine sufficiently-large samples produced dates in proper stratigraphic sequence.

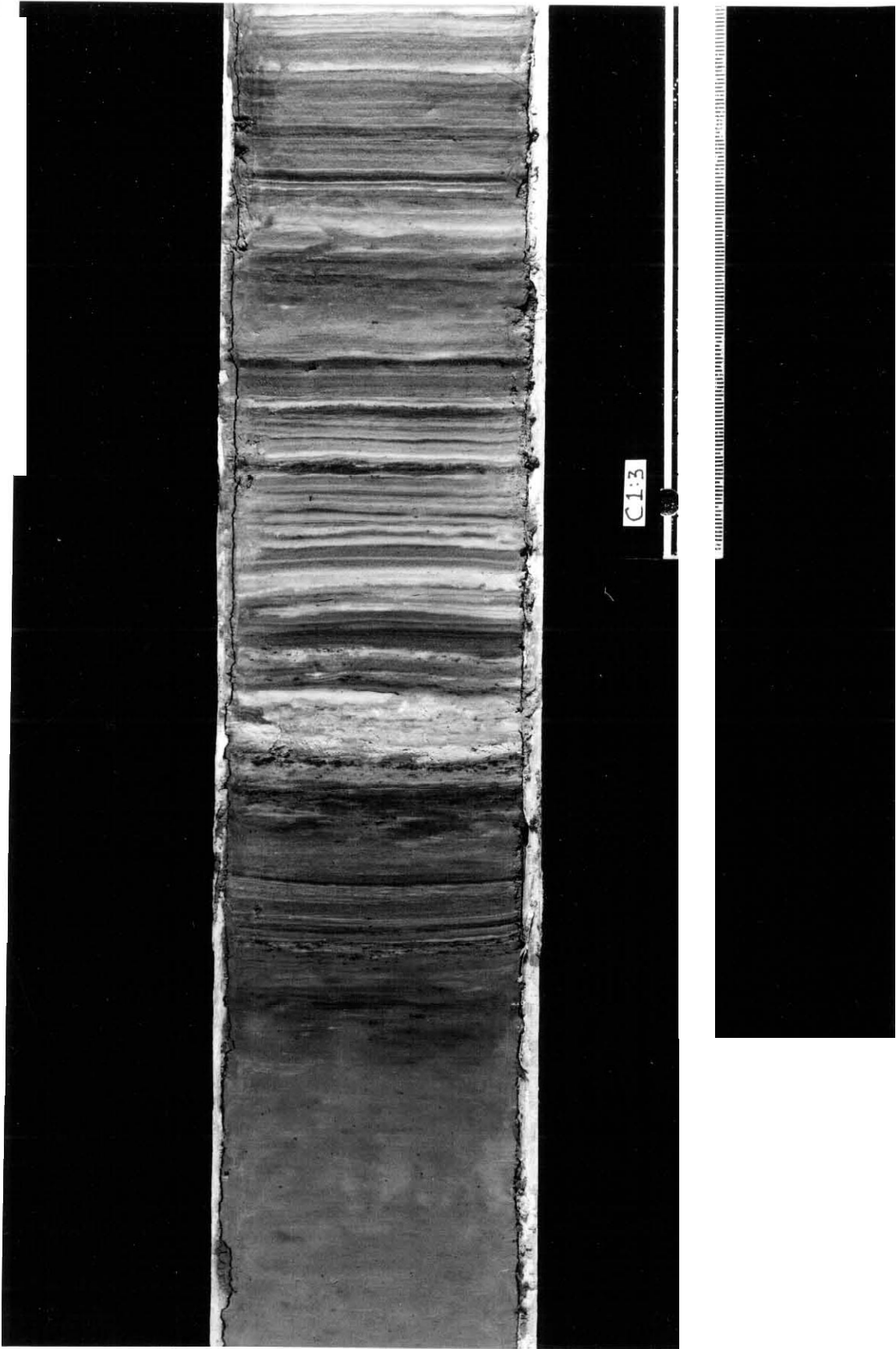
6.3.2 Stratigraphy

All sediment was uniformly black (2.5Y 2/0) and foul-smelling (H_2S) when extruded in the field or opened in the laboratory. With the exception of major textural changes in C2 and C3 no stratigraphic variation was evident. Following exposure to air however, complex colour changes and lamination patterns appeared.

Three main sedimentary units occur in the five cores. Massive silty clay and clayey silt dominates the upper portion of all cores. This sediment varies in colour from very dark grayish brown (2.5Y 3/2) to dark brown (7.5YR 3/2), often displaying olive brown (2.5Y 4/4) mottling. Grain size determinations on 10 samples of massive sediment from core C1 produced an average composition of 47% clay ($<2\mu\text{m}$), 50% silt ($2\mu\text{m}$ – $63\mu\text{m}$), and 3% sand ($63\mu\text{m}$ –2 mm). Silt content increases with depth in these samples.

Units of massive sediment are periodically interrupted by laminated sediment (Fig. 13) in all cores except the near-shore C3. Laminations vary from well-developed, distinct structures with sharp contacts to poorly-defined features with gradational contacts. Lamina colour varies widely, from light olive brown to very pale brown (10YR 7/3), light yellowish brown (10YR 6/4), yellowish brown (10YR 5/4), dark yellowish brown (10YR 4/4), dark reddish gray (5YR 4/2), dark reddish brown (5YR 3/3), reddish brown (5YR 4/3), and very dark grayish brown. Colour variation often highlights the distinctiveness of each laminae.

Figure 13. Core section CI-3, 5.13 m (right) to 4.71 m (left), with finely laminated silty clay underlying massive silty clay.



C13

Single laminae vary in width from <1–5 mm. Grain size analyses of 10 samples of laminated sediment from core C1 produced a mean value of 37% clay, 60% silt, and 3% sand. As is the case with massive sediment, silt content within laminated sequences increases toward the base of the core.

No attempt was made to sample individual laminae for grain size analysis, as the volume of material required for this procedure exceeds the amount available. Close inspection of individual laminae reveals that textural variation sometimes enhances the distinctiveness of adjacent bands. For example, very dark grayish brown silt and fine sand laminae are encountered occasionally in cores L1, C1, and C2.

A final class of sediment recovered from Chappice Lake is dominated by sand of various size classes and confined to the basal section of cores C2 and C3. In 3 samples from core C2 sand content varies between 92–96%. Pebble layers occur at irregular intervals in sand deposits.

Detailed descriptions of cores C1, C2, C3, and L1 are located in Appendix A. The following is a summary of major stratigraphic variations present in each core.

Core C1

The basal portion of C1 displays the greatest variability of all core sections (Fig. 14). The lowermost 0.31 m of predominantly-laminated sediment is capped by 0.20 m of weakly bedded to massive olive brown silt (90% within the $63\mu\text{m}$ – $2\mu\text{m}$ size class), from 8.37 to 8.17 m (Fig. 15). Short sequences of laminated sediment interspersed with massive units of fine sand and silt (between 8.17 and 7.50 m), are interrupted by two distinctive sections of dried, blocky pale olive (5Y 6/4) silty clay at 7.98–7.88 m (bearing contorted, interrupted laminations) and 7.76–7.69 m, as well as a section of pale olive silt and fine sand (64% $63\mu\text{m}$ – $2\mu\text{m}$, 24% 2 mm– $63\mu\text{m}$), between 7.66 and 7.59 m (Fig. 16). From 7.50 to 4.48 m,

Figure 14. Summary diagram of Chappice Lake sediments and radiocarbon dates.

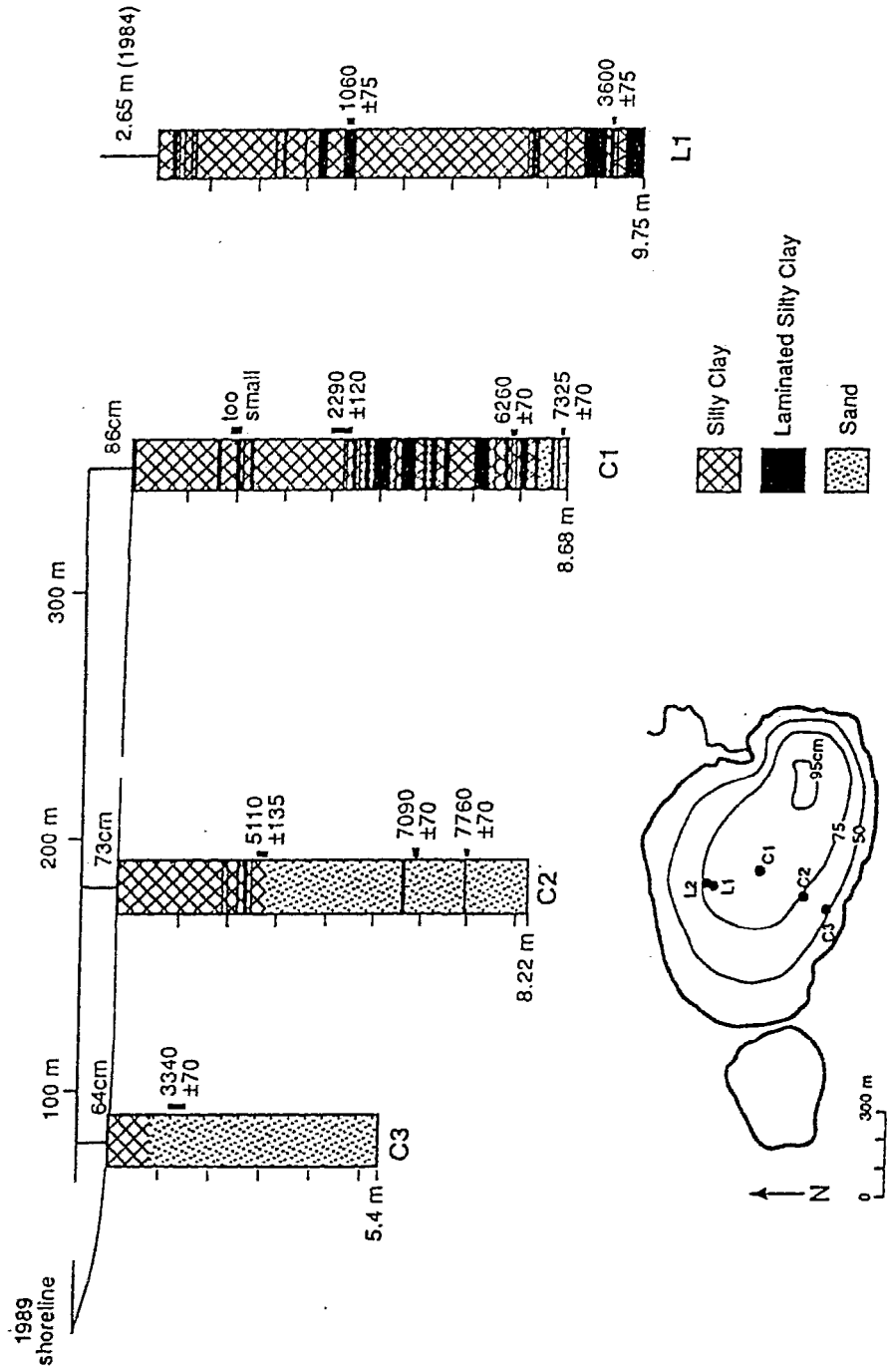


Figure 15. Core section C1-4, 8.44 m (right) to 8.12 m (left), with a weakly bedded to massive silt bed (8.37-8.17 m) overlying laminated silty clay and succeeded by weakly laminated to massive silt clay.

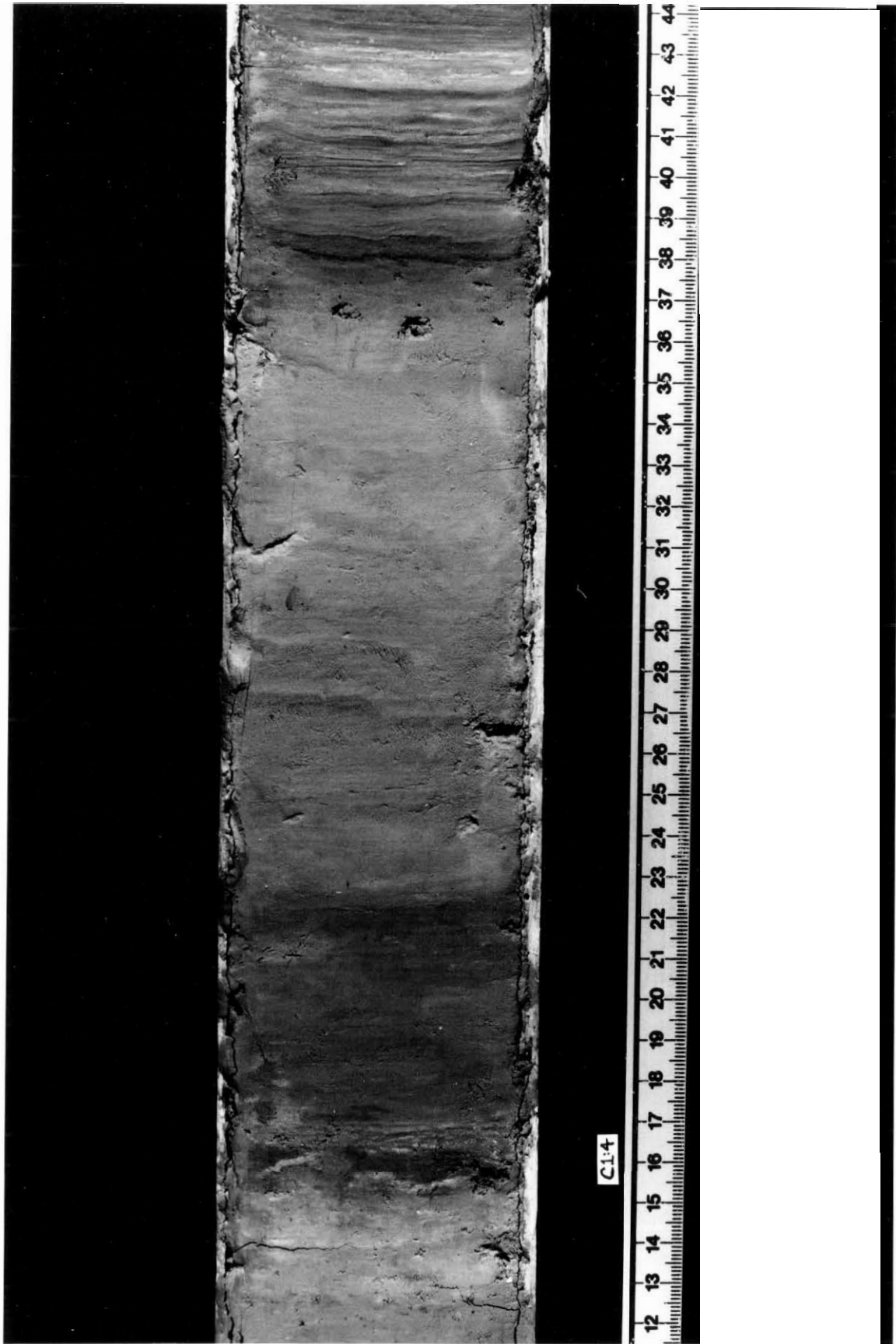
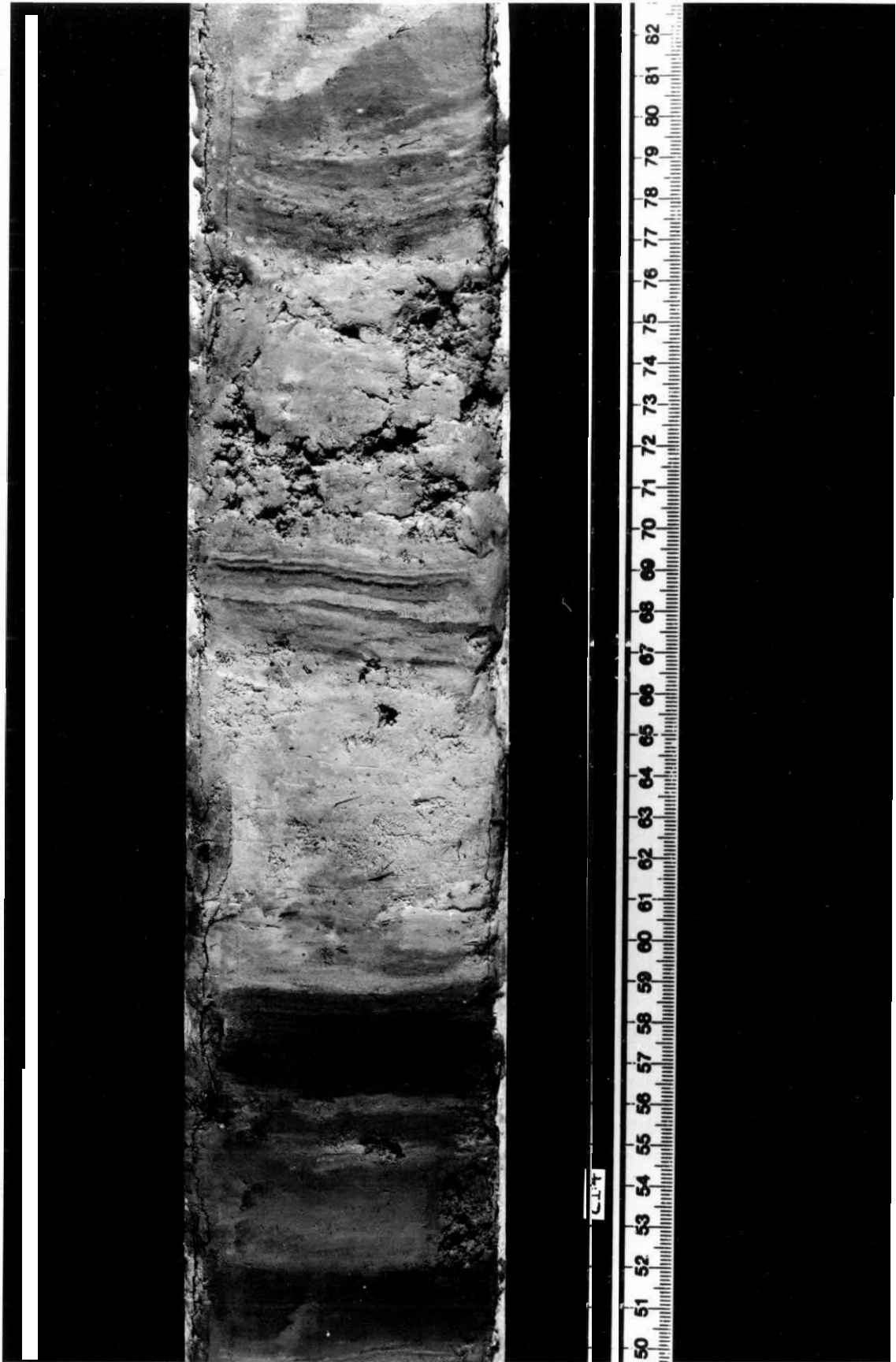


Figure 16. Core section C1-4, 7.82 m (right) to 7.50 m (left), with dried, blocky, silty clay (7.76-7.69 m) overlain by massive fine sand and silt (7.66-7.59 m).



long sequences of finely laminated sediment are interrupted only by short units of massive silty clay and two sections of chalky, light yellowish brown silt (6.26–6.25 m and 6.02–6.00 m).

From 4.48 to 2.34 m massive sediment dominates, at times displaying olive brown mottling as well as occasional thin layers of light yellowish brown silt. This virtually continuous unit of massive sediment is overlain by series of distinctly-laminated deposits between 2.34 and 1.32 m. These laminated sequences are well-defined and display the full range of colour variation described in section 1.3.2. Individual silt and fine sand laminae are included in these sequences.

The upper 1.32 m of C1 consists mainly of massive silty clay, although two distinct laminated sections are found at 0.40–0.35 m and 0.05–0.00 m. Approximately 0.10–0.15 m of salts blanketed the sediment–water interface at the time of coring. X-ray diffraction (whole rock analysis), conducted at the Ottawa Laboratory of the Geological Survey of Canada, showed these salts to be 100% thenardite (Na_2SO_4).

Core L1

Since cores L1 and L2 were taken within 5 m of each other and bear essentially identical stratigraphy, only core L1 is described. The basal 2.5 m of the 9.75 m core is dominated by extensive laminated silty clay sequences. Laminae in the basal 2.5 m of L1 are for the most part well-defined, displaying a wide range of colour variation. A sparse coating of cylindrical salt crystals developed during storage on the exterior of the 9.64 to 9.44 m core section.

Massive silty clay sediment dominates between 7.25 and 3.80 m. Dark grayish brown and very dark grayish brown units grade gradually into one another throughout this section, occasionally bearing olive brown mottling and infrequent grayish brown layers up to 0.05 m thick. Traces of burrowing benthic organisms appear occasionally, in the form of thin winding

trails of light coloured sediment.

Distinctly laminated sediment dominates between 3.82 and 2.37 m. Laminae are best developed in four distinct sections, where they are well-defined and display the full range of colour variation described in section 1.3.2 above.

Massive silty clay sediment dominates the upper 2 m of L1, although short laminated sequences (<5 cm in length) appear occasionally. The most extensive of these are found between 0.75 to 0.61 m and 0.52 to 0.37 m.

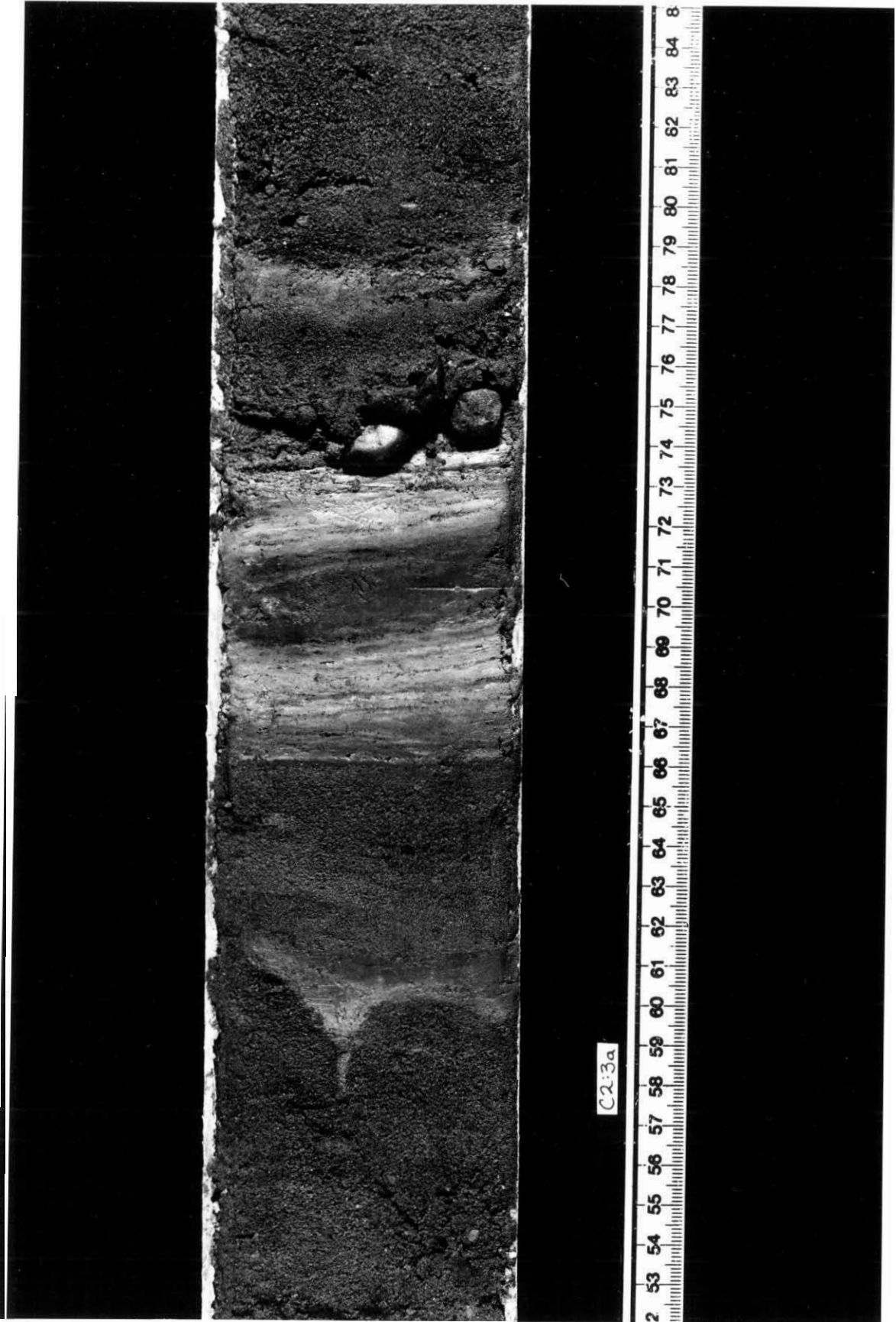
Core C2

The 8.23 m long core C2 consists of two main sedimentary units, 5.48 m of basal sand overlain by 2.75 m of predominantly silty clay deposits. Basal sands are punctuated by fine-grained deposits, most notably a series short clay and silt units from 7.25 to 6.75 m and 5.75 to 5.25 m. Abundant rootlets are found at 6.35 m and fine sand containing pods of organic matter occurs between 6.02 to 5.92 m. A unit of weakly stratified silty clay between 5.75 and 5.60 m displays a sharp, loaded contact (Fig. 17). Several discrete pebble layers are also found in basal C2 sands.

