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LATE QUATERNARY ALPINE AND SUBALPINE LACUSTRINE RECORDS: CANADIAN AND COLORADO ROCKY MOUNTAINS

BY

MEL A. REASONER ©

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The undersigned certify that they have read, and recommended to the Faculty of Gradute Studies and Research, for acceptence, a thesis entitled "Late Quaternary Alpine and Subalpine Lacustrine Records: Canadian and Colorado Rocky Mountains" submitted by Mel A. Reasoner in partial fulfilment of the requirements for the degree of Doctor of Philosophy

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In memory of Marika Daël Laub

ABSTRACT

Improvements for a percussion core sampling system are documented and examples of recovered core are presented. The system has proven effective in remote settings where the weight system may be a constraint. Cores of up to 5.5 m in length have been recovered and the system has functioned in water depths to 200 m. Further, multiple cores penetrating basal diamict have been retrieved. The system weighs approximately 25 kg and costs less than $600.00 (CDN).

Pollen spectra from three discrete intervals, dated by tephras and radiocarbon, are examined in a lake-wide series of nineteen cores from Lake O'Hara, Yoho National Park to investigate two aspects of pollen accumulation: 1. whether pollen taxa have been differentially deposited in sediments in different parts of the lake basin, and 2. whether the entire assemblage has accumulated to a greater extent in deeper parts of the basin. Percentages of the ten major taxa in the assemblages show remarkable basin-wide consistency within each sample set, suggesting that in this case a single core provides a representative pollen record.

Bow Valley ice had receded upvalley from the Crowfoot Lake basin shortly before ca. 11,330 ^{14}C yr BP. Inorganic sediments associated with the Crowfoot moraine were deposited in the basin between ca.11,330 and 10,100 ^{14}C yr BP. The Crowfoot advance is therefore approximately synchronous with the European Younger Dryas cold event. A similar conclusion was reached for the age of a moraine directly upvalley from Sky Pond in the Colorado Front Range. Prior to ca. 10,100 ^{14}C yr BP, the local vegetation in the upper Bow Valley was a sparse shrub - herb community dominated by *Artemisia* and Poaceae. Dramatic changes in the paleobotanical records concomitant with the abrupt onset of organic sedimentation reflect the establishment of an open *Pinus* dominated forest at ca. 10,100 ^{14}C yr BP. By ca. 4160 ^{14}C yr BP, *Picea* and *Abies* were dominant components of the local closed forest and subordinate xerophytic taxa were in decline. Further, the consistent presence of *Selaginella selaginoides* (local) and *Tsuga heterophylla* (regional) after ca. 4160 ^{14}C yr BP suggest increases in precipitation. Declines in local arboreal taxa at ca. 900 ^{14}C yr BP likely reflect expansions of valley floor meadows and descending alpine timberline in the drainage. This late Holocene vegetation response is coincident with renewed glaciogenic sedimentation in the basin.

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Chapter 2

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Chapter 3

We thank C.E. Schweger for encouragement and access to pollen reference material at the Department of Anthropology. MAR thanks the management and staff at Lake O'Hara Lodge for hospitality during fieldwork and Parks Canada for permission to sample in Yoho National Park. MAR thanks N.W. Rutter, C.E. Schweger, and M. Hickman for support from NSERC funds. We thank W. Johnson for drafting Figures 1, 2, and 3 and assembling Figures 6 and 8, and K. Hardie for photograph preparation. We thank Tim Wake (Lake O'Hara Lodge) for permission to use the photograph in Figure 5. Detailed comments by C.E. Schweger, J.R. Vickers, Y. Beaudoin, T. Webb III, and an anonymous reviewer significantly improved earlier versions of the paper.

Chapter 4

Funding was provided by Imperial Oil of Canada Co. Ltd. and Circumpolar Institute research grants awarded to Reasoner. Radiocarbon analyses were directed by Dr. J. Southon at the Lawrence Livermore National Lab, which is funded by the United States Department of Energy. Ron Fanjoy and the Geological Survey of Canada are thanked for providing Rock-Eval analyses. Geochemical analyses were provided by Pat Michael at the University of Calgary. Comments by Dr. P. Birkeland, Dr. P.T. Davis, and an anonymous reviewer significantly improved the manuscript.

Chapter 5

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Chapter 6

Funding for this study was provided by Imperial Oil of Canada Ltd., the Circumpolar Institute, and the Canadian Himalayan Foundation. Radiocarbon analyses were directed by Dr. J. Southon at the Lawrence Livermore National Laboratory, which is funded by the United States Department of Energy. Parks Canada is acknowledged for granting permission to conduct this research in Banff National Park. The Geological Survey of Canada provided Rock-Eval carbon analyses that were carried out by Dr. R. Fanjoy. Pat Michael at the University of Calgary Geology and Geophysics Department conducted the geochemical analyses. Dr. F. J. Hein assisted with petrographic determinations. Greatly appreciated field assistance was provided by Ian Spooner, David Muzzucchi, and Gerry Osborn.

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CHAPTER 1

INTRODUCTION

Since Heusser's original study of late Quaternary environments of the Canadian Rocky Mountains (Heusser, 1956) interest in the paleoenvironmental history of the area has been steadily intensifying. The Rocky Mountain region is attractive for such investigations as most of the area remains in pristine condition, and steep environmental gradients associated with mountainous topography are responsible for the presence of climatically sensitive ecotones. A number of paleoenvironmental studies have been carried out near these ecotones in order to chronicle their past movement in response to late Quaternary climate change (e.g. Schweger, 1985; Luckman and Kearney, 1986; MacDonald, 1989; Reasoner and Hickman, 1989; Beaudoin and King, 1990). Further, studies conducted in proglacial settings have provided information concerning the activity of upvalley alpine glaciers which are also very sensitive to climatic change (e.g. Leonard, 1986; in press). However, few lacustrine records, recovered from climatically responsive ecotonal sites in the region, represent the entire post-glacial history. In addition, the chronological control for many of the longer term records in the area has been limited and, until recently, radiocarbon ages have largely been determined from bulk sediment samples using conventional techniques which are susceptible to contamination with "old carbon"

This dissertation is a compilation of published papers and manuscripts that document aspects of late Quaternary paleoenvironmental investigations. The study is divided into two major categories. The first two chapters address aspects of paleoecological methodology and focus on coring techniques and the verification of fundamental assumptions used in Quaternary palynology. The last three chapters centre on glacial and paleobotanical records derived from cored lacustrine sediments that were retrieved from proglacial ecotonal sites. The specific objectives of this study are: 1) to improve the equipment and procedures for a lightweight percussion coring system in order to recover the entire postglacial record in multiple sediment cores from alpine and subalpine lacustrine basins; 2) to determine the degree to which differential pollen deposition and pollen focussing has occurred in a subalpine lacustrine basin in order to evaluate the reliability of a pollen record derived from a single sediment core; 3) to identify and radiocarbon date glaciogenic sediments associated with the Crowfoot Moraine (at the type locality in the upper Bow Valley) in order to determine the timing of the Crowfoot Advance in the region; 4) to recover core samples from a similar proglacial setting in the Colorado Front Range in order to confirm the interpretation of the Crowfoot Lake record and; 5) to provide a detailed and well dated record of paleoenvironmental change for the upper Bow Valley, Banff National Park, based on sediment, pollen, and macrofossil analyses.

In many cases, retrieval of complete paleoenvironmental records from alpine and proglacial locations has not been accomplished, or even attempted, due to the difficulties associated with the use of existing coring systems in remote settings. Chapter II documents improvements to the construction and operation of a lightweight, percussion core sampling system specifically designed for winter use in remote areas (Reasoner 1986). The system is inexpensive, simple to operate, and can be transported and operated by a small field party. The total weight of the system has been reduced to approximately 25 kg. Cores of up to 5.5 m in length have been recovered and the system has functioned well in water depths up to 200 m. Refinements to the operation of the system have resulted in enhanced core recovery that in several cases has included the retrieval of basal diamict in multiple cores from a single basin. In these cases, it can be confidently assumed that cores represent the entire post glacial record. At least 97 core samples have been recovered to date with this system, including all of the core samples that were analyzed in this study.

The degree to which pollen focussing has influenced the pollen record of a high subalpine lake in the Canadian Rocky Mountains is addressed in Chapter III. A fundamental assumption that is often made when reconstructing vegetation histories from pollen records is that pollen grains of different taxa are not differentially deposited, or focussed, within a lake basin, and therefore, a single core sample contains a pollen record that is representative of sediments within a lake basin as a whole. Few studies in North America, however, have tested this assumption with analyses of multiple sediment cores from a single lacustrine basin, and those that have (e.g. Davis and Ford, 1982; Davis et al. 1984) lack convincing between-core temporal correlation. A suite of nineteen sediment core samples from Lake O'Hara, each with two tephras and a distinctive radiocarbon dated sediment contact, provided an ideal opportunity to evaluate pollen focussing in a subalpine lacustrine basin at three discrete Holocene time intervals. The findings of this study suggest that the record from a single core is representative of the sediments basin-wide, and that a similar vegetation history would result regardless of core location. The results of this study support the paleoenvironmental interpretations derived from the Crowfoot Lake record (Chapter VI), as well as other records that are based on a single core from a lacustrine basin.

Chapter IV and V investigate the age of pre-Neoglacial cirque glaciation in the Rocky Mountains of Alberta, Canada, and Colorado, USA, by identifying and radiocarbon dating glaciogenic sediments in core samples recovered from lacustrine basins located directly downvalley from the moraines that are of interest. The Crowfoot Moraine system is widely recognized throughout the Eastern Canadian Cordillera as well as several areas in Montana (Luckman and Osborn, 1979; Osborn and Luckman 1988). However, the age of the Crowfoot Advance has only been broadly constrained as Late Pleistocene or Early Holocene (Leonard, 1986; Osborn and Luckman, 1988). During the late 1980's the spatial extent of the Late Pleistocene Younger Dryas cold event became the focus of considerable debate that centered on questions of rapid global climate change and their forcing functions. The Younger Dryas event has been traditionally associated with the North Atlantic region (e.g. Broecker et al. 1988). In recent years, however, proxy climate records obtained from a number of widely separated sites have provided supporting evidence for a more widespread cold event (e.g. Chinzei et al. 1987; Overpeck et al. 1989; Kudrass et al. 1991; Mathewes et al. 1993). In this context, resolving the exact timing of the Crowfoot Advance in western Canada has become an important research question. Crowfoot Lake provided an ideal opportunity to address the age of the Crowfoot Advance as the lake is situated directly downvalley from the type locality of the Crowfoot Moraine system. Inorganic sediments in the Crowfoot Lake core samples were associated with the Crowfoot Moraine and were bracketed by AMS radiocarbon ages derived from terrestrial macrofossils. The dating of these lacustrine sediments indicates the Crowfoot Advance occurred during the Younger Dryas Chron and provides further evidence that supports the hemispheric or possibly global influence of the Younger Dryas cold event. The Crowfoot Lake record, however, does not provide the entire sequence of sediments that would be expected from a complete climate oscillation. Apparently, sufficient time was not available for the establishment of organic sedimentation in Crowfoot Lake following deglaciation of the Upper Bow Valley and prior to onset of the Younger Dryas cooling. If this interpretation is correct, sediments representing the entire climate oscillation should be preserved in lake records situated further south where the final stages of deglaciation occurred somewhat earlier than in western Canada. Sky Pond, an alpine lake in the Colorado Front Range, was cored in order to test this hypothesis and the results of this study are presented in Chapter V.

Sky Pond is situated directly downvalley from a moraine that is correlated with the inner Satanta Peak moraine on the basis of pedological features, weathering features, and position and elevation within the cirque. The Satanta Peak moraine pair is recognized throughout the Colorado Front Range (Benedict, 1973; 1981) and, like the Crowfoot Moraine, temporal constraint for this ice advance has been limited. The sedimentary record from Sky Pond spans 12, 040 ^{14}C years and includes an interval of inorganic sediments that was associated with the moraine above Sky Pond. AMS radiocarbon ages derived from terrestrial macrofossils provide firm chronologic control for the deposition of these inorganic sediments and indicate that an advance of cirque glaciers occurred during the Younger Dryas Chron in the Colorado Front Range. Further, an interval of organic sediments was deposited after deglaciation of the Sky Pond basin (ca. 12,040 ± 70 ^{14}C years BP) and prior to the onset of inorganic sedimentation associated with the inner Satanta Peak moraine which supports the interpretation of the Crowfoot Lake record presented in Chapter IV. Both the Crowfoot Lake and Sky Pond sedimentary records add to a growing body of evidence that indicates the Younger Dryas cold event had an appreciable influence on the late Glacial climate of western North America. Moreover, the extent of the Younger Dryas glacial advances in the Canadian and Colorado Rocky Mountains, compared with various proxy records from sites in North America and elsewhere, permit qualitative estimates of the spatial variability of climate change associated with the Younger Dryas event.

A detailed Late Quaternary paleoenvironmental reconstruction for the Upper Bow Valley in Banff National Park, Canada, is presented in Chapter VI. The reconstruction is based on sediment properties, palynology, and macrofossil records. With a total of twelve AMS radiocarbon ages, determined from terrestrial macrofossils, in addition to the presence of three well dated Holocene tephras in the record, the Crowfoot Lake cores provide the longest and most detailed chronological record of vegetation history and paleoenvironmental change in the Canadian Rocky Mountains. Further, the location of the Crowfoot site near the alpine timberline ecotone, and both late Pleistocene and late Holocene moraines, has resulted in records that have been remarkably sensitive to past climatic changes. Pollen percentage, pollen accumulation rate, and macrofossil data were used in the paleoenvironmental reconstruction. Ten points of chronological control were selected for the Crowfoot age-vs-depth model that was used to derive pollen accumulation rates for individual taxa. Pollen records from the Canadian Rocky Mountains tend to be dominated by *Pinus* pollen which, in many cases, obscures changes in less abundant taxa that may be important for paleoenvironmental reconstructions. Consequently, the pollen sum for each level was increased in order to permit a recalculation of the record with *Pinus* excluded from the sum in order to enhance the relative changes of less dominant pollen taxa in the Crowfoot Lake record. The Crowfoot Lake sediment, pollen, and macrofossil records registers local environmental conditions during the late Pleistocene Crowfoot Advance, the early to mid Holocene period of relative warmth and aridity, and the mid to late Holocene period of cooling and increased precipitation that includes a strong "Little Ice Age" signal. The vegetation responses to past climate change are well defined and abrupt and this is particularly evident for changes associated with the termination of the Younger Dryas and the onset of the "Little Ice Age".

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CHAPTER 2

Equipment and procedure improvements for a lightweight, inexpensive, percussion core sampling system.

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INTRODUCTION

In general, percussion coring systems rely on a weighted driver that can be raised and dropped on the top of a core barrel in order to force the barrel into sediments. Recent advances in percussion coring technology have resulted in the development of several closely related coring systems used in both marine and lacustrine coring situations (Gilbert and Glew, 1985; Reasoner, 1986; Nesje *et al.* 1987; Nesje, 1992). The general trend of development has been toward systems that are unrestricted by water depth, lightweight and simple to operate, and more effective than gravity coring systems in terms of penetration and recovery.

Since the initial testing of the prototype corer in 1985 (Reasoner, 1986), at least 92 sediment core samples have been recovered using the original coring system and its modified versions described in this paper. In the process of recovering these cores, a number of refinements have been made to simplify the procedure and facilitate core recovery. Further, several structural modifications have been made to reduce weight and enhance the system's capabilities. The primary purpose of this report is to document these improvements and to point out some potential difficulties that may be encountered during operation. In addition, costs of materials and core splitting methods are discussed. Finally, several examples of recovered core are presented.

The system is ideal for operation in remote field settings because of its light weight and simplicity. It has been used successfully in the American and Canadian Cordillera, on Baffin and Ellesmere Islands in the Canadian Arctic, and at a number of locations throughout the Alberta Plains and Foothills. The system is designed for use on a stable ice pack surface although an anchored platform should be equally effective. The total weight of the system, equipped with lines for water depths of 50 m or less, has been reduced to approximately 25 kg. If all new materials and equipment are purchased, the cost of the system is about $600.00 CDN excluding optional items (Table 2-1). The system is theoretically unrestricted by water depth and has been effective in both shallow lakes and fjords with water depths to 200 m (Lemmen, 1990). The longest core recovered to date measures 5.5 m in length and is thought to represent a continuous 24,000 year record (M. Hickman, pers. commun. 1992). Finally, the corer has supported studies that require multiple cores from one basin (Beaudoin and Reasoner, 1992).

EQUIPMENT AND ASSEMBLY

The coring system is generally very similar to the original prototype (Reasoner, 1986) and consists of three major components; a core barrel, core head, and driver (Figure 2-1). Lengths of 3 m PVC pipe (7.6-cm inside diameter, 0.32-cm wall thickness) are used as core barrels and are fitted with simple basket-type core catchers. The core catchers consist of approximately 10 cm long fingers cut from galvanized sheet metal (Figure 2-2) and are fastened with pop rivets. The rivets are flattened with

Table 2-1. Component costs

Materials	
Core Head	$20.00
Driver	$20.00
Core Barrel (cost/barrel)	$10.00
Duct Tape	$10.00
One-way Valve*	$9.00
Reusable core catcher*	$200.00

Equipment		
50 m Climbing Ropes		$250.00
Ice Screw		$30.00
Carabiners and Slings		$25.00
Ascending Device		$36.00
Pulleys (to fit carabiners)		$36.00
Pop Riveter & Rivets		$45.00
Vice Grip Pliers		$12.00
Screw Driver Set		$12.00
Ice Auger		$80.00
Tin Snips		$8.00
Battery Drill*		$40.00
TOTAL	(CDN)	$595.00
	(U.S.)	$450.00

Optional Items *

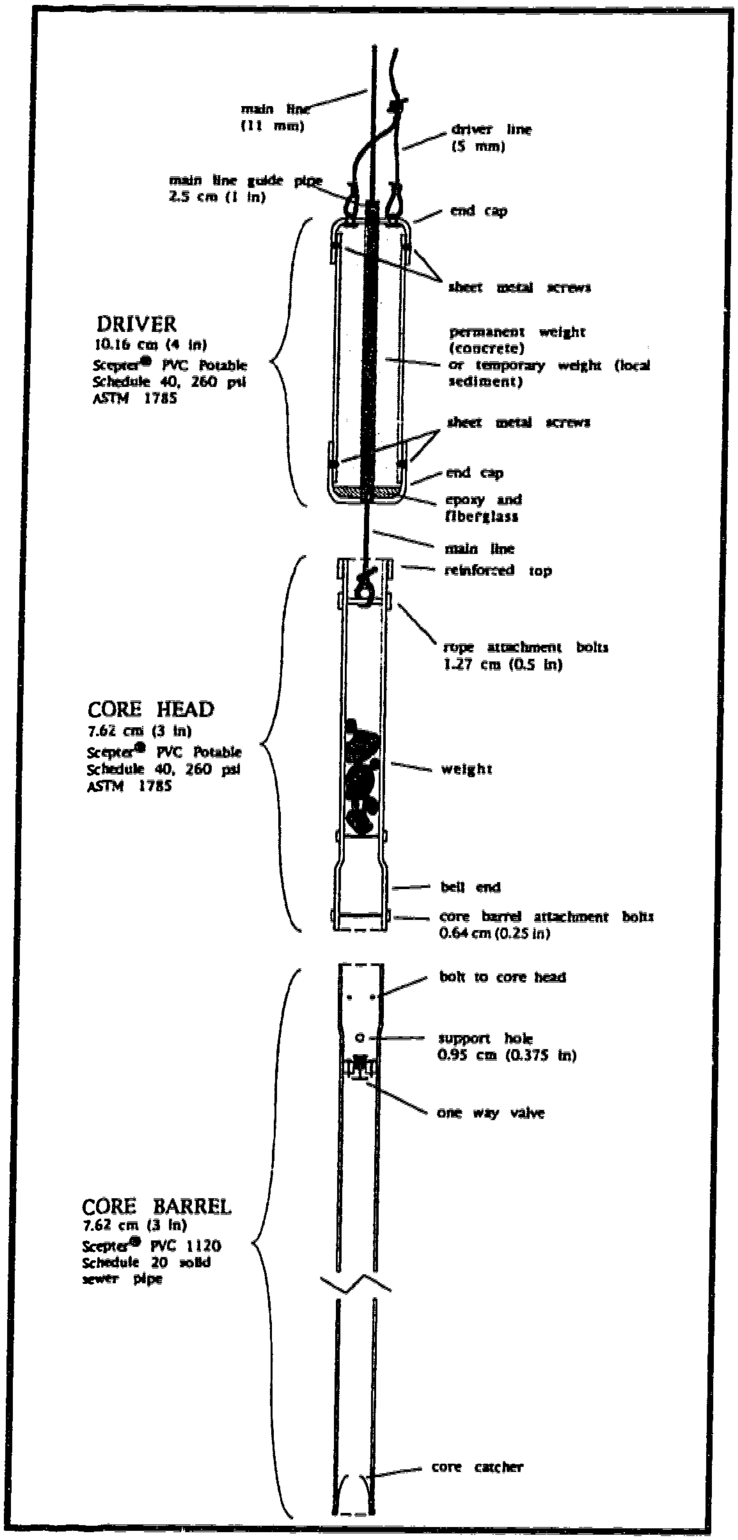

Figure 2-1. Diagram of the coring system. The driver is raised and dropped approximately 2 m to drive the core barrel. The driver is guided to the core head-core barrel assembly by the main line. Imperial units are included because in North America these materials are sold only in Imperial sizes.

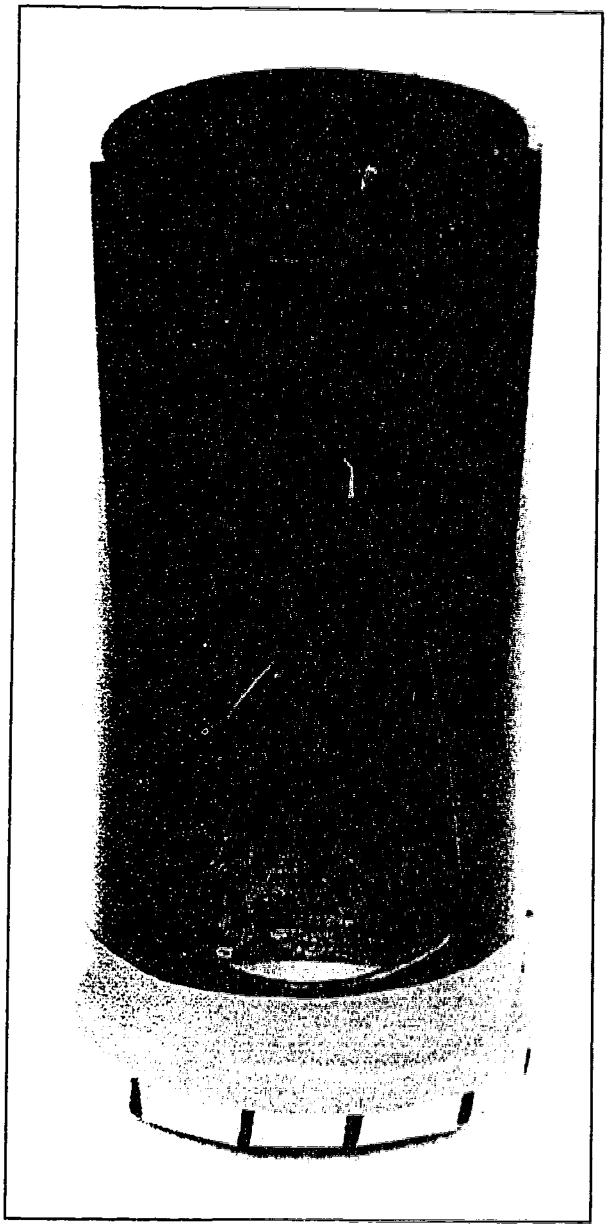

Figure 2-2. Simple galvanized steel core catcher. Core barrel is 7.6 cm inside diameter.

pliers to reduce drag on the core. The bottom of the core catcher is cut into segments and folded over the outside of the core barrel and covered with duct tape.

The core head is a 1 m length of PVC pipe (7.6-cm inside diameter, 0.64-cm wall thickness) modified to fit on the core barrel. The driver is an 0.7-m to 1-m length of PVC pipe (10.16-cm inside diameter, 0.64-cm wall thickness) with end caps and a 2.54 cm internal guide pipe attached through the centre. Climbing ropes (11 mm and 5 mm) are used for the main line and the driver line, respectively. Cores are retrieved by setting up a pulley system on the ice surface that is anchored with ice screws (Figure 2-3) or snow anchors.

Modifications

Improvements to the core barrel include addition of a reusable one-way valve, a reusable basket core catcher-cutting edge assembly, and a support hole for core removal. The one-way 7.6-cm valve, fitted in the top of the core barrel (Figure 2-1), reduces suction on the core during retrieval from the sediments. The valve is held in place by a rubber ring that is compressed against the walls of the core barrel. The only alteration to the valve involved the reversal of the compression screws so it could be easily fitted and removed through the top of the core barrel. The reusable core catcher-cutting edge assembly is shown in Figure 2-4. The cutting edge is machined from stainless steel and provides a 2-mm choke to reduce friction as sediment enters the core barrel. Spring steel is used for the core catcher fingers to prevent the fingers from folding outward during core retrieval. Finally, a 0.95-cm hole is drilled 15 cm below the top of each core barrel (Figure 2-1) for insertion of a support rod during core retrieval (described below).

The addition of a piston has been successfully employed (P.T. Davis, pers. commun. 1992) and is a possible alternative to the one-way valve. This configuration resembles a similar corer developed by Nesje (1987, 1992). The use of a piston, however, requires a third line which increases the complexity and weight of the system. Most piston corers use a steel cable for the piston line. Further, the addition of a third line increases the potential for tangled lines, particularly in deep-water situations.

A simple modification to the driver allows it to be filled with local sediment or soil on site instead of a permanent filling of concrete (Figure 2-1), which significantly reduces the system's total weight. In this case, both end caps are attached with eight sheet metal screws. This allows the driver to be dismantled in the field to facilitate loading and removal of local materials. The internal guide pipe is attached to the base end cap with PVC glue and reinforced with several layers of epoxy and fiberglass (Figure 2-1). The guide pipe is not fixed to the upper end cap, but simply extends about 1 cm through the hole in the cap.

The only significant modification to the core head involves reinforcement to reduce fracture from driver impact. Double wall thickness is achieved by gluing a coupling to the top of the core head

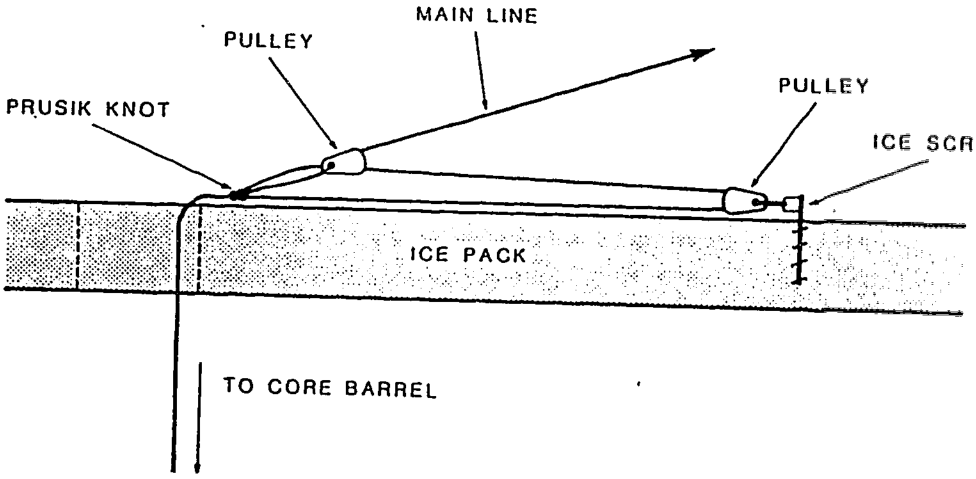

Figure 2-3. Diagram of the core retrieval system. The mechanical advantage required for core removal is achieved using a simple pulley system anchored by an ice screw or snow anchor. The mechanical advantage can be increased by adding pulleys to the system.

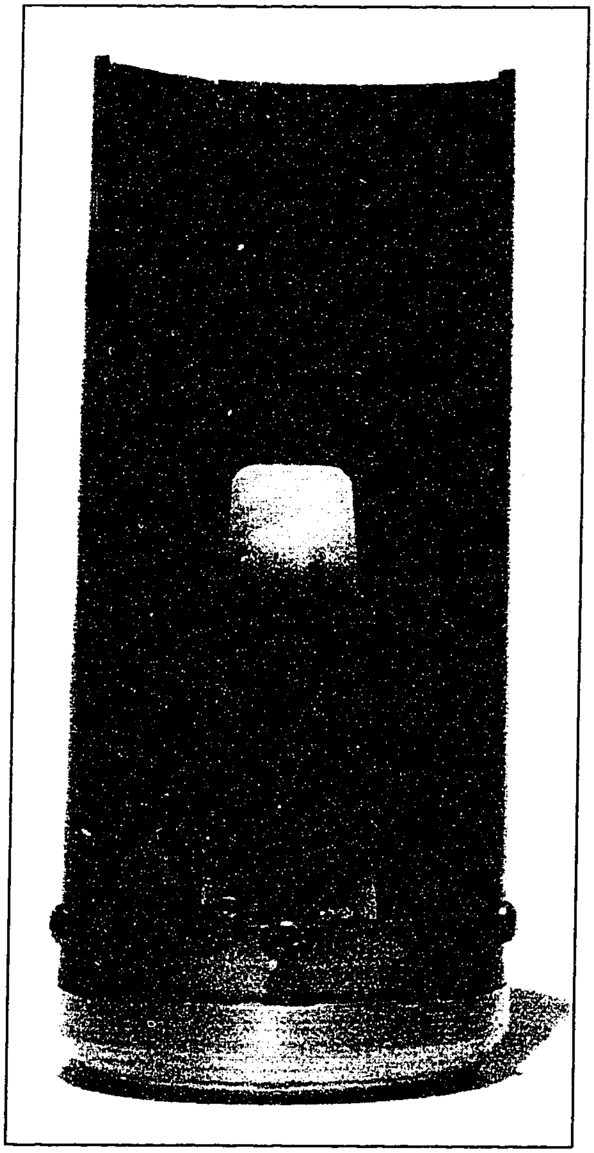

Figure 2-4. Reusable machined core catcher-cutting edge assembly. Core barrel is 7.6 cm inside diameter.

and cutting off the excess material (Figure 2-1). Care must be taken in ensure the cut is close to 90° in order to prevent driver impact from concentrating at one point on the core head. Two attachment bolts serve to centre the main line and further ensure an even impact. In the event that a fracture develops in the core head, the equipment can be salvaged by cutting the pipe below the fracture and re-drilling the bolt holes for the main line attachment. A minor simplification of the core head uses a flared or "bell" end for the core barrel attachment rather than a coupling (Figure 2-1). In either case, the inside of the bell end or coupling requires minor sanding to prevent the core head barrel binding during core removal.

MATERIALS

It is important to ensure that the PVC pipe used for the core head and core barrels is a "high strength" variety manufactured to exceed building standards. In Canada and the United States, Scepter® pipe is available and stands up well to driving impact. Lower grades of PVC pipe should be avoided as they shatter easily. Sheet aluminum was found an unsatisfactory material for the simple core catchers as core loss occurred when the fingers folded out during core retrieval. The use of 22 gauge galvanized sheet metal for core catchers minimizes this possibility. A high-strength rope is required for the main line due to the force required for core retrieval (11 mm climbing ropes are ideal because of their strength and ease of management). Unless the rope is a "static line", however, rope stretch will need to be accommodated by placing an ice screw anchor several meters from the hole. It is crucial that a static line is used for the driver line, particularly in deep water where rope stretch can be significant. Alternatively, plastic coated steel cable lines have been used for the driver and main lines (P.T. Davis, pers. commun. 1992).

DETAILS OF OPERATION

The first step in coring involves drilling a hole in the ice pack. The apparatus described here will fit through a 15.25 cm diameter hole. Extensions for an ice auger are essential in some areas. Ice pack thicknesses in excess of 2.5 m have been encountered in alpine sites in British Columbia and the Canadian Arctic.

Adding a small weight to the core head (Figure 2-1), allows it to also be used as a sounding device. Water depth should be marked directly on the main line using a device that pinches the rope and will not slide on wet and frozen rope (such as a paper clip). The length of core barrel (3 or 6 m) is subtracted from the water depth and marked on the main line in a similar fashion. The core head-core barrel assembly can then be lowered through the water column until the core barrel is at or very near the sediment surface to ready the system for driving. The bottom of the core barrel should be at the sediment surface when the first marker on the main line reaches the water surface. Feeding the main and driver lines from spools should be avoided as line twists can cause tangling. Fairly loose "over the shoulder" coils generally work well for transportation and uncoiling. To minimize tangling, both lines should be uncoiled into separate piles so that the lines feed freely from the top of the piles when the corer is lowered through the water column.

The initial part of the driving process is critical. Care must be taken to ensure that the system drives vertically. This is accomplished by starting with small driver taps while exerting sufficient tension on the main line to keep the system vertical. During this phase, penetration should occur in small increments (< 3 cm). The procedure is analogous to "setting a nail" in carpentry where a few small taps are required before driving hard. Once the core head and core barrel assembly is firmly set into the surficial sediment, the driver is allowed to free fall about 2 m for each subsequent impact. Progress can be monitored by the rate of main line descent. Slight tension on this line is required during driving to guide the driver to the centre of the core head. When firm sediment is encountered, progress will slow to <1 cm per impact and the driver will start to "double bounce" on the core head. The bouncing impacts are transmitted up the rope and are easily felt on the main line. If no progress is made after about thirty impacts, full penetration may be assumed. At this point, three or four bounces of the driver can be detected and the time between bounces noticeably increases. Continued hard driving under these circumstances may result in core head fracture .

During driving, the top marker on the main line arriving at the water surface indicates when the top of the core barrel reaches the sediment-water interface. This, however, does not necessarily indicate that the core barrel is full. In all coring operations to date, the length of recovered core, to a varying extent, has been less than both the length accounted for by the main line and the depth of the mud line on the outside of the core barrel. This phenomenon has been noted in other coring systems (e.g. Wright, 1980) and may be due to: 1) the loss of core from an ineffective core catcher, 2) the compaction of sediment during driving, or 3) the friction of sediment entering the core barrel overcoming the strength of the sediment resulting in penetration without sediment actually entering the core barrel. The first possibility can be eliminated because, in several situations where this discrepancy has been large, the mud line on the *inside* of the core barrel has been only a few centemetres above the recovered surficial sediments. Although compaction cannot be ruled out, the absence of water escape structures in recovered core suggests that the third possibility may be the most likely cause of the problem. Regardless of its origin, the severity of this problem varies considerably between sediment type. Once one core is recovered from a given lake, the degree to which a core can be "overdriven" in order to fill the core barrel can be roughly estimated.

If it is determined that more than 3 m of sediment is present, assembly of a 6 m core barrel is necessary. Two 3-m core barrels are easily connected by joining a regular end and flared or "bell" end with PVC glue. This connection should be reinforced with about ten 0.32-cm x 0.96-cm sheet metal screws that require 0.16-cm guide holes. In this situation, the one-way valve is inserted at the top of the upper barrel and the 0.95-cm support hole in the lower barrel is covered with duct tape. The driving procedure for a 6-m core barrel is the same as for a 3 m core described above.

When the driving process is considered complete, the driver is hauled to the surface and removed from the main line. An ice screw is then inserted into the ice pack to serve as an anchor for the pulley system, providing the mechanical advantage required to lift the core from the sediment (Figure 2-3). Ice screws, carabiners (snap links), slings, and pulleys are common pieces of climbing equipment that are ideal for use in the removal system. Also, replacing the prussik knot (Figure 2-3) with a climber's ascending device significantly simplifies the removal procedure.