Above 3.35 m, medium to coarse sand grades into interbedded fine sand and silt, massive dark grayish brown silty clay and at 2.68 m, laminated silty clay. The 2.68 to 1.75 m section bears massive dark grayish brown silty clay punctuated by units of laminated silty clay, bearing the full range of colours described in section 1.3.2 above.

The uppermost 1.75 m of core C2 consists entirely of massive silty clay, with colour differences defining crude stratification. The sediment/water interface of core C2 is blanketed by a 0.15 m layer of salt, presumed to be thenardite.

Figure 17. Core section C2-3a, 5.84 m (right) to 5.53 m (left), with weakly stratified fine sand, clay, and silt (5.74-5.60 m) displaying a sharp, loaded contact underlain by pebbles (to 2 cm in diameter).



C2:3a

Core C3

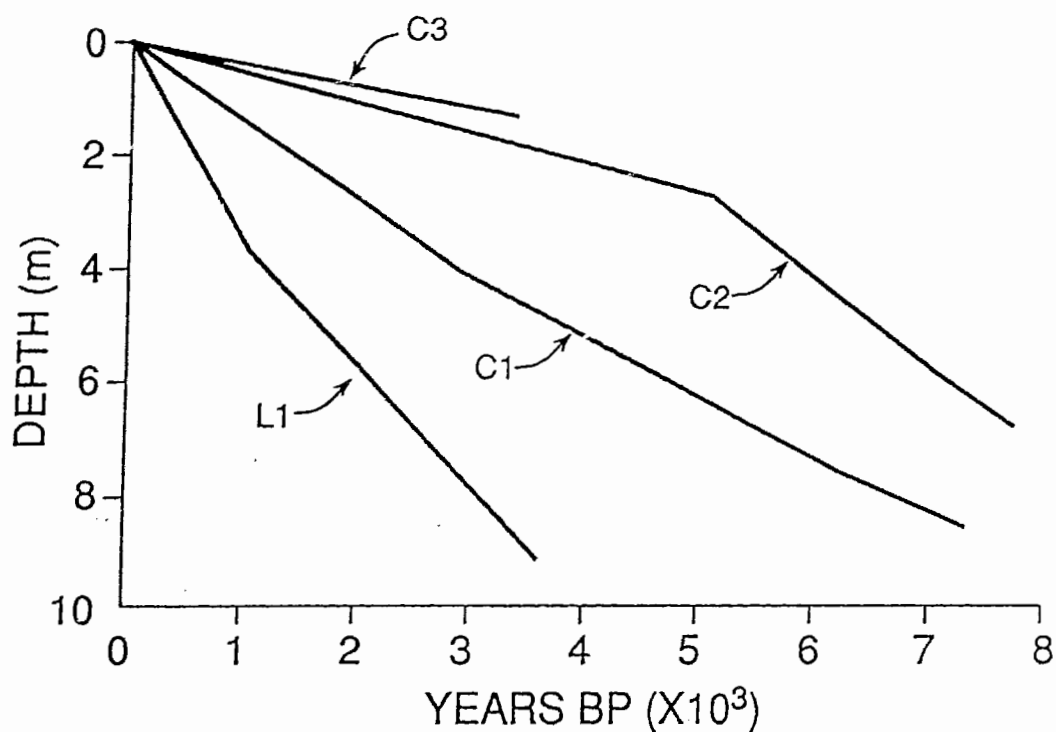
The 5.45 m long core C3 is predominantly sand, with the exception of the upper 1.05 m. Sections of coarse-grained sand are often associated with distinct pebble layers. Units of fine-grained sediment occur occasionally, including an interbedded olive fine sand and silt (containing streaks of dark gray organic matter) between 5.40 and 5.25 m.

The upper 1.05 m of C3 is predominantly massive silty clay. Between 0.60 and 0.40 m an increase in sand, compared to adjacent units, imparts a distinctly gritty feel to the sediment. Plant detritus is abundant in the uppermost 0.14 m. Core C3 is overlain by 0.10 m of salt, presumed to be thenardite.

6.3.3 Sediment Accumulation Rates

A summary of sediment accumulation rates (the slope of a straight line connecting each AMS ^{14}C date) for all dated cores (Fig. 18) shows that the most rapid rates, regardless of sediment type, are found in core L1 (0.356 cm/yr from 0.00 to 3.77 m, 0.213 cm/yr from 3.77 to 9.175 m). Deposition of predominantly silty clay or clayey silt in core C1 occurs at a lower rate, averaging 0.180 cm/yr over the uppermost 4.125 m of sediment, and slows to 0.089 and 0.092 cm/yr in the 4.125 to 7.65 m and 7.65 to 8.63 m sections, respectively. Accumulation rates for the uppermost fine-grained sediment of cores C2 and C3 are lower yet (0.056 and 0.040 cm/yr, respectively), whereas the average accumulation of sand in core C2 is a relatively rapid 0.149 cm/yr. Sedimentation rates portrayed on Figure 18 were used to estimate (by interpolation) ages of events not directly dated by radiocarbon.

Figure 18. Sediment accumulation rates for cores L1, C1, C2, and C3.

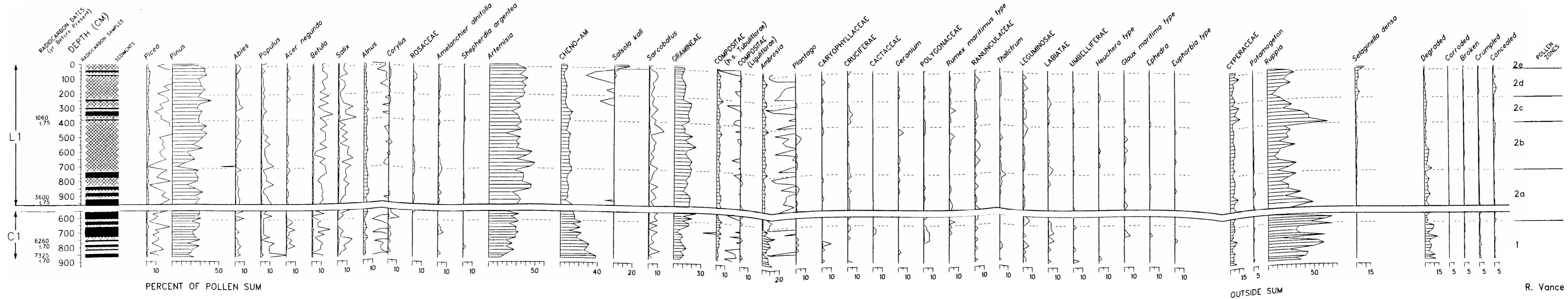



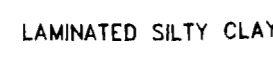

6.3.4 The Composite Pollen Record

The Chappice Lake composite pollen record (Fig. 19) combines results of initial pollen analyses on core section L1 (covering the period from the present to 3800 BP), with the basal 3.16 m of core C1, which extends the record to 7300 BP. Correlation between cores L1 and C1 was accomplished by interpolation between radiocarbon dates. Detailed comparison of sediment stratigraphy, as well as pollen and plant macrofossil assemblages, confirmed the correlation established by interpolation.

Figure 19. Chappice Lake composite percentage pollen diagram.

CHAPPICE LAKE COMPOSITE POLLEN DIAGRAM



 SILTY CLAY
  LAMINATED SILTY CLAY
  FINE SAND AND SILT

 AUXILLARY CURVES SHOWING VALUES AT 10X EXAGGERATION

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Samples spaced 0.25 m apart were utilized in the composite pollen record. A pollen sum of >500 for L1 and >300 for C1 included all upland pollen taxa. As in determination of modern pollen spectra (Chapter 5) Cyperaceae, all aquatic taxa, spores, and unidentifiable pollen grains were tallied and their frequencies calculated outside the pollen sum.

Pollen zone boundaries were determined by first using the program 'zonation' of Michigra 86 (Futyma and Meacham 1986), incorporating all upland pollen types displaying >5% representation. In addition to zone boundaries 1-2 and 2d-2e, 'zonation' identified numerous possible boundaries, each accounting for only small increments of total variation, indicating that differences between these pollen spectra are minor. Thus, additional subzone boundaries were determined by visual inspection, with particular attention paid to *Ruppia* pollen representation, the taxon identified in surface sample analyses as the best indicator of lake level fluctuations (Chapter 5).

Zone 1: 8.66–6.00 m, 7300–4400 BP

Pollen zone 1 is characterized by high Chenopodiaceae (10–40%), high *Ambrosia* (to a maximum of 15%), low *Artemisia* and Gramineae representation (less than 30 and 15%, respectively), and fluctuating but generally high *Ruppia* pollen percentages (20–70%). High degraded pollen counts (to 15%) further distinguish this zone. Sediments deposited throughout pollen zone 1 are variable, ranging from silt and fine sand to laminated silty clay, massive silty clay, and units of blocky, dried silty clay.

Zone 2: 6.00–0.00 m, 4400 BP–present

Declining Chenopodiaceae (to consistently below 15%), *Ambrosia* (<8%), and degraded pollen representation (<7%) distinguish zone 2 pollen spectra from zone 1. Increasing *Artemisia* (>25%) and Gramineae (>10%) frequencies further differentiate pollen zone 2 from assemblages in underlying sediment.

With the exception of occasional fine sand laminae, massive silty clay interspersed with laminated sections characterizes the upper 6.0 m of sediment deposition. Pollen zone 2 has been divided into five subzones, based mainly upon fluctuations in *Ruppia* pollen frequency.

Subzone 2a: 6.00 m (C1)-7.15 m (L1), 4400- 2650 BP

Subzone 2a is distinguished by fluctuating but generally high *Ruppia* (15-50%) and declining Chenopodiaceae (to <15%). Deposition throughout most of 2a is dominated by long sequences of finely laminated silty clay sediment.

Subzone 2b: 7.15-3.80 m, 2650-1060 BP

Subzone 2b is demarcated by a decline in *Ruppia* representation to values below 25% (the lowest of the record), with the exception of two sharp peaks. Following an abrupt peak at the onset of 2b, degraded grains drop in representation to below 5%. *Pinus* percentages decline to the lowest levels of the record early in 2b, but rise steadily throughout. Thus subzone contains the longest unit of massive silty clay in the record.

Subzone 2c: 3.80-2.25 m, 1060-600 BP

Subzone 2c is delineated by an abrupt peak in *Ruppia* representation (to 65%) and a rise in Cyperaceae representation (generally >5%). It also marks a resumption in the deposition of laminae.

Subzone 2d: 2.25-0.35m, 600-100 BP

Ruppia pollen representation drops below 15% and Cyperaceae remains below 5% throughout most of 2d. The lower boundary of this subzone is further delineated by the appearance of *Selaginella* spores in values consistently >3%. Massive silty clay interspersed with short laminated sequences typifies sediment deposition in subzone 2d.

Subzone 2e: 0.35–0.00 m, 100 BP–present

The uppermost sediment deposited in Chappice Lake is distinguished by the presence of *Salsola kali* pollen in abundance (>15%). Massive silty clay is the characteristic sediment of subzone 2e.

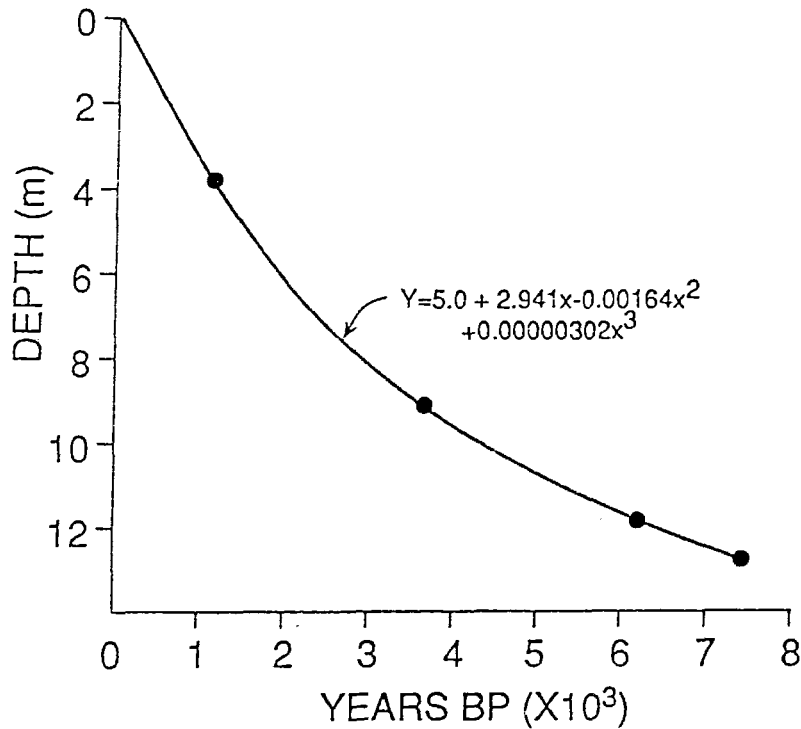
6.3.5 Pollen Influx

A pollen influx record was constructed using sediment accumulation rates determined by the inverse of the first derivative of a curve fitted to the two AMS dates from core L1 and the two basal AMS dates from the basal section of C1 (Fig. 20). The curve-fitting procedure reduces the impact of abrupt changes in pollen influx produced by utilizing straight-line interpolation between radiocarbon dates to determine sediment accumulation rates (Fig. 18). Pollen zone boundaries from the composite percentage diagram have been placed in the same stratigraphic position on the influx diagram.

Zone 1 (Fig. 21) has the lowest total pollen influx of the entire record (<4500 grains/cm²/yr). Low pollen influx obscures indicators originally used to define zone 1, such as high *Ambrosia* and Chenopodiaceae pollen percentages.

Zone 2 contains consistently higher pollen influx than zone 1, varying between 2600 and 20,000 grains/cm²/yr. Influx changes also clarify boundaries between the five subzones. High *Ruppia* pollen influx (generally >5000 grains/cm²/yr) characterizes subzone 2c, while moderate *Ruppia* influx (>2000 grains/cm²/yr) is typical of 2a. Low *Ruppia* pollen influx (<2000 grains/cm²/yr) occurs throughout 2b. Variations in influx also highlight differences in *Artemisia*, Cheno-Am, Gramineae, and Cyperaceae representation within subzones 2a–2c, with each taxon displaying reduced influx in subzone 2b, compared to overlying and underlying pollen assemblages.

Figure 20. Sediment accumulation rate utilized for determination of pollen influx.



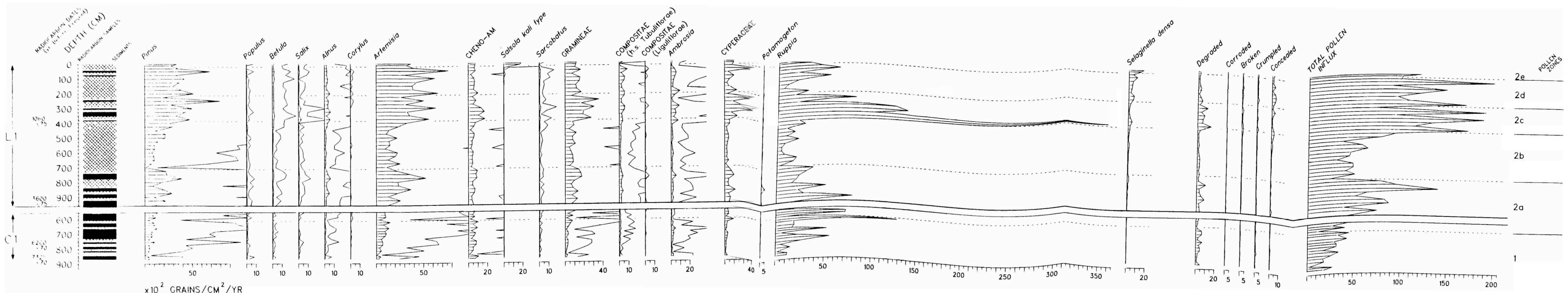
6.3.6 Plant Macrofossil Records


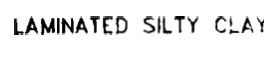


Core L1

Macrofossils recovered from core L1 span approximately the last 3800 radiocarbon years, virtually the duration of pollen zone 2. As in the pollen influx diagram, subzone boundaries derived from the composite pollen percentage diagram have been transferred to their exact stratigraphic location on the macrofossil diagram (Fig. 22).

Figure 21. Chappice Lake composite pollen influx diagram.

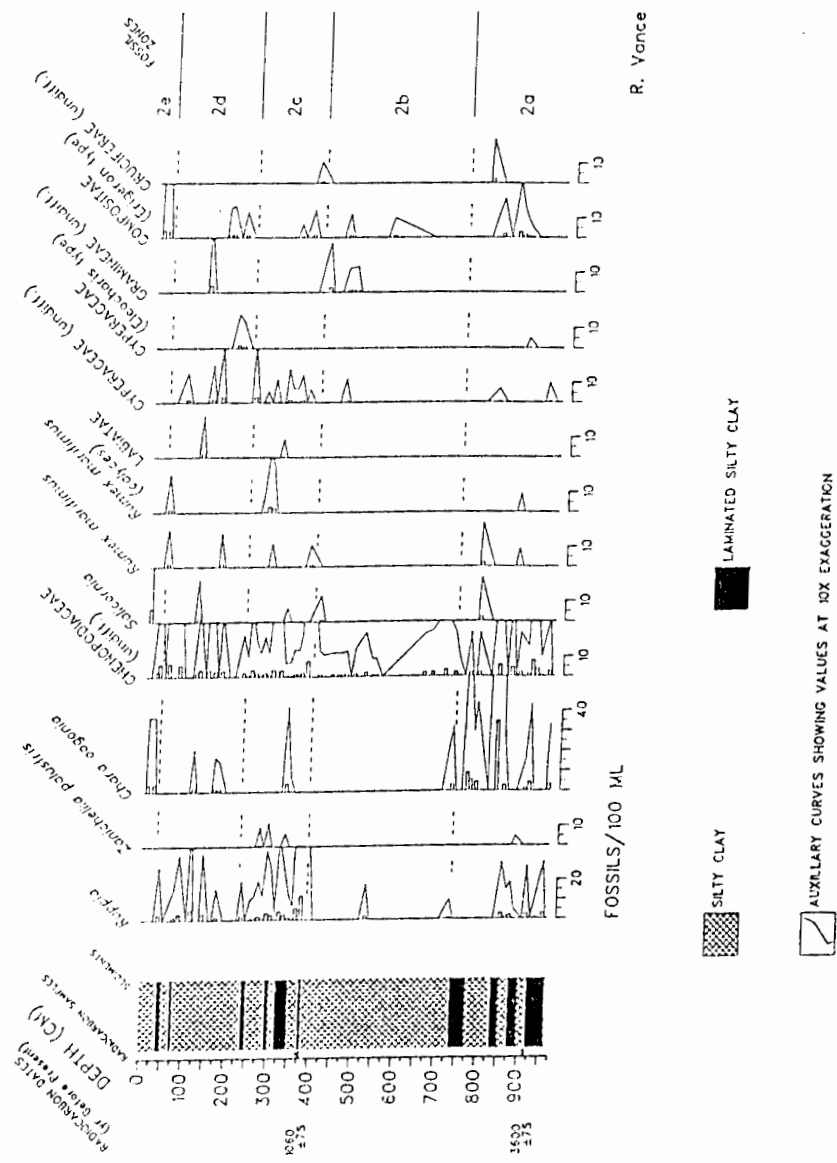
CHAPPICE LAKE COMPOSITE POLLEN DIAGRAM



 SILTY CLAY
  LAMINATED SILTY CLAY
  FINE SAND AND SILT
 AUXILIARY CURVES SHOWING VALUES AT 10X EXAGGERATION

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Figure 22. Core L1 plant macrofossil record.



In general, the L1 macrofossil record offers support for the placement of all subzone boundaries within pollen zone 2. In subzone 2a there is virtually continuous deposition of Chenopodiaceae seeds (2-8 seeds/100 ml of sediment). Low (up to 3/100 ml) but continuous deposition of *Ruppia* seeds is restricted to the lowermost meter. A minor concentration of *Erigeron* type Compositae seeds is found between 8.0 and 9.0 m. Subzone 2a is best distinguished however, by generally high concentrations (up to 34/100 ml) of *Chara* oogonia.

With the exception of low Chenopodiaceae seed concentration in the upper and lowermost levels, sediment deposited within subzone 2b is virtually barren of plant macrofossils. *Chara* oogonia are absent and, with the exception of two samples, no *Ruppia* seeds were recovered. However, vertebral and mandibular fragments of the common deer mouse (*Peromyscus maniculatus*) were found within subzone 2b.

Pollen subzone 2c is characterized by continuous deposition of abundant *Ruppia* and Chenopodiaceae seeds (up to 10 and 8 seeds/100 ml, respectively). Peaks in concentration of Cyperaceae seeds and *Rumex maritimus* calyces are also found within this subzone. Sporadic *Zanichellia palustris* seed deposition further distinguishes the 2c macrofossil assemblage.

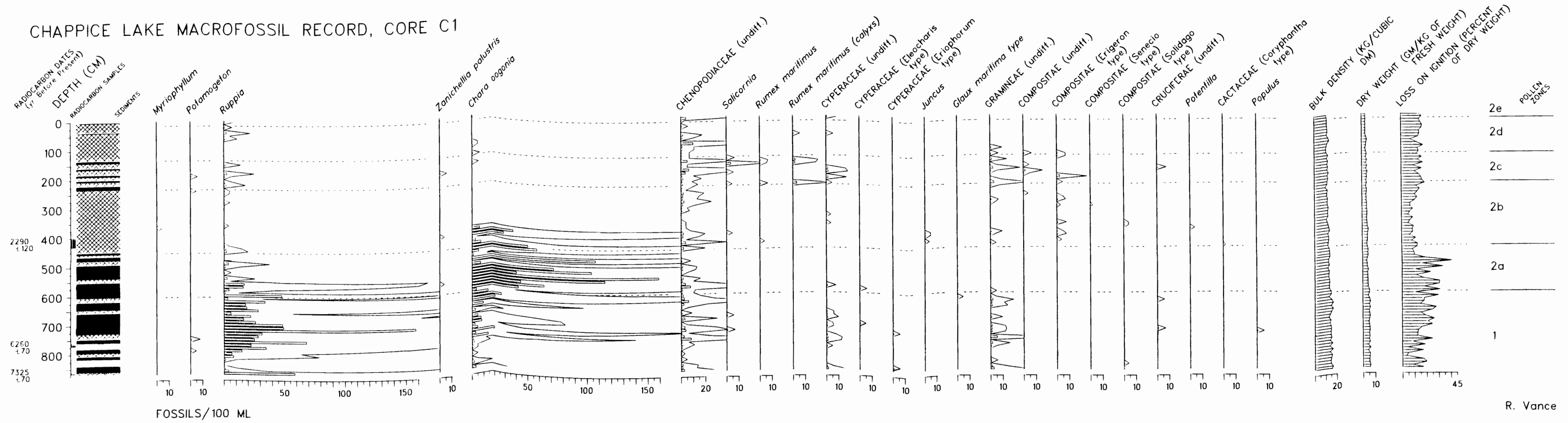
Sporadic, low concentrations of *Ruppia*, Chenopodiaceae, and Cyperaceae seeds as well as *Chara* oogonia are found in sediment deposited throughout subzone 2d. Deposition of *Ruppia* and Chenopodiaceae seeds is reduced, compared to older sediments. High levels of *Chara* oogonia as well as *Salicornia* and *Erigeron* type seeds are found in the uppermost levels of subzone 2e.




Core C1


Core C1 displays pronounced changes in plant macrofossil representation (Fig. 23). As in the core L1 macrofossil record, pollen zones from the the composite record were transferred to the C1 macrofossil record. However, direct transfer was not possible, as

Figure 23. Core C1 plant macrofossil record.

CHAPPICE LAKE MACROFOSSIL RECORD, CORE C1



 SILTY CLAY
  LAMINATED SILTY CLAY
  FINE SAND AND SILT

 AUXILLARY CURVES SHOWING VALUES AT 10X EXAGGERATION

placement of subzone boundaries is complicated by variations in sediment accumulation rates between cores L1 and C1. Thus, subzone boundary locations were determined by first calculating the approximate position using sediment accumulation rates (Fig. 18), then adjusting the position of subzone boundaries 2d, 2c, and 2b slightly depending on *Ruppia* seed abundance, as *Ruppia* pollen abundance was originally used to define these boundaries.