In areas of high snowfall, a thick layer of slush may be present on top of ice surface that is insulated from freezing by the snowpack. Under these conditions, placement of an ice screw may be difficult. An alternative and equally effective anchor can be achieved by simply burying a pair of skis in a snow trench constructed perpendicular to the main line. In many cases, extra pulleys are required to increase the mechanical advantage necessary to retrieve the cores.

Once free of the bottom sediment, the core head and barrel are hauled to the surface and a 80-cm length of steel rod is inserted through the support hole at the top of the core barrel. The core barrel can then rest safely on the ice surface while the core head is removed. At this point a small weight on a fishing line is used to determine the depth to the sediment surface inside the core barrel. The core barrel is then raised to this level and a hole is drilled slightly above the sediment surface to allow water drainage. A battery hand drill is very convenient for drilling holes but a manual drill is sufficient and more reliable in the field. The entire core can then be removed taking care to keep the top of the core barrel at a higher level than the base.

If more than 3 m of core has been recovered, the core barrel is hauled to the surface, the core head removed, and water drained as previously described. The top 3 m core barrel is then raised above the ice surface and a second support rod is inserted through the duct tape covering the support holes in the top of the lower core barrel. This disturbs sediment near the lower support hole but removes the risk of losing the entire lower core section. The 6 m core barrel can now rest on the lower support rod. A cut is made with a hacksaw above the lower support and a piece of sheet metal is inserted into the cut. The upper core barrel is then removed using the sheet metal to prevent sediment loss. The lower 3 m of core is then removed as described above.

During winter coring operations in remote areas, preventing the cores from freezing may not be possible. Once frozen, however, the risk of further disturbance during transportation is eliminated. Generally, the cores are left overnight on the ice surface to freeze and subsequently cut into convenient lengths for transportation. PVC end caps are ideal for sealing the cut ends and can either be glued or taped in place. In situations where temperatures are not cold enough to freeze the cores, sediment disturbance is minimized by gently transporting the core sections vertically. Finally, the core sections are labelled with an indelible marker immediately after cutting.

CORE SPLITTING

Three general techniques have been used for core splitting: splitting with a thin wire, cutting with a band saw, and cutting with a high speed diamond rock saw. The most satisfactory results were obtained by using a high speed rock saw. This method quickly and easily cuts through fine grained organic sediments as well as large clasts. The cores are cut while completely frozen and the surface washed immediately after splitting. Photographing core as soon as possible after the cores are split is desirable as significant changes in sediment colour occur rapidly with oxidation. Splitting with a piano wire requires the core barrel to be cut on opposite sides with a rotary hand saw or table saw. Once the core has completely thawed, the wire is pulled through the sediment and the two halves gently pried apart. Major difficulties arise when large clasts or organic fragments are present. Further, particularly with inorganic sediments, this method can obscure fine details of sedimentary structure. This loss of detail may be circumvented by cutting the core while frozen with a band saw and washing off the smeared surface. However, the main drawbacks of using a bandsaw are that two blades are usually destroyed for each split core and that sediments tend to work into many internal areas of the saw.

EXAMPLES OF RECOVERED CORE

Several examples of recovered core are presented in Figure 2-5 to demonstrate the capabilities of the system and exhibit degrees of core disturbance. All cores were recovered using a simple galvanized basket core catcher (which provides a relatively blunt cutting edge) and without the use of a piston. Presumably, sediment disturbance would have been reduced by using a piston or the reusable cutting edge-core catcher assembly.

In many cases, particularly with sediments of relatively high organic carbon content (gyttja), core disturbance is minimal. The laminated gyttja in Figure 2-5a and upper sections of Figures. 2-5b and 2-5c exhibit ideal core recovery with very little core disturbance. Mazama tephra (*ca.* 6800 yrs BP) is present in Figure 2-5a. In cores containing both high and low organic-carbon sediment, core disturbance in the form of down-warping along the core edge (coning) is primarily restricted to the underlying light grey inorganic sediment (Figure 2-5b, 2-5c). The lower inorganic core section shown in Figure 2-5c shows both moderate coning and disturbance related to subaqueous slumping. Evidence for mass movement in these sediments is also apparent on seismic records from the site (Reasoner and Rutter, 1988). The contact between dark organic and underlying light grey inorganic sediments in

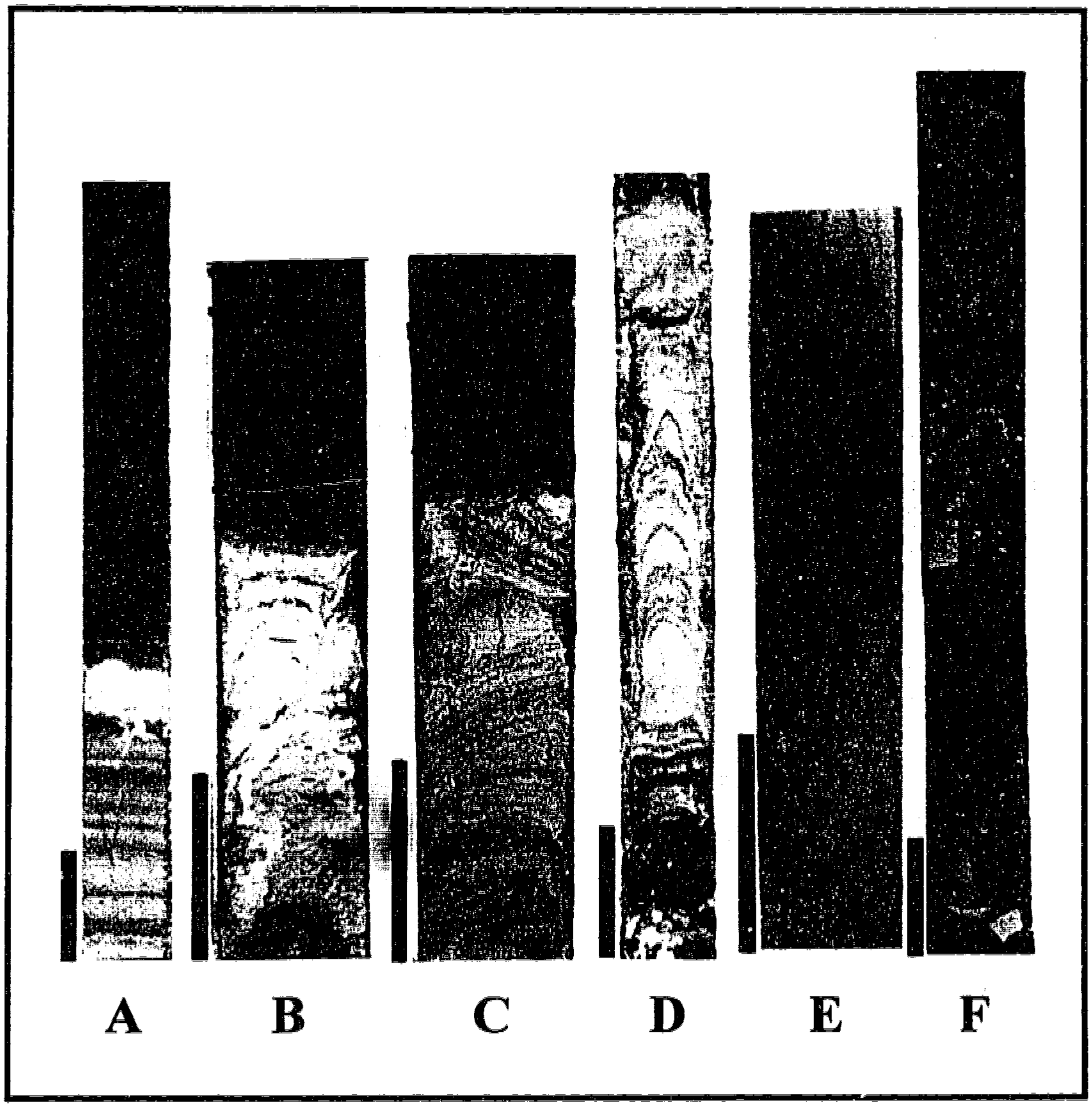

Figure 2-5. Examples of recovered core: 2-5a. minimal coring disturbance of laminated gyttja containing Mazama Tephra (*ca.* 6800 yrs BP); 2-5b and 2-5c undisturbed dark gyttja overlying light grey inorganic sediment that display moderate coning; 2-5c. the underlying light grey inorganic sediment show both moderate coning and disturbance from mass movement; Examples of severe coning in laminated inorganic 2-5d. lacustrine and; 2-5e. marine sediments; 2-5f. massive "basal" diamicton with a portion of the core catcher visible at the base. All cores are 7.6 cm in diameter. Scale bar to the left of each core represents 10 cm.

Figure 2-5b and 2-5c has been radiocarbon dated at *ca.* 10,100 yrs BP (Reasoner and Rutter. 1988). Extreme coning has occurred in lacustrine sediments of low organic-carbon content (Figure 2-5d) and in marine sediments (Figure 2-5e). Figure 2-5f shows an example of a core that penetrated massive "basal" diamicton. In circumstances where "basal" diamicton is recovered in multiple cores from a single lake basin, a high degree of confidence can be placed on the assumption that the cores represent the entire post-glacial record.

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CHAPTER 3

Evaluation of differential pollen deposition and pollen focussing from three Holocene intervals in sediments from Lake O'Hara, Yoho National Park, British Columbia, Canada: intra-lake variability in pollen percentages, concentrations and influx.

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INTRODUCTION

In palaeoecological studies, palynologists conventionally produce pollen records from one coring site within a lake basin. This assumes that differential deposition and pollen accumulation within sediments have not significantly affected the record and therefore that a single coring site is representative for the lake basin. However, once pollen is input to the lake system, it is subject to transport and sedimentation processes so that a single coring site may not provide a representative record for interpretation of vegetation history. We examined this assumption by investigating the impact of two specific processes on the recruitment of pollen to lake sediments - pollen focussing and differential pollen deposition - through a case study of nineteen cores from Lake O'Hara (latitude 51°21'N, longitude 116°21'W, about 2015 m a.s.l., 43 m maximum measured water depth), Yoho National Park, British Columbia (Figures 3-1 and 3-2).

Reasoner and Hickman (1989) used the pollen record of a single core from Lake O'Hara (LOH25; Figures 3-3 and 3-4) as one basis for describing the Late Quaternary vegetation history of the area. Good chronological control, provided by stratigraphic characteristics dated with tephrochronology and radiocarbon, was available for the Lake O'Hara cores. Three well-defined chronostratigraphic markers occur in the sediments: Bridge River tephra (ca. 2350±50 yr BP; Mathewes and Westgate 1980), Mazama tephra (ca. 6845±50 yr BP; Bacon 1983), and a sharp basal contact between overlying organic and underlying clastic sediments, dated about 10,100±200 yr BP (Reasoner and Rutter 1988; Figure 3-4). These chronostratigraphic markers allowed correlation of the cores and pollen sampling at three discrete time horizons: directly above the basal contact, and directly below each tephra. Therefore, we were able to examine the basin-wide pattern of pollen assemblages at three distinct intervals using three discrete sample sets.

Previous studies of differential pollen deposition and pollen focussing have often concentrated on modern pollen caught in traps suspended in the water column in the lake (e.g., Davis 1967, 1968) and short sediment cores from the lake floor (e.g., Davis et al. 1971). Much of this work has been carried out in North America by M.B. Davis and her colleagues, notably at Frains and Mirror Lakes (Davis 1967, 1968; Davis et al. 1971; Davis 1973; Davis and Brubaker 1973; Davis and Ford 1982; Davis et al. 1984). Related studies have been undertaken in the English Lake District (U.K.), for instance at Blelham Tarn, in particular by Pennington (1974, 1979), Peck (1973) and Bonny (1976, 1978, 1980). Chronologic control for sediment cores has sometimes been limited. For example, in Davis and Ford's (1982) study at Mirror Lake, the central core had nine radiocarbon dates, but only two of their four other cores had radiocarbon control. Most studies have looked at surface sediments (e.g., R.B. Davis et al. 1969), or the last few hundred years of sediment deposition (e.g., Burden et al. 1986). Fewer studies (e.g., Davis and Ford 1982) have examined focussing or differential pollen deposition from earlier in the Holocene record (Edwards 1983).

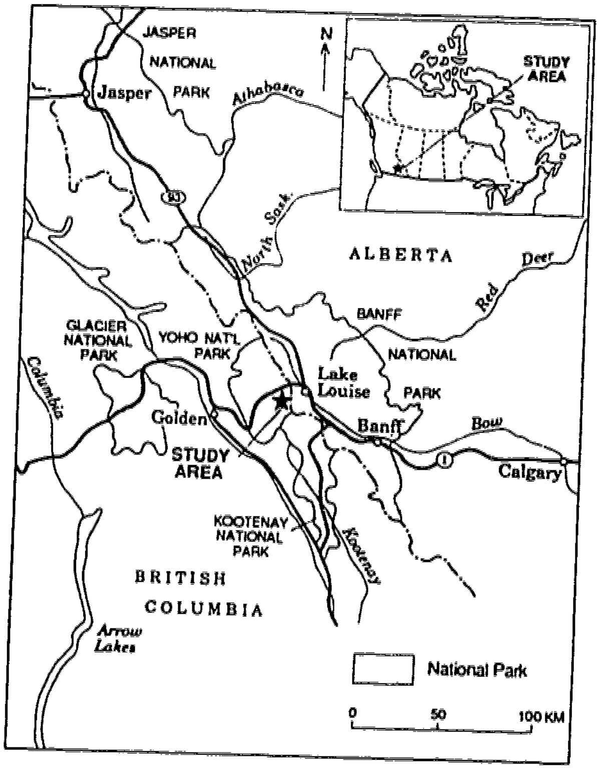

Figure 3-1: The location of Lake O'Hara in Yoho National Park, British Columbia.

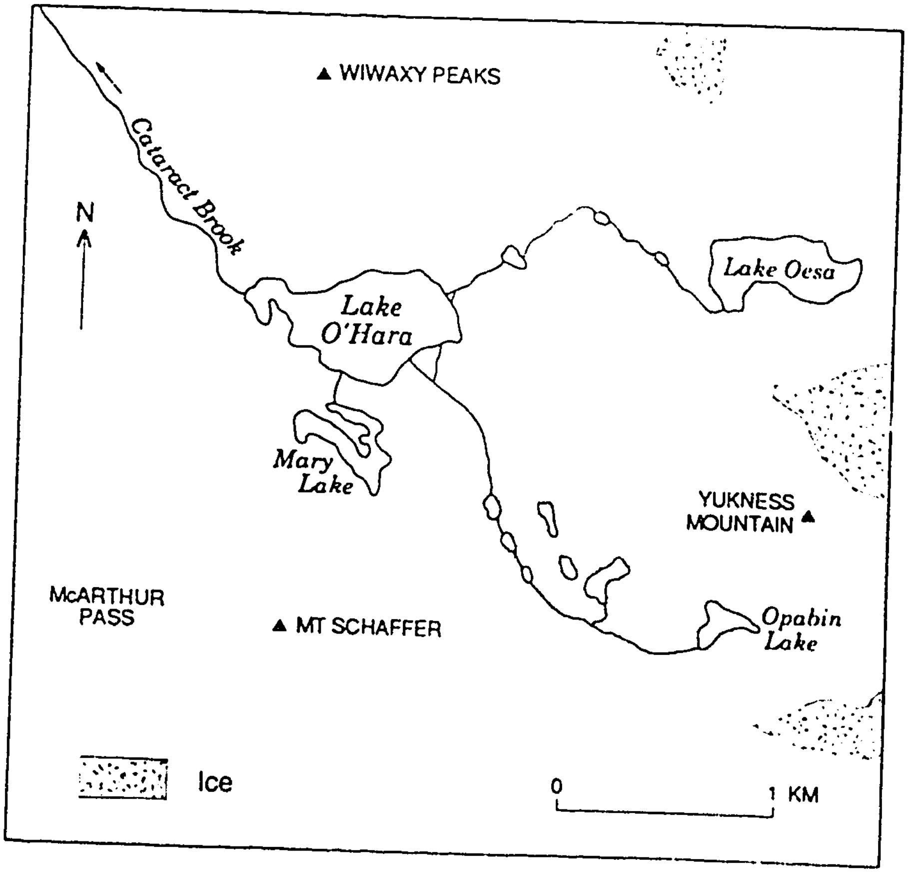

Figure 3-2: Lake O'Hara and its vicinity, Yoho National Park, British Columbia.

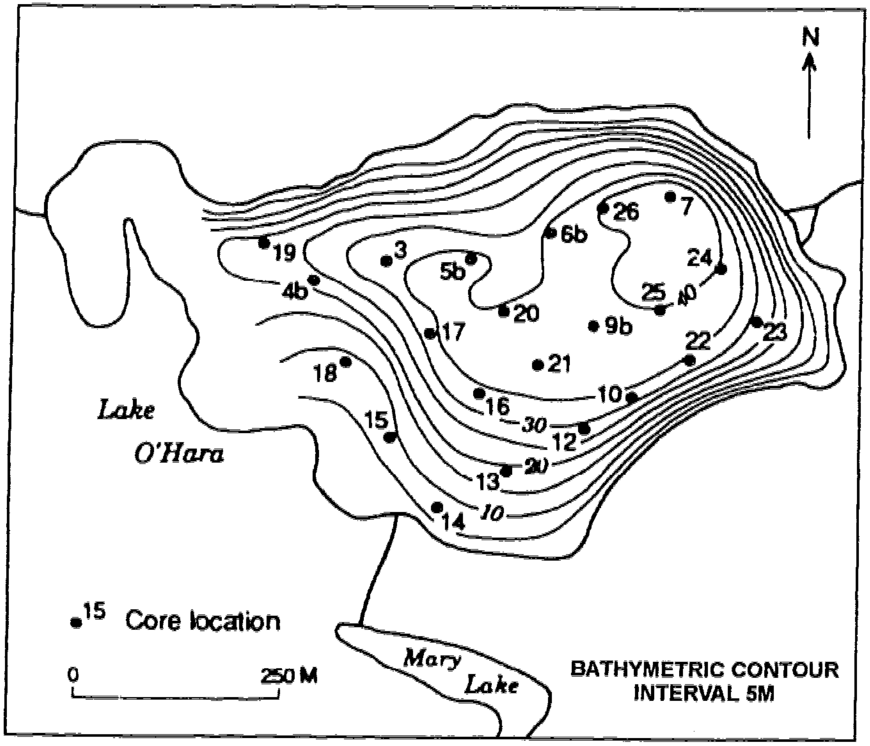

Figure 3-3: The location of cores from Lake O'Hara used in this study and the bathymetry of the lake basin.

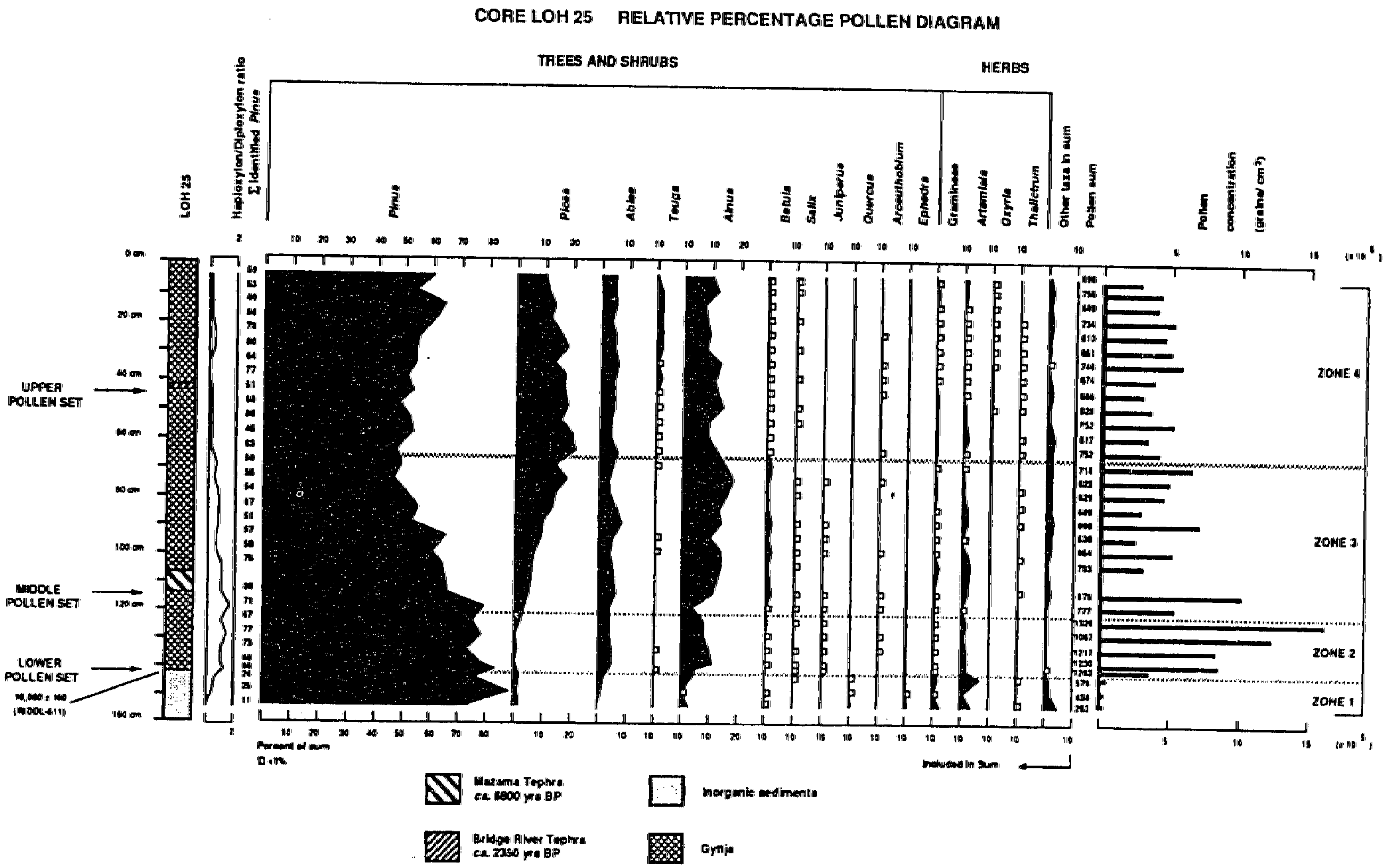

Figure 3-4: Abbreviated pollen percentage diagram for core LOH25, Lake O'Hara. Only selected taxa are shown here; the complete diagram and discussion may be found in Reasoner and Hickman (1989). Raw pollen data are given in Reasoner (1988, Appendix 2).

Pollen focussing is a term that has been implicit in previous work (e.g., Davis and Ford 1982; Davis et al. 1984). Likens and Davis (1975) noted that sediment accumulation may vary throughout a basin, with greater amounts accumulating in deeper parts of the basin; they call this "sediment focussing". Similar mechanisms of transport and deposition may affect the accumulation of pollen within lake sediments (Davis et al. 1984). We call this process "pollen focussing". In this case, pollen of all taxa is moved as one component of the sediments (i.e., there is no sorting) and thus the relative proportions of the various taxa remain unchanged across the basin. That is, the ratios of taxa should remain constant despite sample location. However, total pollen influx values will vary across the basin. Also, if sediment focussing occurs, total pollen influx values may vary over time, although the amount of pollen input to the lake remains unchanged (Davis et al. 1984). Lake-wide pollen concentration values may also vary, depending on local variations in sedimentation rates and the amount of sediment resuspension and mixing (Davis et al. 1984). We examined the patterns of total pollen concentration and total pollen influx in the Lake O'Hara assemblages to see if there had been greater accumulation in deeper water areas, thus showing whether significant pollen focussing had occurred.

Sediment and pollen focussing will be strongly influenced by basin characteristics, such as size and morphology. For instance, Lehman (1975) examined the impact of basin morphology on sediment focussing and concluded that it may be most pronounced in cone-shaped basins. Movement of sediment to deeper parts of the basin leads to greater sedimentation rates in these areas. As the basin fills, sediment is deposited over a larger area. Thus sedimentation rates may decrease over time, and therefore pollen influx values vary, even though the actual amount of sediment deposited within the entire basin does not vary (Davis et al. 1984). At Lake O'Hara, we examined sedimentation rates across the basin, specifically in relation to water depth, to see if there was evidence for significantly greater sediment accumulation in deeper water areas. Lake O'Hara's morphology, with steep sides and a relatively flat floor (Figure 3-3), suggests that sediment may not have been strongly focussed to deeper water areas and therefore that pollen focussing may not have had a major impact on the pollen assemblages. In addition, because the basin has a narrow littoral zone (Figure 3-3), the core sediments are unlikely to show evidence of lake-level fluctuations, in contrast to the more gently-sloping shallower lakes examined by Winkler et al. (1986) for example, and discussed by Digerfeldt (1986).

Pollen grains of different taxa may be differentially deposited at sites within the lake partly because of varying grain masses (Davis and Brubaker 1973). Thus, pollen from particular taxa may be selectively accumulated in sediments of certain parts of the basin. In this case, sediments of similar age should yield basin-wide trends in concentration, influx, and percentage values for individual taxa. As a result, the relative proportions, or ratios, of various taxa would vary with sample location (Davis et al. 1971). Because these trends reflect underlying processes controlling pollen accumulation, they may show relationships with parameters such as water depth or prevailing wind direction, for example

(Davis et al. 1971; Davis and Brubaker 1973). Therefore, a single pollen record may not represent pollen accumulation in the basin as a whole and, consequently, may not accurately reflect Quaternary vegetation history. We thought that if significant differential pollen deposition had occurred in Lake O'Hara, this should be distinguished by distinct areal patterning of the percentage, concentration and influx values of pollen taxa within each sample set. We believed that these patterns would most likely be related to water depth. If, however, a single core represents pollen deposition lake-wide, there should be little core-to-core variability within each sample set.

Lake size may have an important impact on recovered pollen assemblages because it affects the relative proportions of regional and local pollen input to the lake (Jacobsen and Bradshaw 1981). Lake O'Hara is larger (surface area 34.4 ha) than most lakes where differential pollen deposition or sediment focussing has been identified previously (e.g., Frains Lake and Mirror Lake surface areas are 6.7 ha and 15 ha respectively). Therefore regional input is likely to be more strongly reflected in Lake O'Hara's pollen assemblages because of an even input of pollen from extralocal sources over the lake surface. This might also result in lack of differentiation of various pollen taxa in different parts of the Lake O'Hara basin. Any differential pollen deposition would be less apparent in the pollen assemblages, particularly near the lake centre. Differentiation of pollen assemblages might only be found in sites very close to the basin edge where local minor pollen types might be better represented, as found by Davis et al. (1971) at Frains Lake.

Our pollen assemblages provided large samples from three levels within three pollen zones identified by Reasoner and Hickman (1989; Figure 3-4). These zones were defined by fluctuations in the major pollen taxa and were used as a basis for inferring vegetation history. Because the sets have distinct temporal separation, we hypothesized that variability in pollen assemblages between the sets should be greater than variability within each set, if there is no evidence for differential pollen deposition. Thus, the sample sets should be readily distinguished by their pollen assemblages. This would suggest that characteristics on which the zones were defined in the single core are "real" rather than arising from random fluctuations in the data and therefore that the zones are more useful than simply a descriptive device within the single core. If pollen assemblages from these large sample sets are distinct and clearly distinguishable, then this suggests that the zonation could be extended lake-wide. This result would justify making inferences about vegetation history based on zonation from a single core and strengthen the assumption that a single core represents the lake-wide record.

LANDSCAPE CHARACTERISTICS OF THE LAKE O'HARA AREA

Lake O'Hara is in the upper Cataract Brook Valley in Yoho National Park, British Columbia (Figures 3-1, 3-2 and 3-3), just west of the Continental Divide, at an elevation of 2015 m a.s.l. The bedrock immediately around the lake is predominantly quartzitic sandstone with some shales of the

Lower Cambrian Gog Group (Cook 1975). The Lake O'Hara region has a rigorous Cordilleran climate of relatively brief, warm summers and long, cold winters, with estimated mean January and July temperatures of -14.9°C and 11.3°C respectively, based on data for Boulder Creek (51°23'N, 116°32'W, 1219 m a.s.l.; Atmospheric Environment Service 1982). Ice covers the lake from about October to May in most years.

Coniferous upper subalpine forest surrounds Lake O'Hara (Figure 3-5). The forest is dominated by subalpine fir (*Abies lasiocarpa*) and Engelmann spruce (*Picea engelmannii*), with open stands of menziesia (*Menziesia ferruginea*) and a ground cover including feathermosses (Coen and Kuchar 1982). Alpine larch (*Larix lyallii*) and whitebark pine (*Pinus albicaulis*) are both found near timberline, which occurs at about 2150 m a.s.l. (Coen and Kuchar 1982). Whitebark pine is a major component of vegetation on the lower western slopes of Wiwaxy Peaks (Figure 3-2), downvalley from the lake (Coen and Kuchar 1982). Significant areas of alpine vegetation occur upvalley from Lake O'Hara, particularly around Lake Oesa, on the slopes of Odaray Mountain in McArthur Pass, and on Opabin Plateau where vegetation is characterized by mountain heathers (Ericaceae) (Coen and Kuchar 1982).

LAKE O'HARA BATHYMETRY, CORE STRATIGRAPHY AND CHRONOLOGY

In conjunction with the coring programme, the bathymetry of Lake O'Hara was investigated by sounding and acoustical profiling (Reasoner and Rutter 1988). These revealed that, as expected for glaciated terrain, Lake O'Hara has comparatively steep sides and a broad, relatively flat floor (Figures 3-3 and 3-6). The acoustical profiles (Figure 3-6) showed laminated, probably clastic, sediments ponded between several low relief (ca. < 5 m) ridges, possibly of till or bedrock, in the basin floor, particularly towards its western end (Reasoner and Rutter 1988). Ponded sediments and ridges are draped with Holocene gyttja (Reasoner and Rutter 1988). The steep basin walls may be either bedrock or subaqueous talus cones. The deepest part of the basin, where water depths exceed 40 m, occurs near its eastern end (Figures 3-3 and 3-6). The lake is probably dimictic (Hickman, pers. comm.); Smith (1978, 1981) found the similar but larger Hector and Bow Lakes were thermally stratified in summer with fall overturning.

The nineteen core locations were laid out in a rough grid pattern (Figure 3-3) to obtain samples from throughout the lake basin. Most cores were taken from the relatively flat floor of the lake basin. To avoid subaqueous talus slopes, slumped sediments, and areas of significant resuspension, the steep slopes around the basin periphery were not sampled. The acoustical profiles (Figure 3-6 and Reasoner and Rutter 1988) showed evidence of significant slumping in only one location, along the southern margin of the lake, south of core site LOH13. This appears to have involved only the pre-Holocene sediments. Core LOH25 was retrieved from the deepest part of the basin (Figure 3-3), in an area revealed by the acoustical profiles to be one of undisturbed sediment accumulation (Figure 3-6). This

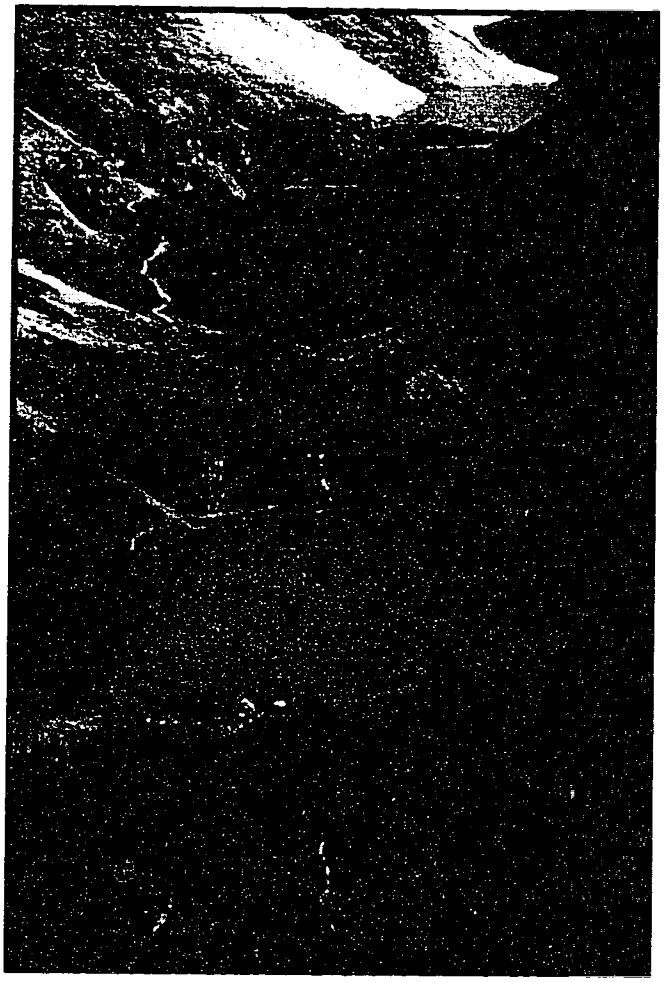

Figure 3-5: View of Lake O'Hara from the west. Several large avalanche tracks are visible along the north shore of the lake. The smaller lake at higher elevation is Lake Oesa. Photographed by Tim Wake, 1985.

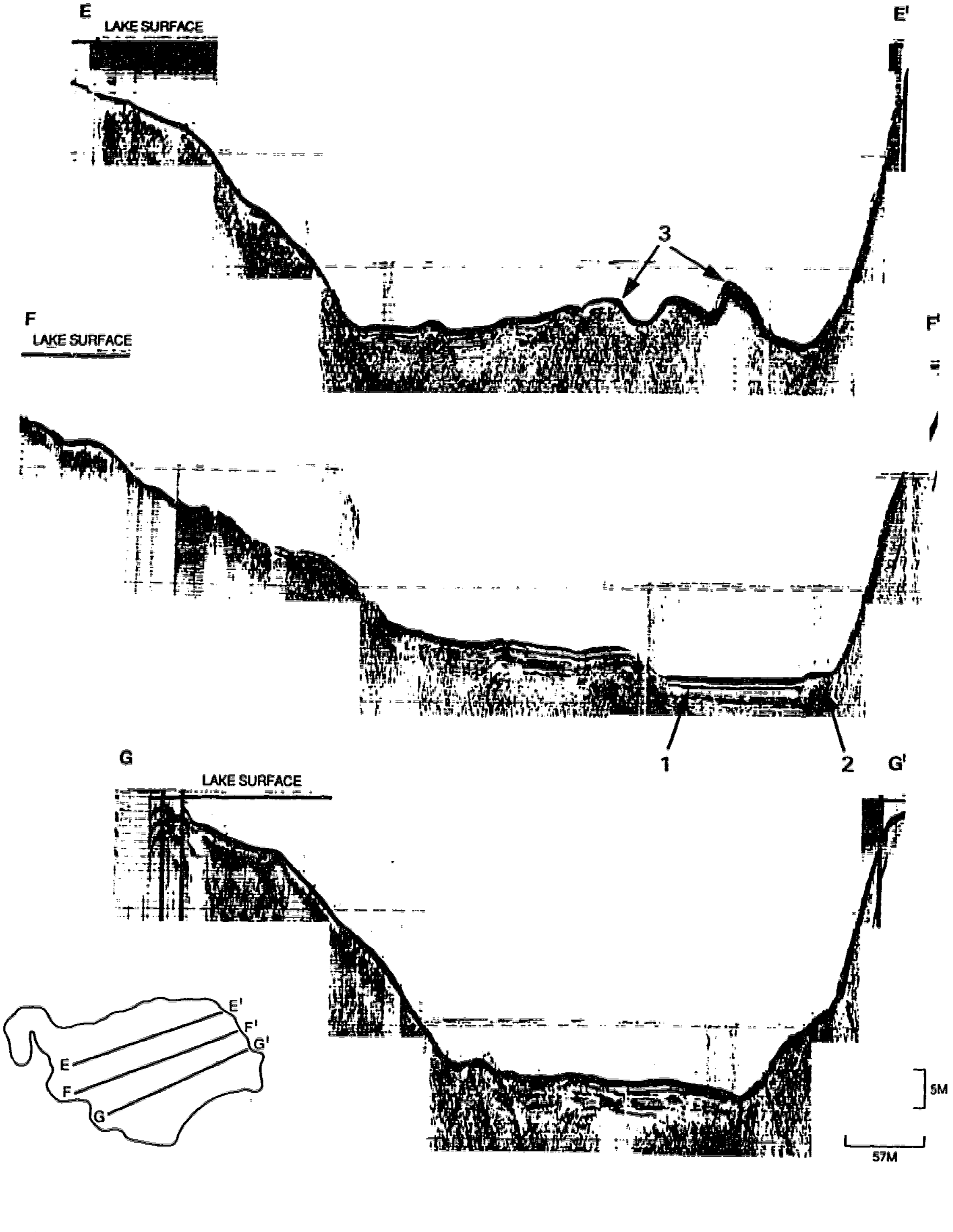

Figure 3-6: Three acoustical profiles from traverses across Lake O'Hara showing bathymetry and lake bottom sediments. 1. First strong reflector probably originating from gyttja/clastic contact. 2. Clastic sediments disturbed by slumping with gyttja draped over the toe of the slump. 3. Cones of coarse debris originating from avalanche slopes along north shore. Note horizontal undisturbed sediments in deepest portion of F-F' profile.

core was sampled for plant macrofossil, pollen, and diatom analysis. The Late Quaternary vegetation history of the Lake O'Hara area, based on this record and a record from the higher elevation Opabin Lake (Figure 3-2), has been described previously by Reasoner and Hickman (1989).

The stratigraphy of all cores recovered from Lake O'Hara is similar. The lake sediments consist of organic-rich material, mainly gyttja, overlying predominantly laminated clastic sediments. The clastic and organic sediments are separated by a sharp distinct basal contact (Reasoner and Rutter 1988). The total sediment thickness overlying the basal contact varies from 1.78 m to 0.4 m. Conifer needles from gyttja at this contact in cores LOH14 and LOH25 gave AMS radiocarbon dates of 10,100±200 (RIDDL-433) and 10,060±160 (RIDDL-511) yr BP (Reasoner and Rutter 1988). The contact appears conformable in all cores except LOH3, where it was undulating and appears to truncate underlying laminae (Reasoner and Rutter 1988). The similarity of the radiocarbon dates obtained from cores in widely separated locations and near minimum and maximum water depths emphasized that the contact is the same age throughout the basin. Thus, it represents an abrupt change in the sediment character in Lake O'Hara. Based on differences in pollen spectra composition above and below the contact, and the dramatic first appearance of abundant conifer needle macroremains above this level, Reasoner and Rutter (1988:1045) suggested that it represents "reduced clastic sedimentation associated with the postglacial stabilization of the landscape by vegetation in the vicinity of Lake O'Hara". Improved climate and reduced turbidity may have caused greater productivity and thus greater accumulation of autochthonous organic material. However, Lake O'Hara has never been highly productive in the Holocene, as shown, for instance, by the diatom record (Hickman, pers. comm. from unpub. data).

The organic-rich sediments in the upper part of the cores contain two distinct tephra layers: Bridge River and Mazama (Figure 3-4). Reasoner and Healy (1986) identified Mazama and Bridge River tephra from nearby Mary Lake, south of Lake O'Hara (Figure 3-2), by electron microprobe analysis of glass shards. The identity of the tephras in the Lake O'Hara cores was confirmed by comparison of core stratigraphy, as well as petrographic characteristics (Reasoner and Rutter 1988). The Mazama tephra (ca. 6800 yr BP) is generally thicker, averaging about 6 cm, whereas Bridge River tephra (ca. 2350 yr BP) is generally about 1 - 2 cm thick. There is no evidence for erosional unconformities underlying the tephras although minor sediment loading was associated with the deposition of Mazama tephra (Reasoner and Rutter 1988). The occurrence of tephra as discrete, well-defined layers in the cores suggests that there has been little post-depositional mixing or tephra redeposition. Therefore, samples immediately below the tephras should be of similar age throughout the basin. As well, samples taken immediately below the tephra layers will not reflect vegetation changes resulting from tephra deposition in the lake catchment.

SAMPLING AND LABORATORY METHODS FOR POLLEN ANALYSIS

The nineteen cores used in this study were obtained during the winter of 1985 using a percussion corer (Reasoner 1986). The cores are 7.6 cm in diameter with a maximum length of 257 cm (core LOH4b). The frozen cores were returned intact to the laboratory where they were thawed, described and later subsampled for pollen analysis.

Not all cores extended to the lower organic/inorganic contact. Thus, the three intervals could not be sampled in every core. The analyzed sample sets included thirteen samples from directly above the basal contact, sixteen samples from beneath the Mazama tephra, and nineteen samples from beneath the Bridge River tephra. This provided 48 samples in total. These sets are hereafter called the lower, middle, and upper sets respectively. They are from three zones identified by Reasoner and Hickman (1989) in core LOH25 (Figure 3-4). The lower set is equivalent to the base of Zone 2, which has an estimated age of ca. 10,100 - 7,000 yr BP. The middle set is equivalent to the base of Zone 3, estimated to span ca. 7,000 - 3,000 yr BP. The upper set matches the middle of Zone 4, estimated to extend from ca. 3,000 yr BP to present.

In the laboratory, pollen samples were prepared from 1 cm^3 sediment subsamples using standard methods, consisting of a 10% hydrochloric acid (HCl) wash to remove carbonates, 10% NaOH wash to disperse the sample, then wet sieving to remove coarse fragments, followed by heavy liquid treatment with zinc bromide ($ZnBr_2$, sp.gr. 2) to remove clastic material, and hydrofluoric acid (HF) treatment to remove remaining fine clastic material. Finally, organic material other than pollen was removed by acetolysis. The samples were then stained with safranine. The samples were dehydrated in an alcohol series, before being suspended in silicone oil. This procedure followed methods used at the Palaeoenvironmental Laboratory, Department of Anthropology, University of Alberta. At the beginning of the preparation procedure, samples were "spiked" with *Lycopodium* spores (2 tablets/sample; 12,490 grains/tablet) to enable pollen concentration to be calculated, as grains/cm^3, following Stockmarr (1971).

Subsamples of the pollen suspensions in silicone oil were mounted on glass slides. Pollen and spores were identified and counted using a Zeiss Standard 14 microscope at 500X and 1250X. A minimum of 500 identifiable grains of taxa included in the pollen sum were counted for each sample, with a mean grain count of 1141, and range from 534 to 5638 grains. Because *Pinus* pollen occurred in high frequency, a minimum target of 100 grains, excluding *Pinus*, was set for each sample. Pollen and spores were identified using published references, including Kapp (1969), McAndrews et al. (1973), Bassett et al. (1978), Lewis et al. (1983), and reference collections.

Pinus pollen was subdivided based on the presence/absence of distal verrucae (Ting 1966) and assigned to subgenera (*Haploxylon* and *Diploxylon* types). At least 50 Pinus grains were identified to type and tallied for each sample where pollen preservation permitted. These values were used to calculate *Haploxylon/Diploxylon* ratios (shown to the left side of Figure 3-7).

The pollen sum includes all identified arboreal pollen (AP) and nonarboreal pollen (NAP) types, except aquatic taxa. Spores, unidentifiable grains, and unknown pollen taxa are not included in the pollen sum. Small (ca. 10 mm) moss spores were numerous in some samples; these were tallied for a few traverses and the final count calculated with respect to the spike count. Selected taxa, generally those occurring in most samples or those greater than 1% of the pollen sum, are displayed on the pollen percentage diagrams (Figure 3-7). Confidence intervals (95%) (Mosimann 1965; Maher 1972), are included for the main arboreal taxa (*Pinus*, *Picea*, *Abies*, and *Alnus*). Mean percentage values for each group for selected taxa are shown in Table 3-1. Among the other taxa encountered were *Cornus canadensis*, Ranunculaceae including *Anemone* and *Delphinium*, *Impatiens*, *Epilobium*, Caryophyllaceae, Saxifragaceae including *Saxifraga aizoides*, Liliaceae including *Allium*, Polygonaceae including *Oxyria digyna*, *Eriogonum*, *Rumex*, and *Polygonum*, Cruciferae, *Sedum*, *Urtica*, *Polemonium*, Umbelliferae including *Heracleum* and *Angelica*, *Plantago*, *Galium*, Primulaceae including *Dodecatheon* and *Androsace*, *Valeriana*, *Campanula*, *Parnassia*, *Pedicularis*, and Labiatae. These were included in the pollen sum but are not displayed on the diagrams because few grains were encountered. Raw pollen data are on file at the Palynology Laboratory, Archaeological Survey, Provincial Museum of Alberta.

To calculate pollen influx values (grains/cm^2/yr), sedimentation rates (cm/yr) were derived by assuming constant sedimentation between dated layers (see Reasoner and Rutter 1988: Table 3-1). Because the samples were taken at or next to dated layers (i.e., the points at which the inferred rates change), the sedimentation rate applied to each set of samples was that for the interval immediately above the sampled layer.

Charcoal fragments and *Pediastrum* (colonial algae) colonies were also tallied (Figure 3-7). In samples where charcoal fragments were numerous, counts were made for a few traverses and the final values calculated with respect to the spike count.

Because sample-to-sample variability was one subject of this investigation, great efforts were made to avoid sources of variance that might be introduced during data acquisition. Therefore, during counting, traverses covering complete slides were tallied to avoid possible bias from differential sorting of grains on the slide, as reported, for example, by Brookes and Thomas (1968). To eliminate any operator bias, the three samples from core LOH25, which were reported by Reasoner and Hickman (1989), were recounted for this study (Table 3-1). Percentages of major taxa (>1% of the pollen sum) in the two counts were in reasonably close agreement (Table 3-1). However, we encountered greater variety and numbers of minor taxa (<1%) in this study. We examined a larger sample from each interval (Table 3-1), and therefore the probability of encountering rare taxa increased.

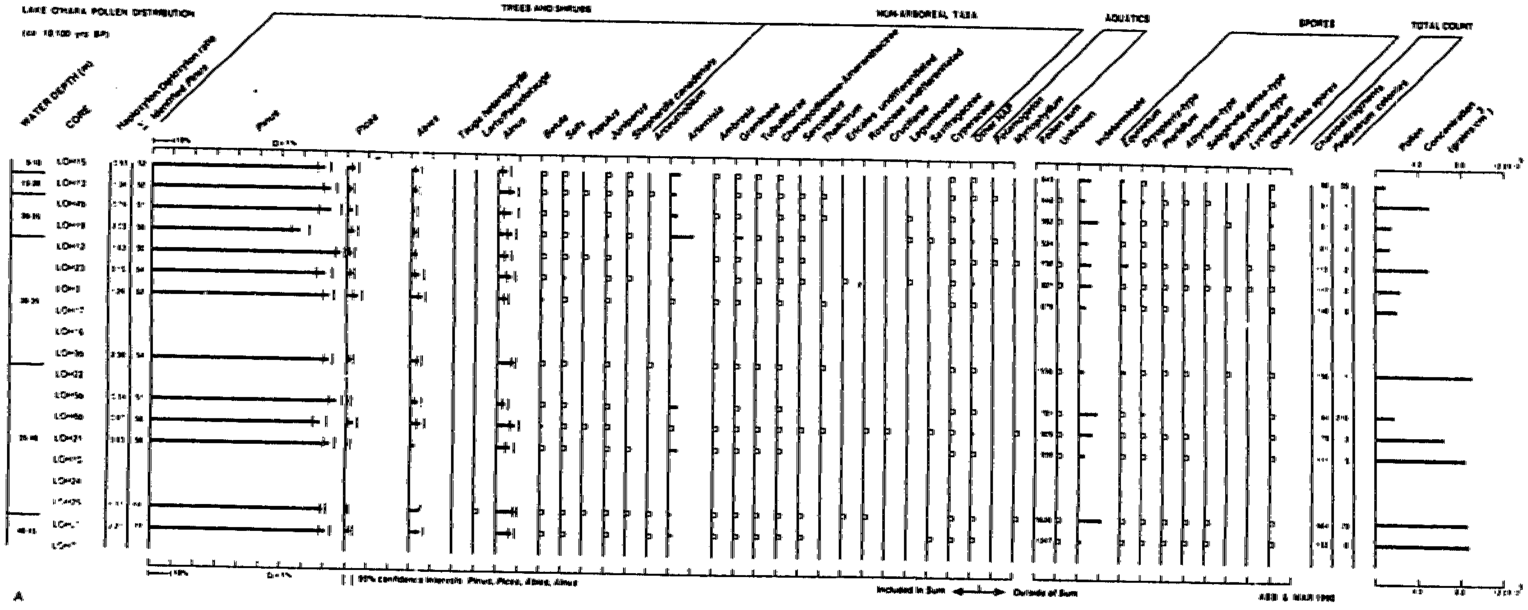

Figure 3-7a: Pollen percentage diagram of selected major taxa from the upper level in Lake O'Hara: Lower sample set (Basal contact, ca. 10,100 yr BP). Samples are grouped according to water depth at the sample site into 5 m classes, shown in the far left column.

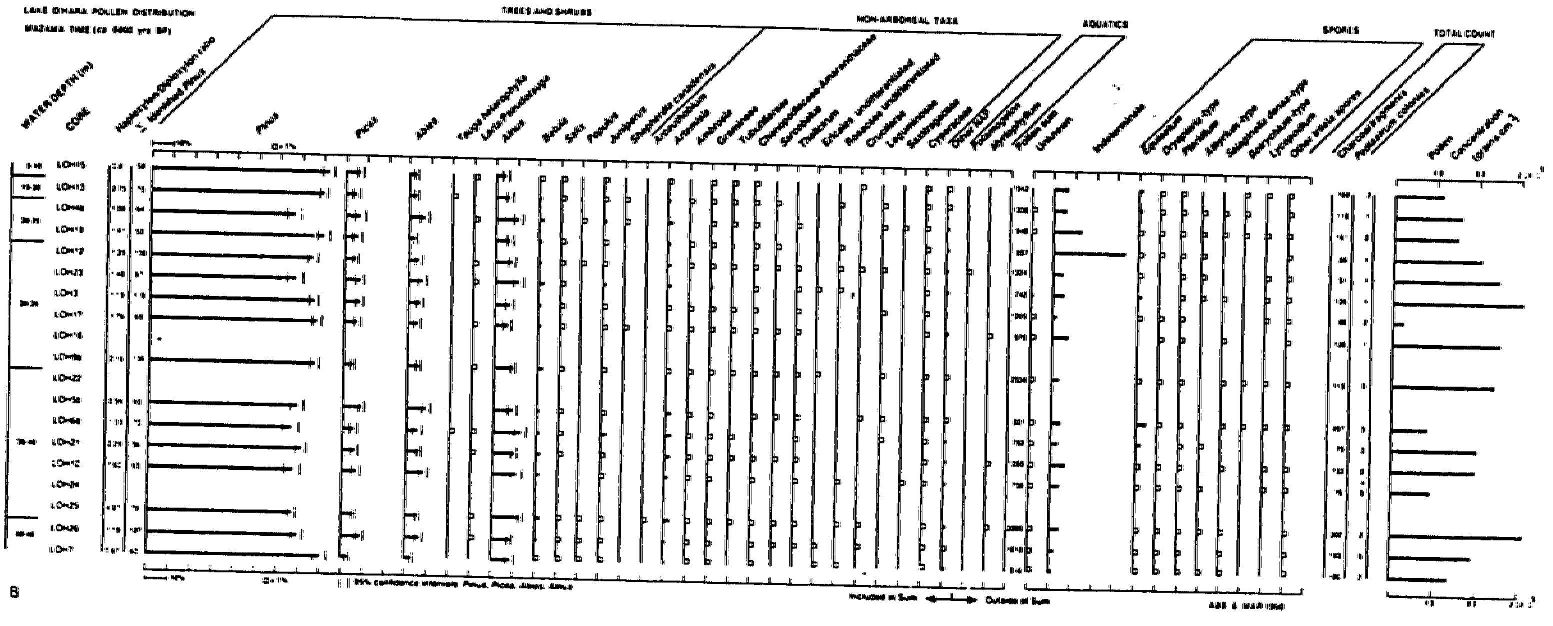

Figure 3-7b: Pollen percentage diagram of selected major taxa from the middle level in Lake O'Hara: Middle sample set (Mazama time, ca. 6800 yr BP). Samples are grouped according to water depth at the sample site into 5 m classes, shown in the far left column.

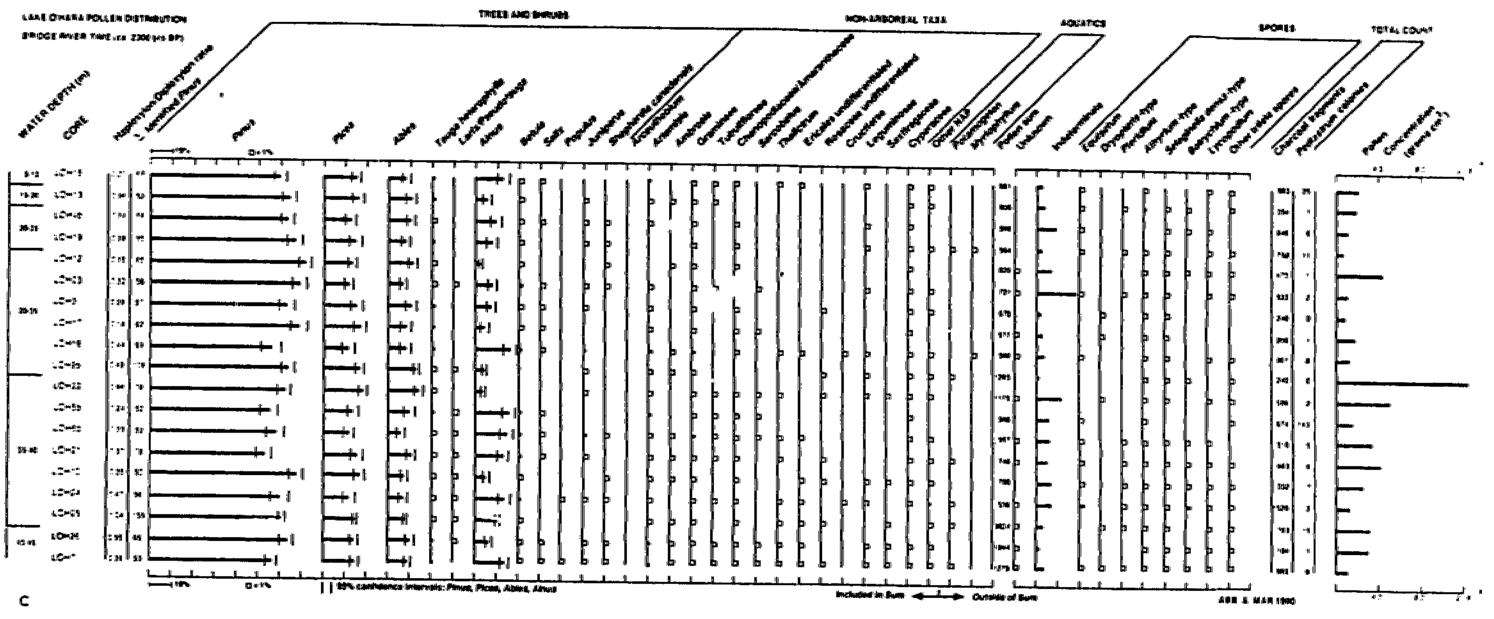

Figure 3-7c: Pollen percentage diagram of selected major taxa from the upper level in Lake O'Hara: Upper sample set (Bridge River time, ca. 2350 yr BP). Samples are grouped according to water depth at the sample site into 5 m classes, shown in the far left column.

Table 3-1: Descriptive Data and Comparison of Individual Pollen Taxa for Three Lake-Wide Sample Sets and Equivalent Pollen Zones from Core LOH25, Lake O'Hara

Taxon	Set	Mean	SD	F^a	Sig.	Zone	Mean	SD	F^b	Sig	LOH25 1988	LOH25 1992
Pinus	Upper	61.81	5.03			4	63.40	4.26	0.87	0.36	60.5	59.0
	Middle	73.05	5.75	55.26	0.00	3	65.46	8.76	7.18	0.01	70.4	67.0
	Lower	80.70	4.34			2	81.35	2.92	0.09	0.76	85.7	80.3
Picea	Upper	14.33	2.04			4	15.63	3.43	1.80	0.19	17.5	14.8
	Middle	6.09	1.48	13.16	0.00	3	10.36	5.10	10.13	0.00	5.2	5.4
	Lower	2.59	1.23			2	2.25	1.22	4.31	0.05	1.0	1.3
Abies	Upper	9.28	2.13			4	4.86	0.61	52.49	0.00	5.4	7.9
	Middle	5.64	1.79	40.42	0.00	3	4.94	1.31	1.15	0.30	5.9	5.8
	Lower	3.59	1.29			2	4.16	0.46	0.92	0.35	3.9	5.1
Tsuga	Upper	1.00	0.38			4	1.24	0.67	1.55	0.22	0.9	1.5
	Middle	0.01	0.04	96.18	0.00	3	0.10	0.16	4.28	0.05	0.0	0.0
	Lower	0.00	0.00			2	0.06	0.09	7.70	0.01	0.2	0.0
Salix	Upper	0.11	0.11			4	0.18	0.20	1.27	0.27	0.0	0.1
	Middle	0.17	0.15	8.22	0.00	3	0.25	0.19	1.51	0.23	0.5	0.1
	Lower	0.43	0.37			2	0.10	0.07	3.83	0.07	0.1	0.1
Betula	Upper	0.70	0.50			4	0.59	0.22	0.53	0.47	0.6	0.7
	Middle	1.57	0.56	16.52	0.00	3	1.33	0.51	1.18	0.29	2.2	2.3
	Lower	0.79	0.29			2	0.55	0.40	1.90	0.19	0.0	0.7
Alnus	Upper	9.03	4.42			4	10.29	2.23	0.91	0.35	11.7	10.4
	Middle	8.89	2.39	3.45	0.04	3	12.89	3.90	10.68	0.00	10.8	13.2
	Lower	6.20	1.71			2	8.49	2.13	5.69	0.03	5.2	8.9
Juniperus	Upper	0.46	0.46			4	0.00	0.00	12.71	0.00	0.0	0.5
	Middle	0.76	0.64	1.63	0.21	3	0.24	0.24	6.01	0.02	0.2	0.5
	Lower	0.57	0.35			2	0.34	0.24	1.74	0.21	0.4	0.3
Artemisia	Upper	0.71	0.50			4	0.76	0.38	0.09	0.77	0.3	1.0
	Middle	1.64	0.69	9.58	0.00	3	1.40	0.72	0.72	0.41	2.1	2.4
	Lower	3.12	2.80			2	1.67	0.29	1.20	0.29	2.1	1.8
Gramineae	Upper	0.55	0.42			4	0.84	0.36	4.10	0.05	1.0	0.7
	Middle	0.58	0.28	0.14	0.87	3	0.76	0.35	2.07	0.16	0.7	0.8
	Lower	0.65	0.84			2	0.66	0.37	0.00	0.97	0.5	0.5

Table shows percentage pollen values and statistics from output produced by SPSS^x MANOVA (SPSS Inc. 1988). Samples: Upper set (19 samples), Middle (16), Lower (13), LOH25 Zone 4 (13), Zone 3 (10), Zone 2 (5).

F^a: Comparison between all three sample sets for each taxon. F^b: Comparison between each individual taxon in each sample set and its equivalent zone from LOH25.

Comparison using univariate F-tests: H_0 (Pollen data do not vary significantly between the groups) rejected when calculated $F >$ critical value $F_{0.01}$ (t_1, t_2).

Degrees of freedom and critical values as follows: Upper, middle and lower sets, (2,45) DoF, F = 5.18; upper set with Zone 4, (1,30) DoF, F = 7.56; middle set with Zone 3, (1,24) DoF, F = 7.82; lower set with Zone 2, (1,16) DoF, F = 8.53.

LOH25 and Lake: Percentage data for samples equivalent to upper, middle and lower set levels from core LOH25 (Reasoner 1988, Appendix 2) and same samples recounted for this study (Lake). Pollen sums are 686, 878, 1283 grains and 3834, 3095, 5638 grains for upper, middle and lower samples respectively (see also Figures 3-4 and 3-7).

METHODS FOR NUMERICAL ANALYSIS OF POLLEN DATA

The pollen data were subjected to multivariate statistical analysis using the SPSS^x version 3.0 statistical package (SPSS Inc. 1988) running on an Amdahl 5870 computer at the University of Alberta. The main statistical techniques used to examine the data were multivariate discriminant function analysis (DFA) and multivariate analysis of variance (MANOVA). In addition, regression analysis was used to examine relationships between various parameters and water depth. Ten taxa were included in the statistical analyses: *Pinus, Picea, Abies, Tsuga, Salix, Betula, Alnus, Juniperus, Artemisia,* and Gramineae (Table 3-1). These taxa were selected because in total they comprised between 96.08% (LOH24, upper set) and 99.55% (LOH3, lower set) of the pollen sum. In addition, they are the only taxa that comprised at least 1% of the pollen sum in at least one sample. The remaining 51 taxa comprised at most about 4% of the pollen sum in total in any sample. Moreover, many minor taxa occurred in only one or a few samples. For these reasons, they were excluded from the statistical analyses. Each technique was applied to pollen percentage, concentration, and influx data.

Two multivariate statistical techniques, DFA and MANOVA, were chosen because both examine the underlying variability in the data sets. First, if differential pollen deposition has occurred, a high degree of variation within each sample set might be expected. If differential pollen deposition has not occurred, variability between the sample sets might be expected to be considerably greater than variability within any set. MANOVA was used to investigate this. Second, because the three sample sets represent three different pollen assemblage zones, they should also be readily distinguishable by their constituent pollen taxa. DFA was used to investigate this. Because the pollen assemblage zones were defined based on fluctuations in the major pollen taxa, this analysis also provides a means of investigating the adequacy of visual zonation, which is inherently subjective.

DFA takes predefined groups, here the three sample sets, and investigates which variables, here pollen taxa, best distinguish between them. SPSS^x DISCRIMINANT permits a choice of method for determining discriminant functions; in this study, WILKS method was chosen. All ten major pollen taxa were included at the beginning of the analysis. The pollen taxa were selected by SPSS^x in a step-wise procedure that minimizes WILKS Lambda, a measure of group discrimination. The three sample sets were used as predefined groups, and two discriminant functions were sufficient to distinguish between them. The results of this analysis are summarized in Tables 3 and 4.

MANOVA is an explicit investigation of variability and tests whether predefined groups, here the three sample sets, are significantly different in terms of their means and variances. All ten pollen taxa were included in this analysis. Hotelling's T^2 and its associated F statistic are used to test group differences (Tables 2 and 4).

Both MANOVA and DFA assume multivariate normality in the data. Individual taxa were tested for least one set. No taxon shows a consistent non-normal distribution across all data sets or for all

Table 3-2: Discriminant Function Analysis of Three Sets of Pollen Assemblages from Lake O'Hara

Percentage data

	Standardized canonical discriminant functions				
	Pinus	*Picea*	*Tsuga*	*Betula*	*Alnus*
Function 1	0.300	1.126	0.371	-0.181	0.89339
Function 2	1.167	0.977	-0.049	1.148	0.92398

	Eigenvalue	Percent of variance	Canonical correlation
Function 1	16.5575	92.42	0.9711
Function 2	1.3586	7.58	0.7590

Concentration data

	Standardized canonical discriminant functions								
	Pinus	*Picea*	*Abies*	*Tsuga*	*Betula*	*Alnus*	*Juniperus*	*Artemisia*	Gramineae
Function 1	0.813	0.169	0.366	-1.157	0.274	-0.893	0.267	0.383	-0.091
Function 2	-0.433	3.171	-2.273	-0.543	0.808	1.146	0.355	-0.703	-0.980

	Eigenvalue	Percent of variance	Canonical correlation
Function 1	2.8112	65.49	0.8588
Function 2	1.4811	34.51	0.7726

Influx data

	Standardized canonical discriminant functions				
	Pinus	*Tsuga*	*Betula*	*Juniperus*	*Artemisia*
Function 1	0.846	-0.880	0.428	0.379	-0.581
Function 2	0.317	0.603	1.039	0.357	-1.232

	Eigenvalue	Percent of variance	Canonical correlation
Function 1	2.8911	77.82	0.8620
Function 2	0.8239	22.18	0.6721

Data taken from output produced by SPSS^x DISCRIMINANT (SPSS Inc. 1988).

Table 3-3: Classification Produced by Discriminant Function Analysis of Three Sample Sets from Lake O'Hara

Percentage data

Actual set	Number of cases	Predicted set Upper	Predicted set Middle	Predicted set Lower
Upper	19	19	0	0
		100%	0%	0%
Middle	16	0	15	1
		0%	93.8%	6.3%
Lower	13	0	1	12
		0%	7.7%	92.3%

Percent of grouped cases correctly classified = 95.83%

Concentration data

Actual set	Number of cases	Predicted set Upper	Predicted set Middle	Predicted set Lower
Upper	19	19	0	0
		100%	0%	0%
Middle	16	1	14	1
		6.3%	87.5%	6.3%
Lower	13	0	1	12
		0%	7.7%	92.3%

Percent of grouped cases correctly classified = 93.75%

Influx data

Actual set	Number of cases	Predicted set Upper	Predicted set Middle	Predicted set Lower
Upper	19	15	0	4
		78.9%	0%	21.1%
Middle	16	0	13	3
		0%	81.3%	18.8%
Lower	13	0	0	13
		0%	0%	100%

Percent of grouped cases correctly classified = 85.42%

Data taken from output produced by SPSS^{X} DISCRIMINANT (SPSS Inc. 1988)

Table 3-4: Multivariate Analysis of Variance of Pollen Assemblages from Lake O'Hara

Hotelling's T^2	Value	F	Hypothetical DoF	Error DoF	Significance of F
A. Using all ten taxa					
Percentages	20.625	36.094	20	70	0.000
Concentrations	4.333	7.582	20	70	0.000
Influxes	4.188	7.329	20	70	0.000
B. Using only taxa that meet criteria for normal distribution					
Percentages	15.142	77.604	8	82	0.000
Concentrations	1.025	11.021	4	86	0.000
Influxes	1.902	13.312	6	84	0.000

Calculated F values for tests on three sample sets are approximate. All values rounded to 3 dp. Values produced by SPSS^X MANOVA (SPSS Inc. 1988).

Test: H_0 (Pollen data do not vary significantly between groups) rejected if calculated $F > F_a (t_1, t_2)$, when a = 0.01.

A. Critical value for percentage, concentration and influx data and three sample sets using all ten taxa, F_a » 2.12.

B. Critical values using only taxa meeting criteria for normal distribution: Percentages (using *Picea*, *Abies*, *Betula*, *Alnus*), F_a » 2.74; Concentrations (using *Alnus*, *Artemisia*), F_a » 3.57; Influx data (using *Pinus*, *Picea*, Gramineae), F_a » 3.04.

measures. This behaviour is expected, particularly for taxa present in small percentage with an irregular pattern of occurrence. For example, for *Artemisia* in the lower set, all samples have values <5%, except two with 6.7% and 10.7% (Figure 3-7a). Departures from normality are prevalent in the concentration and influx data because additional variability is being introduced in the calculations. Both MANOVA and particularly DFA are reasonably robust to departures from normality (Tabachnick and Fidell 1983). To check the influence of non-normality, the MANOVA analyses were re-run including only taxa that met criteria for normality in each data set (Table 3-4). Although numerical values differed between the analyses, the same inferences can be drawn. Therefore we concluded that departures from normality in the data did not affect the validity of our results.

ANALYSIS AND DISCUSSION

Investigation of differential pollen deposition

The most striking feature of these data is the general lake-wide consistency in the proportions or relative percentages of major components of the pollen assemblages at each level (Figures 3-7 and 3-8). This is particularly the case for the major arboreal taxa, such as *Pinus*, and in particular *Picea* and *Abies*. There is some variability within each sample set, but the comparatively small standard deviations, especially for spruce and fir pollen percentages (Table 3-1), suggest that the sample sets are homogeneous. This lake-wide consistency suggests that differential pollen deposition has not occurred. These data show that, for the major pollen taxa, similar pollen assemblages can be found at the same chronostratigraphic level despite core location within the lake basin. Therefore, interpretations about vegetation history based on evidence from major arboreal taxa in a single core are probably reliable. These data therefore support the assumption of representativeness of pollen percentage data from a single core for Lake O'Hara and lakes similar to it.

If differential pollen deposition has occurred, there should be areal variability in pollen assemblages and trends of pollen percentage, concentration and influx data associated with bathymetry. We examined areal distributions in percentages of each major taxon to see if there had been selective deposition of taxa in different areas of the lake. We mapped all ten major taxa individually for each sample set for each type of data (percentage, concentration, influx). In each case, we could not identify consistent lake-wide patterns (e.g., deeper water vs. shallow water sites, western vs. eastern end of lake) in these distributions that might reflect differential pollen deposition. For the percentage data, these distributions, summarized in Figure 3-8, emphasize more strongly than the standard pollen diagrams that there are substantial differences between the sample sets as well as basin-wide consistency in the proportions of the different taxa within each set for the percentage data.