Zone 1 is characterized by generally high concentrations of *Ruppia* seeds and *Chara* oogonia, although both types are rare in the basal 0.60 m. Relatively high concentrations of Chenopodiaceae, Gramineae, and Cyperaceae seeds further distinguish zone 1. Sporadic occurrences of Cruciferae and *Eriophorum* type seeds as well as the sole occurrence of *Populus* in the record are other noteworthy aspects of this assemblage. Generally high but erratic loss-on-ignition is characteristic of zone 1.

Loss-on-ignition and *Chara* oogonia peak in subzone 2a. *Ruppia* seed deposition is virtually continuous throughout 2a, but reduced in concentration compared to zone 1. A similar trend is evident in Chenopodiaceae seed deposition, although the concentration reduction is not as pronounced as that in *Ruppia*. Gramineae seed deposition ceases early in subzone 2a.

With the exception of *Chara* in the basal 0.50 m, low concentrations of plant macrofossils are characteristic of subzone 2b. The lone occurrence of *Myriophyllum* and a *Coryphantha* type Cactaceae seed at the base of this subzone are noteworthy aspects of the sparse record. Lowest loss-on-ignition in C1 is found within subzone 2b.

Increased concentrations of *Ruppia*, Chenopodiaceae (both undifferentiated and *Salicornia* type), Cyperaceae, and Gramineae seeds distinguish subzone 2c from 2b. Relatively high concentrations of *Rumex maritimus* seeds and calyces as well as undifferentiated Compositae and *Erigeron* seeds are also distinctive features. Sporadic occurrences of *Ruppia* and Chenopodiaceae seeds are found within subzones 2b and 2a, but seed concentrations are

reduced in the upper two subzones, compared to 2c.

Core C2

Although core C2 is slightly shorter in length than C1, radiocarbon determinations reveal that it represents a longer time period, extending the overall Chappice Lake record to approximately 8500 BP. As with previous macrofossil records, zones derived from the composite percentage pollen diagram were transferred to the C2 macrofossil diagram (Fig. 24) by interpolation, although changes in sediment type mean that the assumption of uniform sedimentation rates may result in imprecise boundary locations.

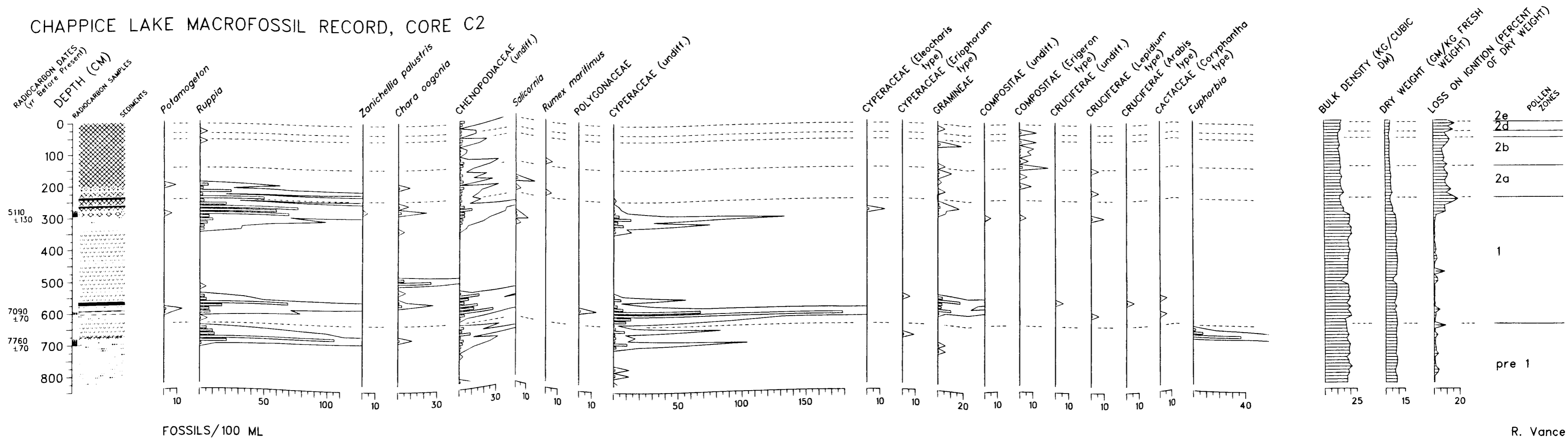
Sand deposited prior to zone 1 displays very low loss-on-ignition and contains a varied macrofossil assemblage dominated by *Ruppia*, Chenopodiaceae, Cyperaceae, and *Euphorbia* seeds. The only gastropod remains of the record were recovered from these sediments. Single complete shells of *Gyraulus* sp. and *Lymnaea* sp. were found between 7.95–7.85 m, 7.00–6.90 m, and 6.90–6.79 m.

Sediment throughout zone 1 (predominantly sand) generally produced low loss-on-ignition, although in the upper 0.50 m loss-on-ignition rises abruptly to levels as high as 19%. Plant macrofossil deposition throughout zone 1 is sporadic, with an essentially barren midsection flanked by pronounced peaks in *Ruppia*, Chenopodiaceae, and Cyperaceae seeds. Gramineae seeds are also a prominent member of the lowermost peak in macrofossil concentration. *Chara* deposition is low and mainly confined to zone 1. The highest concentrations attained (20 oogonia/100 ml of sediment) are found in the basal 1.5 m of zone 1.

Basal C2 sands are succeeded by massive and laminated silty clay in zone 2a. *Ruppia* seeds are common in the basal 0.30 m, an interval also containing sporadic occurrences of *Salicornia* seeds in low concentrations. Chenopodiaceae seed deposition is continuous throughout

Figure 24. Core C2 macrofossil record.

CHAPPICE LAKE MACROFOSSIL RECORD, CORE C2



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- SILTY CLAY
- WEAKLY STRATIFIED SILTY CLAY
- LAMINATED SILTY CLAY
- FINE SAND
- FINE SAND WITH ORGANICS
- MEDIUM TO COARSE SAND
- AUXILIARY CURVES SHOWING VALUES AT 10X EXAGGERATION

2a. Cyperaceae seeds are absent but *Erigeron* type seeds appear occasionally throughout 2a.

The upper three subzones of C2 are confined to merely 1.5 m of sediment, and there is little in the macrofossil record to distinguish these subzones. *Ruppia* seed deposition is generally low, with sporadic occurrences confined to the upper and lower boundaries of 2c. Chenopodiaceae seed deposition is virtually continuous throughout the upper 1.5 m of C2. *Erigeron* type seeds are found at low concentrations throughout 2b, but are only sporadically represented above this subzone.

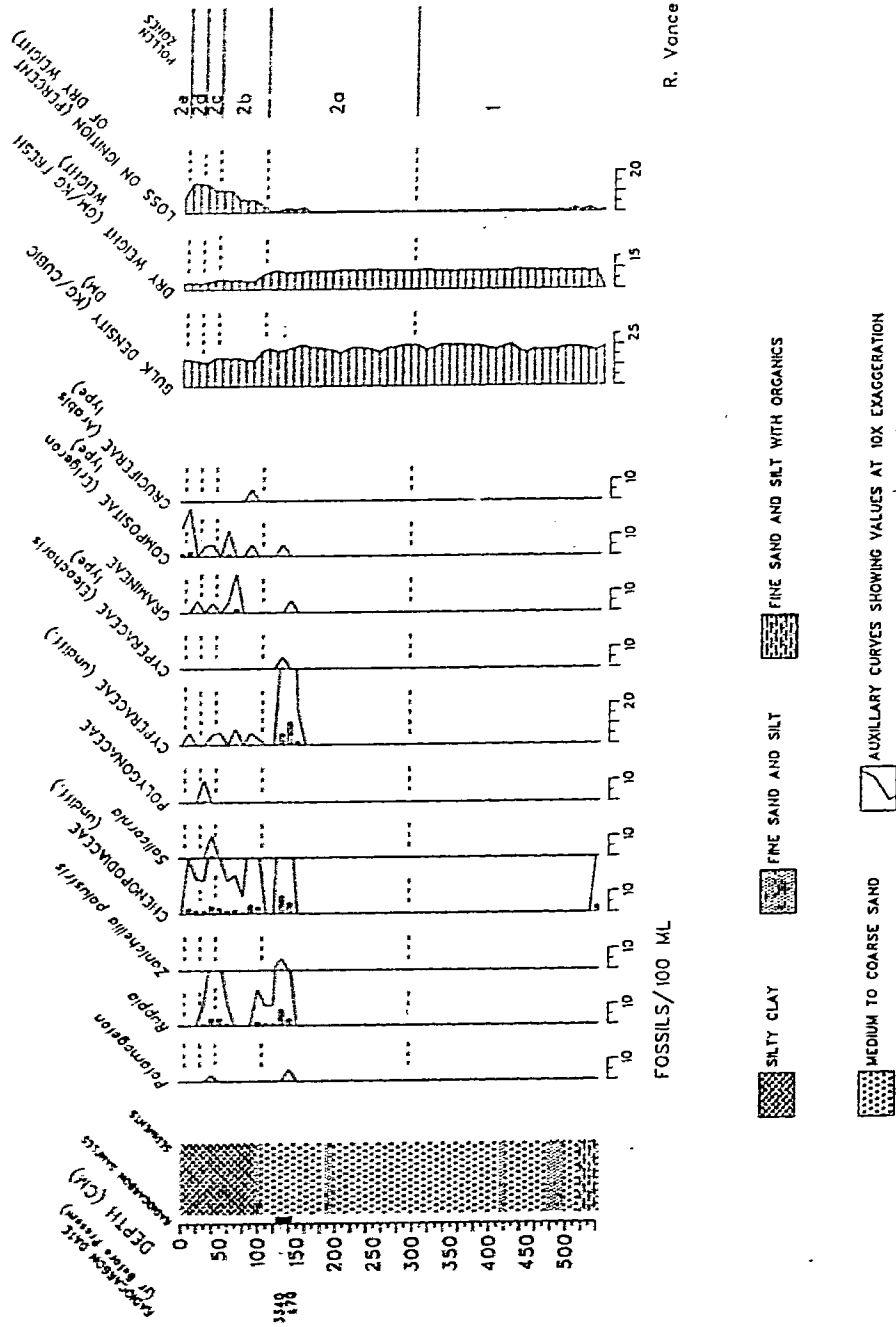
Core C3

Plant macrofossil remains are generally confined to the upper 1.5 m of core C3 (Fig. 25). Using the mean sediment accumulation rate of sand derived from core C2 (0.149 cm/yr), the lone radiocarbon date on basal C3 seeds, and an assumption of constant sedimentation rates within each sedimentary unit, zone boundaries derived from the composite pollen record have been transferred to Figure 25.

Sands deposited in zone 1 produced low loss-on-ignition and are essentially devoid of plant macrofossils, with the notable exception of a minor Chenopodiaceae seed occurrence (<5/100 ml of sediment) at the base of C3. Sand with low loss-on-ignition also characterizes much of subzone 2a. In the upper 0.50 m of 2a, loss-on-ignition rises and a varied macrofossil assemblage, dominated by *Ruppia*, Chenopodiaceae, and Cyperaceae seeds, appears for the first time.

Declining concentrations of *Ruppia* and Cyperaceae seeds, compared to the upper section of 2a, are characteristic of subzone 2b. Silty clay sediment with rising loss-on-ignition, compared to 2a, also typically contains low concentrations of Chenopodiaceae seeds (<5/100 ml). A minor peak in Gramineae seeds (<5/100 ml) occurs in the middle of 2b.

Figure 25. Core C3 plant macrofossil record.



The uppermost three subzones are condensed into less than 0.50 m of sediment in core C3, making distinctions between each tenuous. It is noteworthy however, that subzone 2c is associated with a peak in *Ruppia* seed concentration and the lone occurrences of *Potamogeton*, *Salicornia*, and Polygonaceae seeds in the record. Subzones 2d and 2e contain slightly more *Erigeron* type seeds compared to preceding subzones. The uppermost 0.50 m of sediment in core C3 produced the highest loss-on-ignition of the record.

6.4 Interpretation of Lake Level Fluctuations

6.4.1 8500–7300 BP

The composite pollen record spanning the last 7300 BP, together with plant macrofossil records and sedimentological characteristics of cores L1, C1, C2, and C3, form the basis for historical reconstruction of Chappice Lake. Data on paleohydrological change prior to 7300 BP exist in the basal section of core C2 (Fig. 24), providing some information on lake conditions from 8500 to 7300 BP.

The basal 2.00 m of C2 is predominantly sand, indicating that Chappice Lake was much smaller than present, as sands are the characteristic substrate of dry upland surrounding the lake today (Chapter 4). Thus, coring site C2 was likely situated within a dry prairie environment some distance from shoreline. However, Cyperaceae and Chenopodiaceae seed deposition at the base of C2 suggests that lake level had risen briefly to a position near C2, as seeds of these taxa are abundant in shoreline and near-shoreline environments (Chapter 5). This lake level rise dates to approximately 8500 BP.

The near-basal radiocarbon date from core C2 (7760 ± 70) marks a second rise in lake level, in this case delineated by high concentrations of *Ruppia*, Chenopodiaceae, and Cyperaceae seeds coincident with deposition of fine sand with increased organic matter (as suggested by dark streaks and heightened loss-on-ignition). Seeds of near-shore indicators are

found in association with a pronounced peak in *Euphorbia* seeds, an indicator of dry prairie vegetation (Chapter 5). As this assemblage contains evidence of both shoreline and dry upland environments, it evidently reflects early shoreline oscillations at coring site C2. Lake level subsequently declined, leaving C2 some distance upslope of the lake in dry prairie upland, as indicated by declining concentrations of *Ruppia*, Chenopodiaceae, and Cyperaceae seeds, as well as renewed deposition of sand immediately below the lower zone 1 boundary.

Pronounced changes in lake chemistry accompanying these early lake level fluctuations are suggested by the occurrence of *Lymnaea* and *Gyraulus* shells. Rawson and Moore (1944) note that gastropods are rarely found in lakes with TDS exceeding 3 ppt and are eliminated from lakes when TDS exceeds 15 ppt. Thus, it appears that the lake oscillated between lengthy low water, high salinity stages when water was restricted to the central basin area, and high water, low salinity stands when lake level rose above C2 and salinity declined to <3 ppt.

6.4.2 7300–4400 BP

Paleohydrological change within fossil zone 1 is best interpreted as a continuation of this dynamic, but generally low water setting. Zone 1 was originally defined on the composite pollen percentage diagram (Fig. 19) by generally high Cheno–Am, *Ambrosia*, and *Ruppia* percentages, although all three taxa (in particular *Ruppia*) display considerable fluctuations. A detailed pollen and plant macrofossil diagram spanning zone 1 in core C1 (Fig. 26), illustrates the rapidity with which lake level and salinity changes evidently occurred.

High Cheno–Am pollen percentages are best explained as either representing shoreline location in the vicinity of coring site C1, or simply colonization of the basin floor by annuals able to withstand highly saline conditions on exposed mudflats, such as *Salicornia europaea* var. *rubra* (Ungar 1974). Either explanation embodies low lake levels, with resultant high salinities. A similar explanation also accounts for high *Ambrosia* pollen percentages, since

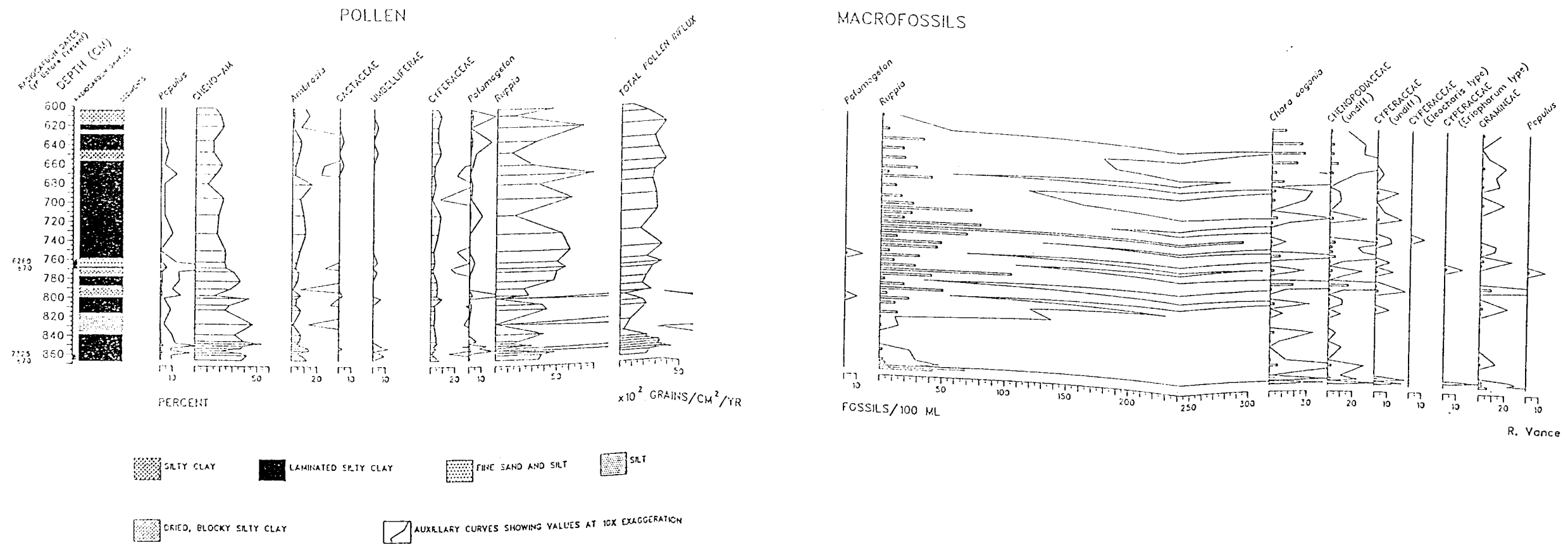
the only species of this genus native to Alberta, *Ambrosia artemisiifolia*, is restricted to drying slough surfaces (Moss 1983). The low total pollen influx of zone 1 is also consistent with this interpretation, as such an environment would likely have a sparse vegetation cover and repeated wet/dry cycles would have created a poor environment for preservation (Holloway 1989).

Watts and Winter (1966) interpret high but fluctuating Chenopod and *Ambrosia* pollen percentages in mid-Holocene sediment from Lake Carlson, Minnesota, as representing low water stands at a time when the lake was experiencing pronounced annual water level fluctuations. Complete dessication of the basin was not implicated, since the expected increased representation of degraded pollen and deposition of distinctive sediment did not occur. Both these features are a component of fossil zone 1 at Chappice Lake, indicating occasional drying did occur.

In addition to high percentages of degraded pollen throughout zone 1, the varied and distinctive sedimentary sequence is suggestive of repeated dessication. The earliest and most compelling evidence is found within a unit of olive brown silt between 8.37 and 8.17 m (Fig. 15), that contains a very sparse botanical record compared to adjacent units (Fig. 26). In this silt, total pollen influx declines to the lowest recorded values and plant macrofossil concentration is consistently low. As well, *Ruppia* pollen and seed representation plummets. The sparse pollen spectra are dominated by Chenopod and only chenopod seed remains are found. Loss-on-ignition is the lowest recorded. This suite of features suggest a period of lake dessication during which the basin was blanketed by aeolian silt.

Because of abrupt sedimentary changes throughout zone 1 and the likelihood that portions of the sedimentary record were removed by deflation, interpolation between radiocarbon dates, assuming constant sedimentation rates, is hazardous but necessary to establish a preliminary chronology. By such a process, the timing of loess deposition in the

Figure 26. Detailed pollen and plant macrofossil record of zone 1, core C1.



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Chappice Lake basin was likely between approximately 7000 and 6800 BP.

Blocky, dried sediment from 7.98–7.88 m containing contorted, interrupted laminations, is likely also a product of dessication. This suggests a setting in which the playa surface was moistened only occasionally, causing intrasedimentary crystallization of salts and disruption of laminations (Last 1984). Alternatively, exposed lake bottom sediments were disturbed by bioturbation or mudcrack formation. The presence of a low water table is also indicated by the botanical assemblage, since Chenopodiaceae and *Ruppia* pollen percentages decline abruptly, as does total pollen influx. The occurrence of Cactaceae pollen and a peak in Gramineae seeds is also suggestive of periodic drying, as these taxa are most commonly found in dry, sandy soil of the prairie upland surrounding the basin today (Chapter 5). Thus, a periodically dry playa surface likely prevailed between approximately 6600 and 6500 BP.

Overlying sediment contains a second unit of dried, blocky sediment (7.76–7.69 m), although in this case laminae are not present (Fig. 16). Pollen influx is low and dominated by Chenopodiaceae and *Ambrosia*. *Ruppia* pollen percentages are high within this unit, but its seeds are rare. Sporadic appearances of Gramineae and Chenopodiaceae seeds suggest these taxa periodically colonized the playa surface. As a result, this unit also appears to record a brief lowering of the water table, from approximately 6375 to 6300 BP.

A silt layer between 7.66 and 7.59 m (Fig. 16), containing an abundance of Chenopodiaceae seeds, a twig fragment (cf. *Potentilla*), and numerous small pebbles, is also indicative of a dessication period. This assemblage may record colonization of the playa surface by shoreline taxa followed by infilling, with slopewash delivering coarse sediment and plant debris to the central basin area. High concentrations of *Ruppia* seeds and *Chara* oogonia deposition in this unit are also suggestive of periodic surface water accumulations. Plant material recovered from this unit produced a ^{14}C age of 6260 ± 70 BP.

Finally, two units of light yellowish brown silt (6.26–6.25 m, and 6.02–6.00 m) interrupting deposition of laminated silty clay are indicative of massive salt precipitation in a hypersaline brine. Since *Chenopodiaceae* seed concentrations in these units indicate reduced lake area, and abundant *Ruppia* and *Chara* propagules suggest water was present, these units are best interpreted as brief intervals of shallow water, hypersaline conditions. These low water high, salinity stages date to approximately 4675 and 4400 BP.

In addition to evidence of draw-down and dessication, zone 1 also contains evidence of periods when basin infilling occurred rapidly. Periodic increases in *Populus* and Umbelliferae pollen percentages are suggestive of occasionally high, relatively freshwater conditions. Similar inferences may be made from the macrofossil assemblage in basal sections of C1, that includes a *Populus* seed and achenes of *Eriophorum* type (Cyperaceae). As these indicators of relatively fresh, highwater conditions appear adjacent to levels containing evidence of draw-down, water table fluctuations may have been rapid throughout zone 1.

The C2 macrofossil record (Fig. 24) also bears evidence of rising water levels between approximately 7100 and 6500 BP. Increased concentrations of *Ruppia*, *Chenopodiaceae*, Cyperaceae, and Gramineae seeds, as well as *Chara* oogonia recovered from within a laminated silty clay matrix, are evidence of shoreline presence at site C2. *Chenopodiaceae* seeds at the base of C3 (approximately 6100 BP) also indicate shoreline proximity. Evidently lake levels rose sufficiently to move shoreline upslope of C2 on more than one occasion, reaching coring site C3 once.

In summary, evidence from fossil zone 1 indicates that from 7300 to 6000 BP Chappice Lake experienced pronounced water level fluctuations, ranging from complete dessication to short-lived high, relatively freshwater stands. Bearing in mind the limitations of calculating the age of basal C2 sediments, it appears that lake level fluctuations documented in fossil zone 1 are a continuation of similar, but less well-documented lake level fluctuations beginning as early as 8500 BP.

Whereas changes in the composite pollen diagram are insufficient to warrant designation of subzones within zone 1, sedimentary characteristics and aspects of the C1 macrofossil record indicate that lake levels at Chappice Lake fluctuated less and began to rise shortly after 6000 BP. Deposition of silty clay sediment began about 6000 BP and continued uninterrupted throughout the remainder of the sedimentary record. In addition, most silty clay sediment in the uppermost portion of zone 1 is distinctly laminated. These changes mark the onset of continuous lacustrine deposition, an event accompanied by near-continuous deposition of *Ruppia* seeds and *Chara* oogonia in concentrations much greater than those found in basal zone 1 deposits. These taxa are found in saline lakes of the northern mixed-grass prairie today (Ungar 1974), although only *Ruppia* persists at high salinities (Hammer and Heseltine 1988).

Members of the genus *Ruppia* currently occur in saline lakes of Europe (Verhoven 1979), Australia (Brock 1981) and North America (Husband and Hickman 1985, 1988). While posing some taxonomic difficulties (see Hammer and Heseltine 1988), the genus is represented by both annual and perennial forms. Annual forms are most tolerant of high salinities. For example, *Ruppia maritima* has been found in South Dakota in water with TDS content as high as 230 ppt (Ungar 1974) and 140 ppt in Saskatchewan (Rawson and Moore 1944). Hammer and Heseltine (1988) note that *Ruppia maritima*, *R. occidentalis*, and *Potamogeton pectinatus* are the only aquatic macrophytes found in hypersaline lakes (TDS >50 ppt) in western Canada. Kantrud et al. (1989), in a survey of North American prairie pothole vegetation, list *Ruppia maritima* as the most salt tolerant species.

Husband and Hickman (1985) studied aspects of *Ruppia* distribution and growth in Alberta, concluding that the genus requires saline water for maximum growth and that numbers of flowering individuals increase with salinity. Brock (1981) notes that annual forms of *Ruppia* produce prolific numbers of seeds. Thus, increased numbers of *Ruppia* seeds and pollen indicate increasingly saline conditions.

The presence of laminated sediment is in itself indicative of increased salinity. O'Sullivan (1983) outlines four requirements for the production of laminated sediment: 1) no disturbance at the sediment surface; 2) morphological suitability (that is, the lake must be flat-bottomed to minimize disturbance from slumping; 3) strong seasonal stratification of the water column, and; 4) scarcity of oxygen in profundal sediment. Chappice Lake is situated within a flat-bottomed basin and during extremely saline phases biological disturbance would cease, as benthic organisms cannot tolerate high salinities. Tims et al. (1986) noted a significant decline in diversity and abundance of benthic organisms in saline lakes of Saskatchewan where TDS exceeded 50 ppt. Rawson and Moore (1944) documented the virtual elimination of benthic organisms at 100 ppt and Hammer et al. (1975) state that no benthic fauna occur in water exceeding 121 ppt in saline lakes of the western Canadian interior.

Once formed, laminations must be preserved, and in shallow basins like Chappice Lake preservation demands protection from wind disturbance. In hypersaline lakes, stratification along concentration gradients has been observed in both shallow (Blinn 1971, Rawson and Moore 1944) and deep water (Hammer et al. 1975, Last and Slezak 1986). A dense hypersaline brine overlying surficial lake sediment would provide the necessary protection from wind required to preserve laminations.

Thus, Chappice Lake likely contained shallow, hypersaline water when laminated sediments were produced, like those found in the upper section of zone 1. A dense underlying layer of brine must exceed 120 ppt but be no greater than 230 ppt (the upper limit of *Ruppia* growth), to produce laminae with *Ruppia* remains.

Considering the abundance of *Ruppia* pollen and seeds in the finely laminated sediment throughout the upper 1.5 m of zone 1, such conditions must have prevailed for a considerable length of time. Evidently the rapid lake level changes outlined for the pre-6000 BP period had ceased, and conditions in the lake remained at a relatively constant shallow water high salinity phase. *Chara* presence suggests that either the overlying brine was less

than 100 ppt on occasion, or that the alga was confined to less saline marginal ponds and its oogonia subsequently washed into the high salinity basin, a process observed today in Australian salt lakes (Burne et al. 1980).

At 5100 BP laminated silty clay began to accumulate at coring site C2, indicating that overall lake level was gradually rising during this period of relative stability. Shoreline passage over site C2 is documented by deposition of abundant *Ruppia*, Chenopodiaceae, and Cyperaceae seeds, followed by accumulation of fine-grained limnic sediment with increased organic matter compared to underlying sands (Fig. 24). Evidently the basin had not yet completely filled, as high Cheno-Am and *Ambrosia* pollen percentages indicate that extensive shoreline zones remained exposed (perhaps reflecting a continuation of pronounced seasonal water level fluctuations) and limnic sediment was not yet deposited at C3.

6.4.3 4400-2650 BP

By the onset of zone 2 (4400 BP), lake levels had risen above coring site C2, eventually reaching coring site C3 at 3350 BP, as indicated by increases in *Ruppia*, Chenopodiaceae, and Cyperaceae seed concentration (Fig. 25). Continued basin infilling accounts for reduced Cheno-Am and *Ambrosia* pollen percentages, since suitable habitat for these taxa was shrinking as lake margin flats were restricted in size and seasonal water level extremes became less pronounced. Throughout zone 2, Chappice Lake was perennial and did not experience major lake level fluctuations of the scale outlined for earlier periods, as indicated by the lack of thick, coarse-grained deposits in all core sections spanning this interval. Nevertheless, subtle changes in sedimentation and fossil assemblages in all cores indicate that less severe lake level fluctuations have taken place over the last 4400 years. The key indicator of reduced lake level and increased salinity is laminated lake sediment containing an abundance of *Ruppia* pollen and seeds.

The Chappice Lake pollen record (Fig. 21) reveals that a rise in total pollen influx occurred in subzone 2a, likely reflecting both a better environment for preservation and an overall increase in local vegetation cover. *Ruppia* pollen influx increases, although percentage representation declines compared to zone 1 (Fig. 19). A gradual decline in *Ruppia* pollen frequency occurs throughout zone 2a.

The subzone 2a macrofossil assemblage of core C1 also documents a gradual decline in *Ruppia* seed concentration (Fig. 23). In addition, Chenopodiaceae seed concentration declines slightly from zone 1. Similar trends are evident in the C2 macrofossil record (Fig. 24). Declining representation of these taxa likely reflects upslope shoreline movement accompanying basin infilling, first above site C1 and later upslope of C2. Moreover, the rise in *Chara* oogonia concentration registered in the L1 (Fig. 22) and C1 (Fig. 23) macrofossil records, suggests changes to other aspects of the lake.

While *Chara* is considered a common component of saline lake flora in North America, it is abundant in shallow water of reduced salinity, compared to *Ruppia* (Kantrud et al. 1989). In Canadian surveys, figures cited for the *Chara* threshold of salinity tolerance vary from 80 ppt (Rawson and Moore 1944) to 25 ppt (Hammer and Heseltine 1988). Burne et al. (1980) report *Chara* in Australian lakes to 70 ppt. Although variable, these values are consistently below the minimum required to eliminate bottom dwellers and to preserve laminae that are so common throughout subzone 2a. As a result, a salinity balance with a sufficiently saline bottom layer to eliminate benthic organisms and less saline marginal brine to allow *Chara* growth must have been maintained between 4400 and 2650 BP.

In addition to these relatively stable conditions, the sudden rise in *Chara* representation suggests water chemistry may have been different in zone 2a from what exists today. Hutchinson (1975) notes that *Chara* has the ability to utilize bicarbonate as a carbon dioxide source in photosynthesis, and cites Vaida's (1966) contention that *Chara* does not do well in sulphate-rich water. Water in Chappice Lake is now dominated by the sulfate anion. Thus, it

is possible that flourishing *Chara* growth throughout subzone 2a may have been the result of bicarbonate-dominated water that prevailed during the gradual overall rise in water level between 4400 and 2650 BP.

High carbonate content in Lake Manitoba sediment is thought to have resulted from dissolution of efflorescent crusts that had formed around the basin during an arid climatic period (Last and Slezak 1988). As Chappice Lake was filling during subzone 2a, its margin encroaching upon former mudflats, a similar process may account for change in water chemistry. In addition to possible dissolution of efflorescent crusts, bicarbonate sedimentation in playas has been attributed to surface runoff and shallow groundwater inflow (Last 1984, 1987). Bicarbonate dominates shallow groundwater in the Chappice Lake area today (Stevenson and Borneuf 1977); Rawson and Moore (1944) suggest that carbonate lake water in an area of sulfate-dominated lakes is indicative of spring activity. Thus, it is possible that groundwater recharge and associated spring activity may have contributed to the gradual increase in lake level recorded during subzone 2a. This suggests that as the water table rose from a mid-Holocene minimum, spring systems and groundwater flow was activated, discharging bicarbonates and dissolving carbonate-rich mudflats. Such processes could have reduced salinity (compared to zone 1) of lake margin water and generated a bicarbonate-dominated water body, thereby creating an ideal environment for *Chara* growth.

6.4.4 2650-1060 BP

Subzone 2b is defined on the basis of declining *Ruppia* pollen percentages. This coincides with the initiation of a long, uninterrupted sequence of massive silty clay deposition. Low total pollen influx and organic matter are also characteristic of sediments deposited during subzone 2b. These changes are best interpreted as representing continued recharge of the basin, eventually producing conditions unsuited for the laminae formation as well as *Ruppia* and *Chara* growth. *Myriophyllum* seed deposition at site C1 (Fig. 23) during subzone 2a supports this hypothesis, as *Myriophyllum* is restricted to less saline water than *Chara* or

Ruppia (Hammer and Heselvine 1988, Kantrud et al. 1989). Increased lake levels further moved shoreline a greater distance upslope from all coring sites (compared to 2a), reducing pollen input from near-shore vegetation and lowering total pollen influx.

The onset of prolonged massive silty clay deposition is dated at 2290 BP in core C1. Assuming constant sedimentation rates from dated horizons in core C2 and L1, laminae deposition ceased at 3100 and 2650 BP, respectively. At C3 the onset of fine lacustrine sedimentation began at 3100 BP. Evidently, a hypersaline brine layer at the base of the water column was gradually restricted in size as lake level rose and salinity declined. By 2650 BP, hypersaline brine was limited to only the deepest portion of the basin, and by 2290 BP completely eliminated.

Low macrofossil concentration at all coring locations is characteristic of sediment deposited during subzone 2b. A minor peak of *Chara* at the base of 2b in core C1 (Fig. 23), and the lone occurrence of *Chara oogonia* at the base of 2b in core L1 (Fig. 22), mark the final appearance of *Chara* in this subzone. Low Chenopodiaceae seed concentration is the only consistent feature of macrofossil deposition throughout the basin. At all coring sites, a marked decline in Chenopodiaceae seed concentration is evident in the central portion of 2b. This likely marks the highest lake level achieved during this period, as at this time distance to shoreline from all coring sites was at a maximum. The timing of this event was approximately 2000 BP (by interpolation, 2100 BP in core L1, 1950 BP in C1, 1800 BP in C2, and 2200 in C3).

Sediment accumulation rates at coring site L1 are consistently greater than nearby C1 (Fig. 18) and the difference is most pronounced throughout subzone 2b (3.45 m of sediment accumulation at L1 compared to 2.15 m at C1). This is likely related to increased erosion of the steep slope bordering the northeastern margin of the basin (Fig. 6, Bar A). During this long highwater stand, lakewater continually undermined the slope, delivering an increased sediment load to nearby coring site L1. Increased clay content in sediment deposited at L1

during this time, compared to sediment that accumulated at site C1, supports this hypothesis and suggests debris flows initiated by erosion of the northeastern shoreline transported an increased proportion of fines to nearby L1, but not C1. The occurrence of *Peromyscus maniculatus* mandibular and vertebral fragments at site L1 may also be attributed to transport from these slopes. In addition, the increased grittiness noted in sediment from the 0.60 to 0.40 m section of core C3 (approximately 2000 to 1300 BP), may be related to either increased slopewash along the southern margin of the lake or increased stream discharge, since streams in this area (Fig. 6) were activated during this prolonged highwater stand. The overall decline in aquatic flora throughout 2b may also be related to an increased load of suspended sediment, effectively reducing light penetration and photosynthesis.

6.4.5 1060–600 BP

The composite pollen diagrams (Figures 19 and 21) record a rapid rise in *Ruppia* representation and rising Cyperaceae values at the onset of 2c. Increased Cruciferae pollen and the occurrence of *Ephedra* pollen further distinguish subzone 2c from 2b. The onset of subzone 2c is also correlative with sedimentary changes in cores L1 and C1, as lengthy sections of massive silty clay are succeeded by units of finely laminated sediment. These changes indicate lake lowering occurred, causing increased salinity, vigorous *Ruppia* growth, and re-establishment of a dense hypersaline brine at the base of the water column. Interpolation from carbon-dated laminae marking the beginning of 2c at coring site L1 (the core containing the greatest detail regarding recent subzones owing to its rapid sediment accumulation rates) indicates that subzone 2c persisted for some four centuries, from 1060 to 600 BP.

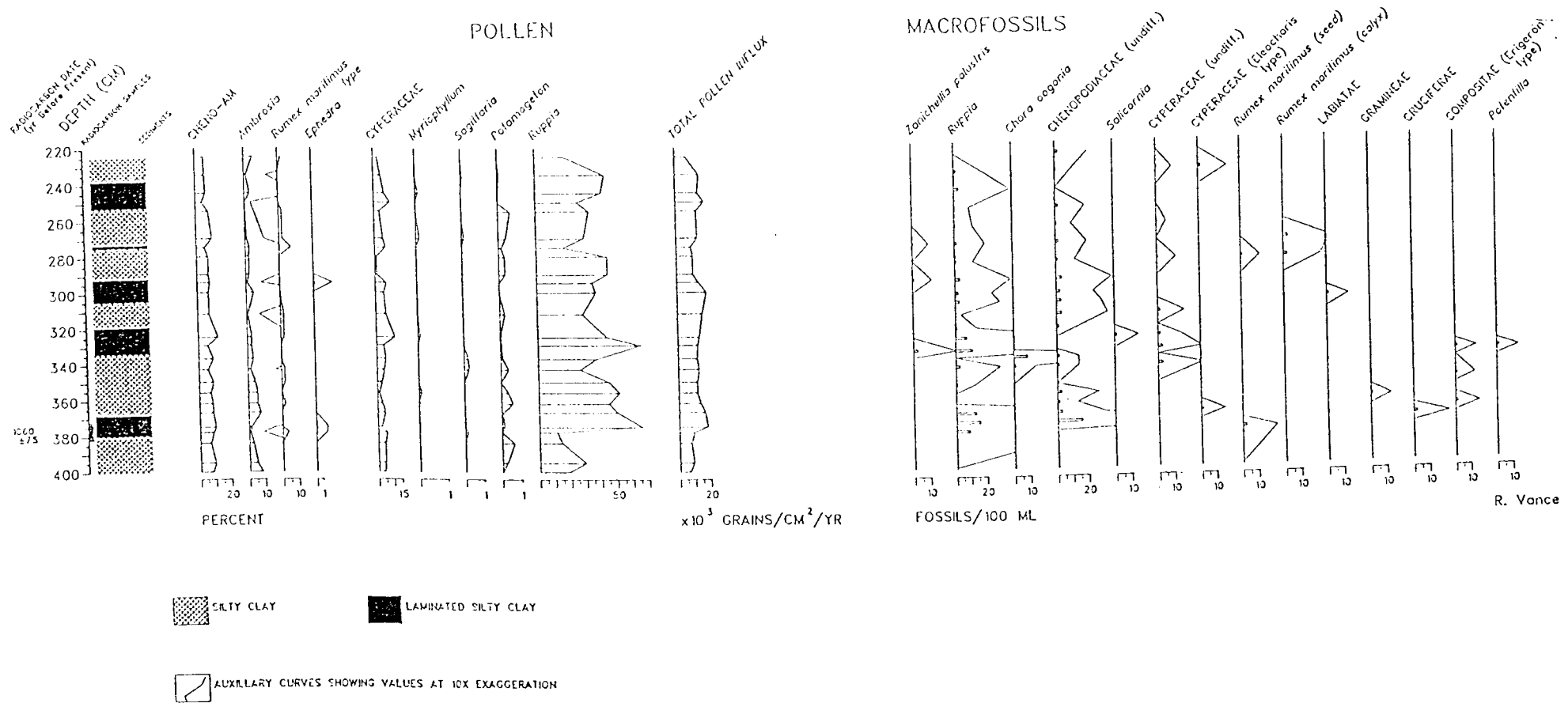
Macrofossil evidence from core L1 supports the interpretation of increased salinity and declining water levels, as *Ruppia* and Cyperaceae seed concentrations increase abruptly at the onset of subzone 2c (Fig. 22). In addition, Chenopodiaceae seed accumulation is marginally increased over 2b, an indication that distance to shoreline had decreased (that is, shoreline

had moved downslope with decreasing lake size to a position closer to L1, compared to 2b). Sporadic occurrences of *Rumex maritimus* seeds and calyces are also suggestive of shoreline closure. Moreover, occasional deposition of *Zanichellia palustris* seeds and *Chara* oogonia suggest water depth decreased at coring site L1.

Draw-down of Chappice Lake during 2c was evidently not as great as at earlier times, since uninterrupted deposition of massive silty clay sediment continued at sites C2 and C3. Although interpolation from radiocarbon dates provides only an approximate chronology of recent events at C2 and C3 (Figures 24 and 25), it is noteworthy that the only occurrence of *Ruppia* seed deposition at each coring locale falls roughly within the boundaries of 2c, supporting the proposed increase in salinity. Further, slight increases in Chenopodiaceae seed deposition within 2c at C2 and C3, as well as the sole occurrence of *Salicornia* and Polygonaceae seed deposition in subzone 2c of C3, are additional indicators of a reduction in distance to shoreline that accompanied lake level draw-down. Evidence from core C1 supports this scenario, as *Ruppia* seed concentration increases in 2c, accompanied by increased Gramineae, *Salicornia*, *Rumex maritimus*, Cyperaceae, and Compositae seed deposition, compared to subzone 2b (Fig. 23).

A detailed record of sedimentary changes and a summary of accompanying variations in fossil assemblages within subzone 2c (Fig. 27) shows that deposition of laminae was not continuous. Although *Ruppia* pollen percentages are high throughout subzone 2c, peaks tend to occur in association with laminations, as do increases in *Ruppia* seed concentration. When massive silty clay sediment is deposited, *Ruppia* pollen and macrofossil abundance declines and *Potamogeton* pollen frequency peaks. In addition, *Sagittaria* and *Myriophyllum* pollen deposition tend to coincide with deposition of massive sediment. Evidently, high water episodes of reduced salinity also occurred within 2c, as these taxa are restricted to less saline water than *Ruppia* (Hammer and Heseltine 1988, Kantrud et al. 1989). Occurrences of *Ephedra* are indicative of long-distance transport to the site, as this genus is not native to Alberta

Figure 27. Detailed pollen and plant macrofossil record of zone 2c, core LL.



(Moss 1983).

Thus, details of subzone 2c show that rather than a prolonged low water stand, a series of low water episodes were punctuated by high water, low salinity stands. Despite the likelihood that fluctuating water levels may obscure details of the fossil record by, for example, transporting remnants of a low water stand from the lake margin to the central lake area when infilling occurs, patterns of fossil distribution appear capable of recording relatively brief water level fluctuations at Chappice Lake.

6.4.6 600–100 BP

Declining *Ruppia* pollen frequency separates subzone 2d from 2c. This change indicates rising lake levels, once again coinciding with elimination of a hypersaline brine layer at the sediment water interface, thus ending laminae formation. Short laminated sequences exist within subzone 2d in cores L1 and C1, but are too brief to be portrayed on Figures 22 and 23. In core L1, weakly laminated sections occur at 2.07–2.06 m, 1.73–1.71 m, 1.41–1.39 m, and 1.19–1.17 m. These features hint at short-lived draw-down events, and may be related to occurrences of *Ruppia*, Chenopodiaceae, Cyperaceae, and *Chara* propagules that appear sporadically in the L1 and C1 macrofossil records (Figures 22 and 23).

As a result of low lacustrine sediment accumulation rates in cores C2 and C3, subzones 2d and 2e are confined to the uppermost few centimeters. This, combined with a lack of radiocarbon dates from these layers, makes distinction between subzones 2d and 2e tenuous. A puzzling aspect of subzone 2d is that it marks the first appearance of *Selaginella densa* spores in significant amounts. This likely records colonization of sandy, exposed habitat in the catchment (Moss 1983), but it is unclear why this species is not as abundant in earlier dry episodes.

In sum, subzone 2d is best interpreted as a high water stand that was punctuated by a series of short-lived lake level declines, the most severe of which (as judged by the length

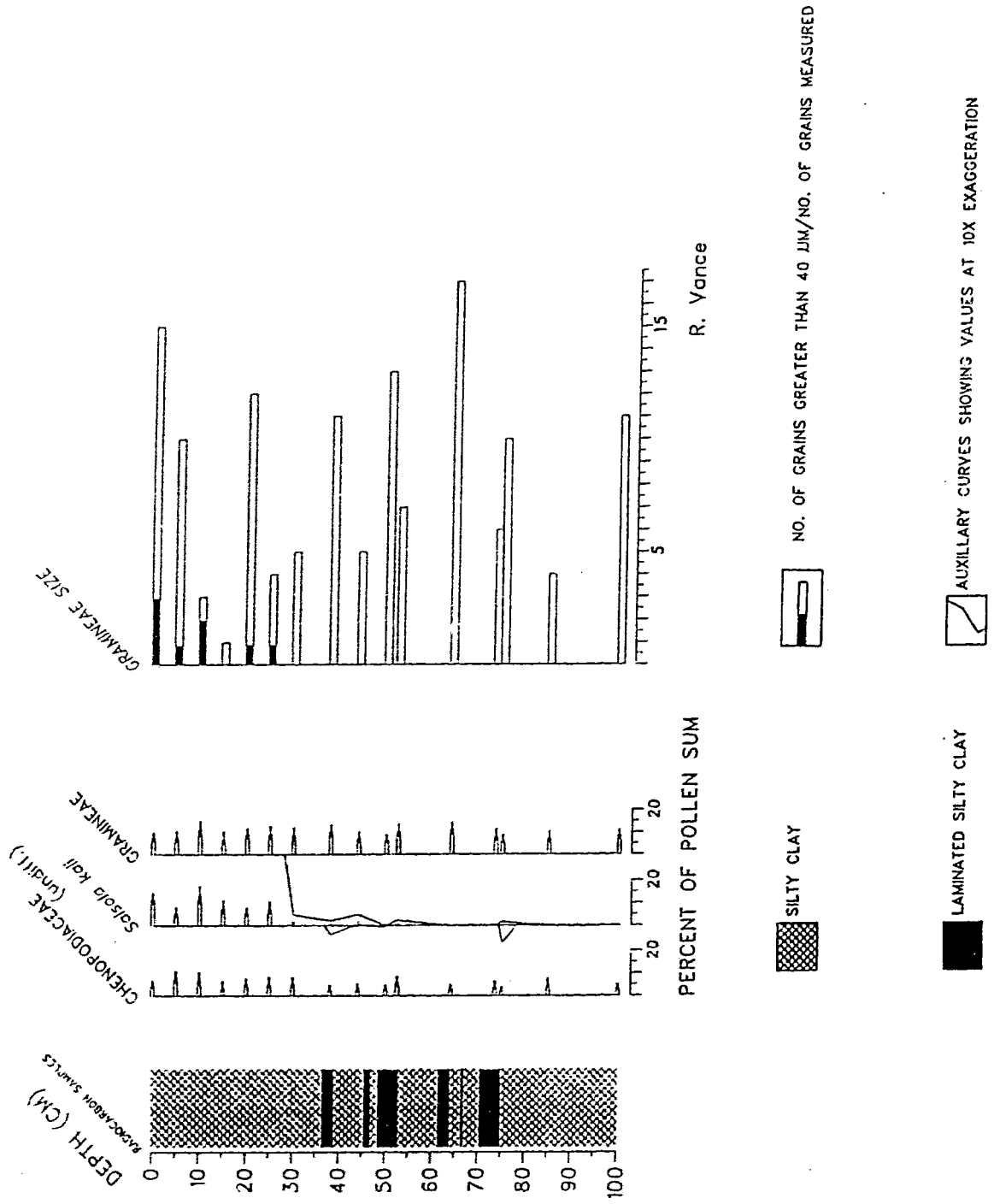
of laminated sections) occurred near the termination of subzone 2d.

6.4.7 100 BP-present

The uppermost portion of the Chappice Lake record is distinguished by significant amounts of *Salsola kali* pollen. As *Salsola kali* is not native to Alberta (Moss 1983), its occurrence is a useful chronological marker for recent events. A detailed diagram of the uppermost meter of L1 sediment in Chappice Lake, including the occurrence of *Salsola kali* pollen Gramineae grain size distribution (Fig. 28), exemplifies the hazards of relying on interpolation from radiocarbon dates to age sediments that are not as yet compacted, as in the case of the uppermost levels of all cores collected.

Figure 28 shows that although *Salsola kali* is abundant only in the upper 0.30 m, it appears occasionally in underlying sediment, making its first appearance between 0.80 and 0.75 m. Gramineae grains $>40\mu\text{m}$ in size, considered to include mainly cultivated grasses (with some exceptions, Faegri and Iversen 1975), are confined to the uppermost 0.30 m of sediment. This suggests that the upper 0.30 m of L1 marks the time that land within the immediate vicinity of Chappice Lake was cultivated, whereas earlier sporadic occurrences of *Salsola kali* mark the beginning of agriculture in the region, as the plant was likely imported to North America mixed with crop seeds (Mack 1986). The first agriculturalists in the Medicine Hat region arrived in 1880 (Jones 1987). *Salsola kali* was noted in abundance in the Chappice Lake area in the mid-1930's, covering fields of deserted farms (Wyatt et al. 1937). Thus, it is possible to revise the chronology of the upper meter of L1 sediment, utilizing the earliest *Salsola kali* occurrence (0.80 m) as a marker for the 1880's. The peak in *Salsola kali* pollen abundance coincident with the occurrence of large diameter Gramineae grains, may therefore date to the 1940's, corresponding to the earliest aerial photography showing cultivated fields directly north of Chappice Lake. If these assumptions are correct, the laminated sections in the upper portion of core L1 represent lake draw-down associated with droughts during the late 1800's, 1920's, and 1930's, thereby establishing a link between

Figure 28. Detailed pollen record of the upper meter of core LI.



prehistoric and historic portions of the record. This tantalizing prospect will await further investigation into cultivation practices within the Chappice Lake watershed, as aerial photography extending to the 1920's and 1930's required to verify the proposed chronology does not exist. Until such time, this account will be adopted as a tentative timescale of recent events.