The mapped distributions of pollen concentration and influx data for individual taxa also lacked lake-wide patterns. The data exhibit considerable variability within each sample set (Figures 3-9 and 3-

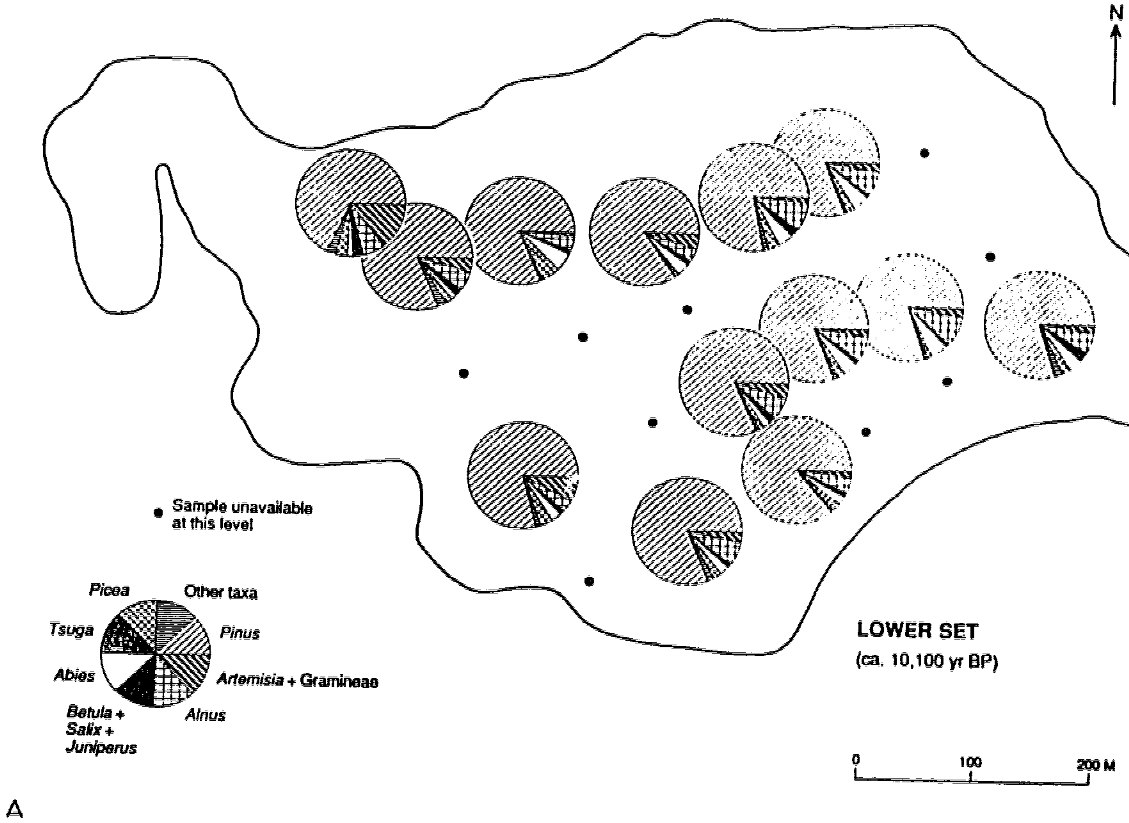

Figure 3-8a: Mapped pollen assemblages at three discrete intervals from nineteen cores from Lake O'Hara: a. Lower sample set (Basal contact, ca. 10,100 yr BP), b. Middle sample set (Mazama time, ca. 6800 yr BP), c. Upper sample set (Bridge River time, ca. 2350 yr BP). These summaries show distributions for pollen percentage data. "Other" taxa include those, apart from major taxa, which comprise the pollen sum; pie charts are therefore directly comparable because the totals on each are 100% and represent the entire pollen

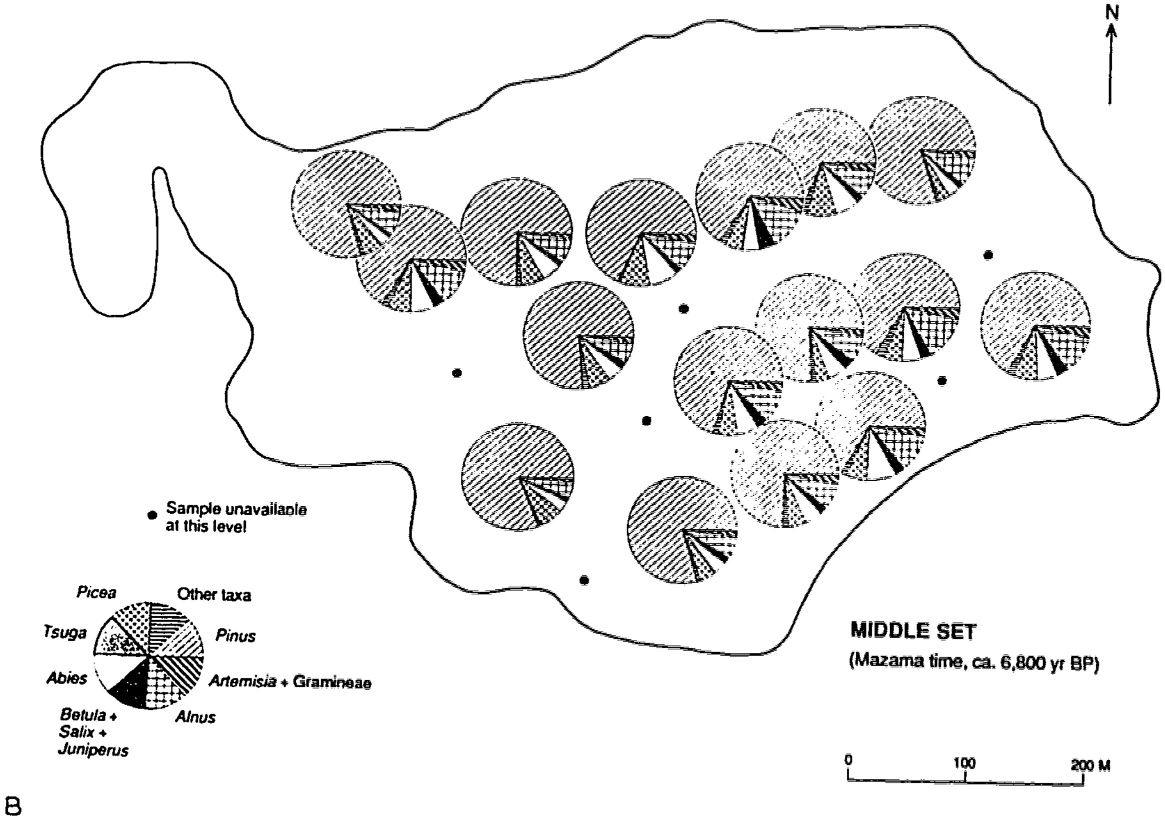
N
Sample unavailable at this level
Picea
Other taxa
Tsuga
Pinus
Abies
Artemisia + Gramineae
Betula + Salix + Juniperus
Alnus
MIDDLE SET
(Mazama time, ca. 6,800 yr BP)
0
100
200 M
B

Figure 3-8b

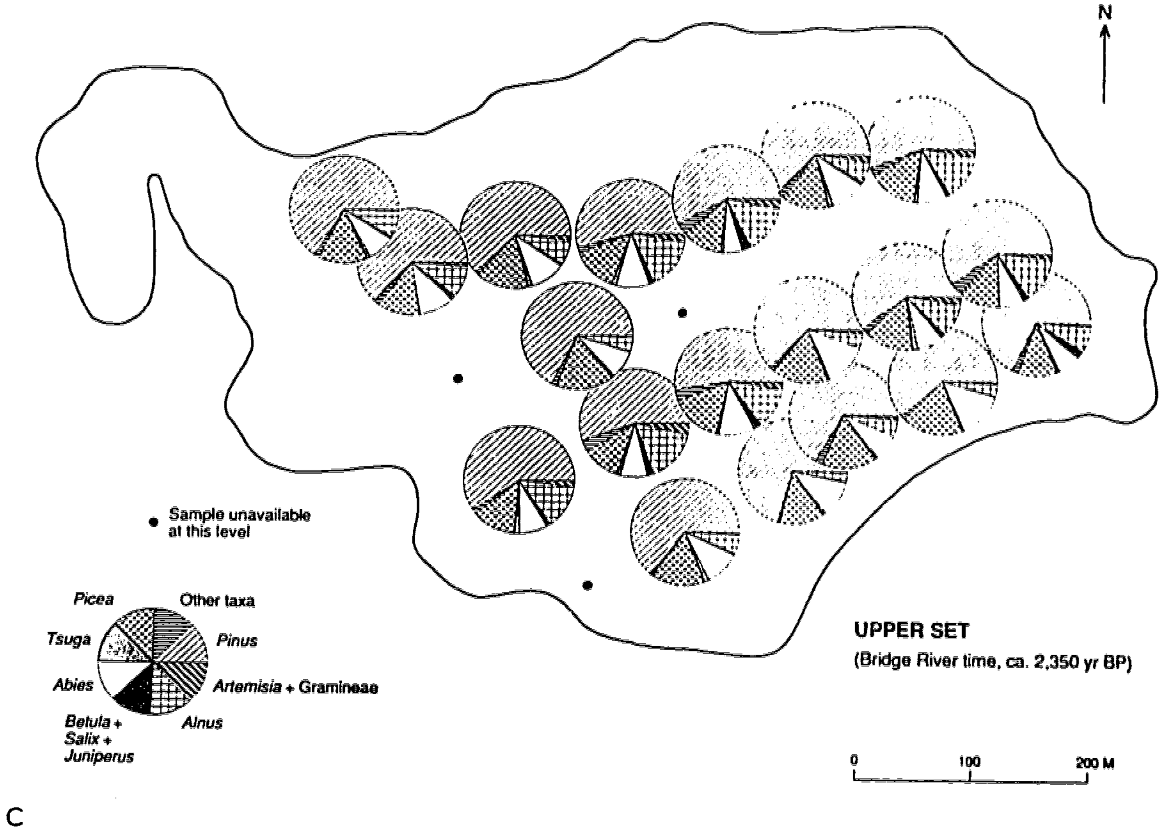
N
Sample unavailable at this level
Picea
Other taxa
Tsuga
Pinus
Abies
Artemisia + Gramineae
Betula + Salix + Juniperus
Alnus
UPPER SET
(Bridge River time, ca. 2,350 yr BP)
0
100
200 M
C

Figure 3-8c

10). These data do not show clear trends for individual taxa or totals in relation to water depth. This lack of lake-wide consistency forms a marked contrast to the percentage data. Unlike the percentage measurements, therefore, we concluded that for these quantitative measures, a single core does not provide representative results for the lake-wide record at any level.

Individual taxa might be expected to show a relationship with water depth, depending whether they are preferentially deposited in shallow or deep water, if they have been affected by differential pollen deposition. To test this, we plotted percentage, concentration and influx values against water depth for all ten major taxa. In each case, the data points were widely scattered. Linear regression showed that in most cases the relationship was positive, i.e., values increased with increased water depth. Often the slope of the regression line was gentle or almost horizontal and most R^2 values were less than 0.1. The greatest R^2 values occurred in the percentage, concentration and influx data from the lower set. Because the relationships are weak, particularly for the percentage values, these data do not show clear evidence for differential deposition of pollen taxa in different areas of the lake.

Our results therefore contrast with some previous research. Several studies (see Faegri et al. 1989; Davis et al. 1971; Davis and Brubaker 1973) have suggested that because bisaccate pollen floats, it should be concentrated in the shallowest part of the lake, around the shoreline. For example, Davis et al. (1971) found that pine pollen occurred in higher frequency (i.e., greater percentage) in shallower than deeper water samples at Frains Lake, which has a maximum water depth of 9.5 m. In contrast, R.B. Davis et al. (1969) found no consistent trend of pine pollen accumulation in the lake sediments they examined in Wisconsin. However, around Frains Lake, the regional vegetation was deciduous and pine pollen, derived from long distance transport, comprised at most 2 - 3% of the pollen sum.

In the Lake O'Hara samples, only weak relationships were identified between total *Pinus* percentages and water depth (R^2 values of 0.04, 0.19, and 0.07 for upper, middle and lower sets respectively), although for the upper and middle sets the relationship was negative. Similarly, *Haploxylon/Diploxylon* ratios and percentages of other bisaccate taxa, specifically *Abies* and *Picea*, did not show strong relationships with water depth when investigated by linear regression. The R^2 values are slightly higher for *Pinus* and *Picea* concentration and influx values in the lower set (between 0.27 and 0.37). The clearest positive relationship occurred between *Abies* percentage, concentration and influx values (R^2 of 0.33, 0.41 and 0.33 respectively) and water depth in the lower sample set (Figures 3-7a and 3-10). That is, greater percentages and influx of *Abies* pollen occur in cores from deeper water, opposite to the pattern found by Davis et al. (1971) for *Pinus* pollen. However, none of these relationships provide particularly strong evidence for differential pollen deposition. At Lake O'Hara, *Pinus* pollen forms most of the pollen sum (Figure 3-7). Its input in the spring is likely to be so massive that it probably overwhelms any tendency to concentrate in shallow water sediments.

The consistency and homogeneity within sample sets exhibited by the major taxa do not hold

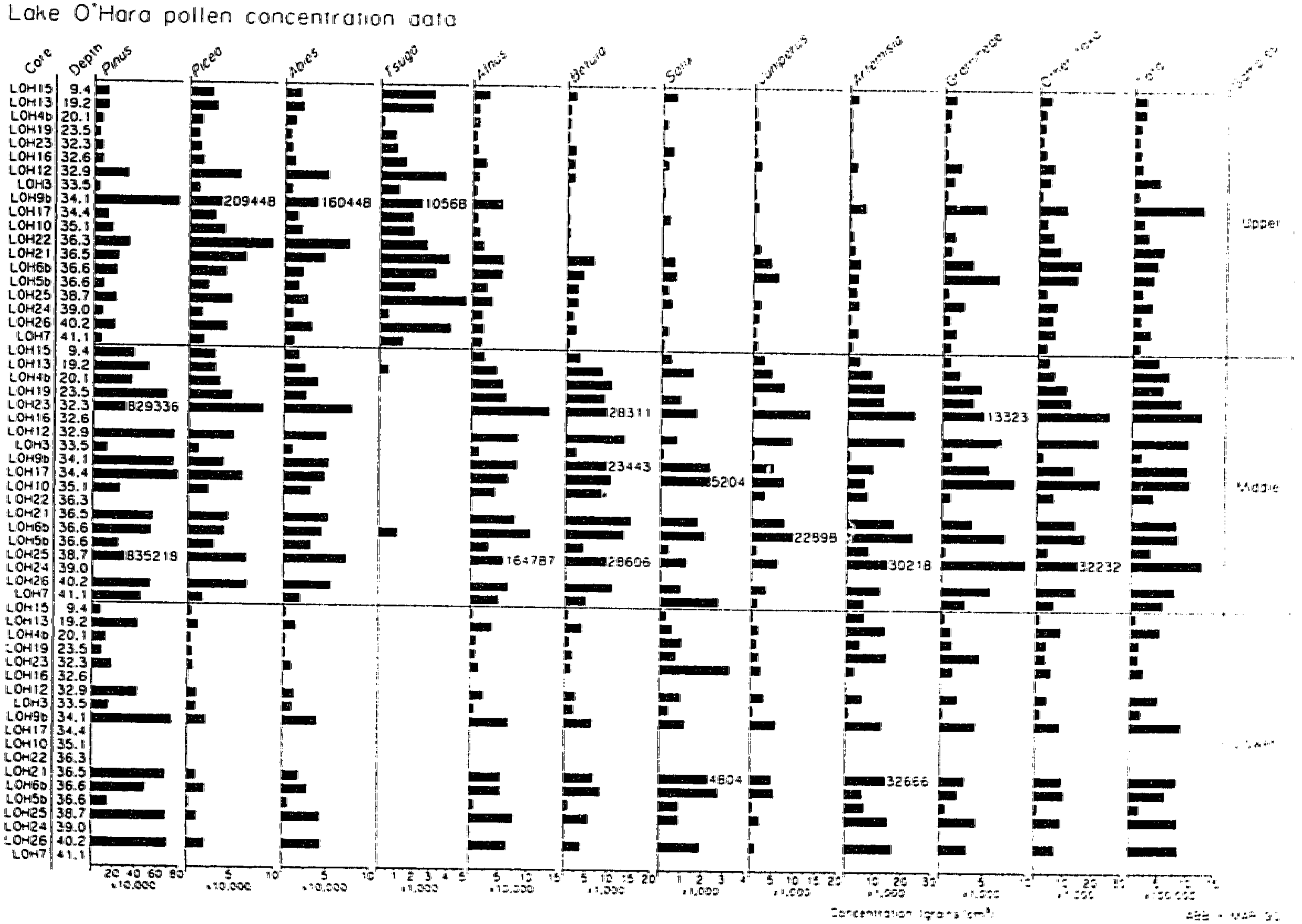

Figure 3-9: Pollen concentration diagram of selected major taxa from the three levels in Lake O'Hara. Samples are grouped into three sample sets: Upper (Bridge River time, ca. 2350 yr BP), Middle (Mazama time, ca. 6800 yr BP), and Lower (Basal contact, ca. 10,100 yr BP). Samples are plotted according to the water depth at the sample site, shown to the left of the diagram. Total pollen concentration is based on the pollen sum.

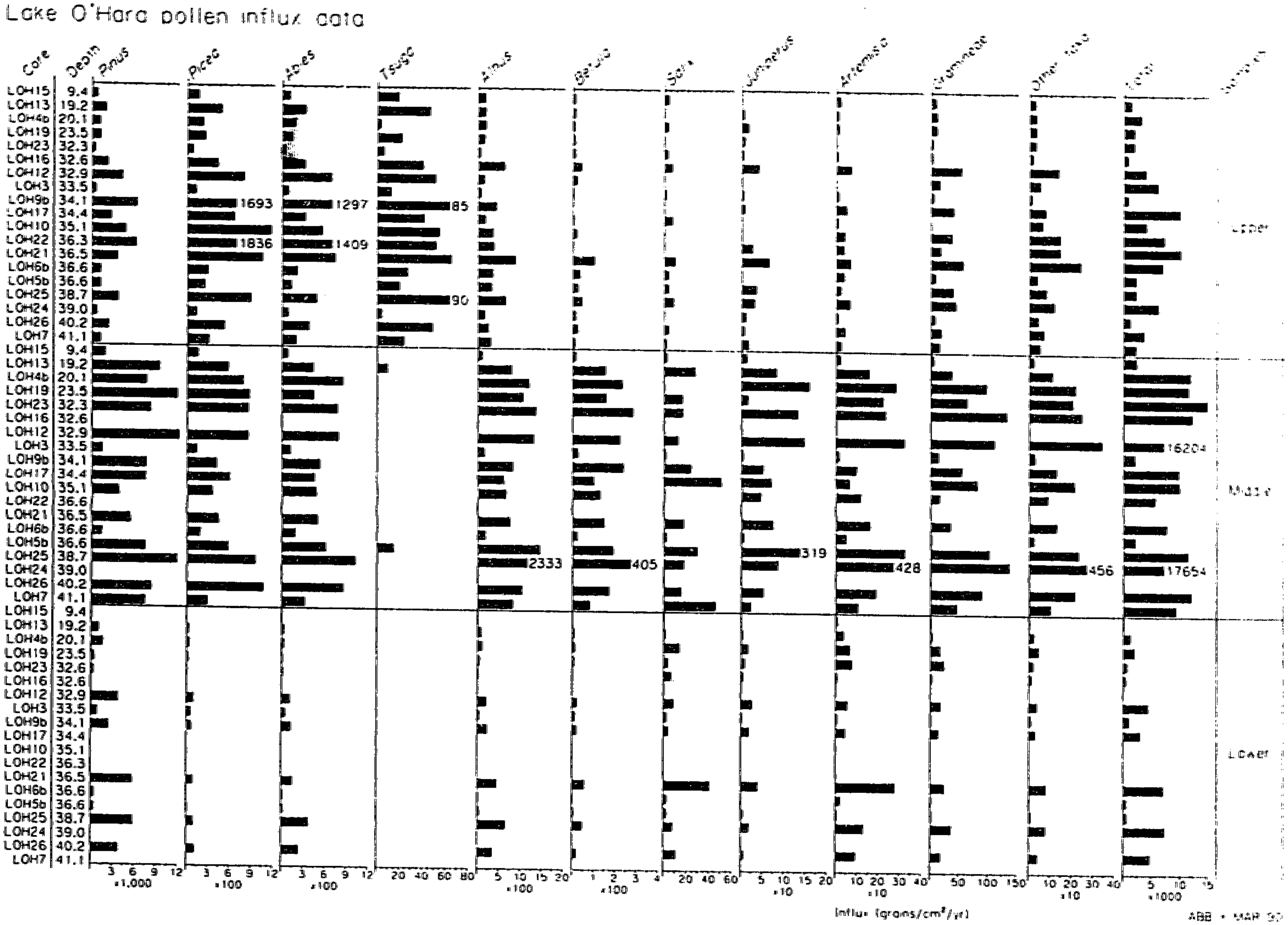

Figure 3-10: Pollen influx diagram of selected major taxa from the three levels in Lake O'Hara. Samples are grouped into three sample sets: Upper (Bridge River time, ca. 2350 yr BP), Middle (Mazama time, ca. 6800 yr BP), and Lower (Basal contact, ca. 10,100 yr BP). Samples are plotted according to the water depth at the sample site, shown to the left of the diagram. Total pollen influx is based on the pollen sum.

true for the minor constituents of the pollen assemblage (Figure 3-7). Minor components (those taxa that are generally less than 1% of the pollen sum), in this instance mainly NAP and spores, show a more variable pattern of occurrence within each set (Figure 3-7). In particular, many minor taxa (e.g., *Thalictrum*) do not occur in every sample. Several minor taxa of interpretive significance in core LOH25 (e.g., *Arceuthobium, Ephedra,* and *Oxyria digyna*) have inconsistent and erratic distributions (Figures 3-4 and 3-7). Some general trends are identifiable, such as greater variety and abundance of minor taxa in the upper and middle sets compared to the lower set. Despite this, we could not discern consistent patterns in the occurrence of minor taxa within sample sets that distinguish them. Therefore, we concluded that inferences about vegetation history based on the occurrence or distribution of minor components (<1%) of the pollen assemblages are undependable. Further, we concluded that minor taxa are unreliable for zonation or for characterizing pollen zones. This suggests that inferences based on absence or disappearance of a minor taxon in the record from a single core are doubtful.

Investigation of pollen and sediment focussing

If pollen focussing has occurred, pollen percentages will remain similar despite core location but we hypothesized that total pollen concentration and pollen influx will vary across the lake basin. This could arise when the entire assemblage had been focussed to deeper parts of the basin. We tested three specific hypotheses to see if the sample sets yielded evidence for pollen focussing.

First, we hypothesized that if pollen has been consistently accumulated as one component of the sediments, there should be a relationship between sedimentation rate (mm/year) and total pollen concentration (grains/cm^3). If the total amount of pollen input to the lake is relatively constant at each interval, and if no focussing has occurred, then higher sedimentation rates should be associated with lower pollen concentrations (a dilution effect). Alternatively, if pollen has been accumulated simply as one component of the sediments, then we should expect a positive relationship between pollen concentration and sedimentation rate. The data provide no evidence for these relationships. Overall, pollen concentrations tend to be greater in the middle sample set but the values show considerable scatter within each sample set (Figures 3-9 and 3-11).

Second, if pollen was deposited in the lake but subsequently moved by resuspension and redeposition towards the deeper areas of the lake, then we hypothesized that greater total pollen concentrations might be expected in cores from deeper water. Thus total pollen concentration should be positively correlated with water depth. If, however, substantial sediment mixing and homogenization had occurred during deposition, then pollen concentrations might be uniform across the basin. However, the data in each group show a wide scatter (Figures 3-9 and 3-12). The strongest relationship occurs in the lower sample set but even here individual values vary substantially from the overall trend given by the regression line (Figure 3-12). For instance, for samples from 30 m - 40 m water depth in

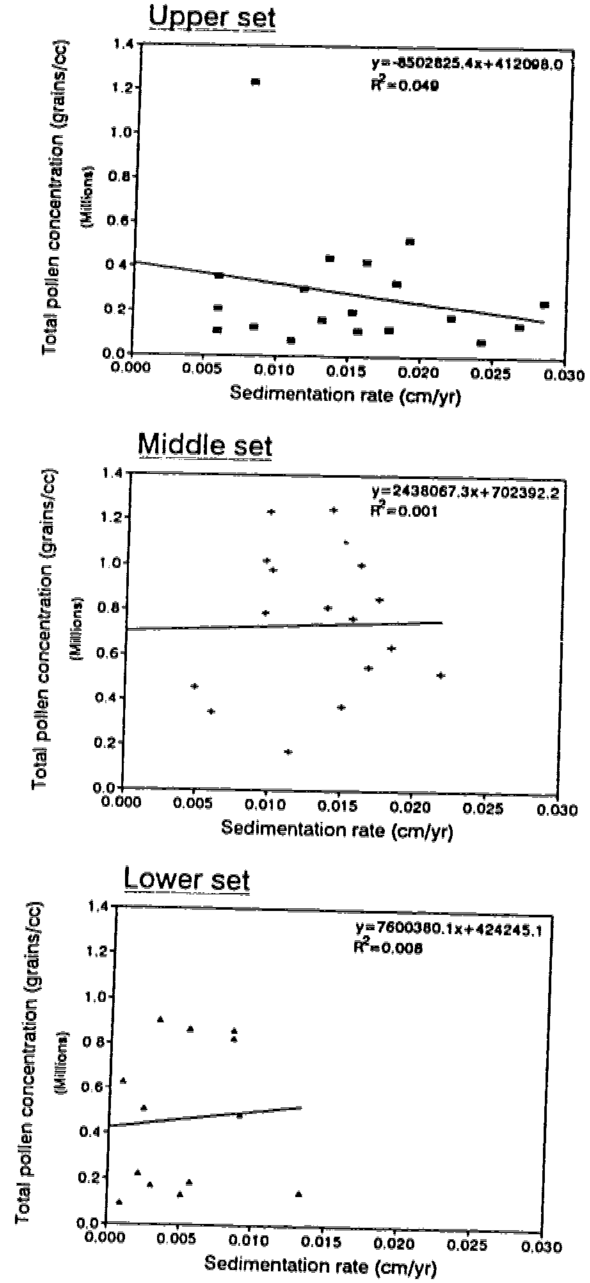

Figure 3-11: Total pollen concentration plotted against sedimentation rate for three sample sets from Lake O'Hara. Total pollen concentration is based on the pollen sum.

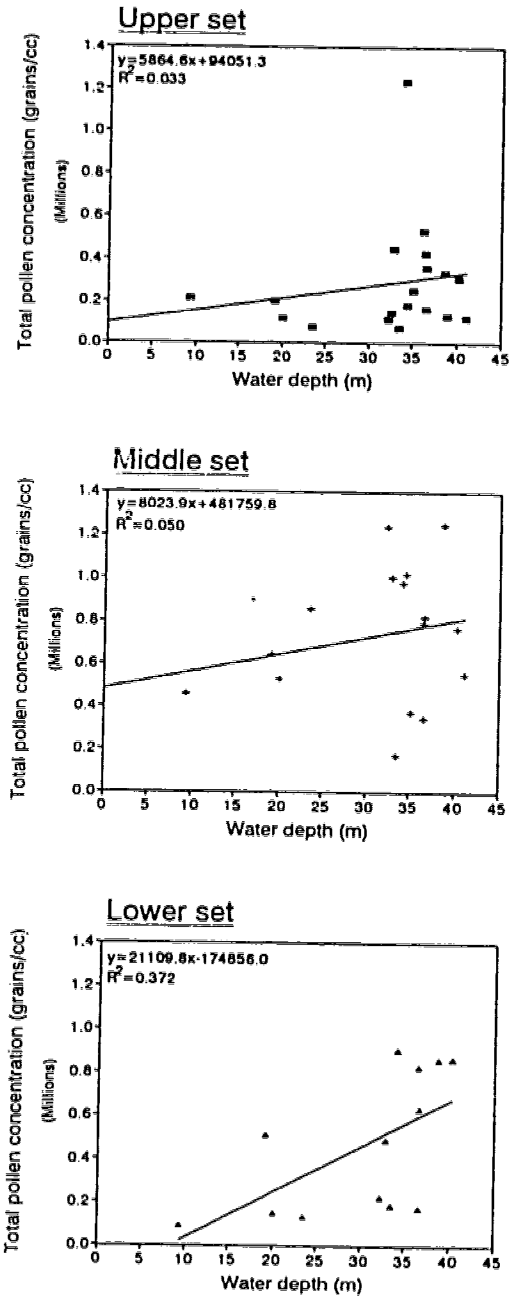

Figure 3-12: Total pollen concentration plotted against water depth for three sample sets from Lake O'Hara. Total pollen concentration is based on the pollen sum.

the lower set (Figure 3-9), total pollen concentration varies between 222,919 grains/cm^3 (LOH23) and 903,927 grains/cm^3 (LOH9b). In contrast, at Mirror Lake, Davis et al. (1984) found that pollen concentration generally was similar in sediments of similar age throughout the basin suggesting the sediment had been homogenized before final deposition.

Third, if pollen focussing has occurred, greater total pollen influx should be expected at sites from greater water depth. For example, Burden et al. (1986) found greater total influx values in samples from deeper water (ca. 9 m depth) than those from shallow water at Gignac Lake, Ontario. Davis and Ford (1982) considered greater pollen influx values in cores from deeper water areas as supporting evidence for sediment focussing at Mirror Lake. In the Lake O'Hara case, total pollen influx values also display a wide scatter (Figures 3-10 and 3-13). Total influx values are generally least in the lower sample set (average value 2542 grains/cm^3/yr). This set shows only a weak relationship with water depth (R^2 of 0.27), however, the relationship is more pronounced than either other set (Figure 3-13). Although the greatest influx values in each set do occur in samples from sites in deeper water at 30 - 40 m depth (e.g., LOH21 and LOH25), other deep water sites (e.g., LOH5b and LOH6b) have very low influx values (Figure 3-10). The average influx value is lower in the upper sample set (3911 grains/cm^3/yr) than in the middle set (9757 grains/cm^3/yr). Davis et al. (1984) suggested that lower Late Holocene influx values at Mirror Lake were associated with decreased intensity of sediment focussing. However, the wide scatter of values within each set from Lake O'Hara (Figures 3-10 and 3-13) do not support this explanation.

We concluded, therefore, that these data from Lake O'Hara do not show strong unequivocal evidence for pollen focussing. We suspected that this might relate to sediment accumulation within the lake basin. Accordingly, we investigated the pattern of sediment accumulation in more detail.

If significant sediment focussing has occurred, then sedimentation rates might be expected to vary with water depth, with greater sedimentation rates in deeper parts of the basin. The data do not show this pattern but instead show a scatter within each sample set (Figure 3-14) without strongly marked trends. Linear regression suggests that there is almost no relationship between sedimentation rate and water depth (Figure 3-14). Overall, sedimentation rates are generally lower for the lower sample set and highest in the upper set (Figure 3-14), perhaps reflecting greater sediment inputs and/or greater lake productivity in the Late Holocene. The geotechnical properties of selected cores revealed that this is not simply a result of compaction (Reasoner and Rutter 1988). Davis and Ford (1982) suggested that sedimentation rates might be expected to decrease over time because, as material accumulates in the basin, sediment is spread over a larger area. This hypothesis is not supported by the Lake O'Hara data, which instead show that sedimentation rates generally increased during the late Holocene, possibly in response to erosion by more extensive upvalley glaciers during the Neoglacial. The very low total amount of sediment accumulation in Lake O'Hara (<2m in the Holocene) means that the overall shape

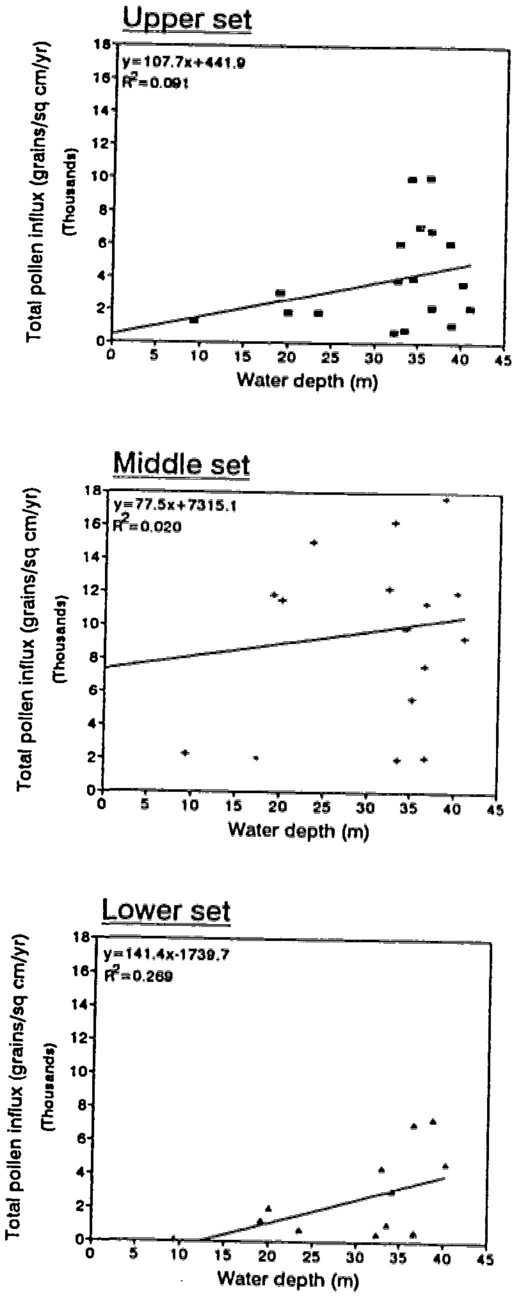

Figure 3-13: Total pollen influx plotted against water depth for three sample sets from Lake O'Hara. Total pollen influx is based on the pollen sum.

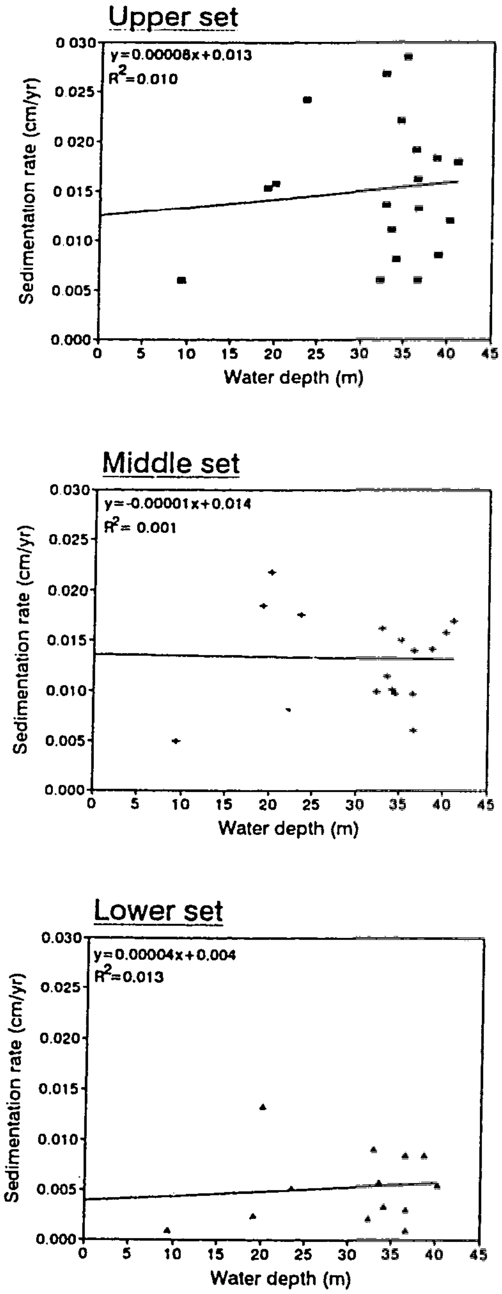

Figure 3-14: Sedimentation rate plotted against water depth for three sample sets from Lake O'Hara.

of the basin has not changed substantially over this interval. This is a different pattern than that described by Davis and Ford (1982) in Mirror Lake, where greater sediment accumulation in the central part of the lake basin since ca. 14,000 yr BP had caused a substantial change in basin morphology with the basin profile becoming less concave. In that smaller and shallower basin, morphology may exert more control over sediment accumulation. Davis and Ford's (1982) suggestion probably applies mainly to smaller basins or ones that are close to being totally sediment-filled.

Generally, sedimentation rates are low (mean values 0.15 mm/yr, 0.13 mm/yr and 0.05 mm/yr for upper, middle, and lower sets respectively). In both the middle and upper sets, the lowest sedimentation rates occur in LOH15, which at 9.4 m is the core from the shallowest water depth. Sedimentation rates are 0.06 mm/yr in cores LOH15, LOH6b and LOH23 in the upper set and 0.05 mm/yr in LOH15 in the middle set. Water depths for cores LOH6b and LOH23 are 36.6 m and 32.3 m respectively (Figure 3-9). The remaining cores show a wide scatter in each set. Webb and Webb (1988:293) suggested that extremely low sediment accumulation rates (defined as <10 cm/1000 yr or 0.1 mm/yr) "are characteristic of nonconstant processes of sediment accumulation". At Lake O'Hara, the stratigraphy of the cores does not show clear evidence of hiatuses in sediment accumulation or sediment removal. Instead, very low lake productivity and a small catchment, consisting partly of bare rock surfaces (Figure 3-5), combine to produce inherently low sedimentation rates.

The site with the maximum sedimentation rate changes between sets, LOH4b (20.1 m water depth) in the middle and lower sets, LOH10 (35.1 m) in the upper set. Reasoner and Rutter (1988) noted higher sedimentation rates for upper set equivalents in core LOH10 than core LOH4b (0.29 mm/yr compared to 0.16 mm/yr), but lower sedimentation rates for middle set equivalents in core LOH10 than LOH4b (0.15 mm/yr compared to 0.22 mm/yr). They noted that LOH10 was close to an inflowing stream and LOH4b was closer to the lake outlet. Thus, the location of maximum sediment accumulation has shifted randomly in the Lake O'Hara basin through the Holocene. These data suggest that there has been no significant systematic sediment focussing from shallow sites towards the deeper part of the Lake O'Hara basin, near site LOH25, during the Holocene.

These results are not unexpected, given the control that basin morphology may exert over sediment accumulation, with steep sides and a gently sloping floor leading to an even distribution of sediment across the basin (Lehman 1975). In addition, much of the sediment in the upper parts of the cores is organic and therefore probably reflects lake productivity. In contrast, Davis and Ford (1982) found that younger organic-rich (Holocene) sediment had been strongly focussed into the deeper parts of the Mirror Lake basin (maximum water depth 10 m), with little or no sedimentation after 11,000 yr BP at shallow sites with less than 5 or 6 m water depth. Generally, decreased sedimentation rates in the upper part of the cores were attributed to developing forest cover and increased landscape stability. However, Mirror Lake is a smaller and shallower lake than Lake O'Hara. Hence the shoreline and near-shore areas, from which redeposited sediment might be derived by current, wave or wind action, are a greater proportion of the total lake area than equivalent areas at Lake O'Hara. If sediment is being redeposited from these areas at Lake O'Hara, the magnitude of total sedimentation in the lake is great enough that this effect is masked. Therefore, one mechanism for sediment focussing suggested by Davis et al. (1984), i.e., redeposition of sediment in deeper water areas after resuspension mainly in the littoral zone, is probably ineffective in Lake O'Hara

The climatic regime at Lake O'Hara probably also influences the annual pattern of sediment deposition. In particular, the long interval of winter ice cover reduces the effectiveness of wind and water erosion for resuspension and redeposition of sediments from shallow water areas. The ice cover protects shallow water sediments from winter storms. Sediment already in suspension can settle over the entire lake floor during the winter when inflows and currents within the lake are reduced. This is similar to the "continuous complete mixing" mechanism for sediment distribution identified by Hilton et al. (1986) from the work of Pennington (1974). Significantly, Pennington (1974) found no evidence for sediment focussing in five Lake District lakes, despite their differing morphometry.

There are several different sources for clastic sediment input to Lake O'Hara. The processes controlling these clastic inputs may vary throughout the watershed over time, leading to different relative magnitudes of sediment inputs from different sources. Debris flows enter the lake mainly along its northern shore and are likely to consist of relatively coarse material that would be deposited close to the lake margin. Similarly, avalanche debris enters the lake close to its northern shore and is likely to be coarse. Sediment inputs from the streams flowing from upvalley lakes may be related to the efficiency of these lakes as traps for fine sediment originating from upvalley ice, particularly during the Neoglacial (see Reasoner and Rutter 1988). Sediment inputs from inflows from Mary Lake (Figure 3-3) are likely to be small but, because the lake is surrounded by forest, may be particularly influenced by fires in the watershed. The occurrence of charcoal in the record (Figure 3-7) shows that fire has occurred intermittently in this area through the Holocene. After fire, erosion in the watershed and therefore sediment input to Lake O'Hara may be increased. Therefore, the location of maximum sedimentation in the lake may vary according to which sediment source is dominant at a particular time.

We concluded that the great variation in sediment accumulation across the Lake O'Hara basin undermines the reliability of pollen concentration data and, in particular, pollen influx data. This has important implications for interpretation of pollen records. For instance, total pollen concentration values from core LOH25 appear to show distinct trends over time (Figure 3-4). The change from very low concentration values in Zone 1 to markedly greater values in Zones 2 and above is probably real, because it is associated with a change in sediment type. Assessment of concentration values' trends in the upper part of the core is more difficult. Although the general pattern from the sample sets (highest values for the middle set, lowest values for the upper set) match the general trends from the single core, the lake-wide sample sets show marked variation in concentration values at each level (Figures 3-4 and 3-9). Thus much of the variation in concentration values in the upper part of core LOH25 may reflect "noise" within the data.

Influx values are often thought to be more stable and hence more reliable than pollen percentage data as a basis for interpreting pollen records because, unlike percentages, values for a taxon are unaffected by changes in other taxa (see Faegri et al. 1989). Our analysis suggests that although the relative proportions of taxa represented in an assemblage based on influx values may be meaningful, the actual values are more difficult to interpret (Figure 3-10). In particular, as also noted by Davis et al. (1984), inferences about changes in the relative magnitude of sediment supply and pollen input to the lake basin from trends in influx values cannot be made from a single core.

Variability of pollen assemblages and implications for zonation

Variations in pollen assemblages are used as criteria for setting up pollen zones, either intuitively or using numerical methods, such as ZONATION (Birks and Gordon 1985). Because the three sample sets are from three pollen zones identified by Reasoner and Hickman (1989), we wanted to examine whether variability within sample sets was less than that between sets. Specifically, if variability within sample sets is so great as to obscure differences between them, then criteria for zonation may also be questionable.

The consistency in the proportions of major taxa across the lake basin (Figure 3-8) was emphasized by the results of the discriminant function analysis (DFA). In the DFA of the percentage data, five pollen taxa (percentages of *Pinus*, *Picea*, *Tsuga*, *Betula* and *Alnus*) were sufficient to define the two discriminant functions that distinguished the three sample sets (Table 3-2). Different combinations of the major taxa were used to define discriminant functions for the influx and concentration data. Most variables selected, such as *Pinus* and *Picea*, are those that show obvious differences between groups from the percentage pollen diagrams (Figure 3-7). For instance, *Tsuga* is chosen as one variable in all three analyses; presumably, it is a good discriminating variable because its values for one group are all zero. Scatterplots of the samples based on their discriminant functions (Figure 3-15) show that the first discriminant function is separating the upper set from the other sets, whereas the second function distinguishes between the middle and lower sets. For the percentage data, the first function accounts for 92.4% of variation between the sample sets, whereas the second function accounts for only 7.58% (Table 3-2). The separation between groups is much less clear when the concentration and influx data are used (Figure 3-15) with less variance accounted for by the first discriminant function (Table 3-2).

SPSS^{x} DISCRIMINANT allows testing the adequacy of the derived discriminant functions by entering a classification phase. In this phase, DISCRIMINANT uses the derived discriminant functions to

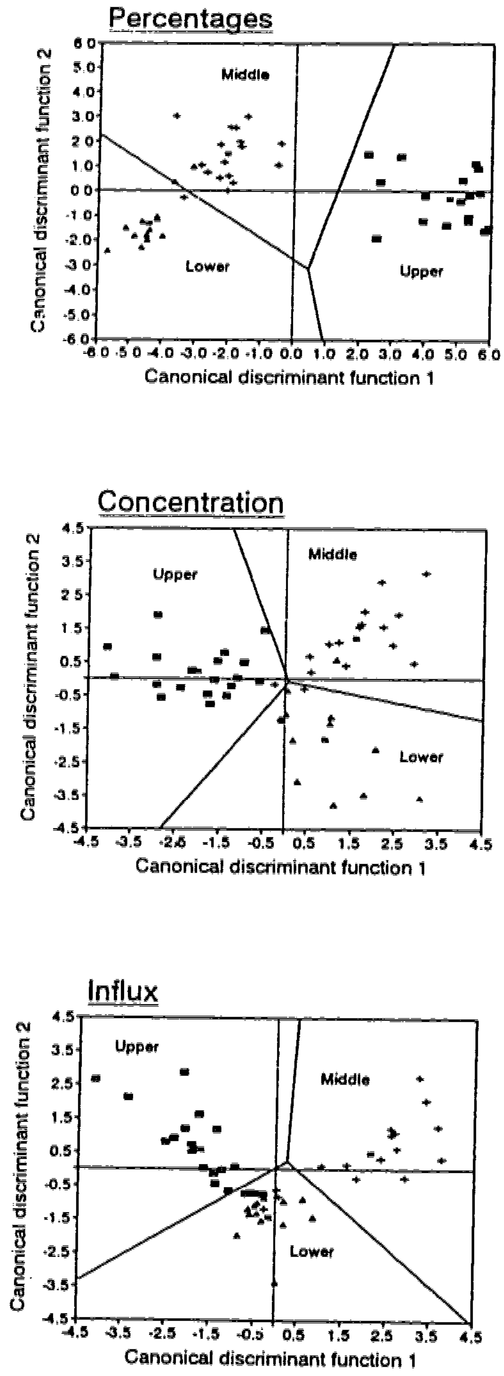

Figure 3-15: Scatterplots showing samples from Lake O'Hara grouped by discriminant function analysis, based on pollen percentages, concentration and influx data for ten selected taxa. Data from output produced by SPSS^{X} DISCRIMINANT. Legend: n = upper set, + = middle set, Ù = lower set, * marks position of group centroid.

classify samples from the defined groups (lower, middle and upper sample sets) as though they were unknown. The analysis results in a table (Table 3-3) indicating what proportion of each sample set is correctly classified. The proportion of each set correctly classified, therefore, provides an indication of the success of discrimination. Using the percentage data, discrimination between the sample sets is good, with almost 96% of cases correctly classified (i.e., assigned to their true group) and only two samples, one each from the middle and lower sets, being incorrectly classified (Table 3-3). The two samples misclassified are LOH3 lower set, classified as middle set, and LOH7 middle set, classified as lower set. The scatterplot (Figure 3-15) shows this more clearly. The two misclassifications are influenced mainly by the amounts of *Artemisia* pollen, which are 0.91% and 2.81% respectively (Figure 3-7a and 3-7b), at the lower and upper extreme values for this taxon. This analysis confirms that for the percentage data the major pollen taxa serve to characterize and distinguish the sample sets and that the sample sets are clearly distinct.

The situation is less clear for the concentration and influx data. The first discriminant function is less successful in separating the sets (Table 3-2, Figure 3-15). For the classification phase, almost 94% of samples are correctly grouped using the concentration data, with three samples misclassified (Table 3-3). For the influx data, however, the percentage of correct classifications drops to about 86%, with six samples misclassified (Table 3-3). This deterioration probably arises because of the additional level of variability and uncertainty in the data introduced from the sedimentation rates when influx values are calculated. This variability is clear from the concentration and influx diagrams (Figures 3-9 and 3-10).

Using MANOVA results for the percentage data, based on the ten pollen taxa included in the analysis, significance tests indicate that the null hypothesis (no significant difference between sample sets) can be rejected (Table 3-4). This shows that the difference between sample sets is greater than the variability within sample sets. The analysis of variance results also show significant differences between the sample sets for percentages of individual taxa, except for *Juniperus*, *Alnus*, and Gramineae pollen (Table 3-2). Gramineae and *Juniperus* are the only taxa whose mean percentage values are less than 1% in each sample set. For *Alnus*, although the means within each sample set are different, the greater standard deviation, particularly for the upper set, reduces the distinction between the sets (Table 3-1). The MANOVA analyses also confirm that there are significant differences among the three sample sets, based on the ten major taxa, when concentration or influx data are examined (Table 3-4).

This analysis has some general implications for zonation. Numerical methods of zonation, such as the ZONATION program of Birks and Gordon (1985), are usually most influenced by the numerically most important taxa. Because these sample sets are clearly distinct based on the major taxa, our case study lends support for numerical zonation methods as a powerful means of classifying pollen data from a single core. In contrast, visual zonation is subjective and may be more affected by, and give more weight to, the stratigraphic patterns shown by minor components of the pollen assemblages.

Because these minor taxa are so variable in their occurrence in our case study, we suggest their patterns of occurrence should not be used as criteria for zonation.

Our main reason for examining differential pollen deposition and pollen focussing was to assess the validity of the Late Quaternary vegetation history derived from a single core from Lake O'Hara. Pollen zones in core LOH25 (Figure 3-4), which were used as the basis for inferring vegetation history, were identified by trends in major taxa, defining areas of relative homogeneity (Zone 2) or systematic change (Zones 3 and 4). Although the three sample sets each represent only one sample per zone, there is a close correspondence between almost all the average percentage values for the major taxa in the sets and the zones (Table 3-1). In addition, the agreement between the mean values for the sets and the equivalent samples from LOH25 are good (Table 3-1). We concluded that the zonation established for a single core from Lake O'Hara based on the major taxa is meaningful. Further, the zonation from the single core could be extended across the lake basin, changing the zones to three dimensional constructs. We concluded that the main characteristics of the pollen assemblages of these three sample sets substantiate inferences about vegetation history made from one core by Reasoner and Hickman (1989).

CONCLUSIONS AND IMPLICATIONS FOR SAMPLING AND INTERPRETATION OF POLLEN RECORDS

1. These data from Lake O'Hara do not show strong evidence for either differential pollen deposition or pollen focussing at the three intervals examined. Specifically, these data show no indication of significantly greater pollen accumulation at the deeper water sites based on pollen influx values, previously used as an indicator of sediment focussing by Davis and Ford (1982). Further, there is no evidence for the preferential sorting and deposition of any individual taxon within the basin. Therefore, in this case the pollen record from a single core represents the pattern of pollen deposition in the lake as a whole.

2. Relative proportions (i.e., percentages) of major components of the pollen assemblages are consistent throughout the basin (Figure 3-8) because there is no evidence for differential pollen deposition or pollen focussing. Hence, a similar pollen record would have been obtained irrespective of core location. Therefore, we conclude that in this case the location of the sampling site within the lake basin is not critical when pollen percentage data are examined.

3. The major pollen taxa are remarkably consistent within a sample set but show significant differences between sample sets. This consistency at a chronostratigraphic level compared to variability between levels implies that qualitative interpretations of pollen records based on fluctuations of the major taxa are likely to be reliable, as Davis and Ford (1982) also concluded.

4. This case study suggests that minor components (<1%) of the pollen assemblage are so variable that they are unreliable indicators of vegetation changes through a core. Therefore, the absence or

disappearance of a minor taxon from the pollen assemblage cannot be used to infer its local absence in vegetation reconstructions. Furthermore, because of their inherent variability, minor taxa are not reliable for zonation of pollen records or for characterizing pollen zones.

5. Quantitative measures of pollen abundance, particularly pollen influx, are highly variable throughout the Lake O'Hara basin, probably because the values are so strongly influenced by sediment fluxes. As pointed out by Davis and Ford (1982), quantitative inferences can only be made after examining several cores to find the pattern of sediment accumulation. Greater chronologic control (radiocarbon dates) would be required for each core to account for temporal and spatial variations in sediment depostion. Therefore, we concluded that little confidence can be placed in pollen influx data from a single core for interpretation of vegetation history, a conclusion also reached by Davis et al. 1984.

6. The distinction among the three sample sets, highlighted by DFA and MANOVA, and comparison between the sample sets and the equivalent zones from core LOH25, supports the validity of the zonation of the single record. However, this analysis shows that zonation will only be meaningful when based on major components of the pollen assemblage.

7. This study suggests that basin morphology may play an important role in the pattern of pollen accumulation within lake sediments, as also suggested by Davis et el. (1984) and Lehman (1975). The shape of the Lake O'Hara basin (steep sides, relatively flat floor) and also its comparatively large size are likely responsible for reducing any impact of pollen and sediment focussing. Consequently, pollen focussing may be a greater concern in smaller, shallower basins, such as Frains and Mirror Lakes, where basin morphology exerts a stronger control over sediment accumulation.

8. These results have implications for site selection for palaeovegetational studies, as discussed by Jacobsen and Bradshaw (1981). Our results suggest that comparatively large deep lakes, like Lake O'Hara (43 m depth, 34.4 ha area), may be ideal sampling sites for the study of vegetation history. Because the Lake O'Hara records show no compelling evidence for differential pollen accumulation or pollen focussing, two sources of variability in the pollen assemblage data from a single core are much reduced or eliminated. It is possible that differential pollen deposition may be a greater concern in smaller, shallower lakes.

9. Finally, this study suggests that consistent changes or trends in major taxa with time through a core are likely to reflect real events, rather than reflect noise within the data. Therefore, we conclude that inferring Late Quaternary vegetation history from a single core within the Lake O'Hara basin is justified, if interpretations are based on pollen percentages for major components of the pollen record. This presumably holds true for lakes of similar size and form.

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CHAPTER 4

Age of the Crowfoot Advance in the Canadian Rocky Mountains: A glacial event coeval with the Younger Dryas oscillation.

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INTRODUCTION

The late Pleistocene Younger Dryas (YD) cold event has been generally dated at ca. 11,000 to 10,000 ^{14}C yr B.P. (Before Present: AD 1950) at a number of European sites (e.g. Berglund, 1979). The duration, severity, and, in particular, the abruptness of termination of the (YD) event preclude Milankovitch (orbital) forcing as a direct cause and suggest that thresholds affecting oceanic and atmospheric circulation patterns may have been surpassed (e.g., Broecker et al., 1988, Wright, 1989; Broecker, 1990; Rooth, 1990; Miller and Kaufman, 1990; Taylor et al., 1993). The YD cold event has traditionally been associated with the North Atlantic region and the absence of a YD cold event from several widely separated areas has been noted (McGlone, 1988; Broecker et al., 1988; Heusser, 1989). However, recent proxy record studies have reported evidence for the YD event in areas outside the North Atlantic region (e.g., Chinzei et al., 1987; Overpeck et al., 1989; Engstrom et al., 1990; Clapperton, 1990; Flower and Kennett, 1990; Linsley and Thunell, 1990; Kudrass et al., 1991; Mathewes et al., 1993), which suggests that the climatic impact of the YD event may have been global. In this paper we present evidence that a regional advance of alpine glaciers in western North America was coeval with the European YD event.

The Crowfoot moraine system represents a distinct glacial advance that is recognized throughout the Canadian Rocky Mountains in areas where the moraines have not been overridden by late Neoglacial ice advances (Luckman and Osborn, 1979; Osborn and Luckman, 1988). Moraines thought to be correlative with the Crowfoot moraines have been identified in several other ranges in western North America (Davis and Osborn, 1987; Osborn and Luckman, 1988). The only chronological constraints on the timing of the Crowfoot Advance have been the presence of Mazama tephra (ca. 6800 yr B.P.) in soils overlying the moraine at the type locality and the absence of Glacier Peak G tephra (ca. 11,200 yr B.P.) on correlative moraines in Montana. Crowfoot Lake is located directly outside of the type Crowfoot moraine, ~10 km downvalley from both the headwaters of the Bow River and the Continental Divide (Figure 4-1). Crowfoot Lake receives meltwater directly from modern glaciers and would have during the Crowfoot Advance also. Nine sediment core samples were recovered from Crowfoot Lake and adjacent Bow Lake to identify and radiocarbon date sediments related to the Crowfoot Advance and to obtain paleoenvironmental records for the site based on pollen, diatom, pigment, and Rock-Eval carbon analyses.

PROCEDURES AND RESULTS

All nine cores (recovered with a portable percussion system; Reasoner, 1993) generally show a consistent lacustrine stratigraphy and penetrated as much as 50 cm of basal diamict (Figure 4-2). The basal diamict in core CRW6 is overlain by up to 20 cm of sandy clay of low organic content that in turn is sharply overlain by highly organic gyttja that contains marl laminae (Figure 4-3). An accelerator

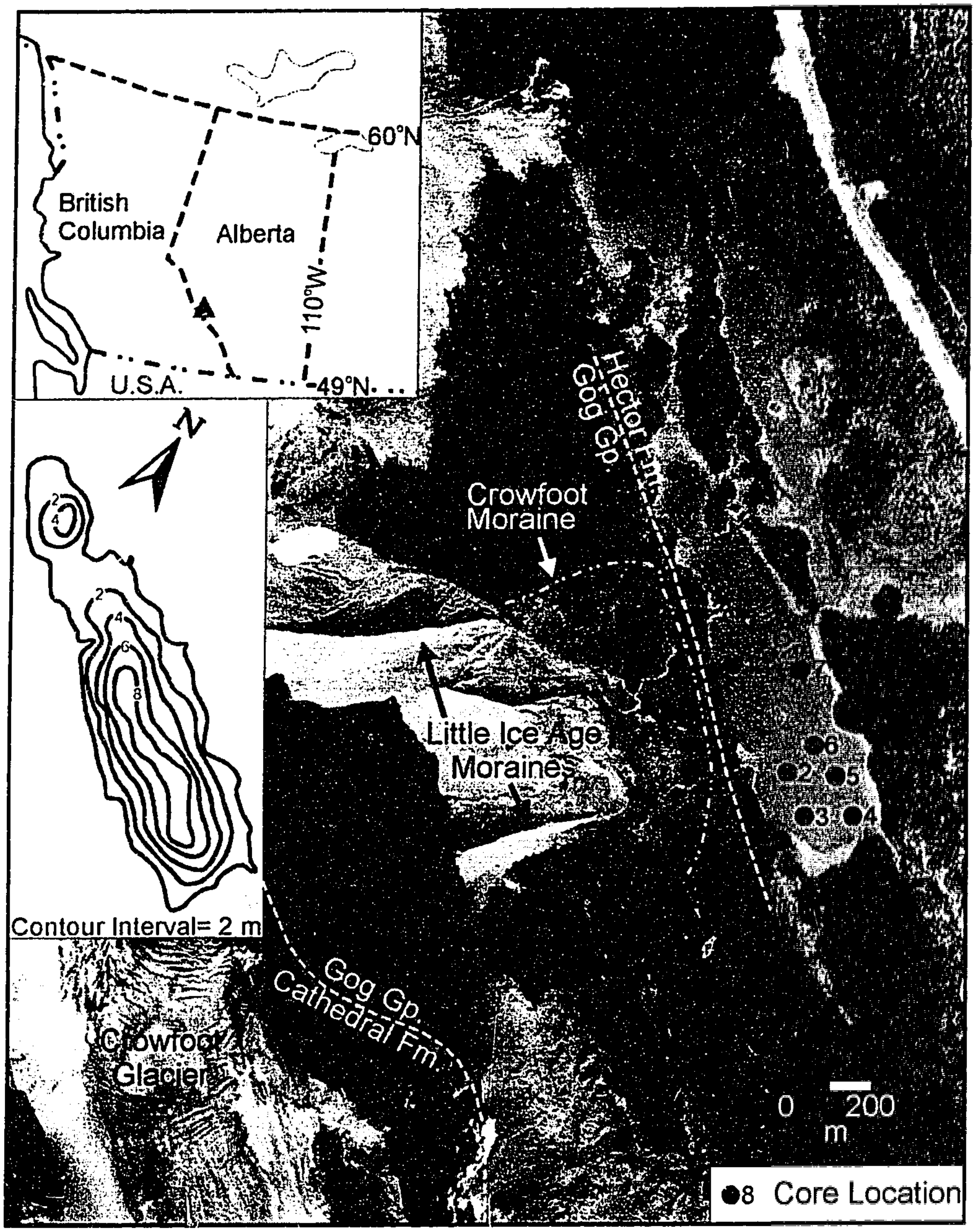

Figure 4-1. Air photograph of the Crowfoot Lake area and bathymetric map of Crowfoot Lake. Solid triangle in inset map shows location of study area in western Canada. Also shown are core locations and approximate contact between Cathedral Fm., Gog Gp., and Hector Fm.. The Bow Lake core was recovered from a location midway between the eastern and western shores and approximately 300m northwest of the top edge of the air photograph.

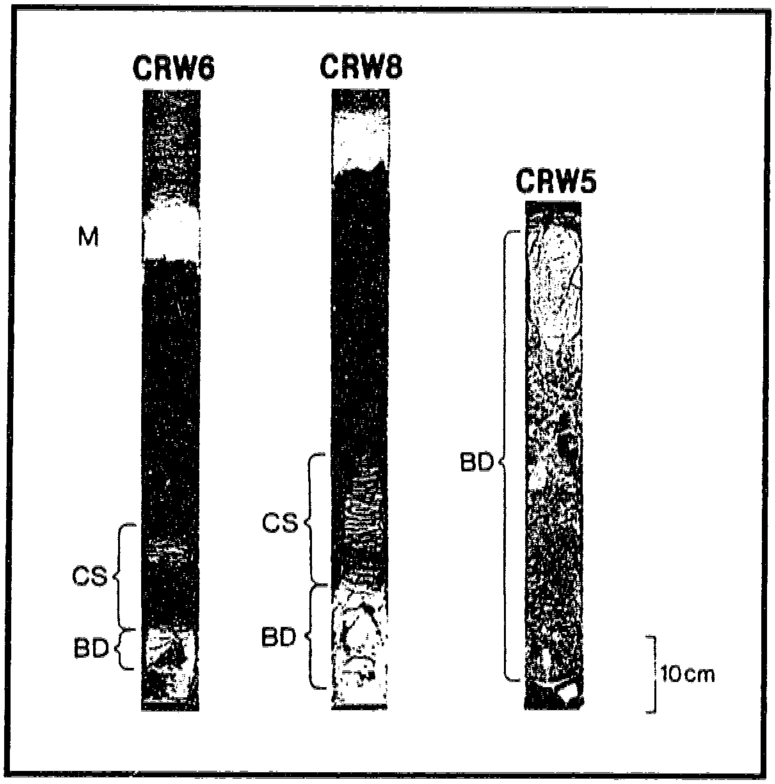

Figure 4-2. Details of lower sections of Crowfoot Lake cores. BD - basal diamict. CS - Crowfoot sediments. MZ - Mazama tephra.

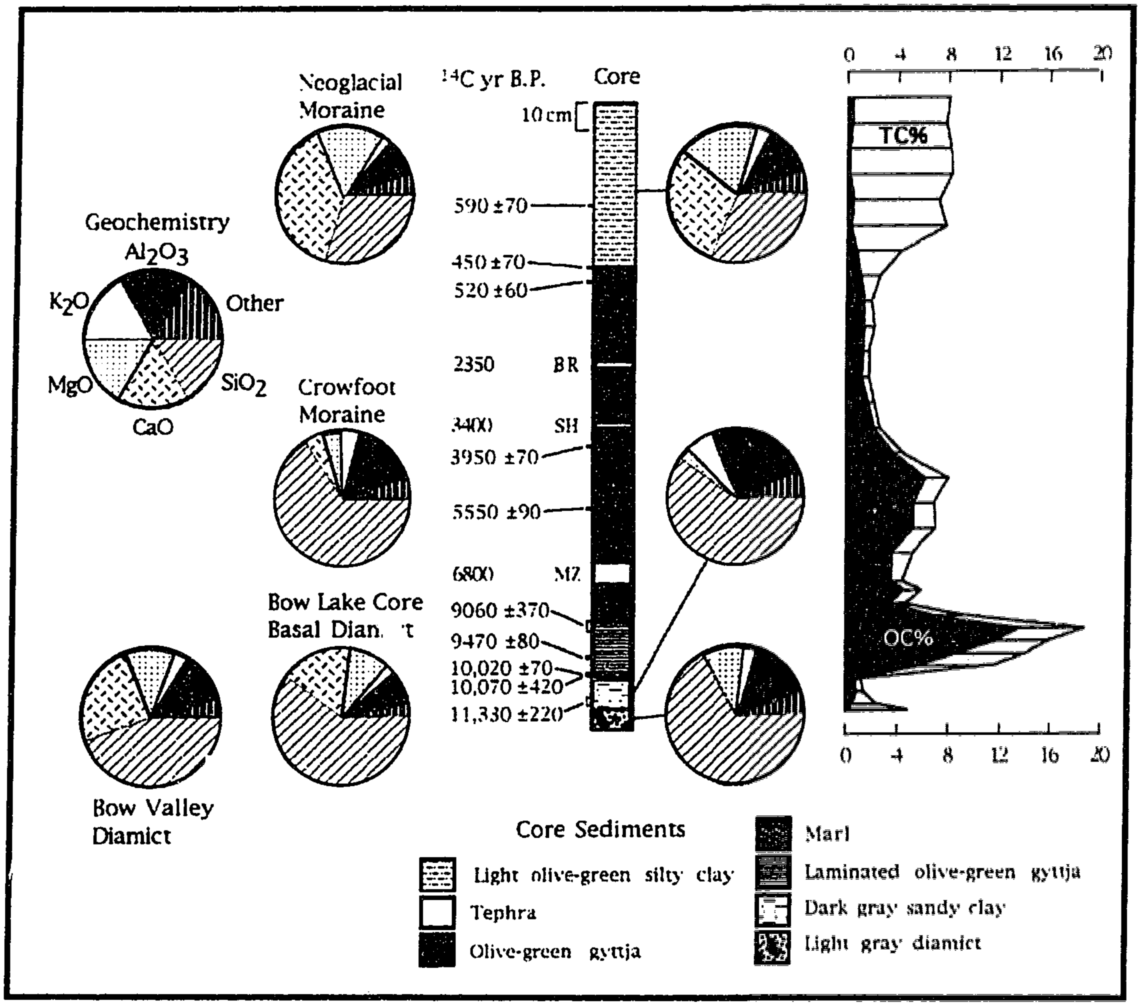

Figure 4-3. Stratigraphy, chronologic control, bulk geochemistry, and carbon content for core CRW6 and bulk geochemistry of diamict samples. M - Mazama tephra. SH - St. Helen's Yn tephra. BR - Bridge River Tephra. TC% - Total carbon content. OC% - Organic carbon content.

mass spectrometry (AMS) radiocarbon age of 11,330 ±330 ^{14}C yr B.P. was obtained from near the base of the sandy clay. Marl-bearing gyttja that directly overlie the inorganic sandy clay provided AMS ages of 10,070 ±420 and 10,020 ±70 ^{14}C yr B.P. (Figures. 4-2 and 4-3). Further Holocene geochronology for core CRW6 is based on an additional seven AMS radiocarbon ages. All radiocarbon ages were derived from terrestrial macrofossils (Table 4-1). Laminated and massive organic gyttja overlie the marl-bearing gyttja and contain Mazama (6800 ^{14}C yr B.P.), St. Helen's Yn (3400 ^{14}C yr B.P.), and Bridge River (2350 ^{14}C yr B.P.) tephras (Figure 4-2). Age-depth relations of the radiocarbon ages and tephras of known age support the validity of the AMS ages (Figure 4-3). With one exception, the uppermost age of 590 ±70 ^{14}C yr B.P., the radiocarbon ages increase in a regular fashion with sediment depth.

The modern Crowfoot Glacier rests almost entirely on dolomites of the Middle Cambrian Cathedral Formation (Figure 4-1). In contrast, the Bow Valley is underlain by quartzite and shales of the Lower Cambrian Gog Group and slates of the upper Precambrian Hector Formation. Samples of core sediments and local till were analysed for clast lithology and bulk geochemistry to use this contast in bedrock lithology in determining the provenance of the sediment. The three intervals of inorganic sediments in the Crowfoot and Bow Lake cores (the basal diamict, sediments dated as ca. 11,300 - 10,100 ^{14}C yr B.P. and sediments post-dating ca. 500 ^{14}C yr B.P.) were sampled for bulk geochemistry, as were the Little Ice Age and Crowfoot moraines and Bow Valley till. Standard X-ray fluorescence procedures were used on the <420 μm fraction of the samples.

The basal diamict in the Crowfoot and Bow Lake cores and Bow Valley diamict contain a variety of lithologies, however, all contain abundant clasts of Hector Formation slates. The bulk compositions of these three diamicts are similar and reflect the variety of clast litholgies (Figure 4-3). Samples of Little Ice Age moraine contain primarily dolomite clasts. The bulk composition of this moraine, and the uppermost silty clay in core CRW6, show high percentages of magnesium and calcium oxides (Figure 4-3). In contrast, the bulk composition of Crowfoot Moraine samples are dominated by silica, and the constituent clasts are primarily quartzite. Inorganic sandy clay that overlies the basal diamict in core CRW6 also shows high silica and low magnesium and calcium oxide percentages (Figure 4-3).

DISCUSSION

Identifying the source of the basal diamict is key to understanding the stratigraphy of the cores. If the basal diamict was deposited by Bow Valley ice, the diamict must predate sediments related to the Crowfoot Advance. Cross-cutting geomorphic relations demonstrate that the Crowfoot Advance represents a distinct event that postdates the recession of ice from the Bow Valley (Luckman and Osborn, 1979). Alternatively, the basal diamict in the cores may represent debris flows that originated from the Crowfoot moraine during or shortly after the Crowfoot event. However, three lines of

Table 4-1. Radiocarbon Ages from Crowfoot Lake Core CRW6

Lab no.	*Core Depth (cm)*	*^{14}C Age (yr B.P.)*	*Dated Material*
CAMS3840	40	590±70	*Picea* needle
CAMS6842	63	450±70	*Picea* needle
CAMS4405	71	520±60	*Picea* needle
CAMS3843	134	3950±70	*Picea* needle
CAMS4406	160	5550±90	*Pinus* needle
CAMS3064	205-208	9060±370	*Abies* needle
CAMS6843	218	9470±70	*Pinus* needle
CAMS3063	226.5	10.020±70	*Pinus* needle
CAMS3177	227	10,070±420	Cyperaceae seed
CAMS3065	235-238	11,330±220	*Salix* (?) twig

Note: An estimated ^{13}C value of -25±2 ‰ was used for all samples.

evidence suggest that the basal dimict in the cores was deposited by Bow Valley ice. Firstly, the bathymetry of Crowfoot Lake before ca. 11,330 ^{14}C yr B.P. suggests that the basal diamict did not originate as debris flows from the Crowfoot moraine. Debris flows would have had to blanket both sub-basins of Crowfoot Lake and extend at least 7 m vertically up the distal slopes without filling in the basins. Further, the stratigraphy of the Crowfoot Lake cores is mirrored in a core from adjacent Bow Lake, which is upvalley from, and well beyond the influence of, the Crowfoot Glacier (Figure 4-1). Secondly, there is a high proportion (59%) of Hector Formation slates in the basal diamict of the Crowfoot Lake cores. The outer edge of the Crowfoot moraine just reaches the contact between the Gog and Hector Formations (Figure 4-1); consequently clasts of Hector Formation are not present in samples of the Crowfoot moraine. Finally, the bulk composition of the basal diamict in Crowfoot Lake cores is generally similar to samples of Bow Valley till and basal diamict from the Bow Lake core (Figure 4-3). The basal diamict therefore did not originate from the Crowfoot moraine and instead represents deposition from Bow Valley ice.

As the Crowfoot Advance postdates the deposition of the basal diamict in core CRW6, lacustrine sediments that are correlative with the advance must overlie the diamict. Marl-bearing gyttja that postdates ca. 10,100 ^{14}C yr B.P. and predates Mazama tephra shows the highest organic carbon content of the entire record (Figure 4-3) and therefore precludes the possibility of a significant advance of the Crowfoot Glacier during this interval. Further, preliminary pollen, diatom, and pigment analyses indicate that these organic sediments represent the most productive period in the the Crowfoot Lake history and suggest the presence of local forest vegetation in the vicinity of the lake (Reasoner et al., unpublished). Consequently, the only sediments from Crowfoot Lake that can be correlative with deposition of the the Crowfoot moraine are the inorganic sediments bracketed by the ca. 11,330 and 10,100 ^{14}C yr B.P. ages, and the bulk composition of these inorganic sediments most closely matches the bulk composition of the Crowfoot moraine (Figure 4-3). These results demonstrate that the inorganic sandy clay bracketed by radiocarbon ages of ca. 11,330 and 10,100 yr B.P. are correlative with the Crowfoot moraine. The timing of the Crowfoot Advance is therefore approximately synchronous with the European Younger Dryas cold event.

An ideal sequence of lacustrine sedimen[illegible]resents the YD oscillation should include sediments that in some way reflect warmer conditions prior to the YD event and that post-date deglaciation. Lacustrine sediments that are correlative with the Crowfoot Moraine, however, almost directly overlie Bow Valley diamict (Figures 4-2 and 4-3) with no intermediate sediments of higher organic carbon content, which suggests the Crowfoot Advance may have occurred shortly after the deglaciation of the upper Bow Valley.

Considering that the Crowfoot moraine extends only ~200-300 m beyond the Little Ice Age moraine, it is surprising that the bulk composition of the two moraines is dramatically different (Figure

4-3). Samples of the Little Ice Age moraine strongly reflect the Cathedral dolomites that underlie the Crowfoot Glacier in both bulk composition and clast lithology. In contrast, the bulk composition and clast lithology of the Crowfoot Moraine reflect quartzites of the Gog Group, which outcrop down-valley from the Cathedral dolomites (Figure 4-1). Regardless of the cause of this difference in moraine composition, the bulk composition of the Little Ice Age moraine and their correlative lacustrine sediments are closely matched (Figure 4-3). Similarly, the sediments most likely related to the Crowfoot moraine are geochemically similar to samples of that moraine.