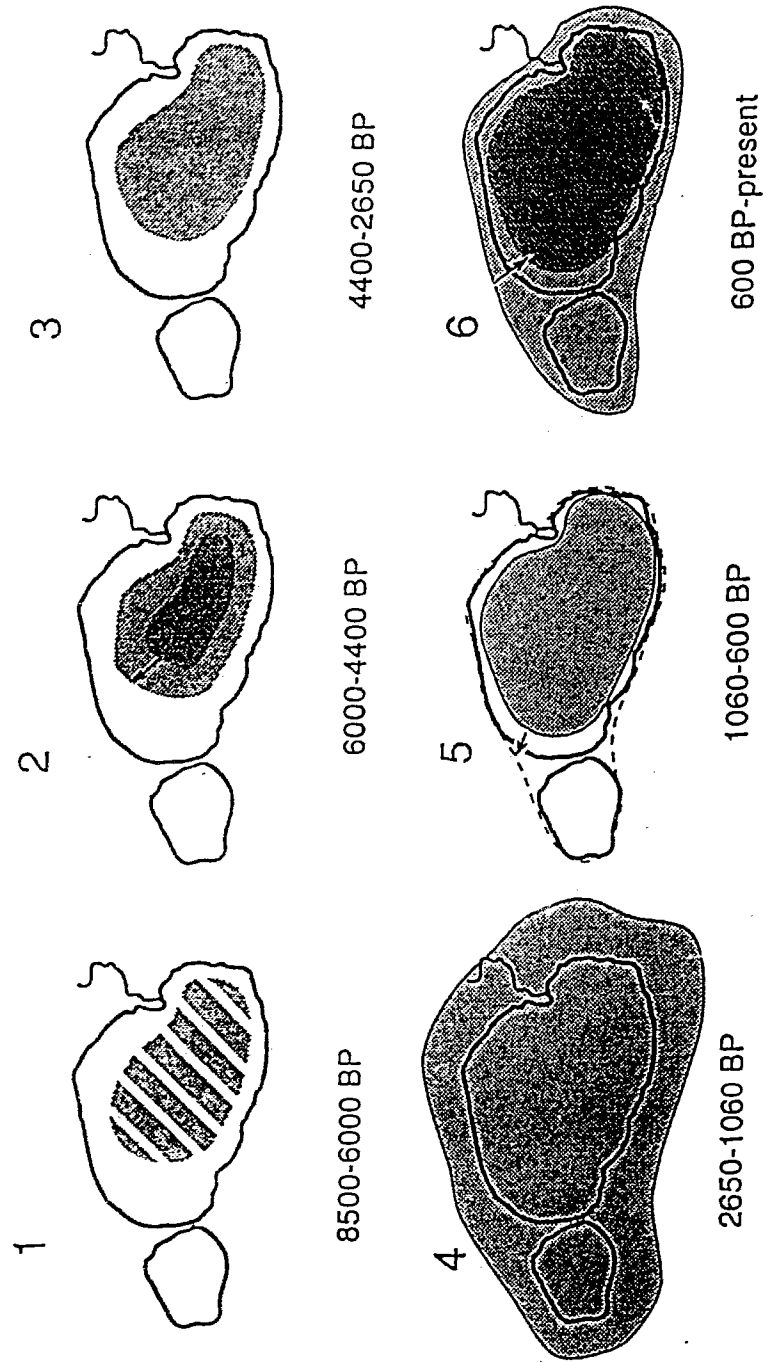
Direct evidence linking deposition of laminated sediment and lake draw-down is available for the 1980's, however. The uppermost portion of core L1 collected in 1984, does not contain laminae. In contrast, the uppermost 0.05 m core C1 collected in 1989, consists of finely laminated sediment. As declining lake levels have been a feature of the last decade of lake history (Chapter 4), it appears that the lake has only recently crossed the lake level/salinity threshold required to stimulate deposition and preservation of these features. Surface water collected from Chappice Lake in August of 1989 has a TDS content of 159 ppt (Chapter 4).

The recovery of thenardite crystals at the sediment/water interface of all cores collected in 1989 indicates sodium sulfate saturation, an aspect of lake chemistry that may stimulate formation of laminae, as well as promote preservation of the features. Thenardite deposited in winter is likely dissolved in spring with rising water temperature, thereby contributing to development of a hypersaline brine layer at the sediment/water interface and protecting laminae from wind disturbance. No salt crystals at the sediment/water interface were encountered during the 1984 coring expedition.

6.5 Conclusion

The Chappice Lake paleobotanical and sedimentary record indicate that the lake has passed through six main phases of development during the Holocene (Fig. 29). During phase 1 (8500-6000 BP) the water table in the Chappice Lake basin was generally low but

Figure 29. Summary of Chappice Lake paleohydrology (refer to text for explanation of shading).



variable. The lake dried completely four times (7000–6800 BP, 6600–6500 BP, 6375–6300 BP, and 6250 BP), but also experienced periods of relatively high, freshwater stands (7100 to 6500 BP and 6100 BP). Since only one of these events (6250 BP) is directly dated by radiocarbon, the remainder are estimates derived by interpolation between dated layers. Errors associated with the assumption of constant sedimentation rates likely account for the overlap of high lake levels and dessication events around 7000 BP. In general, low but variable lake levels dominate the record from 7300 BP (and perhaps as early as 8500 BP, based on fragmentary evidence from one core) to 6000 BP.

During phase 2 (6000–4400 BP) the lake remained at a generally low level, rising gradually (from the dark to light shading on Fig. 29.2) but remaining highly saline. Extreme water fluctuations that characterised phase 1 had ceased, but large seasonal water level fluctuations likely persisted. The lake was reduced to extremely low levels at 4675 BP and 4400 BP.

Chappice Lake continued to rise and freshen throughout phase 3 (4400–2650 BP). Lake water was likely dominated by the bicarbonate anion, as opposed to sulfate today, suggesting that shallow groundwater inflow was an important component of the hydrological budget. In general, lake level throughout phase 3 remained low, but was more stable than in phases 1 and 2.

The highest lake level stand in the record occurred from 2650 to 1060 BP (phase 4). Lake salinity declined from earlier phases. The lake attained its largest size at approximately 2000 BP.

A series of low water, high salinity lake stands punctuated phase 5, from 1060 BP to 600 BP. Lake levels were predominantly low (covering the shaded area of Fig. 29.5), but rose on occasion (its upper limit outlined by the dashed line on Fig. 29.5).

From 600 BP to the present, Chappice Lake has generally remained at a high water stand (shaded area of Fig. 29.6), but has periodically been reduced in size (dark shading in Fig. 29.6). In phase 6, the longest low water stands have occurred in the last century. From 600-100 BP low water intervals were brief and occurred aperiodically.

CHAPTER 7

HOLOCENE PALEOCLIMATOLOGY

7.1 Reliability of Chappice Lake Paleohydrology as a Proxy Record of Climate

As all saline lakes of the northern interior plains require groundwater inflow for survival (Last and Schweyen 1983), the reliability of lake level changes at Chappice Lake as a proxy record of prehistoric climate largely depends upon the role of groundwater in the hydrological balance. In other words, is subsurface inflow directly linked to local groundwater levels, thereby reflecting climatic conditions with only a limited response time, or is the lake fed by deep groundwater, a source little affected by minor climatic oscillations? As the hydrology of northern prairie wetlands is known only superficially and much work needs to be done to identify and evaluate factors that control the interaction of lakes and groundwater (Winter 1977, 1989), this question cannot be answered directly. There are however, geologic and topographic features of the Chappice Lake catchment, as well as aspects of the lake's water chemistry and historic water level fluctuations, that suggest the lake is, and likely has been, sensitive to climatic change.

The most important aquifer in southeastern Alberta is the Oldman Formation (Stevenson and Borneuf 1977). This Cretaceous sandstone and shale unit underlies approximately 50 m of ground moraine in the Chappice Lake watershed (Carlson 1970), and likely contributes to subsurface inflow. The Oldman Formation slopes toward the northwest, in contrast to surface drainage which trends southeast. Although the Oldman Formation underlies most of the Chappice Lake drainage basin, the older Foremost Formation underlies Quaternary deposits on the lake's western margin. As Foremost sandstone, siltstone and shales are known to hold brackish groundwater (Stevenson and Borneuf 1977), they must also be considered a possible source of subsurface flow.

Like bedrock aquifers, ground moraine deposits may be an important component of the local groundwater system. Drift deposits are suspected to form an integral part of groundwater flow to Old Wives Lake (Freeze 1969a, b), a sodium sulfate lake situated some 300 km east of Chappice Lake in a similar geologic setting. In addition to ground moraine overlying the Foremost and Oldman Formations, some 30 m of drift lies upon the Bearpaw Formation directly northwest of Chappice Lake (Carlson 1970). As Chappice Lake is situated at the base of this local upland, subsurface drainage through ground moraine deposits may also direct flow to the lake.

Chappice Lake lies within a meltwater drainage channel, another feature likely directing subsurface flow. This meltwater channel gently slopes toward Chappice Lake from the north. A spring feeds the channel along its steep-sided western flank (Stevenson and Borneuf 1977), and a lush growth of sedges currently residing within the channel bottom indicates a high water table, probably related to near-surface groundwater flow. Sand and gravel deposits within meltwater channels are regarded as groundwater conduits elsewhere on the northern interior plains. For example, groundwater movement through a meltwater channel is considered an important component in the water balance of Big Quill Lake (Whiting 1977). These highly permeable deposits may also drain Chappice Lake.

All of the abovementioned potential groundwater sources are recharged by surface infiltration and, with the exception of the deep Foremost Formation aquifer, should respond rapidly to changes in the balance between precipitation and evaporation. Evaluation of their relative contribution to Chappice Lake is difficult, but since each aquifer bears a distinct chemical signature, the ionic balance of the lake may indicate which groundwater sources are involved. In surficial aquifers, where water enters the system through infiltration and flows to areas of discharge in a relatively short time, calcium and bicarbonate tend to dominate (Stevenson and Borneuf 1977, Last and Slezak 1986). If the surficial system discharges into a regional low, as it does at Chappice Lake, sodium and sulfate are often the dominant ions

(Stevenson and Borneuf 1977), their prevalence likely due to enrichment by cation exchange (Cole 1926) and oxidation of sulfur in weathered till (Hendry et al. 1986). Water rich in sodium and sulfate is also characteristic of the Foremost Formation, but is more saline than near-surface flows (Stevenson and Borneuf 1977). Mixing of groundwater from different sources will of course, obscure chemical differences. As water in Chappice Lake is currently dominated by sodium and sulfate and is at a historic low, some combination of surficial and deep aquifers dominated by sodium sulfate are the most likely sources of current inflow, as most surficial aquifers were probably depleted by recurrent drought in the 1980's.

In addition to local aquifers, there exists a possible connection to a regional groundwater flow system in the form of a preglacial valley that passes beneath Chappice Lake, draining westward to the South Saskatchewan River (Stevenson and Borneuf 1977). Locals consider this to be the main source of subsurface flow to Chappice Lake (J. Barrie, personal communication, 1988). Bedrock channels are considered excellent aquifers (Farvolden 1960) and it has been suggested that they act as conduits delivering groundwater to saline lakes on the northern plains (Last 1989), thereby playing a critical role in determining lake levels, particularly in sodium sulfate lakes (Freeze 1969a, Last and Slezak 1987, Whiting 1977). Several bedrock channels radiate from the Cypress Hills (Borneuf 1971, Freeze 1969a), including the channel passing beneath Chappice Lake. As these channels all originate in the Cypress Hills, recharge must take place on this topographic high. Although the time required for water to pass from the Cypress Hills to Chappice Lake is unknown, it is likely considerable. However, if the flow system is in dynamic equilibrium, then water must be continually added in upland recharge areas to replace water discharged in lowlands (Menely 1963). In this sense, groundwater discharge to Chappice Lake through the bedrock channel may respond rapidly to precipitation changes on the Cypress Hills.

In summary, although the contribution made by groundwater to the Chappice Lake water balance is unknown, potential sources include both shallow and deep local aquifers, as

well as a deep regional system. Interaction among all potential components cannot be ruled out. In fact, Chappice Lake may owe its lengthy existence to the variety of potential groundwater sources in the region. The response time of these subsurface flows to climate is unknown; however, for the purpose of reconstructing a long-term climatic record (particularly when relying on radiocarbon dates with typical confidence limits on the order of 100 years), the lake is considered a reliable barometer of regional climatic change, since local surface and subsurface flow recharges the basin (as well as a possibly more-distant source, the Cypress Hills). These varied sources of input are balanced by evaporative losses and possible groundwater outflow. Finally, although groundwater flow may have been less responsive in prehistoric time, historic records indicate that the lake responds rapidly to climatic change at present (Chapter 4), implying that lake size, depth, and chemistry are governed mainly by prevailing climatic conditions.

A further constraint on the sensitivity of the Chappice Lake sedimentary record to climate concerns the completeness of the record. As previously discussed, prior to 6000 BP the lake dried repeatedly, and loss of record due to deflation likely occurred. Since 6000 BP no dessication has occurred and losses through deflation are not a concern. However, at times when the lake was at or near the concentration threshold required to produce and preserve laminae, it is possible that some laminae were lost through subsequent disturbance. This situation likely prevailed in the last 600 years of lake history, suggesting that the catalogue of low lake level stands for this period records only the most severe events. As recent decades of drought (the late 1800's, 1920's, 1930's, and 1980's) are a part of the record, the reconstruction of the last 600 years is best viewed as one recording only lake draw-down similar to that experienced during these historic drought intervals.

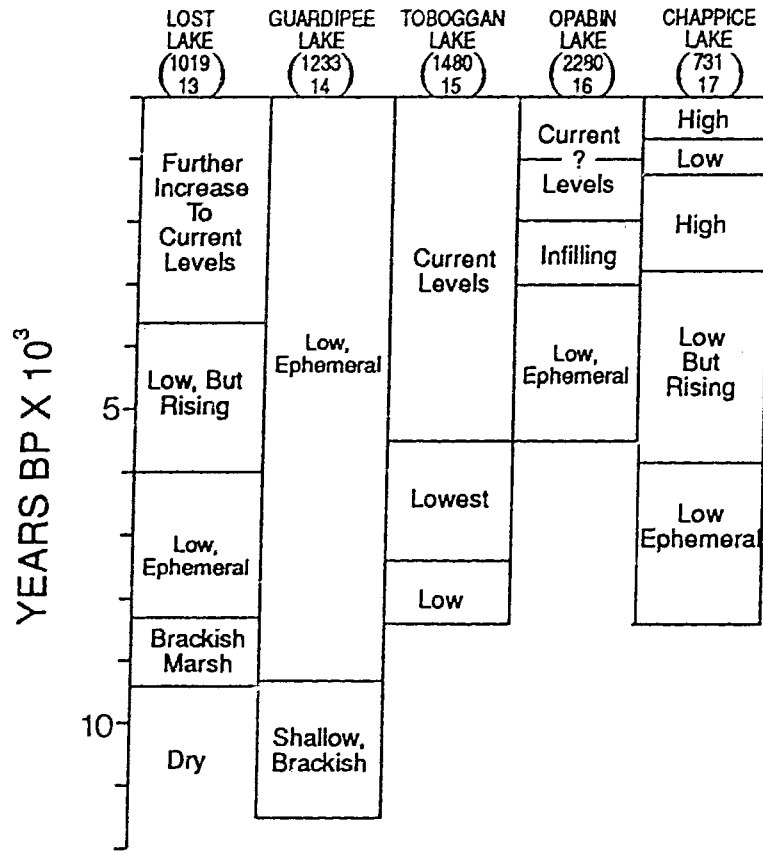
7.2 Prehistoric Climate of the Northern Mixed-Grass Prairie

A further test of the reliability of reconstructed lake levels as a proxy record of climate is how accurately regional paleoclimatic changes derived from other sites are reflected by Chappice Lake paleohydrology. A summary of past lake level changes at Chappice Lake compared to nearby AMS dated sites (Fig. 30), shows that lake level changes at all sites display reasonable synchronicity, suggesting climatic change is the main driving force behind Holocene lake level fluctuations in the area. Lake level changes at Chappice Lake reflect major paleoclimatic events inferred from other proxy records (Chapter 3) and provide additional information, particularly concerning the pace and direction of late Holocene climatic change.

Prior to 6000 BP, the water table in the Chappice Lake area was generally low but variable. Repeated lake dessication and short-lived high water stands were both features of this early stage of lake development. This interpretation indicates several millenia of climate with severe droughts (the most severe of the entire record), punctuated by brief moist intervals.

The early to mid-Holocene is widely regarded as a time of generally warmer and drier conditions throughout North America, although the timing of peak aridity is time-transgressive (Anderson et al. 1989, Baker 1983, Barnosky et al. 1987b, Winkler et al. 1986, Wright 1983). Widespread low lake levels are recorded throughout the mid-latitudes of western North America as early as 9000 BP (Harrison and Metcalfe 1985). The character of this climatic regime is generally attributed to orbital variations, since at 9000 BP perihelion occurred in summer (as opposed to winter today), and axial tilt was at a maximum, producing in high northern latitudes an 8% increase in summer insolation and a corresponding 8% decrease in winter insolation (Kutzbach and Geutter 1986). A direct result of orbital variations would be amplification of high northern latitude seasonality, between 15,000 and 6000 BP (ibid.). Thus,

Figure 30. Comparison of Chappice Lake paleohydrology with lake level fluctuations at nearby AMS dated study sites. Elevation (m asl) and location on Figure 4 indicated in brackets below each site name.



increased summer temperature was likely a factor in the extreme drought events recorded at Chappice Lake prior to 6000 BP.

In addition to changes in the earth's radiation balance, changes in upper atmospheric wind patterns must have occurred during deglaciation. Computer simulation of glacial atmospheric circulation patterns suggests that the orographic barrier of the Laurentide ice sheet

would have created a split westerly flow, with one branch of the jet stream passing to the north of the ice mass and a second branch flowing south (Kutzbach and Geutter 1986). Since global wind circulation patterns provide the most readily observed mechanism of climatic variations (Lamb 1972), changing patterns of high altitude westerly flow must be considered in paleoclimatic reconstructions. Though computer simulations provide insight into global wind patterns, details of regional westerly flow required for climatic reconstructions in a specific geographic region cannot be resolved. Thus, reconstructions for a specific area may most effectively be achieved through synthesis of regional paleocological data in conjunction with consideration of factors affecting westerly configuration.

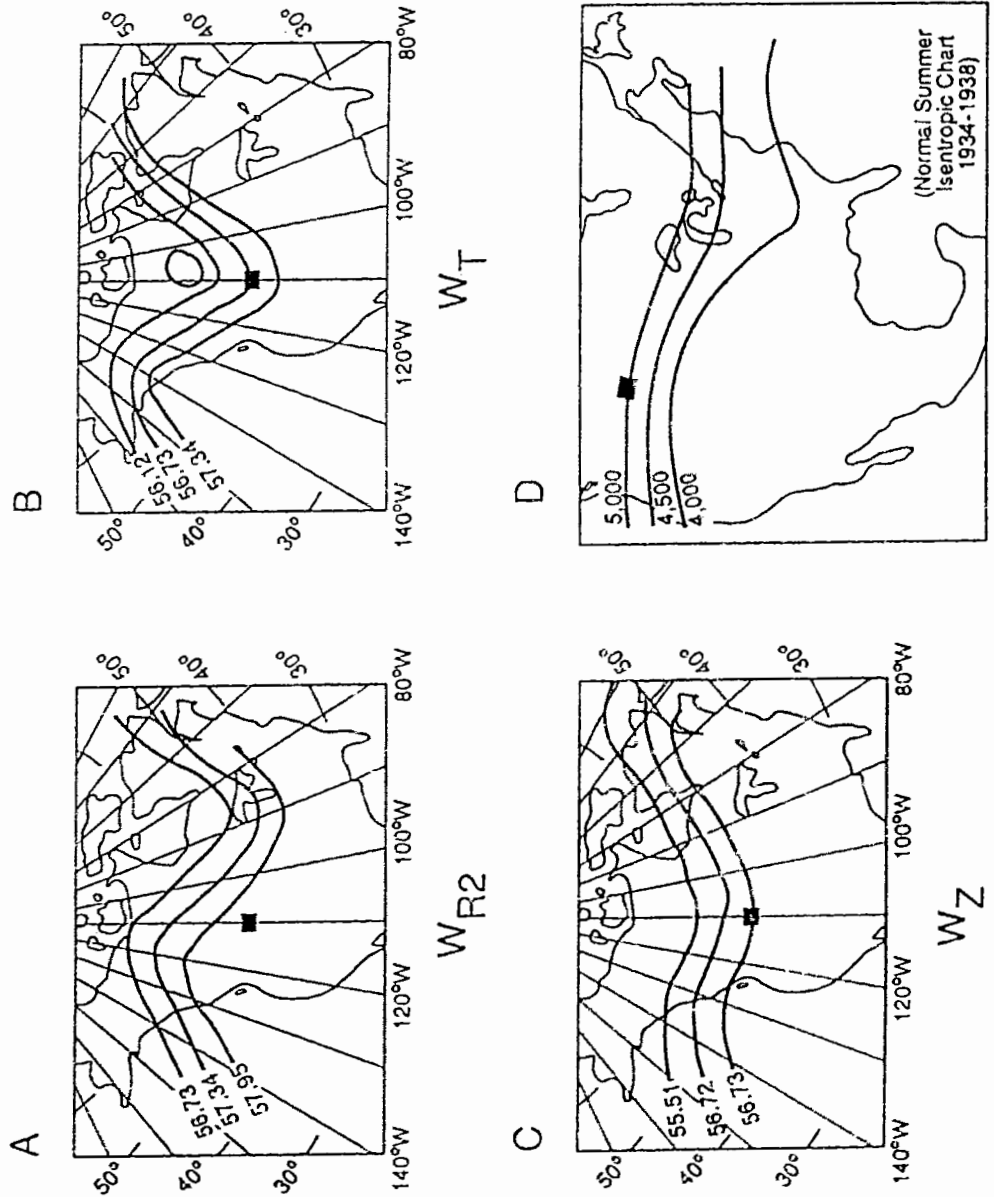
Important early postglacial factors include the Laurentide ice sheet, since in addition to splitting westerly flow, the ice sheet maintained a near-permanent high pressure cell over the main ice mass (Bryson and Wendland 1967, Kutzbach and Geutter 1986, Lamb 1977). This would produce a vigorous southeasterly surface flow over the study area and effectively inhibit movement of moist Pacific air into southern Alberta. An additional factor is the reduced temperature difference between polar and equatorial regions resulting from enhanced insolation in northern latitudes. A low thermal gradient would likely have reduced westerly strength and promoted meridional circulation, as a reduced temperature gradient is thought to do today (Lamb 1972). In combination with the more or less fixed ridging of the high altitude westerlies over the orographic barrier of the Rocky Mountains (*ibid.*), frequent high pressure ridges over northwestern North America likely were a prominent feature of early postglacial climate (Vance 1987). Blocking high pressure cells in the area today produce intense summer drought and clear, cold spells of winter weather, since they inhibit westward movement of mild and moist Pacific air (Dey 1982). Dynamic postglacial climatic conditions likely prevailed however, as recent decades and centuries characterized by frequent high pressure blocking are known have to produced periods of extreme weather events in any one region, due to periodic latitudinal displacement of the blocks (Lamb 1979).

In sum, the frequent appearance of blocking high pressure cells combined with increased summer temperature likely accounts for the extreme drought events that occurred in the Chappice Lake area prior to 6000 BP (Fig 31a). The occurrence of relatively fresh, high water stands may be attributed to periodic displacement of high pressure blocking cells or periodic breakdown of meridional flow, allowing relatively moist Pacific air to move into southern Alberta (Fig. 31c). This hypothesized circulation pattern would have impacted adjacent regions. Thus, traces of its existence should be found in the regional paleoecological record.

Wright (1983), when summarizing the sparse Holocene record of climatic change for the northern Great Plains, noted that it would be possible to decipher precipitation and temperature changes in the area when paleoecological records from high altitude treeline (controlled mainly by summer temperature) and low elevation treeline (controlled mainly by precipitation) were available. Both high and low elevation treeline sites now exist to the west of Chappice Lake (in the path of westerly airflow) and offer a valuable supplement to the Chappice Lake record, particularly since events at these sites are dated by AMS radiocarbon techniques.

The early Holocene record from Opabin Lake (Chapter 3) which currently lies above the upper limit of tree growth, documents upward treeline movement beginning as early as 8500 BP (Reasoner and Hickman 1989), indicating that summer temperature was warmer than present. At 9300 BP, establishment of xeric grassland combined with declining but fluctuating lake levels in central Montana (Barnosky 1989), indicates that a reduced but variable precipitation regime was also a component of early Holocene climate. In the forest-grassland transition of southwestern Alberta, increased aridity is evident by 8000 BP, marking the onset of a prolonged period of low but fluctuating water levels (MacDonald 1989). This is preceded by an interval of increased moisture (*ibid.*), an event recorded at other sites in the southern Canadian Rocky Mountains (Hills et al. 1985, Reeves and Dormaar 1972). As outbreaks of Arctic air during early Holocene winters would have first occurred during retreat of

Figure 31. Synoptic weather types indicating 500-millibar constant pressure level: A) meridional flow with ridge over study area (in black) producing drought; B) meridional flow with ridge over coastal British Columbia resulting in heavy rainfall in the study area, and; C) zonal flow with weak ridge over British Columbia producing moderate rainfall on the northern interior plains. For comparison; D) portrays upper winds prevailing during the zonal circulation of the 1930's (adapted from Dey 1982, Namais 1983).



Laurentide ice (Bryson and Wendland 1967), particularly within a meridional westerly flow pattern, enhanced winter precipitation may also have been an important aspect of early Holocene climate. The seasonal water level fluctuations noted in the early stages of Chappice Lake history may be related to these seasonal temperature extremes.

Although both severe droughts and moist intervals prior to 6000 BP at Chappice Lake are best explained by an extended period dominated by meridional westerly flow, the boundary conditions required to maintain this pattern (that is, the Laurentide ice sheet and seasonal insolation extremes) dissipated during the mid-Holocene. Succeeding westerly flow patterns are more difficult to reconstruct, but traces of subsequent regimes nevertheless appear in the regional paleoecological record. The rapid growth of this data base over the last twenty years has produced a network of proxy 'observatories' in central North America in a density sufficient for a first approximation of long-term tendencies in recent westerly behaviour.

Vigorous westerly flow has been considered the cause of eastward expansion of prairie vegetation in the American midwest, between 8000 and 5000 BP (Bryson and Wendland 1967, Wright 1983), marking the peak of postglacial warmth in the region. In contrast, maximum postglacial warmth was achieved in western North America between 10,000 and 8000 BP (Barnosky 1987b, Mathewes 1985). The delay in the onset of postglacial aridity in the American midwest, compared to western North America, may be explained by a shift from a predominantly meridional pattern to one typified by vigorous zonal flow. Early postglacial high pressure ridges in western North America would have channelled Arctic air to the American midwest (Fig. 31a), where interaction with moist tropical air would have resulted in abundant precipitation, accounting for the consistently high lake levels characteristic of the American midwest at that time (Harrison and Metcalfe 1985). A shift to zonal flow would have replaced Arctic and tropical air masses with a virtually continuous flow of warm and dry Pacific air over the American midwest (Fig. 31d), producing an arid climatic regime and

promoting the observed eastward extension of the prairie border in the American midwest.

It is unknown when the transition from meridional to zonal westerly circulation occurred, since it is not clear when the Laurentide ice sheet had ablated to the point that it no longer influenced westerly behavior. Dune formation in eastern Canada, associated with the glacial anticyclone, ceased at 7500 BP (Filion 1987). This suggests that the influence of the Laurentide ice sheet on atmospheric circulation was reduced by 7500 BP, although it is likely that the transition between circulation patterns was a stepwise shift in dominance from meridional to zonal flow, as opposed to an abrupt transition.

If meridional circulation was indeed becoming less prevalent by 7500 BP, and since seasonal radiation balance was like today's by 6000 BP (Kutzbach and Geutter 1986), evidence of a climatic transition in western North America should appear during the mid-Holocene. At Chappice Lake, 6000 BP marks the termination of playa conditions and the development of a perennial lake, indicating that the regional water table had begun to rise from an early Holocene low. Evidently the early period of climatic extremes had passed. To the south, at Lost Lake, increasing moisture is also apparent at 6000 BP, as lake level rose and low elevation treeline was depressed (Barnosky 1989). On Alberta's eastern slopes, lake levels rose and modern lower treeline elevation was attained by 5500 BP (MacDonald 1989). Temperatures were also declining from the early Holocene maximum, as suggested by reduced upper treeline at Opabin Lake, beginning about 7000 BP (Reasoner and Hickman 1989). Reduced temperature and increased precipitation, compared to the early Holocene, were evidently part of a regional mid-Holocene climatic change in the western interior of North America. In spite of these changes, a well-developed arid zone had formed across the continent (ca. 32°–57°N, Harrison and Metcalfe 1985), with the exception of the far west (Barnosky et al. 1987a, Mathewes 1985). Although early Holocene climatic extremes had passed, conditions evidently remained more arid than present.

In addition to a mid-Holocene increase in zonal flow, it is likely that the core of westerly flow shifted slightly northward compared to present, as grassland extended north of its current limit in the Canadian interior throughout the mid-Holocene (Ritchie 1987). A wedge of warm, dry Pacific air, fuelled by increasing westerly intensity, may have been responsible for altered vegetation boundaries across the central interior of North America and may also have contributed to the mid-Holocene dessication of Lake Manitoba (Teller and Last 1981). Strong zonal flow of the mid-Holocene likely varied in strength however, as paleohydrological evidence from the American midwest suggests periodicity of drought events. Watts and Winter (1966) document four 200-year long drought periods between 7000 and 5000 BP in southern Minnesota, while Dean et al. (1984) provide evidence of a high water stand between 5400 and 4800 BP, interrupting an extended drought interval between 8500 and 3800 BP in northwestern Minnesota.

From 6000 to 4400 BP the Chappice Lake sedimentary record contains no evidence of periodic relief from drought, but rather reflects a prolonged drought interval (that promoted two near-dessication events). High salinity conditions were maintained as lake level rose gradually. This suggests that inflow of water was just sufficient to balance evaporative losses over an extended period. Although temperature had declined from the early Holocene, particularly in summer, it remained higher than present (Reasoner and Hickman 1989, Vance 1986). Precipitation had increased from early Holocene lows, but was still likely lower than present (Barnosky 1989, Reasoner and Hickman 1989, Vance 1986). As droughts in the 1920's and 1930's occurred within a similar type of zonal flow regime (Lamb 1972, 1982, Borchert 1950, 1970), the period between 6000 and 4400 BP may best be viewed as a climatic interval where weather similar to the 1930's was a near-continuous feature. High winds might also have been prevalent during this period, as they were in the 1930's (Borchert 1950, 1970). Frequent chinook winds may also have been a product of a northward shift in westerly flow (Bryson 1980). Although the extreme events typical of the early Holocene no longer occurred, the climate of this period remained warmer and drier than present

throughout much of the year.

From 4400 to 2650 BP, lake levels continued to slowly rise at Chappice Lake, but still remained lower than present. A shift in the aquatic flora (from *Ruppia* to *Chara*) suggests declining salinity combined with the dominance of dissolved bicarbonate, as opposed to today where sulfate is the dominant anion. As surficial groundwater in the Chappice Lake area is dominated by calcium bicarbonate (Stevenson and Borneuf 1977), it is possible that this change in lake chemistry reflects recharge of local surficial aquifers resulting from increased water supply, due to decreased drought severity and frequency. Alternatively, carbonates may have washed into the basin as increased surface runoff mobilized salt crusts surrounding the lake. Sauchyn (1990) suggests that a condition of persistent groundwater recharge prevailed on the Cypress Hills between 4100 and 3600 BP, perhaps renewing subsurface flow along preglacial valleys radiating from the upland and ultimately contributing to the overall rise in Chappice Lake. Although details surrounding lake chemistry changes remain unclear, increased precipitation compared to the pre-4400 BP period is indicated, accompanied by a continued decline in yearly average temperature that began prior to 6000 BP.

Strong mid-Holocene zonal flow was evidently waning and the core of the westerly winds was shifting southward by 4000 BP, as the prairie border in the American midwest began a westward retreat (Dean et al 1984, Baker et al. 1990), lake levels rose in south-central Wisconsin (Winkler et al. 1986), and grassland receded southward in western Canada (Ritchie 1987). Although temperatures were likely still slightly above present (Reasoner and Hickman 1989, Vance 1986), precipitation was increasing, as reflected by rising water levels and freshening of Lost Lake (Barnosky 1989). Lakes in central Alberta were also rising to current levels (Schweger and Hickman 1989), and basins in arid south central Saskatchewan were filling (Last and Schweyen 1983). After 4400 BP at Chappice Lake, lake levels continued to rise and water was becoming less saline.

The transition from mid-Holocene to late Holocene climate is recorded in a variety of proxy records from both Europe and North America (Lamb 1972). Like the onset of early Holocene aridity, initiation of the cool, moist Neoglacial period was time-transgressive on a continent-wide basis (Anderson et al. 1989, Lamb 1972). In central Alberta, climatic deterioration from mid-Holocene aridity occurred between 6000 and 3000 BP (Schweger and Hickman 1989), with modern conditions developing about 3000 BP (Vance et al. 1983).

Bryson et al. (1970) suggested that meridional circulation best explains declining aridity in western North America at 3500 BP, and observed from modern data that a persistent summer high pressure ridge over British Columbia could explain precipitation increases on the Great Plains (Fig. 31b). Harrison and Metcalfe (1985) also invoke west coast ridging and increasing meridional circulation at 3000 BP to explain continent-wide high lake levels. A persistent west coast ridge channels cool and moist air into the study area today. In the last 50 years, this type of circulation pattern typically results in high summer precipitation on the Canadian prairies (Dey 1982). This illustrates how the location of blocking high pressure cells influences the character and occurrence of extreme weather that one particular region will experience during periods of meridional circulation, since location of the blocking high 20°W would have resulted in drought on the northern Great Plains (Fig. 31a and b). Factors influencing the position of high pressure blocking cells are poorly understood, although ocean temperatures are considered to be important (Namais 1983). Whatever the causes may be, recurrence of this type of westerly atmospheric circulation adequately accounts for the late Holocene rise in lake levels at Chappice Lake.

Most paleoecological study sites in the western interior of North America display little response to climatic change over the last three millenia (Barnosky 1989, Schweger and Hickman 1989, Vance et al. 1983). Present day air mass configuration is considered to have been established by 3000 BP (Harrison and Metcalfe 1985). Indeed, compared to earlier events, changes in the late Holocene were minor. As vigorous westerly flow characteristic of

the mid-Holocene came to a close, subsequent changes in upper westerly flow patterns appear to have been short-lived by comparison. However, since analysis of historic climatic records reveals that even short-lived perturbations are capable of producing significant changes in weather (Lamb 1982), late Holocene fluctuations likely produced significant changes in regional climate. Unfortunately, the short-term nature of these recent climatic events and the relatively long response time of vegetation limits the resolution of most paleobotanical records. Since Chappice Lake is located within a drought-prone region that is particularly sensitive to climatic fluctuations, it offers some insight into the pace and intensity of recent events.

Rising lake levels at Chappice Lake beginning as early as 4400 BP, culminated between 2650 to 1060 BP when the lake experienced the longest high, relatively freshwater stand of the entire record. Droughts on the northern mixed-grass prairie were infrequent throughout this period. This prolonged highwater stand may result from a continuation of the blocking pattern described above, aided by declining temperatures associated with Neoglaciation (Leonard 1986). Due to a paucity of proxy records from climatically-sensitive sites however, reconstruction of westerly configuration is not possible. To the east of Chappice Lake, an extended low water stand occurred at Waldsea Lake between 2800 to 2200 BP (Last and Schweyen 1985). This is a puzzling conflict with the Chappice Lake record, suggesting that either central Saskatchewan was beyond the influence of a high pressure cell to the west, or perhaps more probably that the conventional ^{14}C determinations on bulk sediment at Waldsea Lake are in error by some 1000 years.

At 1060 BP, the extended period of cool, moist conditions in southern Alberta ended abruptly. Low levels once again prevailed at Chappice Lake, in this case recurring periodically over the next four centuries. Conversion to calendar years of the single radiocarbon date marking the onset of this period of increased drought frequency produces a date of AD 775 to 1200 (Klein et al. 1982). These low water stands at Chappice Lake therefore fall within the Medieval Warm Period, an interval during which melting of the Arctic ice pack allowed

establishment of the first Norse colonies in Greenland (Gribbin and Lamb 1978). In the American midwest, Medieval warmth is recorded by grassland expansion that is suggested to have been the result of a 25% decline in summer precipitation as part of a 200 year-long drought, beginning in AD 1200 (Bryson et al. 1970, Bryson 1980). Increased zonal flow, displaced south of its current position, has been used to explain this drought (ibid.). This circulation pattern is similar to that outlined for the mid-Holocene, although intense westerlies were not maintained for as long nor had the core of westerly flow shifted north, as it evidently had during the mid-Holocene. This suggests that Medieval warmth should not have produced as severe a drought period as during the mid-Holocene in the Chappice Lake area, since mid-Holocene northward movement of the westerly core would have resulted in increased chinook frequency in southern Alberta (Bryson 1980), thus increasing annual temperature. Evidently, circulation changes were sufficient to affect only the most arid portions of the southwestern Canadian interior. Data from Chappice Lake further suggest that zonal flow was occasionally interrupted, as low water stands were interspersed with brief high water stands throughout the Medieval Warm Period.

Reduced water levels at 1080 BP at Deadmoose Lake, central Saskatchewan (an event dated by a conventional ^{14}C determination of a wood fragment; Last and Slezak 1986) suggest that the effects of Medieval warmth extended to this portion of the Canadian prairies. Nearby Waldsea Lake also bears evidence of a reduced water level at this time (ibid.). Few other paleoecological studies contain evidence of this event. One exception is high altitude Opabin Lake, where increased water turbidity at approximately 1000 BP (Reasoner and Hickman 1989) may have been a product of increased winds. Interpolation from the radiocarbon date marking the onset of the Medieval Warm Period at Chappice Lake suggests the effects of this climatic interval were felt for some 400 years in southeastern Alberta, producing a series of lengthy droughts.

Since the Medieval Warm Period, Chappice Lake has been maintained at generally high levels punctuated by periodic, short-term low lake stands. Using the occurrence of *Salsola kali* as a marker of the historic period, and interpolation from the uppermost ^{14}C determination to date the onset of this most recent high water stand, this recent phase is estimated to span the period from AD 1400 to 1880. This corresponds closely to the dates bracketing the Little Ice Age.

Gribbin and Lamb (1978), after consulting an array of historic data, conclude westerly circulation during the Little Ice Age (AD 1430 to 1850) was predominantly meridional and marked by an abnormal number of blocking events. Bryson (1980) outlines a similar scenario for North America, suggesting that exaggerated north/south meanders in westerly flow produced comparatively heavy rainfall in the interior of the continent. As at the onset of the Neoglacial period, high pressure blocking west of the study area is implicated (Fig. 31b). Data from Chappice Lake indicate that during the Little Ice Age the area was infrequently beset with drought events of short duration. This is consistent with historic meteorological observations of periods typified by frequent blocking, as changing position of the westerly waves can produce opposite weather extremes at any given location (Lamb 1979).

Direct observation of air mass behaviour reveals that meridional circulation during the Little Ice Age was followed by a period of intensified zonal westerly flow from 1900 through 1935 (Dzerdzeevskii 1969, Lamb 1981), marking the return to a circulation pattern similar to that outlined for the Medieval Warm Period. Drought over much of the northern mixed-grass prairie during the 1920's and 1930's occurred as a result (Borchert 1950, 1970), and these events reduced water levels at Chappice Lake. Compared to the first half of the century, weaker westerly flow with more frequent blocking events has prevailed (Gribbin and Lamb 1978). During the 1980's drought again returned to the study area, caused by persistent high pressure blocks over the interior of North America (Trenberth 1989). The aridity of the late 1980's, combined with modifications restricting surface flow to Chappice Lake, has

reduced lake levels to those approximating the late 1800's, 1920's, and 1930's. This suggests that drought in the 1980's was not as severe as the aridity of the late 1800's, 1920's, or 1930's.

7.3 Implications for Archaeology

Although the archaeological data base for southern Alberta has expanded over the last twenty years there remains considerable gaps (Vickers 1986), and any attempt to assess the impact of climatic change on prehistoric populations is premature. In other words, the archaeological record is not amenable to direct observation of population responses to increased drought frequency. The paleoclimatic record of the area does however, make possible a number of predictions concerning the conditions early inhabitants of the area likely encountered.

With the exception of the Fletcher Site (Vickers and Beaudoin 1989), no archaeological sites dating to the early Holocene have been discovered, although scattered surface finds of projectile points indicative of early occupation suggest habitation has been continuous since approximately 11,000 BP (Brumley and Dau 1988). Long-standing debate regarding the impact of postglacial warmth and aridity on human occupation of the northern plains has centered on whether the paucity of sites in the region reflects depopulation or simply difficulty in locating deeply buried sites (Reeves 1973). The Chappice Lake record suggests that the early postglacial period (prior to 6000 BP) presented a challenging environmental setting, with dynamic climatic conditions producing extremes in the availability of natural resources. Periods of intense drought would have created chronic resource shortages, but these relatively short-term events would have been balanced by intervals of abundance. To nomadic peoples used to dealing with periodic shortages in a semi-arid environment, a fluctuating resource base would not likely have created insurmountable problems, but presumably maintained low population densities.

From 6000 BP to 4000 BP, conditions on the northern interior plains changed considerably from the early Holocene. A period of continual resource shortage, maintained by lower-than-present precipitation, high winds and increased temperature compared to today, would have created a very difficult environment for nomadic hunters and gatherers. Few archaeological sites in southern Alberta are known from this period (Vickers 1986). Whether this actually reflects a population decline from the preceding period or simply inadequate site recovery is not clear, but a population decline is one prediction stemming from this study.

During the late Holocene, conditions for human habitation of southern Alberta should have improved considerably, compared to earlier periods. A cooler, moister climatic regime likely established a dependable natural resource base in what had been, until this time, a rather sparse and undependable setting. The earliest archaeological sites on the southern Alberta prairie date to 4200 BP (Brumley and Dau 1988). Oxbow is the dominant projectile point style of the time. Although not conclusive, the climatic reconstruction from Chappice Lake suggests that favourable environmental circumstances may have aided the establishment of populations utilizing these projectile points.

Archaeological remains from subsequent times indicate continuous occupation of the area to the historic period. One distinctive recent projectile point style, Avonlea, is found in sites dating between 1750 and 1150 BP (Vickers 1986). Although not as numerous as sites from preceding or later periods, the sites excavated to date are extensive and have been interpreted as indicating large aggregations that did not disperse frequently (Brumley and Dau 1988). While social and technological developments likely played a part in this distinctive settlement pattern, sites bearing Avonlea points date to an interval of infrequent drought, a period when plant and animal resources would have been more abundant per unit area and dependable than at any other time during the Holocene. The decline of Avonlea corresponds to increasing drought frequency at the onset of the Medieval Warm Period. Thus, it is possible that changing environmental conditions played a role in the expansion and decline of the

aboriginals producing this style of projectile point.

7.4 Implications for Future Climatic Change

The Chappice Lake record suggests that droughts on the northern Great Plains are not cyclical. Rather, there are intervals of frequent, intense drought in contrast to extended periods during which droughts are rare. The record also indicates that historic droughts that have been sufficient to cause economic hardship and depopulation of the region are minor compared to several past drought periods. This conclusion does little to ease concerns related to impending drought, as predicted by computer modelling of atmospheric changes caused by increased concentration of 'greenhouse gases' (Hare 1988).

Throughout much of the early and mid-Holocene, extended drought intervals were the norm rather than the exception. It is likely that a return of the circulation pattern that produced these prolonged drought intervals would result in the elimination of agricultural production throughout much of the northern Great Plains. A critical question arises as to whether these types of circulation patterns are likely to recur, and if so, whether they will persist.

The return of a persistent early to mid-Holocene circulation pattern is unlikely, as extensive continental ice sheets and changes to seasonal insolation, boundary conditions that likely produced persistent high pressure blocks over western North America, no longer exist. In this sense, the reconstruction of this period is not a suitable analogue for the future. The return of increased westerly vigor and zonal circulation patterns during the Medieval Warm Period and the early portion of the present century, suggest this circulation pattern may recur. However, the heightened temperature difference between polar and equatorial regions required to fuel westerly intensity (considered a necessity to maintain this type of circulation pattern) is not part of projected global temperature changes. Predictions based on projected

increases in 'greenhouse gas' concentrations suggest that the temperature difference between polar and equatorial regions will be reduced, due to high latitude warming (Hare 1988). If high latitude warming does in fact transpire, a decrease in westerly intensity could result, producing a continuation of meridional flow established in the 1950's (Gribbin and Lamb 1978). In effect, CO₂-induced warming may produce a circulation regime akin to that outlined for the early Holocene. Variable climatic extremes would result, as they have in the recent past when meridional circulation has persisted (Lamb 1982). The impact of increased meridional circulation patterns on the northern Great Plains would then be dependent upon the location of high pressure blocking cells. As outlined above, blocking over the interior of North America typically produces intense seasonal drought, whereas latitudinal displacement of blocking cells may promote precipitation increases in the interior of the continent. In this regard, further research into factors determining location of blocking high pressure cells may increase the precision of long-term forecasts of climate for the northern interior plains.

The northern Great Plains was the area chosen to conduct investigation into past hydrological and climatic variation because it is a region particularly sensitive to climate. The Chappice Lake record supports this contention, as it bears evidence of recent, relatively minor climatic fluctuations. In terms of monitoring climate for evidence of current change, particularly concerning the possible impact of increased concentration of 'greenhouse gases', the area is not particularly well-suited, as extreme weather appears to be a part of the region's climate. Better suited areas for assessing the influence of 'greenhouse gases' are those situated in more stable climatic settings, where variations may be more readily linked to anthropogenic causes.

CHAPTER 8

CONCLUSION

Saline lakes of the northern Great Plains respond rapidly to climatic change and their sediments contain a detailed record of paleohydrological change. Saline lakes are also repositories of detailed information on past climatic change, although their dependence on groundwater inflow means that the climatic signal may be buffered. With judicious site selection however, a detailed proxy record of climatic change was developed using paleobotanical and sedimentary indicators.

Study of modern vegetation, as well as pollen and plant macrofossil distribution in a hypersaline lake in southeastern Alberta, indicates that plant macrofossils are abundant and are a clear reflection of current local vegetation. Plant macrofossils are therefore good indicators of past hydrological change, clearly tracking past shoreline location. Modern pollen distribution does not as clearly reflect current local vegetation. Although *Ruppia* pollen is a valuable indicator of water depth, salinity, and distance to shoreline, few other pollen taxa convey unequivocal information on shoreline movements associated with hydrological change.

Analyses of pollen, plant macrofossils, and sediments obtained from five cores raised from Chappice Lake, southeastern Alberta, allow the reconstruction of paleohydrological changes over the last eight millennia. Lake levels were generally low but variable, from approximately 8500 BP to 6000 BP. The lake exhibited playa characteristics throughout this period. From 6000 to 4400 BP, high salinity water was maintained during low, but gradually rising, water levels. From 4400 to 2650 BP, the lake continued to rise and freshen. Groundwater flow, reduced to a minimum during the preceding extended low water stand, was likely reactivated during this period. Highest lake levels were attained from 2650 to 1060 BP. This prolonged high stage ended abruptly by 1060 BP, and was followed by four centuries of repeated low water, high salinity stands. From 600 BP to present, relatively high lake levels have been

maintained but punctuated by aperiodic, short-lived low water stands.

Comparison of this paleohydrological reconstruction with regional proxy climate data indicates that water level fluctuations at Chappice Lake accurately reflect major climatic changes and add additional details, particularly concerning the late Holocene. Paleohydrological change is considered a chronicle of long-term drought on the northern Great Plains. Climatic changes are interpreted as representing changing configurations of the high altitude westerly air flow, since persistence of particular patterns has produced drought in the region during the historic period.

It is suggested that a preponderance of blocking high pressure ridges over western North America produced frequent, intense droughts in the study area prior to 6000 BP, although moist periods interrupted drought periodically. From 6000 to 4400 BP, zonal westerly flow predominated, producing an extended though less intense, drought interval. A shift to meridional flow occurred about 3500 BP, with blocking highs situated further west than in the early Holocene, bringing increased precipitation to southern Alberta. Subsequent changes in westerly configuration were less pronounced.

Beginning at 1060 BP, droughts once again increased in frequency. This four-century long interval of periodic drought, correlative with the Medieval Warm Period, was caused by a return to predominantly zonal flow. During the Little Ice Age (AD 1450-1850) a meridional flow, similar to the configuration of 3500 BP, was established. Droughts were once again infrequent. Renewed drought frequency is evident in the late 1800's and early 1900's, when predominantly zonal circulation was established. With certain assumptions, the Chappice Lake sedimentary record bears evidence of these historic drought occurrences. A return meridional flow since 1950 has increased drought frequency once again, particularly during the 1980's.