Several recent studies have highlighted the extremely abrupt nature of the YD termination. For example, data on dust concentration from Greenland ice cores show that the YD - Preboreal transition spanned less than 20 years (Dansgaard et al., 1989), and reconstructions of snow accumulation indicate that the transition may have occurred in less than three years (Alley et al., 1993). In addition, the most dramatic rates of vegetation change identified in records throughout eastern North America occurred at ca. 10,000 ^{14}C yr B.P. (Jacobson et al., 1987). Isotopic data at the YD termination in the Greenland ice cores indicate a warming of about 7 °C (Dansgaard et al., 1989). The Crowfoot and Bow Lake cores all show a very sharp contact, dated at ca. 10,100 ^{14}C yr B.P., that separates the Crowfoot inorganic sediments from overlying highly organic marly gyttja (Figures. 4-2 and 4-3). A similar contact has been identified and dated by two AMS ages of 10,100 ±200 (RIDDL-433) and 10,060 ±160 (RIDDL-511) ^{14}C yr B.P. in cores from Lake O'Hara, which is situated approximately 35 km south of the Crowfoot site (Reasoner and Rutter, 1988). The sharpness of these sedimentary transitions suggests that the termination of the Crowfoot Advance was a significant and abrupt event in at least two sites in western Canada. However, a plateau in the calendrical ^{14}C calibration curve at ca. 10,100 ^{14}C yr B.P. (summarized in Bard, 1993) render correlations near this radiocarbon age less secure.

The Crowfoot Lake record add to a growing body of evidence for a widespread YD cold event. If the triggering of the YD cold event is in fact related to a disruption of thermohaline circulation patterns in the Atlantic Ocean (e.g. Wright, 1989; Broecker, 1990), it appears that the perturbation was transmitted around the Northern Hemisphere at least as far as western Canada. However, equilibrium-line altitude (ELA) depressions associated with the Crowfoot Advance are very near ELA depressions determined for the Little Ice Age advance in western Canada. Hence, we infer that the Crowfoot Advance represents a cold event that was coeval with, but less severe than, the YD event in northwesten Europe.

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CHAPTER 5

Evidence for cirque glaciation in the Colorado Front Range during the Younger Dryas Chronozone

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INTRODUCTION

The Younger Dryas (YD) oscillation is the most recent of a series of abrupt climatic fluctuations that suggest the Earth's climate has repeatedly shifted between semi-stable interglacial and glacial modes during the Late Quaternary (e.g. Alley et al., 1993; Taylor et al., 1993). Resolving the extent, timing, and relative magnitude of the YD event is essential for understanding the mechanisms responsible for abrupt climate change. The event was severe in the North Atlantic region (e.g. NASP Members, 1994; Walker, 1995) and has been well documented in paleobotanical, marine sediment, and Greenland ice core records (e.g. Mott et al., 1986; Lehman and Keigwin, 1992; Alley et al., 1993). Most records indicate that the onset of cooler conditions occurred between 13,000 and 12,000 cal yrs BP and that the event terminated at ca.11,500 cal yrs BP (e.g. Alley 1993, Taylor et al. 1993, Mayle et al. 1993). The very abrupt onset and termination of the YD event, and its occurrence during a period of increasing solar insolation in the Northern Hemisphere, preclude orbital forcing as a direct cause. A number of studies have linked the YD climatic oscillation with a major re-organization of thermohaline circulation in the North Atlantic Ocean (e.g. Wright, 1989; Broecker, 1990; Rooth, 1990; Miller and Kaufman, 1990; Lehman and Keigwin, 1992) and a growing body of evidence suggests the climatic impact extended well beyond the North Atlantic region (e.g. Peteet 1995; Hughen 1996). Recently, several disparate proxy studies have furnished evidence for a YD climatic oscillation in north western North America (Engstrom et al., 1990; Mathewes,et al. 1993; Reasoner et al., 1994; Peteet and Mann, 1994; Hu et al., 1995; Patterson et al., 1995; Gosse et al., 1995). Of these studies, only the Crowfoot Lake record in Alberta (Reasoner et al., 1994), and both cosmogenic and radiocarbon dating of cirque moraines in Wyoming (Gosse et al., 1995; Davis, 1994), imply that glacial advances were associated with the YD event in northwestern North America. In this paper we present evidence that indicates a limited advance of alpine glaciers occurred in the Colorado Front Range Mountains during the YD Chron.

Sky Pond is located within the Colorado Front Range Mountains and is the largest (4.8 ha) and second highest (3320 m) lake within the Loch Vale Watershed, Rocky Mountain National Park (Figure 5-1). Precambrian granites and metamorphic units of biotitic gneiss comprise the bedrock of the watershed (Braddock and Cole, 1990) and are covered in many areas with talus and glacial drift of Holocene and late Pleistocene age. A broad crested, terminal moraine is located between the upvalley shore of Sky Pond and the terminus of a large tongue-shaped rock glacier and a second terminal moraine is located within the Sky Pond basin (Figure 5-1). The elevations of this moraine pair (3320 to 3360 m) correspond well with the elevations (3340 to 3410) of the Satanta Peak moraines in the Colorado Front Range Mountains (Benedict, 1973). Although the commencement of the Satanta Peak Advances is poorly constrained, the termination of the most recent advance occurred at approximately 10,000 ^{14}C yr BP and based on radiocarbon dates obtained from organic matter overlying outwash from

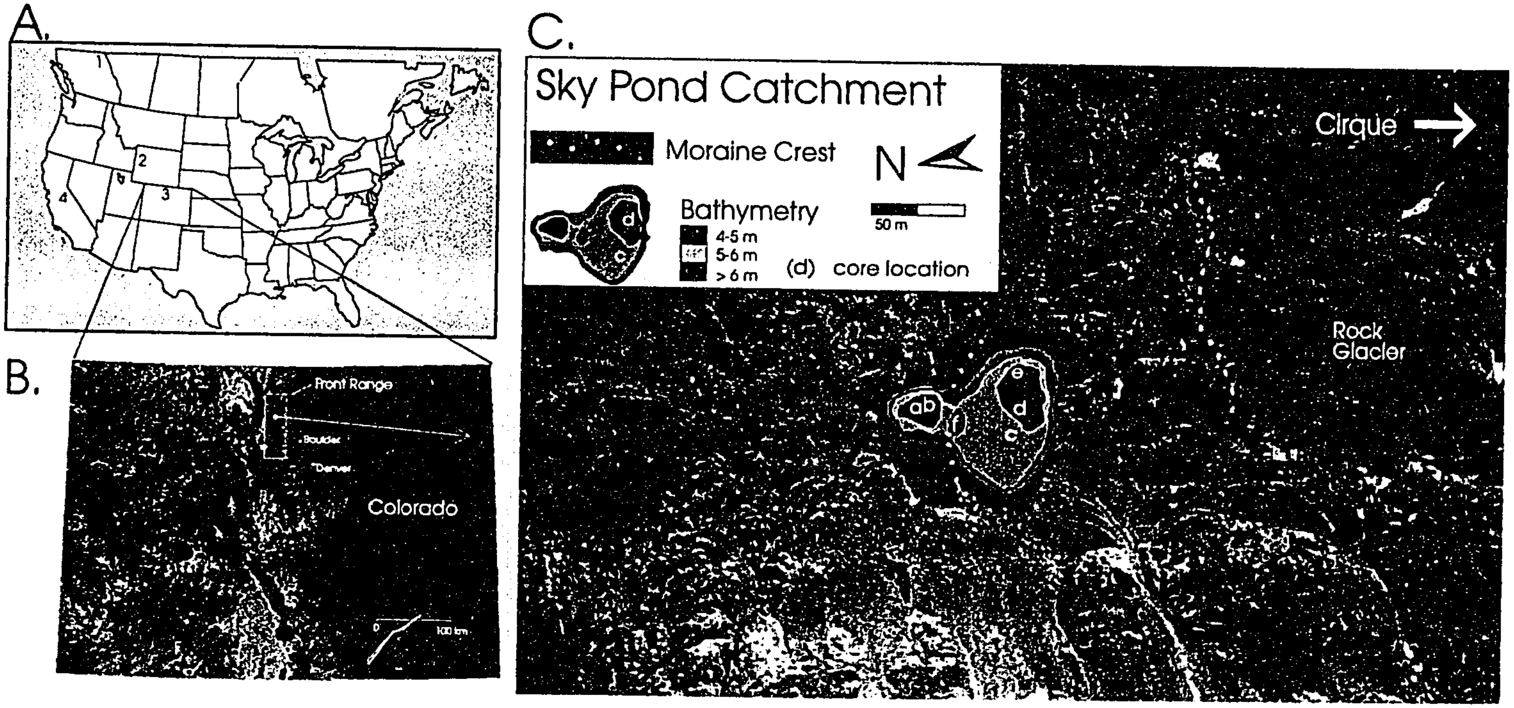

Figure 5-1. A) Location map showing study location (3) and related sites in North America: (1) Crowfoot Lake, Alberta (Reasoner et al., 1994); (2) Titcomb basin, Wyoming (Gosse et al., 1995); and (4) Sierra Nevada Range, California (Clark, 1994). B) The Rocky Mountain Front Range in Colorado. C) aerial photograph of the Sky Pond area, and bathymetric map of the Sky Pond basin showing core locations.

the type Satanta Peak moraines (Benedict, 1973; 1981).

METHODS

Six sediment cores (7.6 cm diameter) were recovered from Sky Pond in March 1995 with a percussion coring system (Reasoner, 1993). The cores were taken on two orthogonal transects; positioned across and in the down valley direction to ensure a consistent sedimentological signal for the lake (Figure 5-1). The cores were frozen, split with a rock saw and sampled for bulk density, field moisture (water content), organic matter content, and magnetic susceptibility. Magnetic susceptibility measurements were made on the < 2 mm size fraction and are useful for inter-core correlation (c.f. Dearing, 1986; King and Channel, 1991). Readily oxidizable organic matter content was determined by peroxide digestion (Walkley, 1946) and converted to percentage organic carbon by means of a conversion factor (1.3). Particle size distributions were determined for sediment subsamples from the longest core (95-01D). The > 63 μm size fraction was dry sieved and the silt (63-3.9 μm) and clay (< 3.9 μm) size fractions were determined with a Micrometrics Sedigraph analyzer. Macrofossils collected for accelerator mass spectrometry (AMS) radiocarbon dating were obtained by gently disaggregating 2 cm thick slices of sediment and washing this material through a 500 μm sieve with distilled water. With the exception of a single radiocarbon age derived from extracted humic acids, all radiocarbon ages were determined from terrestrial macrofossils. Calibrated ages were derived from the ^{14}C ages with the CALIB 3.1 program (Stuiver and Reimer, 1993). The chronology of the following discussion is reported in ^{14}C years to facilitate correlation with other studies.

RESULTS

Consistent stratigraphy of the Sky Pond sediment cores permits confident between-core correlation (Figure 5-2). This correlation is supported by six accelerator mass spectrometry (AMS) radiocarbon ages that increase with sediment depth (Table 5-1; Figure 5-2). Two cores, 95-01C and 95-01D, penetrate as much as 50 cm of basal diamict. An age of 12,040±60 yr BP was obtained from directly above the basal diamict. The diamict is overlain by dark brown organic gyttja (Figure 5-2) and the contact between the diamict and gyttja is gradational. Humic acids extracted from near the top of the gyttja provided an age of 11,070±50 yr BP. The lower gyttja is in turn overlain by an interval of brown clayey silt that contains clasts well in excess of 2 mm. The upper and lower contacts of the clayey silt are gradational and the interval is massive. The clayey silt is of lower organic carbon and water content, and shows higher dry density and magnetic susceptibility, than the underlying gyttja (Figure 5-2). Radiocarbon ages of 10,410±90 and 9970±80 yr BP were derived from within and near the top of the inorganic interval respectively. The remainder of the record is represented by dark brown organic gyttja that shows high water content and low dry density and magnetic susceptibility.

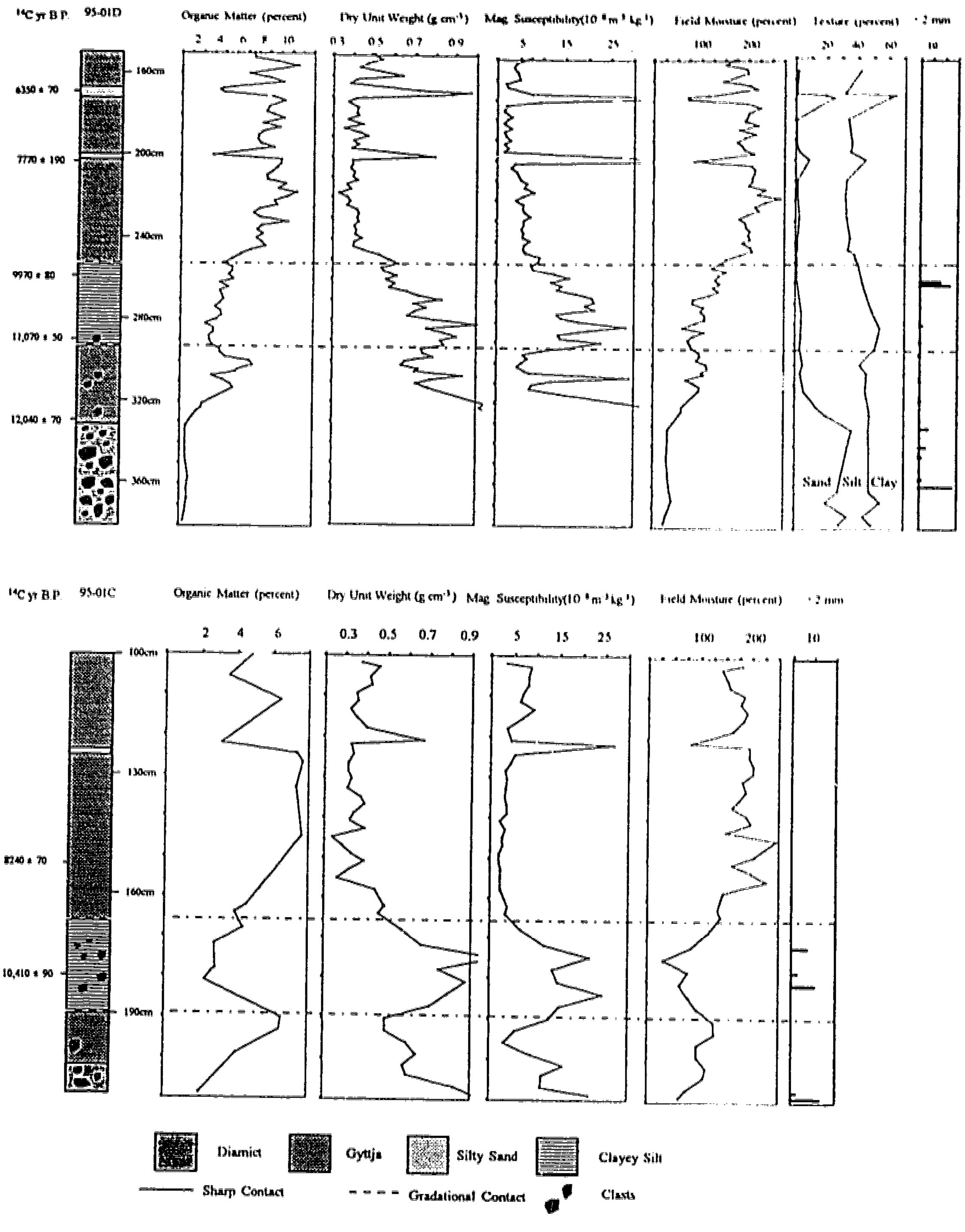

Figure 5-2. Stratigraphy, chronologic control, and bulk physical properties of Sky Pond cores 95-01D and 95-01C.

Table 5-1. Radiocarbon ages and calibrated ages from the Sky Pond cores.

Lab no.	*Core*	*Depth (cm)*	*^{14}C Age (yr BP)*	*Calibrated Age(s) (cal yr B.P.)*	*Material Dated*
CAMS-24128	95-01 C	141	8240±70	9367 (9214) 9000	conifer needle.
CAMS-24130	95-01 C	181	10410±90	12442 (12311) 12152	lepidoptera mandibles, conifer needle
CAMS-22028	95-01 D	165	6350±70	7283 (7223) 7193	wood
AA-19415	95-01 D	202	7770±190	8945 (8501) 8364	conifer needles
CAMS-24129	95-01 D	259	9970±80	11554 (11192, 11187, 11081) 11002	lepidoptera mandibles
CAMS-25554	95-01 D	296	11070±50	13050 (12983) 12913	humic Acids
CAMS-22027	95-01 D	329	12040±60	14184 (14040) 13907	lepidoptera mandibles

Calibrated ages are based on Stuiver and Reimer (1993). The most likely calendric ages are noted in parentheses.

Radiocarbon ages of 8240±70. 7770±190. and 6350±70 yr BP were obtained from gyttja overlying the inorganic interval (Figure 5-2). The upper gyttja contains thin. normally graded sandy laminae that are bounded by sharp lower and upper contacts (Figure 5-2).

Late Pleistocene to early Holocene pedologic and surface boulder weathering features associated with the moraine located directly upvalley (3360 m) of Sky Pond include silt-infiltrated O and A horizons (34 cm total). an underlying weakly developed Bw horizon (33+ cm) and cavernous differential weathering features on biotitic gneiss boulders exceeding 15 cm.

DISCUSSION

The basal age of 12.040±60 yr BP from directly above basal diamict in the Sky Pond record agrees well with limiting dates for the deglaciation of high cirques in the Colorado and Wyoming Rocky Mountains (e.g. Harbor, 1985: Zielinski and Davis, 1987; Davis et al. 1992). Consequently, the basal diamict likely represents till that was deposited during deglaciation of the Sky Pond basin and provides a minimum limiting age for the moraine located in Sky Pond (Figure 5-1). Organic sedimentation in the lake basin started shortly after this time and reached a minor peak prior to 11.070±50 yr BP presumably in response to climatic amelioration that followed deglaciation. Radiocarbon ages associated with the overlying clayey silts (10,410±90, and 9970±80 yr BP) indicate that a period of inorganic sedimentation in the Sky Pond basin occurred during the YD Chron (Figure 5-2). The subsequent deposition of organic gyttja containing conifer macrofossils suggests a return to warmer climatic conditions that favored both higher lacustrine productivity and increased vegetation cover within the catchment. These environmental conditions persisted well into the Holocene for the Sky Pond catchment and correspond well with palaeobotanical records in the southern Colorado Rocky Mountains (Elias 1985; Fall et al., 1995).

The soil development and weathering features suggest the moraine located directly upvalley from Sky Pond is mid-early Holocene in age or older and are consistent with features associated with the type Satanta Peak moraines (Benedict. 1973; 1981; 1985; Birkeland et al., 1987). Further, the moraine is situated at an altitude similar to that reported for late Pleistocene type Satanta Peak moraine (Benedict, 1973). As lacustrine records of basins situated directly downvalley from glaciers faithfully record the activity of upvalley ice (e.g. Leonard, 1986a; b; Reasoner and Hickman, 1989; Desloges, 1994; Souch, 1994), the only clastic sediments in the record that can be associated with the deposition of the moraine above Sky Pond (Figure 5-1) are the clayey silts bracketed by radiocarbon ages of 11.070±50 and 9970±80 yr BP. Further, the abundance of silt in these sediments (50 %) is consistent with textures determined for glacio-lacustrine sediments (e.g. Ostrem, 1975; Karlén 1981; Souch, 1994). The absence of internal structure, the gradational contacts, and the length of time represented by these sediments rule out mass movement as a potential origin (Figure 5-2). These results indicate that the moraine directly upvalley from Sky Pond is correlative with the clastic sediments that were deposited in the basin during the YD Chron. The similarities in soil development, weathering features, and altitude between this moraine and the type Satanta Peak moraines suggest the moraines are correlative, and consequently, the latest Satanta Peak Advance was likely coeval with the European YD event.

The available information from Wyoming (Gosse et al., 1995; Davis et al., 1994), the Canadian Cordillera (Reasoner et al., 1994) and this study, all suggest an ice advance of limited extent occurred during the YD Chron: a possibility originally proposed by Davis and Osborn (1987). Gosse et al. (1995) used cosmogenic isotope age determinations to assign the Titcomb moraine in the Wind River Range of Wyoming to the YD Chron. Although uncertainties exist with the cosmogenic dating method, the ages are in approximate agreement with estimates for the Temple Lake moraine which is also located in the Wind River Range (Davis 1994). Sedimentary records from Crowfoot Lake, Alberta indicate the deposition of the type Crowfoot moraine occurred between ca. 11,300 and 10,100 ^{14}C yr BP (Reasoner et al., 1994). The Crowfoot moraine system is widely recognized throughout the Canadian Rocky Mountains (Luckman and Osborn, 1979) and correlative deposits have been identified in the Purcell Mountains of British Columbia and the northern Rocky Mountains of Montana (Davis and Osborn, 1987; Osborn and Luckman, 1988). It appears that the Crowfoot Advance occurred shortly after the deglaciation of the trunk valleys in the Canadian Rockies and, consequently, organic sediments are not present between the basal diamict and clastic sediments associated with the advance in the Crowfoot Lake record (Reasoner et al., 1994). In contrast, organic sediments accumulated in Sky Pond for approximately 1000 years after deglaciation and prior to the onset of clastic sedimentation associated with the YD event.

In western Canada, the YD advance was similar in extent to the "Little Ice Age" advance whereas, the YD event appears to have been slightly more extensive relative to late Neoglacial advances in Wyoming and Colorado. Equilibrium-line altitude (ELA) depressions associated with the Crowfoot Advance are similar to those determined for the "Little Ice Age" glaciers in the Canadian Rockies. In many cases, the Crowfoot moraines were overrun by the "Little Ice Age" advance (Luckman and Osborn, 1979). The Titcomb and Temple Lake moraines are situated approximately 1-3 km downvalley from late Neoglacial moraines in the Wind River Range. In high cirques within the Colorado Front Range, the inner Satanta Peak moraine is also situated 1-3 km downvalley from late Neoglacial moraines. The relative extent of the ice advance responsible for the moraine directly above Sky Pond and the "Little Ice Age" advance is difficult to estimate due to the rock glacier that occupies the upper cirque (Figure 5-1). However, the maximum ice extent of this advance was less than 3 km from the cirque headwall. In contrast, Clark et al. (1994) present convincing evidence from high cirques of the Sierra Nevada, California that indicate alpine glaciers did not advance during the YD

Chron. These findings imply that glacial responses to the YD event in western North America may have had significant spatial variability. It is likely that major changes in atmospheric circulation resulted from the reorganization of North Atlantic oceanic circulation during the Younger Dryas (e.g. Lehman and Keigwin, 1992), however, the details of these changes are poorly resolved. Further documentation of the extent of alpine glacial advances during the YD Chron, particularly in the American southwest, is essential for verifying the details of regional environmental response to climatic change associated with the YD event.

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CHAPTER 6

Postglacial Vegetation and Climate History of the Upper Bow Valley, Banff National Park, Canada.

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INTRODUCTION

The general details of the postglacial vegetation and climate history of the Canadian Rocky Mountains have been resolved to a greater degree over the last decade (e.g. Luckman and Kearney 1986, Kearney and Luckman 1987, Reasoner and Hickman, 1989, MacDonald 1989, Beaudoin and King 1990, Luckman et al. 1993, Luckman 1994a, in press). The general trends from these records indicate that: 1) the high trunk valleys of the Central Canadian Rocky Mountains were deglaciated prior to 10,100 ^{14}C yrs BP; 2) the Early to Mid Holocene was a period of relative warmth and aridity associated with elevated alpine timberlines and reduced ice extent; 3) Neoglacial cooling and increased precipitation began sometime between 5000 and 3000 ^{14}C yrs BP; and 4) the "Little Ice Age" commenced between ca. 1100 - 1300 AD and culminated in the 19th century. A growing body of work, primarily carried out by Brian Luckman and his colleagues (e.g. Luckman 1993, 1994a, b, 1995, 1996), has added high resolution detail to the glacial and environmental history of the last millennium. However, the chronologies of most of the longer records in the area are generally based on the presence of one or two Holocene tephras and very few radiocarbon ages, and until recently, these radiocarbon ages have largely been determined from bulk samples using conventional techniques. Further, very few studies in the region (Kearney and Luckman 1983a, Beaudoin 1986, Beaudoin and King 1990, Luckman and Kearney 1986, Reasoner and Hickman 1989, MacDonald 1989) have recovered records from ecotonal or proglacial areas that are highly sensitive to past climatic change.

This study presents new palaeoenvironmental information from well-dated sediment core samples that were recovered from a high subalpine lake in the Central Rocky Mountains of Alberta. The Crowfoot Lake cores represent the longest AMS dated record yet reported from the eastern Canadian Cordillera, and provide the oldest limiting date for deglaciation of trunk valleys adjacent to the Continental Divide (Reasoner et al. 1994). The palaeoenvironmental reconstruction for the upper Bow Valley is based on the records of pollen, macrofossil and sediment properties. Crowfoot Lake is ideally situated to record both vegetation and glacial responses to past climate change. The site is located near alpine timberline in the upper subalpine forest and is positioned directly down-valley from late Quaternary glacial deposits (e.g., Reasoner and Hickman 1989, Beaudoin and King 1990, Leonard 1986a, b). Consequently, the sediment cores recovered from the lake provide a sensitive palaeoenvironmental record that spans the last ca. 11,330 ^{14}C years.

SETTING

Crowfoot Lake is situated in the upper Bow Valley approximately 7 km east of the Continental Divide in Banff National Park, Alberta (Figure 6-1). The lake catchment is a heavily glaciated area of high relief that covers approximately 35 km^2. Numerous peaks in the drainage basin exceed 3000 m in elevation above sea level (a.s.l.) and approximately 30% of the catchment is presently ice covered.

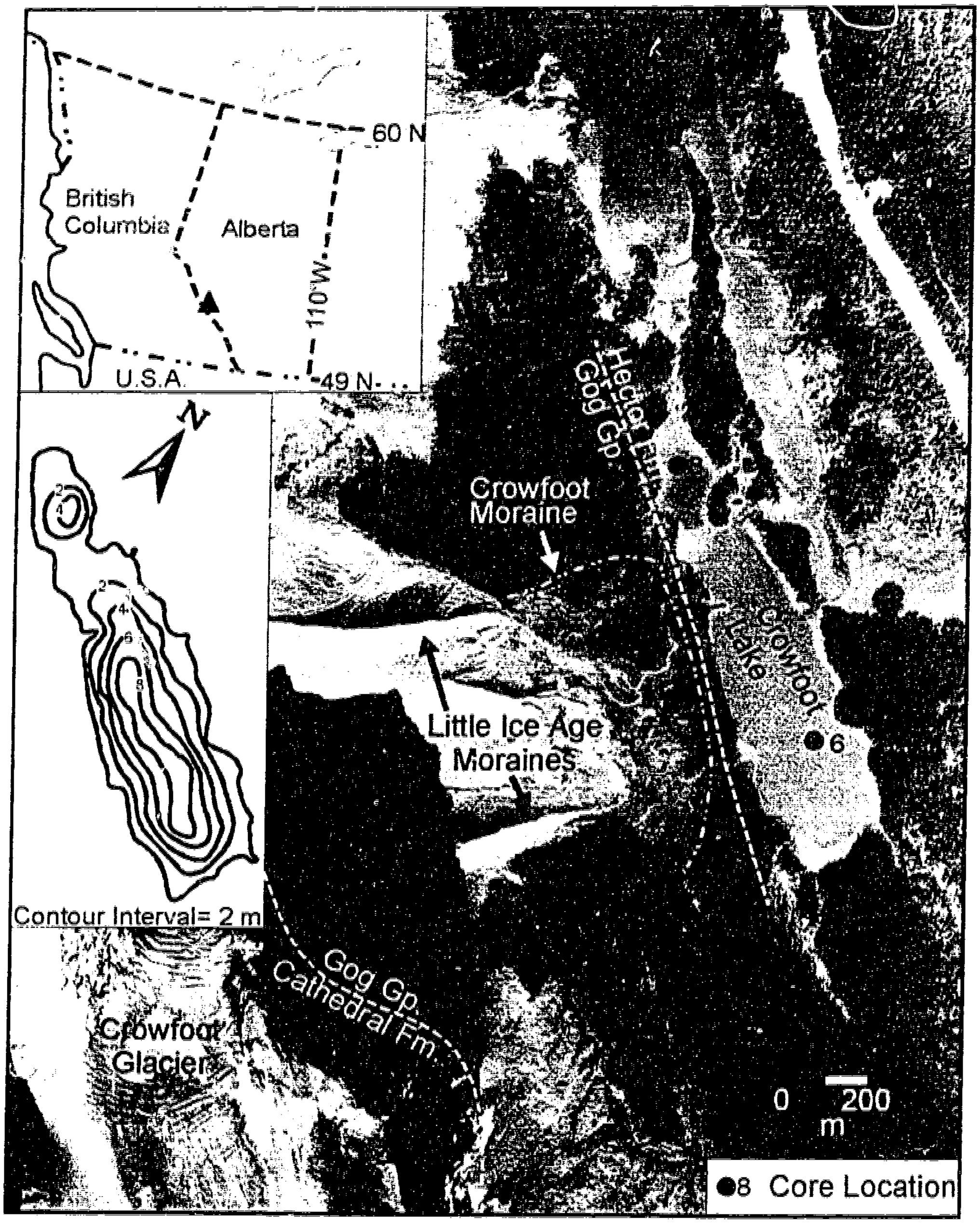

Figure 6-1. Aerial photograph of Crowfoot Lake area and bathymetric map of Crowfoot Lake showing core locations. Solid triangle in inset map shows location of study area in western Canada. Also shown are approximate contacts between Cathedral Fm., Gog Gp., and Hector Fm.

Crowfoot Lake lies at an elevation of 1940 m a.s.l. and has a surface area of approximately 0.23 km^2. The Lake has two principal inputs and both have nearby glacial sources (Figure 6-1): the Bow River enters the lake at two points along the northeastern shore, and two streams that originate from the terminus of the Crowfoot Glacier enter along the western shore. The Bow River outlet is located at the southeastern corner of the lake. Crowfoot Lake is divided into a main southern basin and a small northern basin by a shallow sill (Figure 6-1). The maximum depths of the southern and northern basins are approximately 9 m and 5 m respectively.

Argillites of the PreCambrian Hector Formation underlie the Crowfoot Lake basin (Cook 1975). The approximate location of the Upper Precambrian boundary parallels the western shore of the lake (Figure 6-1). Lower Cambrian Gog Group quartzites and shales outcrop along the western valley wall directly above the lake and Middle Cambrian Cathedral dolomites outcrop at higher elevations. The modern Crowfoot Glacier occupies a hanging valley directly west of Crowfoot Lake and rests almost entirely on Cathedral dolomites.

The type Crowfoot Moraine forms the western boundary of both the northern and southern sub-basins of Crowfoot Lake. Analyses of bulk sediment chemistry, clast lithology, and organic carbon content of Bow Valley, Crowfoot, and Neoglacial diamicts as well as radiocarbon-dated lacustrine sediments has revealed that the Crowfoot Advance occurred between ca. 11,330 and 10,100 ^{14}C yrs BP and was therefore approximately synchronous with the European Younger Dryas cold event (Reasoner et al. 1994). The Crowfoot Moraine was breached in two locations such that subsequent glacial events would have had a significant influence on the sedimentary record of the southern Crowfoot basin only (Figure 6-1). Neoglacial moraines are situated approximately 200 m west of the crest of the Crowfoot Moraine, and the terminus of the Crowfoot Glacier has receded to a position approximately 600 m west of the Neoglacial moraine crest during the 100 years.

The climate along the Continental Divide in the Central Rockies is characterized by cool, wet summers and long, cold winters. Lake ice is often present on Crowfoot Lake until mid June and re-forms in November. Mean January and July temperatures for the upper Bow Valley are -14.0 °C and 11.0 °C respectively (extrapolated from Lake Louise data 51 ° 26'N, 116 ° 11' W, 1520 m a.s.l.). Regression equations for average annual precipitation as a function of elevation (Janz and Storr 1977) suggest annual precipitation in the Upper Bow Valley may be on the order of 750-1000 mm/yr.

Crowfoot Lake is situated in the subalpine forest vegetation zone as defined by Holland and Coen (1982). The dominant arboreal taxa in the vicinity of the lake are *Picea engelmannii* and *Abies lasiocarpa* with lesser *Pinus contorta*. *Pinus albicaulis* is present on southern and western exposures near alpine. Understorey components include *Betula glandulosa*, *Salix* spp., *Menziesia ferruginea*, *Cornus canadensis*, *Vaccinium scoparium*, *Phyllodoce* spp., and *Linnaea borealis*. The alpine timberline ecotone is situated at an elevation approximately 2100-2300m a.s.l. in the upper Bow

Valley. Crowfoot Lake is also located near an ecotone that occurs locally in high subalpine valley floors near the Continental Divide. This ecotone is a "lower" timberline that separates tundra in the valley floors from subalpine forest along the valley walls. Adjacent to, and directly upvalley from the site, the valley-floor forest is less continuous than in down-valley locations, and, within 2 km upvalley from the lake, *Picea* and *Abies* are restricted to isolated stands across the valley floor (Figure 6-2). The dominant vegetation in the valley floor meadows include *Betula glandulosa*, *Salix* spp., *Menziesia ferruginea*, *Cornus canadensis*, *Vaccinium scoparium*, *Phyllodoce* spp., and *Linnaea borealis*. The absence of trees in many of the valley floors near the Continental Divide has been attributed to slightly cooler valley bottom conditions that result from cold air drainage (Gadd 1986) or moist valley-bottom conditions (Kearney and Luckman 1983a, Luckman 1990).

PROCEDURES

A suite of eight sediment cores, 7.6 cm in diameter, were retrieved from Crowfoot Lake during the winter of 1993; seven from the southern sub-basin and one from the northern sub-basin (Figure 6-1). The cores were recovered with a lightweight percussion coring device designed for winter operation (Reasoner 1993). The cores were allowed to freeze in the core barrels and were split with a high-speed diamond rock saw at Core Laboratories Canada Ltd. in Calgary, Alberta. The cores were photographed immediately after splitting and were subsampled for pollen, Rock-Eval carbon, diatom, pigment, and bulk physical analyses when thawed. All sub-samples were taken as 1 cc sediment volumes and were removed from as close to the core centre as possible to avoid contamination. Core CRW6 was chosen for pollen analyses because of minimal core disturbance and the presence in the core of all three Holocene tephras that occur in the region.

The LECO method (Tabatabai and Bremner 1970) was used for total carbon determinations and the Rock-Eval pyrolysis procedure (Espitalié et al. 1977) was used for organic carbon analyses. Hydrogen indices (HI) correspond to the quantity of pyrolyzable organic compounds relative to total organic carbon content of the sediment (Espitalié et al. 1977) and have been accepted as a proxy of primary lacustrine productivity (Ariztegui et al. 1992). Duplicate samples were analyzed for carbon content with standard loss on ignition (LOI) techniques. Although the absolute values differ, the down-core trends in total and organic carbon percentages are very similar for Rock-Eval and LOI analyses. Wet bulk densities and water contents were determined from the mass of wet and oven-dried 1 cm^3 plugs of sediment. Standard X-ray fluorescence procedures were used on the <420 µm fraction of the sediment samples to determine bulk chemical compositions (Reasoner et al. 1994). Holocene tephras in core CRW6 were identified on the basis of field properties and petrographic examination.

Chronological control for the Crowfoot Lake record is based on twelve AMS radiocarbon ages and the presence of three Holocene tephras of known age. In all cases, terrestrial macrofossils were

Figure 6-2. Subalpine meadows in the upper Bow River drainage approximately 2 km upvalley from Crowfoot Lake. L - "lower" timberline, A - alpine timberline.

used for AMS radiocarbon age determinations and care was taken to avoid any material from near the core barrel wall. The macrofossils that were submitted for radiocarbon dating were either extracted directly from the split core surface or collected by wet sieving small intervals of core. A number of studies have focussed on the physical and chemical properties of Holocene tephras in western Canada (e.g., Mathewes and Westgate 1980, Westgate and Gorton 1981, Reasoner and Healy 1986, Beaudoin and King 1994, Beaudoin and King 1986), and both the distribution and ages of these tephras are well documented. Crowfoot Lake lies well within the distributions of both Mazama and Bridge River tephras, and very near the known southeastern extent of the St. Helens Y_n plume.

Pollen samples were processed following standard procedures (Faegri et al. 1989). *Lycopodium* marker grains were added prior to the initial suspension in concentrated HCL (Stockmarr 1971) to permit the calculation of pollen concentrations and pollen accumulation rates (PAR). Following the removal of carbonate material, the samples were suspended in 45% HF to remove silicates, in concentrated HCl to remove colloidal particles, and finally in an acetolysis mixture (10% concentrated sulfuric acid and 90% acetic anhydride) to remove organic particles other than pollen. The samples were then dehydrated in an alcohol series and suspended in silicon oil.