LIST OF REFERENCES

- Anderson, T.W., Mathewes, R.W., and Schweger, C.E. 1989. Holocene climatic trends in Canada with special reference to the Hypsithermal interval. pp. 520-528, In: **Quaternary Geology of Canada and Greenland**, (R.J. Fulton ed.), Geological Survey of Canada, Ottawa.
- Baker, R.G. 1983. Holocene vegetational history of the western United States. pp. 109-127, In: **Late Quaternary Environments of the United States: Volume 2, The Holocene**. (H.E. Wright Jr., ed.), University of Minnesota Press, Minneapolis.
- Baker, R.G., Chumbey, C.A., Witinok, P.M., and Kim, H.K. 1990. Holocene vegetational changes in eastern Iowa. **Journal of the Iowa Academy of Science** 97(4): 167-177.
- Barendregt, R.W. 1988. The Physical Setting of the Southern Alberta Landscape. Unpublished manuscript.
- Barnosky, C.W. 1989. Postglacial vegetation and climate in the northwestern Great Plains of Montana. **Quaternary Research** 31: 57-73.
- Barnosky, C.W., Anderson, P.M., and Bartlein, P.J. 1987a. The northwestern U.S. during deglaciation; vegetational history and paleoclimatic implications. pp. 289-321, In: **North American and Adjacent Oceans During the Last Deglaciation**. (W.F. Ruddiman and H.E. Wright Jr., eds.), Geological Society of America, The Geology of North America, v. K3., Boulder, Colorado.
- Barnosky, C.W., Grimm, E.C., and Wright, H.E. 1987b. Towards a postglacial history of the northern Great Plains: a review of the paleoecologic problems. **Annals of Carnegie Museum** 56(16): 259-273.
- Barrie, J. Conversation with author, May 1988.
- Beaty, C.B. 1972. Geomorphology, geology, and non-agricultural resources. pp. 11-34, In: **Southern Alberta - A Regional Perspective**. (F. Jankunis, ed.), The University of Lethbridge, Lethbridge.
- Beaudoin, A.B. 1986. Using *Picea/Pinus* ratios from the Wilcox Pass core, Jasper National Park, Alberta, to investigate Holocene timberline fluctuations. **Geographie Physique et Quaternaire** 40(2): 145-152.
- Beijerinck, W. 1976. **Zadenatlas der Nederlandsche Flora: Ten Behoeve van de Botanie, Palaeontologie, Bodemcultuur en Warenkennis**. Backhuys & Meesters, Amsterdam.
- Bengtsson, L. and Enell, M. 1986. Chemical analysis. pp. 423-451, In: **Handbook of Holocene Palaeoecology and Palaeohydrology**. (B.D. Berglund, ed.), John Wiley & Sons Ltd., Chichester.
- Beukens, P.R. Letter to author, March 1990.
- Blinn, D.W. 1971. Dynamics of major ions in some permanent and semi-permanent saline systems. **Hydrobiologia** 38(2): 225-238.

- Borchert, J.R. 1950. The climate of the central North American Grassland. *Annals of the Association of American Geographers* 40(1): 1-34.
- Borchert, J.R. 1970. The dust bowl in the 1970s. *Annals of the Association of American Geographers* 61(1): 1-22.
- Borneuf, D.M. 1971. Hydrogeological map—Foremost, Alberta. NTS 72E. Alberta Research Council, Edmonton.
- Borneuf, D.M. 1983. **Springs of Alberta. Earth Sciences Report 82.3.** Alberta Research Council, Natural Resources Division, Edmonton.
- Borneuf, D.M. and Stevenson, D.R. 1970. Hydrogeological map—Medicine Hat, Alberta. NTS 72L. Alberta Research Council, Edmonton.
- Brock, M.A. 1981. The ecology of halophytes in the south-east of South Australia. *Hydrobiologia* 81: 23-32.
- Brown, T.A., Nelson, D.E., Mathewes, R.W., Vogel, J.S., Southon, J.R. 1989. Radiocarbon dating of pollen by accelerator mass spectrometry. *Quaternary Research* 32: 205-212.
- Brumley, J.H. and Dau, B.J. 1988. Historical resource investigations within the forty mile coulee reservoir. **Archaeological Survey of Alberta Manuscript Series No. 13.**, Alberta Environment, Alberta Culture and Multiculturalism, Edmonton.
- Bryson, R.A. 1980. Ancient climes on the Great Plains. *Natural History*. June, pp. 65-73.
- Bryson, R.A. and Wendland, W.M. 1967. Tentative climatic patterns for some late glacial and post-glacial episodes in central North America. pp. 271-298. In: **Life, Land and Water, Lake Agassiz Region.** (W.J. Mayer-Oakes, ed.), Occasional Paper No. 1, Department of Anthropology, University of Manitoba, Winnipeg.
- Bryson, R.A., Baeris, D.A., and Wendland, W.M. 1970. The character of late-glacial and post-glacial climatic changes. pp. 53-73. In: **Pleistocene and Recent Environments of the Central Great Plains.** (W. Dort, Jr. and J.K. Jones, Jr., eds.), The University Press of Kansas, Lawrence.
- Burne, R.V., Bauld, J., and De Dekker, P. 1980. Saline lake charophytes and their geological significance. *Journal of Sedimentary Petrology* 50(1): 281-293.
- Catto, N.R. 1980. **Quaternary geology of the western Cypress Hills region, Alberta and Saskatchewan.** Unpublished MSc. Thesis, The University of Alberta, Edmonton, Alberta.
- Carlson, V.A. 1970. Bedrock topography of the Medicine Hat, Alberta. NTS 72L. Alberta Research Council, Edmonton.
- Chakravarti, A.K. 1970. Precipitation deficiency patterns in the Canadian prairies, 1921 to 1970. *Prairie Forum* 1: 95-110.
- Christiansen, E.A. 1979. The Wisconsinan deglaciation of southern Saskatchewan and adjacent areas. *Canadian Journal of Earth Sciences* 16: 913-937.

- Clayton, J.S., Ehrlich, W.A., Cann, D.B., Day, J.H., and Marshall, I.B. 1977. **Soils of Canada**. Vol. 1. Research Branch, Canada Department of Agriculture, Government of Canada, Ottawa.
- Clayton, L. and Moran, S.R. 1982. Chronology of Late Wisconsinan glaciation in Middle North America. **Quaternary Science Reviews** 1(1): 55-82.
- Cole, I.H. 1926. Sodium sulfate of western Canada. Occurrence, uses and technology. **Canadian Department of Mines Publication 646**. 155p.
- Coupland, R.T. 1950. Ecology of mixed prairie in Canada. **Ecological Monographs** 20(4): 272-315.
- Coupland, R.T. 1961. A reconsideration of grassland classification in the northern Great Plains of North America. **Journal of Ecology** 49: 135-167.
- Cwynar, Les C., Burden, E., and McAndrews, J.H. 1979. An inexpensive sieving method for concentrating pollen and spores from fine-grained sediments. **Canadian Journal of Earth Sciences** 16(5): 1116-1120.
- Dean, W.E., Anderson, R.Y., and Barnosky, C.W. 1984. The variability of Holocene climate change: evidence from varved lake sediments. **Science** 226: 1191-1194.
- Dey, B. 1982. Nature and possible causes of droughts on the Canadian prairies - case studies. **Journal of Climatology** 2: 233-249.
- Digerfeldt, G. 1986. Studies on past lake-level fluctuations. pp. 127-143, In: **Handbook of Holocene Palaeoecology and Palaeohydrology**. (B.D. Berglund, ed.), John Wiley & Sons Ltd., Chichester.
- Digerfeldt, G. 1988. Reconstruction and regional correlation of Holocene lake-level fluctuations in Lake Bysjon, south Sweden. **Boreas** 17: 165-182.
- Dodd, J.D. and Coupland, R.T. 1966. Vegetation of saline areas in Saskatchewan. **Ecology** 47(6): 958-968
- Dodd, J.D., Rennie, D.A., and Coupland, R.T. 1964. The nature and distribution of salts in uncultivated saline soils in Saskatchewan. **Canadian Journal of Soil Science** 44(2): 165-175.
- Dyke, A.S. and Prest, V.K. 1987. Late Wisconsin and Holocene history of the Laurentide ice sheet. **Geographie physique et Quaternaire** 41: 237-263.
- Dziedziewskii, B.L. 1969. Climatic epochs in the Twentieth Century and Some Comments on the Analysis of Past Climates. pp. 49-60, In: **Quaternary Geology and Climate**. (H.E. Wright, Jr., ed.), National Academy of Sciences, Washington, D.C.
- Eugster, H.P. and Hardie, L.A. 1978. Saline lakes. pp. 237-293, In: **Lakes Chemistry Geology Physics**. (A. Lerman, ed.), Springer-Verlag, New York.
- Fægri, K. and Iversen, J. 1975. **Textbook of Pollen Analysis**. 3rd edition. Blackwell Scientific Publications, Oxford.

- Farrow, M. 1989. Saskatchewan hopes, dreams hinge on spring rains. *The Vancouver Sun*, 30 March, pp. B1-B2.
- Farvolden, R.N. 1960. Bedrock channels of southern Alberta. pp. 63-74. In: *Early Contributions to the Groundwater Hydrology of Alberta*. (R.N. Farvolden, W.A. Meneley, E.G. LeBreton, D.H. Lennox, and P. Meyboom eds.), Research Council of Alberta, Bulletin 12, Edmonton.
- Filion, L. 1987. Holocene development of dunes in the central St. Lawrence lowland, Quebec. *Quaternary Research* 28: 196-209.
- Forbes, J.R. and Hickman, M. 1981. Paleolimnology of two shallow lakes in central Alberta, Canada. *Internationale Revue der gesamten Hydrobiologie* 66(5): 1-26.
- Fraser, H.M. 1980. Economic and social effects of the 1980 drought on the prairies. pp. 23-29. In: *Canadian Climate in Review, 1980*. Government of Canada, Ottawa.
- Freeze, R.A. 1969a. *Hydrogeology and Groundwater Resources of the Gravelbourg Aquifer, Saskatchewan*. Scientific Series No. 6. Inland Waters Branch, Department of Energy, Mines, and Resources, Government of Canada. Queen's Printer, Ottawa.
- Freeze, R.A. 1969b. *Regional Groundwater Flow - Old Wives Lake Drainage Basin, Saskatchewan*. Scientific Series No. 5. Inland Waters Branch, Department of Energy, Mines and Resources, Government of Canada, Queen's Printer, Ottawa.
- Fritz, P. and Krouse, H.R. 1973. Wabamun Lake, past and present, an isotopic study of the water budget. pp. 244-258. In: *Proceedings of the Symposium on Lakes of Western Canada*. (E.R. Reinelt, A.H. Laycock, and W.M. Schultz, eds.), University of Alberta Press, Edmonton.
- Futyma, R.P. and Meacham, C.A. 1986. *Michigrana: The Michigan Graphing Program for Pollen Analysis*. Department of Botany, University of Michigan, Ann Arbor.
- Gaillard, M.J. 1985. Postglacial palaeoclimatic changes in Scandinavia and central Europe. A tentative correlation based on studies of lake level fluctuations. *Ecologia Mediterranea* 11(1): 159-175.
- Godwin, R.B. 1988. Drought, the impending crisis? pp. 11-15. In: *Proceedings of the Prairie Drought Workshop*. Saskatoon, October 1988. (D.J. Bauer, ed.). Environment Canada, Government of Canada, Ottawa.
- Gorham, E., Dean, W.E., and Sanger, J.E. 1983. The chemical composition of lakes in the north-central United States. *Limnology and Oceanography* 28(2): 287-301.
- Gove, H.E., Elmore, D., Ferraro, R.D., Beukens, R.P., Chang, K.H., Kilius, I.R., Lee, H.W., Litherland, A.E., Purser, K.H., and Rubin, M. 1980. Radiocarbon dating with tandem electrostatic accelerators. *Radiocarbon* 22: 785-793.
- Gribbin, J. and Lamb, H.H. 1978. Climatic change in historical times. pp. 68-82. In: *Climatic Change*. (J. Gribbin and H.H. Lamb, eds.) University Press, Cambridge.

- Grossman, I.G. 1968. Origin of sodium sulfate deposits of the northern Great Plains of Canada and the United States. pp. B104-B109. U.S. Geological Survey Professional Paper 600-B.
- Hammer, U.T. 1990. Letter to author, 11 April 1990.
- Hammer, U.T. 1978. The saline lakes of Saskatchewan III. Chemical characterization. *Internationale Revue der gesamten Hydrobiologie* 63: 311-335.
- Hammer, U.T., Haynes, R.C., Heseltine, J.M., and Swanson, S.M. 1975. The saline lakes of Saskatchewan. *Verhandlungen der Internationalen Vereinigung für theoretische und angewandte Limnologie* 19: 589-598.
- Hammer, U.T. and Haynes R.C. 1978. The saline lakes of Saskatchewan II. Locale, hydrography and other physical aspects. *International Revue der gesamten Hydrobiologie und Hydrographie* 63(2): 179-203.
- Hammer, U.T. and Heseltine, J.M. 1988. Aquatic macrophytes in saline lakes of the Canadian prairies. *Hydrobiologia* 158: 101-116.
- Hammer, U.T., Shames, J., and Haynes, R.C. 1983. The distribution and abundance of algae in saline lakes of Saskatchewan, Canada. *Hydrobiologia* 105: 1-26.
- Hare, F.K. 1988. Vulnerability to climate. pp. xvii-xxv. In: *The Impact of Climatic Variability and Change on the Canadian Prairies*. (B.L. Magill and F. Geddes, eds.), Symposium/Workshop Proceedings, Prepared by Alberta Department of Environment, Edmonton.
- Hare, F.K. and Thomas, M.K. 1979. *Climate Canada*. 2nd Edition. John Wiley & Sons Canada Limited, Toronto.
- Harrison, S.P. 1988. Lake-level records from Canada and the Eastern USA. *Lundqua Report 29*. Lund University, Department of Quaternary Geology, Tornavägen 13, Lund, Sweden.
- Harrison, S.P. and Metcalfe, S.E. 1985. Variations in lake levels during the Holocene in North America: an indicator of changes in atmospheric circulation patterns. *Geographie Physique et Quaternaire* 34(2): 141-150.
- Haste, R.A. 1908. A warm chinook in sunny Alberta. p. 33 in Jones, D.C. 1987. *Empire of Dust*. The University of Alberta Press, Edmonton. Originally appeared in *Canada West Magazine*. March 19, 1908
- Haynes, R.C. and Hammer, U.T. 1978. The saline lakes of Saskatchewan. IV. Primary production by phytoplankton in selected saline ecosystems. *Internationale Revue der gesamten Hydrobiologie und Hydrographie* 63(3): 337-351.
- Hendry, M.J., Cherry, J.A., and Wallick, E.I. 1986. Origin and distribution of sulfate in a fractured till in southern Alberta, Canada. *Water Resources Research* 22(1): 45-61.

- Hickman, M. and Klarer, D.M. 1981. Paleolimnology of Lake Isle, Alberta, Canada (including sediment chemistry, pigments and diatom stratigraphy). *Archiv fur Hydrobiologie* 91(4): 490-508.
- Hickman, M., Schweger, C.E., and Habgood, T. 1984. Lake Wabamun, Alta.: a paleoenvironmental study. *Canadian Journal of Botany* 62: 1438-1465.
- Hills, L.V., Christiansen, O.A., Fergusson, A., Driver, J.C., and Reeves, B.O.K. 1985. Postglacial pollen and paleoclimate in southwestern Alberta and southeastern British Columbia. pp. 345-396, In: *Climate Change in Canada 5 Critical Periods in the Quaternary Climatic History of Northern North America*. (C.R. Harington, ed.), Syllogeus Series No. 55, National Museums of Canada, Ottawa.
- Holloway, R.G. 1989. Experimental mechanical pollen degradation and its application to Quaternary age deposits. *The Texas Journal of Science* 41(2): 131-145.
- Holloway, R.G., Bryant, V.M., and Valastro, S. 1981. A 16,000 year pollen record from Lake Wabamun, Alberta, Canada. *Palynology* 5: 195-208.
- Houghton, R.A. and Woodwell, G.M. 1989. Global climate change. *Scientific American*. April, pp. 36-44.
- Husband, B.C., and Hickman, M. 1985. Growth and biomass allocation of *Ruppia occidentalis* in three lakes, differing in salinity. *Canadian Journal of Botany* 63: 2004-2014.
- Husband, B.C., and Hickman, M. 1988. The frequency and local abundance of *Ruppia occidentalis* in relation to sediment texture and lake salinity. *Canadian Journal of Botany* 67: 2444-2449.
- Hutchinson, G.E. 1975. *A Treatise on Limnology, Volume III. Limnological Botany*. John Wiley and Sons, New York.
- Jones, D.C. 1987. *Empire of Dust*. The University of Alberta Press, Edmonton.
- Kantrud, H.A., Millar, J.B., and van der Valk, A.G. 1989. Vegetation of wetlands of the prairie pothole region. pp. 132-187, In: *Northern Prairie Wetlands*. (A. van der Valk, ed.), Iowa State University Press, Ames.
- Kearney, M.S. and Luckman, B.H. 1983. Postglacial vegetational history of Tonquin Pass, British Columbia. *Canadian Journal of Earth Sciences* 20: 776-786.
- Kearney, M.S. and Luckman, B.H. 1987. A Mid-Holocene vegetational and climatic record from the subalpine zone of the Maligne Valley, Jasper National Park, Alberta (Canada). *Palaeogeography, Palaeoclimatology, Palaeoecology* 59: 227-242.
- Keith, L.B. 1958. Some effects of increasing soil salinity on plant communities. *Canadian Journal of Botany* 36: 79-89.
- Kerr, R.A. 1984. Climate since the ice began to melt. *Science* 226: 326-327.

- Klassen, R.W. 1989. Quaternary geology of the southern Canadian interior plains. pp. 138-173. In: **Quaternary Geology of Canada and Greenland**, (R.J. Fulton ed.), Geological Survey of Canada, Ottawa.
- Klein, J., Lerman, J.C., Damon, P.E., and Ralph, E.K. 1982. Calibration of radiocarbon dates: tables based on the consensus data of the workshop on calibrating the radiocarbon time scale. **Radiocarbon** 24(2): 103-150.
- Kunkel, K.E. 1989. A drought's unyielding cycle. **Natural History**, Jan., pp. 48-49.
- Kutzbach, J.E. and Guetter, P.J. 1986. The influence of changing orbital parameters and surface boundary conditions on climate simulations for the past 18000 years. **Journal of the Atmospheric Sciences** 43(16): 1726-1759.
- Labagh, J.W. 1989. Chemical characteristics of water in northern prairie wetlands. pp. 56-91. In: **Northern Prairie Wetlands**. (A. van der Valk, ed.), Iowa State University Press, Ames.
- Labagh, J.W., Winter, T.C., Adomaitis, V.A., and Swanson, G.A. 1987. **Hydrology and chemistry of selected prairie wetlands in the Cottonwood Lake area Stutsman County, North Dakota, 1979-82**. U.S. Geological Survey Professional Paper 1431. United States Government Printing Office, Washington. 26p.
- Lamb, H.H. 1972. **Climate: Present, Past and Future** . Vol. 1. Methuen & Co. Ltd., London.
- Lamb, H.H. 1977. **Climate: Present, Past and Future**. Vol. 2. Methuen & Co. Ltd., London.
- Lamb, H.H. 1979. Climatic variations and changes in the wind and ocean circulation: The Little Ice Age in the northeast Atlantic. **Quaternary Research** 11: 1-20.
- Lamb, H.H. 1981. Some aspects of the cold, disturbed climate of recent centuries, the Little Ice Age, and similar occurrences. pp. 628-639. In: **Weather and Weather Maps. A Volume Dedicated to the Memory of Tor Bergeron (15.8.1891-13.6.1977)**. (G.H. Liliequist, ed.), Birkhauser Verlag, Boston.
- Lamb, H.H. 1982. **Climate, History and the Modern World**. Methuen & Co. Ltd., London.
- Lanesky, D.E., Logan, B.L., Brown, R.G., and Hine, A.E. 1979. A new approach to portable vibracoring underwater and on land. **Journal of Sedimentary Petrology** 49: 654-657.
- Langbein, W.B. 1961. **Salinity and hydrology of closed lakes. A study of the long-term balance between input and loss of salts in closed lakes**. U.S. Geological Survey Professional Paper 412. United States Government Printing Office, Washington.
- Last, W.M. 1984. Sedimentology of playa lakes of the northern Great Plains. **Canadian Journal of Earth Sciences** 21: 107-125.
- Last, W.M. 1987. Sedimentology, geochemistry, and evolution of a saline playa in the northern Great Plains of Canada. pp. 319-337. In: **Geochemistry of the Earth Surface and Processes of Mineral Formation**. Granada, Spain, March 26, 1936. (R. Rodriguez Clemente and Y. Tardy, eds.), CEP de la Biblioteca Nacional (Madrid).

- Last, W.M. 1989. Continental brines and evaporites of the northern Great Plains of Canada. Sedimentology and diagenesis of evaporites. *Sedimentary Geology* 64: 207-221.
- Last, W.M. 1990. Paleochemistry and paleohydrology of Ceylon Lake, a salt-dominated play basin in the northern Great Plains, Canada. *Journal of Paleolimnology* 4: 219-238.
- Last, W.M. and Schweyen, T.H. 1983. Sedimentology and geochemistry of saline lakes of the Great Plains. *Hydrobiologia* 105: 245-263.
- Last, W.M. and Schweyen, T.H. 1985. Late Holocene history of Waldsea Lake, Saskatchewan, Canada. *Quaternary Research* 24: 219-234.
- Last, W.M. and Slezak, L.A. 1986. Paleohydrology, sedimentology, and geochemistry of two meromictic saline lakes in southern Saskatchewan. *Geographie physique et Quaternaire* 40(1): 5-15.
- Last, W.M. and Slezak, L.A. 1987. Sodium sulphate deposits of western Canada: geology, mineralogy, and origin. pp. 197-205, In: **Economic Minerals of Saskatchewan**. Proceedings of a symposium held in Regina, Saskatchewan, November 1986. Special Publication Number 8, Saskatchewan Geological Survey, Regina.
- Last, W.M. and Slezak, L.A. 1988. The salt lakes of western Canada: A paleolimnological overview. *Hydrobiologia* 158: 301-316.
- Last, W.M. and Teller, J.T. 1983. Holocene climate and hydrology of the Lake Manitoba basin.
- Leonard, E.M. 1986. Use of lacustrine sedimentary sequences as indicators of Holocene glacial history, Banff National Park, Alberta, Canada. *Quaternary Research* 26: 218-231.
- Lerman, A. and Jones, B. 1973. Transient and steady-state salt transport between sediments and brine in closed lakes. *Limnology and Oceanography* 18(1): 72-85.
- Lichti-Federovich, S. 1970. The pollen stratigraphy of a dated section of late Pleistocene lake sediment from central Alberta. *Canadian Journal of Earth Sciences* 7: 960-971.
- Lieffers, V.J. and Shay, J.M. 1980. Distribution and variation in growth of *Scirpus maritimus* var. *paludosus* on the Canadian prairies. *Canadian Journal of Botany* 60: 1938-1949.
- Lieffers, V.J. and Shay, J.M. 1983. Ephemeral saline lakes on the Canadian prairies: their classification and management for emergent macrophyte growth. *Hydrobiologia* 105: 85-94.
- Longley, R.W. 1968. Climate and weather patterns. pp. 53-67, In: **Alberta - A Natural History**. (W.G. Hardy ed.), M.G. Hurtig Publishers, Edmonton.
- Longley, R.W. 1972. **The Climate of the Prairie Provinces**. Environment Canada, Atmospheric Environment Service, Climatological Studies No. 13.
- Luckman, B.H. 1987.

- Luckman, B.H. and Kearney, M.S. 1986. Reconstruction of Holocene changes in alpine vegetation and climate in the Maligne Range, Jasper National Park, Alberta. **Quaternary Research** 26: 244-261.
- Luckman, B.H., and Osborn, G.D. 1979. Holocene glacier fluctuations in the middle Canadian Rocky Mountains. **Quaternary Research** 11: 52-77.
- MacDonald, G.M. 1982. Late Quaternary paleoenvironments of the Morley Flats and Kananaskis Valley of southwestern Alberta. **Canadian Journal of Earth Sciences** 19(1): 23-35.
- MacDonald, G.M. 1989. Postglacial palaeoecology of the subalpine forest - grassland ecotone of southwestern Alberta: new insights on vegetation and climate change in the Canadian Rocky Mountains and adjacent foothills. **Palaeogeography, Palaeoclimatology, Palaeoecology** 73: 155-173.
- MacDonald, G.M., Beukens, R.P., Kieser, W.E., and Vitt, D.H. 1987. Comparative radiocarbon dating of terrestrial plant macrofossils and aquatic moss from the "ice-free corridor" of western Canada. **Geology** 15: 837-840.
- Mack, R.N. 1986. Alien plant invasion into the intermountain west: a case history. pp. 192-213, In: **Ecology of Biological Invasions of North America and Hawaii**. (H.A. Mooney and J.A. Drake, eds.), Ecological Studies, Volume 58, Springer-Verlag, New York.
- Martin, A.C. and Barkley, W.D. 1961. **Seed Identification Manual**. University of California Press, Berkeley.
- Mathewes, R.W. 1985. Paleobotanical evidence for climatic change in southern British Columbia during late-glacial and Holocene time. pp. 397-422, In: **Climate Change in Canada 5 Critical Periods in the Quaternary Climatic History of Northern North America**. (C.R. Harington, ed.), Syllogeus Series No. 55, National Museums of Canada, Ottawa.
- McAndrews, J.H., Berti, A.A., and Norris, G. 1973. **Key to the Quaternary Pollen and Spores of the Great Lakes Region**. Royal Ontario Museum Life Sciences Miscellaneous Publication, Toronto. 64p.
- Menely, W.A. 1963. Occurrence and movement of groundwater in Alberta. pp. 4-9, In: **Early Contributions to the Groundwater Hydrology of Alberta**. (R.N. Farvolden, W.A. Menely, E.G. LeBreton, D.H. Lennox, and P. Meyboom, eds.), Research Council of Alberta Bulletin 12, Edmonton.
- Montgomery, F.H. 1977. **Seeds and Fruits of Plants of Eastern Canada and Northeastern United States**. University of Toronto Press, Toronto.
- Moore, P.D. and Webb, J.A. 1978. **An Illustrated Guide to Pollen Analysis**. Hodder and Stoughton, London.
- Moss, E.H. 1983. **The Flora of Alberta**. 2nd ed. University of Alberta Press, Edmonton.

- Mott, R.J. 1973. Palynological Studies in Central Saskatchewan. Pollen Stratigraphy from Lake Sediment Sequences. Geological Survey of Canada, Paper 72-49. Government of Canada, Ottawa.
- Mott, R.J. and Christiansen, E.A. 1981. Palynological study of slough sediments from central Saskatchewan. pp. 133-136. In: **Current Research, Part B. Geological Survey of Canada. Paper 81-1B.** Government of Canada, Ottawa.
- Mott, R.J. and Jackson, Jr., L.E. 1982. An 18000 year palynological record from the southern Alberta segment of the classical Wisconsinan "Ice-free Corridor". **Canadian Journal of Earth Sciences** 19(3): 504-513.
- Moulton, G.E. 1983. Introduction. pp. ix-xiii, In: **Mapping the North American Plains. Essays in the History of Cartography.** (F.C. Luebke, F.W. Kay, and G.E. Moulton, eds.) University of Oklahoma Press, Norman.
- Namias, J. 1983. Some causes of United States drought. **Journal of Climate and Applied Meteorology** 22: 30-39.
- Northcote, T.G. and Larkin, P.A. 1966. Western Canada, In: **Limnology in North America.** (D.G. Frey, ed.), The University of Wisconsin Press, Madison.
- Nkemdirim, L.C. 1986. Chinooks in southern Alberta: some distinguishing features. **Journal of Climatology** 6: 593-603.
- O'Sullivan, P.E. 1983. Annually-laminated lake sediments and the study of Quaternary environmental changes - a review. **Quaternary Science Reviews** 1: 245-313.
- Pearce, F. 1991. Warmer winters fit greenhouse model. **New Scientist**, Jan., pp. 20.
- Pip, E. 1988. Niche congruency of aquatic macrophytes in central North America with respect to 5 water chemistry parameters. **Hydrobiologia** 162: 173-182.
- Rawson, D.S. and Moore, J.E. 1944. The saline lakes of Saskatchewan. **Canadian Journal of Research** 22(D): 141-201.
- Reasoner, M.A. and Hickman, M. 1989. Late Quaternary environmental change in the Lake O'Hara Region, Yoho National Park, British Columbia. **Palaeogeography, Palaeoclimatology, Palaeoecology** 72: 291-316.
- Reeves, B.O.K. 1973. The concept of an Altithermal hiatus in northern plains prehistory. **American Anthropologist** 75: 1221-1253.
- Reeves, B.O.K. and Dormaar, J.F. 1972. A partial Holocene pedological and archaeological record from the southern Alberta Rocky Mountains. **Arctic and Alpine Research** 4(4): 325-336.
- Reeves Jr., C.C. 1968. **Introduction to Paleolimnology.** Elsevier Publishing Company, Amsterdam.

- Richardson, J.L. 1969. Former lake-level fluctuations - their recognition and interpretation. *Mitteilungen Internationale Vereinigung für Theoretische und Angewandte Limnologie* 17: 78-93.
- Rissler, P.G., Birney, E.C., Blocker, H.D., May, S.W., Parton, W.J. and Weins, J.A. 1981. *The True Prairie Ecosystem*. Hutchinson Ross Publishing Company, Stroudsburg, Pennsylvania.
- Ritchie, J.C. 1976. The late-Quaternary vegetational history of the western interior of Canada. *Canadian Journal of Botany* 54: 1793-1818.
- Ritchie, J.C. 1983. The paleoecology of the central and northern parts of the glacial Lake Agassiz basin. pp. 157-170. In: *Glacial Lake Agassiz*. (J.T. Teller and Lee Clayton, eds.), Geological Association of Canada Special Paper 26.
- Ritchie, J.C. 1987. *Postglacial Vegetation of Canada*. Cambridge University Press, Cambridge.
- Ritchie, J.C. and Lichti-Federovich, S. 1968. Holocene pollen assemblages from the Tiger Hills, Manitoba. *Canadian Journal of Earth Sciences* 5: 873-880.
- Roots, E.F. 1988. The Study of Drought: Scientific Approaches to a Human Problem. pp. 3-10. In: *Proceedings of the Prairie Drought Workshop*. Saskatoon, October 1988. (D.J. Bauer, Ed.), Environment Canada, Government of Canada, Ottawa.
- Rozkowska, A.D. and Rozkowski, A. 1969. Seasonal changes of slough and lake water chemistry in southern Saskatchewan (Canada). *Journal of Hydrology* 7: 1-13.
- Russell, L.S. and Landes, R.W. 1940. *Geology of the Southern Alberta Plains*. Geological Survey of Canada Memoir 221, Government of Canada, Ottawa.
- Suachyn, D.J. 1990. A reconstruction of Holocene geomorphology and climate, western Cypress Hills, Alberta and Saskatchewan. *Canadian Journal of Earth Sciences* 27: 1504-1510.
- Schweger, C.E. 1972. Quaternary geology and geomorphology between Winnipeg and the Rocky Mountains. pp. 54-61. In: *Field Excursion C-22*. International Geological Congress, twenty-fourth session, Montreal.
- Schweger, C.E., Habgood, T., and Hickman, M. 1981. Late Glacial-Holocene climatic changes of Alberta the record from lake sediment studies. pp. 47-59. In: *The Impacts of of Climatic Fluctuations On Alberta's Resources and Environment*. Proceedings of the workshop and annual meeting of the Alberta Climatological Association, February 1981, Edmonton.
- Schweger, C.E., and Hickman, M. 1989. Holocene paleohydrology of central Alberta: testing the general-circulation model climate simulations. *Canadian Journal of Earth Sciences* 26: 1826-1833.
- Scoggan, H.J. 1978. *The Flora of Canada*. National Museum of Natural Sciences, National Museums of Canada, Ottawa.

- Shay, J.M., and Shay, C.T. 1986. Prairie marshes in western Canada, with specific reference to the ecology of five emergent macrophytes. *Canadian Journal of Botany* 64: 443-454.
- Slezak, L.A. and Last, W.M. 1985. Geology of sodium sulfate deposits of the northern Great Plains. pp. 105-115, In: *Twentieth Forum on the Geology of Industrial Minerals. Proceedings from Industrial Minerals of the Mid-Atlantic States, May 1984.* (J.D. Glaser and J. Edwards, eds.), Maryland Geological Survey, Department of Natural Resources, Maryland.
- Sloan, C.E. 1970. Biotic and hydrologic variables in prairie potholes in North Dakota. *Journal of Range Management* 23: 260-263.
- Stalker, A. MacS. 1977. The probable extent of Classical Wisconsin ice in southern and central Alberta. *Canadian Journal of Earth Sciences* 14: 2614-2619.
- Stark, M. 1987. Soil erosion out of control in southern Alberta. *Canadian Geographic*, June/July, pp. 16-25.
- Stevenson, D.R. and Borneuf, D.M. 1977. Hydrogeology of the Medicine Hat area, Alberta. *Alberta Research Council, Report 75-2.*, Edmonton. 11p.
- Stockton, C.W., and Meko, David M. 1983. Drought recurrence in the Great Plains as reconstructed from long-term tree-ring records. *Journal of Climate and Applied Meteorology* 22: 17-29.
- Street-Perrott, F.A. and Harrison, S.P. 1985. Lake levels and climate reconstruction. pp. 291-340, In: *Paleoclimate Analysis and Modeling.* John Wiley & Sons, New York.
- Street-Perrott, F.A. and Roberts, N. 1983. Fluctuations in closed lakes as an indicator of past atmospheric circulation patterns. pp. 331-345, In: *Variations in the Global Water Budget.* (F.A. Street-Perrott, F.A. Beran, and R.A.S. Ratcliffe, eds.), D. Reidel, Dordrecht.
- Teller, J.T. 1987. Proglacial lakes and the southern margin of the Laurentide Ice Sheet. pp. 39-69, Chapter 3 in: *North America and Adjacent Oceans During the Deglaciation.* (W.F. Ruddiman and H.E. Wright, eds.), Geological Society of America, The Geology of North America, v. K-3, Boulder, Colorado.
- Teller, J.T. and Last, W.M. 1981. Late Quaternary history of Lake Manitoba, Canada. *Quaternary Research* 16: 97-116.
- Teller, J.T. and Last, W.M. 1990. Paleohydrological indicators in playas and salt lakes, with examples from Canada, Australia, and Africa. *Palaeogeography, Palaeoclimatology, Palaeoecology* 76: 215-240.
- Timms, B.V., Hammer, U.T., and Sheard, J.W. 1986. A study of benthic communities in some saline lakes in Saskatchewan and Alberta, Canada. *Internationale Revue der gesamten Hydrobiologie* 71: 759-777.
- Tokarsky, O. 1973. Hydrogeological Map, Lethbridge-Fernie, Alberta. NTS 82G-H. Alberta Research Council, Edmonton.

- Trenberth, K.E. 1989. The Wayward Winds. *Natural History*. Jan., pp. 44-45.
- Trenberth, K.E., Branstator, G.W., and Arkin, P.A. 1988. Origins of the 1988 North American Drought. *Science* 242: 1640-1645.
- Ungar, I.A. 1970. Species-soil relationships on sulfate dominated soils of South Dakota. *The American Midland Naturalist* 83(2): 343-357.
- Ungar, I.A. 1974. Inland halophytes of the United States. pp. 235-305. In: *Ecology of Halophytes*. (R.J. Reinold and W.H. Queen, eds.), Academic Press, Inc., New York.
- Ungar, I.A. 1978. Halophyte seed germination. *The Botanical Review* 44: 233-264.
- Vaida, B.S. 1966. A study of some environmental factors affecting the occurrence of charophytes in western India. *Hydrobiologia* 29: 256-262.
- Vance, R.E. 1986. Aspects of the postglacial climate of Alberta: calibration of the pollen record. *Geographie physique et Quaternaire* 40(2): 153-160.
- Vance, R.E. 1987. Meteorological records of historic droughts as climatic analogues for the Holocene. pp. 17-32. In: *Man and the Mid-Holocene Climatic Optimum*. Proceedings of the 17th Annual Chacmool Conference, Calgary, Alberta, Canada. (N.A. McKinnon and G.S.L. Stuart, eds.), The University of Calgary Archaeological Association.
- Vance, R.E., Emerson, D., and Habgood, T. 1983. A mid-Holocene record of vegetation change in central Alberta. *Canadian Journal of Earth Sciences* 20(3): 364-376.
- van der Valk, A.G. and Davis, C.B. 1976. The seed banks of prairie glacial marshes. *Canadian Journal of Botany* 54: 1832-1838.
- van der Valk, A. and Davis, C.B. 1979. A reconstruction of the recent vegetational history of a prairie marsh, Eagle Lake, Iowa, from its seed bank. *Aquatic Botany* 6: 29-51.
- Verhoven J.T.A. 1979. The ecology of *Ruppia*-dominated communities in western Europe. I. Distribution of *Ruppia* representatives in relation to their autecology. *Aquatic Botany* 6: 197-268.
- Vickers, J.R. 1986. *Alberta plains prehistory: a review*. Archaeological Survey of Alberta. Occasional Papers No. 27., Alberta Culture Historical Resources Division, Edmonton.
- Vickers, J.R. and Beaudoin, A.B. 1989. A limiting AMS date for the Cody occupation at the Fletcher Site, Alberta, Canada. *Plains Anthropologist* 34: 261-264.
- Walker, B.H. and Coupland, R.T. 1968. An analysis of vegetation-environment relationships in Saskatchewan sloughs. *Canadian Journal of Botany* 46: 509-522.
- Wallick, E.I. 1981. Chemical evolution of groundwater in a drainage basin of Holocene age, east-central Alberta, Canada. *Journal of Hydrology* 54: 245-283.
- Warkentin, 1988. In: *Proceedings of the Prairie Drought Workshop*. Saskatoon, October 1988. (D.J. Bauer, ed.). Environment Canada, Government of Canada, Ottawa.

- Watts W.A. and Winter, T.C. 1966. Plant Macrofossils from Kirchner Marsh - A Paleocological Study. **Geological Society of America Bulletin** 77: 1339-1360.
- Webb III, T. and Bryson, R.A. 1972. Late- and postglacial climatic change in the northern Midwest, U.S.A.: quantitative estimates derived from fossil pollen spectra by multivariate statistical analysis. **Quaternary Research** 2: 70-115.
- Webb III, T. Kutzbach, J., Wright, H., Cushing, E., Street-Perrott, A., and Ruddiman, W. 1984. Climate since the ice began to melt. **Science** 226: 326-327.
- White, J.M., Mathewes, R.W., and Mathews, W.H. 1985. Late Pleistocene chronology and environment of the "ice-free corridor" of northwestern Alberta. **Quaternary Research** 24: 173-186.
- Whiting, J.M. 1973. **Big Quill Lake, Annual Report 1972**. Saskatchewan Research Council Engineering Division Report E 73-14.
- Whiting, J.M. 1977. **The Hydrological and Chemical Balance of the Big Quill Lake Basin with a View Towards Economic Potential**. Saskatchewan Research Council, Saskatoon.
- Whiting, J.M. 1986. The effects of climate on prairie lakes. pp. 639-656, In: **Canadian Hydrology Symposium (CHS:86) Associate Committee on Hydrology**. National Research Council of Canada, Regina.
- Williams, W.D. 1981. Inland salt lakes: An introduction. **Hydrobiologia** 81:1-14.
- Winkler, M.G., Swain, A.M., and Kutzbach, J.E. 1986. Middle Holocene dry period in the northern midwestern United States: lake levels and pollen stratigraphy. **Quaternary Research** 25: 235-250.
- Winter, T.C. 1977. Classification of the hydrologic settings of lakes in the north central United States. **Water Resources Research** 13(4): 753-767.
- Winter, T.C. 1989. Hydrologic studies of wetlands in the northern prairie. pp. 17-54, In: **Northern Prairie Wetlands**. (A. van der Valk, ed.), Iowa University Press, Ames.
- Wright, H.E., Jr. 1967. A square-rod piston sampler for lake sediments. **Journal of Sedimentary Petrology** 37: 975-976.
- Wright, H.E., Jr. 1983. Introduction. pp. xi-xvii, In: **Late Quaternary Environments of the United States: Volume 2, The Holocene**. (H.E. Wright Jr., ed.), University of Minnesota Press, Minneapolis.
- Wyatt, F.A. and Newton, J.D. 1926. **Soil Survey of the Medicine Hat Sheet**. Alberta Soil Survey, Report No. 2. University of Alberta, Edmonton.
- Wyatt, F.A., Newton, J.D., Bowser, W.E., and Odynsky, W. 1937. **Soil Survey of the Rainy Hills Sheet**. University of Alberta, College of Agriculture, Bulletin No. 28, Edmonton.

Zoltai, S.C. and Vitt, D.H. 1990. Holocene climatic change and the distribution of peatlands in western interior Canada. **Quaternary Research** 33: 231-240.

APPENDIX A

Core CI

The basal sediment of CI displays the greatest variability of all core sections. The lowermost 7 cm are massive to weakly laminated, but succeeded by 24 cm of distinctly laminated silty clay, displaying a wide range of colours (Chapter 6). This laminated sequence is overlain by weakly bedded to massive olive brown silt (90% within the 63 μm –2 μm size class), from 8.37 to 8.17 m (Fig. 16). Short sequences of laminated sediment interspersed with massive units of fine sand and silt occur from 8.17 to 7.50 m, but are interrupted by two sections of dried, blocky pale olive (5Y 6/4) silty clay at 7.98–7.88 and 7.76–7.69 m (the lowermost unit contains contorted, interrupted laminations), and a section of pale olive silt and fine sand (64% 63 μm –2 μm , 24% 2 mm–63 μm , Fig. 17), from 7.66 to 7.59 m. Above 7.50 m, long sequences of well defined laminae are interrupted only by massive units (6.55–6.46, 6.17–6.05, 5.50–5.40, 4.89–4.79, and 4.79–4.63 m) and two sections of chalky, light yellowish brown silt (6.02–6.00 and 6.26–6.25 m). One sequence of laminated sediment between 6.00 and 5.935 m contains layers with an abundance of dark grains at 5.98, 5.97, and 5.945 m.

From 4.48 to 2.34 m massive sediment dominates, varying between very dark grayish brown and dark grayish brown with sections of olive brown mottling as well as occasional thin layers of light yellowish brown silty clay. A light yellowish brown bed between 4.36 and 4.33 m contains an abundance of dark grains scattered throughout its lowermost 5 mm. Faintly stratified sediment is found between 4.12 and 3.95 m, and a single conspicuous very pale brown layer occurs at 3.775 m.

This lengthy massive unit is succeeded by distinctly laminated silty clay at 2.34–2.18, 2.07–1.98, 1.89–1.80, 1.65–1.57, and 1.43–1.32 m. Individual lamina are well defined and vary

considerably in colour. In addition, layers rich in dark grains occur at 2.33, 2.22, and 1.82 m, as well as individual silt and fine sand laminae at 2.31, 2.23, 2.035, 1.613, and 1.62 m. A concentration of dark grains also occurs within massive sediment at 2.145 m.

The upper 1.32 m of C1 consists mainly of massive sediment, although two distinctly laminated zones occur at 40–35 and 5–0 cm. Layers with abundant dark grains occur in massive sediment at 1.12–1.11 m, 65–61, 42–40, and 5 cm, and are scattered throughout the 25 to 17 cm section.

Approximately 10–15 cm of salts blanketed the sediment–water interface at the time of coring. X-ray diffraction (whole rock analysis), conducted at the Geological Survey of Canada Ottawa Laboratory, determined that these salts are 100% thenardite (Na_2SO_4).

Core L1

As cores L1 and L2 were taken within 5 m of each other and have essentially identical stratigraphy, only core L1 is described. The basal 2.5 m of the 9.75 m core is dominated by extensive laminated silty clay sequences. In addition, the 9.64 to 9.44 m section has a sparse exterior coating of cylindrical salt crystals. Laminae in the basal 2.5 m of L1 are for the most part well defined, displaying a wide range of colour variation. From 8.445 to 8.385 m, laminae are faulted and folded and include two fine sand layers.

Massive sediments dominate between 7.25 and 3.80 m. Dark grayish brown and very dark grayish brown units grade gradually into one another throughout this unit, occasionally bearing olive brown mottling and infrequent grayish brown layers up to 5 cm thick. Evidence of burrowing by benthic organisms appears occasionally. A layer of abundant dark grains occurs between 4.86 and 4.855 m. Four samples from this unit have a mean grain size composition of 60% clay, 38% silt, and 2% sand.

Distinctly laminated sections occur at 3.82–3.72, and 3.045–2.91 m, while weakly laminated to massive silty clay is found between 3.15–3.50 m. These units are generally well-defined and display a wide range of colours, as well as a 2 mm thick very fine sand laminae at 3.24 m, two thin fine sand laminae at 3.10 and 3.025 m, and a concentration of dark grains defining a sharp contact at the base of the 3.045–2.91 m laminated section. Scattered dark grains are also found from 2.985 to 2.965 m.

Similarly well-developed laminations also occur at 2.525–2.37 m, containing a concentration of dark grains between 2.525 and 2.52 m, and scattered dark grains from 2.52 and 2.50 m. Short, poorly defined laminated sediments occur at 2.075–2.065 and 2.75–2.725, both containing well-defined fine sand laminae, and the latter bearing a 1 mm thick layer of dark grains defining its upper contact.

From 2.505 to 0.75 m, massive sediment dominates, although short laminated sequences (<5 cm in length) appear occasionally, including fine sand laminae between 1.73 and 1.715 m, and at 1.39 m. Concentrations of dark grains are found in massive sediment between 2.845 and 2.82 m, 2.10 and 2.08 m, and within a faintly laminated sequence between 2.405 and 2.37 m.

From 61.5 to 75 cm, a series of weakly laminated to well-defined laminated sequences are punctuated by two fine sand lamina (1–2 mm in width), and capped by a thick concentration of dark grains. The uppermost sequence of laminae in core L1 appear intermittently between 37.5 and 52.5 cm.

Core C2

Core C2 consists of two main units. The basal 5.48 m of sand is overlain by 2.75 m of predominantly silty clay deposits. Basal sands range in colour from very dark grayish

brown to dark grayish brown, very dark gray (10YR 3/1), dark olive gray (5Y 3/2), olive gray (5Y 4/2), olive brown, and olive (5Y 5/4). This lowermost sand contains discrete layers of fine grained deposits.

From 8.175 to 8.16 m gray (5Y 5/1) clay is mixed with sand. A gray silt and sand deposit occurs between 7.93 and 7.89 m, and dark gray (5Y 4/1) silty clay is found between 7.83 and 7.76 m. An olive gray clayey silt unit (70% clay, 21% silt, and 9% sand) extends between 7.295 and 7.07 m, and a pale olive silty sand deposit (13% clay, 71% silt, and 16% sand) is found within a very fine sand unit between 6.80 and 6.745 m. The lowermost 70 cm of C2 is riddled with an abundance of coarse vegetative matter, most likely root fragments.

From 6.45 to 5.93 m stratified to weakly stratified dark gray, olive gray, grayish brown, and dark grayish brown silty clay (6.45-6.19, 6.19-6.11, and 5.93-5.88 m) is found within medium to coarse-grained sands. These deposits occur in association with a dark brown organic layer containing abundant rootlets (6.375 m), and a dark grayish brown fine sand unit with pods of organic material (6.025-5.92 m).

Overlying sands are punctuated by dark grayish brown to grayish brown silty clay (5.79-5.77 m) and a sequence of olive to dark olive gray laminated to weakly stratified silty clay, from 5.74-5.60 m (25% clay, 65% silt, and 10% sand), that has a sharp, loaded upper contact (Fig. 18). Thin layers of dark olive gray silt and olive gray silty clay occur at 5.48-5.475 and 5.30-5.295 m, respectively. Massive to weakly stratified olive silt (5% clay, 87% silt, and 8% sand) at 5.095-4.99 m, and an olive fine sand and silt unit (4% clay, 74% silt, and 22% sand), from 4.99 to 4.865 m (containing dark yellowish brown to very dark grayish brown weak lenticular stratification in the basal 5 cm), are the major stratigraphic breaks in the upper 4.75 m of medium (3% clay, 5% silt, and 93% sand) to coarse (3% clay, 2% silt, and 95% sand) grained sand deposits.

Throughout the basal 5.47 m of predominantly sand deposits of core C2 are several discrete pebble deposits. Pebbles from 2–3 cm in diameter occur at 7.40–7.30, 6.19–6.17, 5.74–5.735 (Fig. 18), 4.34–4.26 m, while pebbles to 6 cm and 9 cm in diameter are found from 4.08–4.02 and 3.965–3.90 m, respectively.

Above 3.335 m medium to coarse sands grade into olive coloured fine sands, very dark gray fine silt and sand (8% clay, 19% silt, and 73% sand) at 3.21 m, massive dark grayish brown silty clay between 2.94–2.84 and 2.76–2.68 m, and laminated silty clay sediment at 2.68 m. From 2.68 to 1.755 m, massive dark grayish brown deposits are punctuated by laminated silty clay sediments bearing the typical colour sequences described above. A very dark gray fine sand layer with scattered dark grains occurs between 2.35 and 2.345 m, and concentrations of dark grains occur within laminated silt clay at 2.62–2.59 and 2.135 and 2.125 m.

The uppermost 1.755 m of core C2 consists entirely of dark grayish brown to very dark grayish brown massive silty clay. Colour differences define crude stratification within this unit. All contacts between units in this section of the core are indistinct. A concentration of dark grains occurs at 55–53 cm. The sediment–water interface of core C2 is blanketed by a 15 cm layer of thenardite.

Core C3

From the base of core C3 (5.45 m) to 1.045 m, the dominant sediment is a variety of grades of sand (from fine to very coarse) that vary in colour from olive to olive brown, olive gray, dark gray, dark grayish brown, to very dark grayish brown, with the darkest shades tending to be confined to fine-grained units. Sections of coarse-grained sand contain distinct layers of pebbles to 2 cm in diameter (5.15–5.14, 5.12–5.105, 5.08–5.055, 2.48–2.41 m), to 3.5 cm in diameter (2.85–2.79 and 1.70–1.63 m), and to 4 cm in diameter between 4.30