Palynomorphs were identified and enumerated with reference to various pollen identification keys (e.g., Moore et al. 1991, Faegri et al. 1989) and the North American pollen reference collection at the Institute of Arctic and Alpine Research, Boulder, Colorado. Pollen identification was generally carried out with 400X magnification and critical determinations were made with 1000X magnification under oil immersion objectives. The presence or absence of verrucae on the distal membrane of *Pinus* pollen grains (Ting 1966) was used to distinguish between *P. contorta*-type and *P. albicaulis*-type (including *P. flexilis*). At a number of levels, however, this distinction was possible for very few grains due to poor preservation. *Haploxylon/diploxylon* ratios were calculated for levels where greater than 30 *Pinus* grains were assigned to either subgenera. In cases where vegetation groups could not be differentiated on the basis of pollen morphology, these taxa were classified together (i.e., Chenopodiaceae/Amaranthaceae, *Larix/Pseudotsuga*). Pollen samples were initially counted at 10cm intervals over the core to identify intervals of significant change in pollen spectra. The sample interval was subsequently reduced over these zones of interest. A total of 48 levels were counted in the Crowfoot record.

As *Pinus* pollen dominates the Crowfoot Lake record, and for the most part was probably of extra-local origin, sufficient numbers of pollen grains were enumerated to permit meaningful interpretations of the percentage record with *Pinus* excluded from the pollen sum. The mean grain count for the Crowfoot record is 774 (range: 260-1883) and the mean grain count with *Pinus* excluded is 245 (range: 132-324). The pollen percentage records were based on sums that include trees, shrubs, and herbs, and exclude aquatics, spores, and unidentified grains. Pollen accumulation rates (PAR) were calculated by dividing the pollen concentrations (grains/cm^{-3}) by the deposition time of the sediment (calibrated yr/cm). Deposition times for the CRW6 pollen accumulation rate record were derived from linear interpolation between accepted calibrated radiocarbon ages. Deposition times for the four lowermost levels were calculated using ages extrapolated by linear regression from overlying dated levels.

Conifer needles and other macrofossils were concentrated by wet sieving 10 cm thick core sections with a 250 mesh sieve. Approximately 1 cm of sediment adjacent to the core barrel wall was not included in the samples to avoid contamination. The needles were enumerated with a 40X dissecting microscope. *Pinus* needles were subdivided into *Pinus* cf. *contorta* type and *Pinus* cf. *albicaulis/flexilis* type.

RESULTS

Sediment Characteristics and Chronology

The seven sediment cores recovered from the southern sub-basin of Crowfoot Lake show a consistent lacustrine sedimentary sequence and penetrate up to 50 cm of basal diamict (Figure 6-3). The buff-gray (10YR 6/1) basal diamict in the cores is of very low organic carbon content, approximately 5% total carbon content (Figure 6-3), and contains abundant large clasts of which 59% are Hector Formation slates. In CRW6 the basal diamict is overlain by 14 cm of dark gray (5B 5/1 - 4/1) sandy clay with an organic carbon content of about 1% and a total carbon content of 1-2% (Figure 6-3). The sandy clays are sharply overlain by 18 cm of dark olive gray (5Y 3/1 - 2.5/1 - 3/2) laminated gyttja of relatively high organic carbon content(6-13%) that contains marl laminae in the lower 4 cm. Massive olive-gray (5Y 5/2 - 4/2 - 4/1, 5GY 4/1) gyttja occurs above the laminated gyttja from 204 to 65 cm core depth and contains three Holocene tephras (Figure 6-3). The organic carbon content declines from about 5% to 2% over 150 to 130 cm core depth (Figure 6-3). The upper 50 cm of CRW6 is light olive-green (5Y 6/1 - 5/2, 5BG 5/1) silty clay of very low organic carbon content and approximately 8% total carbon content. The interval from 75 cm to 50 cm is gradational between underlying gyttja and overlying clastic sediments. Core CRW8, the only core recovered from the northern sub-basin of Crowfoot Lake, shows a similar sedimentary sequence with the exception that silty clays do not replace the massive gyttja and organic carbon contents remain at about 2-4% in the upper portion of the CRW8 record.

Uncalibrated AMS radiocarbon ages (^{14}C yr BP) from core CRW6 terrestrial macrofossils are listed in Table 6-1 along with calibrated ages (cal yr BP). Calibrated ages were used to derive the age-vs-depth curve for the record (Figure 6-4), and were determined for both macrofossil and tephra ^{14}C ages with the CALIB 3.1 program (Stuiver and Reimer 1993). The three tephras present in core CRW6 are Mazama (ca. 6845 ^{14}C yr BP; Bacon 1983), St. Helens Y_n (ca. 3400 ^{14}C yr BP; Beaudoin and King

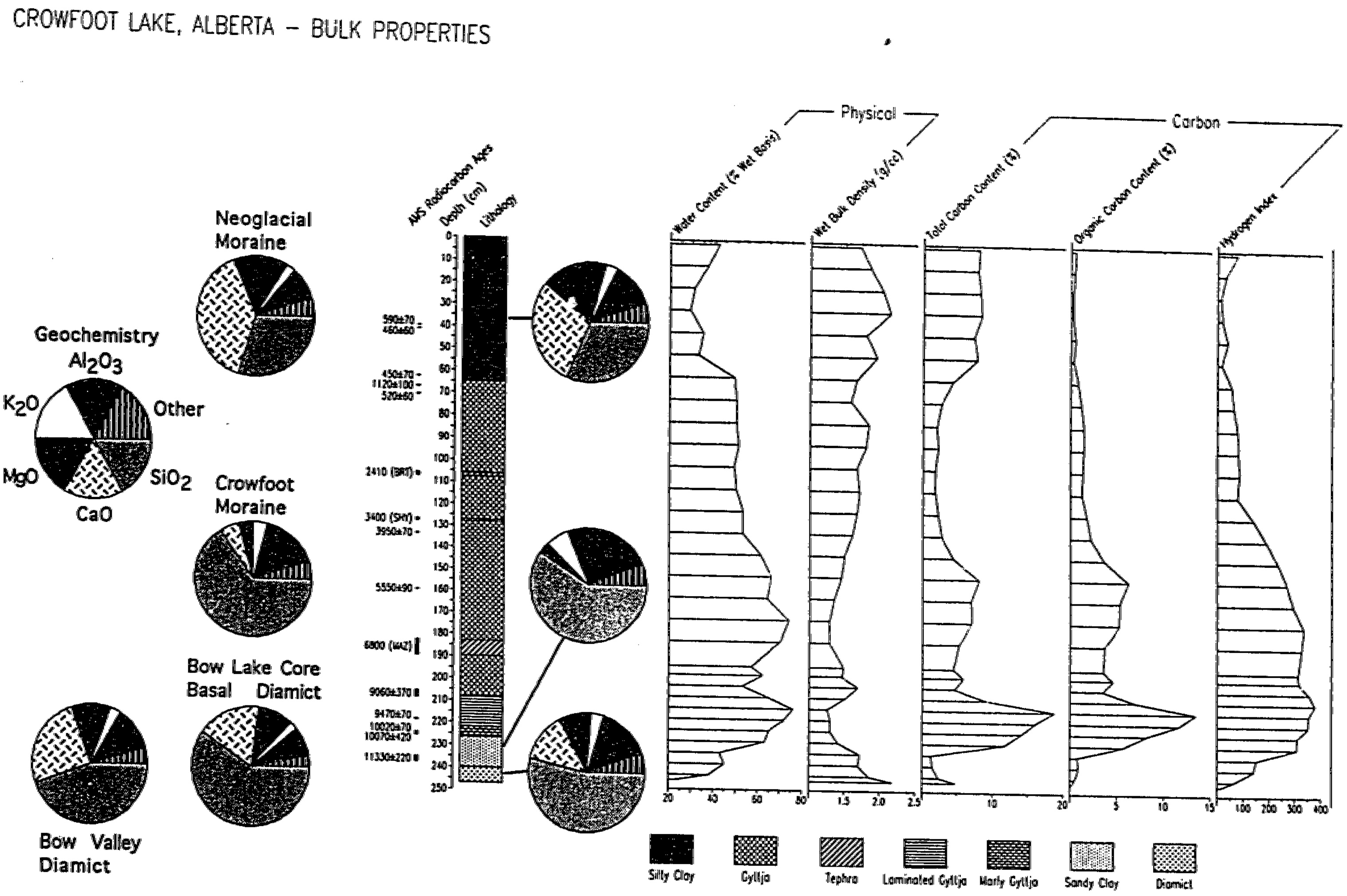

Figure 6-3. Physical and chemical bulk properties of the CRW6 core and related deposits: after Reasoner et al. (1994).. BRT - Bridge River Tephra, SHY - St. Helen's Yn Tephra, MAZ - Mazama tephra.

Table 6-1. Radiocarbon dates from Crowfoot Lake and Lake O'Hara cores CRW6 and LOH14.

Lab Number	*Core Depth*	*^{14}C Date*	*Dated Material*	*Calibrated age (cal yr BP)*
Core CRW6				
CAMS3840	40 cm	590±70	*Picea* needle	648 (552^b) 529
CAMS25429	43 cm	460±60	*Pinus* needle	530 (509^b) 477
CAMS6842	63 cm	450±70	*Picea* needle	530 (506^e) 467
CAMS25430	67.5 cm	1120±100	*Picea* needle	1161 (1052, 1044^a, 989) 993
CAMS4405	71 cm	520±60	*Picea* needle	550 (529^e) 509
CAMS3843	134 cm	3950±70	*Picea* needle	4508 (4410^a) 4287
CAMS4406	160 cm	5550±90	*Pinus* needle	6414 (6309^a) 6283
CAMS3064	205-208 cm	9060±370	*Abies* needle	10,388 ($10{,}007^e$) 9594
CAMS6843	218 cm	9470±70	*Pinus* needle	10,795 ($10{,}472^a$) 10,367
CAMS3063	225.5 cm	10,020±70	*Pinus* needle	11,721 ($11{,}321^c$, $11{,}301^c$, $11{,}213^c$, $11{,}174^c$, $11{,}126^c$) 11,009
CAMS3177	226 cm	10,070±420	Cyperaceae seed	12,418 (11,557, 11,357, 11,343) 10,900
CAMS3065	235-238 cm	11,330±220	*Salix* (?) twig	13,486 ($13{,}238^a$) 13,016
Core LOH14				
RIDDL433	87cm	10,100±200	*Abies* needle	12,169 ($11{,}687^d$) 11,004

Calibrated ages are based on Stuiver and Reimer (1993). ^{a} Calibrated ages used in age-vs-depth curve (Figure 6-4). ^{b}Average of two separate calibrated ages used for age-vs-depth curve. ^{c} Possible calibrated ages averaged and used to extrapolate an age of 11,600 BP for sediment contact at 227 cm in core CRW6. ^{d} Age of sediment contact in core LOH14 (Reasoner and Rutter 1988). ^{e} Possibly contaminated, age not used in age-vs-depth model.

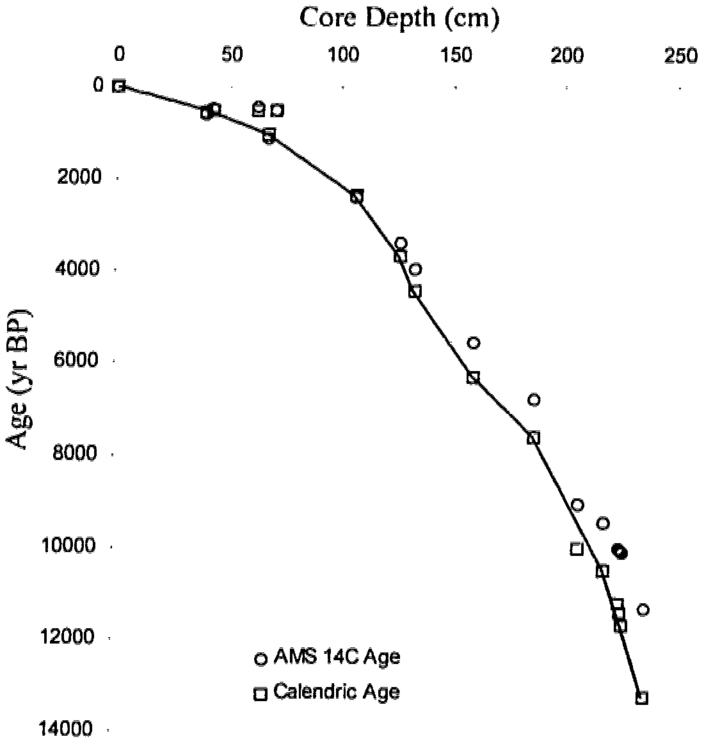

Figure 6-4. Age-vs-depth curve for core CRW6. Solid line connects the 11 calibrated radiocarbon ages used in the age-vs-depth model.

1986), and Bridge River (ca. 2360 ^{14}C yr BP; Clague et al. 1995, Leonard 1996). With the exception of the 450 ± 70, 520 ± 60, and 9060 ± 370 ^{14}C yr BP ages, the radiocarbon ages increase in a regular fashion with core depth, and are consistent with the positions of well dated Holocene tephras in the core (Figure 6-4). The age of 9060 ^{14}C BP is considered an outlier from the age-vs-depth curve and is excluded from the chronological model. The three possible calendric ages for the 10,020 ± 70 ^{14}C yr BP age (225 cm) were averaged, and used to extrapolate an age of ca. 11,600 cal yr BP for the sedimentary contact at 226.5 cm. A similar sedimentary contact in the record from nearby Lake O'Hara (Reasoner and Rutter 1988) has been dated to 10,100 ^{14}C yr BP (11,678 cal yr BP). The two ages of 450 ± 70 and 520 ± 60 ^{14}C yr BP fall off the age-vs-depth curve between the core top and Bridge River Tephra (Figure 6-4). If accepted as correct, these ages imply a dramatic change in deposition rates in the upper portion of the record that results in inexplicable pollen accumulation rates. In addition, the acceptance of the above ages constrain the transition to glacigenic sedimentation in the record to ca. 460 cal yr BP which is inconsistent with other records from the area (Luckman 1995, Leonard in press). Consequently, the ages of 450 ± 70 and 520 ± 60 ^{14}C yr BP are excluded from the Crowfoot chronology, and the late Holocene age-vs-depth relations are based on Bridge River Tephra and the ages of 460 ± 60, 590 ± 70, and 1120 ± 100 ^{14}C yr BP (Figure 6-4). Either bioturbation or contamination during sampling procedures may have lead to radiocarbon ages that are apparently younger than the deposition age of the enclosing sediment.

Crowfoot Lake Paleobotanical Data

Pollen percentage, pollen accumulation, and macrofossil data are used to reconstruct the vegetation history of the Crowfoot Lake area. Figure 6-5 is a relative percent pollen diagram that shows down-core variations of the major taxa. The relative pollen percentages were recalculated with *Pinus* excluded from the pollen sum (Figure 6-6) to show enhanced percentage variations of the less common taxa. A stratigraphically constrained cluster analysis (CONISS; Grimm 1988) subdivides this record into pollen assemblage zones. The zone divisions are located at first, second, and third order CONISS dissimilarity boundaries. A high lake-wide consistency in pollen percentage data has been demonstrated in records from Lake O'Hara (Beaudoin and Reasoner 1992), and consequently, inferences of vegetation history were primarily based on changes in the relative percent pollen records and supplemented by pollen accumulation rate (Figure 6-7) and macrofossil (Figure 6-8) information.

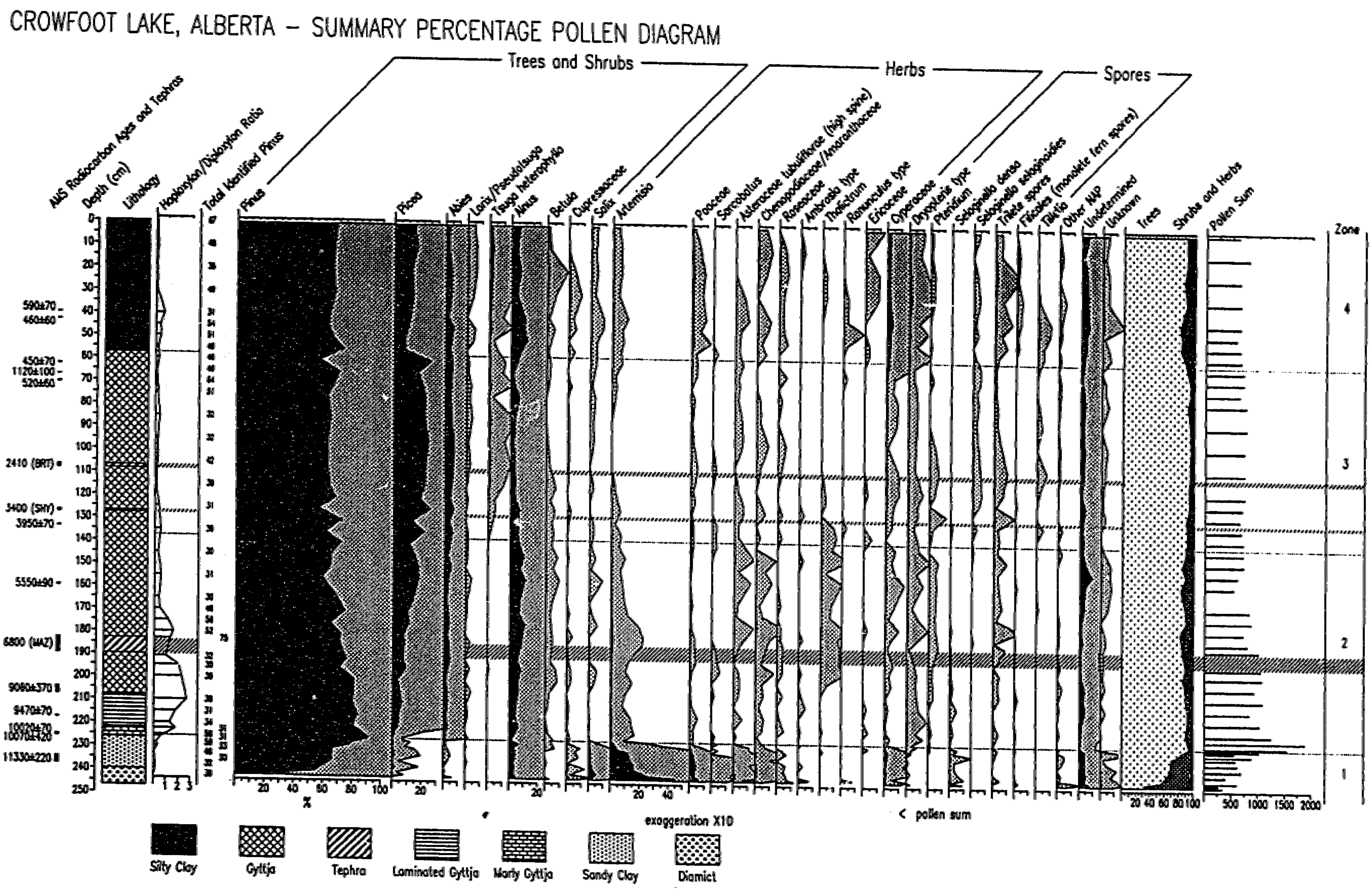

Figure 6-5. Summary pollen percentage diagram for the CRW6 core. BRT - Bridge River Tephra, SHY - St. Helen's Yn Tephra, MAZ - Mazama tephra.

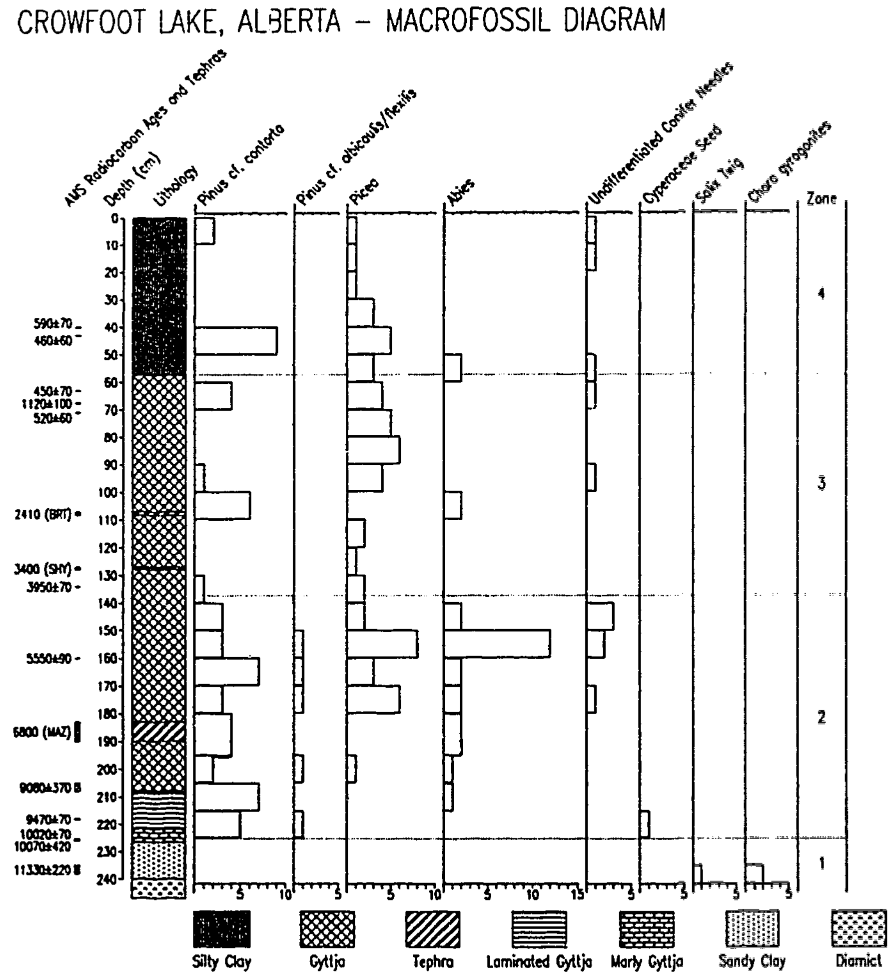

Figure 6-8. Macrofossil diagram for the CRW6 core. Bars represent the number of whole and macrofossil fragments recovered from the interval of core represented by the thickness of each bar.. BRT - Bridge River Tephra, SHY - St. Helen's Yn Tephra, MAZ - Mazama tephra.

CRW6 Zone 1, 245-225.5 cm. **Artemisia-Alnus-Poaceae-Salix** *assemblage zone.*

The base of Zone 1 predates ca. 11,330 ^{14}C yr BP and the top of the zone terminates at ca. 10,100 ^{14}C yr BP. The sub-zones within Zone 1 are defined by second and third order CONISS dissimilarity partitions. The dominant features of Zone 1 are increasing *Pinus* pollen percentages, decreasing *Artemisia* pollen percentages, and the presence of several NAP taxa (Poaceae, Asteraceae/Tubuliflorae, Chenopodiaceae/Amaranthaceae, *Sarcobatus*, *Salix*, Cupressaceae, and Cyperaceae) in percentages between 1% and 5%. Haploxylon/diploxylon ratios indicate that pine pollen is primarily diploxylon type. Also present in trace amounts throughout Zone 1 are *Picea*, *Abies*, *Betula*, *Shepherdia canadensis*, Caryophyllaceae, Rosaceae, *Potentilia*, *Ambrosia*, *Dryopteris*-type, *Equisetum*, and *Selaginella densa*. Average *Pinus* pollen percentages increase from 30% in Zone 1a, to 68% in Zone 1b, to 89% in Zone 1c. In contrast, average *Artemisia* pollen percentages decline from 46% in Zone 1a, to 13% in Zone 1b, to 3% in Zone 1c. *Alnus* pollen is consistently present in Zone 1 at about 4.5%. Taxa that occur in trace amounts in Zone 1a include *Fraxinus*, *Franseria*-type, Asteraceae/Tubuliflorae low-spine, Saxifragaceae, *Lithocarpus*-type, *Cornus stolonifera*, and Umbelliferae. Trace taxa in Zone 1b include *Ephedra*, *Quercus*, *Populus*, *Franseria*-type, Tubuliflorae low-spine, Umbelliferae, *Campanula*, *Elaeagnus commutata*, *Lycopodium*, and *Arceuthobium*. Trace taxa in Zone 1c include *Androsace*, *Labiatae*, Liguliflorae, *Rumex*-type, *Thalictrum*, Ericaceae, *Lycopodium*, and *Pteridium*. Also, *Betula*, *Shepherdia canadensis*, Cyperaceae, and *Dryopteris*-type grains occur in slightly higher percentages in Zone 1c than in the lower two sub-zones. Total PAR are at a minimum in Zone 1 but increase sharply over subzone 1c primarily in response to increasing accumulation rates of *Pinus* and *Alnus*. *Pinus*, *Artemisia*, *Alnus*, *Salix*, and Poaceae show the highest PAR in subzones 1a and 1b. Macrofossils are absent from Zone 1 with the exception of a *Salix* twig and *Chara* gyrogonites.

CRW6 Zone 2, 225.5-137.5 cm. **Picea-Alnus-Abies-Artemisia** *assemblage zone.*

Zone 2 dates from 10,100 ^{14}C yr BP to an interpolated age of ca. 4160 ^{14}C yr BP. The base of Zone 2 is a first-order CONISS dissimilarity partition. Sharp increases of *Pinus*, *Picea*, *Abies*, and *Alnus* pollen percentages, and declines in *Artemisia*, Poaceae, Asteraceae/Tubuliflorae high-spine, Chenopodiaceae/Amaranthaceae, *Sarcobatus*, *Salix*, Cupressaceae, and Cyperaceae pollen percentages, occur at or near the base of this zone. Zone 2 is characterized by steady increases in *Picea* pollen percentages (< 2% to about 18%) and decreases in *Pinus* pollen percentages (> 90% to about 65%). Haploxylon pine pollen is dominant throughout the zone. *Abies* pollen occurs consistently throughout the zone at 2-6%, Alnus at 3-11%, and *Betula*, Poaceae, Asteraceae/Tubuliflorae high-spine, and Chenopodiaceae/Amaranthaceae at <1%. Both *Artemisia* and Cyperaceae pollen also occur throughout the zone at <1-3%, however *Artemisia* declines slightly and Cyperaceae increases slightly over the upper portion of the zone. *Larix/Pseudotsuga*, *Salix*, Rosaceae, and Ericaceae pollen and *Dryopteris*, *Athyrium* and *Pteridium* spores are present throughout the zone in trace amounts. Trace amounts of *Shepherdia canadensis*, Cupressaceae, *Sarcobatus*, and *Ambrosia* pollen and *Selaginella densa* and *Lycopodium* spores occur in the lower half, whereas *Valerian*, *Rumex*-type, and *Sambucus* pollen are present in the upper half of the zone. Total PAR vary considerably over Zone 2 from 1000 to 5000 grains /cm^2/yr and are primarily represented by *Pinus*, *Picea*, *Abies*, and *Alnus*. A number of herbaceous taxa show a peak in PAR in mid-zone above Mazama Tephra. *Pinus* cf. *contorta* needles are common in Zone 2 and *Pinus* cf. *albicaulis/flexilis* needles are exclusive to this zone. *Picea* and *Abies* needles are more abundant in the post Mazama portion of Zone 2.

CRW6 Zone 3, 137.5-57.5 cm. **Picea-Abies-Alnus-Tsuga** ***assemblage zone.***

Zone 3 spans the interval from ca. 4160 - 900 ^{14}C yr BP. The lower boundary of Zone 3 is a second-order CONISS dissimilarity partition. *Pinus*, *Picea*, and *Abies* pollen occur in uniform percentages of about 65%, 20%, and 6% respectively throughout Zone 3. Haploxylon/diploxylon ratios record a shift to predominately diploxylon pine pollen near the base of the zone. The base of the zone is also marked by the appearance in low but consistent percentages of *Tsuga heterophylla* pollen and *Selaginella selaginoides* spores. *Thalictrum* pollen is present in low percentages only in the lower portion of Zone 3. *Pteridium* spores are present in low percentages in the lower two thirds of the zone. *Alnus* pollen declines from about 7% near the base, to 3-5% in the middle of the zone and rises to 8% at the top of the zone. *Artemisia* pollen is present at about 0.7% near the base and top of Zone 3 and is absent from several levels in the middle of the zone. Similarly, *Betula* pollen is present in small percentages throughout the zone with the exception of its absence in three levels in the zone's upper third. Cyperaceae and Poaceae pollen are present between <1% and 2% throughout Zone 3 and *Larix/Pseudotsuga*, Cupressaceae, *Salix*, Asteraceae/Tubuliflorae high-spine, Chenopodiaceae/Amaranthaceae, and Rosaceae pollen and *Dryopteris* and *Athyrium* spores are consistently present in trace amounts. *Sarcobatus*, *Franseria*, *Ambrosia*, *Cornus canadensis*, *Valeriana*, *Sambucus*, *Ranunculus*-type, and Ericaceae pollen and *Lycopodium*, *Sphagnum* and *Tilletia* spores occur sporadically throughout the zone in trace amounts. Total PAR are consistently high over the upper portion of Zone 3 primarily due to high accumulation rates of *Pinus*, *Picea*, and *Abies*. PAR of a number of shrubs and herbs are low in this zone. *Picea* needles are common in this zone with sporadic occurrences of *Pinus* cf. *contorta* needles and a single *Abies* needle present in mid-zone.

CRW6 Zone 4, 57.5-0 cm. **Picea-Alnus-Abies-Cyperaceae** ***assemblage zone.***

Zone 4 postdates ca. 900 ^{14}C yr BP. The base of Zone 4 is a third-order CONISS dissimilarity partition. The most striking feature of the zone is a sharp decline in *Picea* pollen from over 26% to 8% across the lower zone boundary. This decline in *Picea* pollen is accompanied by increases in *Alnus*, *Betula*, Cupressaceae, *Salix*, *Artemisia*, Poaceae, Chenopodiaceae/Amaranthaceae, *Ranunculus*-type, Ericaceae, and Cyperaceae pollen and *Dryopteris* and *Tilletia* spores. Asteraceae/Tubuliflorae high-spine and *Ranunculus*-type pollen and spores of *Tilletia* are present in low percentages in only the lower half of the zone. *Pinus* pollen percentages do not vary significantly from the previous zone and haploxylon/diploxylon ratios remain low throughout the zone. *Alnus* pollen percentages decline slightly from 5-6% near the lower zone boundary to about 4% by mid zone and remain stable thereafter. *Tsuga heterophylla* pollen increases steadily from 0.4% to 3% and *Abies* pollen declines slightly over the zone. Rosaceae pollen and spores of *Athyrium*, *Pteridium*, and *Selaginella selaginoides* occur in uniform, low percentages in the zone. *Shepherdia canadensis*, *Sarcobatus*, *Populus*, *Franseria*-type, *Potentilla*, *Ambrosia*-type, *Rumex*-type, *Cornus canadensis*, Cruciferae, Leguminosae, *Sambucus*, *Dodecatheon*, *Heracleum*, *Potamogeton* and *Arceuthobium* pollen as well as spores of *Botrychium*, *Lycopodium*, and *Sphagnum* occur sporadically in trace amounts throughout the zone. The PAR of *Pinus*, *Picea*, and *Abies* decrease sharply in the lower portion of Zone 4 and generally remain at levels below the previous zone. Several shrubs and herbs including *Betula*, *Salix*, *Artemisia*, Poaceae, Ericaceae, and Cyperaceae show increases in PAR from levels in Zone 3. *Picea* needles are abundant in the lower half of Zone 4 and subsequently decline in numbers. *Pinus* cf. *contorta* needles occur sporadically in the zone with a peak abundance at 40 - 50 cm.

DISCUSSION

Glacial History

The Crowfoot Lake record provides the oldest limiting age for the deglaciation of the Upper Bow Valley at a point approximately 10 km downvalley from the Continental Divide. Further, the record provides evidence for the timing of both the Crowfoot Advance at the type moraine, and the subsequent Late Holocene "Little Ice Age" advance.

Identifying the source of the basal diamict in the Crowfoot Lake sediment cores is key to understanding the stratigraphy of the record. These sediments may have been deposited by Bow Valley ice or, alternatively, may represent mass movements that originated from the Crowfoot moraine (Leonard 1986b). Hector Formation argillites outcrop extensively in the floor of the Bow Valley, however, the Crowfoot moraine just reaches the contact between the Gog and Hector Formations (Figure 6-1), and accordingly, clasts of Hector Formation argillite are absent from samples of the Crowfoot moraine. In contrast, 59% of the clasts in the basal diamict of the Crowfoot Lake cores are Hector Formation argillites. Further, the bulk composition of the basal diamict in the Crowfoot cores closely matches that of diamict samples taken from a nearby highway section located on the opposite side of the valley and from the basal diamict in a core from Bow Lake (Figure 6-1, 6-3). The basal diamict in the Crowfoot sediment cores therefore did not originate from the Crowfoot moraine but rather represents deposition from the retreat of the last trunk glacier in the Bow Valley. The lowest radiocarbon age in the Crowfoot record suggests that this basal diamict was deposited by Bow Valley ice prior to ca. 11,330 ^{14}C yr BP.

Crosscutting geomorphic relations indicate that the Crowfoot Advance represents a distinct, regional advance of alpine glaciers that postdates the recession of ice from the trunk valleys of the Canadian Rocky Mountains (Luckman and Osborn, 1979). The presence of Mazama tephra on the type Crowfoot moraine confirms that the advance occurred prior to ca. 6800 ^{14}C yr BP. Marl-bearing gyttja that post-date ca. 10,100 ^{14}C yr BP and underlie Mazama tephra in Core CRW6 are of relatively high organic carbon content and show high hydrogen indices (Figure 6-3) which together rule out the possibility of a significant advance of the Crowfoot glacier during this interval. In addition, several palaeoenvironmental reconstructions from alpine sites in the region indicate that ca. 9000 - 7000 ^{14}C yr BP was a period of relative warmth associated with elevated alpine timberlines (Luckman 1988, Reasoner and Hickman 1989, Beaudoin and King 1990). Consequently, the only sediments in the Crowfoot Lake record that could possibly be associated with the deposition of the Crowfoot moraine are the sandy clays that are bracketed by radiocarbon ages of 10,020 ± 420, and 10,070 ± 70 directly above and 11,330 ± 220 near the base (Figure 6-3). The bulk composition of these sediments very closely matches the composition of the <42 µm fraction of the Crowfoot moraine (Figure 6-3). These results indicate that the timing of the Crowfoot Advance is approximately synchronous with the European Younger Dryas cold event. However, the extent of the Crowfoot Advance is comparable to the Late Holocene Cavell advance in the Canadian Cordillera (Luckman and Osborn, 1979), whereas Younger Dryas ice advances in Europe are significantly more extensive than Late Holocene advances.

Several recent studies have determined that the termination of the Younger Dryas cold event was abrupt (e.g., Dansgaard et al. 1989, Alley et al. 1993). The Crowfoot Lake core samples all show an abrupt sedimentary transition dated at ca. 10,100 ^{14}C yr BP that separates the inorganic sandy clays associated with the Crowfoot Advance from overlying organic, marly gyttja (Figure 6-3). A similar sharp contact was identified in sediment cores from Lake O'Hara which is located approximately 35 km south of Crowfoot Lake. Terrestrial macrofossils recovered from this sedimentary contact provided two AMS radiocarbon ages of 10,100 ± 200 (RIDDL-433) and 10,060 ± 160 (RIDDL-511) ^{14}C yr BP (Reasoner and Rutter, 1988). These sharp sedimentary transitions suggest that the termination of the Crowfoot Advance was a significant and abrupt climatic event at least two sites in western Canada. Unfortunately, a plateau in the calendric ^{14}C calibration curve at ca. 10,000 ^{14}C yr BP (e.g. Bard et al. 1993) hinders accurate age determinations for these sedimentary transitions. The range of possible calendric ages (Stuiver and Reimer 1993) derived from on, or directly above, this contact in cores from Crowfoot Lake and Lake O'Hara (10,020 -10,100 ^{14}C yr BP) span ca. 550 cal years (11,126 to 11,678 cal yr BP).

The onset of Neoglacial cooling in western Canada is poorly constrained although there is a general consensus that the transition from the Early Holocene period of relative warmth and aridity to cooler and wetter conditions was underway between 5000 and 3000 ^{14}C yr BP. Palaeobotanical records indicate that timberline in the Maligne Range of Jasper National Park was in decline between ca. 5200 - 4500 ^{14}C yr BP (Kearney and Luckman 1983b, Luckman and Kearney 1986). Gardner and Jones (1985) document an advance of the Boundary Glacier in Northern Banff National Park between 4000 and 3800 ^{14}C yr BP. Further, "kill ages" for sheared and detrital wood from glacier forefields in the Canadian Rockies (Luckman et al. 1993) indicate that the Neoglacial Peyto Advance occurred between ca. 3300 and 2800 ^{14}C yr BP. Advanced ice positions have also been documented between ca. 3300 and 2000 ^{14}C yr BP in the Purcell and Coast Mountains of British Columbia (Osborn and Karlstrom 1988, Ryder and Thomson 1986). A ca. 4500 yr varve record from Hector Lake (Leonard, in press) records a significant increase in varve thickness from ca. 3500 to 3000 cal yr BP (approximately 3320 to 2860 ^{14}C yr BP). Glaciogenic sediments are absent from the Mid Holocene portion of the Crowfoot record. However, the decrease in organic carbon content from ca. 4500 to 3600 ^{14}C yr BP in core CRW6 likely reflects the onset of Neoglacial cooling.

A sharp decline in organic carbon and water content is coincident with increases in inorganic carbon content and wet bulk density between approximately 65 and 50 cm in core CRW6. These upper inorganic sediments are identical in bulk composition to samples of the "Little Ice Age" moraine (Figure 6-3) and thus represent a significant advance of the Crowfoot glacier in the Crowfoot Lake drainage. Although this sedimentary transition is abrupt, the exact timing of the transition is complicated by reversals involving the five uppermost radiocarbon ages (Figure 6-4). Accepting the 460 ± 60, 590 ± 70, and 1120 ± 100 ^{14}C yr BP ages and rejecting the two outlying ages (450 ± 70 and 520 ± 60 ^{14}C yr BP) from the age-vs-depth curve (Figure 6-4) implies the onset of Late Holocene glaciogenic sedimentation in Crowfoot Lake occurred between ca. 1060 and 740 ^{14}C yr BP (1000-1280 AD). Although the main phase of "Little Ice Age" moraine building culminated in the 18th and 19th centuries, evidence for a less extensive advance between 1140 and 1370 AD has been reported by Luckman (1995, in press). Osborn (1993) documents an advance of the Stutfield glacier at ca. 940 ^{14}C yr BP (1040 -1150 AD) and Clague and Mathewes (1992) suggest an ice advance in the Coast Mountains of British Columbia may have commenced as early as ca. 1000 AD. Finally, an interval of increased varve thickness suggests the presence of more extensive ice in the Hector Lake drainage between ca. 1240 and 1370 AD (Leonard, in press).

A comparison of the sedimentary records from core CRW6 and CRW8 (Figure 6-1) provides a qualitative evaluation of the sensitivity of Crowfoot Lake lacustrine records to upvalley glacial activity. Meltwaters from the Crowfoot Advance entered both sub-basins of Crowfoot Lake, and consequently, the pre-Mazama sedimentary records of CRW6 and CRW8 are essentially identical. Streams breached the Crowfoot moraine such that meltwaters from subsequent glacial advances influenced only the southern sub-basin of the lake and did not directly enter the northern sub-basin (CRW8) which is partially isolated by a shallow sill. In contrast to the sedimentary record of core CRW6. the CRW8 record does not include an influx of inorganic sediment related to Late Holocene advances of the Crowfoot glacier. This suggests that these lacustrine records have faithfully recorded upvalley glacial activity as long as meltwaters have directly entered the basins.

Vegetation History

Late-glacial: >ca. 11,330 - ca. 10,100 ^{14}C yr BP.

The very low total pollen concentrations and accumulation rates that characterize Zone 1 and the presence of "long distance" pollen such as *Ephedra*, *Sarcobatus*, *Fraxinus*, and *Quercus* suggest that this zone represents an interval of very low local pollen production. The predominately clastic sediments in this portion of the record likely reflect a sparsely-vegetated, unstable landscape associated with recently deglaciated terrain in the upper Bow Valley and the deposition of the Crowfoot moraine along the western shore of Crowfoot Lake. A thin or discontinuous shrub - herb vegetation cover, dominated by *Artemisia*, *Salix*, and Poaceae, is inferred for the local vegetation during this interval of the record. Minor components of this community likely included Asteraceae/Tubuliflorae high spine. Caryophyllaceae, Cyperaceae, Rosaceae, Cupressaceae, *Selaginella densa* and the nitrogen-fixers *Shepherdia canadensis* and *Alnus*. Trees were probably absent from this community with the possible exception of *Pinus contorta* which may have been present in the vicinity of the lake in isolated stands. Macrofossils of *P. contorta* and other conifers are absent from this zone and the tendency of *P. contorta* pollen to be transported long distances and over-represented in modern pollen spectra from tundra environmentsis well documented (e.g., Faegri et al. 1989, Fall 1994). The presence of *P. contorta* has been inferred in the lower Bow Valley by ca. 11,200 ^{14}C yr BP (MacDonald and Cwynar 1985). Although modern analogues for this vegetation community are absent from the region, the pollen spectra from CRW6 Zone 1 bear some striking similarities to modern pollen percentages derived from alpine vegetation in the Wind River Mountains of Wyoming (Fall, 1994). As the lower three samples from the zone represent sediments that directly overlie the basal diamict, it is possible that the pollen spectra from Zone 1a includes a proportion of pollen grains that were released from retreating Bow Valley ice and do not reflect the earliest local vegetation in the vicinity of the site. Pollen Zone 1c appears to be transitionary to Zone 2 and may reflect increases in *Alnus* and *P. contorta* at or near the site.

This shrub and herb dominated community is similar in composition to the inferred early postglacial vegetation a number of sites in the Rocky Mountains and adjacent foothills (e.g., MacDonald 1982, 1987a, 1989, White and Mathewes 1986, Schweger et al. 1981, Mott and Jackson

1982, Reasoner and Hickman 1989, Beaudoin and King 1990, Kearney and Luckman 1983a). Palaeoclimatic interpretations for this early treeless vegetation community are varied due to several complicating factors. Many components have wide environmental tolerances and the community as a whole was unlikely to have been highly sensitive to climate change. Further, the pollen of several major components of the early pollen spectra are difficult to subdivide below major taxonomic levels (e.g., Poaceae, Chenopodiaceae/Amaranthaceae, Cyperaceae). Finally, the early vegetation community was probably very responsive to edaphic conditions and site availability (Stork 1963) as well as climate change. Various workers have suggested that the early treeless communities may have been seral in nature and convey little climatic information (e.g. Lichti-Federovich 1970, Beaudoin and King 1990). Relatively cool climatic conditions have been associated with an early treeless vegetation community in several records from northwestern North America (e.g. Mack et al. 1978, 1983, Fergusson and Hills 1985, MacDonald 1989) although the region received greater summer insolation than today during the late-glacial and Early Holocene (Verneker 1972, Kutzbach and Guetter 1986).

Recently, a number of disparate proxy studies have furnished evidence in support of cooler climatic conditions in western North America during the Younger Dryas Chron (Reasoner et al. 1994, Gosse et al. 1995, Menounos and Reasoner 1996, Engstrom et al. 1990, Mathewes et al. 1993, Peteet and Mann 1994, Hu et al. 1995, Patterson et al. 1995, Mann and Hamilton 1995). Of the Rocky Mountain paleobotanical records, only the Toboggan Lake (MacDonald 1989), Lake O'Hara (Reasoner and Hickman 1989), and Crowfoot Lake records have sufficient chronological control to clearly demonstrate that intervals of sediment were deposited during the Younger Dryas Chron. However, a number records from the area very likely contain sediments of equivalent antiquity (e.g., Beaudoin and King 1990, Kearney and Luckman 1983a, MacDonald 1982). In all cases, the pre-ca. 10,100 ^{14}C yr BP portions of these paleobotanical stratigraphies do not chronicle clear responses to an oscillation in climate. It appears that the Younger Dryas event resulted in relatively minor advances of alpine glaciers in western North America that shortly followed the deglaciation of trunk valleys (Reasoner et al. 1994, Gosse et al. 1995, Menounos and Reasoner 1996) and, consequently, sufficient time may not have been available following deglaciation for the local vegetation to respond to pre-Younger Dryas climatic conditions. In addition, due to the broad ecological ranges of the dominant taxa, this early shrub - herb community was unlikely to have been highly sensitive to climate changes associated with the Crowfoot Advance.

Early - Mid Holocene: ca. 10,100 - ca. 4160 ^{14}C yr BP.

The sharp increase in arboreal PAR and abrupt shift from clastic to organic sedimentation at the base of Zone 2 represents the stabilization of the landscape and the appearance of abundant forest vegetation in the vicinity of Crowfoot Lake. The presence of *P. contorta* and *P. albicaulis/flexilis* macrofossils in the early portion of the zone accompanied by higher haploxylon/diploxylon ratios suggest that both *Pinus* subgenera were present in the early forest. *Picea* and *Abies* were also present in the early forest, however, the PAR and macrofossil records suggest that these taxa became more abundant during the latter portion of the zone. Also present as elements of the understorey flora were Polypodiaceae including *Dryopteris, Athyrium, Pteridium*, as well as *Lycopodium* and *Selaginella densa*. Although the percentages of a number of shade-intolerant taxa such as *Artemisia*, Poaceae, Chenopodiaceae/Amaranthaceae, and *Salix* decline sharply at the base of Zone 2, PAR for these taxa remain relatively high or increase slightly from levels in Zone 1. The high PAR for these taxa indicate that the early *Pinus* dominated forest in the upper Bow Valley included persistent areas of open vegetation. The abundance of xerophytic taxa such as *Pinus albicaulis/flexilis, Artemisia, Selaginella densa*, and *Ambrosia*, particularly in the pre-Mazama portion of the record, suggest the conditions responsible for maintaining a partially open forest may have been drought-related.

The rapid expansion of forest vegetation was ubiquitous in the Rocky Mountains and adjacent foothills during the Early Holocene (MacDonald 1989, Vance 1986). *Picea* forests occupied a number of northern foothill locations at ca. 10,000 ^{14}C yr BP (MacDonald 1987a, White and Mathewes 1986, Schweger et al. 1981) whereas the early forests in the south-central mountains and foothills appear to have been *Pinus* dominated (Beaudoin and King 1990, Kearney and Luckman 1983, Reasoner and Hickman 1989). Forest development along the eastern slopes and in the extreme southern Rocky Mountains may have occurred somewhat later (MacDonald 1989, Hills et al. 1985). Temporal variations of forest expansion and differences in early Holocene forest composition are probably due to migrational lags (MacDonald and Cwynar 1985) and/or local environmental conditions near the forest - grassland ecotone (MacDonald 1989). Lacustrine records from Lake O'Hara indicate that *Pinus* dominated forest vegetation developed rapidly along the Continental Divide ca. 10,100 ^{14}C yr BP and coincided with the abrupt onset of organic sedimentation (Reasoner and Hickman, 1989, Reasoner and Rutter 1988).

Although a number of studies in the central Rocky Mountains record the presence of forests at higher-than-present elevations during the Early Holocene (Kearney and Luckman 1983b, Luckman and Kearney 1986, Beaudoin and King 1990, Beaudoin 1986, Reasoner and Hickman 1989, Luckman 1988), only two continuous pollen records from alpine sites furnish evidence for the timing of forest expansion to elevations above modern timberline. Both the Wilcox Pass (Beaudoin and King 1990) and the Opabin Lake (Reasoner and Hickman 1989) records demonstrate striking similarities in the lower two pollen zones, however, temporal constraint in these records suggests the zones were not coeval. The record from alpine Opabin Lake implies that the early *Pinus*-dominated forest did not reach elevations above modern timberline until ca. 8500 ^{14}C yr BP whereas the Wilcox Pass record indicates the presence of a local *Pinus* dominated forest prior to 9600 ^{14}C yr BP. Unfortunately, radiocarbon control is sparse in both records and potential "old carbon" errors cannot be ruled out for the Wilcox Pass chronology. In addition, dramatic fluctuations in Opabin Lake water levels observed in recent years suggest that intervals of non-deposition or erosion may be present in the Opabin Lake record.

The presence of forest vegetation at sites above the elevation of modern timberline implies warm summer conditions prevailed in the region during the early Holocene. Palaeolimnological records from the western Alberta plains (Schweger and Hickman 1989) and paleobotanical records from along the subalpine - grassland ecotone in the foothills (MacDonald 1989) indicate that aridity reached a maximum during the early Holocene. Both increased aridity and higher summer temperatures during this interval are consistent with higher summer insolation associated with orbital configurations (Guetter 1986, COHMAP 1988).

Mid - Late Holocene: ca. 4160 - ca. 900 ^{14}C yr BP.

Zone 3 represents a period of closed coniferous forest in the vicinity of Crowfoot Lake. Pollen percentages, PAR, and macrofossil abundance indicate that *Picea* was abundant in the local forest during Zone 3. Moreover, the PAR of all three arboreal taxa are consistently high, particularly in the latter half of this interval. In contrast, pollen percentages and PAR are generally low for a number of shade-intolerant shrubs and herbs such as *Artemisia*, Cupressaceae, Chenopodiaceae/Amaranthaceae, Poaceae, and *Salix*. The absence of *P. albicaulis/flexilis* macrofossils and relatively low haploxylon/diploxylon ratios suggest that *P. contorta* was the dominant *Pinus* taxa in the local forest during this period. Declines in xerophytic taxa at or prior to the base of this interval coupled with the consistent presence of *Selaginella selaginoides* imply that a local reduction in effective evaporation occurred during this period. As *Tsuga heterophylla* is not presently, and was unlikely to have been, a component of the local forest, the consistent presence of its pollen in this portion of the record probably reflects a mid to late Holocene eastward expansion of this and perhaps other mesophytic taxa from coastal areas (Baker 1983, Cwynar 1993).

In general, records from the Canadian Rocky Mountains indicate that modern forest compositions were achieved in subalpine vegetation zones by approximately 5000 - 3000 ^{14}C yr BP (Luckman and Kearney 1986, Kearney and Luckman 1987, Reasoner and Hickman 1989, Beaudoin and King 1990). These records suggest that *Picea* had become well established in the mid Holocene forest and that the importance of *Pinus flexilis/albicaulis* had diminished in comparison to the early Holocene. Both *Pinus/Picea* ratios and conifer macrofossil abundance from alpine sites in the region indicate that the elevation of timberline was in decline during the period ca. 4000 to 1000 ^{14}C yr BP (Luckman and Kearney 1986, Beaudoin 1990, Reasoner and Hickman 1989). Alpine records from the Maligne Range (Luckman and Kearney 1986) suggest that the onset of this decline may have occurred shortly after the deposition of Mazama Tephra whereas the Wilcox Pass record (Beaudoin 1986) indicates rapid declines between ca. 2500 and 1000 ^{14}C yr BP.

A variety of independant proxy climate records from the Alberta plains, foothills and Rocky Mountains indicate that a shift to cooler and moister conditions occurred at approximately 4000 - 3000 ^{14}C yr BP. Palaeolimnological records from the central Alberta Plains, for example, indicate that present water levels had been attained by ca. 3000 ^{14}C yr BP (Schweger and Hickman 1989). Pollen records from sites throughout Alberta suggest that a decline in temperature concomitant with increasing regional precipitation began after 4000 ^{14}C BP (Vance 1986, Vance et al. 1995). A significant expansion of alpine glaciers designated as the Peyto Advance was underway by ca. 2500 and 3100 ^{14}C yr BP in the Rocky Mountains (Luckman et al. 1993). This advance is reflected in an increase in varve thickness from ca. 3500 to 3000 cal yr BP in the Hector Lake record (Leonard, in press) and this period of thicker varve deposition persisted until ca. 1800 cal yr BP. Finally, correlative zones in other Rocky Mountain pollen records show the consistent presence of *Tsuga heterophylla* and/or *Selaginella selaginoides* in low percentages (Beaudoin and King 1990, Reasoner and Hickman 1989, Luckman and Kearney 1986, Kearney and Luckman 1983a, Kearney and Luckman 1987); the former representing regional, and the latter reflecting local mesic conditions.

Late Holocene: ca. 900 ^{14}C yr BP - Present.

The PAR of all local arboreal taxa indicate a marked decline in forest density occurred at the base of Zone 4. This decline is also evident in the pollen percentages of *Picea* and to a lesser extent *Abies* although *Picea* macrofossils persist in relatively high numbers until mid-zone. Percentages and PAR of a number of taxa common to modern alpine tundra and subalpine meadow communities such as Cyperaceae, Ericaceae, *Salix*, *Artemisia*, Poaceae, and *Ranunculus*-type also increase in Zone 4. The decline in arboreal pollen taxa likely reflect both declining alpine timberline elevation and more extensive valley floor meadows in the Crowfoot Lake drainage. Consequently, more rigorous climatic conditions in the Crowfoot Lake drainage are inferred for this period. Coincident with this major change in the paleobotanical records is the abrupt influx of inorganic sediments that are associated with the "Little Ice Age" advance of the Crowfoot glacier (Figures. 6-1, 6-3).

Several other palynological records from the central Rockies record late Holocene vegetation changes that may be associated with the "Little Ice Age", however, confident correlations between records are generally hampered by poor chronological control. The Wilcox Pass (Beaudoin and King 1990), Opabin Lake (Reasoner and Hickman 1989), and the Maligne Pass (Kearney and Luckman 1987) records all suggest that declines in *Picea* and *Abies* were accompanied with expansions of Cyperaceae and one or more shade intolerant taxa during the last ca. 1500 ^{14}C years. Similar increases in Cyperaceae as well as several shrubs and herbaceous taxa are also evident in the Tonquin Pass and

Maligne Range records (Kearney and Luckman 1983a, Luckman and Kearney 1986). The increase in Cyperaceae in the uppermost zones of the Crowfoot and other records in the region (Beaudoin 1990, Kearney and Luckman 1987) likely reflect wetter conditions during this period. A growing body of dendrochronological and glacial studies in the region suggest that the onset cooler conditions associated with significant late Holocene glacier advances may have occurred during the 12th and 13th centuries (e.g. Luckman 1993, 1994a, b, 1995, 1996) and perhaps as early as ca. 1100 A.D. (Osborn 1993, Clague and Mathewes 1992).

SUMMARY

The Crowfoot Lake sediment core samples provide a >11,330 ^{14}C year continuous record of glacial events and vegetation change for the upper Bow Valley in the Canadian Rocky Mountains. Deglaciation of the Upper Bow Valley occurred before ca. 11,330 ^{14}C yr BP. Inorganic sediments associated with an advance of the Crowfoot glacier were deposited in the Crowfoot basin between 11,330 and 10,100 ^{14}C yr BP. The Crowfoot Advance was therefore approximately coeval with the European Younger Dryas cooling event. The maximum extent of the Crowfoot Advance is comparable to late Neoglacial ice advances in the Rocky Mountains whereas European Younger Dryas ice advances were significantly more extensive than late Neoglacial advances. The pre - 10,100 ^{14}C yr BP vegetation in the valley comprised a sparse shrub - herb community dominated by *Artemisia*, Poaceae and *Salix*. Paleobotanical evidence is lacking for any vegetation response associated with the onset of the Crowfoot Advance. It appears, however, that the Crowfoot Advance commenced shortly after deglaciation of the upper Bow Valley, and consequently, little time was available for the establishment of vegetation during this interval. Further, the early shrub - herb treeless vegetation was unlikely to have been highly responsive to climate changes associated with a relatively minor glacial advance. The abrupt onset of organic sedimentation in Crowfoot Lake and the rapid expansion of the pioneering *Pinus*-dominated forest in the area coincides approximately with the Younger Dryas termination at ca. 10,100 ^{14}C yr BP. This transition has also been well dated at Lake O'Hara (Reasoner and Hickman 1989), a subalpine site located approximately 35 km south of Crowfoot Lake. This early open *Pinus* forest included a number of subordinate xerophytic understorey components. A closed *Picea* - *Abies* forest dominates the record from ca. 4160 to ca. 900 ^{14}C yr BP. Declines in xerophytic taxa accompany the transition to modern subalpine forest compositions and increases in *Selaginella selaginoides* (local) and *Tsuga heterophylla* palynomorphs (regional) during this interval suggest increases in precipitation. The Crowfoot Lake paleobotanical records register sharp declines in local arboreal taxa at ca. 900 ^{14}C yr concomitant with the onset of late Neoglacial glaciogenic sedimentation in the basin. The uppermost zone of the pollen record likely reflects a decline in the elevation of alpine timberline in the drainage as well as an expansion of valley floor meadows that occurred in response to a combination of cold air drainage and moist valley-bottom conditions.

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CHAPTER 7

SUMMARY

This dissertation doucments aspects late Quaternary paleoenvironmental research that include both developents in paleoecological methodology as well as glacial and paleobotanical records derived from cored lacustrine sediments that were retrieved from subalpine and alpine sites in the Canadian and Colorado Rocky Mountains. Chapters 2 and 3 provide the methodological background for the findings reported in subsequent chapters. These two procedural chapters document equipment and technique improvements for a lightweight percussion coring system and accessed the degree to which differential pollen deposition and pollen focussing has occurred in the records from a subalpine lacustrine basin. Chapters 3 and 4 identify and radiocarbon date glaciogenic sediments associated with alpine glaciers in the Canadian and Colorado Rocky Mountains in order to determine the timing of late Pleistocene advances in both regions. Finally, Chapter 6 provides a detailed and well dated record of paleoenvironmental change for the upper Bow Valley, Banff National Park, based on sediment, pollen, and macrofossil analyses.

Methodology

Refinements concerning the equipment and operation of a lightweight, inexpensive, percussion sediment coring system (Reasoner, 1986) have resulted in substantial improvements to the ease of operation and both the number and length of sediment cores that may be recovered. The system has functioned well in water depths of approximately 200 m and is theoretically unrestricted by water depth. Sediment cores of up to 5.5 m in length have been recovered. In addition, as much as 50 cm of "basal" diamict have been retrieved in multiple cores from a single lacustrine basin. In these cases, it can be assumed with confidence that the cores represent the entire post-glacial record. The relatively large core barrel diameter (7.6 cm) provides ample material for multiple analyses and increases the likelihood of recovering terrestrial macrofossils for radiocarbon dating. The weight of the system equipped for 50 m water depth has been reduced to approximately 25 kg and the materials and equipment can be purchased for less than $600.00 (CDN). The system is ideal for operation in remote field settings because of its light weight and simplicity. It has been used successfully in the American and Canadian Cordillera, on Baffin and Ellesmere Islands in the Canadian Arctic, and at a number of locations throughout the Alberta Plains and Foothills.

A suite of 19 core samples were recovered from Lake O'Hara, Yoho National Park, to examine details of pollen deposition within a subalpine lacustrine basin in the Canadian Rocky Mountains. Each core contained two well dated Holocene tephras (Bridge River and Mazama tephras) and a radiocarbon dated sedimentary contact that could be correlated between cores throughout the basin. Consequently, pollen analysis of samples extracted from these discrete horizons provided a means of evaluating the degree to which differential pollen deposition or pollen focusing may have influenced the recruitment of pollen within the sediments. A fundamental assumption that is often made when reconstructing vegetation histories from pollen records is that pollen grains of different taxa are not differentially deposited, or focussed, within a lake basin, and therefore, a single core sample contains a pollen record that is representative of sediments within a lake basin as a whole.

The data from Lake O'Hara do not show strong evidence for either differential pollen deposition or pollen focussing at the three intervals examined. Specifically, there is no evidence for the preferential sorting and deposition of any individual taxon within the basin or significantly greater pollen accumulation at the deeper water sites based on pollen influx values. Therefore, the pollen record from a single core represents the pattern of pollen deposition in the lake as a whole. Further, the relative proportions (i.e., percentages) of major components of the pollen assemblages are consistent throughout the basin and a similar pollen record would have been obtained irrespective of core location. These findings support the Late Quaternary vegetation history inferred from a single core within the Lake O'Hara basin.

The major pollen taxa are remarkably consistent within a sample set but show significant differences between sample sets. This consistency at a chronostratigraphic level compared to variability between levels implies that qualitative interpretations of pollen records based on fluctuations of the major taxa are likely to be reliable, a conclusion also reached by Davis and Ford (1982). The distinction among the three sample sets, highlighted by DFA and MANOVA, and comparison between the sample sets and the equivalent zones from a single core from the basin, supports the validity of the zonation of the single record. However, this analysis shows that zonation will only be meaningful when based on major components of the pollen assemblage. The minor components (<1%) of the pollen assemblage are so variable that they are unreliable indicators of vegetation changes through a core. Therefore, the absence or disappearance of a minor taxon from the pollen assemblage cannot be used to infer its local absence in vegetation reconstructions.

Quantitative measures of pollen abundance, particularly pollen influx, are more variable than percentage data throughout the Lake O'Hara basin, because the values are strongly influenced by sediment fluxes. Greater chronological control (radiocarbon dates) would be required for each core in the Lake O'Hara basin to account for temporal and spatial variations in sediment deposition. Consequently, it may be generally concluded that detailed chronological control is required before any confidence can be placed in pollen influx data for the purpose of interpreting vegetation history from a single core.

Paleoenvironmental reconstructions

Since the originally mapping of the Crowfoot moraine system in the Canadian Cordillera (Luckman and Osborn, 1979), the age of the Crowfoot Advance has been a point of debate. Estimates for the timing of the advance have ranged from early Holocene to late Pleistocene in age (Leonard, 1986; Davis and Osborn, 1987; Osborn and Luckman, 1988), and prior to this study, the only firm chronological limitation was the presence of Mazama tephra (ca. 6800 ^{14}C yr BP) in soils overlying the type moraine at Crowfoot Lake, Alberta. The extent of the late Pleistocene Younger Dryas cold event was initially thought to have been restricted to the North Atlantic region (e.g. Broecker et al. 1988). However, an increasing number of investigations conducted during the last fifteen years have provided evidence that chronicled the influence of the event in widely separated areas of the globe (e.g. Chinzei et al.. 1987; Overpeck et al., 1989; Kudrass et al., 1991; Mathewes et al., 1993). Consequently, an exact determination of the age of the Crowfoot moraine was required in order to ascertain weather or not the Younger Dryas cold event had resulted in an advance of alpine glaciers in western Canada. A suite of nine lacustrine sediment cores was recovered from Crowfoot Lake in order to address the question of the age of the Crowfoot Advance. Crowfoot Lake provided an ideal opportunity to investigate the age of the Crowfoot Advance as the lake is situated directly downvalley from the type locality of the Crowfoot Moraine system.

The Crowfoot Lake cores show a consistent lakewide stratigraphy consisting of basal diamict overlain by silty clays which are in turn overlain by highly organic gyttja. The gyttja contains three Holocene tephras, Mazama, St. Helen's Yn, and Bridge River tephras, that have been dated at ca. 6800, 3400, and 2350 ^{14}C years BP respectively. Analyses of bulk sediment chemistry, clast lithology, as well as the basin-wide distribution of the basal diamict, indicate that the diamict was deposited from the retreat of Bow valley ice and did not originate from mass wasting of the Crowfoot moraine. As crosscutting geomorphic evidence dictate the Crowfoot Advance post-dated deglaciation of the upper Bow Valley, sediments related to the advance must overlie the basal diamict in the cores. The only pre-Mazama inorganic sediments in the record that could possibly be associated with the Crowfoot moraine are the silty clays that overlie the basal diamict. The bulk composition of these sediment are closely matched with the composition of the Crowfoot moraine and are distinct from other inorganic sediment in the record. The high organic carbon content of the pre-Mazama gyttja that overlies the silty clays precludes the possibility of a significant advance of the Crowfoot glacier during this interval. Accelerator mass spectrometry radiocarbon dating of macrofossils extracted from directly above, and from near the base, of the inorganic silty clays provide bracketing ages of ca. 10,100 and 11,330 ^{14}C years BP respectively. Consequently, the sediments in the Crowfoot Lake record associated with the deposition of the Crowfoot moraine were deposited during the Younger Dryas Chron (10,000 - 11,330 ^{14}C yrs BP). These findings illustrate that the climatic changes associated with the Younger Dryas cold event extended well beyond the North Atlantic region.

An ideal sequence of lacustrine sediments that represents the Younger Dryas oscillation should include sediments that post-date deglaciation and in some way reflect warmer conditions prior to the event. Lacustrine sediments that are correlative with the Crowfoot Moraine, however, directly overlie Bow Valley diamict with no intermediate sediments of higher organic carbon content. Apparently, sufficient time was not available for the establishment of organic sedimentation in Crowfoot Lake following deglaciation of the Upper Bow Valley and prior to onset of the Younger Dryas cooling. If this interpretation is correct, sediments representing the entire climate oscillation should be preserved in lake records situated further south where the final stages of deglaciation occurred somewhat earlier than in western Canada. Accordingly, an alpine cirque lake in the Colorado Rocky Mountains was cored in order to verify the interpretation of the Crowfoot Lake record.

Sky Pond is situated at an elevation of 3320 m in an alpine cirque in the Colorado Front Range. The setting of Sky Pond is similar to that of Crowfoot Lake in that it lies directly downvalley from a deeply weathered moraine that predates Neoglacial deposits. A second moraine occurs within the lake basin. Pedological and weathering features, as well as elevation and position within the cirque, suggest that the moraine directly above sky pond is equivalent to the inner Satanta Peak moraine. The Satanta Peak moraine pair is recognized throughout the Colorado Front Range (Benedict, 1973; 1981), and, like the Crowfoot Moraine, temporal constraint for this ice advance has been limited. A suite of five sediment cores were recovered from Sky Pond and two of these cores recovered up to 50 cm of basal diamict. An accelerator mass spectrometry age of 12,040±70 was obtained from directly above the diamict and agrees well with limiting ages for deglaciation of other cirques in the region (e.g. Harbor, 1985, Zielinski and Davis, 1987, Davis et al. 1992). The basal diamict is overlain by an interval of gyttja that is in turn overlain by inorganic clayey silt. The remainder of the record is represented by gyttja that is interrupted by thin, sharply bounded sandy beds. The inorganic clayey silts show lower organic carbon and water contents together with higher magnetic susceptibility and dry density values that the underlying and overlying gyttja. This interval of inorganic sediments are bracketed by AMS radiocarbon ages of 11,070±50 and 9970±80 ^{14}C yr BP. An additional AMS age of 10,410±90 ^{14}C yr BP was obtained from within the interval of clastic sediments. Hence, the interval of inorganic sedimentation in the Sky Pond basin occurred during the Younger Dryas Chron. Further, as lacustrine records of basins situated directly downvalley from glaciers faithfully record the activity of upvalley ice, the only clastic sediments in the record that are likely to be associated with the deposition of the moraine above Sky Pond are the inorganic clayey silts of Younger Dryas age. The similarities in soil development, weathering features, and altitude, between this moraine and the type Satanta Peak moraines suggest these moraines are correlative, and consequently, the latest Satanta Peak Advance was likely coeval with the European Younger Dryas event.

The identification of the entire Younger Dryas climatic oscillation in the Sky Pond sedimentary record supports the reconstruction of the late Pleistocene glacial history of the Upper Bow Valley in the central Canadian Rocky Mountains. The stratigraphy of the Crowfoot Lake cores suggests that the Crowfoot Advance occurred during the Younger Dryas Chron shortly after deglaciation of the Upper Bow Valley. It appears that sufficient time was not available for the establishment of organic sedimentation following deglaciation of the Crowfoot Lake basin and prior to the onset of glacigenic sedimentation associated with the Crowfoot Advance. If this sequence of events is correct, a record of the entire Younger Dryas oscillation should be preserved in lacustrine records from basins that were deglaciated earlier. Organic sediments accumulated in Sky Pond for approximately 1000 years after deglaciation and prior to the onset of clastic sedimentation associated with the Younger Dryas event.

The Crowfoot Lake and Sky Pond studies provide evidence that suggests alpine glaciers in the North American Rocky Mountains advanced during the Younger Dryas Chron. In the Canadian Rocky Mountains, the Crowfoot Advance similar in magnitude to late Neoglacial ice advances whereas in the Colorado Rockies, this late Pleistocene advance resulted in the deposition of moraines 1 - 3 km downvalley from late Neoglacial ice positions. In contrast, evidence from the Sierra Nevada of California demonstrate that cirque glaciers did not advance during the Younger Dryas Chron (Clark, 1994). The available evidence suggests that glacial responses to the Younger Dryas event in western North America may have had significant spatial variability. Additional documentation of the extent of alpine glacial advances during the Younger Dryas Chron, particularly in the American southwest, is essential for verifying the details of regional environmental response to climatic change associated with the Younger Dryas event.

A detailed Late Quaternary paleoenvironmental reconstruction for the Upper Bow Valley in Banff National Park, Canada is based on sediment, pollen, and macrofossil analyses of a sediment core recovered form Crowfoot Lake, Alberta. Chronological control for the record is based on a twelve AMS radiocarbon ages determined from terrestrial macrofossils, as well as the presence of three well dated Holocene tephras. The Crowfoot Lake core provide the longest and most detailed chronological record of vegetation history and paleoenvironmental change in the Canadian Rocky Mountains. Further, the location of the Crowfoot site near the alpine timberline ecotone, and both late Pleistocene and late Holocene moraines, has resulted in records that have been remarkably sensitive to past climatic changes. The Crowfoot Lake record registers local environmental conditions during the late Pleistocene Crowfoot Advance, the early to mid Holocene period of relative warmth and aridity, and the mid to late Holocene period of cooling and increased precipitation that includes a strong "Little Ice

Age" signal. The vegetation responses to past climate change tend to be well defined and abrupt and this is particularly evident for changes associated with the Termination of the Younger Dryas and the onset of the "Little Ice Age".

Pioneering vegetation was established in the upper Bow Valley by ca. 11,330 ^{14}C yr BP following the retreat of the trunk valley glacier. This early community was present during the Crowfoot Advance and persisted untill ca. 10,100 ^{14}C yr BP. The pioneering vegetation was a thin or discontinuous shrub - herb community dominated by *Artemisia*, Poaceae and *Salix*. Paleobotanical evidence is lacking for any vegetation response associated with the onset of the Crowfoot Advance. It appears, however, that little time was available for the establishment of vegetation between deglaciation of the upper Bow Valley and the onset of the Crowfoot Advance. Further, the early shrub - herb treeless vegetation was unlikely to have been highly sensitive to climate changes associated with a relatively minor glacial advance. The abrupt onset of organic sedimentation in Crowfoot Lake and the rapid expansion of the pioneering *Pinus* dominated forest in the area coincides approximately with the Younger Dryas termination at ca. 10,100 ^{14}C yr BP. This transition has also been well dated at Lake O'Hara (Reasoner and Hickman 1989), a subalpine site located approximately 35 km South of Crowfoot Lake. This early open *Pinus* forest included a number of xerophytic components including as *Pinus albicaulis/flexilis*, *Artemisia*, *Selaginella densa*, and *Ambrosia*. A closed *Picea - Abies* forest dominates the record from ca. 4160to ca. 900 ^{14}C yr BP. Declines in xerophytic taxa accompany the transition to modern subalpine forest compositions and increases in *Selaginella selaginoides* (local) and *Tsuga Heterophylla* (regional) during this interval suggest increases in precipitation. The Crowfoot Lake paleobotanical records indicate sharp decline in local arboreal taxa at ca. 900 ^{14}C yr concomitant with the onset of late Neoglacial glaciogenic sedimentation in the basin. The uppermost zone of the pollen record likely reflects a decline in the elevation of alpine timberline in the drainage as well as an expansion of valley floor meadows that occurred in response to a combination of cold air drainage and moist valley-bottom conditions.

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